Predictive Stratigraphic Analysis—Concept and Application

U.S. GEOLOGICAL SURVEY BULLETIN 2110

Cover. Calcic paleo-Vertisol underlying the resistant transgressive marine limestone Little Stone Gap Member of the Hinton Formation (Upper Mississippian) in southwestern West Virginia. This paleosol is indicative of a relatively dry climate when evapotranspiration exceeded rainfall for more than 6 months out of the year. The light-gray color at the level of the photograph scale (center) is the result of gleying (bleaching) after burial. A calcified root system, located in the proximity of the scale, branches downward and suggests a well-developed root system for a plant whose stem may have been up to 15 centimeters in diameter. Numerous mineralized fossil roots at this level indicate that land plants were very well adapted to seasonally dry conditions in nonwaterlogged environments by Late Mississippian time. Cross-cutting fractures, known as mukkara structures and caused by seasonal expansion (wet) and contraction (dry), are visible throughout the outcrop beneath the resistant limestone layer except where interrupted or destroyed by paleoroot systems.

Predictive Stratigraphic Analysis—Concept and Application

Edited by C. Blaine Cecil and N. Terence Edgar

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A collection of extended abstracts of papers presented at two workshops on the title subject

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PREFACE

Two workshops were held to develop a Predictive Stratigraphic Analysis (PSA) research initiative. The workshop cooperators were the U.S. Geological Survey, the West Virginia Geological Survey, Morgantown, West Virginia (April 22–24, 1991), and the Kansas Geological Survey, Lawrence, Kansas (June 24–26, 1991). This volume is a compilation of extended abstracts of each of the scheduled presentations.

Larry Woodfork, State Geologist and Director of the West Virginia Geological and Economic Survey, and Lee Gerhard, State Geologist and Director of the Kansas Geological Survey, set the tone of each meeting by discussing the potential benefit of the program to the scientific community, Federal and State agencies, and the public. The need was stressed for a continental-scale program through multidisciplinary research in order to evaluate and resolve eustatic and tectonic controls on transgressive-regressive cycles and to determine paleoclimatic latitudinal gradients, orography, and paleoclimatic cycles as controls on sediment flux including organic productivity. The well-known cyclothems of the Pennsylvanian System of North America were cited as one of the most ideal records with which to evaluate global change and the effects of climate, eustasy, and tectonics on the origin of sedimentary rocks in foreland basin (Appalachian), epeiric sea (midcontinent), deep sea (Ouachita), epicontinental shelf (western United States), and lacustrine basin (Appalachian) environments. The results of PSA research on the mid-Pennsylvanian rocks of North America will have application on a global scale and for older and younger intervals of time.

C. Blaine Cecil
N. Terence Edgar
Editors

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PREDICTIVE STRATIGRAPHIC ANALYSIS—CONCEPT AND APPLICATION

Edited by C. Blaine Cecil and N. Terence Edgar

Introduction to Predictive Stratigraphic Analysis

C. Blaine Cecil and N. Terence Edgar

The objective of the Predictive Stratigraphic Analysis project is to develop a methodology that (1) can distinguish and systematically evaluate and integrate the three allocyclic (changes in energy and materials external to the sedimentary system) factors that control sedimentation—tectonics, sea level, and climate, which (2) can then be integrated into a predictive stratigraphic model. Tectonic changes and eustasy can be evaluated through interbasinal correlation. By correlating among basins, the probability is high that tectonic effects, which generally are intrabasinal, can be distinguished from eustatic processes, which are interbasinal. The paleoclimatic influence varies primarily with paleolatitude and is modified by orbital forcing, paleo-orography, and other effects such as paleo-oceanography. The methodology is used in the study of sedimentation and stratigraphy and applied to modeling of energy and mineral resource occurrences.

Initially, the project focused on rocks of Carboniferous age (360 to 280 Ma) because climatic, tectonic, and glacio-eustatic changes are particularly well expressed in rocks of this age in sedimentary basins across the continent (Appalachian, midcontinent, Paradox, and Great basins), and major hydrocarbon deposits have been discovered in Carboniferous rocks of North America. A large data base has been created as a result of coal and petroleum exploration in these basins, and a detailed study by this project will contribute to a thorough understanding of the effects of climate on sedimentation, stratigraphy, and the distribution of energy resources in the Carboniferous System. In future years the methods and techniques developed in the initial study will be applied to strata of different ages, to other economic sedimentary minerals, and on a global scale.

Even though it was recognized as early as 1875 by Lyell, in his *Principles of Geology*, as one of the primary controls on stratigraphy, interpretation of the relationships between paleoclimate and sedimentation has generally been limited to sequences that contain strata such as evaporites, eolianites, tillites, and coal. Such interpretations are commonly used to infer paleolatitudes (Witzke, 1990). With notable exceptions (Huntington, 1907; Wanless and Shepard, 1936; Glennie, 1984; Perlmutter and Matthews, 1989), there appears to have been a lack of appreciation and understanding of the frequency and intensity of climate change and the resulting impact of such change on sedimentation and stratigraphy. In the last decade or two, excellent progress has been made on our understanding of the factors that control climate change. In addition, the ability to recognize the effect of paleoclimates on cyclic sedimentation and stratigraphy from lithologic and paleontologic climatic signatures also has been developed recently (Cecil and others, 1985; Perlmutter and Matthews, 1989; Cecil, 1990).

These advances have given us new insight into the first-order effect of climate on stratigraphy and sedimentation as a result of movement of continents through latitudinal climatic belts with time (Glennie, 1984; Schutter and Heckel, 1985; Perlmutter and Matthews, 1989). There are modifying effects of other factors such as orbital-forcing cycles, mountain building, ocean circulation, and variations in atmospheric "greenhouse" gases. We are now in a position to develop new geologic models, which integrate climate change with tectonic and eustatic controls on sedimentation and stratigraphy, that will improve our ability to predict stratigraphic sequences and evaluate their inclosed resources. More reliable predictive stratigraphic modeling appears to be possible because, to the degree that climate change is deterministic and therefore predictable, change in sediment flux to depocenters is also deterministic and therefore predictable.

Climatic effects on stratigraphy and resource occurrence and quality have been documented in Upper Pennsylvanian strata (Cecil and others, 1985). Climate is a major control on organic productivity; therefore, ancient climate setting can be used to predict the occurrence of conditions necessary for the formation of type III kerogen, the basic precursor to coal and certain petroleum deposits. The nature of potential petroleum reservoir rock (siliciclastic or carbonate) is also predictable when one has an accurate understanding of the paleoclimate. In the Appalachian basin, the driving mechanism for the origin of coal and lacustrine limestone beds appears to have been related to moisture changes of tropical climate in which peat formed during relatively wet intervals and limestone was deposited during drier periods (Cecil, 1990). Wet and dry climatic cycles are further documented by the types of paleosols, which developed laterally to precursors of beds of coal and limestone.

Coeval rocks in the western plains (Colorado, Nebraska, and Wyoming) were deposited in a dry belt at a paleolatitude of approximately 20°N. In contrast to the coal-bearing section of the eastern United States, the marine and continental strata deposited in that epicontinental dry belt are petroliferous but do not contain coal resources. In the midcontinent, which was in the transition zone between the relatively wet eastern United States and the arid west, changes of climate may have been a primary factor in the long-debated origin of the Pennsylvanian cyclothems. These strata are petroliferous, and they also contain mineable coal beds.

Climate also effects deposition of sedimentary mineral resources. Those minerals that are concentrated by weathering, erosion, and redeposition are mobilized under specific climatic conditions. For example, uranium is mobilized under oxidizing arid or semiarid conditions of weathering and then concentrated under reducing conditions in ground water systems. Therefore, an understanding of the climatic conditions at any given locality and time enables us to enhance prediction of conditions favorable for uranium occurrence.

An integrated multidisciplinary approach to the origin of sedimentary rocks is necessary for this study. Study elements include, but are not limited to, the following disciplines: biostratigraphy and paleobiology (both invertebrate paleontology and paleobotany), paleotectonics, paleomagnetism, paleogeography and paleo-oceanography, paleoclimatology, geochemistry (both organic and inorganic), paleopedology, sedimentary petrology (of both chemical and siliciclastic rocks), sedimentology, and stratigraphy. Because of the broad scope of the project in stratigraphic range, geographic area, and disciplines, selected intervals within Carboniferous strata of the United States and Canada serve as the initial bases for the development of models of sedimentary sequences and resource occurrence. The project interfaces with and complements similar projects in State geological agencies and other U.S. Geological Survey projects that involve sedimentary basin analysis, energy and mineral resources studies, global change (through modern analogue studies), and geologic mapping. Because tectonic, eustatic, and climatic controls on sedimentation and stratigraphy can rarely be distinguished by studying individual sedimentary basins, the results of this interbasinal study will have a profound effect on our understanding of sedimentary sequences and resources at local, regional, and continental scales. Cooperation among Federal and State agencies, industry, and academia is, therefore, appropriate, justified, and essential to the success of the project.

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Carboniferous Paleocontinental Reconstructions

Christopher R. Scotese

Presented are five paleocontinental reconstructions illustrating the collision of Laurussia (North America and Europe) with Gondwana during the Early and Late Carboniferous. Progressive closure of the Appalachian and Hercynian seaways resulted in the formation of the western half of Pangea by the Late Carboniferous to Early Permian. During this interval, the midcontinent of North America moved from the subtropics (20°S.) to a position astride the Late Carboniferous equator (DiVenere and Opdyke, 1990). The reconstructions presented here are based primarily on the paleomagnetic data base compiled by Van der Voo (1993), combined with paleomagnetic results from Siberia (Khramov and Rodionov, 1980). The orientation of Gondwana is modified after Scotese and Barrett (1990).

There is little disagreement among a variety of authors (Van der Voo and others, 1984; Rowley and others, 1985; Ziegler, P.A., 1989; Scotese and McKerrow, 1990; Witzke, 1990; Ziegler, A.M., 1990; Van der Voo, 1993) concerning the configuration of the continents during the Late Carboniferous and Early Permian. Paleomagnetic, paleoclimatic, and biogeographic evidence all indicate that North America straddled the equator in an orientation similar to that shown in figures 1 to 3.

There is, however, some debate concerning both the latitudinal orientation and relative positions of the major continental blocks during the Early Carboniferous. The distribution of paleoclimatic indicators, such as evaporites and reefs, led Witzke (1990) to suggest that the Early Carboniferous (Visean) equator passed through northern Greenland and British Columbia and that most of North America was located in the southern subtropics. A second group of authors (Ziegler, P.A., 1989; Kelley and others, 1990) favors a more northerly orientation of Pangea in which the equator runs across the midcontinent of North America. Kelley and others (1990) prefer this equatorial orientation because it best explains latitudinal patterns of brachiopod diversity. Though Ziegler, P.A. (1989) also showed North America astride the equator during the Early Carboniferous, his reconstruction is based on paleomagnetic data (Scotese and others, 1979) that have been superseded by recent summaries (Van der Voo, 1990, 1993).

The Early Carboniferous position for Pangea given here (figs. 4, 5) lies between these two orientations and is in agreement with new paleomagnetic results from Maritime Canada (DiVenere and Opdyke, 1990), assuming firm attachment to cratonic North America. If, however, there has been significant left-lateral motion (>500 km) between Maritime Canada and cratonic North America since the Namurian, then a more northerly position of cratonic North America might be indicated.

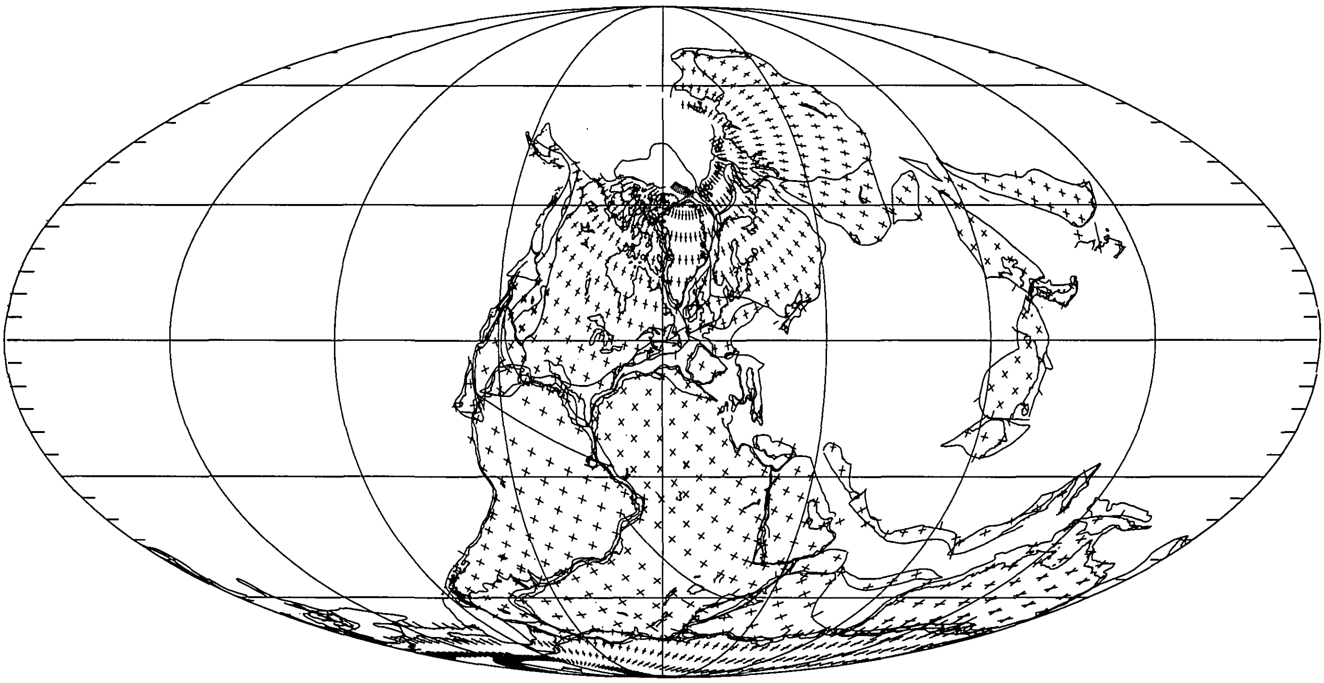

Figure 1. Paleocontinental reconstruction of Laurussia and Gondwana during the Stephanian.

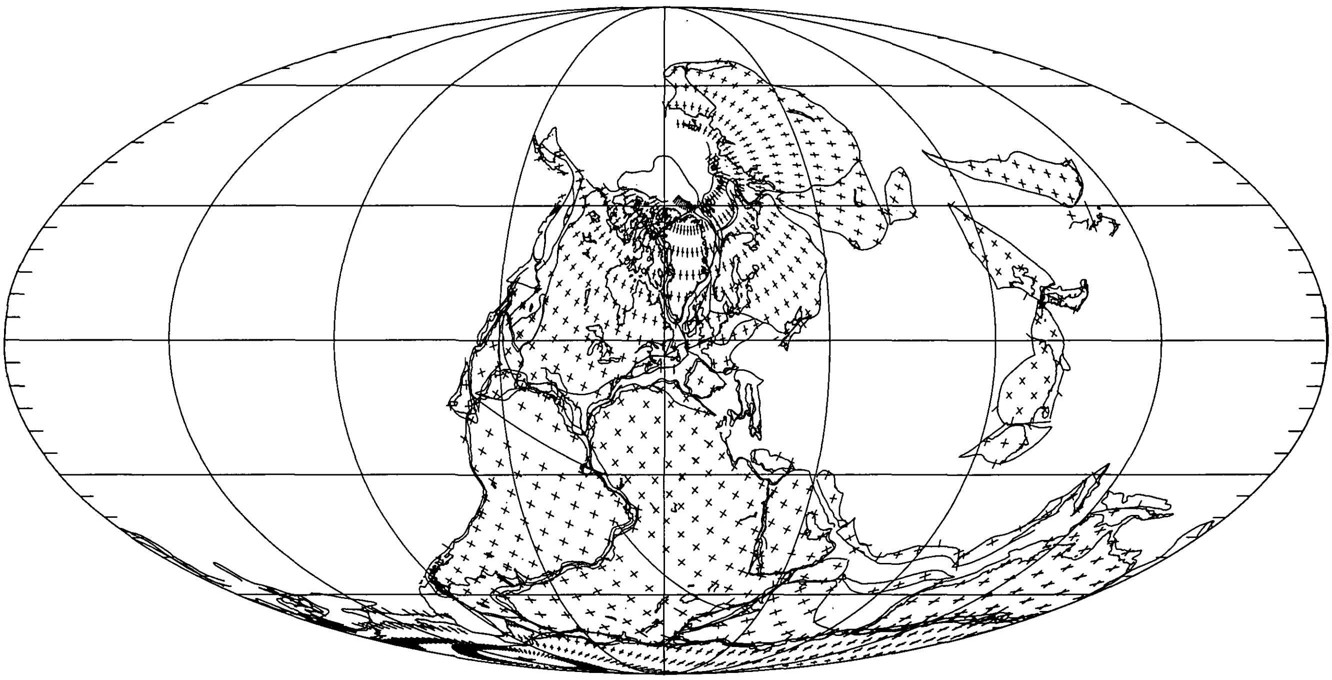

Figure 2. Paleocontinental reconstruction of Laurussia and Gondwana during the Westphalian.

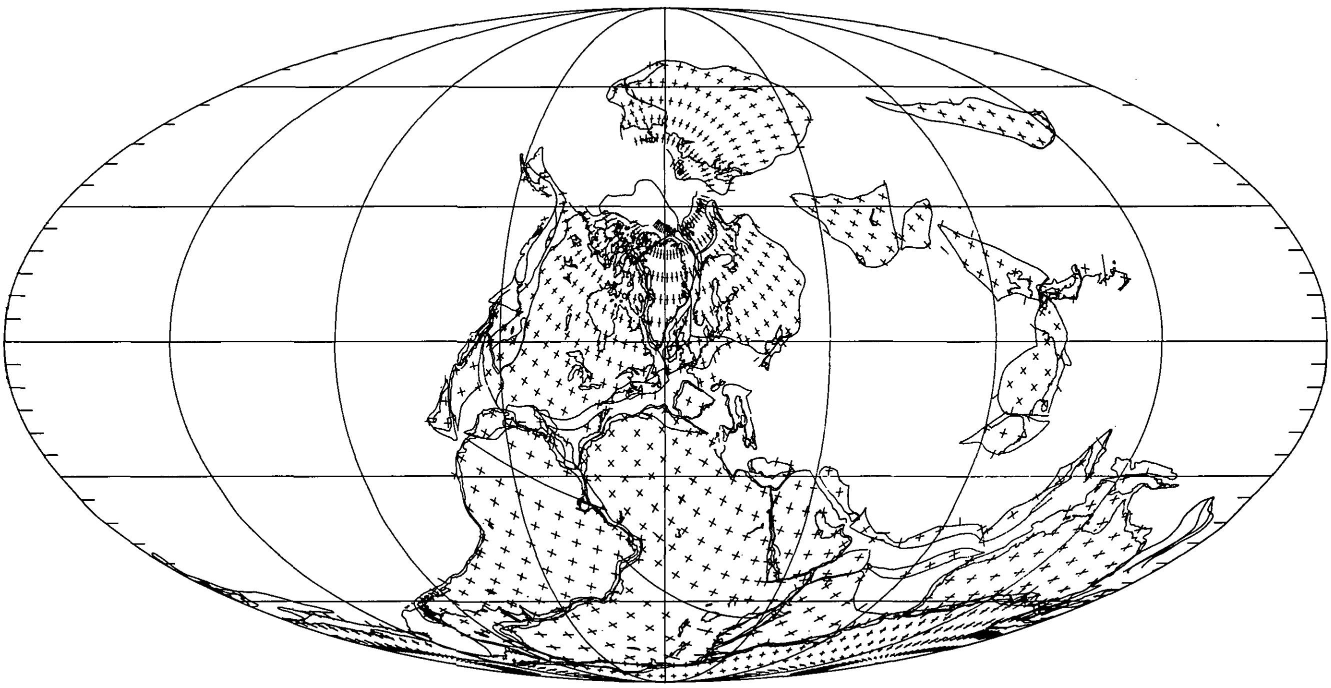

Figure 3. Paleocontinental reconstruction of Laurussia and Gondwana during the Namurian.

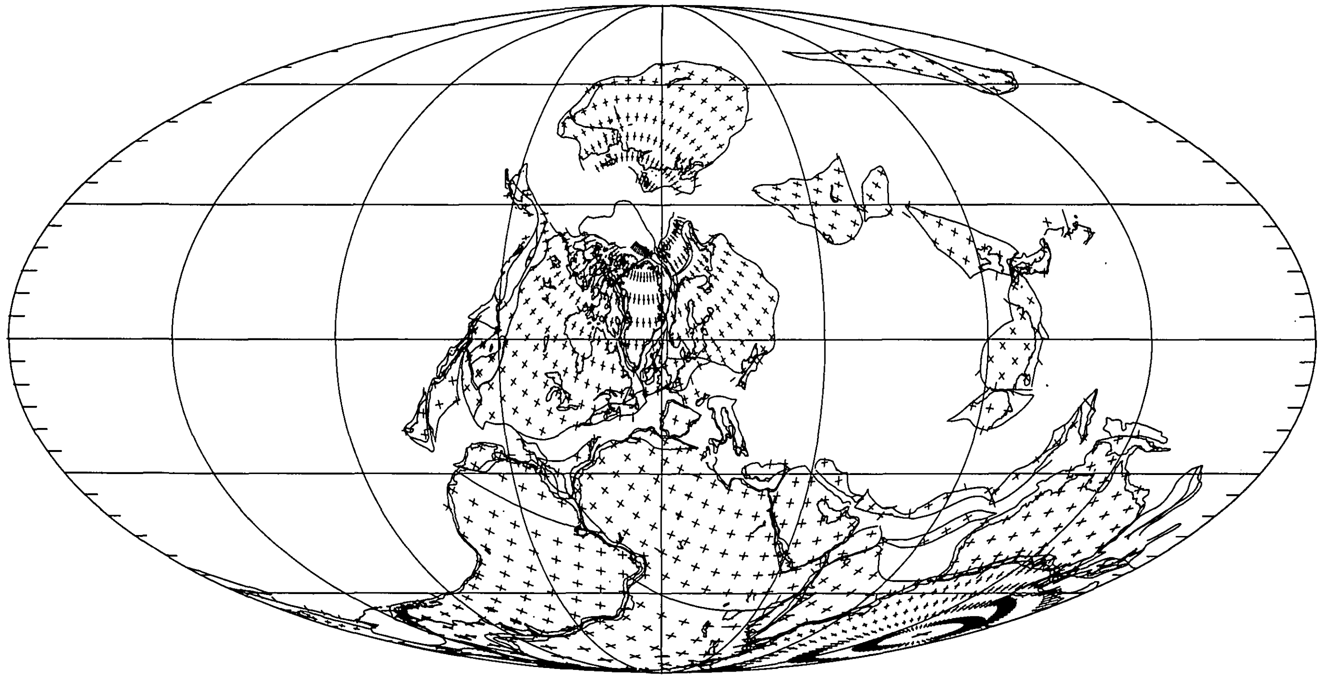

Figure 4. Paleocontinental reconstruction of Laurussia and Gondwana during the Visean.

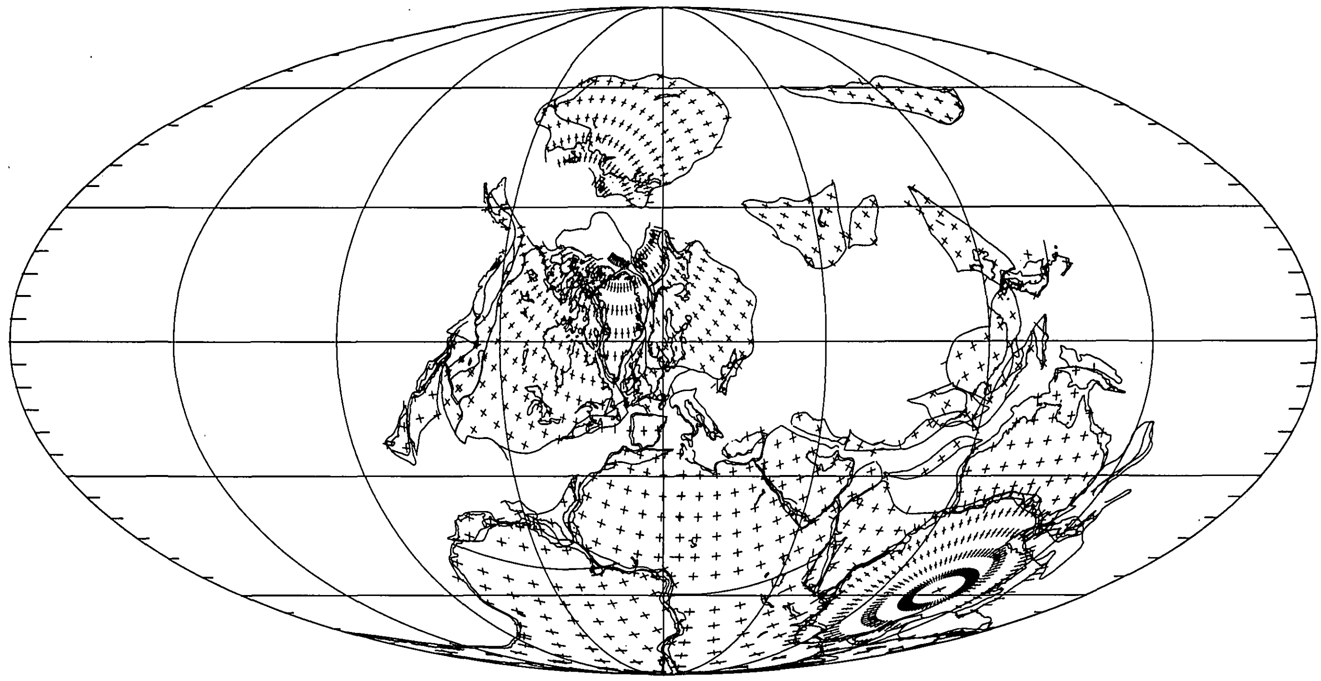

Figure 5. Paleocontinental reconstruction of Laurussia and Gondwana during the Tournaisian.

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Global Cyclostratigraphy

M.A. Perlmutter and M.D. Matthews

Global cyclostratigraphy, a conceptual semiquantitative model, can be used to predict stratigraphy by evaluating the response of depositional systems to the long-term tectonostructural evolution of a basin and short-term orbitally forced climatic changes (Perlmutter and Matthews, 1990, 1992). Long-term conditions and processes are considered as being constant or having constant rates of change compared to short-term conditions and processes, and thus the long-term geologic system provides a stable framework for dynamic short-term variations in depositional conditions.

Conceptual and dynamic climate models of a geologic epoch or age, combined with the distribution of paleoclimate indicators for that period of time, are used to analyze the spatial and temporal variation of global climate during a Milankovitch cycle. Sediment flux in relation to climate variation is modeled by assessing the effects of provenance, topography, temperature, humidity, runoff, and soil binding on sediment production and transport. Stratigraphy is then predicted by integrating sediment flux with conceptual and dynamic models of the development of accommodation space. Global cyclostratigraphy enables stratigraphy to be forecast in regions where there is little or no direct data.

GLOBAL CLIMATE

Global climate exhibits a latitudinally zoned pattern caused by the thermal gradient between the equator and the poles and the variation of precipitation and evaporation (humidity and runoff) related to the Hadley circulation (fig. 1; Perlmutter and Matthews, 1990, 1992). Regional and local effects cause the global climate pattern to be azonal. Azonal effects include (1) a poleward shift of the intertropical convergence zone caused by the differential heating of land and sea, which can produce monsoons in equatorial areas and shift and compress climates in adjacent poleward regions, (2) circulation around ocean-centered midlatitude high-pressure cells, which on east sides of continents produce onshore winds in equatorward areas and offshore winds in poleward areas, and on west sides of continents produce onshore winds in poleward latitudes and offshore winds in equatorward latitudes, (3) warm ocean currents and sea surface temperatures, which provide heat and moisture onshore to lower midlatitude eastern coasts and upper midlatitude western coasts, (4) cool ocean currents and sea surface temperatures, which can cool midlatitude western coasts, (5) upwelling of cold, deep ocean water, which can cool adjacent coasts and, (6) elevation, which causes cooler and wetter conditions on windward sides of mountains and warmer and drier conditions on leeward sides of mountains.

MILANKOVITCH-INDUCED GLOBAL CLIMATE CHANGE

The global climate pattern migrates during a Milankovitch cycle as Hadley circulation, and regional and local azonal climatic modifications respond to changes in the seasonal distribution of insolation caused by orbital oscillations (fig. 1; Perlmutter and Matthews, 1990). Climates can shift relatively large distances (up to 30° latitude) in geologically short intervals of time (<10,000 years). Evaluating or predicting the global distribution of endmember climates over a Milankovitch cycle (climatic maximum versus climatic minimum) permits identification of the sequence of climates in any particular location for any time period (Perlmutter and Matthews, 1990, 1992).

Although the Earth generally has been considered as cooler and drier during the climatic minimum and warmer and wetter during the climatic maximum, this generalization does not hold for the midlatitudes because of the reverse nature of Ferrel cell circulation compared to Hadley and polar cell circulation. The Ferrel cell moves upwelled air toward the equator rather than the poles. As a result, the wettest areas in the Ferrel cell are poleward, and an equatorward move in the position of the Ferrel cell during the climatic minimum causes midlatitude areas to become more humid.

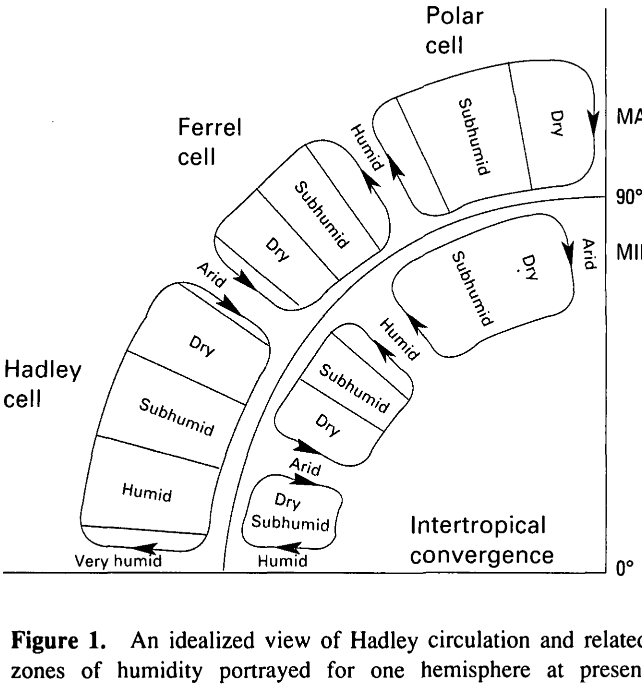

Figure 1. An idealized view of Hadley circulation and related zones of humidity portrayed for one hemisphere at present (climatic maximum; shaded area) and an idealized shift in position and size of Hadley circulation cells and related zones of humidity at a climatic minimum (modified from Perlmutter and Matthews, 1990).

EFFECT OF MILANKOVITCH CYCLES ON SEDIMENT FLUX

The timing of sediment input relative to the stages of lake and sea levels can be different in regions with different (orbitally forced) climatic sequences. In general, fluvial runoff and sediment yield are highest during the wettest climatic phases and tend to coincide with transgressions or highstands in lacustrine basins. Note that highstand lake levels and periods of maximum sediment yield may not be globally synchronous on a Milankovitch time scale in widely separated continental basins because of the effects of the Ferrel cell (Perlmutter and Mathews, 1990). In marine basins, however, the highest sediment delivery to continental margins may occur at any phase of glacioeustasy because the atmospheric pattern associated with the Ferrel cell causes runoff and sediment yield to be out of phase with glacioeustasy in midlatitudes. A global cyclostratigraphic evaluation of sediment flux indicates variation by as much as an order of magnitude during Milankovitch climate cycles. The specific magnitude and timing of the changes in sediment supply in a particular region depend on the climatic range and topography of the provenance area. Grain size and mineralogy also vary as a function of the climatic succession.

EFFECT OF SEDIMENT FLUX ON STRATIGRAPHIC INTERPRETATION

As a consequence of the variability of sediment flux, interpretations of the magnitude and direction of fourth- and fifth-order eustatic changes may be influenced by paleogeography. Regions with high sediment supply in phase with highstand sea level may bias the resulting stratigraphic record and interpretation toward highstand conditions over lowstand conditions because of higher preservation potential and seismic resolution. Conversely, high yield at lowstand sea level will tend to bias the record and interpretation toward lowstand conditions. Additionally, interpretation of the direction of longer term sea level changes (second- and third-order scales) also may be biased by a progressive shift in timing of sediment supply relative to sea level that can occur as a result of continental drift of the drainage basin through a series of different climatic zones.

The stratigraphic pattern of both continental and marine sequences can be strongly influenced by the climatic pattern in the drainage area of the associated fluvial regime. Neglecting to account for variations in sediment flux may cause misinterpretation of the occurrence of potential reservoir sands associated with both shelf and turbidite and fan systems and misidentification of the occurrence of potential source rock associated with condensed sections.

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Paleomagnetism and Carboniferous Climate

N.D. Opdyke and V.J. DiVenere

A large-scale climatic change is recorded in Carboniferous and Permian rocks of North America. The middle Mississippian to Early Pennsylvanian sediments of central and eastern North America are characterized by a dry, oxidizing climate reflected in an abundance of red beds. The climate turned wet in the Pennsylvanian, evidenced by a lack of red beds and an abundance of coal. Iron is mostly held in the reduced forms, largely siderite. Dry conditions returned in the latest Pennsylvanian and into the Permian as shown by the return of red beds and reduction of coals, particularly domed peat types. One of the primary factors influencing climate was the paleolatitude of ancient sedimentary environments. Paleomagnetic data compiled from North America may be used to determine the expected paleolatitude of any site on the continent. Figure 1 shows the expected paleolatitude for Lawrence, Kans. (present lat 39°N., long 264.9°E.). Paleolatitudes are calculated from North American mean paleopoles of Van der Voo (1990). Carboniferous and Permian mean paleopoles are based on 13 to 15 individual poles. The geographic position of Kansas varies throughout the Paleozoic, reaching a position of about lat 40°S. during the Early Devonian. Kansas then moved toward the north during the Mississippian and crossed the equator in the Late Pennsylvanian. It remained on the equator until the Late Triassic when Kansas began to move steadily north, reaching its present latitude in the Cretaceous. This latitude change may be compared with that of Morgantown, W. Va. (fig. 1). The paleolatitude curves of Lawrence, Kans., and Morgantown, W. Va., are similar; however, West Virginia crossed the paleoequator later than Kansas in the Early Permian.

It is obvious from the paleomagnetic data that more factors than latitude change are involved in the large-scale Carboniferous climatic change. The initial dry to wet sequence correlates well with the approach to the equator in the Mississippian, but this correlation ends in the Late Pennsylvanian. It is during this period that North America and Africa converged and collided and left a large orogen to the east of the midcontinent. It seems reasonable, therefore,

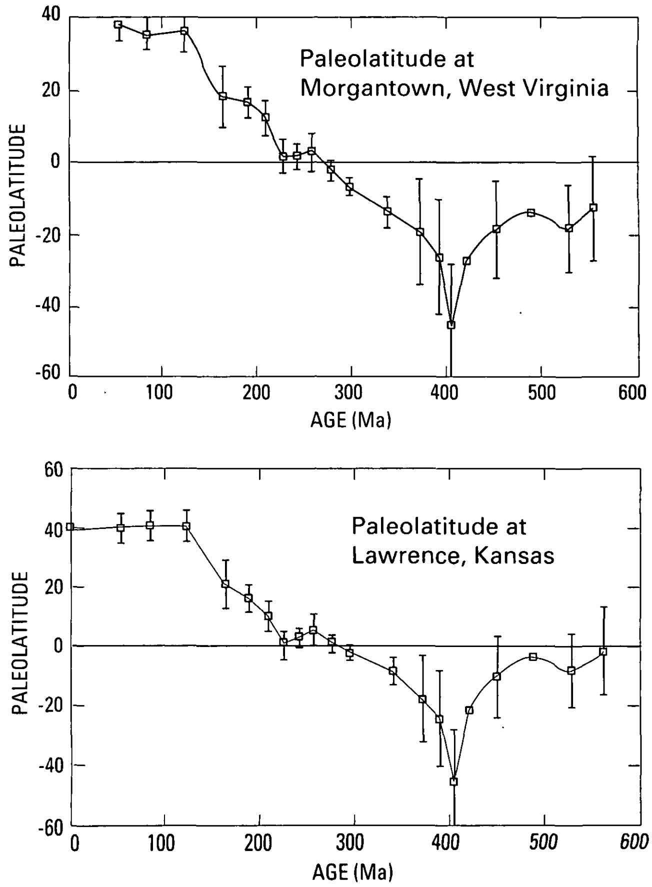

Figure 1. Comparative change in paleolatitude at Morgantown, West Virginia, and at Lawrence, Kansas.

that the dry conditions evidenced in the late Paleozoic, as seen in the voluminous red beds of the time, may have been partly the result of a significant rain shadow effect from the growing Appalachians. It is also during the late Paleozoic that there is evidence for glaciation in Gondwanaland. These glacial cycles are probably reflected in the midcontinent cyclothems.

Magnetostratigraphy offers the possibility of establishing synchronous timelines in the stratigraphic record. Recent work in Upper Mississippian and Lower Pennsylvanian sediments from Pennsylvania and the Canadian Maritime Provinces (DiVenere and Opdyke, 1990, 1991) has resulted in the beginnings of a detailed magnetostratigraphy for this period. Our work, and previous work by Roy and Morris (1983), suggests that the base of the Permo-Carboniferous reverse superchron (PCRS) is within the Westphalian B to C interval. Figure 2 gives the results of our own work (present best estimate) as well as a survey of the paleomagnetic data in the literature for the Carboniferous. A preliminary magnetic polarity time scale for the Carboniferous is presented.

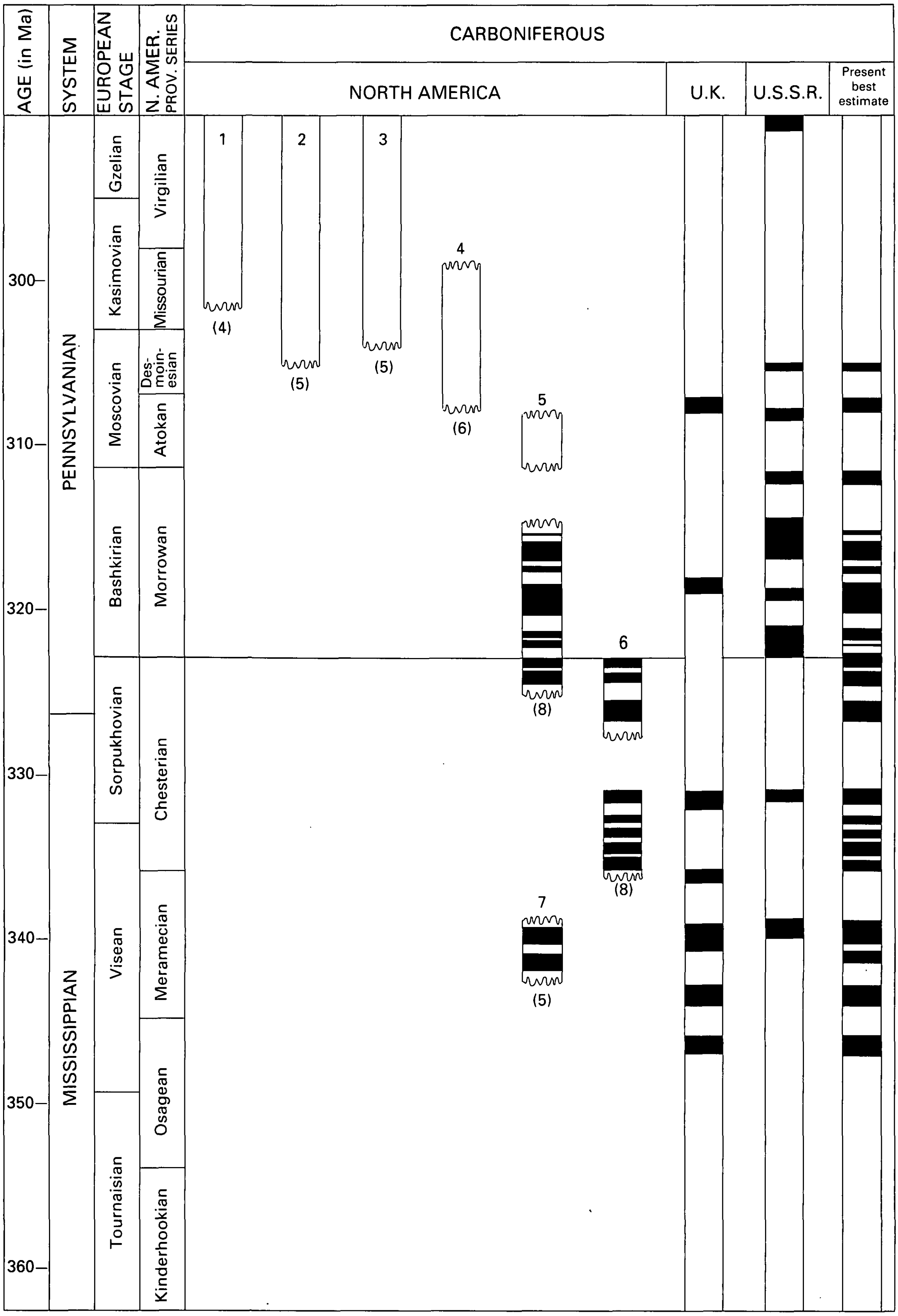

Figure 2. Summary of world magnetostratigraphic data for the Carboniferous. The reliability of each study is rated on a scale of 1 to 10, and the appropriate score is indicated below each study in parentheses. The columns labeled U.K. and U.S.S.R. are data compilations from these regions. The last column on the right is the present best estimate of the reversal history for the Carboniferous.

Due to the single polarity nature of the PCRS, magnetostratigraphy probably will not be of use during the Middle Pennsylvanian and most of the Permian, though there are claims for some short normal polarity intervals (Helsley, 1965). We believe that cyclostratigraphy may offer the best chance of high resolution stratigraphic correlations during the Middle Pennsylvanian.

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Relative Effects of Tectonism, Eustasy, Paleoclimate, and Paleo-Oceanography on Atlantic Passive-Margin Sedimentation

C. Wylie Poag

Seismostratigraphic analysis has stimulated vigorous reexamination of the relationships between depositional patterns in marine sedimentary basins and three primary regulating agents: tectonism, eustasy, and paleoclimate. Most authors acknowledge the interplay of these regulators, but few try to integrate them. Fierce controversies have arisen over such questions as whether eustasy or tectonism is the main forcing agent for cyclic depositional episodes and what is the relative importance of sediment supply or sediment load in controlling depositional patterns. The role of paleoclimate usually gets minimal attention.

A particularly encouraging trend of late, however, is the appearance of more case studies of individual basins (Galloway, 1989; Harris and Grover, 1989; Fulthorpe, 1991) that provide field tests of the conceptual models. As recent examples, Poag and Sevon (1989) and Poag (1991, 1992) synthesized the main aspects of postrift deposition along the middle U.S. Atlantic margin (coastal plain to continental rise; fig. 1). These authors used greater than 10,000 km of multichannel seismic reflection profiles correlated to 88 key boreholes to map the distribution and thickness of 23 postrift allostratigraphic units (unconformity-bounded sedimentary deposits) across this region of about 500,000 km^2. They focused on five main aspects of sedimentation: (1) net volumetric siliciclastic accumulation rates, (2) latitudinal migration of depocenters, (3) bathymetric migration of depocenters, (4) gross lithofacies composition, and (5) systems-tract development.

From variation around the mean siliciclastic accumulation rate (9,000 km^3/m.y.), Poag and Sevon (1989) recognized five temporal phases (I–V) of deposition (fig. 2). During each phase, sedimentation was regulated by a complex interplay (spatially and temporally variable) between tectonism (source-terrain uplift and basin subsidence), eustasy, paleoclimate, and paleo-oceanography. These agents regulated the location of terrigenous source terrains, dispersal routes, and depocenters and controlled net accumulation rates, gross lithofacies composition, and systems-tract development (fig. 3).

Alternating uplift and quiescence of three source terrains ranks first as the most consistently effective regulator, mainly by controlling the supply of siliciclastic sediments. A notable exception occurred during phase IV, when a tropical rainforest (extensive vegetation; Wolfe, 1978) severely limited siliciclastic accumulation (Cecil, 1990) in spite of source-terrain uplift. Eustasy (inferred from the Exxon model of sequence stratigraphy; Haq and others, 1987) ranks second, primarily because sediments are distributed and redistributed once they reach the basins. Sea level was particularly effective during short-term rises or falls by determining the bathymetric position of depocenters and by timing the succession of systems tracts. A marked increase in sediment supply, however, triggered by source-terrain uplift (such as in the Late Cretaceous; phase III), could override eustatic effects. Likewise, in the absence of source-terrain uplift or continental glaciation, major eustatic falls did not accelerate siliciclastic accumulation (for example, phase IV).

Paleoclimate ranks third, having helped to regulate (to varying degrees) every aspect of sedimentation studied, except systems-tract development and latitudinal depocenter migration. Paleoclimate was particularly effective in its extremes [for example, unusual aridity during early phase I (evaporite deposition), everwet rainforests during phase IV, and extensive continental glaciation during phase V] and was a prime control of carbonate platform development (phases I, II).

Basin subsidence ranks fourth; it significantly enhanced siliciclastic accumulation, gross lithofacies, systems-tract development, and bathymetric position of

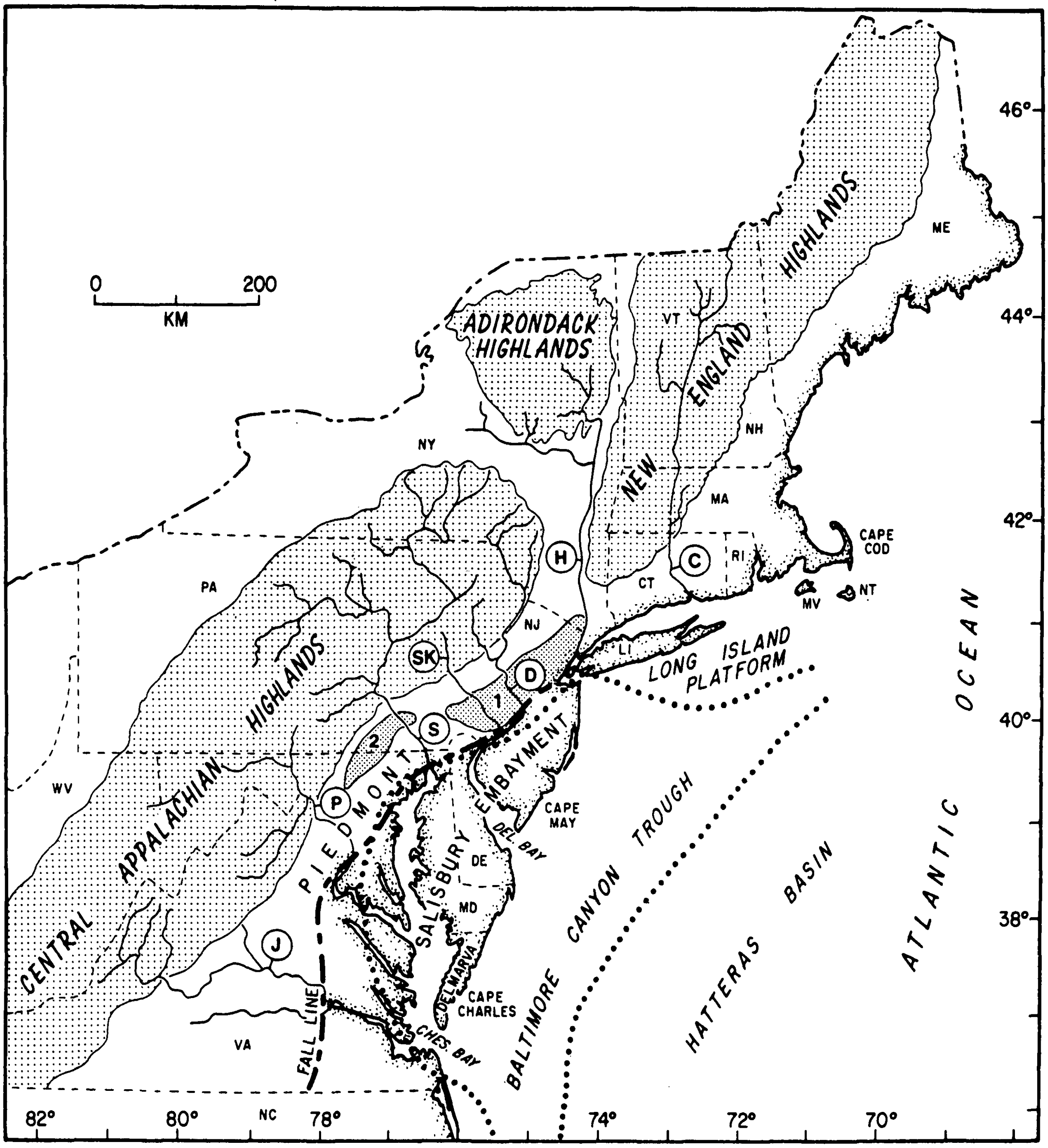

Figure 1. Location map of the U.S. middle Atlantic continental margin. Rivers denoted by letters: C, Connecticut; H, Hudson; D, Delaware; SK, Schuylkill; S, Susquehanna; P, Potomac; J, James. Triassic rift basins denoted by numbers 1 and 2. Islands denoted by letters: LI, Long Island; MV, Marthas Vineyard; NT, Nantucket. Bays denoted by abbreviations: Ches., Chesapeake; Del., Delaware.

depocenters during early phase I but played only a minor role thereafter. Paleo-oceanography ranks fifth, as the least effective regulator, mainly having affected gross lithofacies composition, but it also modified systems-tract development and altered the latitudinal location of some depocenters.

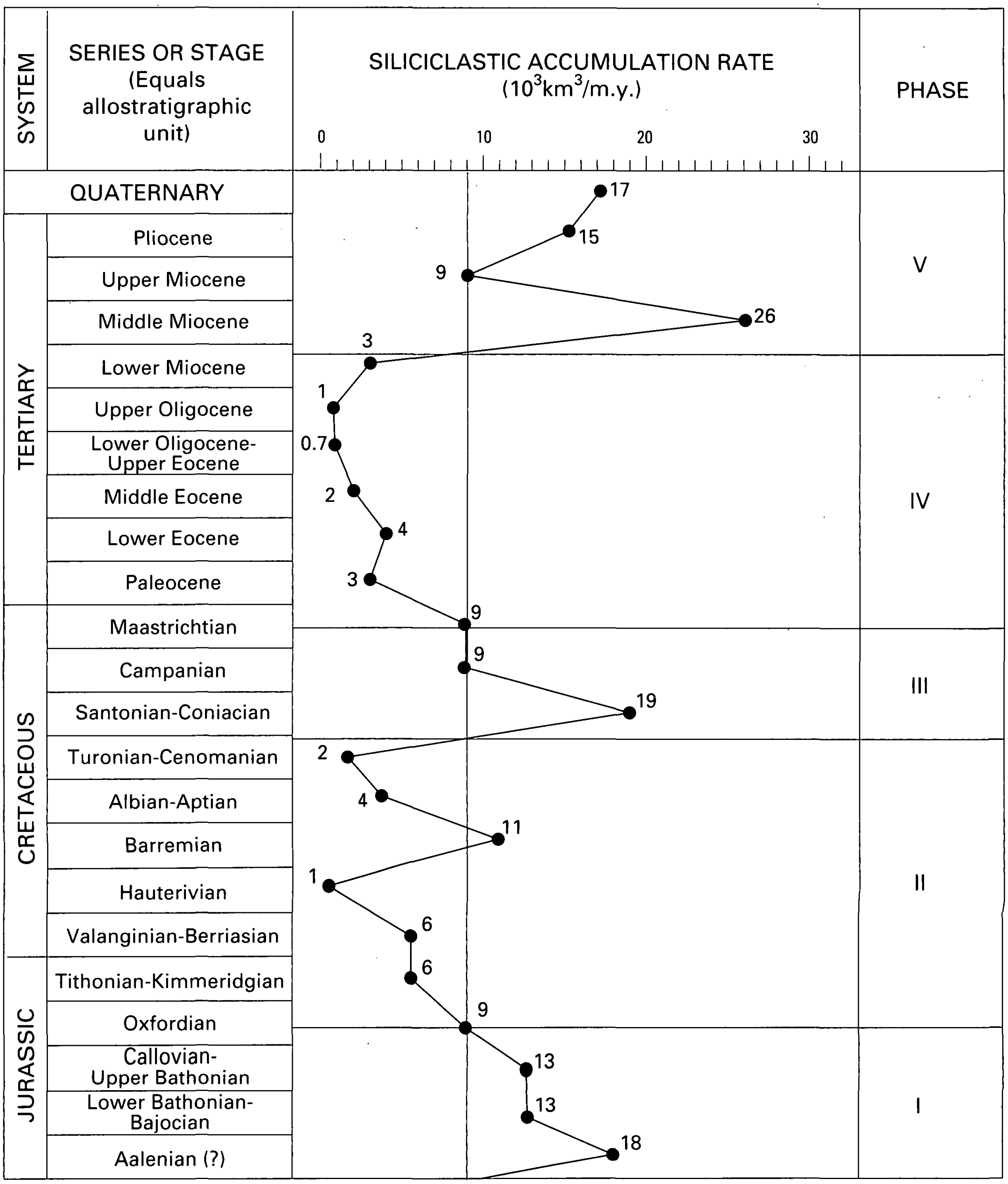

Figure 2. Net siliciclastic sediment-accumulation rates for 23 postrift allostratigraphic units of the study area. Values are volumetric rates given in thousands of cubic kilometers per million years. Raw values for middle Miocene through Quaternary units have been reduced by 30 percent to make them compatible with volumes of older, more deeply buried and compacted units. Roman numerals indicate five depositional phases derived from variation of accumulation rate about the mean value (9,000 km³/m.y.; from Poag, 1992).

Aspect of sedimentation	Phase	Regulator: Source-terrain uplift	Eustasy	Paleoclimate	Basin subsidence	Paleo-oceanography
Siliciclastic accumulation rate	V	●	●	●		
	IV		○	●		
	III	●		○		
	II	●	●			
	I	●	●	●	●	
Latitudinal depocenter migration	V	●				○
	IV	●				
	III	●				○?
	II	●				
	I	●				
Bathymetric depocenter migration	V	●	●	●		
	IV		●			
	III	●	●	○?		
	II	●	○	●	●	
	I			○	●	
Gross lithofacies composition	V	●		●		●
	IV		●	●		●
	III	●				
	II	●	●			●
	I	●	●	●	●	●
Systems-tract development	V	●	●		○	○
	IV		●		○	○
	III	●	●		○	
	II	●	●		○	
	I	●	●	○	●	○

Figure 3. Summary of relative effects of five regulating agents on five aspects of sedimentation in the study area during each of five phases of deposition (I–V). Solid circles, major effect; open circles, minor effect; blank spaces, no appreciable effect.

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Pennsylvanian Vegetation and Soils

Gregory J. Retallack

Fossil tree lycopsids of Pennsylvanian swamps were unlike modern swamp plants botanically, but how different were Pennsylvanian vegetation types as ecosystems, as soil binders, as producers of carbon, as consumers of nutrients, and as regulators of water? Were the habitats outside the swamp vegetated at all? To what extent had the evolution of forests, initiated during Devonian time, progressed to create the variety of woody vegetation found today? These questions are difficult to impossible to answer from the evidence of fossil plants, which were preserved mainly as fragments in swampy environments where aerobic decay was inhibited by anoxia. Fortunately, there is a new line of evidence that is being applied to these and related problems: the evidence from fossil soils.

Fossil soils are by definition in the place they form, unlike many fossil plants and animals. Root traces in paleosols can be evidence of vegetation in habitats not suitable for preservation of fossil plants, including climatically dry and locally well drained sites (Retallack, 1984). The stature, biomass, and economy of ecosystems can be interpreted within broad limits from such paleosol features as the size and penetration of root systems, the degree of development of soil horizons and soil structure, and the depletion of alkaline earth and other elements that are major cationic nutrients for plants (Retallack, 1990). A variety of plant formations can be recognized from the evidence of paleosols (table 1), and only in some cases is their botanical composition known. In the ensuing discussion these vegetation types are grouped into general environmental categories of waterlogged, climatically wet, climatically dry, and frigid. Evidence from paleosols indicates that all of these varied environments supported woody vegetation by the Pennsylvanian.

The best known Pennsylvanian vegetation is that of waterlogged habitats, especially swamps of tree lycopsids

Table 1. Geological antiquity of plant formations based mainly on features of paleosols.
[E, Bt, Bs, and Bk designations from U.S. Department of Agriculture, 1975].

Plant formation	Characteristic paleosol features	Age	References
Open grassland	Red, brown, or gray paleosol with abundant fine root traces and granular soil peds, sometimes with a shallow horizon of calcareous nodules (Bk)	Eocene	Retallack (1990).
Wooded grassland	Red, brown, or gray paleosol with abundant fine root traces and granular soil peds, and scattered large woody root traces, sometimes with subsurface clayey horizon (Bt) and deeper calcareous nodules (Bk)	do.	Do.
Sea grassland	Root traces in shallow subtidal sediments, often associated with distinctive suite of large foraminifera	Late Cretaceous	Brasier (1975).
Fireprone shrubland	Red or brown paleosol with moderate-sized woody root traces and abundant fossil charcoal	Late Triassic	Harris (1957).
Heath	Sandy, noncalcareous paleosol with moderate-sized woody root traces and shallow siderite nodules or other indicator of high water table	Early Triassic	Retallack (1977).
Desert scrub	Red or brown paleosol with sparsely scattered large woody root traces or rhizoconcretions, and calcareous nodules (Bk horizon) close to the surface	Early Permian	Loope (1988).
Taiga	Paleosol with large woody root traces and frost-heave structures in periglacial deposits	Latest Pennsylvanian	Retallack (1980).
Tundra	Paleosol with small root traces and frost-heave structures in periglacial deposits	do.	Do.
Bog	Black or gray shale, coal, or chert with abundant fossil plants lacking true roots, such as mosses or liverworts	do.	Anderson and Anderson (1985).
Shrubland	Red or brown paleosol with clumped woody root traces of moderate size, common easily weathered minerals such as feldspar and a shallow subsurface horizon of calcareous nodules (Bk)	Pennsylvanian	Loope (1988).
Rainforest	Red or brown paleosol mainly of kaolinite or other deeply weathered clay, with large woody root traces and little feldspar or carbonate	do.	Keller and others (1954); Retallack (1990).
Oligotrophic forest	Red or brown paleosol principally of quartz, with large woody root traces and little clay, feldspar, or carbonate	Mississippian	Percival (1986); Retallack (1990).
Dune binders	Small but deeply penetrating root traces in eolian or fluvial sand	do.	Ettensohn and others (1988); Loope (1988).
Fen	Black or gray paleosol, sometimes coal bearing, with small root traces and calcareous nodules	do.	Rex and Scott (1987).
Carr	Black or gray paleosol, sometimes coal bearing, with large woody root traces and calcareous nodules	do.	Retallack and Dilcher (1988).
Swamp	Black or gray paleosol, sometimes coal bearing, with large woody root traces, lacking pyrite or carbonate	Late Devonian	DiMichele and others (1987).
Wooded shrubland	Red or brown paleosol with scattered large woody root traces and stump casts and abundant smaller woody root traces, as well as easily weathered minerals such as feldspar and subsurface calcareous nodules (Bk)	do.	Retallack (1985).
Dry woodland	Thick red or brown paleosol with large woody root traces and stump casts and common easily weathered minerals such as feldspar, as well as deep subsurface calcareous nodules (Bk)	do.	Retallack (1985, 1990).

Table 1. Geological antiquity of plant formations based mainly on features of paleosols—Continued.

Plant formation	Characteristic paleosol features	Age	References
Forest	Thick red or brown paleosol with large woody root traces and stump casts, and common easily weathered minerals such as feldspar, as well as development of subsurface leached (E), clay-enriched (Bt), or ferruginized (Bs) horizons	Late Devonian	Retallack (1985).
Mangal	Black or gray paleosol, sometimes coal bearing, with large woody root traces and marine body and trace fossils, sometimes also pyrite nodules	Middle Devonian	DiMichele and others (1987); Retallack (1990).
Marsh	Black or gray shale, coal or chert containing abundant herbaceous plants with rhizomes or true roots	Early Devonian	Kidston and Lang (1921); Krassilov (1981).
Brakeland[1]	Red or brown paleosol, with small root or rhizome traces of herbaceous, but not sod-forming plants	Late Silurian	Retallack (1990).
Salt marsh	Black or gray paleosol with small root or rhizome traces and marine body and trace fossils	Early Silurian	Schopf and others (1966).
Polsterland	Red or brown paleosol, with burrows, lichen stromatolites or reduction spotted, erosion resistant mounds, as might form under plants without true roots	Late Ordovician	Retallack (1990).
Microbial rockland	Rock surface with weathering rind, endolithic microbial trace fossils or biotic isotopic depth function	Precambrian (1.2 Ga)	Beeunas and Knauth (1985).
Microbial earth	Red or thick and leached paleosol with microfossils, microbial trace fossils, soil structure, or element or isotopic depth function suggestive of life	Precambrian (3 Ga)	Grandstaff and others (1986); Retallack (1986b, 1990).
Sabkha stromatolites	Algal lamination, often with domed form, and with pseudomorphs or crystals of evaporite minerals	Precambrian (3.5 Ga)	Schopf (1983).
Aquatic stromatolites	Algal lamination, often with domed form, crossed by traces of cyanobacterial sheaths	do	Do.

[1]Brakeland denotes a formation of numerous individual plants of similar physiognomy.

(*Lepidodendron*) and marattialean tree ferns (*Psaronius*). The plants of these former peat swamps are known from fossils in coals and enclosing shales. Vegetation of swamps not so waterlogged as to encourage peat formation is found in the form of stumps and leaf litters preserved in carbonaceous shale surface horizons of gleyed soils. The plants of acidic, mineral soils ("clastic swamps" of DiMichele and others, 1987; Gastaldo and others, 1989) were similar, though more diverse, than those of peat swamps. Woody vegetation of local alkaline wetlands, or carr, also may be known from Pennsylvanian coals with calcareous nodules, or "coal balls" (Retallack, 1986a; Retallack and Dilcher, 1988). The fossil flora of these eutrophic wetlands includes a very diverse flora, but shares many species with the flora of acidic swamps (Phillips, 1980).

Marine influenced woody vegetation, or mangal, of low diversity and dominated by *Cordaites* also is known from Pennsylvanian coal-bearing paleosols having abundant pyrite and sparse marine fossils (Raymond and Phillips, 1983). Herbaceous vegetation of wetlands, such as salt marsh, marsh, and fen probably also existed during Pennsylvanian time (DiMichele and others, 1979), especially considering geologically more ancient occurrences (table 1). Indeed, many of the nonmarine limestones of the Monongahela Formation of West Virginia and Ohio (see p. 19, app. 1, locs. 1–3), which have fine root traces, abundant brecciation, and local lamination, are similar to lime muds accumulating under periphyton algal fens of the modern Florida Everglades (as described by Spackman and others, 1969). Wetland vegetation of mosses and other plants lacking true roots are well known from rocks as ancient as Early Permian (Meyen, 1982), but some moss-filled carbonaceous shales within Gondwanan glacial deposits could be as old as Late Pennsylvanian (Anderson and Anderson, 1985).

Climatically humid, well-drained soils were forested well before Pennsylvanian time (table 1), but little is known botanically about this ancient vegetation. Such noncalcareous, red paleosols with large root traces and persistent weathering-susceptable minerals (Alfisols) have been reported from Pennsylvanian rocks in England (Besly and Fielding, 1989), and similar profiles exist in the Pennsylvanian and Permian Fountain Formation of Colorado, the Permian Hermit Shale of the Grand Canyon, Arizona, and the Permian Vale Formation near Lake Abilene, Texas (app. 1, locs. 4–6). All have copiously branching root traces like those of woody gymnosperms, but only in the Permian examples is there associated evidence of the plants of these humid well-drained forests, including a variety of broad-leaved seed ferns (*Supaia*, *Evolsonia*: White, 1929; Mamay, 1989). Fossil floras dominated by broad-leaved seed ferns (*Megalopteris*) also are known from sediments within ravines of an Early Pennsylvanian tropical paleokarst in northeastern Illinois (Leary, 1981). These fossil plants probably were derived from well-drained soils higher in the landscape and are further evidence of wet broad-leaved forests at that time.

There is also evidence from paleosols that forest cover extended during Pennsylvanian time onto nutrient-poor clayey soils (Ultisols) and sandy soils (quartzipsamments, dystrochepts, and perhaps also Spodosols) of humid climates. Sandy, nutrient-poor paleosols are widely known as ganisters, a Welsh mining term for these refractory quartzites in coal measures. Many of the ganister-bearing paleosols had a shallow water table as indicated by siderite nodules, but both these and thick, deeply rooted and well-drained ganisters commonly include *Stigmaria*, the root system of tree lycopsids (Percival, 1986; Retallack, 1990). Possible Pennsylvanian Ultisols have been known for some time from the diaspore clay district of the Missouri Ozark Mountains (Keller and others, 1954). The Farnberg pit (app. 1, loc. 7) contains profiles with both a horizon of clay enrichment (argillic or Bt horizon) and large woody root traces of gymnosperms. In addition to deeply penetrating root traces, one of these paleosols in the Farnberg pit also shows a surficial mat of roots. This soil is similar to those now found under Guineo-Congolian and Amazonian rainforest, but little is known about the botanical affinities of this possible Pennsylvanian rainforest, a topic long of interest to paleobotanists (Krassilov, 1975).

Dry woodland also is known earlier than Pennsylvanian time (table 1), and many slickensided, red, red-mottled, and calcareous nodular paleosols have been reported from Pennsylvanian rocks, even within major coal basins (Joeckel, 1988). The problem of calcareous, red-mottled paleosols of dry climates alternating in sequences with noncalcareous, red-mottled paleosols and thick coal beds of wet climates, has recently been attributed to Milankovitch variation in climate, and this also explains other features of Pennsylvanian cyclothemic sedimentation (Cecil, 1990). Such paleoclimatically distinct, superposed paleosols can be seen on either side of a thin and shaly margin of the Pittsburgh coal near Burnsville and Sissonville in West Virginia and also within the upper part of the Bonner Springs Shale near Holliday, Kans. (app. 1, locs. 8–10). Deeply penetrating root traces, low-angle slickensided cracks, and calcareous rhizoconcretions in these paleosols indicate that they were generally well drained, but the pattern of gray and red mottling is similar to that formed in modern soils by seasonal waterlogging ("groundwater gley" of Retallack, 1990). These paleosols, which are generally similar to those of the Indogangetic Plains of India, receive more than 1,000 mm mean annual rainfall (Hanrgram Series soils of Murthy and others, 1982) for the noncalcareous profiles and some 700 to 1,000 mm for the calcareous profiles (Sadhu Series soils). Comparable modern soils support lowland, evergreen, wet forest, and deciduous seasonally dry, monsoon forest, respectively (Champion and Seth, 1968; Retallack, 1991).

Pennsylvanian vegetation of the noncalcareous, red-mottled paleosols may have been similar to that of other gleyed paleosols ("clastic swamps" of DiMichele and others, 1987; Gastaldo and others, 1989) or to the broad-leaved wet forests of seed ferns already discussed. Vegetation of the dry-climate phase also may be known. Pith casts of calamites occur in one of these calcic-vertic-hydromorphic paleosols above the Sewickley coal near Macksburg in Ohio (app. 1, loc. 11), and casts of large gymnospermous roots and stumps in two superimposed paleosols of this kind occur below the thin Williamsburg coal in Clinton Lake Spillway, Kansas (app. 1, loc. 12). A systematic search for plant fossils in these paleosols may be quite revealing.

The well-known conifer (*Walchia*) and seed-fern (*Callipteris*) vegetation of fossil localities near Hamilton (Leisman and others, 1988; Rothwell and Mapes, 1988) and Garnett (Winston, 1983), both in Kansas, could also represent vegetation at the dry extreme of Milankovitch cycles. I could not find paleoclimatically instructive paleosols at either site, but both localities include evidence of channel incision and a position within their respective cyclothems that is compatible with this view. The Garnett site is south of, and at the same stratigraphic level as, a widespread calcareous paleosol to the north (Joeckel, 1988). These xeromorphic conifer-callipterid floras have in the past been taken to indicate Permian rather than Pennsylvanian time, upland rather than lowland floras, or extrabasinal rather than basinal floras (Pfefferkorn, 1980). None of these much-argued alternatives works well for the Hamilton or Garnett localities. Both sites are now known to be Pennsylvanian (Virgilian and Missourian, respectively). Both include gray to black shales with exceptional preservation of organic remains and no clear sign of red beds or well-drained paleosols. Both are also well within the boundaries of their depositional basin. Schutter and Heckel (1985) were closer to the mark in proposing Pennsylvanian "conifer savannas," but that is not a term I would use (Retallack, 1990), especially without evidence from fine root traces and granular mull humus in the paleosols for a continuous herbaceous ground cover like that provided by grasses (well demonstrated in some Miocene paleosols; Retallack, 1991). The nature of the paleosols does not exclude wooded shrubland, but the distribution of root traces in the calcareous paleosols mentioned here is more like that of open woodland or dry forest.

It could be that the conifer-callipterid dry woodland expanded at the expense of wetland vegetation of tree lycopsids and tree ferns by Milankovitch-driven climatic fluctuation the same as the more recent full glacial expansion of grassland expanded at the expense of interglacial rainforest in Africa and Amazonia. This new view of Pennsylvanian climatic variability calls for detailed reevaluation of the fossil records of both soils and plants.

A variety of woody desert vegetation types also had evolved by Pennsylvanian time, as indicated by paleosols in sequences of eolian dunes with calcareous rhizoconcretions and horizons of calcareous nodules close to the former soil surface (Loope, 1988). Pennsylvanian to Permian examples of these aridland paleosols exist in the Sangre de Cristo Formation near Howard and Coaldale in Colorado (app. 1, locs. 13–14). Their vegetation may have been structurally similar to pinyon or juniper woodlands of the North American desert Southwest. In the Permian Cutler Formation near Gateway in Colorado (app. 1, loc. 15), only small root traces were seen in thin calcareous paleosols, which may have supported vegetation structurally similar to the bluebush shrublands of central Australia. The botanical nature of Pennsylvanian woody desert vegetation remains completely unknown. There have not yet been identified any Pennsylvanian analogues of fireprone shrubland (also called chaparral, maquis, or matorral), nor desert succulent vegetation, nor any herbaceous aridland vegetation analogous to grassland.

Woody vegetation of frigid climates also may have evolved by Pennsylvanian time, judging from fossil root traces in paleosols associated with Gondwanan glacial deposits. In Carboniferous and Permian glacial deposits in the Sydney basin of southeastern Australia, there are remains of tundralike vegetation dominated by *Botrychiopsis* and taigalike vegetation dominated by *Gangamopteris* (Retallack, 1980). If woody vegetation extended to such high-latitude permafrosted soils, then it probably clothed high mountains as well.

Although counterparts of many modern vegetation types can be recognized from Pennsylvanian fossil plants and soils, they were certainly distinct botanically from modern vegetation, especially in lacking angiosperms. Pennsylvanian vegetation also may have been distinctive in some ways at the functional or ecosystem level, although only subjective impressions and conjecture can currently be offered in support of this idea. For example, the abundance of coal balls in Pennsylvanian coal seams contrasts with the extreme rarity of calcareous peat today and during the Mesozoic and Cenozoic. The rarity of ferruginous zones (spodic horizons) in nutrient-poor sandy paleosols in Pennsylvanian rocks contrasts with their abundance in sandy soils today and in those as old as Eocene. Perhaps Pennsylvanian vegetation was less acidifying and iron mobilizing than modern conifers of swamps and oligotrophic forests. Many Pennsylvanian trees were less densely woody than modern conifers of these habitats, and their foliage may have yielded fewer phenolic compounds to the leaching effects of rainwater. Flying insects appear in the fossil record during Late Mississippian time, and Pennsylvanian trees may not yet have evolved such an array of acidic and mildly toxic secondary plant metabolites to deter their herbivory as have modern conifers after several hundred million years of coevolution with insects. The degree of acid leaching and of iron redistribution in Pennsylvanian compared to modern soils could bear closer examination and may be only one of a number of differences between modern and ancient ecosystems that will become apparent from the study of fossil soils.

A complex picture is emerging of Pennsylvanian vegetation and its variation with climate, drainage, substrate, and time. Even now it is possible to create maps of Pennsylvanian vegetation and to recognize changes in vegetation related to cyclothemic sedimentation. An appreciation of vegetation beyond Pennsylvanian peat swamps is growing with the examination of paleosols and their associated fossil plants often too poorly preserved to have previously commanded much attention. Nevertheless, much remains to be done, both in the gathering of primary data and the reassessment of preexisting data to accommodate new views of Milankovitch climatic variation and models of the soil-vegetation system.

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Appendix 1. Mentioned localities of paleosols not studied in detail.

Loc. no.	Locality
1	Morgantown Mall, W. Va. (grid reference 854869, Osage 7.5-minute quadrangle, Monongalia County), in 600-m-long east-west cut for new mall, 1 mile west-northwest of intersection of highways I–79 and U.S. 19, are exposed Pittsburgh, Redstone, Fishpot and Sewickley coals of the Monongahela Formation, and gray calcareous paleosols above and below the nonmarine unranked Redstone limestone (Late Pennsylvanian): examined March 1, 1991.
2	Weston, W. Va. (grid reference 195496, Weston 7.5-minute quadrangle, Lewis County), in roadcuts of ramp within northwestern quadrant of cloverleaf at intersection of highways I–79 and U.S. 33 is exposed unranked Redstone limestone and Pittsburgh coal and its underclay, all of the Monongahela Formation (Late Pennsylvanian): examined March 2, 1991.
3	Bridgeport, Ohio (SE¼NE¼SE¼NW¼ sec. 33, T. 3 N., R. 2 W., Lansing 7.5-minute quadrangle, Belmont County), in roadcuts south of eastbound highway I–70 near base of descent into Ohio Valley is exposed a sequence of nonmarine limestones disrupted by well-preserved root traces and cradle knolls at several levels above the Sewickley coal of the Monongahela Formation (Late Pennsylvanian): examined March 3, 1991.
4	Manitou Springs, Colo. (SW¼SE¼NE¼NW¼ sec. 5, T. 14 S., R. 67 W., Manitou Springs 7.5-minute quadrangle, El Paso County), in gully 100 m north and up from Alpine Trail Road near Williams Canyon, are exposed several ganister-bearing paleosols with stigmarian root systems and siderite nodules, overlain by a sequence of red paleosols with deeply penetrating, drab-haloed, woody root traces, all in the Fountain Formation (Pennsylvanian and Permian): examined July 2, 1979.
5	Grand Canyon, Ariz. (grid reference 018917, Phantom Ranch 7.5-minute quadrangle, Coconino County), in red beds west of the Kaibab Trail are a sequence of red paleosols with ferruginized woody root traces and concretions forming the entire Hermit Shale (Permian): examined March 12, 1978.
6	Lake Abilene, Tex. (grid reference 158685, View 7.5-minute quadrangle, Taylor County), roadside gully south of unsealed road climbing the hill north of the dam wall exposes two thick red paleosols with gymnospermous roots, divided by bedded siltstone and sandstones covering and containing locally abundant remains of the seed fern *Evolsonia texana* in the Vale Formation of the Clear Fork Group (Early Permian): examined October 28, 1990.
7	Drake, Mo. (NE¼NW¼SW¼ sec. 30, T. 43 N., R.5 W., Goerlisch Ridge 7.5-minute quadrangle, Gasconade County), in the southeastern corner and uppermost levels of Farnberg pit, 3 miles southwest of Drake, are exposed several red clayey paleosols with woody root traces, in the Cheltenham Clay, which lies unconformably on paleokarst into the Ordovician Jefferson City Dolomite and is overlain conformably by the Fort Scott Limestone (Middle Pennsylvanian): examined November 3, 1989.
8	Locality 8. Burnsville, W. Va. (grid reference 314025, Burnsville 7.5-minute quadrangle, Braxton County), in roadcut to east of northbound highway I–79 are exposed a thin Pittsburgh coal overlying a noncalcareous green- to red-mottled paleosol and overlain by a calcareous red-mottled paleosol with mukkara structure, all Monongahela Formation (Late Pennsylvanian): examined March 2, 1991.
9	Sissonville, W. Va. (grid reference 445662, Sissonville 7.5-minute quadrangle, Kanawha County), roadcut south of frontage road on hill in southeast quadrant of cloverleaf at intersection of highway I–77 and Haines Branch Road exposes a thin gray claystone equivalent to the Pittsburgh coal, here forming the surface horizon of a noncalcareous red-mottled paleosol and overlain by a calcareous red-mottled paleosol with large cradle knolls and root traces, also of the Monongahela Formation (Late Pennsylvanian): examined March 2, 1991.
10	Holliday, Kans. (center sec. 6, T. 12 S., R. 24 E., Edwardsville 7.5-minute quadrangle, Johnson County), roadcuts to the east of southbound highway I–435, 1 mile east of Holliday, expose a maroon-colored noncalcareous paleosol with woody root traces overlain by the lower portion of a gray calcareous paleosol all in the upper part of the Bonner Springs Shale, which is truncated by the marine Merriam Member of the Plattsburg Limestone, Lansing Group (Late Pennsylvanian, Missourian): examined June 25, 1991.
11	Macksburg, Ohio (SW¼SE¼NW¼ SE¼ sec. 18, T. 5 N., R. 8 W., Lower Salem 7.5-minute quadrangle, Washington County), roadcut 1 mile south of Macksburg on highway I–77 exposes noncalcareous underclay to the Sewickley coal, which is overlain by a red-mottled calcareous paleosol with mukkara structure, all Monongahela Formation (Late Pennsylvanian): examined March 3, 1991.
12	Clinton Lake Spillway, Kans. (NW¼NE¼ NW¼ sec. 8, T. 13 S., R. 19 E., Lawrence West 7.5-minute quadrangle, Douglas County), in northern bank of spillway is exposed the Oread Limestone of the Shawnee Group overlying the Lawrence Formation of the Douglas Group, the latter including a thin Williamsburg coal overlying two thick gray- to red-mottled calcareous paleosols with deep cracks and large casts of woody roots and stumps (Late Pennsylvanian, Virgilian): examined June 25, 1991.
13	Howard, Colo. (grid reference 248578, Howard 7.5-minute quadrangle, Fremont County), in roadcut east of highway U.S. 50 and the Arkansas River, 2 miles west of Howard, are exposed steeply dipping red paleosols with carbonate nodules in the Sangre de Cristo Formation (Pennsylvanian to Permian): examined May 13, 1979.
14	Coaldale, Colo. (grid reference 312514, Howard 7.5-minute quadrangle, Fremont County), in roadcut south of highway U.S. 50 and Arkansas River, 4 miles west of Coaldale, are steeply dipping red paleosols with carbonate nodules in the Sangre de Cristo Formation (Pennsylvanian to Permian): examined May 13, 1979.
15	Gateway, Colo. (NW¼NE¼NW¼SE¼ sec. 21, T. 51 N., R. 19 W., Gateway 7.5-minute quadrangle, Mesa County), low in cliffs above the dry wash 1 mile southwest of Gateway are thin red calcareous paleosols with abundant small root traces in the Cutler Formation (Permian): examined May 14, 1979.

Geochemical Variations in Pennsylvanian Black Shales May Reflect Changes in Climatic Conditions

Raymond M. Coveney, Jr.

Besides coals, which long have been of economic importance, perhaps the most distinctive rock type formed during the Pennsylvanian is the black shale that commonly overlies coal. Bertram Woodland (in Zangerl and Richardson, 1963) was first to report the extraordinary metal contents, for metals other than uranium, contained by thin Pennsylvanian black shales of the central United States. In the black portions of the thin (<1 m) Mecca Quarry Shale Member of the Linton Formation of Indiana, Woodland noted the presence of many metals in enriched amounts including up to several thousands of parts per million molybdenum, which is within the range of typical grades for conventional Climax-type ores. Other heavy elements that are commonly enriched above average crustal values in thin organic-rich shales of the Pennsylvanian include zinc and vanadium, which commonly occur at the thousands of parts per million level, and selenium, cadmium, and uranium at the tens to hundreds of parts per million level. Phosphate contents are high in some shales. Recent geochemical studies include those of Vine and Tourtelot (1970), Cubitt (1979), Desborough and others (1990), Schultz (1991), and the author and colleagues. Coveney and others (1991) discerned at least two types of thin black metal-rich shale in the Pennsylvanian. One type, similar to the Mecca Quarry Shale Member, consists of molybdenum-rich, nonphosphatic shale containing more than 20 weight percent organic carbon, mainly in terrestrial organic matter; the other type, similar to the phosphatic Heebner Shale Member of Kansas, contains less molybdenum and less organic matter chiefly of marine origin. Mecca-type shales are inferred to have originated near an ancient shoreline lined by peat swamps and may be restricted to the Middle Pennsylvanian; Heebner-type shales are thought to have formed mainly offshore.

Organic matter, an abundant component of the shales in all cases, is of mixed origin. Wenger and Baker (1986) found a tendency for terrestrial-type organic matter to be especially concentrated in the lowermost laminae of one shale. Marine organic matter becomes relatively more abundant near the top of the same unit, implying a lesser input of terrestrial organic debris after the initial transgression of epeiric seas that inundated peat swamps. Coveney and others (1987) found more terrestrial-type organic matter nearshore for one Middle Pennsylvanian shale in Indiana and mainly marine-type organic matter in its stratigraphic equivalent offshore. Nevertheless, Desborough and others (1990) found some samples from thin laminae to contain a predominance of terrestrial organic matter in an Upper Pennsylvanian core shale in eastern Kansas in an area likely to have been far from the ancient shoreline. All three sets of investigators have examined only a few samples from the numerous black shales that occur in the Pennsylvanian of the Midwest; however, since much of the organic matter was derived from land (Coveney and others, 1987; Desborough and others, 1990), it would not be surprising if the drastic decline in the proportions of woody and semiwoody plants, which Phillips and Peppers (1984) note at the transition from the Middle Pennsylvanian (Desmoinesian) to the Upper Pennsylvanian (Missourian), was reflected in a diminished content for terrestrial organic matter in Upper Pennsylvanian black shales, even though they would remain as one of the main repositories of organic carbon.

Despite extensive geochemical efforts concerning inorganic geochemistry to date, much research remains to be done on midwestern Pennsylvanian black shales. According to the maps of Wanless and Wright (1978), there are at least 15 thin shales containing black facies, each extending over more than 40,000 km^2. Most of these shales are probably metalliferous, but this is not known with certainty because not all have been analyzed by quantitative techniques. The total concentrations of metals may have been proportional to the duration of starved basin conditions; if so, it would be important to understanding the history of sedimentation in epeiric seas. Basinal brines, implicated in the formation of Mississippi Valley-type (MVT) ores that, according to paleomagnetic studies (Symons and Sangster, 1991), formed near the end of the Pennsylvanian (Kiaman), may have provided metals for Pennsylvanian black shales (Coveney, in press).

Grossman and others (1991) have painstakingly examined the oxygen isotopes in carbonates of the Pennsylvanian and obtained interesting results. However, comprehensive isotopic investigations remain to be done. Coveney and Shaffer's (1988) preliminary analyses of 77 samples of pyrite and sphalerite separated from black shales suggest a −15 per mil shift from an average $\delta^{34}S$ value of −12.9 per mil (relative to the troilite of Canyon Diablo) in Desmoinesian beds to an average of −27.2 per mil for Missourian shales, which they suggested may reflect deepening of the Pennsylvanian seas. Other interpretations include vegetation changes in the source area, changes in the input rate of clastics, or intensification of hydrothermal activity as possible alternatives. The data for Missourian black shales are sparse (seven samples from two shales, all in the Kansas City area).

The black shales of the Pennsylvanian are likely to have recorded much of the history of the basins in which they lie. Throughout the Pennsylvanian, the epeiric basins have been affected by distant tectonic effects (Klein and Willard, 1989), changes in climate (Cecil, 1990), changes in vegetation (Phillips and Peppers, 1984), changes of sea level resulting from the waxing and waning of glacial epochs (Boardman and Heckel, 1989) or from global tectonism (Sloss, 1991), and migrating basinal brines (Bethke and Marshak, 1990). Each of these processes may have had a discernible impact on black shales. For example, variations in sediment flux as a result of rainfall patterns may have induced changes in sedimentation style causing the development of cyclothems and their black shales without the need for eustatsy (Cecil, 1990). Comprehensive studies of fresh samples of equivalents of black shales may help to resolve these questions of the importance of terrestrial processes and should also help to illuminate larger questions concerning the evolution of the broad basins that contain metalliferous black shales and climate changes during the Pennsylvanian.

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Chemical and Mineral Variations in Pennsylvanian Black Shales—Depositional and Diagenetic Indicators in Marine Evaporite Cycles, Hermosa Formation, Paradox Basin, Utah

Gene Whitney, Michele L. Tuttle, Timothy R. Klett, Dirck E. Tromp, and Mark Richardson

The clay mineralogy and the distribution of sulfur, carbon, and iron in black shales of the Pennsylvanian Hermosa Formation, Paradox basin, Utah, record depositional and diagenetic processes affecting petroleum source rocks in marine carbonate-evaporite black shale cycles. The Paradox Member of the Hermosa Formation contains a total of 29 evaporite cycles, which are correlative across the basin and reflect changes in water level driven by tectonics, climate, and (or) freshwater recharge. Black shales in cycles 3 and 5 were analyzed to detect mineralogical and geochemical trends related to depositional environments. Black shales from one core were deposited in a shelf environment (shelf facies; State core). In two other cores equivalent shales were deposited in an evaporite basin and are interbedded with anhydrite (saline facies; Norton core) and halite

(hypersaline facies; Shafer core). The following data were acquired: bulk and clay mineralogy; abundance of sulfur, carbon, and iron species; isotopic composition of sulfur in sulfate (SO_4), disulfide, and organosulfur; and hydrogen and oxygen indices of the organic matter.

The so-called black shales, deposited during times of maximum transgression and the beginning of regression, contain significant amounts of dolomite and quartz, as well as clay minerals and organic matter. The predominant clay mineral in the shelf facies of the black shale is illite (or illitic interstratified illite-smectite). In more saline parts of the basin, interstratified chlorite-smectite (C–S) predominates. The C–S varies systematically in the proportions of chlorite and smectite layers in the structure; within a particular black shale, the most highly expandable C–S is commonly found near the center of the shale, and some shales exhibit more than one expandability maximum (fig. 1). The C–S near the contacts between the black shale and enclosing units is richer in chlorite layers. There is some variation in the relative abundances of illite and C–S, but these variations are not correlative with the expandability of the C–S.

We attribute the systematic changes in the expandability of C–S to variations in salinity of the water during the transgression-regression cycle. During deposition, detrital clay (primarily illite) was altered by contact with the Mg-rich saline or hypersaline brines into C–S or some Mg-rich clay precursor. This reaction is observed in modern hypersaline environments. The expandability (proportion of smectite layers) of the C–S is greatest when the brine is most dilute (lowest salinity) because pH is lowest and Mg concentrations are relatively low, thus retarding the formation of chlorite layers. As salinity and pH increase with increasing evaporation, more chlorite layers are formed in the C–S. Thus, the C–S appears to be a sensitive indicator of brine composition and may point to the time of maximum transgression or freshening due to increased rainfall or some other mechanism. The amount of detrital illite that reacts to C–S soon after sedimentation may be related to the initial amount of detrital clay or the amount of time in contact with the brine rather than the salinity or Mg content of the brine. Furthermore, the preservation of expandability trends in the C–S suggests that postdepositional changes in clay mineralogy have been relatively minor.

The abundances of organic carbon, hydrogen indices, and types and abundance of sulfur species in the black shales also reflect changes in the depositional environment. The amount of organic carbon in the evaporite-basin shales is greater than in those from the shelf and increases toward the basin center where the most hydrogen-rich organics are found (fig. 2). Hydrogen indices in the shales from the hypersaline facies systematically increase during transgressive deposition (fig. 2). Disulfide contents in the shales are proportional to reactive-iron contents (fig. 3) and are greatest in argillaceous shales from the transgressive stage (fig. 2). Although sulfur-containing phases become systematically enriched in ^{34}S during transgressive deposition (fig. 2), they are depleted in ^{34}S relative to the bedded anhydrite between the black-shale intervals ($\delta^{34}S$ values between +12 and +16 ‰).

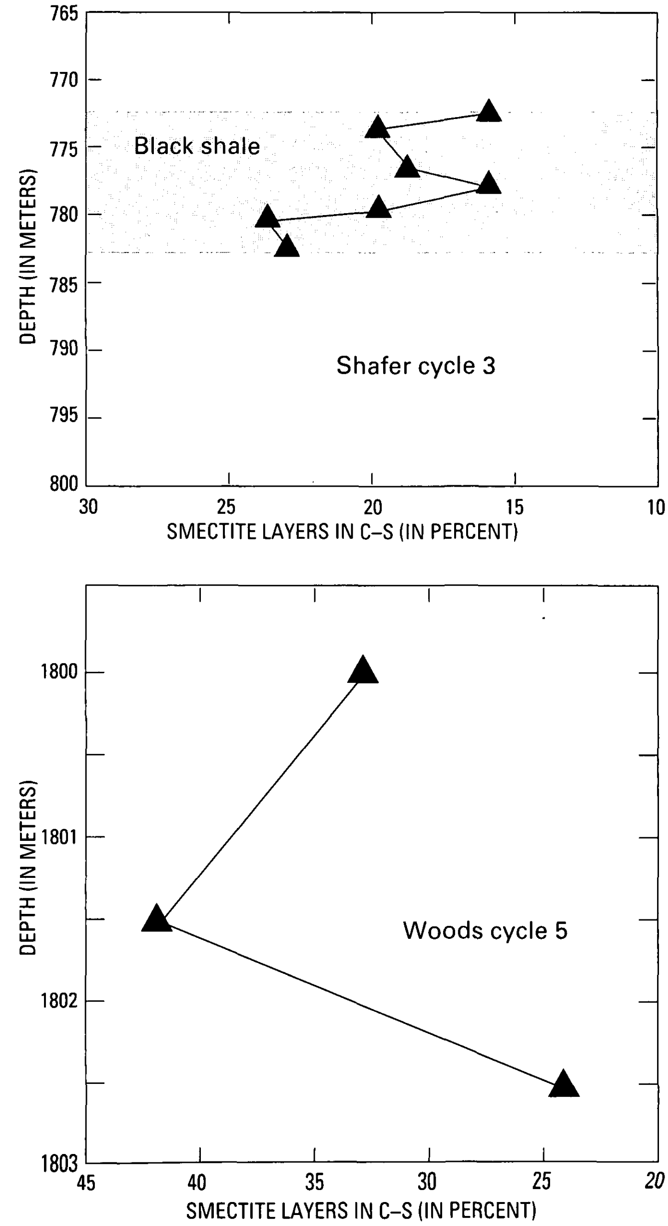

Figure 1. Distribution of expandability of interstratified chlorite-smectite (C–S) clays in black shale. Maximum expandability corresponds to the time of maximum transgression or brine dilution. The Woods drill hole is close to the Norton hole.

Differences in carbon, sulfur, and iron geochemistry among the depositional environments and within the transgressive-regressive cycles reflect differences in organic-matter preservation, biological productivity rates, the source of clastic material and its affect on iron sulfidization, and sedimentation rates. The sulfur, carbon, and iron data indicate that (1) sulfide mineral formation was iron

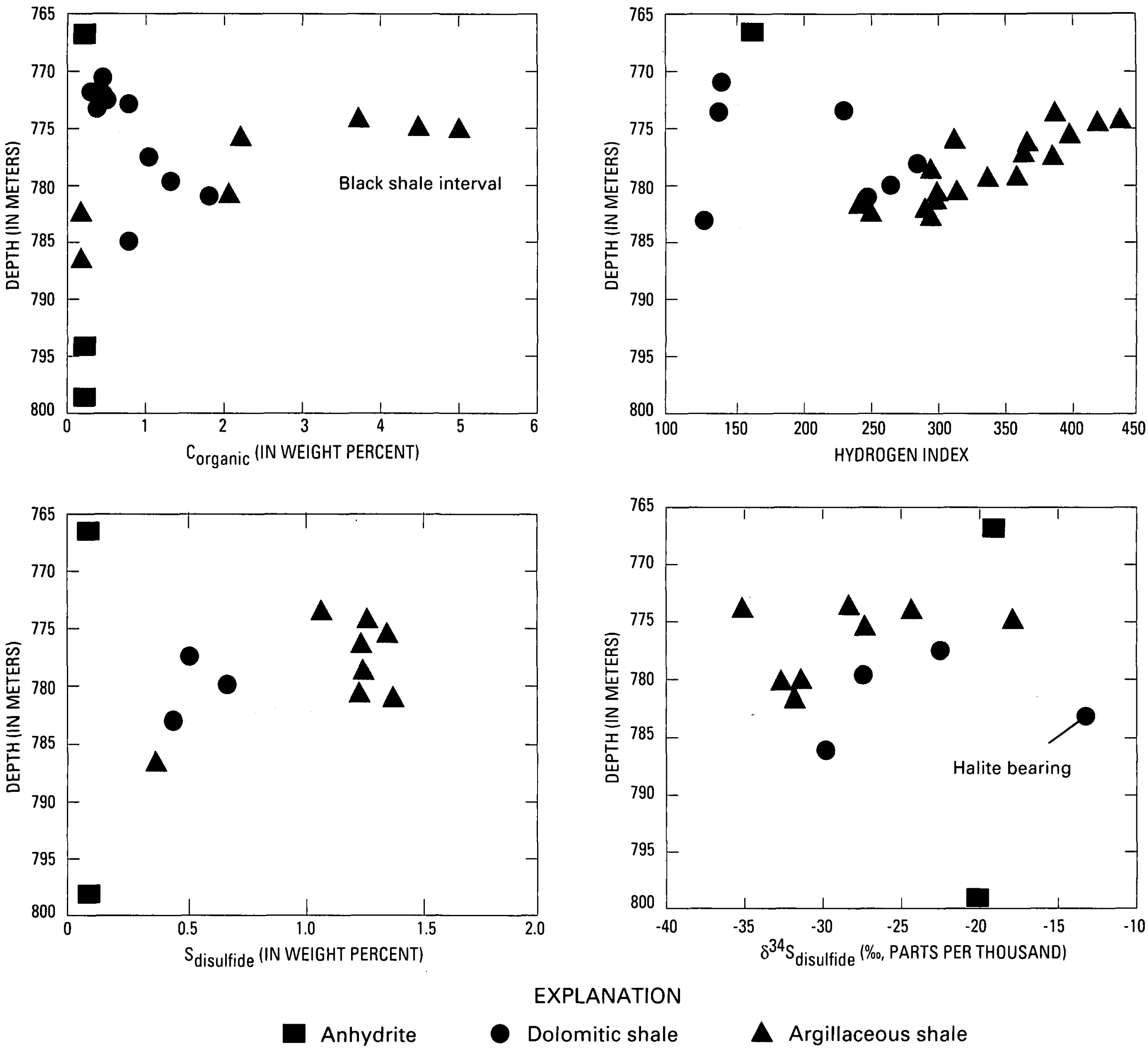

Figure 2. Depth profiles of $C_{organic}$ concentrations, hydrogen indices, $S_{disulfide}$ concentrations, and disulfide isotopic compositions in cycle 3 black shales from the hypersaline facies. Maximum transgression estimated to be about 775 m depth.

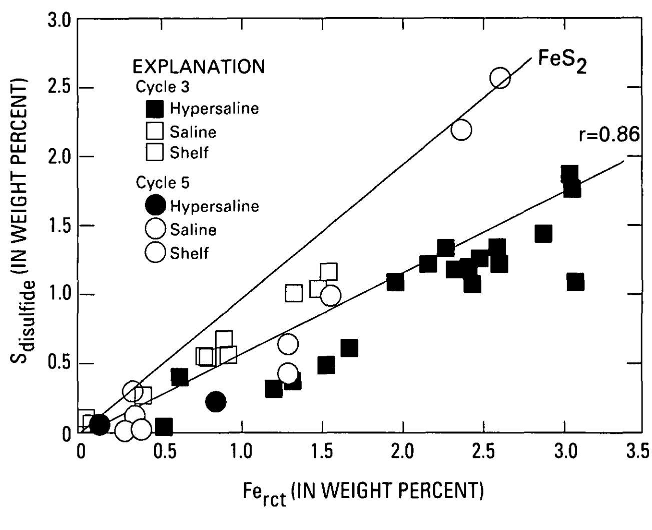

Figure 3. Plot of $S_{disulfide}$ concentrations versus reactive iron (Fe_{rct}) concentrations of two black shale intervals (cycles 3 and 5) from three depositional environments.

limited, (2) more organic matter accumulated during transgressive stages of shale deposition, (3) organic matter was better preserved in the evaporite-basin facies, and (4) the smallest ^{34}S depletion of sulfide coincided with dilution of the brines because the sulfate concentrations decreased relative to the amount of sulfide produced (H_2S:SO_4 ratio increased). These hypotheses are supported by systematic trends in the geochemistry of carbon and sulfur within individual transgressive-regressive cycles and are consistent with documented effects of these cycles on geochemical processes involving carbon, sulfur, and iron.

Contributions of Dependent and Independent Paleontological Data to Predictive Stratigraphic Analysis

Christopher G. Maples and Ronald R. West

Paleontology can contribute to predictive stratigraphic analysis through the use of data that fall into two general categories: dependent and independent. Dependent paleontological data result from biotic responses to externally mediated physical parameters such as climatic fluctuations, sea-level change, and sediment accumulation rate. Therefore, dependent paleontological data are constrained by, or dependent on, the boundaries of the stratigraphic units under consideration. Independent paleontological data are generated from taxonomic and evolutionary investigations. Therefore, independent paleontological data are independent of any lithostratigraphic or genetic- or sequence-stratigraphic boundaries; indeed, they are the basis of biostratigraphic boundaries. These two types of paleontological data address very different questions. Dependent paleontological data address questions such as "What were the biotic responses to climate change, water temperature fluctuations, marine transgression, or episodic sedimentation?" Thus, dependent paleontologic data cannot be used to test chronostratigraphic correlations but can be used as additional information for making such correlations. Independent paleontological data address questions such as "Are these units the same age?" Thus, independent data are the only data that can be used to test chronostratigraphic correlations.

The goal of any biostratigrapher is to make more refined correlations. Nonetheless, biostratigraphy, as the cornerstone of independent paleontologic data, has lagged behind high-resolution sequence and event-genetic stratigraphy in its ability to provide high-resolution correlations over wide areas. Within-basin correlations attain higher resolution than interbasinal correlations, and interbasinal correlations attain higher resolution than intercontinental correlations. If the ultimate goal of predictive stratigraphic analysis is to correlate on an intercontinental basis, then subsurface-compatible biostratigraphy must be taken into account (for example, use of last appearance datums or LAD). Most biostratigraphers use their own particular fossil group for correlation. Zonation schemes that use multiple fossil groups are uncommon and, when published, often rely on collaboration, which may force compromise. However, a holistic biostratigraphy can result in a better overall biostratigraphic framework. But just naming more new taxa is not the answer. Quantitative biostratigraphy (graphic correlation, morphometrics, ranking and scaling, and probabilistic stratigraphy) is the future of high-resolution biostratigraphy. These types of analyses must use populations of taxa, numerous sections, and numerous measurements. Industry, because of their access to larger data sets, is far ahead of most other biostratigraphers in this approach. More powerful still is the combination of biostratigraphy with datable chronostratigraphic indicators such as tonsteins, magnetic reversals, or chemical events.

Two examples of dependent paleontological data are trace fossils (which have little biostratigraphic utility and no high-resolution biostratigraphic utility) and epiboles (three-dimensional biotic events). Trace fossils almost invariably are found where they were formed; therefore, some of the taphonomic problems associated with transport, size sorting, and burial are avoided. Trace fossils in the Paleozoic are especially useful indicators of marine-flooding sequences in nonfossiliferous siliciclastic units. Trace fossils used in this way can be studied by using both outcrop and core. Epiboles can form through different processes; for instance, myalinid clam epiboles in the Pennsylvanian of Kansas form through a time-averaging effect that results in accurate tracking of nearshore facies. Coral epiboles, also in the Pennsylvanian of Kansas, form in predominantly siliciclastic packages when siliciclastic input decreases. Thus, trace fossils and epiboles in general, and coral epiboles in particular, may be sensitive paleoclimatic indicators.

The old computer adage "garbage in, garbage out" is particularly applicable to paleontological contributions to predictive stratigraphic analysis. Independent paleontological data are only as good as the taxonomy and amount of input. Dependent paleontological data are only as good as the controls that enable a paleontologist to recognize the biotic events. In short, there is absolutely no substitute for detailed, fine-scale sampling and evaluation. High-resolution correlation does not come from low-resolution sampling schemes. Paleontology has much to offer along these lines but only through careful, detailed work.

Tectonic Framework of the Appalachian Basin

Robert C. Milici

The Appalachian basin, an elongate asymmetric synclinorium, extends from Lake Ontario southward for 1,600 km through New York, Pennsylvania, Ohio, West Virginia, eastern Kentucky, Virginia, Tennessee, and northwestern Georgia to Alabama. The basin extends from the Appalachian Blue Ridge westward to the Cincinnati arch, where its fill of Paleozoic strata ranges from 600 to 900 m thick. On the east, basin fill is thickest in central Pennsylvania, where it exceeds 13,700 m.

Geophysical evidence indicates that the eastern edge of the Appalachian basin lies buried beneath crystalline thrust sheets from New England southwestward into the southern Appalachian Piedmont of Virginia, the Carolinas, and Georgia (Cook and Oliver, 1981; Harris and others, 1982). Farther to the southwest in Mississippi, Alabama, Georgia, and South Carolina, Appalachian magnetic trends are terminated by the Charleston magnetic terrane (Higgens and Zietz, 1983). A tectonic suture occurs nearby between the North American craton to the north and rocks with African affinities to the south; the suture lies astride Mesozoic synrift basins beneath the Atlantic Coastal Plain (Chowns and Williams, 1983). In central Mississippi, carbonate strata of the Appalachian platform terminate along a buried shelf edge and are replaced in western Mississippi and Arkansas by basinal siliciclastic deposits of Paleozoic age.

Strata of the Chilhowee Group and Shady Dolomite (Early Cambrian) record the opening of the Iapetus Ocean and thermal subsidence of the passive margin of eastern North America (Pfeil and Read, 1980; Read and Pfeil, 1983; Walker and Simpson, 1991). Inundation of the craton was accompanied by widespread carbonate deposition of the Knox and Beekmantown Groups and their equivalents in Late Cambrian and Early Ordovician time. Regional uplift and erosion of these carbonate strata occurred over all but the deepest parts of the basin, with the development of a major regional unconformity at the beginning of Middle Ordovician time (Bridge, 1955; Milici, 1973; Harris and Repetski, 1983). This uplift marks the onset of renewed tectonic activity along the continental margin and the beginning of the Taconic orogeny (Rodgers, 1971), when the continent may have collided with a magnetic arch that developed above an eastward-dipping subduction zone (Wagner and others, 1991). Later in the Ordovician, tectonic lands were thrust upward along the eastern margin of the Appalachians by the collision and shed siliciclastics, some very coarse grained, into several adjacent subsiding foreland basins. These basins extend along the western side of the Valley and Ridge from Alabama to Pennsylvania (Rodgers, 1953; Kellberg and Grant, 1956; Kreisa, 1981; Shanmugam and Lash, 1982; Rader and Gathright, 1986).

The final episode of the Taconic orogeny is marked by the widespread deposition of the red beds and shales of the Queenston delta. These sediments also were shed westward from the tectonic uplands and then spread southward, blanketlike, over much of the Appalachian basin (Dennison, 1976).

As the mountains continued to wear away during the early part of the Silurian, more siliciclastic rocks were carried westward to where they were deposited in alluvial fans on alluvial plains and in coastal environments (Cotter, 1983). Later in the Silurian, an area extending from the western part of the Appalachian basin to the Michigan basin became restricted on the east by these encroaching deltaic deposits and on the west by carbonate banks and reefs. The restricted circulation, and perhaps the influence of climatic variations, resulted in accumulations of thick evaporites in this area (Alling and Briggs, 1961; Richard, 1969; Smosna and Patchen, 1978).

The Acadian orogeny began in Middle Devonian time when a shallow, slowly subsiding foreland basin formed and began to fill with sediments. This basin extended generally from southeastern New York through West Virginia, western Virginia, eastern Kentucky, and Tennessee into northwestern Georgia and Alabama. In general, the Acadian delta is a great fill sequence consisting of black shales, siltstones, sandstones, and turbidites and exceeding some 3,600 m in thickness in eastern Pennsylvania (deWitt, 1975). The delta thins to the southwest to a nondepositional wedge in southern Tennessee and adjacent Alabama. Lower Mississippian formations occupy the top of the deltaic mass and occur in two lobes: the Pocono delta in eastern Pennsylvania, adjacent West Virginia, and northwestern Virginia, and the Price delta in southwestern Virginia, eastern Tennessee, and eastern Kentucky. Coal occurs in the upper part of the Price and, indeed, fueled the ironclad *Merrimac* (*Virginia*) during the War Between the States (Kreisa and Bambach, 1973).

The Price is capped by the red beds and evaporites of the Maccrady Formation in southwestern Virginia, which perhaps reflects both the formation of an isolated, restricted basin and the changing climatic conditions of the time. Acadian tectonism may have been caused by collision of an ancient landmass, Armorica, with ancestral North America and with northern European crust (Perroud and others, 1984).

A great marine transgression occurred in the Late Mississippian when thick sequences of marine carbonate strata accumulated over much of the basin. In the northeast, thousands of meters of red beds (Mauch Chuck Formation) were deposited in eastern Pennsylvania as they were shed from uplifting sedimentary and metamorphic provinces in the Piedmont nearby (Hatcher and others, 1989). Similarly, great thicknesses of Mississippian and Pennsylvanian siliciclastic sediments were introduced into the Pocahontas basin of southwestern Virginia from whence they spread northwestward through West Virginia into Pennsylvania (Ferm and Cavaroc, 1969) and southwestward into Tennessee (Milici and others, 1979). Almost everywhere, vertical sequences extend from red and green shales and carbonate rock at the base through orthoquartzites interpreted to be beach-barrier deposits into deltaic deposits, which are characterized by the predominance of subgraywacke sandstones. Thick orthoquartzites containing few coal beds commonly are overlain by subgraywacke sequences containing abundant coal beds.

A third source for the siliciclastic sediments of the southern part of the Appalachian basin lies to the south beneath the Coastal Plain of Alabama and Mississippi and to the west off the edge of the buried carbonate platform in

Mississippi and Arkansas. From this source, shales and turbidite sequences of the Floyd and Parkwood Formations spread generally northeastward into Alabama and southernmost Tennessee and grade upward first into orthoquartzites and then into graywacke sandstones and conglomerates (Ferm and Ehrlich, 1967).

Throughout the Appalachians, the lithologic variations and stratigraphic sequences of the Carboniferous reflect both the effects of climate, geologic terrane, topography, and tectonics during the formation of the sediments in the source area and the imprint of climate, tectonics, and depositional environments on the sediments within the basin of deposition.

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Carboniferous Paleoclimates, Sedimentation, and Stratigraphy

C. Blaine Cecil, Frank T. Dulong, N. Terence Edgar, and Thomas S. Ahlbrandt

Long- to short-term paleoclimate changes (table 1) were primary controls on changes in chemical and mechanical weathering, terrestrial organic productivity, and sediment transport to both epicontinental and continental margin depocenters during the Carboniferous across the United States. Changes in Carboniferous climates were primarily governed by (1) zonal atmospheric circulation as North America moved northward across paleolatitudes, (2) orbital forcing, and (3) orographic controls. Climatic controls on stratigraphy are, therefore, recorded both temporally and spatially from regional to continental scales.

During much of the Mississippian, eastern North America was in or near the Southern Hemisphere high-pressure cell, the dry tropics (Scotese, this volume), and was, therefore, relatively dry as evidenced by evaporites, carbonates, and paleo-Vertisols and paleo-Aridosols. These soil orders are indicative of dry yet somewhat seasonal conditions in which evapotranspiration exceeds precipitation for most months of the year. During the Mississippian, the western United States was climatically drier than the eastern United States as evidenced by evaporites and marine carbonates. By the Late Mississippian much of the western United States was near the paleoequator (Scotese, this volume).

During the latest Mississippian and earliest Pennsylvanian, much of North America was exposed during a pronounced lowstand in sea level, which resulted in the middle Carboniferous unconformity. Prevailing winds during this time were probably from east to west. These winds apparently lost much of their moisture in the east as there is an east to west gradient in types of paleosols on the unconformity. Ultisols, which form under wet climatic conditions (rainfall exceeds evapotranspiration for most or all months of the year), occur at the unconformity in the east, whereas Aridosols occur in the west.

Long-term climate change continued in the Early and early Middle Pennsylvanian. The climate of eastern North America became increasingly wet and less seasonal as it moved northward into the low-pressure paleoequatorial rainy belt (Cecil and others, 1985; Cecil, 1990). Short- to intermediate-term climate cycles were also important during this time as evidenced in the Appalachian basin of eastern North America where coal beds and coeval, chemically weathered, upland paleosols (Ultisols) record the more pluvial periods of climate cycles. During the pluvial parts of climate cycles, increased terrestrial organic productivity restricted erosion and, hence, restricted siliciclastic and dissolved load input from fluvial systems. Drier and more seasonal parts of these climate cycles resulted in increased siliciclastic flux to terrestrial and marine depocenters in the Appalachian basin (Cecil, 1990). Pennsylvanian source rocks in the eastern United States are gasprone due to the climatically induced high input of terrestrial organic matter.

By the late Middle Pennsylvanian (Desmoinesian), a long-term dry climate is recorded in the western United States by sand seas, marine carbonates, and evaporites. However, pluvial periods that were part of intermediate- and short-term climate cycles (1) stabilized dune fields through increased terrestrial organic productivity, (2) increased fluvial siliciclastic flux, and (3) affected circulation in epeiric seas where oilprone, black shale source rocks were deposited. Extremely dry conditions existed in the Paradox basin of southeastern Utah during the Middle Pennsylvanian as evidenced by evaporites including potash. The extreme aridity appears to be the result of both paleolatitudinal location of the region in the dry tropics and a rain shadow effect of the Uncompahgre uplift.

Table 1. Tropical and subtropical climate-change classification.
[Modified from Cecil, 1990]

Relative duration	Cause	Time (years)
Long term..........	Movement of continents across latitudes; orogenesis, "greenhouse" gases (?)	10^6–10^8 10^5–10^7
Intermediate term.	100,000- and 400,000-year cycles of orbital eccentricity, "greenhouse" gases (?)	10^5
Short term	Cycles in axial tilt and precession	10^4
Very short term...	Solar variation (?)	10^3
Instantaneous	Weather systems	10^{-2} (months, weeks, days, hours).

In addition to long-term climate change associated with the northward movement of North America, short- and intermediate-term climate cycles (table 1) had a pronounced effect on chemical (dissolved inorganic and organic sediment load) and siliciclastic sediment flux in fluvial systems. For example, Late Pennsylvanian paleoclimate cycling was a major factor that controlled sedimentary cycles in the Appalachian basin. The effects of climate cycles are recorded stratigraphically by geochemical signatures such as paleosols, coal beds, and nonmarine limestones. The wetter phases of paleoclimate cycles are recorded as laterally extensive coal beds that were derived from lowland topogenous peat and coeval upland paleosols, the structure, chemistry, and mineralogy of which are similar to modern Ultisols. Ultisols result from leaching by chemical weathering in warm climates where rainfall exceeds evapotranspiration for most months of the year. The coal beds and associated paleosols indicate that rainfall exceeded evapotranspiration for most months of the year. The drier parts of climate cycles are recorded stratigraphically as nonmarine limestone beds that grade laterally into highly calcareous paleo-Vertisols. Brecciation and subaerial-exposure crusts within these limestone beds suggest fluctuation in the water table. A paucity of terrestrial organic matter, the occurrence of nonmarine limestone with multiple subaerial-exposure features, and the characteristics of the coeval upland paleosols are indicative of relatively dry phases of climatic cycles when evapotranspiration exceeded rainfall for more than 6 months of the year.

Paleoclimate change, therefore, appears to have caused changes in weathering, sediment flux, and organic productivity throughout the Carboniferous. Long-term climate change is recorded on a continental scale as North America moved from the dry tropics of the Southern Hemisphere, through the equatorial rainy belt into the dry tropics of the Northern Hemisphere. Climate cycles, which may have been intermediate and short-term in response to orbital forcing, contributed to cyclic stratigraphy as evidenced by paleosols and other climatically sensitive strata.

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Applications of Coal Palynology To Biostratigraphic and Paleoecologic Analyses of Pennsylvanian Coal Beds

Cortland F. Eble

Palynological studies of Pennsylvanian bituminous coal beds fall into two broad categories: biostratigraphic and paleoecologic. As a biostratigraphic tool, spores with restricted stratigraphic ranges, and occasionally the abundances of spore taxa, are used to help define the relative age of coal and coal-related strata. As palynostratigraphic zonations have now been formulated for many of the major coal-bearing basins (fig. 1), one can define the age of a coal bed or coal zone by using several chronostratigraphic nomenclatures. For example, the Fire Clay coal bed of the central Appalachian basin (Breathitt Formation, middle Middle Pennsylvanian) is equivalent with the upper part of the Morrowan Provincial Series of the Eastern and Western Interior basins and the upper part of the Westphalian B Stage of the Maritime coal basins and Western Europe (fig. 1).

Another application of palynostratigraphy is the correlation of individual coal beds or zones both on an intra- and interbasinal scale. The Lower Kittanning coal bed of the northern Appalachian basin (northern West Virginia, southwestern Pennsylvania, and eastern Ohio) is correlative with the No. 6 Block (southern West Virginia) and Princess No. 6 (eastern Kentucky) coal beds of the central Appalachian basin, the Colchester coal bed of the Eastern Interior basin, and the Whitebreast coal bed of the Western Interior basin (fig. 1). Although this application of palynology is more tenuous because of varying environmental and ecologic factors both within and between coal-forming peat systems, it nonetheless is a proven correlation method.

Coal palynology also is used to help understand the paleoecology of ancient swamp floras. This application relies on several factors. First, a majority of the producers (parent plants) of spore taxa from Pennsylvanian coal beds are known to, at least, the group level. Some groups, notably the lycopsids, have received a great deal of study, and spore-plant associations for this Pennsylvanian plant group can be made at the genus and, in some cases, species level (Willard, 1989a, b). Second, the ecological preferences and latitudes of the five major Pennsylvanian plant groups (lycopsids, ferns, pteridosperms, calamites, and cordaites) are partially known (DiMichele and others, 1985). Third, the spore-pollen rain in large modern equatorial swamps, thought to be good analogues for the swamps that produced thick, aerially extensive Pennsylvanian coal beds, is largely autochthonous (Anderson and Muller, 1975). Because of this, the dispersed spore-pollen record accurately reflects the local vegetation, at least qualitatively.

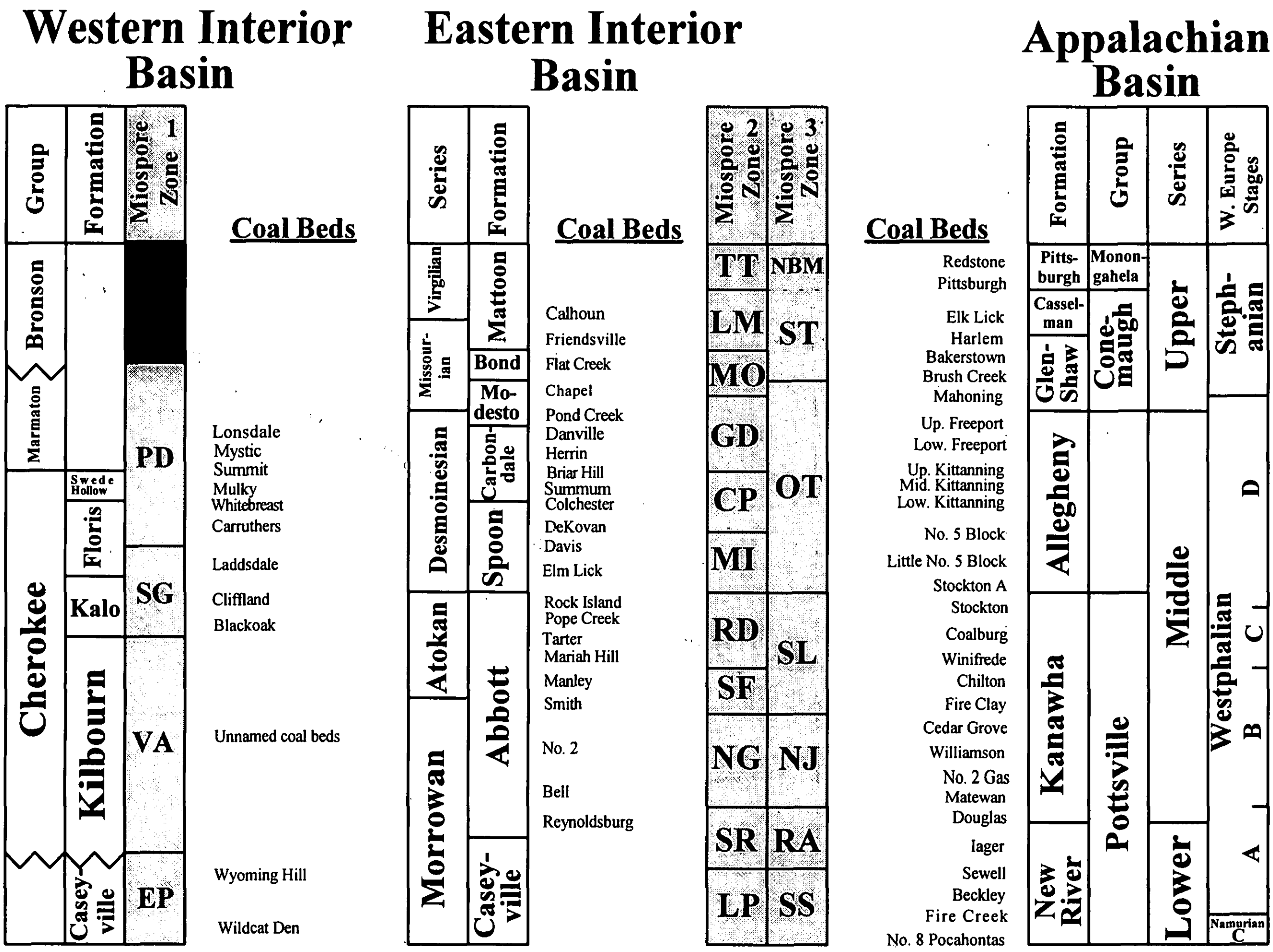

Figure 1. Correlation chart for the Western Interior, Eastern Interior, and Appalachian basins. The placement of the Morrowan-Atokan boundary in the Appalachian basin section is based on invertebrate data (T.W. Henry, oral commun., 1991) and may differ from the boundary placement adopted by other authors (Peppers, 1994). Miospore zone 1 (Ravn, 1986): EP, *Schulzospora elongata, Reticulatisporites papillata;* VA, *Grumosisporites varioreticulatus, Densosporites annulatus;* SG, *Torispora securis, Laevigatosporites globosus;* PD, *Thymospora pseudothiessenii, Schopfites dimorphus.* Miospore zone 2 (Peppers, 1984): LP, *Lycospora pellucida;* SR, *Schulzospora rara, Laevigatosporites desmoinensis;* NG, *Microreticu, latisporites nobilis, Endosporites globiformis;* SF, *Torispora securis, Vestispora fenestrata;* RD, *Radiizonates difformis;* MI, *Cadiospora magna, Mooreisporites inusitatus;* CP, *Schopfites colchesterensis, Thymospora pseudothiessenii;* GD, *Lycospora granulata, Cappasporites distortus;* MO, *Punctatisporites minutus, Punctatisporites obliquus;* LM, *Apiculatasporites lappites, Latosporites minutus;* TT, *Thymospora thiessenii.* Miospore zone 3 (Clayton and others, 1977): SS, *Cirratriradites saturni, Triquitrites sinani;* RA, *Radiizontes aligerans;* NJ, *Microreticulatisporites nobilis, Florinites junior;* SL, *Torispora securis, Torispora laevigata;* OT, *Thymospora obscura, Thymospora pseudothiessenii;* ST, *Angulisporites splendidus, Latensina trileta;* NBM, *Potonieisporites novicus, bharadwajii, Cheiledonites major.*

Combined, these three factors allow us to use coal palynology to assess the types of plants that inhabited peat-forming swamps. We also can make inferences as to the environmental and edaphic conditions under which the plants grew. By applying palynology in this manner, one can begin to see vegetational patterns on interbed (stratigraphic) and intrabed (ecological) scales. The observed patterns are very useful as plants are sensitive indicators to prevailing environmental, especially climatic, conditions.

EARLY AND EARLY MIDDLE PENNSYLVANIAN

Lower and lower Middle Pennsylvanian peat-swamp floras were dominated by large, arboreous lycopsids (*Lepidophloios* and *Lepidodendron*), and pteridosperms (for example, *Lyginopteris*) were a persistent, and in some cases common, accessory plant. The lycopsid trees, except for *Sigillaria*, which probably was not a swamp-centered plant, preferred wet areas (areas of standing water or supersaturated peat substrates) for growth and reproduction. In fact, *Lepidophloios*, which commonly dominates Lower Pennsylvanian coal ball assemblages (Phillips and others, 1985), had developed a reproductive mechanism (*Lepidocarpon*) specifically designed for water dispersal. This suggests that during the Early and early Middle Pennsylvanian, wet to very wet conditions prevailed. It also follows that, because of this type of climate, domed peat swamps, perhaps similar to the ones presently developing in portions of equatorial Indonesia and Malaysia, probably were the dominant swamp morphology. This contention is supported by the overall low ash yields and sulfur contents of Lower and lower Middle Pennsylvanian coal beds (Cecil and others, 1985).

The early Middle Pennsylvanian marks two important events: the increasing abundance of ferns and the occurrence of a "drier" interval (Phillips and others, 1985). Ferns, which are all but absent from Lower Pennsylvanian spore assemblages, became a small but persistent element of swamp floras. This group expanded throughout the rest of the Middle Pennsylvanian and became a major peat-forming group in late Middle Pennsylvanian swamps. During the early Middle Pennsylvanian, calamites and cordaites, plants that may have preferred lowland, more nutrient-rich clastic substrates, are observed to become more common in peat swamps. These two groups continue to be accessory during the Middle Pennsylvanian and often occur in abundance in association with high ash yield coal layers and inorganic partings in coal beds.

MIDDLE MIDDLE PENNSYLVANIAN

The beginning of a trend toward less-wet, more seasonal environments is believed to have started in the middle Middle Pennsylvanian and continued throughout the remainder of the Middle Pennsylvanian. Evidence for this comes from palynology, coal petrography, and coal geochemistry. The expansion of ferns in peat-swamp floras, mentioned previously, may have been a response to drier (less-wet?) conditions that also would have prohibited large-scale arboreous lycopsid expansion and domination. Unlike the lycopsid trees, ferns (especially the treelike varieties, which were the only major peat formers), could flourish in a number of environments, including peat swamps. Ferns apparently could readily occupy areas in peat swamps where lycopsids could not become established (Phillips and others, 1985). This factor may have been especially important during times of a stressed hydrologic budget within the swamps.

Coal petrographic analyses of Middle Pennsylvanian coal beds (Sprunk and others, 1940; Grady, 1979, 1983) show an increase in "splint coal" layers toward the top of the middle Middle Pennsylvanian (top of Kanawha Formation; see fig. 2). Splint coal layers contain increased percentages of inertinite macerals, especially those of inferred degradation (as opposed to fire) origin. These data are interpreted to represent increased frequency of peat surface exposure, oxidation, and the production of peat layers rich in inertinite or inertinite precursors.

Geochemically, these splint coal layers are low in ash yield (<10 percent) and sulfur content (<1 percent). Mineralogically, the ash is composed dominantly of kaolinite and quartz, suggesting that moderate to severe mineral leaching of the peat accompanied periodic exposure of the peat surface. Rainwater flushing of the surficial peat may be one way to accomplish both (aerobic exposure and leaching) simultaneously. If the water table of a domed peat swamp is lowered, perhaps because of less-wet conditions, oxygenated rain water would be able to percolate through peat layers above the water table, oxidizing and leaching the peat at the same time. The result of this would be the formation of inertinite, or preinertinite macerals, and a mineral assemblage composed of the least soluble elements, aluminum and silicon, which are the building blocks of kaolinite and quartz.

LATE MIDDLE PENNSYLVANIAN

The Allegheny Formation (correlative with the Charleston Sandstone) comprises the upper Middle Pennsylvanian in the central and northern Appalachian basin. Coal beds in this formation are transitional in nature. Ones in the bottom contain abundant splint layers and are compositionally similar to those just described from the top of the middle Middle Pennsylvanian. In contrast, coal beds that occur toward the top of the formation are dominantly bright

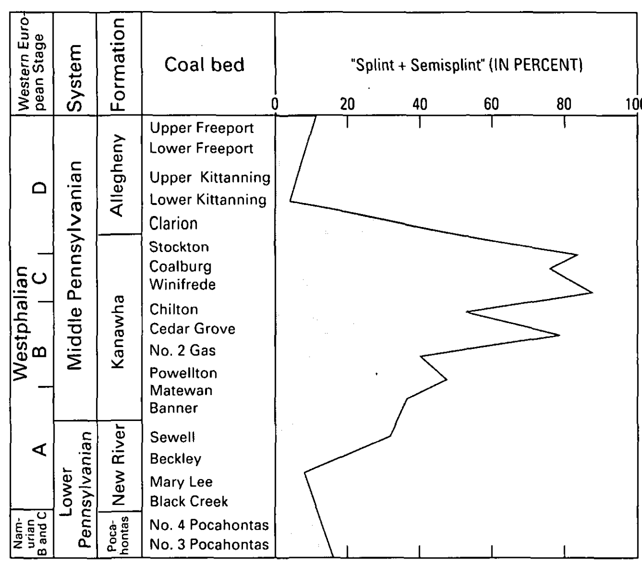

Figure 2. Occurrence of "splint + semisplint" coal in the Appalachian basin. Note the vertical increase in splint coal throughout the Kanawha Formation (Middle Pennsylvanian), followed by a rapid decline in the Allegheny Formation (data from Sprunk and others, 1940).

banded and higher in ash yield and sulfur content (Cecil and others, 1985). This change in coal composition may reflect a more pronounced seasonality in the paleoclimate. A drier climate would favor the development of planar, rather than domed, peat swamps. Planar swamps are considered to promote the development of a bright (vitrinite-rich, inertinite-poor) coal, but one that may contain increased amounts of ash and sulfur (Cecil and others, 1985).

This part of the Pennsylvanian in North America also represents a time of great floral diversity within the peat swamps. All five major plant groups were present and contributed to peat formation (DiMichele and others, 1985). Although arboreous lycopsids are still an abundant, and often dominant, floral component in the swamps, conditions that allowed the proliferation of other plant groups became more prevalent during late Middle Pennsylvanian time.

LATE PENNSYLVANIAN

The inferred drying trend that started during the middle Middle Pennsylvanian culminated at the Middle-Late Pennsylvanian (Westphalian-Stephanian) transition. In North America all but one of the major arboreous lycopsid genera (*Sigillaria*) became extinct, presumably because of their reproductive and morphologic adaptation to areas containing an abundant water supply (Phillips and others, 1974). This void in swamp vegetation was readily filled by tree ferns and, probably to a lesser extent, calamites. As such, swamp floras of the Late Pennsylvanian were different taxonomically and architecturally from those of the Early and Middle Pennsylvanian.

Wet times during the Late Pennsylvanian apparently were wet enough to allow for the development of widespread peat swamps, which in turn gave rise to numerous coal beds of minable thickness. In fact, the most widespread coal bed in the Appalachian basin, the Pittsburgh coal bed, occurs in the Upper Pennsylvanian Monongahela Group (Stephanian B and C). The compositional characteristics of these coal beds suggest that most of the beds were derived from planar peat swamps, as evidenced by their bright (vitrinite-rich) appearance and moderate to high ash yields and sulfur contents. Curiously, these wet times, which resulted in peat formation, are often directly juxtaposed with dry times, indicated by red beds and freshwater limestones. Cecil (1990) has proposed a model stressing a climate component for the origin of sediments. This proposal appears to have direct application in the Late Pennsylvanian by explaining why sediments occurring in a stratigraphic sequence may have highly disparate origins.

SUMMARY

Pennsylvanian coal palynology has application to both biostratigraphy and paleoecology. Spore analyses assist in establishing relative age assignments and in inter- and intrabasinal correlation efforts. As the affinities of a majority of dispersed Pennsylvanian spore genera become known, reconstructions of ancient swamp floras can be partially generated, on both semiqualitative and quantitative bases. By using palynologic data in tandem with coal petrographic and geochemical data, inferences can be made as to the paleoecology of the swamp, the level of preservation or degradation of the peat, and the type of swamp (domed or planar) in which the peat accumulated. These inferences have climatic ramifications because rainfall abundance is the primary driving factor as to whether a swamp will be domed or planar (fig. 3, p. 32).

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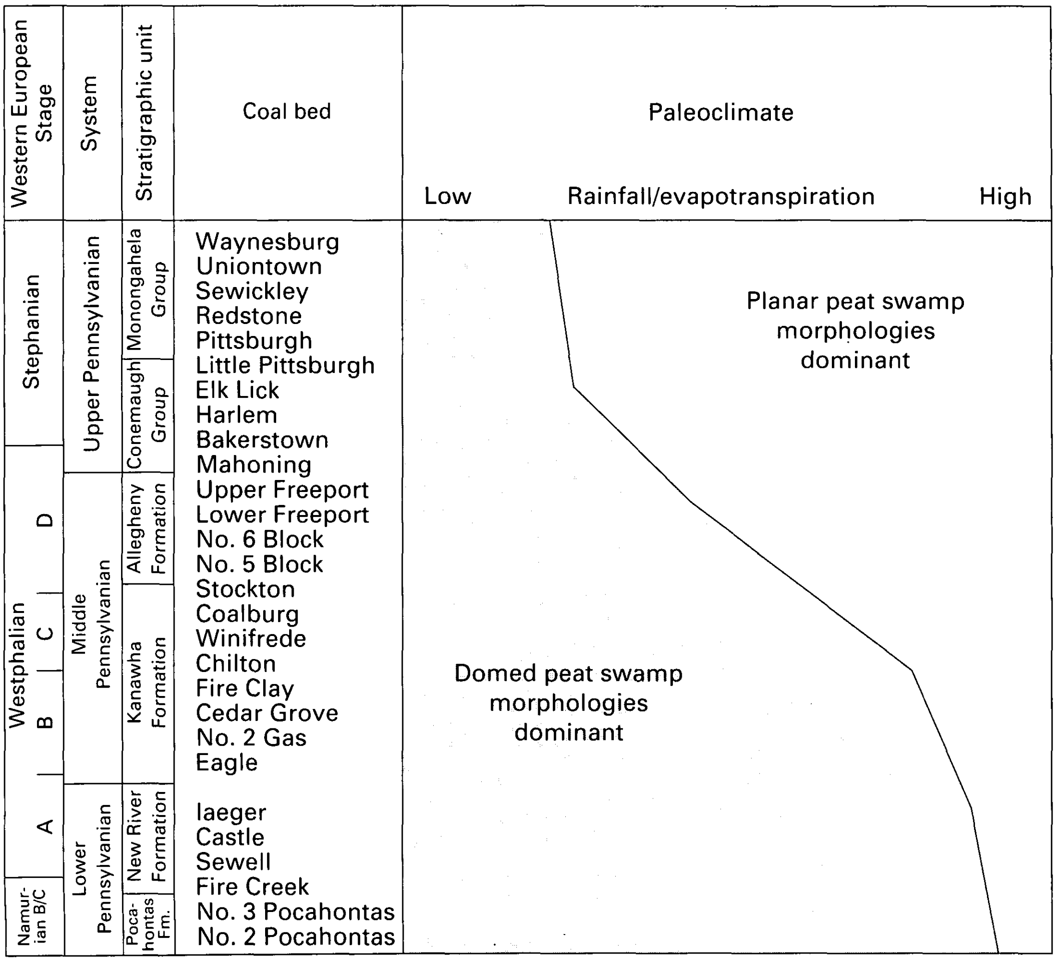

Figure 3. Summary of paleoclimates during the Pennsylvanian. Relative wetness of climate is indicated by vertical line. Note the inferred dominance of domed peat swamp types during the Early through middle Middle Pennsylvanian, reflecting a wet climate, in contrast to the Late Pennsylvanian, which is inferred to have been drier, supporting dominantly planar peat swamp types (modified from Cecil and others, 1985).

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Diverse Factors Controlling Sedimentation in the Northern Appalachians During the Pennsylvanian

Viktoras W. Skema and Leonard J. Lentz

The exact nature of depositional environments of Pennsylvanian rocks in the Appalachian region has always defied easy explanation. A number of different, and in some cases seemingly conflicting, scenarios have been proposed. An initial attempt to explain the obvious repetitive nature of the coal-bearing lithologic sequences in the Illinois basin led to the cyclothem model (Udden, 1912; Weller, 1930; Wanless, 1931; Wanless and Weller, 1932). Weller (1930) and Wanless (1931) both recognized the importance of base level changes to development of the cycles; Weller attributed the fluctuation to tectonism, and Wanless favored eustatic control. The model worked well in the epeiric cratonic basins and quickly gained widespread acceptance. Attempts at applying the cyclothem model to the foreland Appalachian basin proved less successful (Ashley, 1931). Even though a general cyclicity seemed apparent, efforts to formulate representative cyclothems proved to be elusive. The character and sequence of individual lithosomes were too inconsistent to reduce to any simple order.

To provide an adequate explanation of the complex, discontinuous nature of the rocks, later workers focused on the mechanics of sediment emplacement within the basin. Comparison of the autocyclic processes controlling the nature and distribution of sediments in modern-day depositional settings to their Pennsylvanian analogues permitted a more rational explanation of the complex lithologic variation encountered in the Pennsylvanian rocks (Donaldson, 1969, 1974, 1979; Ferm, 1970, 1974). The "deltaic" model developed by these workers, however, did not satisfactorily explain the existence of some of the widespread continuous nonmarine beds, which appear to span an area larger than a typical delta.

Most recently, the focus has shifted back to the broader allocyclic controls required for deposition of such extensive units. Widespread repetitive transgressive-regressive events have been identified in the lower half of the Conemaugh Group covering the northern Appalachian basin (Busch and Rollins, 1984). These workers emphasized the apparent rhythmic nature of these deposits and suggested that the timing of deposition coincided with climate-controlling periodic perturbations in the motion of the Earth. The role of climate as a primary allogenic control has received additional scrutiny, and the apparent alternating repetitive nature of Pennsylvanian chemical and siliciclastic deposits present in the northern Appalachians has been attributed to paleoclimatic cycles (Cecil and others, 1985; Cecil, 1990).

The concepts discussed above contribute greatly to the understanding of Pennsylvanian stratigraphy. Depositional patterns attributable to all of the above-mentioned controls are evident in Pennsylvanian rocks of the Appalachians. This is certainly true in the northern Appalachian basin. In many ways the stratigraphy suggests the presence of wide-ranging factors controlling sedimentation throughout and possibly beyond the basin. Cyclic sedimentation indicative of periodic episodes of transgression and regression similar to the ones found in the midcontinent are present in parts of the section. In Pennsylvania, this allocyclicity is most apparent in the Conemaugh Group. Some of the marine zones in the lower half of the Conemaugh appear to be correlative with similar marine zones in the midcontinent. Their great lateral extent and nearly rhythmic reoccurrence strongly suggest that they are products of eustasy. The regressive siliciclastic component of each of these cycles in the Conemaugh Group is often capped by a relatively thick paleosol. The character of these paleosols changes both laterally and vertically through the section; however, they are often overlain by coal or carbonaceous shale.

These coals reoccur throughout the Pennsylvanian, and their spacing is remarkably even. The interval between major coals or their equivalent carbonaceous horizons is roughly 50 to 80 feet within much of the section throughout most of the bituminous coal fields of Pennsylvania. If, indeed, the same rhythm of deposition is present in both the lower part of the Conemaugh Group, which is marked by thin widespread marine units, and in the more terrestrial parts of the section, such as the upper part of the Allegheny Group, which lacks any trace of marine or brackish fossils, then the presence of a wide-ranging fundamental controlling factor has to be considered. Periodic change in global climate might be such a factor. Cyclical global temperature fluctuations could cause waxing and waning of glaciers, which would produce eustatic changes with resultant transgressions and regressions in coastal areas. The same temperature fluctuations could, at the same time, cause changes in amounts of rainfall and concomitant changes in amounts of plant growth, erosion, and sediment transport regionally, including inland areas unaffected by transgression.

If this were the paramount control operating during the Pennsylvanian, the end result would be deposition of a layered repetitive sequence of rocks. The predominantly marine Pennsylvanian section in the midcontinent closely approximates this condition. The situation in the northern Appalachian basin is considerably different. Individual beds are commonly discontinuous, and facies changes abound. Even key beds, in which horizons are generally recognizable over a wide area, are commonly discontinuous and change in character. This is especially true in the lower part of the Pennsylvanian in the Pottsville and Allegheny Groups. Local stratigraphy can be chaotic. Intrabasinal controls affecting depositional patterns on a local level are evident. For the most part, these controls are ones associated with deposition on a fluvially prograding coastal plain.

Fluvially deposited sandstones are present throughout the Pennsylvanian. In the Pottsville Group, deposition of these sandstones was predominantly in the form of multilateral sand-filled channels in which volume greatly exceeded that of finer grained overbank deposits. They appear to have been deposited by fast-flowing streams in wide belts that coalesced to form both laterally and vertically extensive bodies. Their maximum grain size is very coarse to pebbly. The coals associated with these Pottsville sandstones are discontinuous and thin. Marine to brackish fossil-bearing shales are also present, but, like the coals, they are scattered and difficult to correlate.

Upward through the sequence, these fluvial sandstones gradually become less dominant volumetrically. Major sandstone bodies become isolated from each other laterally and clearly define linear meander belts bounded by overbank sediments. These fluvial systems were separated by quiet interdistributary areas that received little siliciclastic sediment. Instead they were sites of quiet embayments and lakes in which low-energy deposits such as clay, lime, and peat accumulated. A good example of this setting occurs high in the Pennsylvanian rocks overlying the Pittsburgh coal in southwestern Pennsylvania. In places the coal is overlain by a thick sandstone that aerially occurs along a sinuous to linear belt several miles wide (see isopach map in Roen and Kreimeyer, 1973). Its pattern of distribution and internal geometry indicates deposition by a north to northwest flowing river. Downcutting by the river is clearly evident in places where the Pittsburgh coal has been scoured and replaced by sandstone. The sandstone is bounded by a band of fine-grained overbank deposits. A few miles to the west the lateral equivalent of this channel deposit comprises a thick deposit of lacustrine limestones and calcareous claystones (see columnar section *B–B'* in Roen and Kreimeyer, 1973).

The fluvial sandstone deposits most commonly overlie coal horizons but do occur elsewhere in the section. In places they are laterally adjacent to coals, and in some cases deposition appears to be contemporaneous. In these situations major coal seams adjacent to meander belts thin and split in the direction of the sandstones. An occurrence of this type is present in northeastern Greene County and the adjacent part of Washington County (Skema and others, 1982). In this locale the Upper Freeport coal, stratigraphically situated at the top of the Allegheny Group, is a roughly pod-shaped deposit, much of which is greater than 60 inches thick. Some of the thicker inner portions of the pod contain no partings. Westward of this central area, the coal gradually thins, develops a number of thin partings, and at its western edge grades into a channel deposit of sand and silt. In one place along this western edge the coal splits and is separated by as much as 15 feet of overbank deposits composed of sand and silt. A similar split can be seen on the other side of the channel (fig. 1).

Occasionally, during times of flooding, crevasses opened in the levees allowing fluvial sediment to temporarily spread across the adjacent swamp. These crevasse splay deposits are thickest and most coarse nearest the breach in the levee and become thinner and finer grained as they fan out into the swamp. This type of autogenically controlled deposition was responsible in places for major coal splitting. A dramatic example of such an occurrence can be seen in the Redstone coal in southern Somerset County (fig. 2, p. 36). The lenticular central portion of the splay is composed of sand and silt and has a maximum thickness of 30 feet. The lens is approximately 800 feet wide and abruptly pinches into a thin claystone parting at each end. This thin parting, which represents the distal portion of the splay deposit, continues for an indeterminate distance.

The fluvial processes of meandering, avulsion, and crevassing and the process of compactional subsidence can account for many of the patterns of sedimentation seen on a local scale in the Pennsylvanian of the northern Appalachians. However, these intrabasinal processes cannot adequately explain the regional persistence of some of the coals, freshwater limestones, and paleosols. Exclusion of siliciclastics over as wide an area as some of these low-energy deposits occupy seemingly can be understood only in terms of broader basinwide controls. Some conceivable mechanisms responsible for these deposits could have been either tectonically induced diversion of major drainage systems creating a much wider than normal quiet interdistributary area (Belt and Lyons, 1989) or a climatic change substantially reducing erosion and (or) sediment transport over a large area. The apparent rhythmic reoccurrence of coal throughout the Pennsylvanian, if real, also strongly suggests an extrabasinal, probably climatic cause.

All of these controls, the influences of tectonism, eustasy, and climate external to the basin and the internal fluvial, deltaic, and shoreline processes, were operating in the Appalachians during the Pennsylvanian to a significant degree, many of them probably simultaneously. The critical factor in understanding the resultant complex assortment of sedimentary deposits is careful consideration of the relative influence of each of these controls through time.

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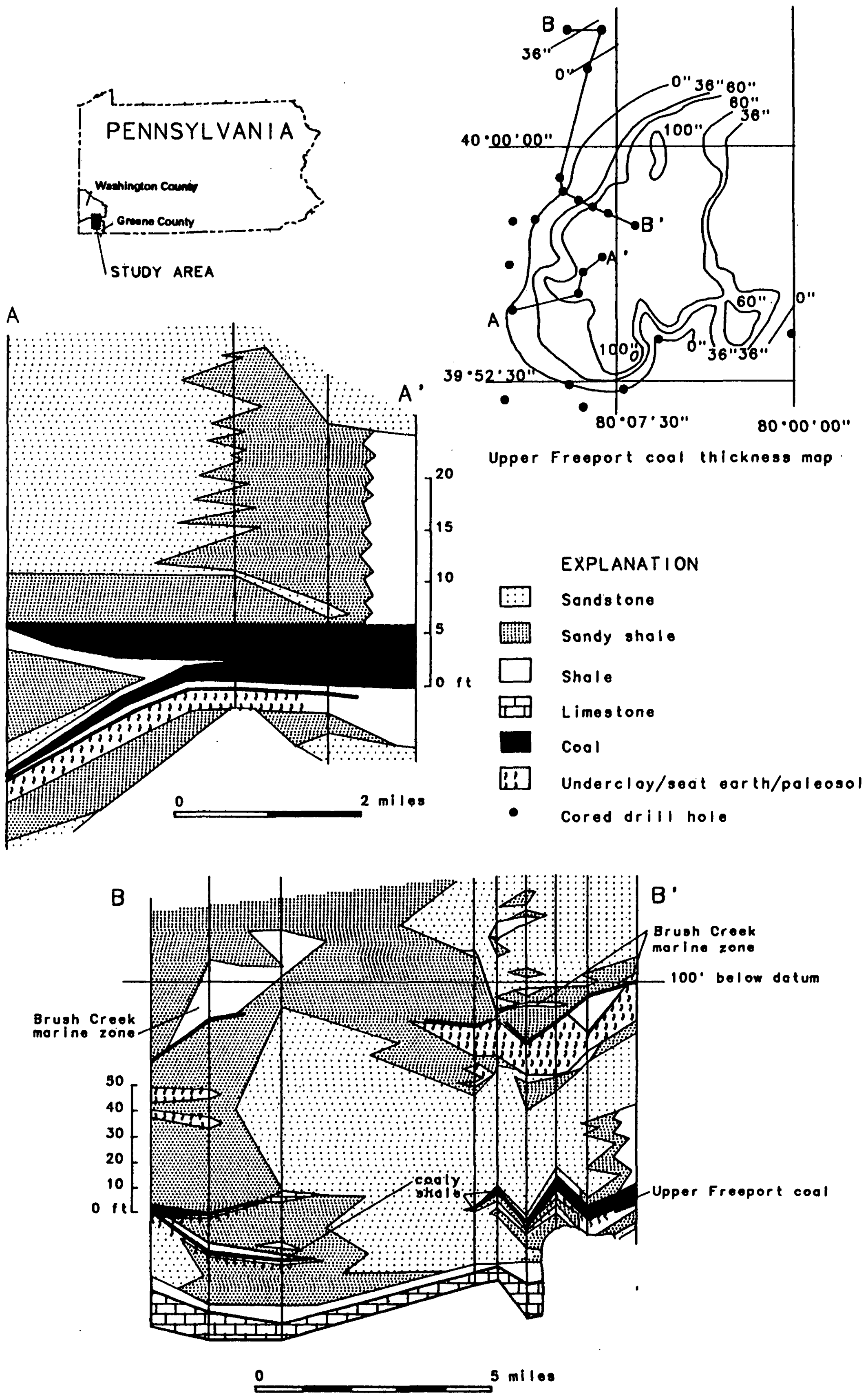

Figure 1. Upper Freeport coal in parts of Greene and Washington Counties, Pa.

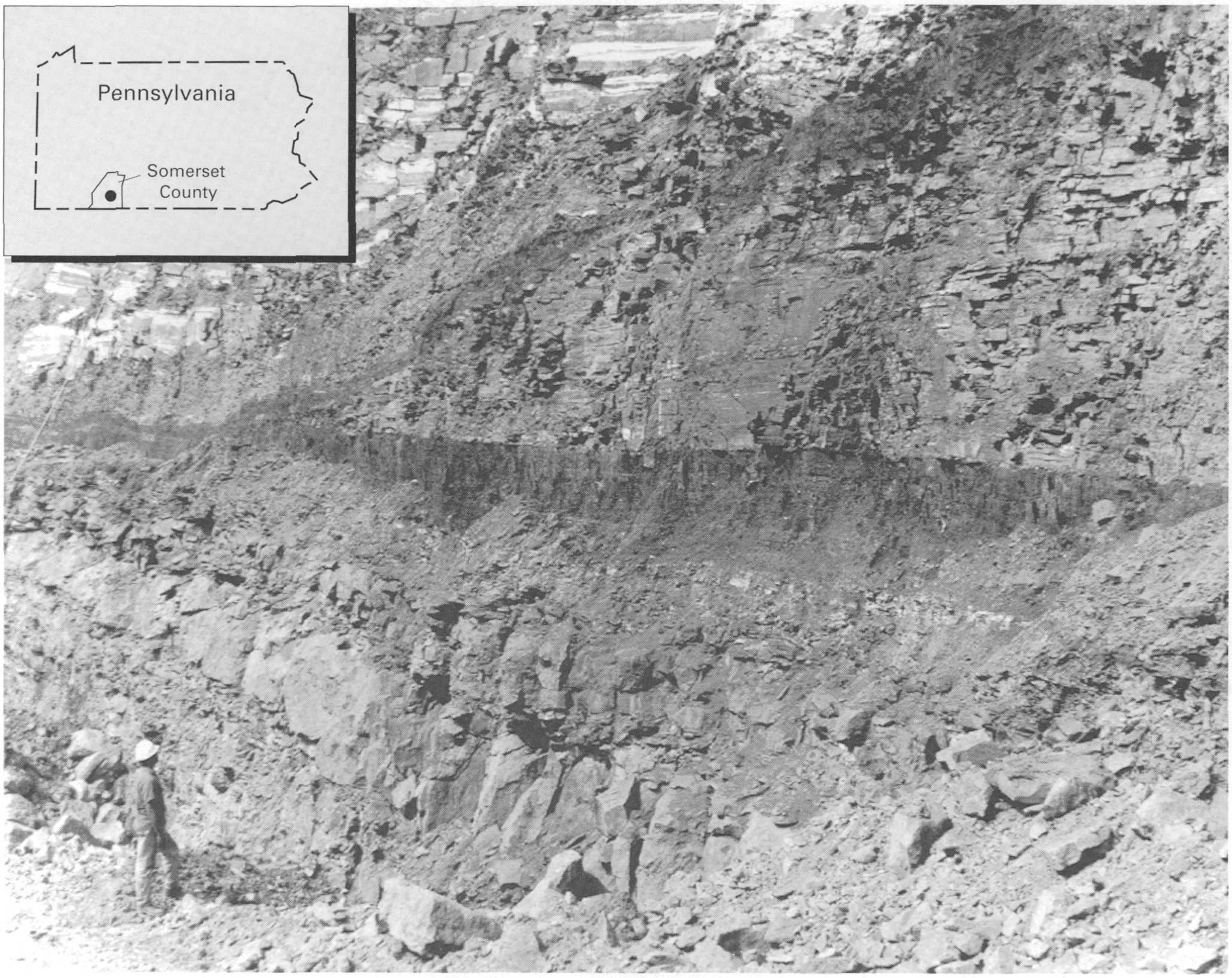

Figure 2. Redstone coal split in Somerset County, Pa.

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Significance of Midcontinent Pennsylvanian Cyclothems to Deciphering Global Pennsylvanian Stratigraphy

Philip H. Heckel

Each marine cyclothem in the classic upper Middle to Upper Pennsylvanian cyclic succession on the northern midcontinent shelf (fig. 1), north of the foreland basinal region of central Oklahoma, is characterized by a distinctive vertical sequence (fig. 2) consisting of (1) a thin (<1 m) transgressive limestone overlain by (2) a thin (~1 m) nonsandy offshore shale often containing a black phosphatic facies overlain by (3) a thicker (2–10 m) regressive limestone consisting of a classic shallowing-upward sequence overlain by (4) a terrestrial to nearshore detrital unit ranging from thin paleosols to thick alluvial and deltaic deposits; this capping unit is overlain by the transgressive limestone (unit 1) of the succeeding cyclothem (Heckel, 1977). All these units are essentially laterally continuous along 500 km of outcrop and into the subsurface of easternmost Colorado. Thus, each cyclothem is a marine transgressive-regressive (T–R) stratigraphic sequence recording a single inundation and withdrawal of the sea across the entire northern midcontinent shelf, an area covering about 500,000 km^2 in the States of Kansas, Missouri, Nebraska, and Iowa (Heckel, 1980).

The offshore shale across the entire shelf is characterized by an abundant fauna of conodonts dominated by *Idiognathodus* and closely related *Streptognathodus* (plus *Neognathodus* in the Desmoinesian) and including *Idioprioniodus* and *Gondolella* (fig. 2); the nearer shore parts of the cyclothem contain sparser conodont faunas that lack the latter two genera but include *Hindeodus* (=*Anchignathodus* of previous work) and become dominated by *Adetognathus* in the shallowest, most nearshore environments. Differences in species composition of *Idiognathodus*, *Streptognathodus*, and *Gondolella* in the dark offshore shales from cyclothem to cyclothem allow individual depositional cycles to be traced into the greatly thickened detrital succession in the foreland basin of Oklahoma (fig. 1), where the two limestone members are replaced by detrital clastic deposits from the Ouachita orogenic source. Patterns of both lithology and generic distribution of conodonts in stratigraphic units interspersed with the major cyclothems illustrated in figure 2 allow recognition of less widespread T–R marine cycles on the shelf and allow construction of a sea-level curve (fig. 3) for the mid-Desmoinesian to mid-Virgilian succession from the foreland basin to the northern limit of outcrop on the shelf (Heckel, 1986).

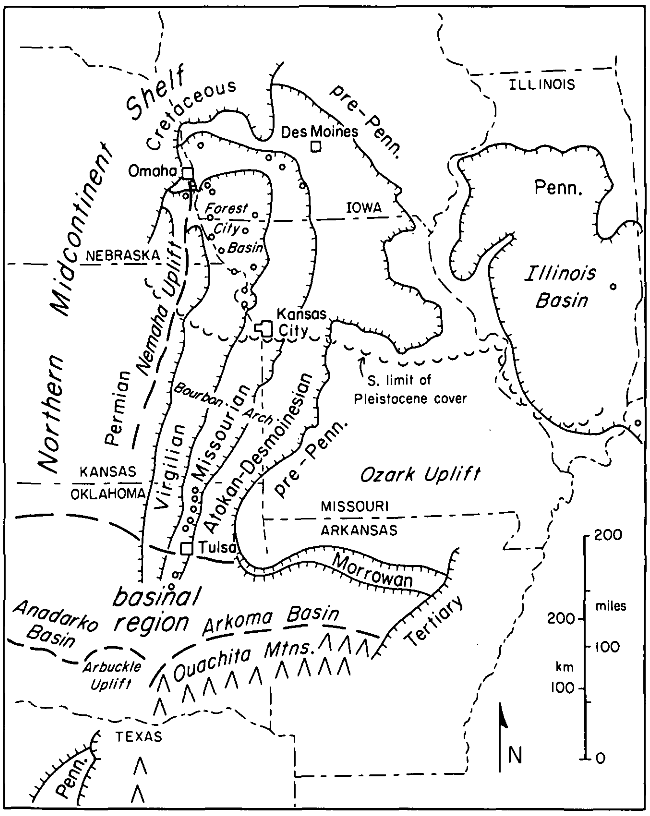

Figure 1. Midcontinent Pennsylvanian outcrop belt, with hachures in direction of dip, showing generalized Pennsylvanian structural features. Nemaha uplift and Forest City basin were formed during early Pennsylvanian time; Forest City basin became largely filled by the end of Middle Pennsylvanian (Desmoinesian) time, and the entire region north of the basinal region of Oklahoma then acted as the northern midcontinent shelf. Circles denote locations of long cores (from Heckel, 1990).

The sparsity to total lack of deltas between the cyclothems on the northern midcontinent shelf north of Kansas City (fig. 4) eliminates delta shifting as a general cause for these cyclothems. The appearance of all cyclothems from the upper mid-Desmoinesian Fort Scott Formation to the mid-Virgilian Howard Limestone upon the Nemaha uplift in southeastern Nebraska as well as in the adjacent Forest City basin eliminates differential local tectonism as a cause. The periodicity of the cyclothems and the interspersed smaller T–R cycles falls within the 20,000- to

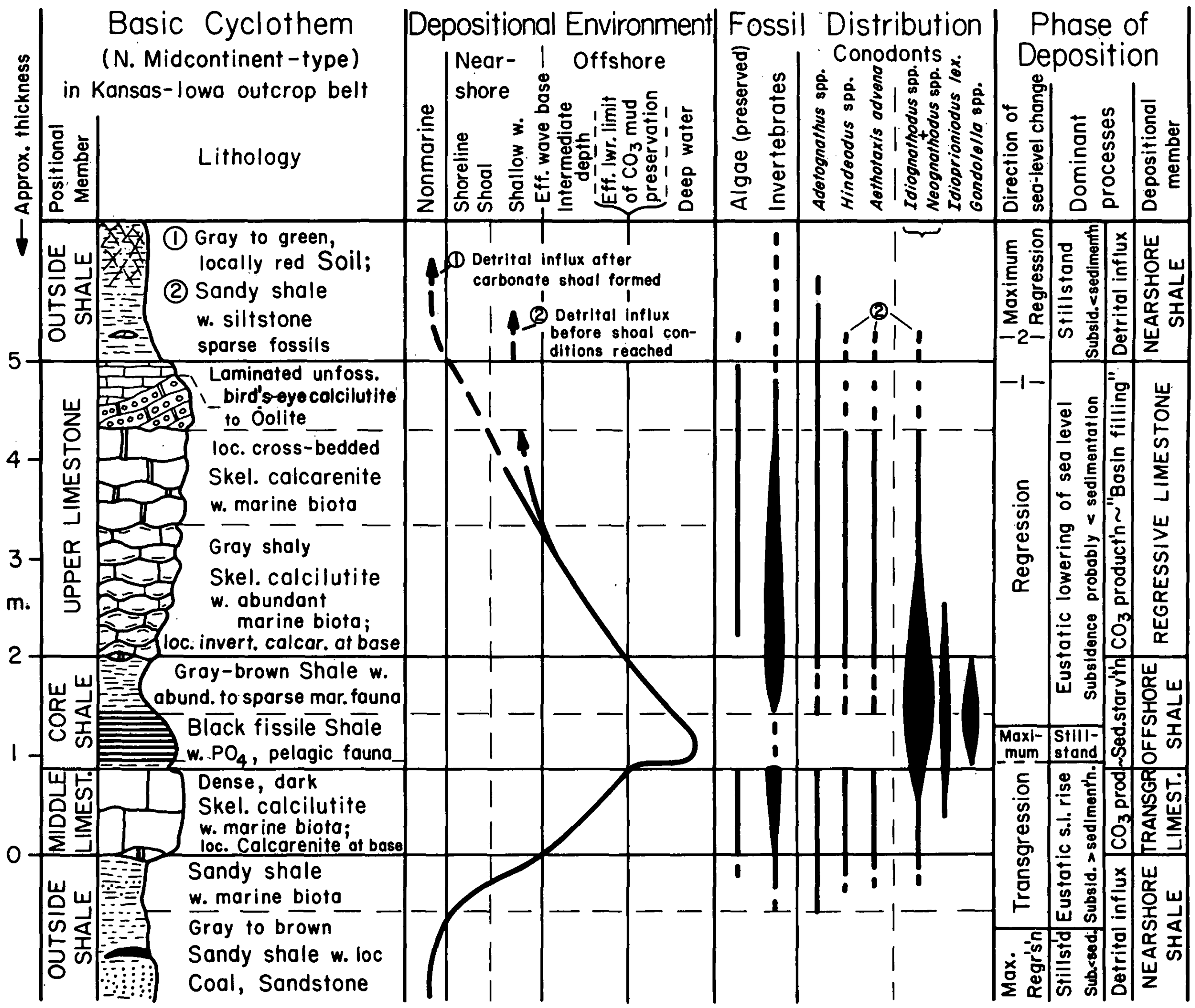

Figure 2. Basic northern midcontinent cyclothem representing one complete marine transgression and regression across the northern midcontinent shelf (Iowa, Nebraska, Missouri, and Kansas) during late Middle and early to middle Late Pennsylvanian (mid-Desmoinesian to mid-Virgilian) time and showing typical distribution of conodont genera (modified from Heckel, 1977).

400,000-year range of the periodicity of the Earth's orbital parameters (Heckel, 1986). This supports the idea of primary glacial-eustatic control over formation of the cyclothems because these periods control variation in mid-latitude solar insolation and are statistically significant in the cyclicity of Pleistocene glaciation (Imbrie, 1985). This short-term periodicity severely constrains the possibility of cyclic tectonic models (developed so far) as basic controls over cyclothem formation. It also has been determined that only transgressive rates on the order of postglacial sea-level rise of greater than 5 mm/yr are sufficient to outstrip the rates of tropical carbonate sediment accumulation consistently and produce a thin transgressive limestone overlain by a subthermocline black shale deposited in water depths on the order of 100 m (Heckel, 1984). This requirement eliminates all known cyclic tectonic models that could have acted over a broad shelf for which the maximum calculated rates of sea-level change are 0.1 mm/yr, a rate of sea-level rise that would be readily compensated by carbonate sediment accumulation. Evidence for tectonism during Late Pennsylvanian cyclothem formation in the midcontinent is reflected mainly in the increasing thickness of the cyclic sequence toward the downwarping foreland basin in Oklahoma (fig. 4).

Gondwanan glaciation as a cause of glacial eustasy is well documented (Veevers and Powell, 1987). Eustasy is necessary to explain the distinctive characteristics of northern midcontinent cyclothems; it also controlled the cyclic

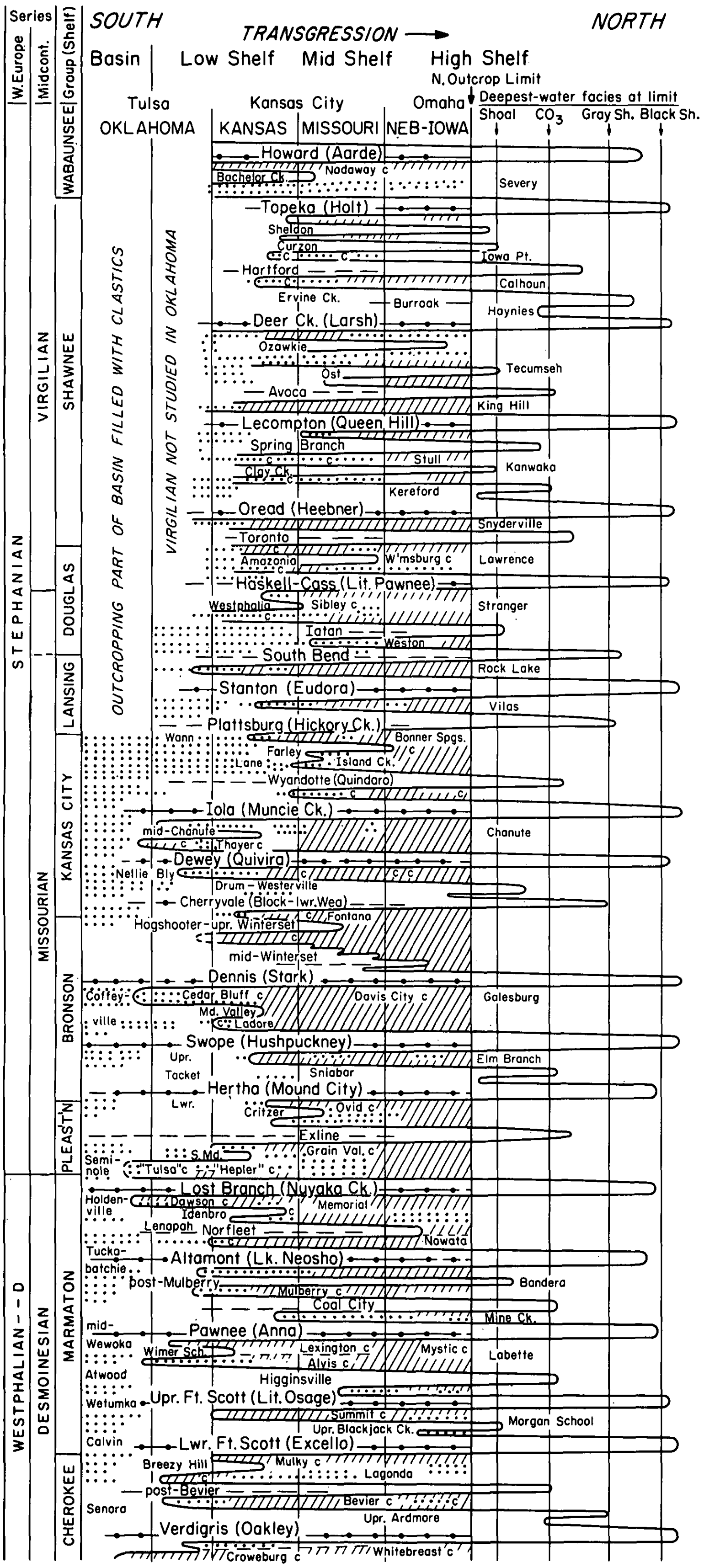

Figure 3. Sea-level curve (modified from Heckel, 1986) showing all scales of marine transgressive-regressive cycles of deposition that extend from the foreland basin of central Oklahoma various distances across the northern midcontinent shelf. Scales range from major cyclothems illustrated in figure 2 (largest letters on left side of curve), through intermediate cycles (middle-sized letters), many of which resemble the cyclothem illustrated in figure 2 but lack black facies in offshore shale, to minor cycles (smallest letters on left side of curve). Offshore shales (names in parentheses) are shown by lines (dashed where gray); large dots indicate nonskeletal phosphorite. Lowstand deposits include exposure surfaces and paleosols (oblique lines), fluvial-detaic complex (dots), and coal beds (c). PLEAS'T'N, Pleasanton.

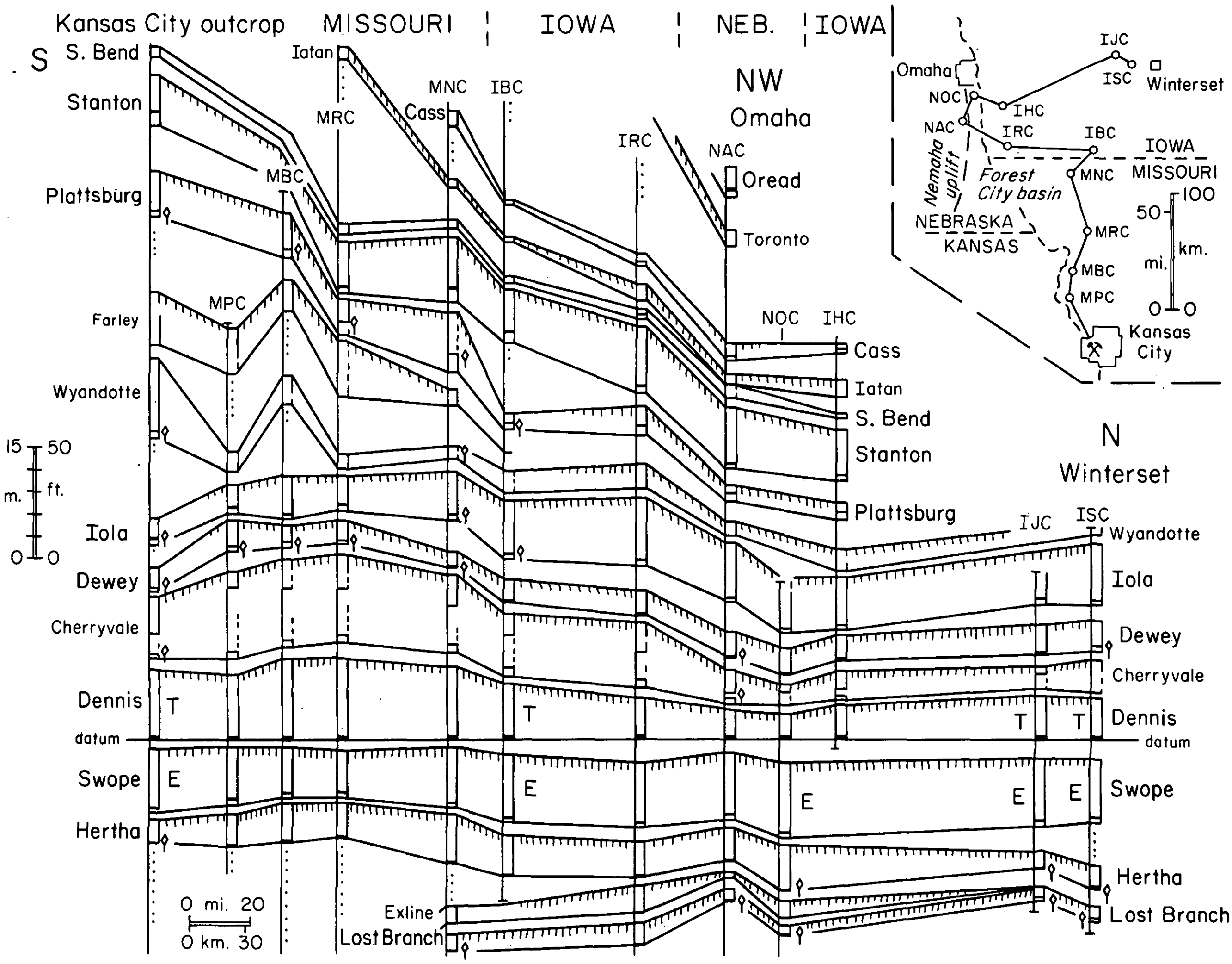

Figure 4. Generalized correlation cross section of lower Upper Pennsylvanian succession (Missourian provincial series = Exline Limestone—Iatan Limestone) on northern midcontinent shelf, based on long cores held by respective State geological surveys. Named units with largest letters are major marine cyclothems (fig. 2); black lines near base mark black to dark-gray phosphatic shales, and hachures on top mark exposure surfaces usually overlain by paleosols; lack of hachures indicates nearshore detrital deposits culminating in deltaic or other sandstones (dots). Southward thickening of succession reflects increase in tectonic subsidence in that direction. Tailed diamond symbols for conodont faunas and letters for fusulinids (E, *Eowaeringella ultimata;* T, lowest *Trticites*) show biostratigraphic control for correlation (modified from Heckel, 1990).

pattern of marine inundation and withdrawal (within these frequencies) that is recorded in Pennsylvanian deposits worldwide. It also means that the glacial-eustatic T–R sequences should be able to be correlated eventually on a worldwide scale. The initial stage of correlation across the northern midcontinent shelf by means of lithic sequence confirmed by conodonts throughout and fusulinids at a critical interval is shown in figure 4 (Heckle, 1990). Correlation of much of the succession (fig. 3) has been extended to the north Texas shelf (fig. 5) by Boardman and Heckel (1989) using conodonts, fusulinids, and ammonoids, and it shows that the potentially confounding effects of both tectonism and the more overwhelming detrital influx from the nearby Ouachita orogenic source are minimal. Correlation is now being established with the cyclic succession in Illinois and in the Appalachian basin, where deltas are even more conspicuous but are recognized now to be typically overlain by widespread paleosol horizons that attest to periodic widespread withdrawal of the sea from there as well as from the midcontinent. This places the delta-shifting model of Ferm (1970) into a more realistic perspective as a control over local detrital cycles that formed mostly during the phase of eustatic regression in areas near a detrital source (fig. 6). Climate change driven by the interactions of the orbital parameters, of course, is the basic control over the cyclothems because it controls glaciation and deglaciation, but local shorter cycles of climatic control over sedimentation described by Cecil (1990) in nonmarine sequences may in part have been controlled by glacial-eustatic fluctuations in marine shoreline position from nearby to distant, which

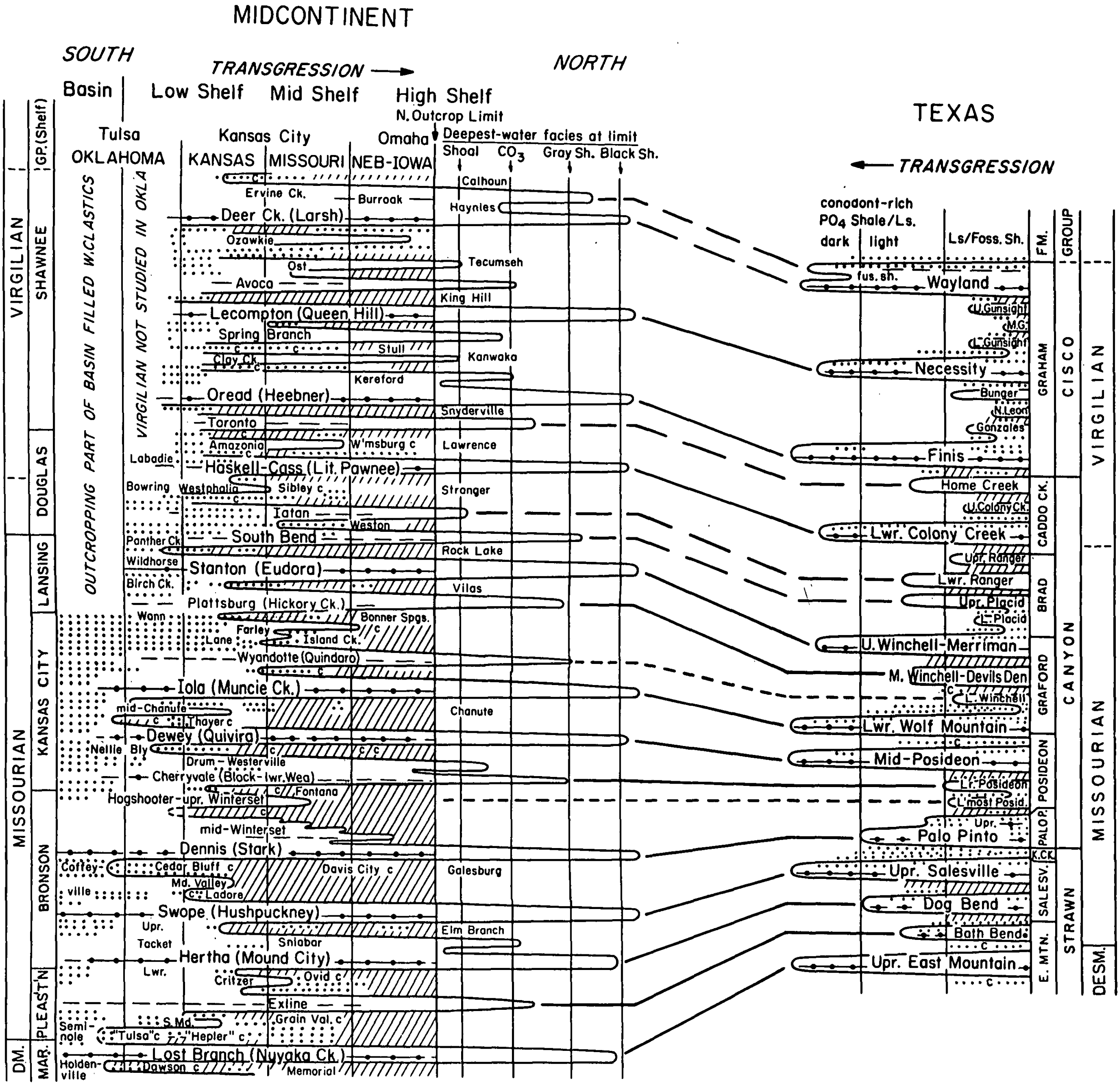

Figure 5. Correlation of part of midcontinent eustatic curve shown in figure 3 with eustatic curve determined independently for north-central Texas (lithic symbols same as on fig. 3). Correlations reflect (1) diagnostic faunas (solid lines) with first, last, sole, or acme occurrence of certain taxa (conodonts, ammonoids, fusulinids), (2) compatible faunas (long dashes), and (3) positional matches (short dashes in selected examples) (Boardman and Heckel, 1989). DM., DESM, Desmoinesian; MAR., Marmaton; PLEAS'T'N, Pleasanton; GP., Group; MG., Middle Gunsight; E. MTN., East Mountain; SALESV., Salesville; K.CK., Keechi Creek; PALOP., Palo Pinto; FM, Formation

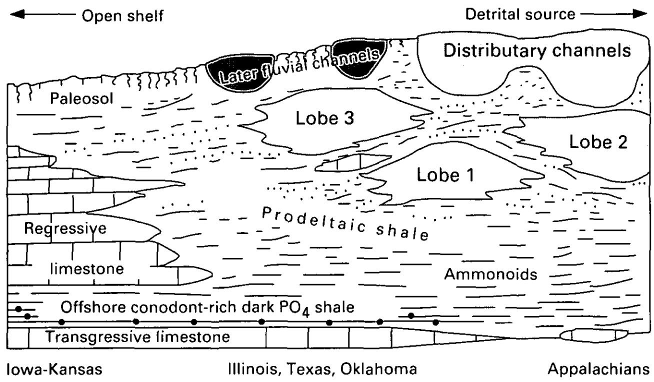

Figure 6. Depositional model for eustatic Pennsylvanian cyclothem approaching a humid shoreline dominated by detrital influx and incorporating delta shifting (lobes 1, 2, and 3 in succession) as a control over formation of minor local detrital cycles near the shoreline during eustatic phase of regression. Areas named below figure roughly show the position of major Pennsylvanian marine cyclothems on this idealized transect (modified from Boardman and Heckel, 1989).

would have controlled local sources of rainfall and temperature modulation.

Eventual establishment of a firm framework of biostratigraphic correlation of the major glacial-eustatic marine T–R events (the classic midcontinent cyclothems of fig. 2 as modified by detrital influx in fig. 6) should allow longer term, more subtle tectonic signals to be deciphered in more orogenic areas such as the Appalachians. It also has the potential to achieve correlation at a scale finer than the half-million years that is normally attributed to biostratigraphy because the eustatic T–R cycles of lesser magnitude (fig. 3) can be matched up on an event basis between the biostratigraphically correlated major cycles (fig. 5). This can lead to the development of paleogeographic maps for phases of deposition during marine transgression, regression, highstand, and lowstand, from the major cycles to progressively more minor cycles between them, which may achieve resolution of intervals, perhaps as short as thousands of years, and allow quite accurate delineation of the succession of ancient large-scale geographic and climatic patterns during the Pennsylvanian.

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Pennsylvanian Cyclic Deposition, Paradox Basin, Southwestern Colorado and Southeastern Utah

A.C. Huffman, Jr., S.M. Condon, and K.J. Franczyk

Thirty-three evaporite cycles have been identified in the Pennsylvanian of the Paradox basin (fig. 1). An idealized cycle begins with anhydrite overlain, in ascending order, by silty dolomite, calcareous black shale, dolomite, anhydrite, and halite (with or without potash at the top) and is separated from the overlying cycle by a sharp, possibly erosional contact. This basinal sequence is reflected on the shelves to the southeast, south, and southwest by carbonate cycles containing local concentrations of algal mound buildups and widespread evidence of subaerial exposure. Many of the black shales are traceable from the evaporite into the carbonate facies and have thus been used to correlate the Paradox basin cycles with those of the San Juan basin to the southeast and Black Mesa basin to the south.

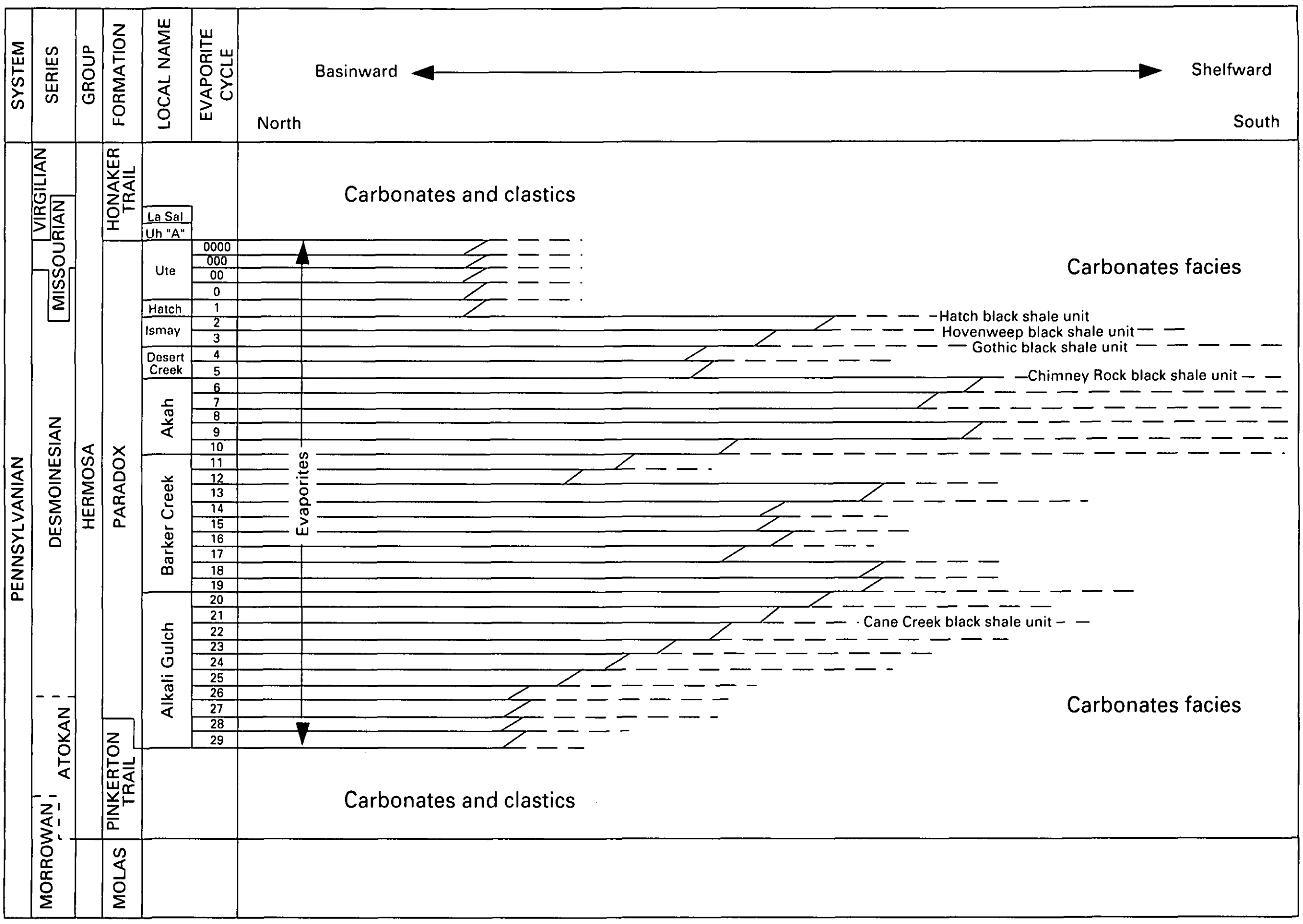

Figure 1. Diagrammatic stratigraphic section showing relationships of Pennsylvanian black shales to evaporite and carbonate facies in the region of the Paradox basin, southwestern Colorado and southeastern Utah. (Modified with permission of Rocky Mountain Geological Databases and R.J. Hite, U.S. Geological Survey.)

Areal distribution of lithologies within each cycle can be mapped with some precision through much of the basin because of the large amount of subsurface data available from petroleum exploration and development. For the purposes of regional mapping, cycles are commonly divided at the base of the black shale interval, even though these boundaries do not coincide with regional hiatuses or abrupt changes in depositional conditions. Figure 2 is a generalized map of lithologies in the lower Desert Creek zone (50–200 ft thick) above the Chimney Rock black shale unit in cycle 5. The map demonstrates the asymmetry of the basin, and the deepest, most saline part is in proximity to the Uncompahgre uplift on the northeastern margin. Less saline deposits grade southwest to the shelf carbonate facies with areas of algal mound buildups. During times of maximum lowstand, the basin was probably cut off entirely from marine waters, and much of the shelf area was exposed. Other cycles vary in detail, extent, and thickness but display the same general patterns and relationships of lithologies. A proximal arkosic facies along the northeastern basin margin, which has not been well defined because of the lack of drilling close to the Uncompahgre, is not included here.

Underlying the evaporites and carbonates of cycle 5 is the Chimney Rock black shale unit, varying in thickness from 0 to 50 feet (fig. 3). This interval is commonly calcareous or dolomitic and silty with a high organic carbon content. The shale contains a mixture of types II and III kerogen, as is the case in most of the shales in the upper part of the Paradox Formation and Honaker Trail Formation. In order to explain the high terrestrial kerogen content and general thickness patterns of these shales, previous workers have proposed the existence of a large fan delta issuing from the Uncompahgre highland on the southeastern margin of the basin. Clastic and organic material would have been delivered to the basin from wetter highlands, stored in the fan during lowstands, and then distributed throughout the basin by the next transgressive event and incorporated into the silty dolomite and black shale. Several lines of investigation are currently underway to determine the existence and possible extent of such a feature but are hampered by the sparsity of drilling close to the basin margin.

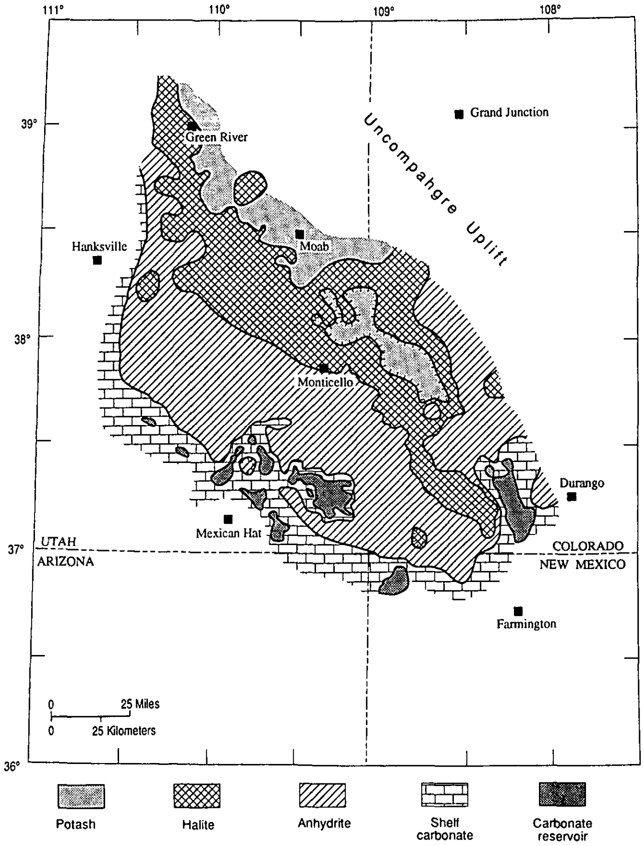

Figure 2. Distribution of lithologies making up cycle 5 above the Chimney Rock black shale unit in the Paradox Formation. Arkosic deposits along the eastern margin of the basin are not shown.

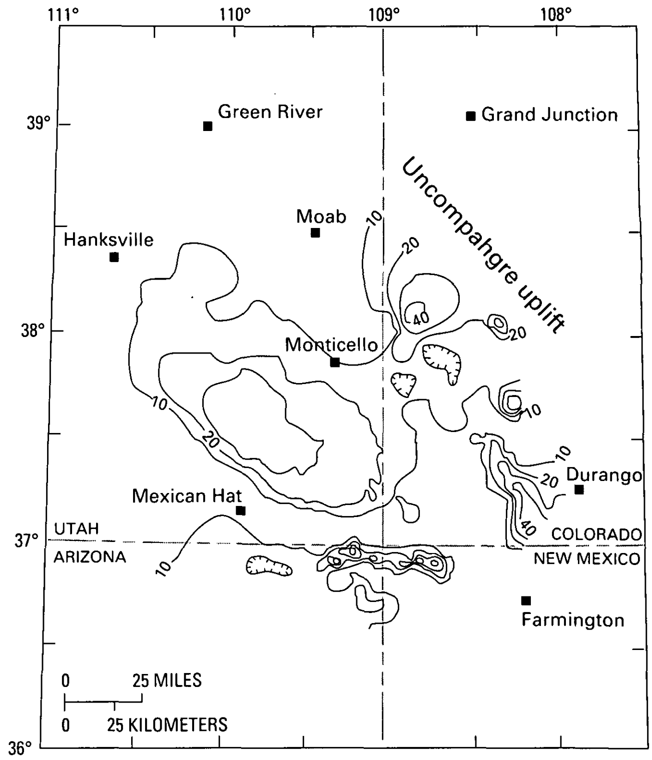

Figure 3. Isopach map of the Chimney Rock black shale unit. Pinchouts on northwestern, western, and southern margins of the basin are not shown. Contour interval is 10 ft.

The Uncompahgre uplift formed a highland of significant elevation from early Desmoinesian to early Wolfcampian. A large volume of arkosic and silicic clastics was shed from the highland, but most was trapped along the rapidly subsiding eastern margin of the basin and subsequently overridden by the basinward thrusting of the Uncompahgre during the Late Pennsylvanian to Early Permian. Correlation of basinal cycles with those exposed in outcrops along the southeastern part of the highland has been difficult partially because of the thrusting but primarily due to the general lack of evaporites and black shales close to the uplift. Cyclic sedimentation along the southeastern basin margin is expressed by a succession of carbonate-clastic pairs. Correlations are complicated by exposures that are locally superb but commonly inaccessible or widely separated from other exposures, by the lack of well control close to the outcrop, and by rapid changes in depositional environments between the basinal deposits and those adjacent to the highlands.

Throughout most of the Pennsylvanian, the Paradox basin was moving north-northeast from close to the equator to approximately 10°N., a paleolatitude position characterized by humid climate with prevailing winds from the east. The orientation of the Uncompahgre highlands and ancestral Front Range effectively blocked most of the moisture and produced predominantly arid to semiarid conditions with northerly winds in the basin. These paleoclimatic conditions in conjunction with eustatic fluctuations have been invoked as the principal controls on deposition, particularly the evaporite cycles. However, local features such as active faulting and folding as well as possible salt movement also strongly influenced deposition throughout the basin. Work is underway to determine if climatic cyclicity can be detected within the depositional cycles. Ongoing paleontologic, geochemical, and sedimentologic investigations are examining the interplay of climate, eustasy, and tectonism and will provide valuable insights not only into the development of the Paradox basin, but also into regional and continental paleogeography and paleoclimatology.

Pennsylvanian and Early Permian Paleogeography of Northwestern Colorado and Northeastern Utah

Samuel Y. Johnson, Marjorie A. Chan, and Edith H. Konopka

Northwestern Colorado and northeastern Utah include parts of four major sedimentary provinces active during the late Paleozoic ancestral Rocky Mountain orogeny: the Eagle basin, the northern part of the Paradox basin, the southern Wyoming shelf, and the southeastern part of the Oquirrh basin (fig. 1). Early Pennsylvanian to Early Permian sedimentation patterns in these provinces have been mapped on four time-slice paleogeographic map sets that show deposition during maximum transgressions and regressions. In general, clastic deposition (largely sandstone in shallow-marine, fluvial, deltaic, and eolian systems) dominated during regressions; deposition of marine limestones and clastics dominated during transgressions. The map sets are the basis for interpreting controls on depositional patterns, which include repetitive eustatic and climatic fluctuations, tectonics, and sediment supply.

Morrowan and early Atokan strata across most of the study area consist of fine-grained clastic rocks (regressive deposits) and more abundant limestones (transgressive deposits). Delta systems prograded out of the basement-cored Front Range and Sawatch uplifts into the Eagle basin. Regolith was deposited on a low-relief platform that extended westward from the western flank of the Eagle basin to the eastern edge of the Callville shelf. Fragmenta-

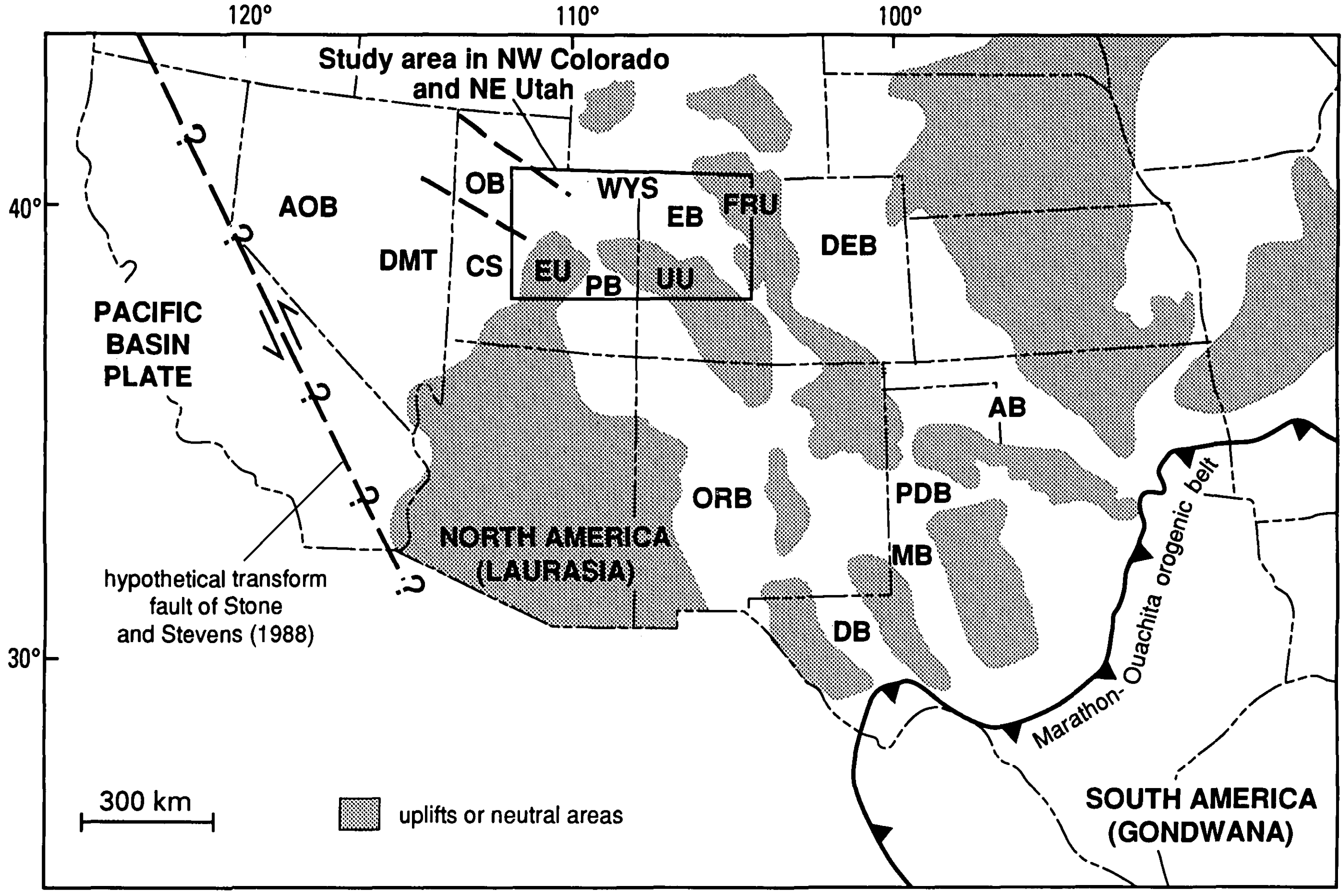

Figure 1. Schematic paleogeographic map showing the location of the Uinta-Piceance basin region, adjacent basins and uplifts of the ancestral Rocky Mountain orogeny, and inferred continental margins to the west and southeast. Map does not restore Mesozoic shortening and Cenozoic extension in the western United States. AB, Anadarko basin; AOB, Antler overlap basins; CS, Callville shelf; DB, Delaware basin; DEB, Denver basin; DMT, Dry Mountain trough; EB, Eagle basin; EU, Emery uplift; FRU, Front Range uplift; MB, Midland basin; OB, Oquirrh basin; ORB, Orogrande basin; PB, Paradox basin; PDB, Palo Duro basin; UU, Uncompahgre uplift; WYS, Wyoming shelf. Areas characterized by minimal uplift or subsidence are considered "neutral." (Modified from McKee and Crosby, 1975; Ross, 1986; Stone and Stevens, 1988; Smith and Miller, 1990; Johnson and others, 1992.)

tion of this platform occurred during the late Atokan by uplift of the basement-cored Uncompahgre highland and subsidence of the Paradox basin.

During regressions in the late Atokan to Desmoinesian, the combined effects of tectonics and lowered sea level led to restricted circulation and evaporite deposition in the Eagle and Paradox basins, while eolian sands prograded south-southwestward across the Wyoming shelf en route to the Oquirrh basin. Limestone deposition was dominant across the study area during transgressions.

During the Missourian and Virgilian, decreases in subsidence rate and associated progradation of clastic sediments into the central parts of the Eagle and northern Paradox basins resulted in termination of evaporite deposition. During regressions, fluvial and eolian deposition dominated in the Eagle basin, while sabkha and (or) shallow marine deposition became dominant in the northern Paradox basin. Clastic sediment continued moving southward in eolian dune fields across the Wyoming shelf to the Oquirrh basin. Deposition of limestones during transgressions was limited to the western part of the study area and a small area in the eastern Eagle basin. The Oquirrh basin was marked by a transition from shallow- to deep-water deposition. This transition apparently resulted from a decrease in sediment supply and not an increase in subsidence rate.

The Emery uplift became fully or mostly submerged in the early Wolfcampian, ending the history of the Paradox basin as a discrete geomorphic element. Deep-water clastic deposition in the Oquirrh basin continued from the latter part of the Missourian and early Virgilian, as did depositional patterns in the Eagle basin and on the Wyoming shelf. Deposition of transgressive limestones was limited to the southwestern part of the study area.

Regionally, the rates and magnitudes of subsidence were greatest in the Oquirrh basin, intermediate in the Eagle and northern Paradox basins, and lowest on the Wyoming shelf. In these four sedimentary provinces, rates of subsidence were lowest in the Early Pennsylvanian, highest in the Middle Pennsylvanian, and intermediate in the Late Pennsylvanian and Early Permian. The timing and geometry of uplift and subsidence in the study area suggest the overlapping influences of interactions along a more distant, convergent, continental margin to the southeast, and a more proximal, transtensional margin to the west.

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Cyclostratigraphic Correlation of Desmoinesian-Lower Missourian Shelf Carbonates (Horquilla Limestone) of the Pedregosa Basin with Midcontinent Cyclothems

W. Marc Connolly

Late Paleozoic cyclothems of the midcontinent have been studied for more than 60 years, and recent work has resulted in sea-level curves of Milankovitch scale periodicity for the Pennsylvanian System (Heckel, 1986, 1989; Ross and Ross, 1987). These presumably record eustatic events driven by orbital forcing of climate and concomitant Gondwanan glaciation. Eustasy as the driving mechanism can be best established from cycle-by-cycle correlation between different basins, demonstrating that sea-level fluctuations are essentially synchronous. The high-frequency (~400 ka) eustasy curves documented for the midcontinent have great potential for interbasinal correlation with precision exceeding that presently available from biostratigraphy alone. A eustatic signal is best preserved in the stratigraphic record of platform or shelf carbonates in settings that were isolated from clastic influx and tectonism. Southeastern Arizona is such an area (fig. 1) and provides the opportunity to further test the potential of cycles for long-distance interbasinal correlation. During Middle and early Late Pennsylvanian time, the cyclic strata of the Horquilla Limestone were deposited on the broad tectonically stable Pedregosa shelf. Subsidence rates in the Pedregosa basin were linear and uniform. Clastic sediments are rare in all but the deeper water facies.

The correlations, within stratigraphic intervals defined by conventional biostratigraphic datums, are based on two sections in Cochise County that span early Atokan through middle Missourian time (figs. 1, 2) and on coeval strata in the midcontinent section. The Dry Canyon section is on the eastern flank of the Whetstone Mountains (Tyrrell, 1957; Wrucke and Armstrong, 1987; Connolly and Stanton, 1990). The Gunnison Hills section, 50 km to the northeast, is on the western side of Gunnison Peak (Estes, 1968).

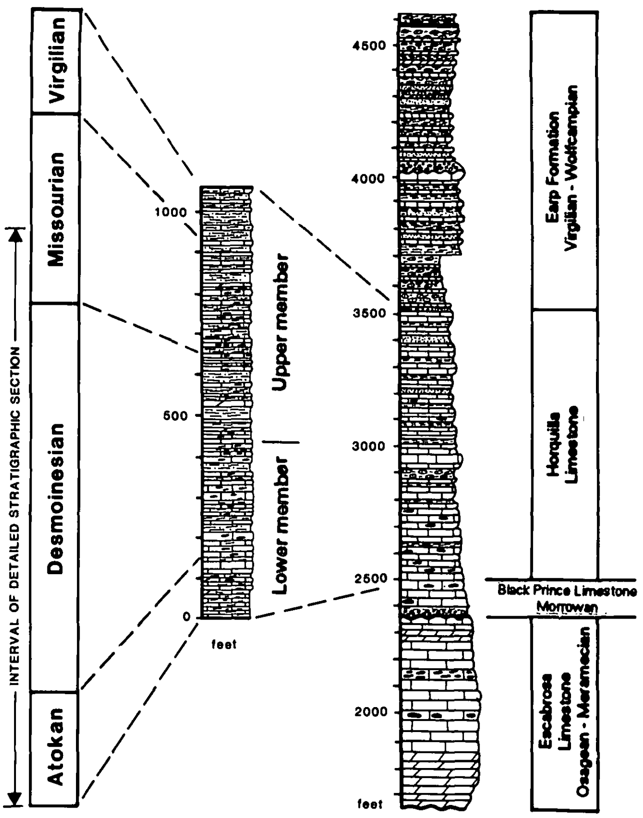


Figure 1. Generalized columnar section of upper Paleozoic carbonate strata exposed on the southern side of Dry Canyon in Arizona (modified from Ross and Tyrrell, 1965; Wrucke and Armstrong, 1987; additional locality data are available in Schreiber and others, 1990).

CARBONATE CYCLES

The conspicuous transgressive-regressive (T–R) cyclicity of the Horquilla Limestone is reflected in the bench and slope topography. These cycles are equivalent to the midcontinent cyclothems but are more asymmetric, lithologically simpler, and generally thinner. Individual T–R cycles are bounded by marine flooding surfaces and consist of three units: (1) a basal marl, (2) a wavy thin-bedded limestone, commonly nodular near the base, and (3) a massive thick-bedded bench-forming limestone (fig. 3).

The basal marl generally forms the lower third to lower half of each cycle. The initial transgression, to a depth sufficient to inhibit the production of carbonate sediment, was rapid. Gray calcareous siltstone in the lower third of the basal marl grades into a relatively pure claystone, about 15 cm thick, followed by gray calcareous siltstone in the upper two-thirds of the unit (fig. 3). The claystone probably represents the maximum flooding horizon. The marl unit consists of argillaceous sediment bypassed into deeper water under somewhat starved conditions, eolian silt, and carbonate sediment that may be largely allochthonous.

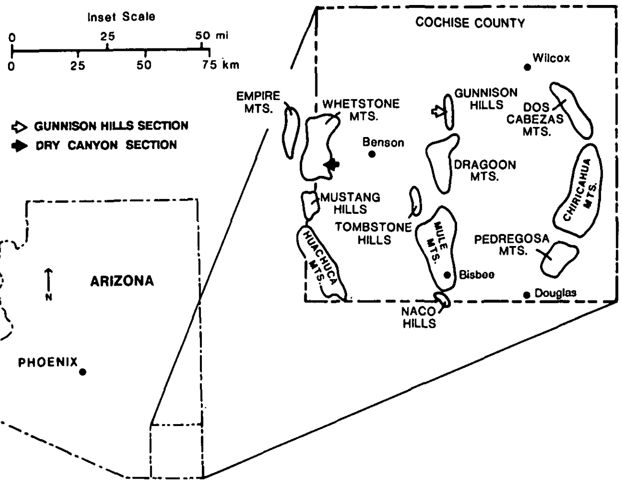


Figure 2. Index map of southeastern Arizona with locations of the Dry Canyon and Gunnison Hills measured sections (from Blakey and Knepp, 1989).

The middle unit, a wavy thin-bedded argillaceous limestone, formed as regression brought the depositional surface into the bathymetric range of carbonate production below fair-weather wave base. The upper unit formed during the later stages of regression as carbonate production overwhelmed clastic influx. It grades from less argillaceous wackestones and packstones to shoal-water grainstones. *Chaetetes*, diagnostic of shallow water, forms biostromes in the upper half in some cycles (Connolly and others, 1989). Within the upper 50 cm of many cycles, both primary and diagenetic fabrics indicate peritidal conditions and (or) pedogenesis (Connolly and Stanton, 1990). Nonmarine facies are apparently absent, and maximum regression is commonly represented by a subaerial exposure surface. The cyclic sequence is best explained by relatively constant argillaceous influx coupled with carbonate sedimentation that decreased with depth to a critical threshold depth. The magnitude of the transgressions is difficult to assess, but sea level fluctuated from subaerial exposure to depths sufficient to terminate autochthonous carbonate production.

INTRABASINAL CORRELATIONS

Local correlations are based on lithologic criteria, fusulinid biostratigraphy, and larger scale variations in the cyclic succession. Pedogenic horizons that presumably represent basinwide subaerial exposure events, proved useful for correlating cycles. By using the framework established from the lithologic and faunal datums, the sequences of about 50 T–R cycles were compared on the basis of stratigraphic position, prominent and distinctive cycles, and the groupings of these cycles into larger cycles (fig. 4).

The lowest biostratigraphic datum is the first appearance of primitive species of *Beedeina* diagnostic of the base of the Desmoinesian in southeastern Arizona (Ross and Sabins, 1965). The first appearance, *B. arizonensis* in Dry-Canyon (Ross and Tyrrell, 1965) and *B. hayensis* in the Gunnison Hills (Estes, 1968), is about the same distance and the same number of T–R cycles above the base of the Horquilla Limestone (fig. 4). The highest occurrence of *Beedeina* in the Gunnison Hills is an indeterminate species that occurs above *B. acme*, an advanced late Desmoinesian species (Estes, 1968). The highest occurrence of *Beedeina* in Dry Canyon is *B. rockymontana*, a middle Desmoinesian species (Ross and Tyrrell, 1965). Because more advanced species have yet to be found in Dry Canyon, the highest occurrence of *Beedeina* is not a reliable datum for correlating the two sections (fig. 4). The lowest occurrence of *Triticites* in Dry Canyon is two cycles (about 50 feet) below the lowest occurrence in the Gunnison Hills, based on the correlation of a distinctive pedogenic surface at the tops of unit T–125 in Dry Canyon and unit E–67 in the Gunnison Hills (fig. 4).

INTERBASINAL CORRELATIONS

Cyclostratigraphic correlation with the midcontinent is expedient because the Pennsylvanian type series and the eustasy curves have been established there. The midcontinent cyclothems differ from the Horquilla cycles in being more symmetrical, having a better developed transgressive phase, and in the presence of outside shales that commonly contain thick prodeltaic to nonmarine clastic facies (Heckel, 1989).

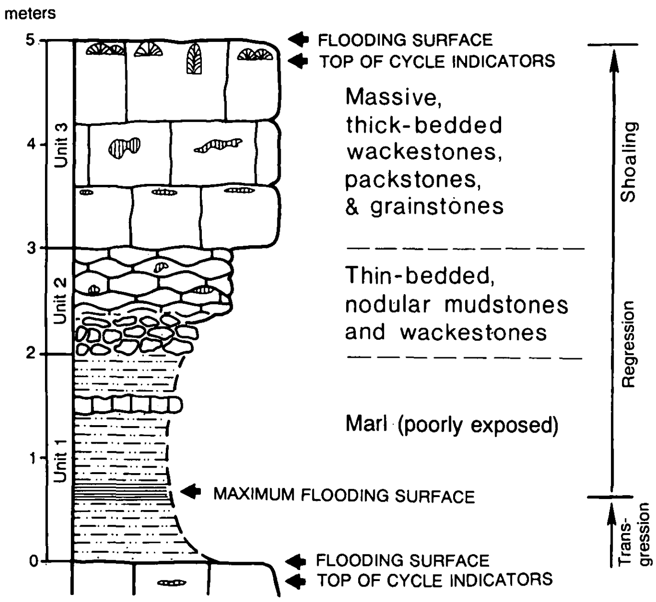


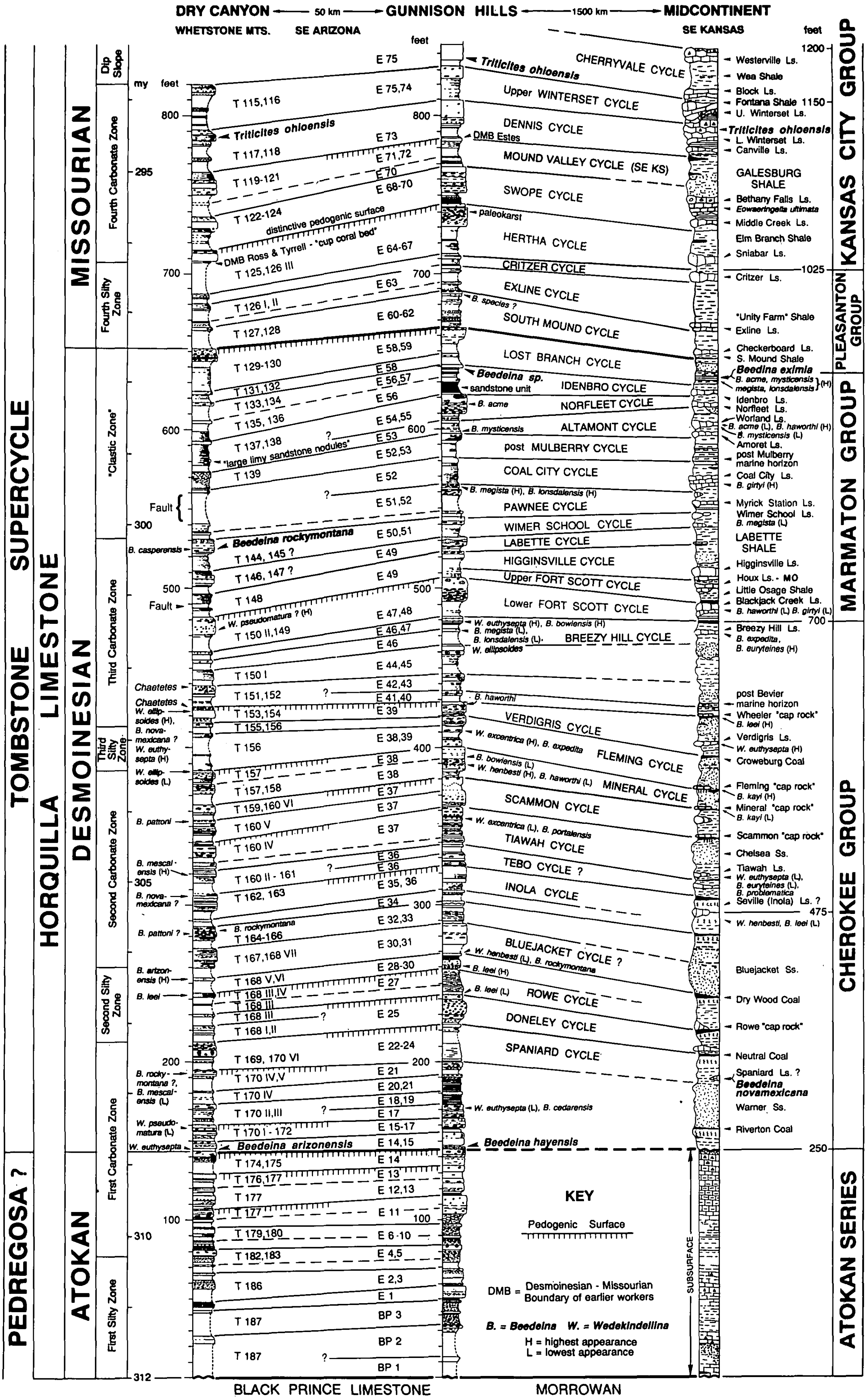
Figure 3. Typical carbonate cycle (Horquilla Limestone) idealized from Desmoinesian strata in Dry Canyon. Top of cycle indicators include oolitic grainstones, *Chaetetes* biostromes, and subaerial exposure surfaces (peritidal fabrics and pedogenic fabrics).

The first correlation among distant basins constrains, and in a sense forces, subsequent correlations, and therefore is the most critical. The first appearance of *Triticites* was the principal horizon selected for interbasinal correlation. The lowest occurrence of *Triticites* in the Arizona sections (unit T–117 in Dry Canyon, Tyrrell, 1957) was correlated with the lowest midcontinent occurrence (lower part of the Winterset Limestone of the Dennis cycle, Thompson, 1957). On the basis of this datum, the biostratigraphic succession lower in the sections was utilized to evaluate subsequent cyclostratigraphic correlations.

The lowest occurrence of *Beedeina* in Arizona cannot be correlated directly with the midcontinent section because the lower part of the Cherokee Group is largely nonmarine and because the most primitive species of *Beedeina* have not been reported from the midcontinent (fig. 4). *Beedeina eximia*, an advanced late Desmoinesian species from the Cooper Creek Limestone of the Lost Branch cycle, represents the highest reported occurrence in the midcontinent (Thompson and others, 1956; Heckel, 1991). In Arizona, *Beedeina* sp. has been reported from two cycles in the Gunnison Hills that bracket the inferred correlative of the Lost Branch cycle (Estes, 1968; fig. 4). Both occurrences lie above *B. acme*, which occurs with *B. eximia* in the Lonsdale Limestone of the Illinois basin (Dunbar and Henbest, 1942). The Lonsdale Limestone has been correlated with the Cooper Creek Limestone on the basis of conodonts (Heckel, 1991).

The highest reported occurrence of *Beedeina* in Arizona (unit E–62, Estes, 1968) warrants additional scrutiny. It may be diachronous with the last appearance in the midcontinent (fig. 4). This serves to illustrate the current resolution of fusulinid biostratigraphy at the series level. The range zones of individual species are poorly constrained, are facies dependent within a given basin, and differ from basin to basin. Existing zonations for the Desmoinesian are informal in nature, agree in general but differ in detail, and

Figure 4. Cyclostratigraphic correlation of T–R cycles within ▶ the Pedregosa basin (Dry Canyon and Gunnison Hills) and interbasinal correlations with the midcontinent (southeastern Kansas). Correlation criteria are discussed in the text. Time scale is interpolated from Ross and Ross (1987) after Harland and others (1982); T-numbers, unit numbers of Tyrrell (1957); E-numbers, unit numbers of Estes (1968). Kansas section is after Howe (1956) and Zeller (1968) with revisions from Heckel (1986, 1989, 1991). Vertical scale of the Kansas section is normalized to the Arizona sections. Fusulinid biostratigraphy is from Alexander (1954), Thompson and others (1956), Thompson (1957), Bebout (1963), Ross and Tyrrell (1965), and Estes (1968).


DRY CANYON — 50 km — GUNNISON HILLS — 1500 km — MIDCONTINENT
WHETSTONE MTS. SE ARIZONA
SE KANSAS
feet
my feet
PEDREGOSA ?
TOMBSTONE SUPERCYCLE
HORQUILLA LIMESTONE
MISSOURIAN
DESMOINESIAN
ATOKAN
Dip Slope
Fourth Carbonate Zone
Fourth Silty Zone
"Clastic Zone"
Third Carbonate Zone
Third Silty Zone
Second Carbonate Zone
Second Silty Zone
First Carbonate Zone
First Silty Zone
Triticites ohioensis
DMB Ross & Tyrrell - "cup coral bed"
distinctive pedogenic surface
"large limy sandstone nodules"
Fault
Beedeina rockymontana
B. casperensis
W. pseudomatura ? (H)
Chaetetes
Beedeina arizonensis
W. euthysepta
B. patroni
B. leei
B. rockymontana
B. mescalensis
B. novamexicana ?
Beedeina hayensis
B. = Beedeina W. = Wedekindellina
H = highest appearance
L = lowest appearance
KEY
Pedogenic Surface
DMB = Desmoinesian - Missourian Boundary of earlier workers
SUBSURFACE
CHERRYVALE CYCLE
Upper WINTERSET CYCLE
DENNIS CYCLE
MOUND VALLEY CYCLE (SE KS)
SWOPE CYCLE
HERTHA CYCLE
CRITZER CYCLE
EXLINE CYCLE
SOUTH MOUND CYCLE
LOST BRANCH CYCLE
IDENBRO CYCLE
NORFLEET CYCLE
ALTAMONT CYCLE
post MULBERRY CYCLE
COAL CITY CYCLE
PAWNEE CYCLE
WIMER SCHOOL CYCLE
LABETTE CYCLE
HIGGINSVILLE CYCLE
Upper FORT SCOTT CYCLE
Lower FORT SCOTT CYCLE
BREEZY HILL CYCLE
VERDIGRIS CYCLE
FLEMING CYCLE
MINERAL CYCLE
SCAMMON CYCLE
TIAWAH CYCLE
TEBO CYCLE ?
INOLA CYCLE
BLUEJACKET CYCLE ?
ROWE CYCLE
DONELEY CYCLE
SPANIARD CYCLE
DMB Estes
paleokarst
sandstone unit
Westerville Ls.
Wea Shale
Block Ls.
Fontana Shale
U. Winterset Ls.
L. Winterset Ls.
Canville Ls.
GALESBURG SHALE
Bethany Falls Ls.
Eowaeringella ultimata
Middle Creek Ls.
Elm Branch Shale
Sniabar Ls.
Critzer Ls.
"Unity Farm" Shale
Exline Ls.
Checkerboard Ls.
S. Mound Shale
Beedeina eximia
Idenbro Ls.
Norfleet Ls.
Worland Ls.
Amoret Ls.
post Mulberry marine horizon
Coal City Ls.
Myrick Station Ls.
Wimer School Ls.
LABETTE SHALE
Higginsville Ls.
Houx Ls. - MO
Little Osage Shale
Blackjack Creek Ls.
Breezy Hill Ls.
post Bevier marine horizon
Wheeler "cap rock"
Verdigris Ls.
Croweburg Coal
Fleming "cap rock"
Mineral "cap rock"
Scammon "cap rock"
Chelsea Ss.
Tiawah Ls.
Seville (Inola) Ls. ?
Bluejacket Ss.
Dry Wood Coal
Rowe "cap rock"
Neutral Coal
Spaniard Ls. ?
Beedeina novamexicana
Warner Ss.
Riverton Coal
KANSAS CITY GROUP
PLEASANTON GROUP
MARMATON GROUP
CHEROKEE GROUP
ATOKAN SERIES
BLACK PRINCE LIMESTONE
MORROWAN

range from five to seven subzones (Douglass, 1987; Ross and Ross, 1987; Wilde, 1990). Nevertheless, a succession of primitive to intermediate to advanced species is well documented in both Arizona and the midcontinent. This hypothesis supports the cyclostratigraphic correlations but lacks the resolution for independent corroboration (compare fig. 4 with cited zonal schemes). Conodont biostratigraphy probably holds greater potential for evaluating and independently substantiating the correlations.

PERIODICITY OF CYCLES

The average periodicity of T–R cycles in the Arizona sections, ranging from 280,000 years (lower Desmoinesian) to 475,000 years (lower Missourian), is 352,000 years (table 1). This is comparable to the 235,000- to 393,000-year range for the major cycles of the midcontinent (Heckel, 1986) and to the 330,000- to 370,000-year range for Desmoinesian cycles in the Orogrande basin of New Mexico (Algeo, 1991). Estimated cycle periods for this time interval in the Appalachian basin are 400,000 to 450,000 years (Busch and Rollins, 1984). The common range of periodicities for Pennsylvanian T–R cycles in areas that are both widely separated and characterized by differing degrees of tectonic influence and styles of sedimentation (325,000–425,000 years), and the fact that the periodicities fall within the Milankovitch band, provides circumstantial evidence of orbital forcing of glacial eustasy as the cause of the sea-level fluctuations.

Cycles with periods longer than the T–R cycles are present in the Arizona sections. These reflect variations in the proportions of the marl and carbonate units. Silty zones in which the marl member is relatively thick alternate with carbonate zones, in which the bench-forming limestone units are relatively thick (fig. 4). These long-term cycles have a period of about 4 million years, and they may prove to be useful for interbasinal correlations. Their origin is unknown however; they may not be genetically related to the cyclothem scale T–R cycles.

Table 1. Average cycle periodicities in the Horquilla Limestone.

[Cycle periods calculated for intermediate (cyclothem scale) T–R cycles in the Gunnison Hills section (fig. 2)]

Interval	Time (ka)	No. of cycles	Period (thousand years)
Lower Missourian	3,800	8	475
Upper Desmoinesian	4,300	10	430
Lower Desmoinesian	7,000	25	280
Atokan	3,200	9	355
Total	18,300	52	
Average			385

APPLICATION TO PREDICTIVE STRATIGRAPHIC ANALYSIS

Correlation of cyclic sequences over a distance of 1,500 km, and across significantly different facies, indicates the value of cyclostratigraphy for detailed correlation. The cycle-by-cycle correlation among different basins and depositional settings emphasizes the widespread eustatic nature of the sea-level fluctuations and provides compelling evidence for an interregional synchronous allogenic mechanism for the generation of these cratonic T–R cycles. Detailed interbasinal correlations can help resolve the complex interplay of eustasy, tectonics, and climate. These variables strongly influence the stratigraphic record and vary in relative importance, both temporally and spatially.

Analysis of a thin stratigraphic interval, or time slice, can minimize temporal effects and enhance the variation induced by geographic gradients (for example, the sedimentary response to climatic change along a latitudinal or orographic gradient). Cyclostratigraphic correlation of coeval T–R cycles has the potential to generate time slices on the order of a few hundred thousand years over long distances. For example, unit E–59, a regressive limestone in the Gunnison Hills, has been correlated with the upper part of the Lost Branch Formation of Kansas and Missouri (fig. 4). On the basis of stratigraphic, biostratigraphic, and cyclostratigraphic criteria, the upper part of the Lost Branch Formation (and by extrapolation unit E–59) has been correlated with the upper part of the East Mountain Shale (Texas), the Glenpool Limestone (Oklahoma), the Cooper Creek Limestone (Iowa), the Lonsdale Limestone (Illinois), the West Franklin Limestone (Illinois and Indiana), and possibly the Madisonville Limestone Member of the Sturgis Formation of Kentucky (Douglass, 1987; Boardman and Heckel, 1989; Heckel, 1991; Heckel and others, 1991).

Cyclostratigraphic correlation can facilitate the discrimination of global allogenic, regional allogenic, and local autogenic agents that contribute to contrasting styles of sedimentation; it exceeds the current resolution of biostratigraphy; and it provides an improved framework for paleoenvironmental, paleogeographic, paleoclimatic, and basin history analyses. It has broad application to the evaluation of the interrelationship between paleoclimate and sediment flux and represents an important approach to predictive stratigraphic analysis.

ACKNOWLEDGMENTS

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Evidence of Climate Change in the Lower and Middle Carboniferous Shallow-Water Carbonate Rocks of Arctic Alaska, New Mexico, and Arizona

Augustus K. Armstrong and Bernard L. Mamet

An extensive shallow-water carbonate shelf developed in arctic Alaska during the Carboniferous. According to paleomagnetic data, the region was then 28 to 43° north of the equator (C.R. Scotese, University of Texas at Arlington, written commun., June 1991). Twenty-two measured sections (700–3,000 ft thick) from the Lisburne Group (Mississippian to Permian) in the Brooks Range of arctic Alaska and three from the Yukon Territory, Canada (figs. 1, 2) contain microfossil assemblages assigned to zones of late Tournaisian (Osagean) through early Westphalian (Atokan) age (Armstrong and Mamet, 1977).

Representatives of both Eurasiatic and American cratonic microfaunas permit correlation with the original Carboniferous type sections in Western Europe as well as with the standard Mississippian and Pennsylvanian sequences in the midcontinent region of North America. The carbonate petrology of the Lisburne Group is composed of predominantly bryozoan-pelmatozoan wackestones and packstones and lesser amounts of lime mudstones, diagenetic dolomites, and pelmatozoan and ooid grainstones. The Lisburne Group was deposited on a slowly subsiding shallow-water carbonate shelf. The stratigraphic succession is commonly cyclic, alternating from open marine to subtidal. A carbonate-platform depositional model for these carbonate rocks, illustrating the spatial distribution of the organic remains and microfacies to water depth and salinity, shows that the corals and foraminifers are common near the shoaling-water facies, rare in the basinal and subtidal facies, and absent in the intertidal or supratidal facies. The pelmatozoan-bryozoan wackestone packstone facies, an open-marine facies, contains a sparse foraminifer fauna.

The microfauna belong to the Alaska and Taimyr subrealms, and a temperate warm environment is indicated by low abundance, low species diversity, high genus to species ratio, high rate of cosmopolitanism, and incomplete phylogenies. Some 61 genera and 130 species of Carboniferous foraminifers and algae are recognized. Algae are not diverse, although Palaeosiphonocladales are prolific at some levels (for example, *Donezella* bands in zone 21, Mamet and de Batz, 1989). Dasycladales, considered good indicators of tropical-equatorial waters are quite scarce (base of the Alapah Limestone and top of the Wahoo Limestone).

Lithostrotionoid corals can be identified to the species level, in part provincial to northern Canada and Alaska. The stratigraphic range of individual coral species and faunal assemblages extends throughout two to four microfossil zones. Rugose corals are rare in the Wachsmuth Limestone, zones 8 to 9 (Tournaisian). Colonial rugose corals are abundant in zones 11 through 15 (Visean) in the Alapah Limestone and disappear near the base of zone 16i (Visean). They reappear in significant numbers in the shoaling ooid sands of zone 20 to 21 (Morrowan and Atokan), Middle Carboniferous, Wahoo Limestone (Armstrong, 1972). The Tournaisian carbonate rocks contain poor faunas of foraminifers, and the sedimentary deposits are dominated by echinoderm-bryozoan fragments. Oolites are absent in these rocks. The upper Visean-Namurian rocks have red beds and evaporites that are now preserved as collapse breccias (zones 16i to 19). The microfauna and microflora are poorly diversified.

These shallow-subtidal fossils and microfacies are interpreted as representing cooler and deeper water (fig. 3) in the Tournaisian, whereas the abundance of foraminifers and robust colonial corals and patch reefs indicate warmer waters in zones 10 into 15 (Visean). A warmer and wet climate in the early Visean is supported by thin coal beds in the terrigenous Kekiktuk Conglomerate in the subsurface of the Prudhoe Bay region and in outcrops in the eastern Brooks Range (Armstrong and Mamet, 1975; Bloch and others, 1990).

In extreme northwestern Alaska on the sea cliffs south of Cape Lisburne, a thick sequence of lower Visean, terrigenous clastics rocks has 1- to 4-ft-thick coal beds (Collier, 1906). The Cape Lisburne exposures contain numerous silicified upright-standing tree stumps and trunks that extend into the overlying argillaceous-carbonaceous siltstones and sandstones. These Visean coal seams are part of the terrigenous sequence beneath the diachronous marine carbonate transgression.

The abrupt disappearance of the colonial rugose corals early in zone 16i (Visean) represents a cooling and local increase of aridity that remained through zone 18 (Namurian). The Middle Carboniferous (zones 20 and 21, upper Namurian and lower Westphalian) rocks of the Wahoo Limestone, which contains calcareous algae flora and rugose corals, reflect sedimentation in warmer water and repeated, thin, shoaling-upward cycles from oolitic sands to intertidal dolomites. This distinctive sequence of cyclic facies of shoaling-upward sequences of beds extends over a wide area in the subsurface of the North Slope from the Sadlerochit Mountains to the Prudhoe Bay oil fields.

According to paleomagnetic reconstructions, New Mexico (figs. 4, 5) was some 8° south of the paleomagnetic equator at the beginning of the Carboniferous. By Westphalian (Middle Carboniferous) time, the paleomagnetic equator crossed New Mexico from the northeast to the southwest (Habicht, 1979; C.R. Scotese, written commun., June 1991). Some 94 genera and 113 species of Early Carboniferous foraminifers and algae are recognized in these (100- to 1,400-ft-thick) cratonic carbonate rocks. This is considerably less than the Tethyan fauna and flora in which well

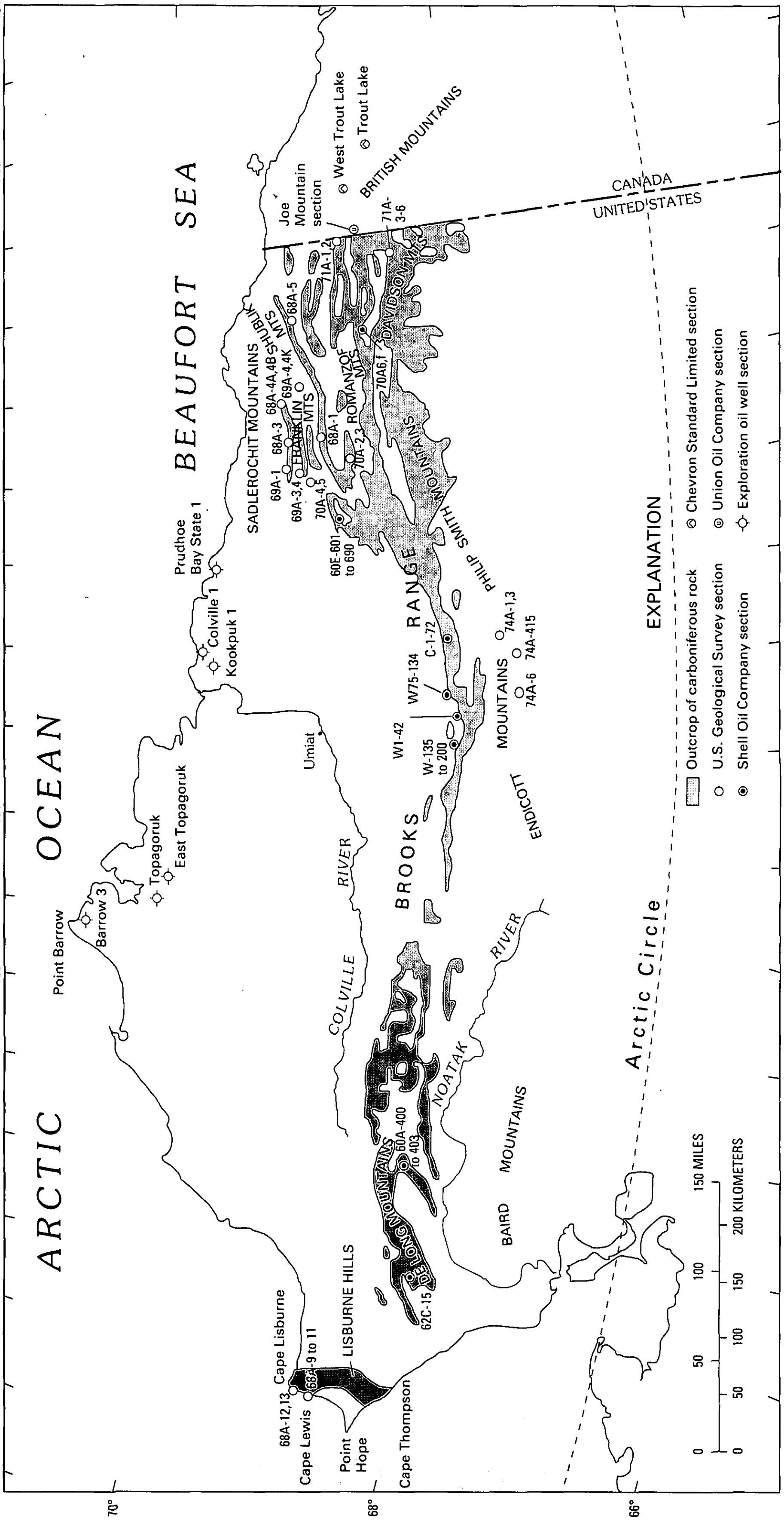

Figure 1. Index map of arctic Alaska and the Brooks Range illustrating Carboniferous outcrops and location of the measured sections described in this report (modified from Armstrong and Mamet, 1977).

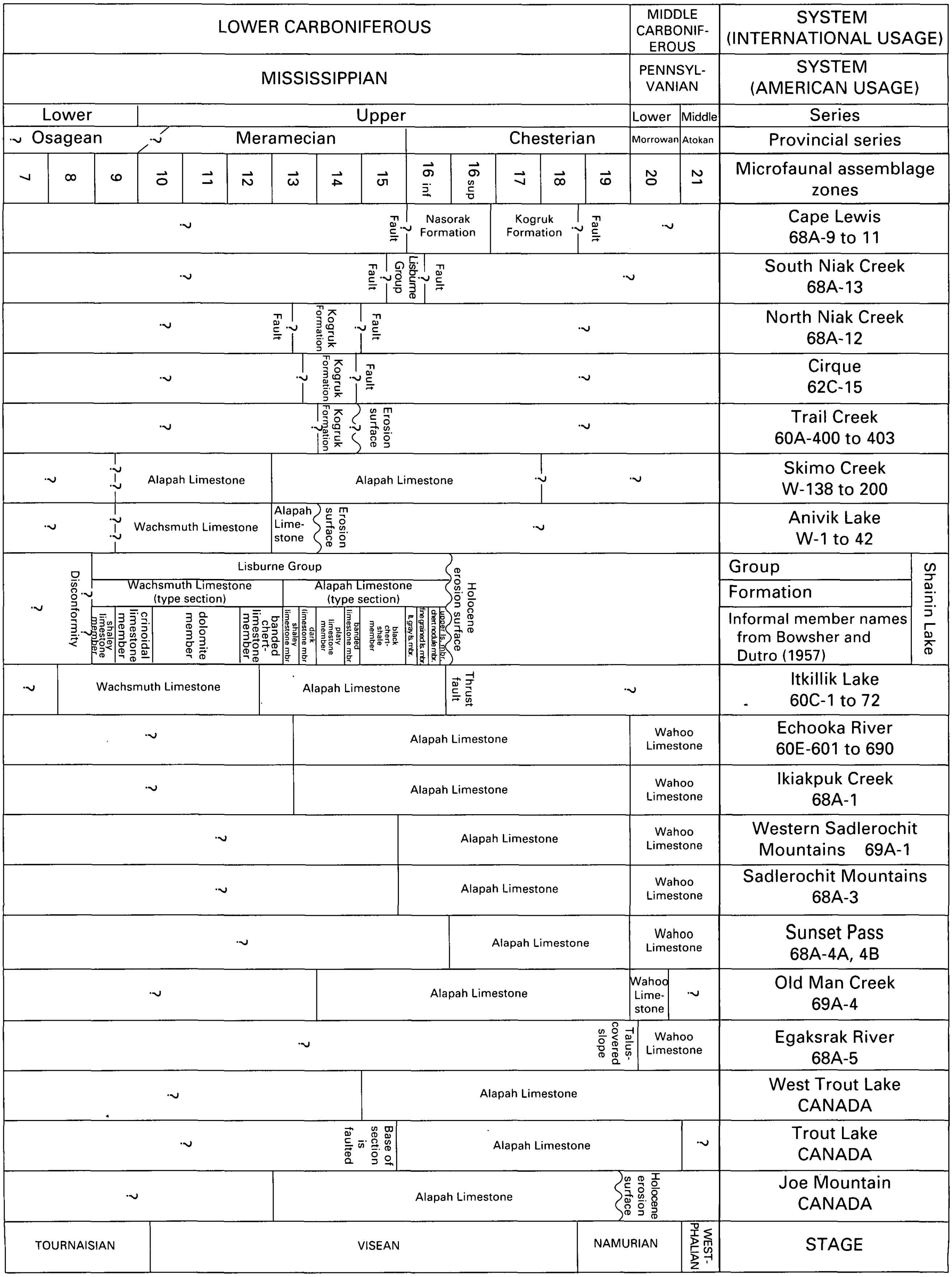

Figure 2. Regional correlation diagram for the Lisburne Group (from Armstrong and Mamet, 1977).

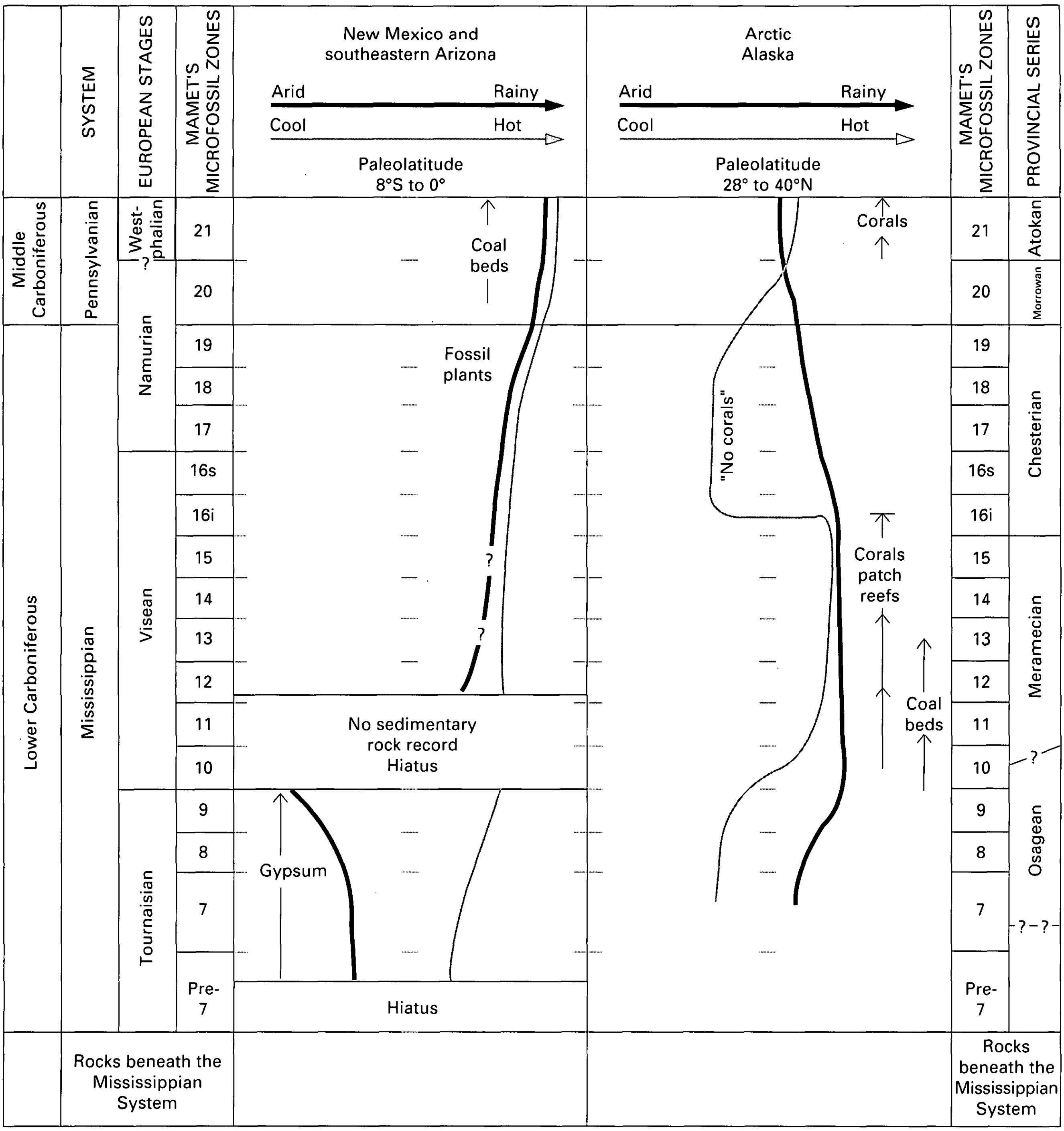

Figure 3. Model for the relative rainfall and temperature for the Early and part of the Middle Carboniferous for arctic Alaska and for New Mexico and southeastern Arizona. The model is subjective and is based on field and petrographic studies and the analysis of carbonate rock microfacies, including mineralogical composition, fossil content, and distribution.

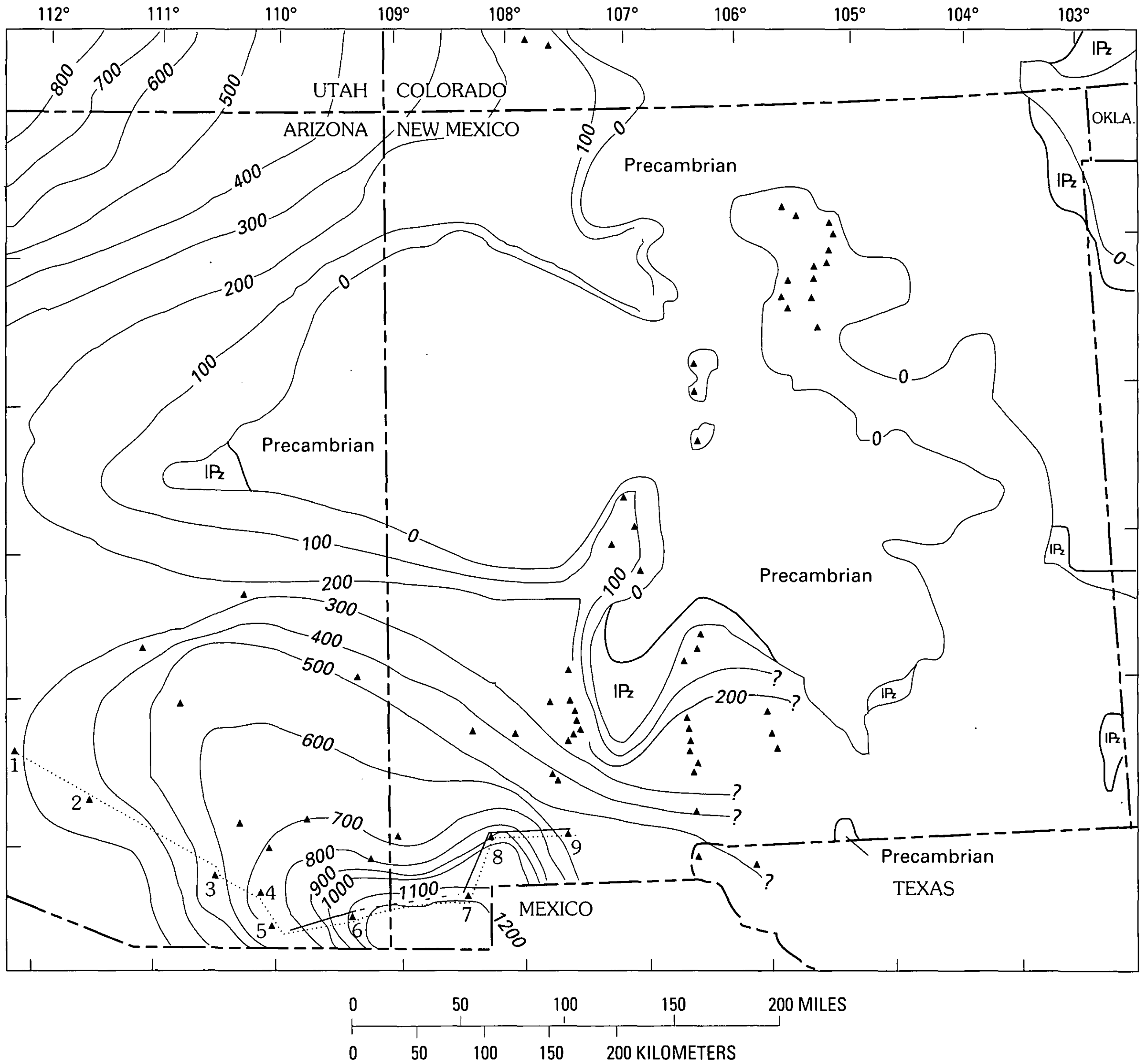

Figure 4. Index map of New Mexico and eastern Arizona showing location of outcrop sections used in this report (triangles) and isopachs (interval=100 ft) of the Lower Carboniferous (Mississippian) (from Armstrong and Mamet, 1988).

over 200 genera are recognized. Dasycladales are more abundant than those in arctic Alaska (*Alberaporella, Perkiskopora, Columbiapora*), indicating higher temperature, but again they are considerably less diverse than those in the Tethys. Thus equatorial surface temperatures should be excluded.

Rugose corals are abundant only at a specific level in the pre-7 zone and are associated with oolitic grainstones. Lithostrotionoid corals are rare in Visean carbonate rocks. During the Tournaisian and Visean, this peneplained part of the craton was fairly stable, and the region was alternately slightly emergent or submergent as a result of eustasy and (or) mild tectonism. Initial Lower Carboniferous deposits of southwestern New Mexico and southeastern Arizona (Tournaisian, pre-7 zone) are subtidal to intertidal carbonate rocks, which rest unconformably on rocks of Late Devonian age.

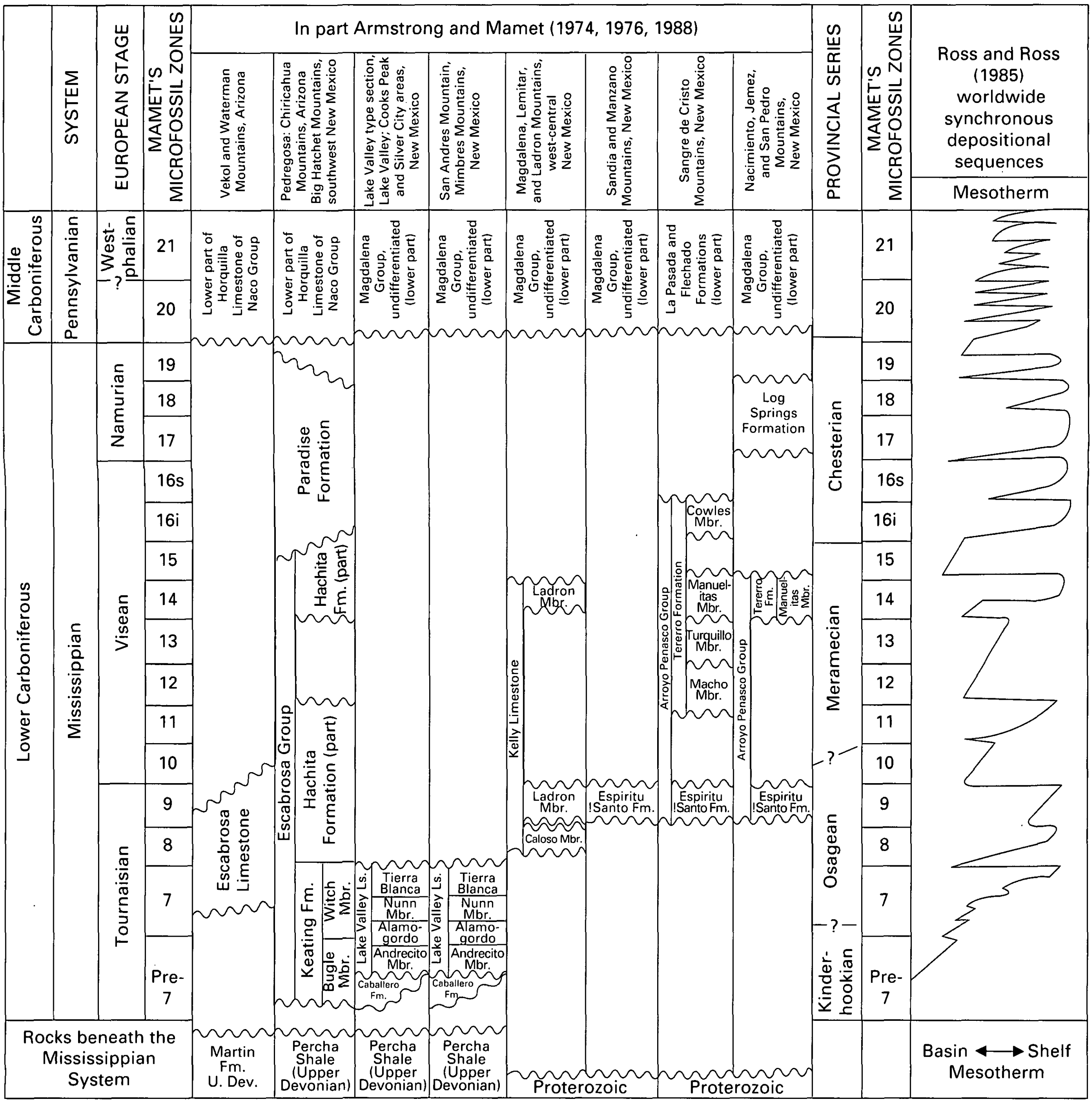

Figure 5. Correlation chart for the Lower Carboniferous (Mississippian) of New Mexico and southeastern Arizona (modified from Armstrong and Mamet, 1988). Ross and Ross (1985) diagram of mesotherms for the Lower and Middle Carboniferous are shown for comparison.

A marine transgression in the early and middle Tournaisian began in southwestern New Mexico and adjacent parts of Arizona, depositing the pelmatozoan-brachiopod packstones of the Keating Formation of the Escabrosa Group (Osagean to Meramecian). Further west in Arizona, the lower part of the Escabrosa Limestone was deposited on a shoaling subtidal to supratidal carbonate platform. Some of the pre-7 zone supratidal dolomites have zebra structures and dolomite pseudomorphs after gypsum. These indicate a hot arid climate. The Hachita Formation of the Escabrosa Group in the Pedregosa basin was deposited in shoaling water as crinoidal sand on a wide carbonate shelf. Further west in Arizona, the stratigraphic equivalent of the lower part of the Hachita Formation is the upper part of the Escabrosa Limestone, which is composed of peloid-calcareous-algal wackestone, packstones, and dolomites deposited in subtidal to supratidal environments in a hot arid climate.

By the end of late Tournaisian time, epicontinental seas had flooded much of New Mexico and Arizona. In northern New Mexico, a thin veneer of subtidal to supratidal lime mudstone and gypsum was deposited over the southern end of the Proterozoic transcontinental arch. The sedimentary record indicates that the climate in New Mexico was hot and arid in zones 8 to 9 (Tournaisian). The supratidal sedimentary deposits include an abundance of calcite pseudomorphs of gypsum and the deposition of bedded gypsum (Vaughan, 1978; Ulmer and Laury, 1984). A major regional marine regression and ensuing transgression took place during early Visean time. This event is represented by a hiatus in the upper part of massive encrinite of the Hachita Formation in southwestern New Mexico. Ross and Ross (1985) show a worldwide marine regression at the end of the Tournaisian, zone 9. The hiatus, which spans zones 11 and 13 in New Mexico, is recorded in an unconformity between shelf carbonates of the Lake Valley Limestone (Lower Mississippian) and deeper water basinal carbonate rocks of the lower part of the Rancheria Formation (Osagean to Chesterian) in south-central New Mexico. This hiatus is found in northern New Mexico in the upper part of the Arroyo Penasco Group (Osagean to Chesterian) and the Kelly Limestone (Osagean to Meramecian) of west-central New Mexico.

A major transgression occurred over the region in zone 14. In southwestern New Mexico and adjacent parts of Arizona, the cyclic, subtidal to intertidal Paradise Formation, zone 15 (Visean) through zone 19 (Namurian), contains oolitic grainstone to dolomite and interbedded plant-bearing quartz sandstones and shales. The withdrawal of marine waters off the craton in late Visean and the influx of terrigenous sediments into Pedregosa basin record the tectonism and rising highland to the north, which were the beginning of the Ouachita orogeny and the cyclic sedimentation that was to characterize the Middle Carboniferous. The lack of pseudomorphs of gypsum and the abundance of plant remains suggest a warmer and wetter climate for the latest Mississippian. The basal sandstones and shales of the Pennsylvanian in the southern Sangre de Cristo Mountains of New Mexico (zones 20 and 21) have 5-ft-thick coal beds (Gardner, 1910), which indicates a warm and humid climate during late Namurian. This is further substantiated by the reappearance of Dasycladales algae in shallow-water carbonate rocks.

These two examples of rocks taken from a rather complete geologic column and deposited on comparable carbonate platforms bordering the same continental block show how difficult it is to estimate global paleotemperatures. Local basinal variations leave paramount imprints in the record and appear to be as important as global trends.

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Climatic Influence on Basin Sedimentation—Application to the Ouachita Basin

C. Blaine Cecil and N. Terence Edgar,

A modern example of climatic influence on deep-sea sedimentation is found in the Peru-Chile trench, off the west coast of South America, and a possible example in the ancient record may occur in the Paleozoic strata of the Ouachita Mountains. The Peru-Chile trench extends from the equatorial wet zone (lat 8°N. to 5°S.), through the subtropical desert belt (lat 5°S. to 30°S.) to the northern limit of the humid temperate zone at about lat 30°S. (fig. 1). The maximum depth of the trench is 8,400 m, whereas the Andes Mountains, 350 km east of the trench, are over 7,000 m high—one of the steepest gradients in the world. Although the crustal depression formed at the subduction zone continues south of lat 30°S. to the southern tip of South America, the depression is filled with sediment and bathymetrically does not extend south of lat 30°S. into the humid temperate zone.

The coastal region of South America between lat 5°S. and 32°S. is the Atacama Desert, one of the driest regions in the world. This extreme dryness results from the combination of the presence of a high pressure subtropical belt of dry descending air, the upwelling of deep cold ocean waters off the coast, and the orographic effect of the Andes Mountains, which block the moist air of the Amazon Valley. Because South America has remained within 10° of its present latitude since Early Jurassic time (Smith and Briden, 1977), these conditions have probably been in effect since the rise of the Andes Mountains, which started in Late Cretaceous time (Andean batholith), but the mountains possibly did not become sufficiently high to be an effective barrier to moisture until the middle of the Tertiary.

Adjacent to the desert, the sediment-starved Peru-Chile trench contains pelagic, deep-sea clay so thin (<50 m) that it is not recorded by any of the seismic systems used and is totally devoid of terrigenous clastic sediment. Cores collected from the bottom of the trench consist of deep-sea clay (Lamont-Doherty Geological Observatory, written commun., 1990). Without rainfall and runoff, terrigenous sediment from the mountains cannot be transported to the sea. Sediment carried by streams resulting from infrequent rainstorms does not transport to the bottom of the trench. South of about lat 32°S., winds blow from the west to the western flanks of the mountains where they deposit moisture that flows to the Pacific Ocean. In these latitudes, rivers carry sediment to the sea, filling the trench with terrigenous sediment, and, in some places, sediment overflows the trench westward on to the Pacific Ocean floor. North of the Chile rise, the sea-floor spreading rate is about 10 cm/yr eastward (Nazca plate); south of the rise it is only about 3 to

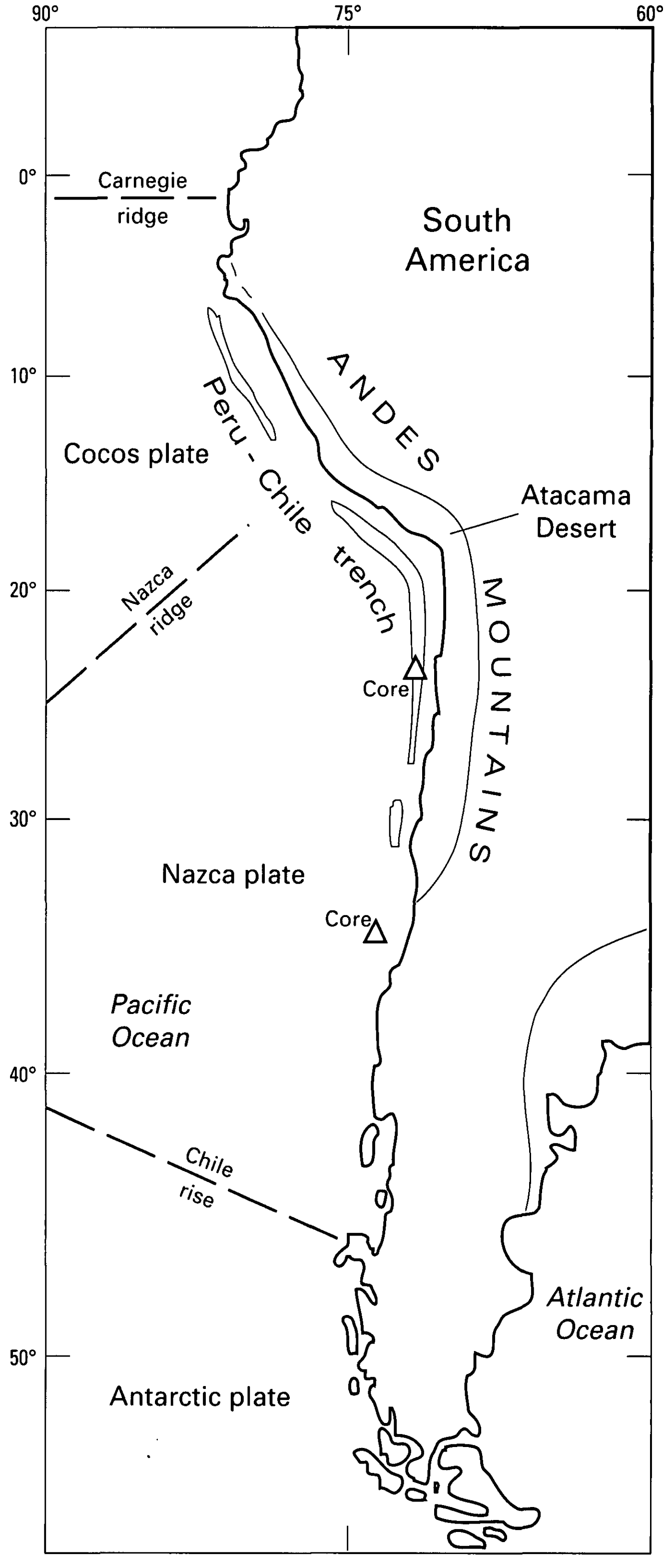

Figure 1. Map of western South America and adjacent Pacific Ocean showing location of the Peru-Chile trench, desert areas, and piston cores. The contour line in the trench is at 8,000 m water depth. The north end of the trench shallows slowly to about lat 4°S., and the south end terminates more sharply at about lat 32°S.

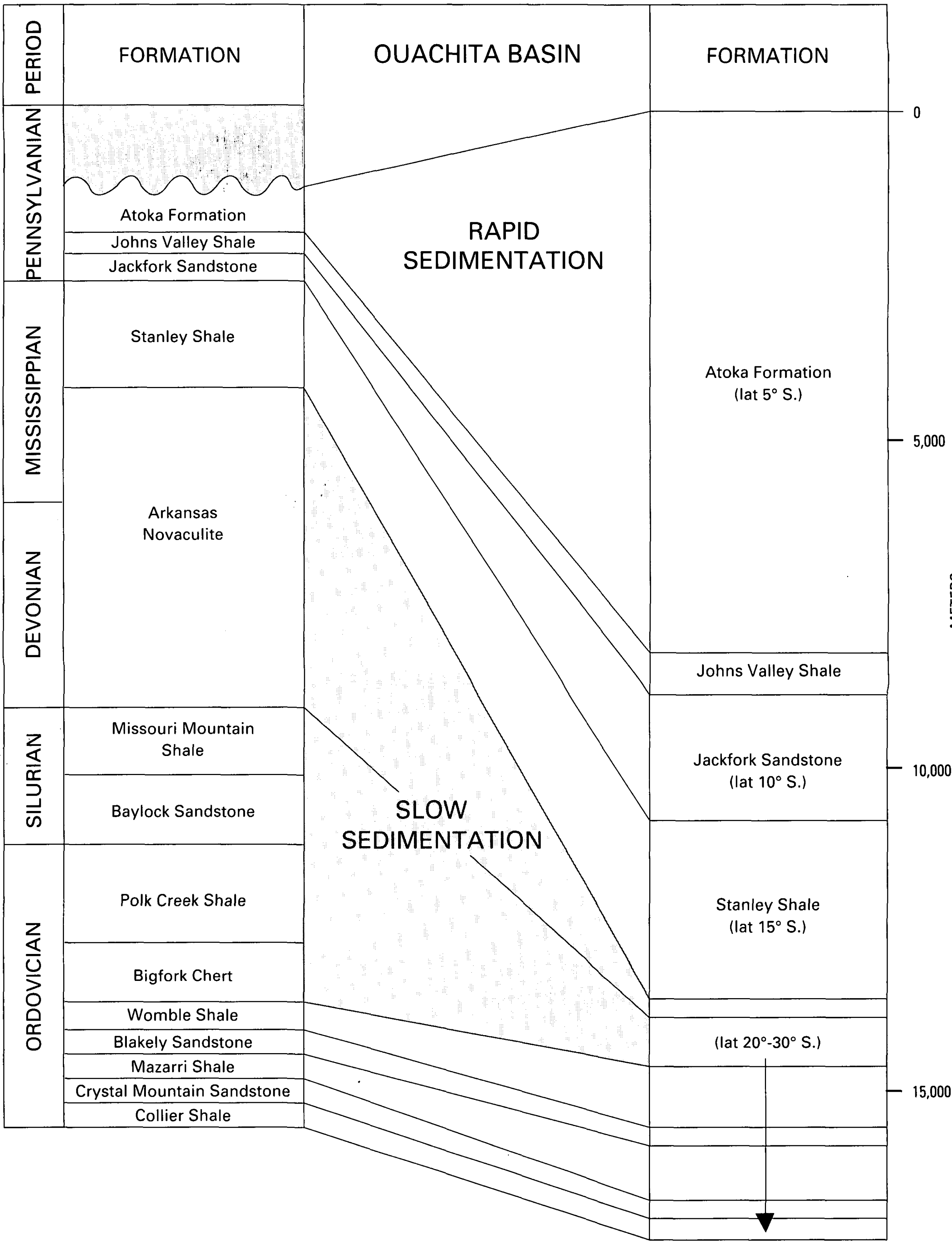

Figure 2. Relationship of time versus sediment thickness from the Ordovician through the Pennsylvanian in the Ouachita basin (modified from Stone and others, 1986). Latitude shown is at the time of deposition (Scotese and McKerrow, 1990).

4 cm/yr eastward (Antarctic plate). The rise intersects the margin of South America at about lat 46°S., which is 15° of latitude south of the transition from a sediment-free to a sediment-filled trench. It is apparent that the amount of sediment in the trench is unrelated to the change in sea-floor spreading rate across the Chile rise, but it is controlled primarily by climate (Galli-Olivier, 1969; Scholl and others, 1970; Hayes, 1974). North of the Atacama Desert, the trench is only partly filled with terrigenous sediment because the drainage area is small and heavy vegetation cover restricts the transport of sediment.

If the desert conditions of the Peru-Chile coasts were replaced by conditions of seasonal rainfall, a thick clastic sequence of sediments would be deposited on the pelagic clays at the bottom of the trench, such as that found south of lat 30°S. Such a sequence deposited in a basin and preserved in the geologic record would probably be misinterpreted as representing slow deep-water deposition followed by an orogeny that resulted in the rapid deposition of a tectonostratigraphic wedge. The interpretation is based on the perception that a large influx of terrigenous sediment is related solely, or largely, to the uplift of mountains. In the case of the Peru-Chile trench, the interpretation would be incorrect; the deposition of a thick siliciclastic wedge would be the result of the postulated change in climate.

We have demonstrated with this example, that the lack of sediment in the Peru-Chile trench is related primarily to an arid climate, which is, in part, a secondary rainshadow effect of the mountains. We have also demonstrated that by changing only the climate while maintaining the mountain system, large volumes of terrigenous sediment can be transported and deposited. Therefore, the origin of a sedimentary wedge depends as much on a wet seasonal climate as it does on tectonics; both are necessary.

It also can be demonstrated that sediment flux from a mountain range in an everwet climate is restricted by rainforests, which trap sediment beneath a blanket of organic material (Cecil, 1990 and this volume). Under these conditions, climate is the controlling factor in determining sediment flux. Therefore, we conclude that the rate of sediment flux derived from the Andes Mountains is dependent on the prevailing climatic regime, eustasy notwithstanding (Vail and others, 1977). Eustasy affected the entire west coast of South America, but there is no evidence of a change in clastic sediment input to the trench that is related to eustasy.

Sediments similar to those deposited in the Peru-Chile trench adjacent to the desert are found in the Ouachita Mountains. A thin section of shales, cherts, volcanic ash, and rare siltstones and sandstones (turbidites) was slowly deposited in a deep basin in the period from Ordovician into Mississippian time (for a summary see Lowe, 1989 and Thomas, 1989) as North America moved northward (Scotese and McKerrow, 1990) through the dry belt, from about lat 30°S. to about 10°S. (fig. 2). Beginning in Late Mississippian time, there was a significant increase in coarse clastic sedimentation as the Ouachita basin began to move into the tropical seasonally wet belt (between lat 10°S. and 5°S.), and, in the Middle Pennsylvanian (Atokan) time, about 8,500 m of coarser clastic sediment was deposited (Stone and others, 1986; Houseknecht, 1987). The clastic sediment deposition is generally attributed solely to tectonic activity. However, we believe that the movement of the deposystem into a wetter regime at the time of increased siliciclastic deposition may also have been a cause for the introduction of abundant terrigenous sediment and thus paleoclimate may have played a significant role in influencing the resulting thicker stratigraphic sequence. This interpretation is supported by the following factors: (1) the influx of siliciclastic material, (2) input of clean (mature) quartz sandstone (for example, Jackfork Sandstone), and (3) the influx of abundant terrestrial organic matter.

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Cyclic Eolian Sedimentation—A Climatic Response

Thomas S. Ahlbrandt

The purpose of this paper is to discuss the origin and significance of cyclic eolian sedimentation in both modern and ancient eolian sequences. The concept of cyclic eolian sedimentation and the influence of climate change can be demonstrated in both modern and ancient eolian examples.

MODERN EOLIAN DEPOSITIONAL CYCLES

Many climatic change indicators are recorded in eolian deposits. Eolian deposits are extremely sensitive to many types of climate change that may be reflected in terms of wind direction and strength, temperature, vegetation, and moisture changes, all of which are readily reflected in eolian sequences. For example, moisture conditions—be they dry, wet, or damp—produce diagnostic primary and secondary sedimentary structures in eolian deposits (Ahlbrandt and Fryberger, 1982; Kocurek and Nielson, 1986); temperature conditions (McKee, 1966, 1979; Ruegg, 1983; Nissen and Mears, 1990) and wind strength and duration variations are recorded in dune forms (Fryberger, 1979; McKee, 1979). As has been pointed out by Ahlbrandt and Fryberger (1980), wind energy is high in the Nebraska Sand Hills. This largest dune field in the Western Hemisphere is now stabilized but would become active if annual precipitation were to decrease to 50 percent of its present amount (Swinehart, 1989), demonstrating potential rapid destabilization of a major sand sea by a rather modest climatic change.

Extremely arid conditions are recorded in eolian deposits by detrital evaporites in the dune itself as well as intercalated allocyclic deposits such as halite, anhydrite, and gypsum. For example, Fryberger and others (1983) document the incorporation of detrital gypsum and anhydrite in the Jafurah Sand Sea adjacent to the Arabian Gulf. The preservation of the full range of eolian to nearshore marine sequence in a sediment cycle in an arid sabkha and coastal dune setting is described by McKenzie and others (1980) for the Arabian Gulf. This sabkha cycle commences with subaerial eolian quartzose-carbonate sand that grades upward into a transgressive sandstone and then to lagoon-intertidal algal mats. These sequences in turn are overlain by subtidal carbonate mud and sands that are overlain by an upper intertidal gypsum mush with laminated algal mats. In turn, the gypsum mush grades upward in supratidal deposits that incorporate minor amounts of eolian sand, overlain by supratidal chickenwire anhydrite, returning to eolian deposits initially crusted with halite and finally buried completely by eolian sand. The association of black muds with these modern sabkha sequences has been demonstrated by Ahlbrandt and Fryberger (1982) and provides a modern analog for ancient eolian deposits, which is discussed in the following section.

Cyclic sedimentation in eolian deposits is demonstrated in interior continental settings as well as the coastal setting described above. Eolian sequences must have had rapid climatic variations from dry to wet for radiocarbon dating to be used. Disruptions in the eolian sequence, which result from transitions between arid and semiarid conditions, may not be recorded in the form of organic material but are commonly recorded as a hiatus or unconformity, which in eolian sequences is called a bounding surface.

Bounding surfaces are divided into first, second, and third order. The most extensive first-order bounding surfaces are also called super surfaces. There is much discussion concerning their origin related either to rising water tables (Stokes, 1968; Fryberger and others, 1988), deflation associated with climbing bedform migration (Kocurek, 1988), or climatic control of the highest order eolian bounding surfaces (Talbot, 1985). This argument will not be settled here; however, rising and falling water tables considered as a controlling mechanism for eolian sequences reflect the significant role of climate change in eolian sequences even in interior settings.

From the available chronologic studies of dune fields, it is clear that climatic changes are rapid even in interior continental settings. Two recent North American examples are given. In a study of the Nebraska Sand Hills and several intermontane basins of the Rocky Mountains, Ahlbrandt and others (1983) demonstrated a series of at least four major pulses of eolian activity separated by wetter, pluvial phases with associated alluvial deposits in the past 11,000 years. The arid cycles range from 500 to 2,000 years in duration at the level of accuracy of the study. More recently, Gaylord (1990) incorporated new ^{14}C data into paleoclimatic interpretations of an eolian sequence in the Ferris Dunes of Wyoming and defined more accurately the duration of arid and moister conditions that are represented by dune and interdune sediments, respectively. Gaylord (1990) documents six paleoclimatic intervals in the last 7,500 years, including 7,545 to 7,035 years ago, arid; 7,035 to 6,460, transitional; 6,460 to 5,940, moist; 5,940 to 4,540, arid; 4,540 to 2,155, transitional; and 2,155 years ago to the present, moist and semiarid. Duration of these cycles respectively are 510, 575, 520, 1,400, 2,385 and 2,155 years.

Climatic fluctuations ranging from decades to centuries to millennia across North Africa are further evidence of modern cyclic climatic changes in eolian terrains (Petit-Maire, 1990). For example, spectacular changes in surface hydrology throughout the Sahara from the Atlantic Ocean to the Red Sea occurred in a 500-year interval from 9,000 to 8,500 years ago. Talbot (1985) argues that in the Sahel and southern margin of the Sahara, large areas of dunes were active in the late Pleistocene and are now stabilized due to a climatic change to more humid, less windy conditions in the Holocene. New dating techniques will likely increase resolution. Nonetheless, it is becoming clear that Quaternary eolian deposits record a cyclic sedimentation pattern that is related to if not dominated by climate change.

ANCIENT EOLIAN DEPOSITIONAL CYCLES

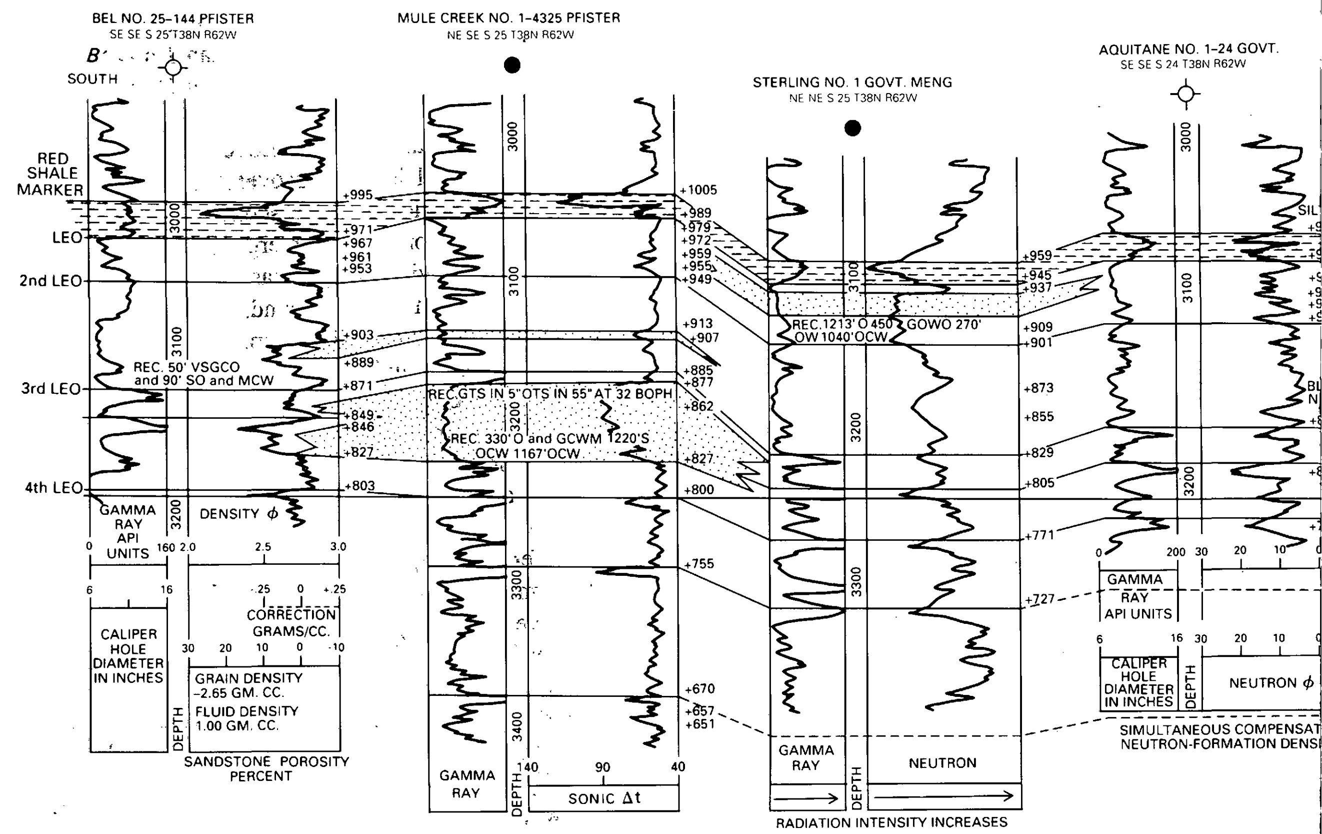
Because datable horizons are uncommon in some eolianites, much effort has been directed toward correlating bounding surfaces (that is, diastems or unconformities) in such rocks (for example, Kocurek, 1988). Inherent in most such analyses is the philosophy that an understanding of eolian processes alone is adequate to explain the nature of deposits preserved in the eolianite. Eolian sequences can range from entirely dune-interdune sequences to dune-extradune sequences (Lupe and Ahlbrandt, 1979). The preservation of individual bedforms among time equivalent non-eolian deposits (extradune) represents an endmember of the eolian system that allows greater correlation of eolian and non-eolian depositional events and the controlling mechanisms such as climatic or eustatic change, which affects them. Understanding such sequences becomes economically important where the dune sandstones become hydrocarbon reservoirs and the associated extradune sediments become the source of the hydrocarbons. Such is the case for the Pennsylvanian "Leo Formation," which is a hydrocarbon-producing eolianite in eastern Wyoming and western South Dakota as discussed by Tromp (1981), Cardinal and Holmes (1984), Desmond and others (1984), and McBane (1984). The "Leo Formation" is an unranked unit in the middle part of the Minnelusa Formation.

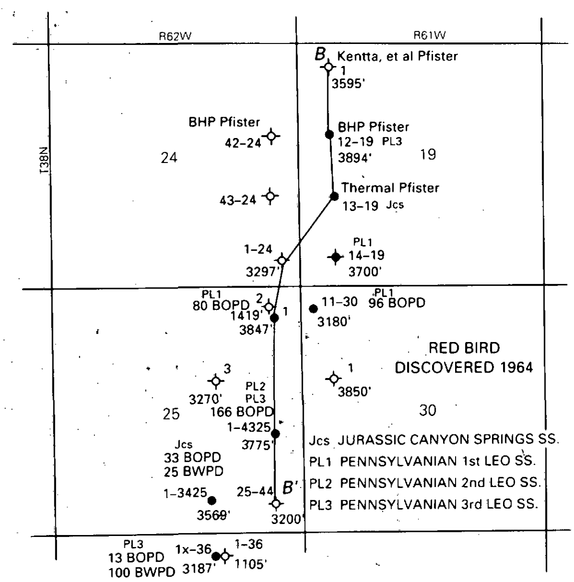
For this paper, the Leo Formation of Pennsylvanian age (Desmoinesian-Virgilian) in the vicinity of the Red Bird oil field in the southeastern part of the Powder River basin in Niobrara County, Wyoming, is used to demonstrate the nature of a cyclic dune-extradune sequence and to discuss its origin and importance (fig. 1). The stratigraphic nomenclature for this Pennsylvanian unit is geographically controlled. In the Black Hills, these rocks are called the Leo Formation, and this terminology is used throughout the subsurface of the southeastern Powder River Basin. Elsewhere in the Powder River Basin, these rocks are referred to as the middle Minnelusa Formation, and, where they outcrop in the Hartville uplift of southeastern Wyoming, they are called the middle member of the Pennsylvanian Hartville Formation (Tromp, 1985). The Leo Formation or Hartville Group was considered to represent middle marine shelf deposition (Condra and others, 1940; Love and others, 1953; Momper, 1963) until it was discovered that eolian deposits occurred in the middle of the sequence (Tromp, 1981). Considerable revision of the interpretation of the original depositional environments has subsequently occurred. Confusion arose from the abundant fossils that occur in the carbonate units with little appreciation being given to intercalated clastic units. The entire sequence is now viewed as nearshore marine to supratidal to subaerial (for example, Desmond and others, 1984), and a eustatic-controlling mechanism for the origin of the cycles is suggested by Cardinal and Sherer (1984). However, a eustatic model alone does not explain significant lithologic and compositional differences observed among the Leo cycles.

The Leo Formation is overlain by a thin (<10 m) but regionally extensive red shale, locally referred to as the "red shale marker," which is considered to represent the Pennsylvanian-Permian boundary on the basis of fusulinids found above and below the unit (Desmond and others, 1984). As discussed by Cardinal and Sherer (1984), the red marker is generally soft, micaceous, and brick red, and its origin is unknown, although interpretations range from a paleosol to an alluvial or estuarine deposit. The Leo Formation includes Desmoinesian, Missourian, and Virgilian sediments overlying deep reddish to maroon shales of Atokan age.

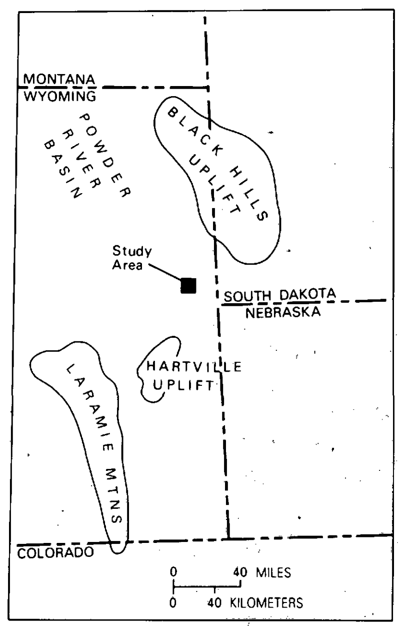
The Leo Formation can be informally subdivided into at least six intervals (cycles) on the basis of laterally extensive black shales, which have a strong gamma-ray signature on well logs (fig. 1). Although there are probably several orders or cycles of differing time scales represented in the Leo (fig. 1), this paper focuses on several-million-year cycles including two Virgilian cycles (1st, 2nd Leo), two Missourian cycles (3rd, 4th Leo), and two Desmoinesian cycles (5th, 6th). Interpreted cycle lithology and depositional setting are shown particularly for a Missourian cycle (3rd Leo, fig. 1).

Hydrocarbons have been produced from sandstones in the upper three cycles in the Red Bird area (fig. 1). The interpretation of these cycles has significant consequences for location and extraction of hydrocarbons as well as for chemical attributes of the oils. For example, hydrocarbons have been produced from the lower Leo cycles (4th, 5th, 6th, fig. 1) in fields 6 to 10 miles west of the Red Bird field. Unfortunately some of these lower sands have high H_2S content (as high as 48 percent) resulting in considerable





RED BIRD FIELD AREA, NIOBRARA COUNTY, WYOMING



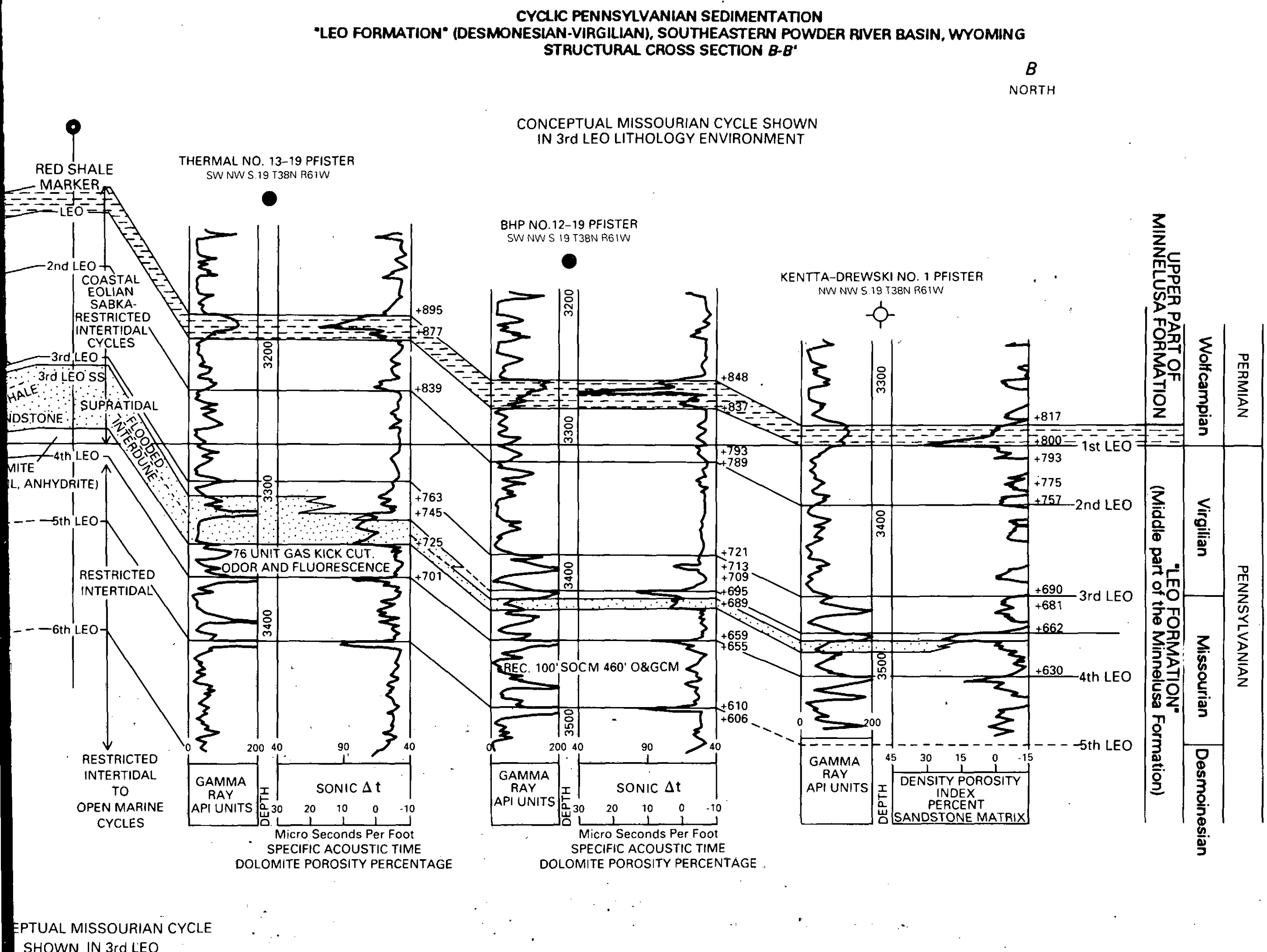


Figure 1. Structural cross section *B–B'* demonstrating six Pennsylvanian age cyclic hydrocarbon reservoir sequences of the "Leo Formation," Red Bird oil field area, eastern Wyoming. Reservoir cycles, bounded by radioactive shales, range in age from Desmoinesian (PL5, PL6) to Missourian (PL3, PL4) to Virgilian (PL1, PL2) based upon surface and subsurface faunal zonations in the Hartville uplift area (Love and others, 1953) and in part upon recent conodont studies (B. Wardlaw, U.S. Geological Survey, oral commun., 1994). Three separate Leo hydrocarbon accumulations are shown on *B–B'*; a PL1 accumulation is bordered by PL3 accumulations to the north and to the south as also outlined on the map view of cross section *B–B'*. Cycles commence with a black shale and are overlain successively by carbonate, siliciclastic, evaporite, carbonate, and finally another shale deposit, which marks the beginning of the next cycle. The varying depositional environments of the successive cycles are demonstrated in the conceptual well (fourth from the right) in cross section *B–B'*; eolian reservoirs of PL3 are the thickest and of the best reservoir quality, and eolian reservoirs of PL1 are of the next best quality of the six reservoir cycles. Note the intertonguing of a black shale and an eolian sand within the 3rd Leo (PL3) reservoir cycle suggesting alternate wet and dry climatic conditions.

expense to operators to protect both employees and the environment from such toxic gases. Tromp (1985) argues that the lower Leo (5th and 6th, Desmoinesian) cycles represent more open marine settings relative to younger Pennsylvanian cycles.

The generalized vertical sequence of lithofacies of a Leo cycle varies by region and author. For example, Desmond and others (1984) describe the following Leo cycle (that is, transgressive sandstone, subtidal, intertidal, supratidal, eolian dune, and transgressive sandstone) for the southwestern Black Hills. In the northwestern Black Hills, Desmond and others (1984) modify the cycle to transgressive sandstone, flooded interdune (lenticular evaporites and black shale), eolian dune, transgressive sandstone, eolian dune, transgressive sandstone, subtidal, and intertidal. The interpreted vertical sequence for the Leo Formation in the Red Bird area of the southeastern Powder River Basin is shown on figure 1 for a Missourian cycle (3rd Leo) that contains the thickest eolian dune deposits in the Leo Formation (this is the 2nd Leo of Cardinal and Sherer, 1984, and the lst Leo B of McBane, 1984). Because black shales interfinger (flooded interdune) laterally with dune sands in the 3rd Leo interval, a modified form of the northwestern Black Hills Leo cycle is used on figure 1. The Missourian Leo (3rd Leo) is a prolific hydrocarbon reservoir with 788 barrels of oil/acre-ft recoveries at Alum Creek field in South Dakota and high recoveries at Buck Creek field in Niobrara County (about 12 miles west of Red Bird). As shown on figure 1, 3rd Leo sandstones of virtual original porosity (for example, 3rd Leo sandstone in the Pfister 1–4325 well has 28–30 percent porosity) demonstrate minor diagenetic modification of the original deposits.

The black shales that bound the cycles are of considerable scientific and economic interest. These shales are described as being thin (generally less than a meter), carbonaceous, and radioactive and contain organic constituents including acritarchs, gymnosperms, pregymnosperm spores, cuticle fragments, amorphous fecal material, wood fragments, resin, and charcoal (Tromp, 1981; Desmond and others, 1984). These shales are rich hydrocarbon source beds, have organic carbon contents as high as 20 percent, and are believed to be the hydrocarbon source for oils produced from the prolific Minnelusa oil fields in the Powder River Basin (Clayton and Ryder, 1984; Desmond and others, 1984). The black shales are most commonly found above fossiliferous carbonates and below laminated dolomite and chickenwire anhydrite (Desmond and others, 1984). However, in outcrop as well as in the subsurface (for example, see the black shale in the 3rd Leo cycle in fig. 1), black shales are interbedded directly with eolian dunes as in the "flooded interdunes" of Desmond and others (1984) northwestern Black Hills Leo cycle model.

The origin of the black shales, whether eustatically or tectonically controlled (for example, deep basin origin as is postulated in the midcontinent for correlative units), is of significance for exploration models. The tectonic or eustatic model alone fails to explain the flooded interdune black shales that intercalate with eolian deposits as described by Desmond and others (1984) in outcrop and are also observed in the subsurface as shown on figure 1.

Much remains to be explained, and climate considerations could greatly aid in our interpretation of this and other sequences. It could be as easily argued that there were periodic wet and dry cycles on a low relief, marginal coastal setting (sabkha) that resulted in a series of repetitive sedimentary sequences along a gradually subsiding coast. Visits to localities in the Hartville uplift show silcretes preserved in the upper part of the carbonate sequences underlying the dunes. The formation of the highest order bounding surfaces in eolian sequences in response to climatic change (Talbot, 1985) is strongly supported in my view not only for the Leo but many ancient eolian sequences. Preconceived notions inhibited our understanding of the Leo for many years, and, until we can accommodate a model that explains all the rocks preserved, new mechanisms will be called upon. The role of climatic cycles I believe is recorded in the Leo and provides a viable option to relying exclusively upon eustatic models for this particular eolianite as well as other sequences that contain less easily recognizable climatic changes. New concepts are required to explain the differences among the Leo cycles, for eustatic changes alone provide little explanation for lithologic and chemical changes among the cycles.

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Paleoclimatic and Sea-Level Effects on a Range of Metallic Mineral-Deposit Types

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Changes of climate and (or) sea level have a wide range of effects on the deposition of metallic minerals. These effects are discussed by deposit type. For brevity and simplicity, nonmetallic deposits are omitted from this discussion. However, somewhat similar considerations apply to barite, phosphorite, evaporites, clays, and other industrial minerals and to the fuels uranium and thorium.

Sedimentary deposits.—The influence of climate and sea-level change is of course greatest in truly sedimentary (syngenitic) deposits. For fluvial deposits, climate is of greater importance, but sea level is the base level for the system. For those deposits that are of marine or shoreline origin, climate and sea level—commonly coupled (Fischer, 1982)—are among several variables such as basin geometry, geochemical input, and hydrographic behavior that control oceanography of a watermass. Some climatic and chemical factors may be imported to a given sedimentary environment from the far margins of a basin, which may be in different climate zones.

Sedimentary mineral deposits may be divided into mechanical (placer) and chemical deposits. Placer deposits of metals include those of gold, tin, titanium, and platinum. Economic placer minerals are not only dense but also are relatively inert to weathering. Weathering beneficiation occurs after deposition also. Weathering history is sufficiently important to limit economic Quaternary shoreline ilmenite deposits to latitudes of less than 35° (Force, 1991). Different atmospheric compositions as well as different climates may influence the weathering process, as in Precambrian gold placers and Cretaceous monazite-rich placers.

In fluvial placer deposits, an effect of dry climate is to prevent the intricate sorting required because the suspension thickens as a flow loses water into its bed. In shoreline placers, alternation between storm-dominated and fair-weather shore profiles permits placer preservations. Sea-level change activates different sediment-supply systems on the shoreline, some deleterious and some beneficial to a valuable mineral assemblage (Force, 1991).

For the chemical sedimentary deposits in the marine environment (iron, manganese, some base metals), climate and sea level may influence water composition by controlling evaporation rate, weathering rate, and dissolved-gas content. Water-column stratification is also of great importance because of the sensitivity of the metals to redox changes. Manganese, for example, characteristically occurs in shallow-marine deposits where the precipitate that forms on a water-column redox interface is preserved on the basin margin (Force and Cannon, 1988). The necessary stratification generally results from both warm climate and high sea-level stands. Thus, black shale pinchouts are common stratigraphic settings of these manganese deposits.

Sedex deposits.—Hydrothermal venting in the marine environment forms sedex (sedimentary-exhalative) deposits of zinc and other base metals. Formation of such deposits may be independent of climate and sea level, but preservation generally requires anoxic bottom waters (Force and others, 1983) that form as a function of climate, sea level, and other variables.

Diagenetic deposits.—Deposits of diagenetic origin may show direct influence of climate and (or) sea-level change as a repartitioning of interstitial water between connate and meteoric sources (Force and others, 1986). More commonly, these deposits involve factors inherited from the sedimentary environment. For example, a recent model (Maynard, 1991) for diagenetic copper-cobalt deposits requires eolian sands with soluble metal-oxite coatings, overlain by black shale on which the mobilized metals may precipitate. This model implies a rather intricate climate and sea-level history that precedes final metal precipitation.

Hydrothermal deposits.—Like diagenetic deposits, hydrothermal deposits hosted by older sediments may follow sedimentary features that were influenced by sea level and paleoclimate. For example, Titley (1991) finds a strong preference of epithermal gold deposits for older sedimentary hosts formed during periods of ocean stratification. Whether the host sediments functioned only as a trap or also as a gold source is presently unclear.

Even the igneous-hosted hydrothermal deposits have not escaped paleoclimatic imprints. Stable-isotope work has shown great degrees of entrainment of meteoric water by hydrothermal systems (Taylor, 1979). The mass balance established by availability of meteoric water may determine the geometry and grade of any resulting deposition.

Supergene enrichments.—Some metal deposits such as those of aluminum, iron, nickel, gold, and copper-molybdenum are enriched on weathering surfaces. In some such deposits, several weathering surfaces are separated by depositional units, thus recording the relative importance of several paleoclimates.

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Petroleum Resource Evaluation in the Predictive Stratigraphic Analysis Program

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An important goal of our research at the Kansas Geological Survey is the application of our knowledge of midcontinent Pennsylvanian strata to assist industry in more efficiently developing resources from these rocks. The future of petroleum research and the nature of our involvement will be affected by several factors. (1) Independent oil and gas companies are and will remain the primary domestic oil and gas operators in this region and are among the audiences that we will need to address. (2) The continuing low prices for petroleum make it essential that new strategies for petroleum exploration and development be implemented to keep the domestic oil and gas industry viable. (3) The Carboniferous will continue to be a target of opportunity in the midcontinent and in other Paleozoic cratonic and foreland basins; the search and development of Pennsylvanian reservoirs has continued to be successful, and the potential remains for additional development. (4) Predictive geologic models must be developed. To be of practical value for locating and exploiting oil and gas reservoirs these models must be carefully constrained and must include all the factors that significantly affect sedimentation, such as climate, eustasy, and tectonics. The parameters will be obtained through interdisciplinary studies, such as predictive stratigraphic analysis (PSA), on a scale and effort similar to the Deep-Sea Drilling Project. Also, the building and testing of a predictive model will require geologic data bases of increasingly greater precision and accuracy. (5) The modeling of time-series changes in parameters and their interaction will require dynamic models that incorporate the rates, durations, and magnitudes of input parameters. The results of modeling will aid in testing and experimentation and will further dictate the types of data that must be collected.

Oil production from larger established fields continues to dominate our oil and gas production base. Our larger established fields may offer the best hope of sustained domestic production because they offer a significant potential for timely increases in production through infill drilling and the application of improved oil recovery methodology. For example, if oil production from existing fields in Kansas was increased from the present average of 33 percent of the original-oil-in-place (OOIP) to 54 percent, an additional 2.5 billion barrels of oil would be recovered. The additional recovery would come at the expense of increased use of materials, manpower, and energy. Improved efficiency of operations is of immediate concern and should be focused on tailoring the latest extraction technology to a field's reservoir heterogeneity.

The Mississippian and Pennsylvanian sandstones and carbonate reservoirs in the midcontinent are now the primary targets for exploration and development drilling. While OOIP and ultimate production are weighted equally between the upper (post-Devonian) and lower Paleozoic strata, most current oil and gas production comes from upper Paleozoic strata. Marked vertical and lateral reservoir compartmentalization in relatively thick stratal intervals typifies the Mississippian to Pennsylvanian interval of the oil-producing section in the southern midcontinent. Complex vertical stacking of strata, a hallmark of the late Paleozoic, is due to widespread, generally abrupt shifts in depositional facies and early diagenetic patterns in time and space. These shifts are attributed by most workers to varying combinations of change in the mechanisms responsible for sedimentation, such as climate and associated oceanographic effects, eustasy, and tectonism.

Structural considerations need to be factored into predictive models of petroleum reservoirs. Subsidence and accommodation space for sediment vary, in part, with distance from orogenic activity and along sites of reactivation of zones of basement weakness, and such variability is manifested in changes in thickness and lithofacies in Paleozoic strata from central Kansas. Variations in subsidence and sediment accommodation space can occur across local structures leading to significant changes in lithofacies and diagenesis.

Qualitative and quantitative, hierarchical, time-series stratigraphic studies have shown that the scale of changes in facies and processes in time and space are complex. Accordingly, the characterization and prediction need to be done at ranges of distance and time scales sufficient to capture these changes in the rocks. The broad nature of the problem and complexity of data requirements necessitate an interdisciplinary effort in order to accomplish these objectives in a realistic time frame.

One method of integrating this varied information is through dynamic stratigraphic modeling (simulation). To date, a variety of models have been developed that operate on processes of sedimentation (through the use of first principles or approximations), global changes in sea level, and subsidence. Climate has been considered in only an ancillary way but needs to be more heavily factored into these efforts as more is discovered about climatic influences on the development of sedimentary sequences via research efforts such as the PSA program. Stratigraphic simulation is a vehicle to quantitatively analyze interactions of mechanisms (that is, conduct experiments) to visualize what may not be intuitively obvious and to help to focus further research. The modeling also can facilitate communication among contributors that have varied expertise.

An example of a near-surface and surface reservoir analogue study from southeastern Kansas illustrates the utility of continuous coring in the development of a data base suited for two-dimensional simulation modeling. The area is characterized by a succession of Upper Pennsylvanian cyclothems that undergo abrupt lateral facies and other stratigraphic changes due to relief along a shelf-to-basin transition. Reconstruction of observed cyclothemic sedimentation and oolitic reservoir development is being accomplished using a two-dimensional stratigraphic simulation program written by J.A. French. The modeling has proven to be a useful tool in testing controls on reservoir development and heterogeneity and will eventually be useful to transfer the results of the analogue study to actual field situations.

We believe that this is a good opportunity to draw together workers with expertise in different areas in order to develop better, more constrained geological models of

Pennsylvanian stratal sequences. The creation of a robust sedimentation-stratigraphic-geochemical-structural model of Pennsylvanian cyclothems is likely too involved an effort for any one individual. Geological surveys are especially well structured to undertake an integrated approach to such efforts.

Use of a General Circulation Model to Simulate Paleoclimates and Evaluate Economic Resources

George T. Moore, Darryl N. Hayashida,
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Paleogeography, together with thermally and tectonically induced topography, sea-level fluctuations, Milankovitch cycles, and the chemical state of the atmosphere control paleoclimate. In turn, paleoclimate creates the environments in which sediments were deposited on continents and their margins. Since the acceptance of the theory of a plate-tectonically active Earth, various methods have been used to study the relationships among paleoclimate and differing paleogeography in various time intervals.

Computerized general circulation models (GCM) can be valuable tools in recreating the paleoclimate of brief geologic time intervals that are significant in the Earth's history. Any three-dimensional GCM is a complex integrated collection of mathematical formulae, algorithms, and parameterizations. The models were designed originally to simulate present-day global climate. Testing the model with present-day boundary conditions shows that GCM's reproduce the climate of today's world rather well. Chevron uses a GCM, termed the community climate model (CCM), which was developed at the National Center for Atmospheric Research in Boulder, Colo. The CCM uses a nine-level atmosphere that is coupled thermally, but not dynamically, to a one-level ocean. The CCM uses a 4.5° latitude by 7.5° longitude grid. Part of the fully resolved hydrologic cycle is dynamic, and part is parameterized. The CCM has been adapted by several geologists to model paleoclimates. Using a version of the CCM, we report the results from multiple seasonal simulations of the Late Permian and Late Jurassic. Certain time frames exist in which the arrangement of tectonic plates relative to one another has changed slowly enough that we can model the temporal evolution of paleoclimate by using several simulations at critical thresholds. The rationale and approximation are reasonable considering the fragmentary nature of the geologic record, the imprecision of plate location, the poorly constrained paleoatmospheric greenhouse gas concentrations, and the large grid cell size of current GCM's. The formation, existence, and disintegration of Pangea lends itself to such a study.

Pangea's early Late Permian (255–252 Ma, Kazanian) formation created a chain reaction of events that produced an inhospitable climate by disrupting zonal atmospheric circulation, an elevated greenhouse effect that warmed the planet, and falling sea level that severely restricted the shallow-water marine environments. Collectively they produced an extinction event that progressed over 10 to 15 million years and was the most severe of the Phanerozoic Eon. Disintegration of Pangea occurred approximately 100 million years later. The Late Jurassic Kimmeridgian and Tithonian Stages (154.7–145.6 Ma) represent a time when seaway connections became established between northern and southern Gondwana, North America, and Eurasia. Rising global sea level throughout the Jurassic flooded large parts of the continents. These conditions contributed significantly to climate amelioration by the middle of the Mesozoic.

The simulation of a Late Permian warmer Earth with an elevated greenhouse effect fits geologic observations and isotope signals. The entire planet warms; the greatest temperature increases are north of 50° latitude and the least in the tropics. Warming causes the poleward retreat of sea ice in both hemispheres. Precipitation and evaporation increase, and runoff is confined only to areas of heavy rainfall. The majority of Kazanian coals occurs where seasonal precipitation would support the biomass. Monsoons are limited to the Southern Hemisphere. Southern Pangea receives year-round rainfall. The restricted Zechstein, Perm (U.S.S.R.), and Permian (U.S.A) basins record times of evaporite deposition and are characterized by negative precipitation-evaporation (P–E) rates. Interior Pangea at middle to high latitudes endures torrid summers (60°C) and frigid winters (–40°C).

Two Late Jurassic simulations examine different paleoatmospheric greenhouse conditions. The geologic record favors a warm world and an elevated greenhouse effect. Sea ice is restricted to high latitudes, making landfall only in restricted areas. The trade winds bring heavy seasonal rainfall to eastern Gondwana and to the Tethys Sea margins. A strong summer monsoon occurs over southeast Asia. The distribution of coal-forming environments correlates with precipitation sufficient to maintain gymnosperm forests. Evaporites are localized to areas of negative P–E. Runoff is restricted to regions of intense precipitation. A strong positive correlation occurs between model-generated wind-driven coastal upwelling and the distribution of oil-prone marine kerogens.

The experiments on Pangean formation and disintegration show that the CCM can be used to generate paleoclimatic simulations for different plate tectonic settings and to evaluate the distribution of economic resources. These simulations predict the distribution of large-scale zonal phenomena (for example, evaporites, carbonates, coals) and more complex, environmentally sensitive deposits (for example, corals and oilprone kerogens). The CCM results are impressive considering the model's simplistic nature, coarse grid, and many parameterizations.

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