Palaeogeography, Palaeoclimatology, Palaeoecology 268 (2008) 205–221

Contents lists available at ScienceDirect

Palaeogeography, Palaeoclimatology, Palaeoecology

journal homepage: www.elsevier.com/locate/palaeo

The Late Pennsylvanian Midcontinent Sea of North America: A review

Thomas J. Algeo [a,*], Philip H. Heckel [b]

[a] Department of Geology, University of Cincinnati, Cincinnati, OH 45221-0013, USA
[b] Department of Geoscience, University of Iowa, Iowa City, IA 52242, USA

ARTICLE INFO

Article history:
Accepted 26 March 2008

Keywords:
Hydrography
Bathymetry
Pycnocline
Stratification
Anoxia
Primary productivity
Cyclothem
Black shale

ABSTRACT

The Late Pennsylvanian Midcontinent Sea (LPMS) of North America reached its greatest extent (~2.1×10^6 km^2) during glacioeustatic highstands from the Middle Pennsylvanian to the Early Permian. At these times, the sea was strongly stratified, with a subpycnoclinal layer that was anoxic and intermittently sulfidic. The development of widespread benthic anoxia in the LPMS was due to a combination of factors, including some found in most modern epicontinental seas, e.g., relatively shallow bathymetry, elevated runoff into a largely landlocked basin, a strong pycnocline, and estuarine-type circulation. However, two factors that contribute significantly to the development of benthic anoxia in such settings, i.e., a shallow marginal sill to limit deepwater renewal, and high marine primary productivity rates to stimulate benthic oxygen demand, were absent in the LPMS. Rather, a key factor controlling benthic redox conditions was lateral advection of "preconditioned" intermediate waters from Panthalassa. As in the modern eastern tropical Pacific, the oxygen-minimum zone (OMZ) may have risen to depths <100 m in the Late Pennsylvanian eastern tropical Panthalassic Ocean, allowing oxygen-depleted and intermittently denitrified waters to flood deeper basins on the southwestern margin of Laurentia. Slow transit of these waters through the ~1000-km-long, stratified Greater Permian Basin Seaway maintained the oxygen-poor status of these waters prior to upwelling out of the Anadarko and Arkoma basins onto the Midcontinent Shelf of the LPMS. Despite low levels of primary productivity and benthic oxygen demand, deepwater anoxia was maintained and even intensified into interior regions of the LPMS due to its strong pycnocline and proximal tapering of the subpycnoclinal layer. The intensity of benthic anoxia in the LPMS was a function of the strength and lateral extent of its pycnocline and, hence, of regional precipitation and continental runoff. Consequently, the LPMS highstand depositional system was highly sensitive to climate fluctuations at intermediate timescales (i.e., hundreds to tens of thousands of years). Controls on benthic redox conditions in the LPMS and similar ancient seas are not well understood owing to a paucity of appropriate modern analogs. Because existing models for anoxia in epicontinental seas do not invoke some of the key controls identified in this study, we propose a new *superestuarine circulation model* for which the LPMS may be considered the type example.

Contents

* Corresponding author. Tel.: +1 513 556 4195; fax: +1 513 556 6931.
E-mail address: Thomas.Algeo@uc.edu (T.J. Algeo).

0031-0182/$ – see front matter © 2008 Elsevier B.V. All rights reserved.
doi:10.1016/j.palaeo.2008.03.049

1. Introduction

During the Middle Pennsylvanian to Early Permian, the interior region of North America was repeatedly flooded when Gondwanan icesheets melted in the Southern Hemisphere and eustatic elevations rose (Heckel, 1994). At its maximum extent, this Late Pennsylvanian Midcontinent Sea (LPMS) covered an area of ~2.1×10^6 km^2, making it larger than modern epicontinental seas such as Hudson Bay (1.2×10^6 km^2), the Gulf of Carpentaria (0.51×10^6 km^2), and the Baltic Sea (0.42×10^6 km^2). Whereas these modern seas have either permanently or intermittently oxic bottomwaters (Algeo et al., 2008), LPMS deepwaters were anoxic for extended intervals during highstands, resulting in accumulation of laminated, organic-rich black shales (Heckel, 1977, 1980, 1991; Algeo et al., 1997; Hoffman et al., 1998; Algeo et al., 2004, in review). These "core" or offshore shales, representing the deepest water facies of Kansas-type cyclothems, are commonly thin (<1 m) but laterally extensive, in some cases being traceable from Oklahoma and Kansas northeastward to Iowa and Illinois (Heckel, 1977, 1994; Youle et al., 1994; Heckel, 1995; Watney et al., 1995). They are characterized by fine grain size, abundant phosphatic granule layers, lack of biota other than nektic and pelagic organisms, and strongly ^{34}S-depleted authigenic sulfides (Heckel, 1977; Coveney and Shaffer, 1988; Schultz and Coveney, 1992). These features are thought to reflect slow accumulation in a sediment-starved, distal offshore setting with oxygen-depleted (and possibly sulfidic) bottomwaters.

Recent studies of Kansas-type cyclothemic core shales (Algeo et al., 1997; Genger and Sethi, 1998; Hoffman et al., 1998; Algeo and Maynard, 2004; Algeo et al., 2004; Cruse and Lyons, 2004; Algeo and Maynard, in press; Algeo et al., 2008, in press) have provided new insights into environmental conditions and dynamics of the highstand LPMS, necessitating revision of earlier depositional models (Heckel, 1977, 1980, 1991; Hatch and Leventhal, 1992; Heckel, 1994). First, organic petrographic and geochemical data have shown that some core shales contain large quantities of organic matter (OM) of terrigenous origin (cf. Coveney et al., 1987); such OM is refractory in character and provides little "fuel" to stimulate benthic oxygen demand. Second, trace-metal data have shown that LPMS deepwaters were not restricted, consistent with earlier inferences of good communication between this epicontinental sea and the global ocean via a deepwater corridor. These observations have generated a conundrum: normally, a combination of limited benthic oxygen demand and unrestricted deepwater exchange would result in oxygenated bottomwaters, as in modern Hudson Bay and the Gulf of Carpentaria. The fact that the LPMS was strongly anoxic and, at times, euxinic despite these conditions suggests that its redox status was influenced by other, less common factors. Recent work has identified the "preconditioned" oxygen-poor status of deepwaters that were laterally advected to the LPMS as a key factor (Algeo et al., 2008, in press). The present contribution summarizes recent work by the authors and their collaborators aimed at reconstructing the environmental conditions and dynamics of the LPMS and, in particular, the factors that contributed to widespread benthic anoxia.

The data presented herein and the conclusions drawn therefrom are based largely on studies of core shales of Missourian Stage (lower

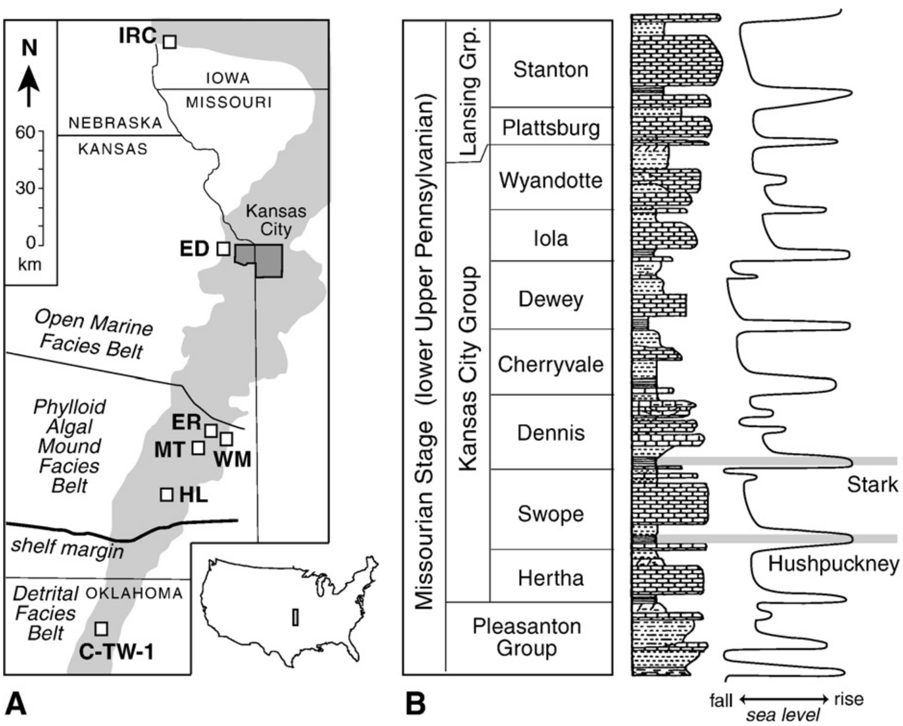


Fig. 1. (A) Location map of drillcores containing Kansas-type cyclothemic core shales cited in this study: ED = Kansas Geological Survey (KGS) Edmonds #1A; ER = KGS Ermal #1; HL = KGS Heilman #1; MT = KGS Mitchellson #1; WM = KGS Womelsdorf #1; IRC = Iowa Geological Survey Riverton Core; C-TW-1, Tulsa County, OK. Missourian Stage outcrop belt (shaded) and facies belts from Heckel (1977). (B) Missourian Stage stratigraphy of eastern Kansas; the Hushpuckney and Stark shales were the main focus of work by Algeo et al. (1997, 2004, 2008), Hoffman et al. (1998), and Algeo and Maynard (2004). Sea-level curve modified from Heckel (1986).

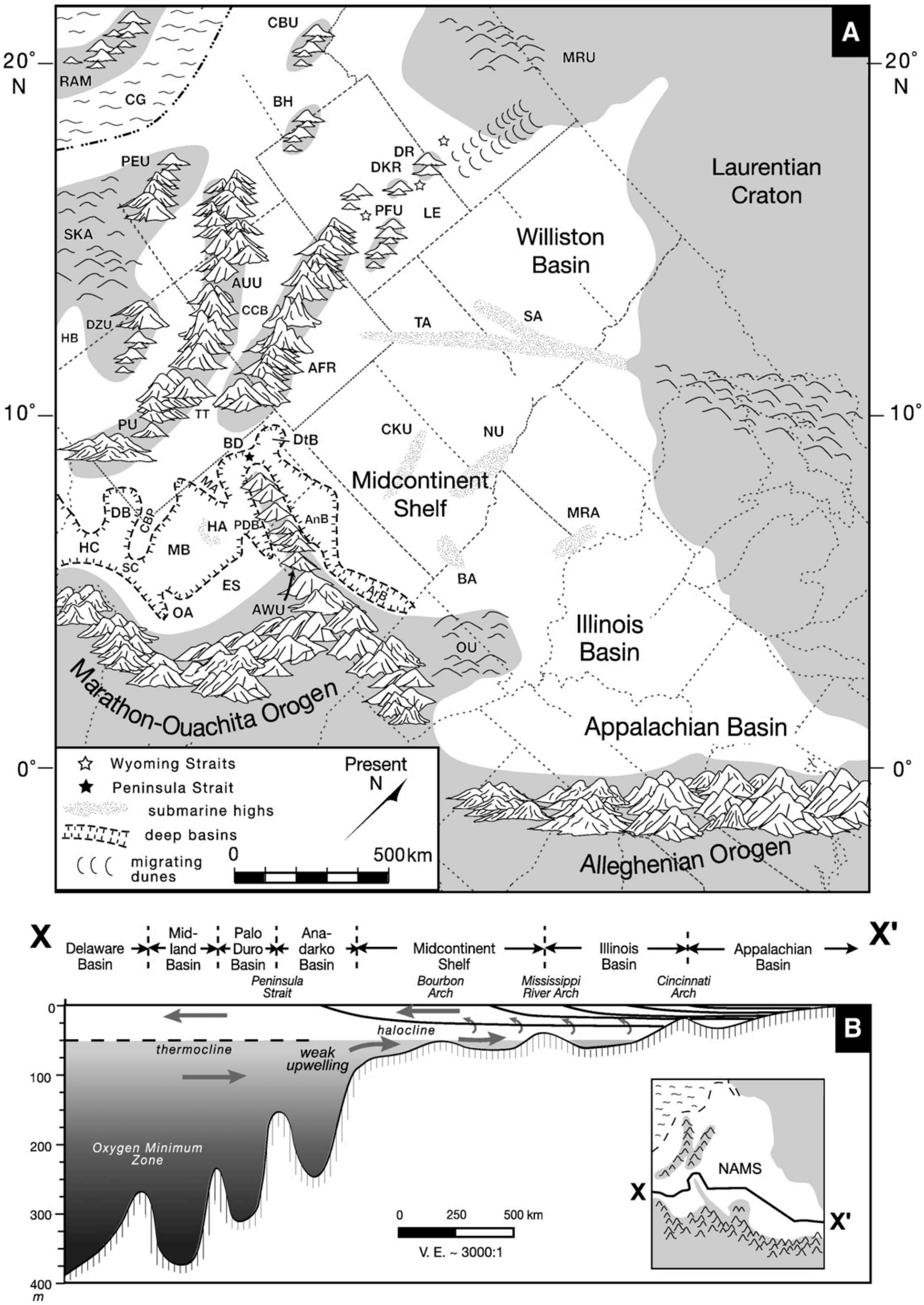


Fig. 2. (A) Paleogeography of the Late Pennsylvanian Midcontinent Sea (LPMS) of North America. Abbreviations: AnB = Anadarko Basin, ArB = Arkoma Basin, AFR = Ancestral Front Range, AUU = Ancestral Uncompahgre Uplift, AWU = Amarillo–Wichita Uplift, BA = Bourbon Arch, BD = Bravo Dome, BH = Bannock High, CBP = Central Basin Platform, CBU = Copper Basin Uplift, CCB = Central Colorado Basin, CG = Cordilleran Geosyncline, CKU = Central Kansas Uplift, DB = Delaware Basin, DtB = Dalhart Basin, DKR = Dull Knife Ridge, DR = Darton Ridge, DZU = Defiance–Zuni Uplift, ES = Eastern Shelf, HA = Horseshoe Atoll (or Scurry Reef), HB = Holbrook Basin, HC = Hovey Channel, LE = Lusk Embayment (or Alliance Basin), MA = Matador Arch, MB = Midland Basin, MRA = Mississippi River Arch, MRU = Milk River Uplift, NU = Nemaha Uplift, OA = Ozona Arch, OU = Ozark Uplift, PDB = Palo Duro Basin, PEU = Piute–Emery Uplift, PFU = Pathfinder Uplift, PU = Pedernal Uplift, RAM = Remnant Antler Mountains, SA = Siouxian Arch, SC = Sheffield Channel, SKA = Sedona–Kaibab Arches, TA = Transcontinental Arch, TT = Taos Trough. (B) Schematic cross-section from Hovey Channel (X) to the eastern interior region of LPMS (X′) showing inferred watermass circulation patterns, redox conditions, and seafloor bathymetry. Modified from Algeo et al. (2008).

Upper Pennsylvanian) Kansas-type cyclothems, especially the Hushpuckney and Stark shales in eastern Kansas and adjacent areas (Fig. 1; Algeo et al., 1997; Hoffman et al., 1998; Algeo and Maynard, 2004; Algeo et al., 2004; Algeo and Maynard, in press; Algeo et al., 2008, in press); further work will be required to determine the degree to which these units are typical of other Midcontinent cyclothemic core shales. Furthermore, the present study considers conditions within the LPMS only during eustatic highstands (and associated late-transgressive and early-regressive phases that are also represented by core core-shale sedimentation), hence coinciding with interglacial stages of Late Pennsylvanian glacio-eustatic cycles. It does not address conditions during eustatic lowstands nor does it comment on the recent debate regarding the extent of Missourian Stage glaciation and the attendant scale of contemporaneous eustatic fluctuations (e.g., Isbell et al., 2003; Jones and Fielding, 2004), although it is worth noting that large changes in continental ice volume are the only means of generating the sizeable fluctuations in sea-level elevation (~60–150 m) recorded by Midcontinent cyclothems (Heckel, 1977, 1986, 1991, 1994; Soreghan and Giles, 1999; Joachimski et al., 2006).

2. The Late Pennsylvanian Midcontinent Sea of North America

2.1. Geographic, tectonic, and climatic boundary conditions

The LPMS was nearly surrounded by landmasses, although these varied greatly in character (Fig. 2A). To the north and northeast, the Laurentian craton was emergent but of low relief (n.b., all directional references are to paleocoordinates). The Williston Basin, partially isolated by the submergent Trancontinental and Siouxian arches, represented the northern margin of the LPMS. To the south and

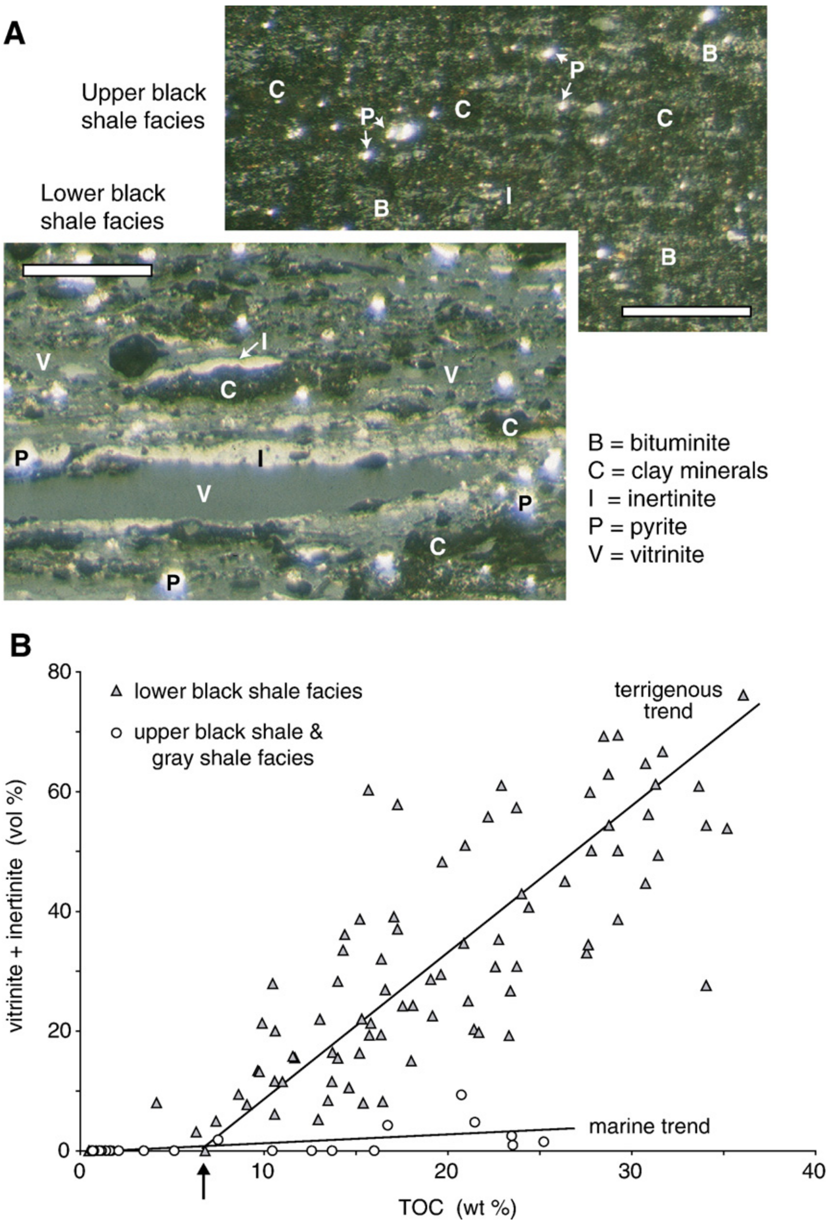


Fig. 3. Organic fraction of cyclothemic core shales. (A) Petrographic photomicrographs of lower and upper black shale facies (Hushpuckney Shale, KGS Edmonds #1A); note differences in abundance of vitrinite and inertinite between facies. Scale bars = 1 mm. (B) TOC vs. total terrigenous macerals (vitrinite + inertinite) for Hushpuckney and Stark shales (all locales). Note that samples from the lower black shale facies exhibit significant covariation, implying that TOC is dominated by terrigenous OM. The "terrigenous trend" exhibits a positive *x*-axis intercept (arrow), which represents the average amount of un(der)counted marine alginitic and bituminitic OM (~7%). Given an average TOC content of 19%, this implies that OM in the organic-rich lower black shale facies is typically about 30–40% marine and 60–70% terrigenous in origin. In contrast, OM in the comparatively organic-lean upper black shale and gray shale facies is overwhelmingly of marine origin ("marine trend").

southeast, the Appalachian–Ouachita–Marathon orogens described a long, nearly continuous mountain arc. The Appalachian portion of this chain rose to high elevations, whereas the Ouachita–Marathon section, which was part of a broad, not fully closed zone of convergence between Laurentia and the South American margin of Gondwana, was lower (Arbenz, 1989; Ye et al., 1996; Speed et al., 1997). An outlier of this orogenic complex, the ~500-km-long Amarillo–Wichita Uplift, was active during the Middle and Late Pennsylvanian and separated the LPMS from the Palo Duro, Midland, Delaware, and Val Verde basins to the southwest (i.e., the Greater Permian Basin region; Handford et al., 1981; Budnik, 1989). To the northwest, the Ancestral Rocky Mountains and associated orogens (e.g., Sierra Grande, Pedernal Uplift) rose to moderate elevations during two phases of uplift in the Desmoinesian (Middle Pennsylvanian) and Virgilian to Wolfcampian (Late Pennsylvanian to Early Permian; Kluth, 1986; Trexler et al., 1991; Burchfiel et al., 1992; Miller et al., 1992; Yang and Dorobek, 1995). West of the Ancestral Rocky Mountains, the western edge of the continent was an obliquely convergent plate margin (Wallin et al., 2000).

The LPMS extended from the humid tropical zone at paleolatitudes of ~0–5°N to the dry tropical zone at paleolatitudes of ~15–20°N (Fig. 2A; Heckel, 1977, 1980). The orogenic arc along its southern margin was within the paleo-intertropical convergence zone (ITCZ; Scotese, 1998). This resulted in a monsoonal climate, with moisture-laden air masses drawn from the proto-Tethyan embayment to the east over these orogens during the summer (Crowley et al., 1989; Parrish, 1993; Crowley et al., 1996; Montañez et al., 2007; Poulsen et al., 2007), resulting in high levels of precipitation and continental runoff into the LPMS. General atmospheric circulation (GCM) climate models suggest that the Midcontinent region had a tropical climate (10–25 °C), limited seasonal temperature range (<15 °C), and moderate annual precipitation (<73 cm yr^{-1}; Crowley et al., 1989, 1996). A persistent subtropical high-pressure system resulted in dominance of trade winds from the northeast (Parrish and Peterson, 1988). Relative to the everwet conditions of the Middle Pennsylvanian, the Late Pennsylvanian climate of the Midcontinent became drier and more seasonal as North America drifted northward out of the humid equatorial zone (Schutter and Heckel, 1985; Cecil, 1990; DiMichele and Phillips, 1996; Tabor and Montañez, 2002; Cecil et al., 2003a; Tabor, 2007). However, climate conditions were relatively more humid during Missourian Stage interglacials, especially the late-transgressive to early-highstand phases, than during the intervening glacial stages (Cecil, 1990; Heckel, 1995; Soreghan et al., 2002; Algeo et al., 2004). Major changes in regional climate and vegetation accompanied large-scale collapse of continental icesheets in the Early Permian (mid-Sakmarian; Montañez et al., 2007; Poulsen et al., 2007).

2.2. Hydrography

2.2.1. Freshwater influx

Moderately humid conditions (during Missourian Stage interglacials) resulted from a monsoonal climate, location of the Appalachian–Ouachita orogen within the paleo-ITCZ, and, perhaps, intensified evapotranspiration associated with widespread highstand coastal coal swamps (Cecil, 1990; Archer and Greb, 1995; DiMichele and Phillips, 1996; Soreghan et al., 2002; Cecil et al., 2003a; Algeo et al., 2004). The LPMS received an estimated discharge volume of ~800–1500 km^3 yr^{-1}, equivalent to that of modern Hudson Bay (~975 km^3 yr^{-1}) and the Baltic Sea (~485 km^3 yr^{-1}) when normalized to drainage area (Algeo et al., 2008). Approximately half of this amount was associated with a major foreland basin drainage system entering the eastern end of the LPMS (Gibling et al., 1992; Archer and Greb, 1995), a discharge somewhat lower than that of the modern Ganges–Brahmaputra (970 km^3 yr^{-1}; Cecil et al., 2003b). Strong continental runoff into the LPMS produced a reduced-salinity surfacewater layer extending from its eastern interior region to at least the Midcontinent Shelf and, at times, possibly into the Greater Permian Basin region (Fig. 2B). However, the subpycnoclinal watermass of the LPMS was close to normal-marine salinity owing to good exchange with open-ocean waters (Section 2.2.5). The average residence time of water in the LPMS was ~70–130 yr, intermediate between that for the modern Baltic Sea (45 yr) and Hudson Bay (130 yr; Algeo et al., 2008).

Independent evidence of strong continental runoff into the LPMS includes (1) large concentrations of terrigenous OM (locally 80–100% of total organic macerals) in some cyclothemic core shales (Fig. 3), reflecting strong export from coastal coal swamps (cf. Greb et al., 2003; Algeo et al., 2004); (2) strong lateral variation in benthic redox proxies and other sediment parameters (Fig. 4), indicative of regional gradients in pycnocline strength and other watermass properties (Fig. 2B; cf. Coveney et al., 1991; Algeo et al., 1997; Hoffman et al., 1998); and (3) uniformity of sediment ε_{Nd} values across the Appalachian and Ouachita basins, suggesting extensive reworking and transport of sediment westward

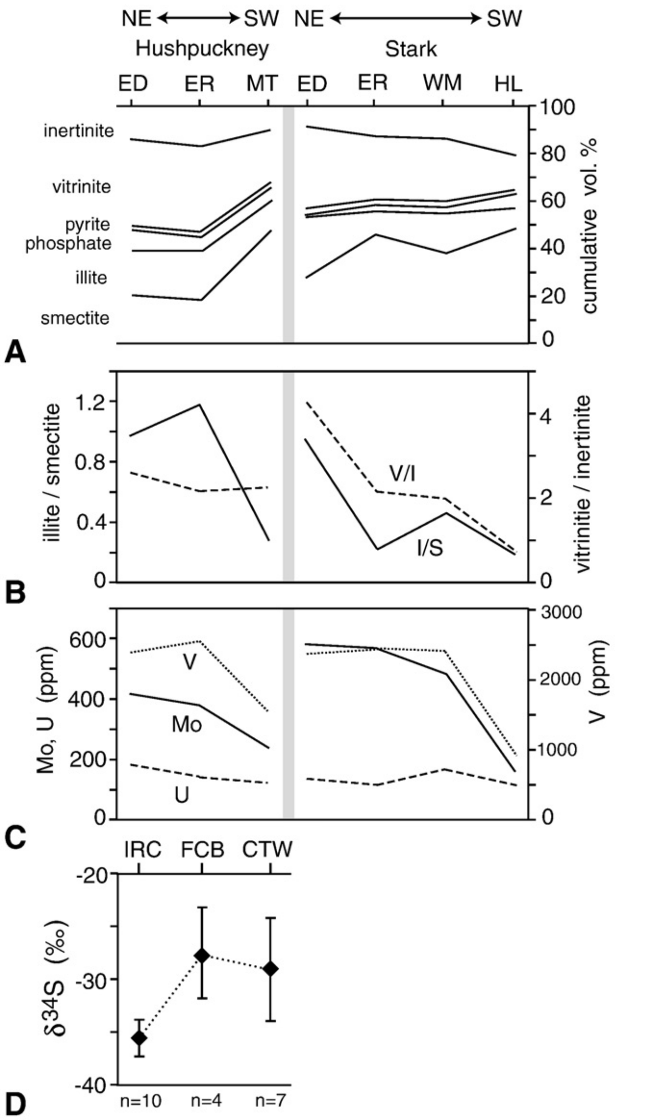
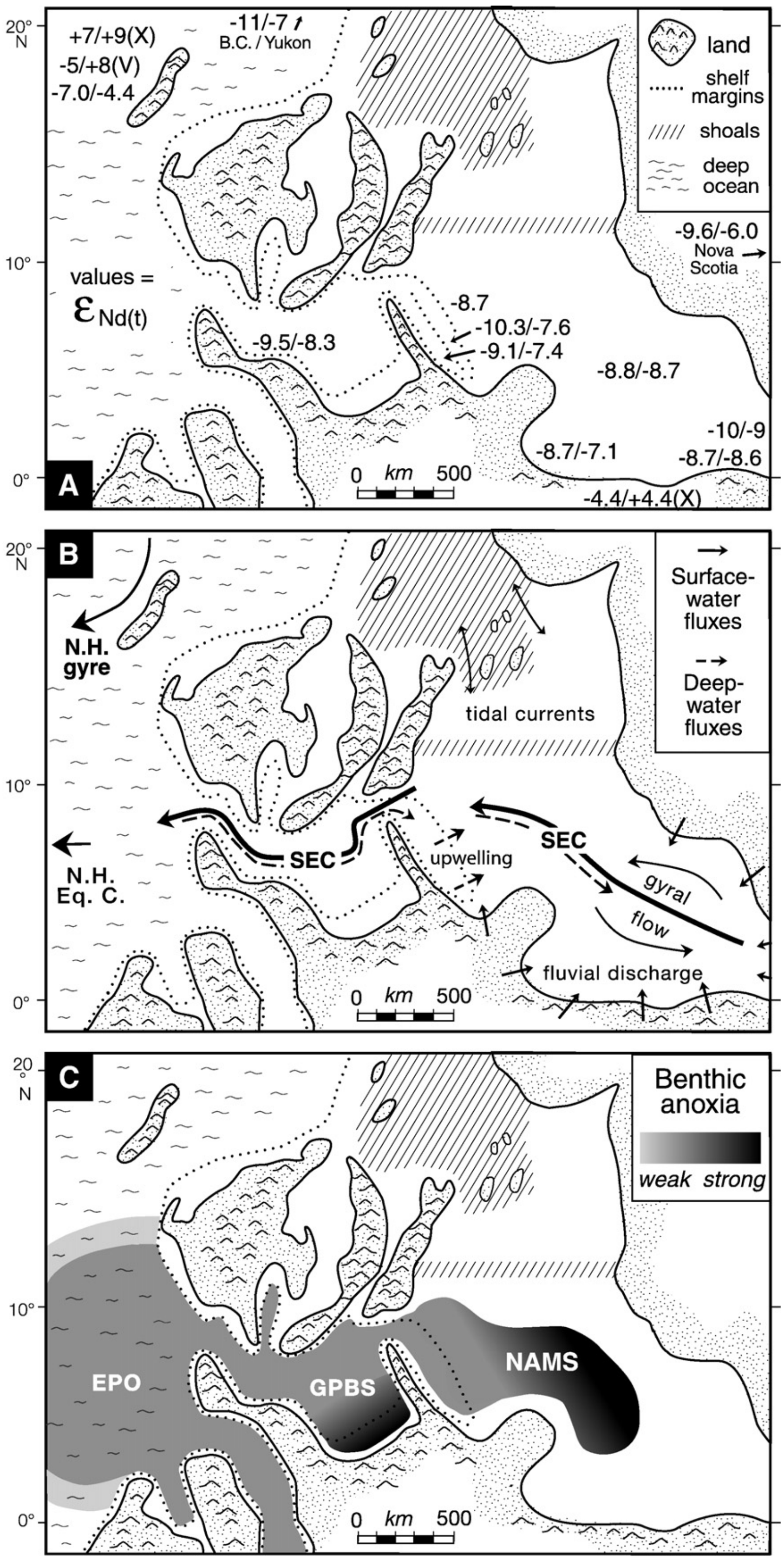


Fig. 4. Geographic gradients for sedimentary parameters of the Hushpuckney and Stark shales in the Midcontinent region: (A) major component abundances; (B) illite/smectite (I/S) and vitrinite/inertinite (V/I) ratios; (C) Mo, U, and V concentrations (mean of lower black shale facies); (D) pyrite S-isotope data (mean and standard deviation range; number of analyses shown below). Note that proximal (northeastern) sites are enriched in illite and vitrinite (relative to smectite and inertinite), trace metals, and pyrite $\delta^{34}S$ compared to distal (southwestern) sites. These patterns imply a directional flux of both detrital material and surface waters from the northeast, with deposition of coarser clays and organic macerals closer to the terrigenous source and enrichment of trace metals and pyrite $\delta^{34}S$ in sediments beneath a stronger pycnocline. See Fig. 1 for site abbreviations. A and B are from Algeo et al. (1997), C is unpublished data, and D is from Coveney and Shaffer (1988) and Cruse and Lyons (2004).

across the Midcontinent region (Fig. 5A; Gleason et al., 1994; Patchett et al., 1999). The paleogeography of the LPMS, with freshwater input focused at its eastern end and communication with the global ocean through the Greater Permian Basin Seaway at its western end, favored the development of estuarine-style circulation and strong spatial gradients in environmental and sedimentary parameters (Fig. 5B and C; Algeo et al., 1997; Hoffman et al., 1998; Cruse and Lyons, 2004; Algeo et al., 2008).



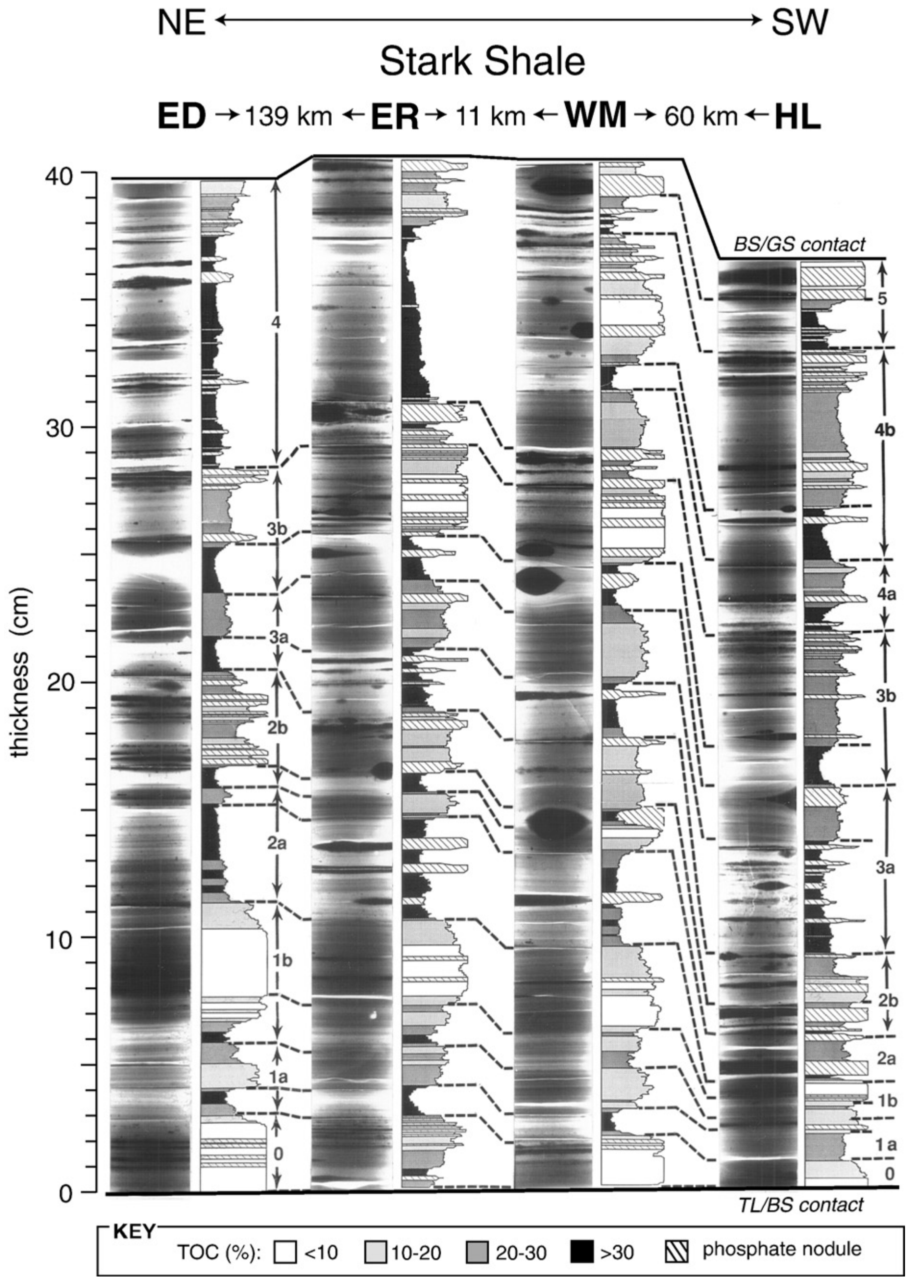


Fig. 6. Correlated black shale facies of the Stark Shale from four sites in eastern Kansas; see Fig. 1 for site abbreviations. For each site, the left column is a positive X-radiograph print in which lighter and darker hues represent lower and higher densities, respectively; low-density layers are generally TOC-rich, and discrete, oval-shaped or granular high-density features are phosphate nodules. The relief on the right column of each site is a continuous trace of gray-scale density measurements from the X-radiograph on the left; shaded fill patterns indicate layer composition (see key at bottom). The base and top of each set of columns represent the contact of the black shale (BS) facies with the underlying transgressive limestone (TL) and the overlying gray shale (GS) facies, respectively. Dashed lines show internal correlations, and numbered arrows indicate dm-scale cycles (see text for discussion).

2.2.2. Pycnocline

Water depths within the LPMS were relatively shallow, even during highstands, as shown by several lines of evidence. First, the presence of red algae throughout the thick regressive limestones of Kansas-type cyclothems indicate water depths within the photic zone (probably < 100 m; Fornós and Ahr, 1997; Basso, 1998). Green algae are most abundant at water depths < 30 m in modern marine environments (Multer, 1977), and green dasycladacean and phylloid algae are generally concentrated in the upper part of regressive limestones of Kansas-type cyclothems, suggesting shallowing to this depth range

Fig. 5. Spatial patterns within LPMS: (A) Neodymium isotopes ($\varepsilon_{Nd(t)}$); (B) watermass circulation patterns; (C) paleoredox conditions. In A, no label = sedimentary rocks, V = volcanic rocks, and X = crystalline rocks. Sediment $\varepsilon_{Nd(t)}$ values are uniformly −10 to −7 across LPMS (t = 300–320 Ma; Gleason et al., 1994, 1995; Schatzel and Stewart, 2000; t = 360, Murphy, 2000; Dickinson et al., 2003; Eriksson et al., 2004) as well as within the western Canadian miogeocline (Boghossian et al., 1996; Garzione et al., 1997). In contrast, $\varepsilon_{Nd(t)}$ values are mostly between −4.4 and +4.4 for Pennsylvanian crystalline rocks from the Southern Appalachians (Samson et al., 1995; Coler et al., 1997) and mostly >+3 for coeval volcanics and volcaniclastic sediments along the western continental margin (Smith and Lambert, 1995; Blein et al., 1996; Childe and Thompson, 1997; Patchett and Gehrels, 1998; Simard et al., 2003; Schwartz et al., 2005). These patterns imply that Late Paleozoic clastic sediments were sourced mainly in the Appalachian–Caledonian orogens and subsequently dispersed westward across the North American craton (Gleason et al., 1994; Patchett et al., 1999). In C, redox patterns are based on spatial variation in trace-metal concentrations and other proxies (e.g., Fig. 4C and D). Abbreviations: EPO = Eastern Panthalassic Ocean, GPBS = Greater Permian Basin Seaway, LPMS = North American Midcontinent Sea, N.H. Eq. C. = Northern Hemisphere Equatorial Current, and SEC = superestuarine circulation.

during the latter stages of regression of the LPMS. Second, a number of major cyclothems are capped by paleosols in northern Kansas, Nebraska, and Iowa, indicating subaerial exposure during lowstands (Heckel, 1977, 1980; Schutter and Heckel, 1985; Joeckel, 1994, 1999). The amplitude of Late Pennsylvanian glacio-eustatic fluctuations was ~60–150 m (median estimate ~100 m) based on analyses of contemporaneous icesheet mass (Crowley and Baum, 1991; Soreghan and Giles, 1999) and conodont $\delta^{18}O$ variation (Joachimski et al., 2006). These amplitudes are equivalent to maximum highstand water depths in areas of the Midcontinent Shelf that were exposed during lowstands. Water depths were generally greater (perhaps several hundred meters deep) in the Anadarko and Arkoma foredeep basins in Oklahoma but varied considerably through time in response to episodes of basin subsidence and fill (Arbenz, 1989; Elmore et al., 1990).

Earlier workers proposed that water-column stratification was maintained in eastern (interior) areas of the LPMS by a halocline and in western (exterior) areas by a thermocline (Heckel, 1991; Hatch and Leventhal, 1992; Hoffman et al., 1998). While basically correct, this view is oversimplified in that vertical stratification was controlled by a combination of salinity and temperature variation everywhere within the LPMS, with salinity being more important in interior areas and temperature gradually becoming more important distally. However, modern cratonic seas demonstrate that, given a sufficiently positive water balance, a halocline can be supported across wide areas, even at relatively shallow water depths (~15–30 m for Hudson Bay and ~40–80 m for the Baltic Sea; Kullenberg, 1981; Pett and Roff, 1982; Algeo et al., 2008). Highstand water depths in the LPMS of ~100 m were sufficient for the development of a permanent halocline that may have extended from interior areas of the Appalachian Basin at least as far west as Kansas. As in all estuarine systems, the depth of the halocline would have shallowed distally. If the shallow- and deepwater layers had temperatures of 25 and 15 °C and salinities of 35–40 and 45–50‰, respectively (n.b., Paleozoic open-ocean seawater salinities were higher than today; Knauth, 1998; Hay et al., 2001, 2006), then watermass densities (σ_t) were ~23–27 and 34–38, respectively (Algeo et al., in press; n.b., σ_t is watermass density in units of kg m^{-3} minus 1000; all values calculated from the international equation of state of seawater, UNESCO, 1981, as given in Chester, 1990). This calculation yields a deep-to-shallow density differential ($\Delta\sigma_t$) of ~11±4, of which ~70% is attributable to salinity variation, consistent with a halocline, and ~30% to temperature variation, consistent with a thermocline. This $\Delta\sigma_t$ value is significantly larger than that for the modern Hudson Bay and Baltic Sea (~4) although less than that found in some modern estuaries (>20; Kennish, 2001). Maintenance of a strong pycnocline in the LPMS was facilitated by (1) strong continental runoff into a largely landlocked basin (Fig. 5B; Section 2.2.1), (2) microtidal conditions (Section 2.2.3), and (3) unrestricted influx of cold deepwaters of normal-marine salinity below a distal thermocline (Figs. 2B and 5B; Section 2.2.5).

The existence of a strong pycnocline (vertical density gradient) is supported by several lines of sedimentologic evidence. First, the black shale facies of core shales exhibits sharp lower and upper contacts, suggesting rapid redox changes on the LPMS seafloor (Heckel, 1977; Algeo et al., 2004; Cruse and Lyons, 2004; Algeo et al., in review). Such changes are thought to have occurred when the pycnocline, which separated an oxygenated surfacewater layer from an oxygen-depleted deepwater layer, migrated first toward and then away from the craton during each eustatic transgressive–regressive cycle (Heckel, 1977, 1991). There is no precise modern analog to this process, but glacioeustatically-driven vertical fluctuations of pycnoclines within large estuaries such as Chesapeake Bay provide some guidance (Bratton et al., 2003; Hobbs, 2004). It is also possible that the sharpness of core-shale contacts was influenced by temporal variations in pycnocline strength, and that onset of black shale deposition reflects a sudden intensification of benthic anoxia due to pycnocline strengthening associated with enhanced precipitation and runoff during deglaciations (Algeo et al., 2004). Second, centimeter-scale (~2 to 10 cm thick) compositional cycles within the black shale facies of core shales are correlatable over distances of at least a few hundred kilometers (Fig. 6; Algeo et al., 1997), a degree of bedding continuity known only from the modern Black Sea (Lyons, 1991, his Fig. 4). Lateral continuity of fine sediment layers at this scale implies uniformity of environmental conditions and synchroneity of changes in watermass properties over wide areas, a pattern that is likely to develop only beneath a persistent, strong pycnocline. Third, sulfidic conditions existed at least intermittently in subpycnoclinal waters of the highstand LPMS, as shown by exceptionally high levels of trace-element enrichment and the possible presence of syngenetic pyrite in cyclothemic core shales (Fig. 4C; Algeo et al., 1997; Hoffman et al., 1998; Algeo and Maynard, 2004; Algeo et al., 2004). A strong pycnocline limited vertical mixing, an important factor in maintaining oxygen-deficient conditions over wide areas of the LPMS despite low levels of primary productivity and benthic oxygen demand (Section 2.3).

2.2.3. Tides

Tidal currents can be a potent force in maintaining a well-mixed watermass. This is evident in the modern Gulf of Carpentaria, where strong tidal currents through Torres Strait inhibit formation of an extended halocline in the northern half of this gulf despite elevated river discharge from the New Guinea highlands (Somers and Long, 1994; Cecil et al., 2003b; Algeo et al., 2008). In Hudson Bay, a macrotidal range on the western basin margin contributes to high benthic dissolved oxygen levels and accumulation of coarse, TOC-lean sediments (Pelletier et al., 1968, 1986; Roff and Legendre, 1986; Biksham and d'Anglejan, 1989; Algeo et al., 2008). More completely landlocked seas such as the Baltic are subject to lesser tidal influence. Although the LPMS may have had a deepwater connection to the global ocean, the serpentine path of this corridor through the Greater Permian Basin region effectively prevented penetration of open-ocean tides into its interior (Fig. 2A). Internally, the LPMS was probably microtidal (Kennett, 1982; Wells et al., 2005a,b), with a tidal range too small to promote vertical mixing of the water column and, hence, having little if any influence on pycnocline strength (n.b., reports of large tidal ranges from the LPMS are limited to coastal embayments, in which tides would have been amplified by coastline morphology, e.g., Adkins and Eriksson, 1998; Kvale and Mastalerz, 1998).

2.2.4. Gyral circulation

Circulation patterns in the LPMS can be inferred based on hydrologic and climatic factors and on regional variation in sediment properties. Hydrologic considerations suggest that continental runoff into the LPMS was concentrated to some extent at its eastern (interior) end (Fig. 5B), resulting in a westward net flow of surface waters (Heckel, 1977, 1980). In addition to net transport, most inland seas also exhibit gyral circulation; for example, the modern Hudson Bay and Baltic Sea circulate cyclonically (counterclockwise in the Northern Hemisphere; Alhonen, 1966; Prinsenberg, 1986). Cyclonic circulation may have existed in the LPMS also (e.g., Heckel, 1980, his Fig. 5). This inference is supported by lateral compositional trends for the Hushpuckney and Stark core shales. Illite increases relative to smectite (Fig. 4A) and vitrinite increases relative to inertinite (Fig. 4B) in a northeastward direction, suggesting a larger flux of coarser, water-borne clays and terrigenous OM from that direction relative to the finer, possibly wind-blown fraction that dominated in more distal areas to the southwest (Algeo et al., 1997). Redox proxies indicate stronger benthic anoxia to the northeast (Fig. 4C and D), consistent with greater freshwater influx and a stronger pycnocline in that direction (Fig. 5C; Algeo et al., 1997; cf. Coveney et al., 1991; Hoffman et al., 1998; Cruse and Lyons, 2004; Algeo et al., 2008). Cyclonic circulation may have developed in the LPMS owing to the influence of the Easterly trade winds and/or the Coriolis effect, both of which would have pushed river waters entering the LPMS from the north in a westward direction, parallel to its northern coastline (Fig. 5B).

2.2.5. Deepwater exchange

Connections of the LPMS to the global ocean were limited by its largely landlocked character. The Ancestral Rockies and associated uplifts formed a barrier extending from southern Wyoming to southern New Mexico, restricting watermass exchange to areas north and south of these orogens (Fig. 2A). The northern passage through central Wyoming (the "Wyoming Straits") was wide (200 to 300 km) but shallow (probably < 10 m) and was studded with shoals and islands and frequently choked with sands from migrating dune fields (Mankiewicz and Steidtmann, 1979; Maughan, 1984; Blakey et al., 1988; Kerr and Dott, 1988). Its existence was finally terminated by late Pennsylvanian–early Permian uplift of the Wyoming Arch (Maughan, 1993; Luebking et al., 2001). Late Pennsylvanian facies of the Williston and Alliance basins are of restricted-marine character (Peterson and MacCary, 1987), suggesting poor communication with the global ocean to the west. An additional barrier to watermass exchange via the northern passage was the Transcontinental Arch, a shallow submarine ridge that extended from Minnesota to Colorado (Garfield et al., 1988).

The more important connection to the global ocean was located in the area south of the Ancestral Rockies, where a serpentine, ~1000-km-long corridor traversed the Greater Permian Basin region (Fig. 2A). This corridor commenced in the vicinity of the Hovey Channel, between the southern end of the Pedernal Uplift and the north-western margin of the Ouachita–Marathon Front, at a paleolatitude of ~5–8°N, and then passed successively through the Midland, Palo Duro, and Dalhart basins, where it entered the southwestern end of the LPMS. Bathymetric variation in the Greater Permian Basin Region was considerable, with deep basins (i.e., hundreds of meters) separated by shallower sills formed over structural highs (Fig. 2B; Handford and Dutton, 1980; Handford et al., 1981; Walker et al., 1995; Hill, 1999). If watermass exchange through this corridor was unimpeded during both highstands and lowstands (as argued below), then sills in this region must have had a minimum highstand depth of ~100 m (based on glacioeustatic amplitudes; Section 2.2.2). The narrowest section of the corridor was probably at the "Peninsula Strait," a ~30–40-km-wide passage located between the western end of the Amarillo–Wichita Uplift and the eastern margin of the Bravo Dome during the Late Pennsylvanian and Early Permian (Handford and Fredericks, 1980; Handford et al., 1981; Budnik, 1989). Flow through this strait was probably on the order of $0.1–0.2 \times 10^6$ m^3 s^{-1}, which is comparable to estimates of flow in the straits regulating watermass exchange for the modern Hudson Bay and Baltic Sea (Algeo et al., 2008).

Several lines of evidence favor strong watermass exchange through the Greater Permian Basin Seaway, consistent with the inference of a continuous deepwater corridor between the eastern tropical Panthalassic Ocean and the LPMS (Fig. 2B). First, a low-diversity, euryhaline fauna is characteristic of hyposaline watermasses (<20‰ salinity), as in many modern estuaries and the Baltic Sea (Friedrich, 1969; Hällfors et al., 1981; Raffaelli, 1996). Faunal assemblages in the LPMS provide no evidence for freshwater or brackish conditions during either eustatic highstands or lowstands. Although Kansas-type cyclothemic core shales, which contain a pelagic-nektic biota consisting of conodonts, ammonoids, fish, and orbiculoid brachiopods, are faunally impoverished, the lack of a diverse benthic biota was due to widespread bottomwater anoxia rather than to reduced salinities (Heckel and Baesemann, 1975; Heckel, 1977, 1980; Boardman et al., 1984). Pennsylvanian marine units in the Appalachian Basin to the east contain a diverse "normal-marine" fauna consisting of molluscs, brachiopods, echinoderms, bryozoans, corals, sponges, and conodonts (Bennington, 1996; Fahrer, 1996; Stamm and Wardlaw, 2003). This same fauna is also associated with gray shales and carbonates deposited during regressive phases of the LPMS (Malinky and Heckel, 1998), when sea-level elevations were lower by up to ~100 m relative to highstands (Crowley and Baum, 1991; Soreghan and Giles, 1999; Joachimski et al., 2006). The faunal evidence is inconsistent with hyposaline conditions in the LPMS and, hence, with a silled perimeter and restricted watermass exchange.

A second argument against restricted watermass exchange in the LPMS is based on trace-metal concentration patterns. Redox-sensitive trace metals such as Mo, U, V, and Zn tend to become depleted in the subpycnoclinal watermass of restricted silled anoxic basins owing to the removal of the sediment without adequate resupply, as in the modern Black Sea and Framvaren Fjord (Dyrssen et al., 1984; Emerson and Huested, 1991). This signal can be transferred to the sediment as a decrease in trace-metal concentrations per unit organic matter (n.b., although still elevated relative to background detrital concentrations; Algeo and Lyons, 2006). Because trace-metal concentrations can also vary as a function of changes in benthic redox status, documentation of a decrease in trace-metal fluxes to the sediment is insufficient to demonstrate aqueous trace-metal depletion as a consequence of deepwater restriction. In this regard, patterns of trace-metal covariation are potentially informative: in environments subject to deepwater restriction, the various trace metals present in the deep watermass are commonly depleted at different rates (Dyrssen et al., 1984; Emerson and Huested, 1991), resulting in divergent concentration trends in the sediment (Fig. 7A; Algeo, 2004; Algeo and Maynard, in press). On the other hand, where concentration variations are due largely or entirely to fluctuations in benthic redox status (via its influence on aqueous trace-metal reactivity and uptake by sediments), the concentrations of various trace metals in the sediment are likely to track each other closely, as is the case for Kansas-type cyclothemic core shales (Fig. 7B; Algeo et al., 2004). Strong covariation of trace-metal concentrations in these units suggests (1) control primarily by redox variation and (2) minimal variation in the concentrations of dissolved trace metals in the host watermass, and the latter inference is consistent with good exchange of LPMS waters with the global ocean.

A third argument against restricted watermass exchange in the LPMS is based on the degree of environmental "dynamism" exhibited by the depositional system. Silled anoxic basins tend to be relatively stagnant systems, in which sediment trace-metal concentrations can remain elevated (relative to background detrital levels) over long stratigraphic intervals owing to the existence of a fairly stable pycnocline and continuous deepwater anoxia. Sediment trace-metal concentrations in modern silled basins such as the Black Sea and Cariaco Basin have remained elevated since the establishment of benthic anoxia during the post-Wisconsinan sea-level rise (Hirst, 1974; Lyons et al., 2003; Algeo and Lyons, 2006), and ancient silled basins in which deepwater anoxia persisted for extended intervals (sometimes upward of a million years) can exhibit continuous trace-metal enrichment over considerable stratigraphic intervals (e.g., the ~50-m-thick Upper Devonian Ohio Shale of the Central Appalachian Basin; Fig. 7A). In contrast, unrestricted anoxic marine systems such as those developed in modern continent-margin upwelling zones tend to exhibit strong environmental variation at relatively short timescales (e.g., decades to millennia), recorded as high-frequency compositional variation of the sediment (e.g., Emeis et al., 1991; Hendy and Pedersen, 2006). Kansas-type cyclothemic core shales exhibit pronounced variation in TOC and trace-metal concentrations at a centimeter scale (Figs. 6 and 7B), patterns that are distinctly more similar to those observed in modern unrestricted continent-margin settings than in restricted silled basins. Such high-frequency compositional variation implies changes at relatively short (~10^3 yr) timescales in the physico-chemical properties of the LPMS watermass, especially with regard to redox conditions, which is consistent with a "dynamic" (i.e., unrestricted) depositional system.

2.2.6. Upwelling

Upwelling can be an important factor in the formation of organic-rich deposits on modern continental margins (Hay, 1995). Earlier studies of the LPMS proposed upwelling of nutrient-rich deepwaters onto the southern margin of the Midcontinent Shelf (Heckel, 1977, 1991). However, various climatic and sedimentologic considerations support a mechanism and intensity for this process different from those of modern continent-margin upwelling systems. First, such upwelling systems are the product of offshore, wind-driven Ekman transport of surface waters on arid tropical shelves. Climatic aridity is a key factor as strong fluvial discharge would produce a density-stratified inner-shelf watermass that could suppress upwelling. Deepwater flow in the LPMS was unrelated to Ekman transport; rather, it was a result of entrainment of the subpycnoclinal watermass by a large-scale estuarine-type circulation system in a largely landlocked epicontinental sea (Figs. 2B and 5B). Second, continent-margin upwelling systems are characterized by a patchy, eddy-controlled distribution of upwelling and downwelling cells (Summerhayes et al., 1995), resulting in laterally discontinuous deposits of organic-rich sediment (Bailey, 1991; Smith, 1992; Glenn et al., 1994). This process is unlikely to generate the widely correlatable centimeter-thick layers of cyclothemic core shales observed in interior areas of the Midcontinent Shelf (Fig. 6; Algeo et al., 1997), although a certain patchiness of black and gray shale facies exists along the southern shelf margin near the Kansas–Oklahoma border (Wanless and Wright, 1978; Heckel, unpubl. data). Third, estimates of primary productivity and organic C burial fluxes for the LPMS are much lower than for modern upwelling systems (Section 2.3), implying a more limited flux of nutrient-rich waters onto the Midcontinent Shelf margin. These considerations suggest that upwelling-driven marine productivity played a secondary role at most in the development of benthic anoxia in the LPMS.

2.3. Primary productivity

Although difficult to estimate, primary productivity appears to have been rather low in the LPMS. Organic C sinking fluxes can be estimated on the basis of (1) organic C burial fluxes, which are calculated from average TOC values and sedimentation rates, and (2) inferred preservation efficiencies (Canfield, 1994; Hay, 1995). TOC values in Missourian Stage cyclothemic core shales are generally in the range of 5–30 wt.% (max. 40 wt.%), but much of this OM is of terrigenous origin and, hence, unrelated to marine productivity. The proportion of marine algal matter in any given sample is highly variable but averages about 30–40% in the highly organic-rich lower black shale facies and ~90% in the only moderately organic-rich upper black shale facies (Fig. 3), or about 50(±10)% overall.

There is ample evidence that sedimentation rates were exceptionally low during deposition of Kansas-type cyclothemic core shales: (1) an abundance of authigenic phosphate, with some core shales containing >50 phosphatic granule layers (Fig. 6; Heckel, 1977; Kidder, 1985; Heckel, 1991; Kidder et al., 1996; Algeo et al., 2004), each possibly requiring hundreds or thousands of years to form (cf. Föllmi, 1996; Filippelli, 1997); (2) high concentrations of higher land plant debris (i.e., vitrinite and inertinite), which must have been rafted or blown hundreds of kilometers across the LPMS (Figs. 3 and 4A; Algeo et al., 1997; Hoffman et al., 1998; Algeo et al., 2004); and (3) strong

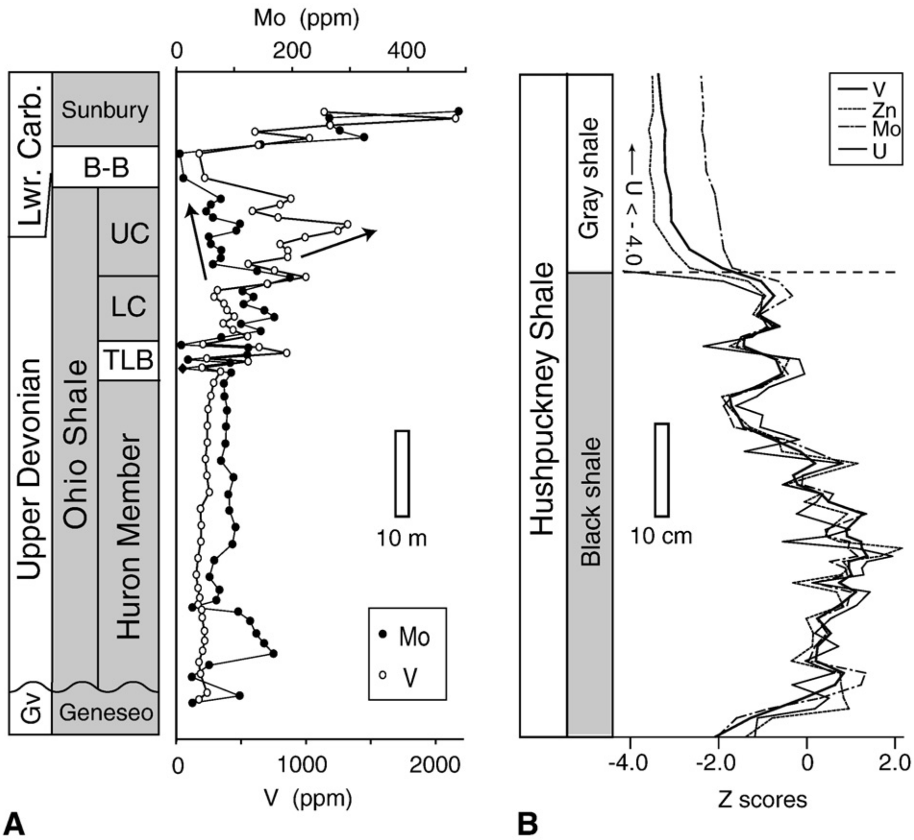


Fig. 7. Chemostratigraphic trends in trace-metal concentrations of (A) Upper Devonian–Lower Mississippian black shales of the Central Appalachian Basin, and (B) the Upper Pennsylvanian Hushpuckney Shale in the KGS Edmonds #1A drillcore. Abbreviations: B–B = Bedford–Berea, Gv = Givetian (late Middle Devonian), LC = Lower Cleveland, TLB = Three Lick Bed, UC = Upper Cleveland. Gray shading indicates organic-rich black shales. In A, increasing watermass restriction during deposition of the Cleveland Shale resulted in an increase in V (tracking variation in both TOC and DOP; see Algeo, 2004) but a decrease in Mo, the latter due to drawdown of the aqueous Mo reservoir through sedimentation (n.b., divergent trends shown by arrows). Note also the nearly uniform trace-metal concentrations of the 40-m-thick Huron Member of the Ohio Shale, implying relatively static watermass conditions during ~4 Myr of sedimentation. In B, trace-metal concentrations are given as "Z scores" (i.e., normalized to a mean of 0 and a standard deviation of 1) for the purpose of comparing multiple elements with widely varying abundances. Note the concurrent shifts in concentrations of the four trace metals shown at a centimeter scale. Strong covariation among these metals implies control of accumulation mainly by benthic redox status.

enrichment of redox-sensitive trace metals such as Mo, U, V, and Zn (Fig. 4C; Coveney et al., 1987; Hatch and Leventhal, 1992; Genger and Sethi, 1998; Algeo and Maynard, 2004; Cruse and Lyons, 2004; Algeo and Maynard, in press). These observations are consistent with a sediment-starved, distal offshore environment in which terrigenous OM and hydrogenously sourced trace metals became highly concentrated owing to limited siliciclastic influx. A predominantly hydrothermal source for trace metals, as proposed by Coveney and Glascock (1989), is unlikely in view of their wide geographic area of enrichment and their strong covariation with TOC at a fine stratigraphic scale (Figs. 5C and 7B). Sedimentation rates can be tentatively estimated on the basis of cyclostratigraphic analysis, if small-scale compositional cycles in the core shales represent 20-kyr orbital precession cycles (Fig. 8). This assumption yields an average (compacted) sedimentation rate estimate of ~0.002–0.008 mm y^{-1} (Heckel, 2004; Algeo et al., 2004, 2008). This estimate is slightly lower than Holocene sedimentation rates on the Black Sea abyssal plain, which are ~0.1 mm y^{-1} at present-day porosities of ~90% and equivalent to ~0.01 mm y^{-1} on a fully dewatered basis (Shimkus and Trimonis, 1974; Karl and Knauer, 1991; Arthur et al., 1994).

Combining a sedimentation rate estimate of ~0.002–0.008 mm y^{-1} with a measured bulk density of 2200 kg m^{-3} and average TOC values of 5–30% yields calculated organic C burial fluxes for Kansas-type cyclothemic core shales in the range of 0.2–5.3 g C m^{-2} y^{-1} (median: 1.1 g C m^{-2} y^{-1}), of which only ~50% or 0.1–2.6 g C m^{-2} y^{-1} (median: 0.6 g C m^{-2} y^{-1}) represents the marine organic C flux. Because preservation efficiencies are relatively high (~10–30%) in sulfidic environments, even at low sedimentation rates (Canfield, 1994), the organic C burial fluxes calculated above imply organic C sinking fluxes of <26 g C m^{-2} y^{-1}. All of these values are lower by an order of magnitude or more than those reported for modern continent-margin upwelling systems (Calvert and Price, 1983; Marlow et al., 2000) and epicontinental seas such as the Baltic Sea and Hudson Bay (Algeo et al., 2008). Independent evidence of low organic carbon fluxes to the sediment is provided by the strongly ^{34}S-depleted compositions of pyrite in cyclothemic core shales relative to contemporaneous seawater sulfate (Fig. 4D; $\Delta^{34}S = 40 \pm 5‰$; Coveney and Shaffer, 1988; Schultz and Coveney, 1992). Strongly ^{34}S-depleted pyrite is indicative of low bacterial sulfate reduction rates, suggesting a limited flux of labile, easily degradable marine OM to the sediment (Anderson and Pratt, 1995). However, it should be noted that fractionation associated with bacterial sulfate reduction depends on a series of metabolic steps that can yield variable fractionations, and that disproportionation of intermediate products can influence resulting sulfur-isotopic compositions (Canfield, 2001; Brunner and Bernasconi, 2005; Canfield et al., 2006). Primary productivity per se cannot be estimated, but, in view of the strongly reducing bottomwater conditions and shallow pycnocline of the LPMS, the fraction of surface-water productivity represented by the organic carbon sinking flux is likely to have been relatively high and, hence, primary productivity is likely to have been quite low.

2.4. Benthic redox conditions

During highstands of the LPMS, deepwaters were anoxic and at least intermittently sulfidic over an area of $>0.3 \times 10^6$ km^2 in the Midcontinent region (Fig. 5C). Anoxia persisted for the duration of deposition of the black shale facies of Kansas-type cyclothemic core shales, estimated at ~100 kyr for the Hushpuckney and Stark shales (Fig. 8). Strong benthic anoxia existed in the LPMS despite good deepwater exchange (Section 2.2.5) and weak primary productivity

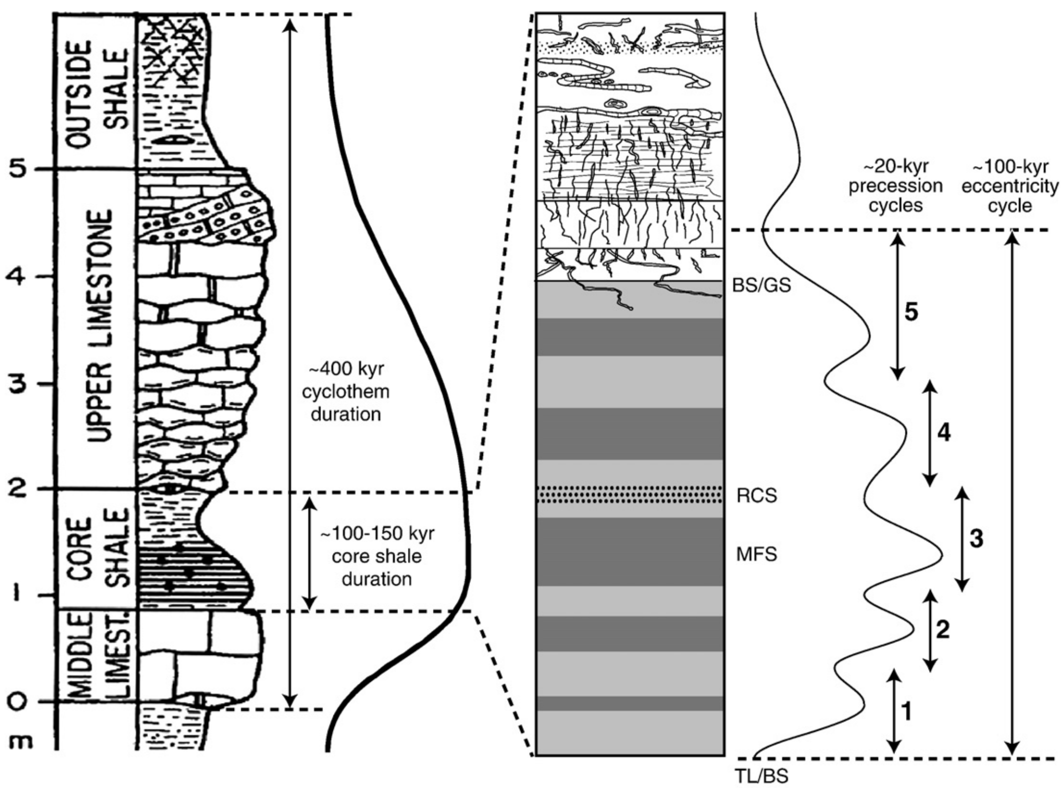


Fig. 8. Cyclostratigraphy of Kansas-type cyclothemic core shales and inferred internal distribution of time. Major cyclothems are thought to represent ~400-kyr long-eccentricity cycles (Heckel, 1986, 2004). The Hushpuckney and Stark shales are both characterized by dm-scale internal cyclicity, with about five well-defined small-scale cycles present in the black shale facies of each (e.g., Fig. 6). If these cycles represent ~20-kyr orbital precession cycles, then the highstand conditions associated with core core-shale deposition persisted for ~100 kyr at a stretch, or a bit longer if the time content of the gray shale facies is included. Abbreviations: BS = black shale facies; GS = gray shale facies; TL = transgressive limestone; MFS = maximum flooding surface; RCS = regressive condensation surface, formed through winnowing caused by enhanced vertical mixing that resulted from the onset of eustatic fall and concurrent pycnocline weakening (Algeo et al., 2004). Cyclothem model at left from Heckel (1977); core core-shale stratigraphy at right modified from Algeo et al. (2004).

(Section 2.3). These conditions yield a conundrum, since the development of benthic anoxia commonly depends on a combination of high organic carbon sinking fluxes and reduced rates of deepwater renewal (Hay, 1995; Algeo et al., 2008). Given the absence of these key factors in the LPMS, it is necessary to consider the role of other, less common controls on watermass redox conditions.

One important control may have been the redox status of the deep watermass that was advected into the LPMS through the Greater Permian Basin Seaway. In the modern eastern tropical Pacific Ocean, the oxygen oxygen-minimum zone (OMZ) rises to shallow water depths (<100 m, versus 500–1000 m in the open ocean) due to a combination of enhanced, upwelling-driven productivity associated with Easterly trade winds, a long watermass residence time occasioned by the failure of the subtropical anticyclones to penetrate this area, and the generally low dissolved oxygen content of Pacific seawater (Fig. 9; Roden, 1964; Wyrtki, 1967; Hay, 1995; Ganeshram and Pedersen, 1998). During the early stages of Quaternary deglaciations, strong upwelling resulted in elevated primary productivity in surface waters and oxygen demand in intermediate waters of the eastern tropical Pacific, leading to enhanced rates of denitrification and development of a ^{15}N-enriched bioavailable nitrogen pool (Pride et al., 1999; Kienast et al., 2002; Thunell and Kepple, 2004). Similar conditions prevailed in the eastern tropical Panthalassic Ocean during the Permo-Pennsylvanian (Fig. 5C), including shallowing of the OMZ during the early stages of Late Pennsylvanian deglaciations and development of ^{15}N-enriched nitrogen pools that were advected into the LPMS, as described below (Algeo et al., in review).

Lateral advection of oxygen-deficient eastern tropical Panthalassic intermediate waters to the LPMS was possible owing to the deepwater corridor that traversed the Greater Permian Basin Seaway (Figs. 2 and 5). Hovey Channel, the gateway to the Greater Permian Basin Seaway, was located at ~5–10°N paleolatitude, the area in which the eastern tropical Panthalassic OMZ shallowed to the maximum degree (Algeo et al., in review). This situation allowed oxygen-deficient (and intermittently denitrified) waters to transit the ~1000-km-long Greater Permian Basin Seaway in ~100 days (based on a deep flow rate of 0.1–0.2 m s^{-1}; Algeo et al., 2008). The oxygen-deficient character of this watermass was maintained during transit through a combination of water-column stratification and moderate benthic oxygen demand within the Greater Permian Basin Seaway, as evidenced by the organic-rich character of deepwater shales and siltstones in the Delaware and Midland basins (Adams et al., 1951; Jackson, 1964; Cys and Gibson, 1988; Landis et al., 1992) and by thin phosphatic black shales (similar to Kansas-type cyclothemic core shales) on the Midland Basin's Eastern Shelf (Jackson, 1964; Boardman and Heckel, 1989). After entering the LPMS through the Peninsula Strait (Fig. 2A), low-intensity upwelling out of the deeper Anadarko and Arkoma basins introduced this watermass onto the shallow Midcontinent Shelf (Figs. 2B and 5B). Owing to the “preconditioned” oxygen-deficient character of the upwelled watermass, a weak organic carbon sinking flux (Section 2.3) was sufficient for the development of widespread benthic anoxia in the LPMS.

An additional factor contributing to benthic anoxia in more proximal portions of the LPMS was the interplay of seafloor bathymetry and pycnocline depth, which controls the volume of the deep watermass and thereby influences its redox status. When the volume of the sub-pycnoclinal layer is large, the deep watermass is buffered to a greater degree against respiratory oxygen demand; conversely, when its volume is small, the deep watermass is more prone to depletion of dissolved oxygen. These relationships are well illustrated by modern epicontinental seas. In Hudson Bay, pycnocline depth (15–30 m) is shallow compared to average water depth (120 m), contributing to only limited depletion of dissolved oxygen in deepwaters. In contrast, the Baltic Sea

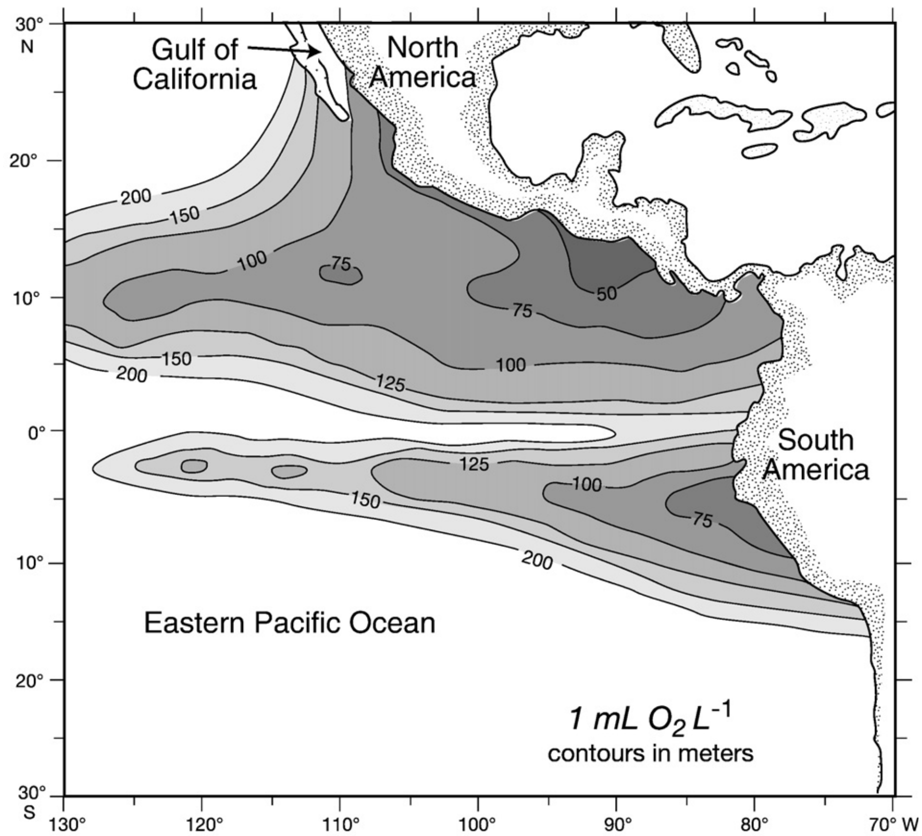


Fig. 9. Depth to the top of oxygen oxygen-minimum zone (OMZ) in the modern eastern tropical Pacific, as defined by the 1 mL L^{-1} dissolved oxygen isocon; contours in meters. Note the extreme shallowing of the OMZ at latitudes of ~3–12°S and 5–22°N. The contoured surface is coextant with a strong thermocline. Data from Levitus and Boyer (1994).

has a deeper pycnocline (40–80 m) and lesser average water depth (55 m), making its deepwaters more prone to oxygen depletion through respiratory demand. In simplified form, the paleobathymetry of the LPMS can be viewed as an inclined ramp with a ~100- to 150-m-deep outer margin that shallowed gradually into its interior regions (Fig. 2B). Given a pycnocline depth of ~15–30 m that shallowed distally (Section 2.2.2), it is evident that the subpycnoclinal layer of the LPMS may have been >100 m deep in distal areas but thinned to a feather edge proximally. Combined with a stronger pycnocline in interior regions of the LPMS (due to proximity to the main sources of continental runoff), the reduced volume of the deep watermass in proximal areas may have facilitated benthic oxygen depletion.

From this analysis, it appears that a unique set of boundary conditions and environmental dynamics resulted in the development of benthic anoxia across extensive portions of the LPMS during glacioeustatic highstands. The key boundary conditions were (1) extended orogenic highlands within the paleo-ITCZ, (2) a largely landlocked setting, (3) shallow seafloor bathymetry, (4) an elongate, serpentine deepwater connection to the Panthalassic Ocean, and (5) a gateway at 5–10°N paleolatitude, within a region of extreme shallowing of the paleo-OMZ. Important environmental features resulting from these boundary conditions included (1) a strong regional halocline, reducing vertical mixing, (2) the limited volume of the subpycnoclinal watermass, facilitating benthic oxygen depletion despite limited respiratory demand, and (3) the "preconditioned" oxygen-deficient character of deep watermasses that were laterally advected to the LPMS.

This combination of boundary conditions and environmental responses does not exist in any large modern epicontinental sea, e.g., Hudson Bay, the Baltic Sea, and the Gulf of Carpentaria (Algeo et al., 2008). Although the Gulf of Carpentaria exhibits geographic, climatic, and tectonic boundary conditions similar to the LPMS (Edgar et al., 2003), differences in hydrography are important: the Gulf of Carpentaria is less landlocked and more subject to open-ocean tides, which result in greater vertical mixing rates, and its restricted interior (southern) end receives comparatively little runoff from the northern Australian margin, preventing formation of a permanent halocline. Hudson Bay is the closest analog to the LPMS in terms of hydrographic characteristics, having good communication with the global ocean via an elongated and deep but narrow corridor (Hudson Strait; Drinkwater, 1986) and similar freshwater and deepwater fluxes and watermass residence times (Algeo et al., 2008). However, in Hudson Bay these conditions result in only limited depletion of deepwater oxygen and OM accumulation, the critical difference being the redox status of deepwaters advected through the Hudson Strait, which are well oxygenated. The Baltic Sea, the only large modern epicontinental sea with anoxic deepwaters, shares certain features with the LPMS, including high levels of freshwater runoff, a strong pycnocline, and microtidality. However, benthic anoxia in the Baltic is critically dependent on a shallow marginal sill (~10–20 m deep at the Danish Straits) and moderately high levels of primary productivity, stimulated in recent decades by anthropogenic eutrophication. In contrast, the LPMS lacked a shallow sill (Section 2.2.5) and exhibited low levels of primary productivity (Section 2.3). The only large modern marine basin to have a gateway located in a region subject to OMZ shallowing, similar to the LPMS, is the Gulf of California, which is not an epicontinental sea but an early-stage ocean basin (Fig. 9). Advection of oxygen-deficient deepwaters contributes to the development of benthic anoxia locally within this basin, although its intensity is limited by climatic and hydrologic factors, i.e., low runoff from the relatively arid surrounding landmasses, exacerbated by anthropogenic reduction in the discharge of the Colorado River at the northern (interior) end of the Gulf of California basin. Thus, the development of widespread benthic anoxia in the LPMS appears to reflect a unique combination of paleogeographic and paleoceanographic controls.

Although there are many models of anoxic marine systems (Demaison and Moore, 1980; Wignall, 1994; Arthur and Sageman, 1994; Hay, 1995), relatively few have been developed with specific reference to epicontinental seas, reflecting the dearth of appropriate modern analogs. Recent anoxic marine environments are mostly located in oceanic or continent-margin settings and classified either as *silled basins*, such as the Black Sea, Cariaco Trench, and Santa Barbara Basin, or as continent-margin *upwelling zones*, such as the Peru Shelf, the Namibian Shelf, and the Arabian Sea. Because none of these models appear to fit the LPMS, we propose that it be the "type example" of a *superestuarine circulation system*, characterized by (1) large-scale estuarine-type circulation and halocline formation, (2) unrestricted deepwater exchange, and (3) lateral advection of "preconditioned" oxygen-deficient deepwaters, facilitating the development of benthic anoxia (Fig. 10A). By analogy with true estuaries (Kennish, 2001), "estuarine circulation" denotes strong surfacewater flow away from coastal point-sources of freshwater influx, with entrainment of the shoreward-flowing deepwater mass along a well-defined pycnocline (cf. "quasi-estuarine circulation" model of Witzke, 1987). Superestuarine circulation systems are distinct from silled basins in lacking topographic restriction of deepwater exchange, and

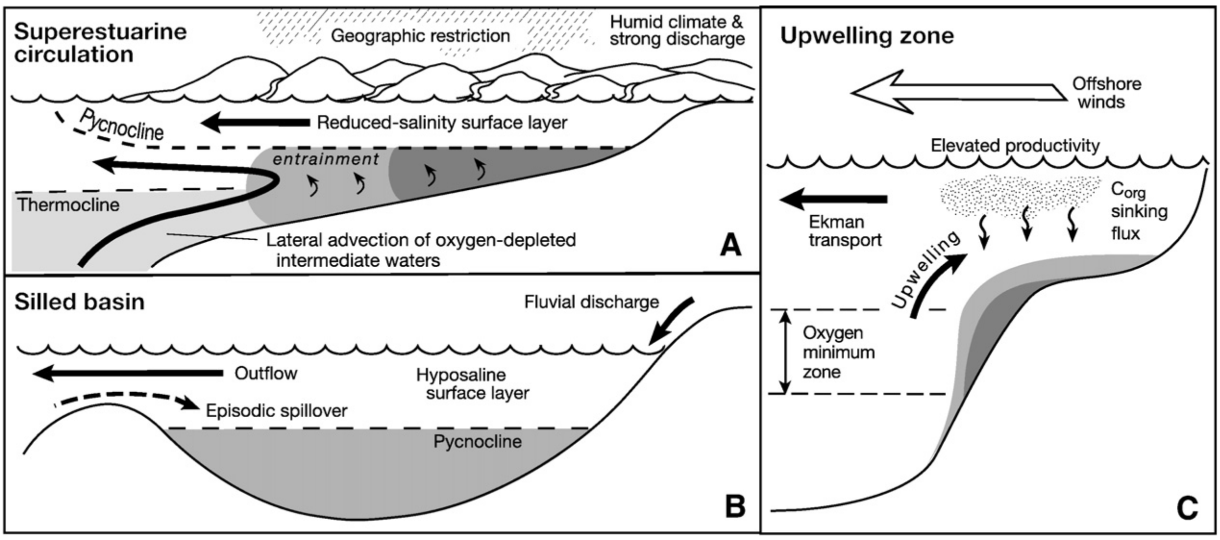


Fig. 10. Models of marine anoxia, contrasting boundary conditions and environmental dynamics of (A) epicontinental-marine *superestuarine circulation systems* versus (B) *silled basins* and (C) continent-margin *upwelling zones*. From Algeo et al. (2008).

from continent-margin upwelling zones with regard to geographic setting and in exhibiting relatively low levels of primary productivity (Fig. 10B and C). Although estuarine-type circulation and oxygen-deficient deepwater sources are elements of some anoxic marine models (Demaison and Moore, 1980; Wignall, 1994; Arthur and Sageman, 1994; Hay, 1995), these features are not generally regarded as the key elements of any existing model for anoxia in epicontinental seas.

With regard to paleoenvironmental analysis, superestuarine marine systems potentially can be distinguished from silled basins and upwelling systems on the basis of characteristic patterns of compositional variation in sediments. *Spatial variation* in sediment composition reflects geographic variation in environmental conditions. Silled basins generally exhibit "bulls-eye" patterns centered on the deepest and most anoxic portion of the water body, as in the modern Black Sea or Baltic Sea (Fig. 8C; Shimkus and Trimonis, 1974; Glasby et al., 1997). Continent-margin upwelling systems tend to exhibit patchy concentrations along a linear trend, reflecting local zones of upwelling on a continental margin (Calvert and Price, 1983; Reimers and Suess, 1983). In contrast, superestuarine marine systems are characterized by fairly uniform spatial gradients in environmental conditions (e.g., pycnocline strength, benthic redox status) and, hence, in sediment composition over broad areas (Fig. 4; Algeo et al., 1997; Hoffman et al., 1998). *Stratigraphic variation* in sediment composition reflects the degree of environmental dynamism of a given depositional system. Silled basins tend to exhibit relatively stable watermass conditions owing to the basinwide extent of their pycnoclines and the short-term immutability of key boundary conditions (e.g., sill depth) (Scranton et al., 1987; Murray, 1991). Continent-margin upwelling systems are generally highly dynamic at short timescales (i.e., months to years) owing to the lack of a stable pycnocline and the movement of large, chemically variable watermasses along continental margins (Bailey and Chapman, 1991; Emeis et al., 1991); such systems can also exhibit systematic variation in environmental conditions at longer timescales (Ganeshram and Pedersen, 1998). Superestuarine marine systems are intermediate in terms of their environmental dynamism. The existence of a strong pycnocline over large areas dampens environmental variability somewhat relative to continent-margin upwelling systems, but the "open-ended" (laterally unconfined) character of pycnoclines in such systems makes them inherently more variable than pycnoclines in silled basins. Dependence of the strength and areal extent of pycnoclines on climatic variables such as precipitation and fluvial runoff makes superestuarine marine systems particularly susceptible to environmental fluctuations at intermediate timescales (i.e., hundreds to tens of thousands of years; Figs. 7 and 8). Careful analysis of spatio-temporal patterns of compositional variation in the sediments of ancient marine systems should facilitate accurate discrimination among these contrasting environmental models.

3. Conclusions

Although modern epicontinental seas can provide insights regarding controls on benthic redox conditions, none represents a close analog to the Late Pennsylvanian Midcontinent Sea (LPMS) of North America. The LPMS was unique in developing anoxic conditions over a large area (>0.3×10^6 km^2) despite lacking a marginal sill to restrict deepwater exchange and having levels of primary productivity too low to impose a significant demand on benthic oxygen. The key boundary conditions promoting widespread benthic anoxia in the LPMS were (1) a humid paleoclimate, (2) a largely landlocked setting, (3) shallow seafloor bathymetry, (4) an elongate, serpentine deepwater connection to the global ocean, and (5) location of the entrance of this deepwater corridor in a region of extreme shallowing of the oxygen-minimum zone in the eastern tropical Panthalassic Ocean. Important features of the LPMS environment deriving from these boundary conditions include (1) a strong regional halocline, reducing vertical mixing of the water column, (2) limited volume of the subpycnoclinal watermass, facilitating dissolved O_2 depletion despite low benthic oxygen demand, and (3) the "preconditioned" oxygen-deficient character of laterally advected deepwaters. In contrast to the modern Baltic Sea, which exhibits a basin-centered pattern of benthic anoxia in accord with the silled basin model, the LPMS exhibits a strong lateral gradient in benthic redox conditions with the development of the most intense anoxia in shallower interior regions of the sea. This pattern reflects the importance of large-scale estuarine circulation in a laterally unconfined epicontinental sea, representing a type of anoxic marine system herein designated the *superestuarine circulation model*. Because benthic redox conditions in the LPMS were dependent on the strength and lateral extent of its pycnocline and, hence, on regional precipitation and continental runoff, the system was highly sensitive to climate fluctuations at intermediate timescales (i.e., hundreds to tens of thousands of years).

Acknowledgments

We thank W. Lynn Watney and the Kansas Geological Survey for access to the study cores, Tim Phillips for drafting services, Tom Robl and Sue Rimmer for geochemical data from the Kentucky Geological Survey's D-series drillcores, and Ray Coveney, Tim Lyons, and an anonymous reviewer for constructive reviews of the manuscript. This project was supported by grants to TJA from the National Science Foundation (EAR-0310072, EAR-0618003, and EAR-0745574) and the University of Cincinnati Research Council.

References

Adams, J.E., Frenzel, H.N., Johnson, D.P., Rhodes, M.L., 1951. Starved Pennsylvanian Midland Basin (Texas). Am. Assoc. Petrol. Geol. Bull. 35, 2600–2607.

Adkins, R.M., Eriksson, K.A., 1998. Rhythmic sedimentation in a Mid-Pennsylvanian delta-front succession, Magoffin Member (Four Corners Formation, Breathitt Group), eastern Kentucky; a near-complete record of daily, semi-monthly, and monthly tidal periodicities. In: Alexander, C.R. (Ed.), Tidalites: Processes and Products. SEPM-SSG Spec. Publ., vol. 61, pp. 85–94.

Algeo, T.J., 2004. Can marine anoxic events draw down the trace-element inventory of seawater? Geology 32, 1057–1060.

Algeo, T.J., Maynard, J.B., 2004. Trace element behavior and redox facies in core shales of Upper Pennsylvanian Kansas-type cyclothems. Chem. Geol. 206, 289–318.

Algeo, T.J., Lyons, T.W., 2006. Mo-total organic carbon covariation in modern anoxic marine environments: implications for analysis of paleoredox and paleohydrographic conditions. Paleoceanography 21, PA1016. doi:10.1029/2004PA001112.

Algeo, T.J., Maynard, J.B., in press. Trace-metal covariation as a guide to watermass conditions in ancient anoxic marine environments. Geosphere.

Algeo, T.J., Hoffman, D.L., Maynard, J.B., Joachimski, M.M., Hower, J.C., Watney, W.L., 1997. Environmental reconstruction of anoxic marine systems: core black shales of Upper Pennsylvanian Midcontinent cyclothems. In: Algeo, T.J., Maynard, J.B. (Eds.), Cyclic Sedimentation of Appalachian Devonian and Midcontinent Pennsylvanian Black Shales: Analysis of Ancient Anoxic Marine Systems—A Combined Core and Field Workshop. Joint Meeting of Eastern Section AAPG and The Society for Organic Petrography (TSOP), Lexington, Kentucky, Sept. 27–28, 1997, pp. 103–147.

Algeo, T.J., Schwark, L., Hower, J.C., 2004. High-resolution geochemistry and sequence stratigraphy of the Hushpuckney Shale (Swope Formation, eastern Kansas): implications for climato-environmental dynamics of the Late Pennsylvanian Midcontinent Sea. Chem. Geol. 206, 259–288.

Algeo, T.J., Heckel, P.H., Maynard, J.B., Blakey, R., Rowe, H., 2008. Modern and ancient epicratonic seas and the superestuarine circulation model of marine anoxia. In: Holmden, C., Pratt, B.R. (Eds.), Dynamics of Epeiric Seas: Sedimentological, Paleontological and Geochemical Perspectives. Geological Association of Canada Special Publication, vol. 48, pp. 7–38.

Algeo, T.J., Rowe, H., Hower, J.C., Schwark, L., Hermann, A., Heckel, P.H., in press. Changes in ocean denitrification during Late Carboniferous glacial-interglacial cycles. Nature Geoscience.

Alhonen, P., 1966. Baltic Sea. In: Fairbridge, R.W. (Ed.), The Encyclopedia of Oceanography. Reinhold, New York, pp. 87–91.

Anderson, T.F., Pratt, L.M., 1995. Isotopic evidence for the origin of organic sulfur and elemental sulfur in marine sediments. In: Vairavamurthy, M.A., Schoonen, M.A.A. (Eds.), Geochemical Transformations of Sedimentary Sulfur. Am. Chem. Soc. Symp. Ser., vol. 612, pp. 378–396. Washington, D.C.

Arbenz, J.K., 1989. The Ouachita System. In: Bally, A.W., Palmer, A.R. (Eds.), The Geology of North America—An Overview. Geol. Soc. Am., Geology of North America, v. A, pp. 371–396.

Archer, A.W., Greb, S.F., 1995. An Amazon-scale drainage system in the Early Pennsylvanian of central North America. J. Geol. 103, 611–627.

Arthur, M.A., Sageman, B.B., 1994. Marine black shales: depositional mechanisms and environments of ancient deposits. Annu. Rev. Earth Planet. Sci. 22, 499–551.

Arthur, M.A., Dean, W.E., Neff, E.D., Hay, B.J., King, J., Jones, G., 1994. Varve-calibrated records of carbonate and organic carbon accumulation over the last 2000 years in the Black Sea. Glob. Biogeochem. Cycles 8, 195–217.

Bailey, G.W., 1991. Organic carbon flux and development of oxygen deficiency on the modern Benguela continental shelf south of 22°S: spatial and temporal variability. In: Tyson, R.V., Pearson, T.H. (Eds.), Modern and Ancient Continental Shelf Anoxia. Geol. Soc. London Spec. Publ., vol. 58, pp. 171–183.

Bailey, G.W., Chapman, P., 1991. Short-term variability during an anchor station study in the southern Benguela upwelling system: chemical and physical oceanography. Prog. Oceanogr. 28, 9–37.

Basso, D., 1998. Deep rhodolith distribution in the Pontian Islands, Italy: a model for a paleoecology of a temperate sea. Palaeogeogr. Palaeoclimatol. Palaeoecol. 137, 173–187.

Bennington, J.B., 1996. Stratigraphic and biofacies patterns in the Middle Pennsylvanian Magoffin marine unit in the Appalachian Basin, U.S.A. Int. J. Coal Geol. 31, 169–193.

Biksham, G., d'Anglejan, B., 1989. Rate of sedimentation and geochemistry of south-eastern Hudson Bay, Canada. In: Hadley, R.F., Ongley, E.D. (Eds.), Sediment and the Environment. Internat. Assoc. Hydrol. Sci. Publ., vol. 184, pp. 27–36.

Blakey, R.C., Peterson, F., Kocurek, G., 1988. Synthesis of late Paleozoic and Mesozoic eolian deposits of the Western Interior of the United States. Sediment. Geol. 56, 3–125.

Blein, O., Lapierre, H., Schweicker, R.A., Monié, P., Maluski, H., Pêcher, A., 1996. Remnants of the northern Sierra Nevada Paleozoic island arc in western Nevada? J. Geol. 104, 485–492.

Boardman II, D.R., Heckel, P.H., 1989. Glacial-eustatic sea-level curve for early Late Pennsylvanian sequence in north-central Texas and biostratigraphic correlation with curve for midcontinent North America. Geology 17, 802–805.

Boardman II, D.R., Mapes, R.H., Yancey, T.E., Malinky, J.M., 1984. A new model for the depth-related allogenic community succession within North American Pennsylvanian cyclothems and implications on the black shale problem. In: Hyne, N.J. (Ed.), Limestones of the Mid-Continent. Tulsa Geological Society Spec. Publ., vol. 2, pp. 141–182.

Boghossian, N.D., Patchett, P.J., Ross, G.M., Gehrels, G.E., 1996. Nd isotopes and the source of sediments in the miogeocline of the Canadian Cordillera. J. Geol. 104, 259–277.

Bratton, J.F., Colman, S.M., Thieler, E.R., Seal II, R.R., 2003. Birth of the modern Chesapeake Bay estuary between 7.4 and 8.2 ka and implications for global sea-level rise. Geo-Mar. Lett. 22, 188–197.

Brunner, B., Bernasconi, S.M., 2005. A revised isotope fractionation model for dissimilatory sulfate reduction in sulfate reducing bacteria. Geochim. Cosmochim. Acta. 69, 4759–4771.

Budnik, R.T., 1989. Tectonic Structures of the Palo Duro Basin, Texas Panhandle. . Report of Investigations, vol. 187. Texas Bureau of Economic Geology. 43 pp.

Burchfiel, B.C., Cowan, D.S., Davis, G.A., 1992. Tectonic overview of the Cordilleran orogen in the western United States. In: Burchfiel, B.C., Lipman, P.W., Zoback, M.L. (Eds.), The Cordilleran Orogen: Conterminous. U.S. Geol. Soc. Am., Geology of North America, vol. G-3, pp. 407–479.

Calvert, S.E., Price, N.B., 1983. Geochemistry of Namibian Shelf sediments. In: Suess, E., Thiede, J. (Eds.), Coastal Upwelling: Its Sediment Record, pt. A: Responses of the Sedimentary Regime to Present Coastal Upwelling. Plenum, New York, pp. 337–375.

Canfield, D.E., 1994. Factors influencing organic carbon preservation in marine sediments. Chem. Geol. 114, 315–329.

Canfield, D.E., 2001. Biogeochemistry of sulfur isotopes. In: Valley, J.W., Cole, D.R. (Eds.), Stable Isotope Geochemistry, Reviews in Mineralogy. Mineral. Soc. Am., vol. 43, pp. 607–636.

Canfield, D.E., Olesen, C.A., Cox, R.P., 2006. Temperature and its control of isotope fractionation by a sulfate-reducing bacterium. Geochim. Cosmochim. Acta 70, 548–561.

Cecil, C.B., 1990. Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks. Geology 18, 533–536.

Cecil, C.B., Dulong, F.T., Edgar, N.T., Stamm, R.G., Wardlaw, B.R., West, R.R., 2003a. Climate controls on the stratigraphy of a Middle Pennsylvanian cyclothem in North America. In: Cecil, C.B., Edgar, N.T. (Eds.), Climate Controls on Stratigraphy. SEPM-Soc. Sedim. Geol. Spec. Publ., vol. 77, pp. 151–180.

Cecil, C.B., Dulong, F.T., Harris, R.A., Cobb, J.C., Gluskoter, H.G., Nugroho, H., 2003b. Observations on climate and sediment discharge in selected tropical rivers, Indonesia. In: Cecil, C.B., Edgar, N.T. (Eds.), Climate Controls on Stratigraphy. SEPM-Soc. Sedim. Geol. Spec. Publ., vol. 77, pp. 29–50.

Chester, R., 1990. Marine Geochemistry. Unwin Hyman, London. 698 pp.

Childe, F.C., Thompson, J.F.H., 1997. Geological setting, U–Pb geochronology, and radiogenic isotopic characteristics of the Permo-Triassic Kutcho Assemblage, north-central British Columbia. Can. J. Earth Sci. 34, 1310–1324.

Coler, D.G., Samson, S.D., Speer, J.A., 1997. Nd and Sr isotopic constraints on the source of Alleghenian granites in the Raleigh metamorphic belt and Eastern slate belt, southern Appalachians, U.S.A. Chem. Geol. 134, 257–275.

Coveney Jr., R.M., Shaffer, N.R., 1988. Sulfur-isotope variations in Pennsylvanian shales of the midwestern United States. Geology 16, 18–21.

Coveney Jr., R.M., Glascock, M.D., 1989. A review of the origins of metal-rich Pennsylvanian black shales, central USA, with an inferred role of basinal brines. Appl. Geochem. 4, 347–367.

Coveney Jr., R.M., Leventhal, J.S., Glascock, M.D., Hatch, J.R., 1987. Origins of metals and organic matter in the Mecca Quarry Shale Member and stratigraphically equivalent beds across the Midwest. Econ. Geol. 82, 915–933.

Coveney Jr., R.M., Watney, W.L., Maples, C.G., 1991. Contrasting depositional models for Pennsylvanian black shale discerned from molybdenum abundances. Geology 19, 147–150.

Crowley, T.J., Baum, S.K., 1991. Estimating Carboniferous sea-level fluctuations from Gondwanan ice extent. Geology 19, 975–977.

Crowley, T.J., Hyde, W.T., Short, D.A., 1989. Seasonal cycle variations on the super-continent of Pangaea. Geology 17, 457–460.

Crowley, T.J., Yip, K.-J., Baum, S.K., Moore, S.B., 1996. Modelling Carboniferous coal formation. Paleoclimates 2, 159–177.

Cruse, A.M., Lyons, T.W., 2004. Trace metal records of regional paleoenvironmental variability in Pennsylvanian (Upper Carboniferous) black shales. Chem. Geol. 206, 319–345.

Cys, J.M., Gibson, W.R., 1988. Pennsylvanian and Permian geology of the Permian Basin region. In: Sloss, L.L. (Ed.), Sedimentary Cover—North American Craton. U.S. Geol. Soc. Am., Geology of North America, vol. D-2, pp. 277–289.

Demaison, G.J., Moore, G.T., 1980. Anoxic environments and oil source bed genesis. Am. Assoc. Petrol. Geol. Bull. 64, 1179–1209.

Dickinson, W.R., Patchett, P.J., Ferguson, C.A., Suneson, N.H., Gleason, J.D., 2003. Nd isotopes of Atoka Formation (Pennsylvanian) turbidites displaying anomalous east-flowing paleocurrents in the frontal Ouachita Belt of Oklahoma: implications for regional sediment dispersal. J. Geol. 111, 733–740.

DiMichele, W.A., Phillips, T.L., 1996. Climate change, plant extinctions, and vegetational recovery during the Middle-Late Pennsylvanian transition: the case of tropical peat-forming environments in North America. In: Hart, M.L. (Ed.), Biotic Recovery From Mass Extinctions. Geol. Soc. London Spec. Publ., vol. 102, pp. 201–221.

Drinkwater, K.F., 1986. Physical oceanography of Hudson Strait and Ungava Bay. In: Voipio, A.V. (Ed.), The Baltic Sea. . Oceanogr. Ser., vol. 30. Elsevier, Amsterdam, pp. 237–264.

Dyrssen, D., Hall, P., Haraldsson, C., Iverfeldt, Å., Westerlund, S., 1984. Trace metal concentrations in the anoxic bottom water of Framvaren. In: Kramer, C.J.M., Duinker, J.C. (Eds.), Complexation of Trace Metals in Natural Waters. Martinus Nijhoff/Dr W. Junk Publishers, The Hague, Netherlands, pp. 239–243.

Edgar, N.T., Cecil, C.B., Mattick, R.E., Chivas, A.R., De Deckker, P., Djajadihardja, Y.S., 2003. A modern analogue for tectonic, eustatic, and climatic processes in cratonic basins: Gulf of Carpentaria, northern Australia. In: Cecil, C.B., Edgar, N.T. (Eds.), Climate Controls on Stratigraphy. SEPM-Soc. Sedim. Geol. Spec. Publ., vol. 77, pp. 193–205.

Elmore, R.D., Sutherland, P.K., White, P.B., 1990. Middle Pennsylvanian recurrent uplift of the Ouachita fold belt and basin subsidence in the Arkoma basin, Oklahoma. Geology 18, 906–909.

Emerson, S.R., Huested, S.S., 1991. Ocean anoxia and the concentrations of molybdenum and vanadium in seawater. Mar. Chem. 34, 177–196.

Emeis, K.-C., Whelan, J.K., Tarafa, M., 1991. Sedimentary and geochemical expressions of oxic and anoxic conditions on the Peru Shelf. In: Tyson, R.V., Pearson, T.H. (Eds.), Modern and Ancient Continental Shelf Anoxia. Geol. Soc. London Spec. Publ., vol. 58, pp. 155–170.

Eriksson, K.A., Campbell, I.H., Palin, J.M., Allen, C.M., Bock, B., 2004. Evidence for multiple recycling in Neoproterozoic through Pennsylvanian sedimentary rocks of the Central Appalachian Basin. J. Geol. 112, 261–276.

Fahrer, T.R., 1996. Stratigraphy, petrography and paleoecology of marine units within the Conemaugh Group (Upper Pennsylvanian) of the Appalachian Basin in Ohio, West Virginia and Pennsylvania. Unpubl. Ph.D. dissertation, University of Iowa, 298 pp.

Filippelli, G.M., 1997. Controls on phosphorus concentration and accumulation in oceanic sediments. Mar. Geol. 139, 231–240.

Föllmi, K.B., 1996. The phosphorus cycle, phosphogenesis and marine phosphate-rich deposits. Earth-Sci. Rev. 40, 55–124.

Fornós, J.J., Ahr, W.M., 1997. Temperate carbonates on a modern, low-energy, isolated ramp: the Balearic Platform, Spain. J. Sediment. Res. 67, 364–373.

Friedrich, H., 1969. Marine Biology (English translation of German original). Sidgwick & Jackson, London. 474 pp.

Ganeshram, R.S., Pedersen, T.F., 1998. Glacial–interglacial variability in upwelling and bioproductivity off NW Mexico: implications for Quaternary paleoclimate. Paleoceanography 13, 634–645.

Garfield, T.R., Scott, A.J., Walker, A.L., 1988. Overview: Late Paleozoic geologic history of the northern Denver Basin. In: Goolsby, S.M., Longman, M.W. (Eds.), Occurrence and Petrophysical Properties of Carbonate Reservoirs in the Rocky Mountain Region. Rocky Mountain Assoc. Geol., Denver, Colorado, pp. 1–18.

Garzione, C.N., Patchett, P.J., Ross, G.M., Nelson, J.A., 1997. Provenance of Paleozoic sedimentary rocks in the Canadian Cordilleran miogeocline: a Nd isotopic study. Can. J. Earth Sci. 34, 1603–1618.

Genger, D., Sethi, P., 1998. A geochemical and sedimentological investigation of high-resolution environmental changes within the Late Pennsylvanian (Missourian) Eudora core black shale of the Mid-Continent region, U.S.A. In: Schieber, J., Zimmerle, W., Sethi, P.S. (Eds.), Shales and Mudstones, vol. 1. Schweizerbart'sche, Stuttgart, pp. 271–293.

Gibling, M.R., Calder, J.H., Ryan, R., van de Ploo, H.W., Yeo, G.M., 1992. Late Carboniferous and early Permian drainage patterns in Atlantic Canada. Can. J. Earth Sci. 29, 338–352.

Glasby, G.P., Emelyanov, E.M., Zhamoida, V.A., Baturin, G.N., Leipe, T., Bahlo, R., Bonacker, P., 1997. Environments of formation of ferromanganese concretions in the Baltic Sea: a critical review. In: Nicholson, K., Hein, J.R., Bühn, B., Dasgupta, S. (Eds.), Manganese Mineralization: Geochemistry and Mineralogy of Terrestrial and Marine Deposits. Geol. Soc. London Spec. Publ., vol. 119, pp. 213–237.

Gleason, J.D., Patchett, P.J., Dickinson, W.R., Ruiz, J., 1994. Nd isotopes link Ouachita turbidites to Appalachian sources. Geology 22, 347–350.

Gleason, J.D., Patchett, P.J., Dickinson, W.R., Ruiz, J., 1995. Nd isotopic constraints on sediment sources of the Ouachita–Marathon fold belt. Geol. Soc. Am. Bull. 107, 1192–1210.

Glenn, C.R., et al., 1994. Phosphorus and phosphorites: sedimentology and environments of formation. Eclogae Geol. Helv. 87, 747–788.

Greb, S.F., Andrews, W.M., Cecil, C.B., DiMichele, W., Eble, C.F., Hower, J.C., 2003. Desmoinesian coal beds of the Eastern Interior and surrounding basins; the largest tropical peat mires in Earth history. In: Chan, M.A. (Ed.), Extreme

Depositional Environments; Mega End Members in Geologic Time. Geol. Soc. Am. Spec. Publ., vol. 370, pp. 127–150.
Hällfors, G., Niemi, A., Ackefors, H., Lassig, J., Leppäkoski, E., 1981. Biological oceanography. In: Voipio, A.V. (Ed.), The Baltic Sea. Oceanogr. Ser., vol. 30. Elsevier, Amsterdam, pp. 219–274.
Handford, C.R., Dutton, S.P., 1980. Pennsylvanian–Early Permian depositional systems and shelf-margin evolution, Palo Duro Basin, Texas. Am. Assoc. Petrol. Geol. Bull. 64, 88–106.
Handford, C.R., Fredericks, P.E., 1980. Lower Permian Facies of the Palo Duro Basin, Texas: Depositional Systems, Shelf-margin Evolution, Paleogeography, and Petroleum Potential. Report of Investigations, vol. 102. Texas Bureau of Economic Geology. 31 pp.
Handford, C.R., Dutton, S.P., Fredericks, P.E., 1981. Regional Cross Sections of the Texas Panhandle: Precambrian to Mid-Permian, 8. Texas Bureau of Economic Geology, Austin, Texas. + 7 plates.
Hatch, J.R., Leventhal, J.S., 1992. Relationship between inferred redox potential of the depositional environment and geochemistry of the Upper Pennsylvanian (Missourian) Stark Shale Member of the Dennis Limestone, Wabaunsee County, Kansas, U.S.A. Chem. Geol. 99, 65–82.
Hay, W.W., 1995. Paleoceanography of marine organic-carbon-rich sediments. In: Huc, A.-Y. (Ed.), Paleogeography, Paleoclimate, and Source Rocks. Am. Assoc. Petrol. Geol., Studies in Geology, vol. 40, pp. 21–59.
Hay, W.W., Wold, C.N., Söding, E., Flögel, S., 2001. Evolution of sediment fluxes and ocean salinity. In: Merriam, D.F., Davis, J.C. (Eds.), Geologic Modeling and Simulation: Sedimentary Systems. Kluwer Academic/Plenum Publishers, New York, pp. 153–167.
Hay, W.W., Migdisov, A., Balukhovsky, A.N., Wold, C.N., Flögel, S., Söding, E., 2006. Evaporites and the salinity of the ocean during the Phanerozoic: implications for climate, ocean circulation and life. Palaeogeogr. Palaeoclimatol. Palaeoecol. 240, 3–46.
Heckel, P.H., 1977. Origin of phosphatic black shale facies in Pennsylvanian cyclothems of mid-contintent North America. Am. Assoc. Petrol. Geol. Bull. 61, 1045–1068.
Heckel, P.H., 1980. Paleogeography of eustatic model for deposition of Midcontinent Upper Pennsylvanian cyclothems. In: Fouch, T.D., Magathan, E.R. (Eds.), Paleozoic Paleogeography of the West-Central United States. SEPM-Rocky Mountain Section, Denver, Colorado, pp. 197–215.
Heckel, P.H., 1986. Sea-level curve for Pennsylvanian eustatic marine transgressive–regressive depositional cycles along midcontinent outcrop belt, North America. Geology 14, 330–334.
Heckel, P.H., 1991. Thin widespread Pennsylvanian black shales of Midcontinent North America: a record of a cyclic succession of widespread pycnoclines in a fluctuating epeiric sea. In: Tyson, R.V., Pearson, T.H. (Eds.), Modern and Ancient Continental Shelf Anoxia. Geol. Soc. London. Spec. Publ., vol. 58, pp. 259–273.
Heckel, P.H., 1994. Evaluation of evidence for glacial-eustatic control over marine Pennsylvanian cyclothems in North America and consideration of possible tectonic effects. In: Dennison, J.M., Ettensohn, F.R. (Eds.), Tectonic and Eustatic Controls on Sedimentary Cycles. . Concepts in Sedimentology and Paleontology, vol. 4. SEPM, pp. 65–87.
Heckel, P.H., 1995. Glacial-eustatic base-level climatic model for late Middle to Late Pennsylvanian coal-bed formation in the Appalachian Basin. J. Sediment. Res. B65, 348–356.
Heckel, P.H., 2004. Updated cyclothem grouping chart and observations on the grouping of Pennsylvanian cyclothems in Midcontinent North America. Newslett. Carbonif. Stratigr. 22, 18–22.
Heckel, P.H., Baesemann, J.M., 1975. Environmental interpretation of conodont distribution in Upper Pennsylvanian (Missourian) megacyclothems in eastern Kansas. Am. Assoc. Petrol. Geol. Bull. 59, 486–509.
Hendy, I.L., Pedersen, T.F., 2006. Oxygen minimum zone expansion in the eastern tropical North Pacific during deglaciation. Geophys. Res. Lett. 33, L20602. doi:10.1029/2006GL025975.
Hill, C.A., 1999. Reevaluation of the Hovey Channel in the Delaware Basin, West Texas. Am. Assoc. Petrol. Geol. Bull. 83, 277–294.
Hirst, D.M., 1974. Geochemistry of sediments from eleven Black Sea cores. In: Degens, E.T., Ross, D.A. (Eds.), The Black Sea—Geology, Chemistry, and Biology. Am. Assoc. Petrol. Geol. Mem., vol. 20, pp. 430–455.
Hobbs III, C.H., 2004. Geological history of Chesapeake Bay, USA. Quat. Sci. Rev. 23, 641–661.
Hoffman, D.L., Algeo, T.J., Maynard, J.B., Joachimski, M.M., Hower, J.C., Jaminski, J., 1998. Regional and stratigraphic variation in bottomwater anoxia in offshore core shales of Upper Pennsylvanian cyclothems from the Eastern Midcontinent Shelf (Kansas), USA. In: Schieber, J., Zimmerle, W., Sethi, P.S. (Eds.), Shales and Mudstones, vol. 1. Schweizerbart'sche, Stuttgart, pp. 243–269.
Isbell, J.L., Miller, M.F., Wolfe, K.L., Lenaker, P.A., 2003. Timing of late Paleozoic glaciation in Gondwana: was glaciation responsible for the development of Northern Hemisphere cyclothems? In: Chan, M.A., Archer, A.A. (Eds.), Extreme Depositional Environments: Mega End Members in Geologic Time. Geol. Soc. Am. Spec. Pap., vol. 370, pp. 5–24.
Jackson, W.E., 1964. Depositional topography and cyclic deposition in west-central Texas. Am. Assoc. Petrol. Geol. Bull. 48, 317–328.
Joachimski, M.M., von Bitter, P.H., Buggisch, W., 2006. Constraints on Pennsylvanian glacioeustatic sea-level changes using oxygen isotopes of conodont apatite. Geology 34, 277–280.
Joeckel, R.M., 1994. Virgilian (Upper Pennsylvanian) paleosols in the Upper Lawrence Formation (Douglas Group) and in the Snyderville Shale Member (Oread Formation, Shawnee Group) of the northern Midcontinent, USA: pedologic contrasts in a cyclothem sequence. J. Sediment. Res. 64A, 853–866.
Joeckel, R.M., 1999. Paleosol in Galesburg Formation (Kansas City Group, Upper Pennsylvanian), northern Midcontinent, U.S.A.: evidence for climate change and mechanisms of marine transgression. J. Sediment. Res. 69, 720–737.
Jones, A.T., Fielding, C.R., 2004. Sedimentological record of the late Paleozoic glaciation in Queensland, Australia. Geology 32, 153–156.
Karl, D.M., Knauer, G.A., 1991. Microbial production and particle flux in the upper 350 m of the Black Sea. Deep Sea Res., Part A 38, S921–S942.
Kennett, J., 1982. Marine Geology. Prentice-Hall, Englewood Cliffs, New Jersey. 813 pp.
Kennish, M.J. (Ed.), 2001. Practical Handbook of Marine Science, 3rd ed. CRC Press, Boca Raton, Florida. 876 pp.
Kerr, D.R., Dott Jr., R.H., 1988. Eolian dune types preserved in the Tensleep Sandstone (Pennsylvanian–Permian), north-central Wyoming. Sediment. Geol. 56, 383–402.
Kidder, D.L., 1985. Petrology and origin of phosphatic nodules from the midcontinent Pennsylvanian epicontinental sea. J. Sediment. Petrol. 55, 809–816.
Kidder, D.L., Hussein, R.A.M., Mapes, R.H., Eddy-Dilek, C.A., 1996. Regional diagenetic variation in maximum-transgression phosphates from Midcontinent Pennsylvanian shales. In: Witzke, B.J., Ludvigson, G.A., Day, J. (Eds.), Paleozoic Sequence Stratigraphy: Views from the North American Craton. Geol. Soc. Am. Spec. Pap., vol. 306, pp. 351–358.
Kienast, S.S., Calvert, S.E., Pedersen, T., 2002. Nitrogen isotope and productivity variations along the northeast Pacific margin over the last 120 kyr: surface and subsurface paleoceanography. Paleoceanography 17, 1–17.
Kluth, C.J., 1986. Plate tectonics of the Ancestral Rocky Mountains. In: Peterson, J.A. (Ed.), Paleotectonics and Sedimentation in the Rocky Mountain Region, United States. Am. Assoc. Petrol. Geol. Mem., vol. 41, pp. 353–369.
Knauth, L.P., 1998. Salinity history of the Earth's early history. Nature 395, 554.
Kullenberg, G., 1981. Physical oceanography. In: Voipio, A.V. (Ed.), The Baltic Sea. Oceanography Series, vol. 30. Elsevier, Amsterdam, pp. 135–183.
Kvale, E.P., Mastalerz, M., 1998. Evidence of ancient freshwater tidal deposits. In: Alexander, C.R. (Ed.), Tidalites: Processes and Products. SEPM-SSG Spec. Publ., vol. 61, pp. 95–107.
Landis, C.R., Trabelsi, A., Strathearn, G., 1992. Hydrocarbon potential of selected Permian Basin shales as classified within the organic facies concept. In: Johnson, K.S., Cardott, B.J. (Eds.), Source Rocks in the Southern Midcontinent, 1990 Symposium. Oklahoma Geol. Surv. Circ., vol. 93, pp. 229–247.
Levitus, S., Boyer, T., 1994. World Ocean Atlas 1994, Volume 2: Oxygen. . NOAA Atlas NESDIS, vol. 2. U.S. Department of Commerce, Washington, D.C. (http://iridl.ldeo.columbia.edu/sources/.levitus94/.annual/.O2/).
Luebking, G.A., Longman, M.W., Carlisle, W.J., 2001. Unconformity-related chert/dolomite production in the Pennsylvanian Amsden Formation, Wolf Springs fields, Bull Mountains basin of central Montana. Am. Assoc. Petrol. Geol. Bull. 85, 131–148.
Lyons, T.W., 1991. Upper Holocene sediments of the Black Sea: summary of Leg 4 box cores (1988 Black Sea Oceanographic Expedition). In: Izdar, E., Murray, J.W. (Eds.), Black Sea Oceanography. Kluwer, Dordrecht, pp. 401–441.
Lyons, T.W., Werne, J.P., Hollander, D.J., Murray, R.W., 2003. Contrasting sulfur geochemistry and Fe/Al and Mo/Al ratios across the last oxic-to-anoxic transition in the Cariaco Basin, Venezuela. Chem. Geol. 195, 131–157.
Malinky, J.M., Heckel, P.H., 1998. Paleoecology and taphonomy of faunal assemblages in gray "core" (offshore) shales in Midcontinent Pennsylvanian cyclothems. Palaios 13, 311–334.
Mankiewicz, D., Steidtmann, J.R., 1979. Depositional environments and diagenesis of the Tensleep Sandstone: eastern Big Horn Basin, Wyoming. In: Scholle, P.A. (Ed.), Aspects of Diagenesis. SEPM Spec. Publ., vol. 26, pp. 319–336.
Marlow, J.R., Lange, C.B., Wefer, G., Rosell-Melé, A., 2000. Upwelling intensification as part of the Pliocene–Pleistocene climate transition. Science 290, 2288–2291.
Maughan, E.K., 1984. Paleogeographic setting of Pennsylvanian Tyler Formation and relation to underlying Mississippian rocks in Montana and North Dakota. Am. Assoc. Petrol. Geol. Bull. 68, 178–195.
Maughan, E.K., 1993. The Ancestral Rocky Mountains in Wyoming. In: Snoke, A.W., Steidtmann, J.R., Roberts, S.M. (Eds.), Geology of Wyoming. Geol. Surv. Wyoming Mem., vol. 5, pp. 188–207.
Miller, E.L., Miller, M.M., Stevens, C.H., Wright, J.E., Madrid, R., 1992. Late Paleozoic paleogeography and tectonic evolution of the western U.S. Cordillera. In: Burchfiel, B.C., Lipman, P.W., Zoback, M.L. (Eds.), The Cordilleran Orogen: Conterminous. U.S. Geol. Soc. Am., Geology of North America, vol. G-3, pp. 57–106.
Montañez, I.P., Tabor, N.J., Niemeier, D., DiMichele, W.A., Frank, T.D., Fielding, C.R., Isbell, J.L., Birgenheier, L.P., Rygel, M.C., 2007. CO_2-forced climate and vegetation instability during Late Paleozoic deglaciation. Science 315, 87–91.
Multer, H.G. (Ed.), 1977. A Field Guide to Some Carbonate Rock Environments: Florida Keys and Western Bahamas. Kendall/Hunt Publishing, Dubuque, Iowa. 415 pp.
Murphy, J.B., 2000. Tectonic influence on sedimentation along the southern flank of the late Paleozoic Magdalen basin in the Canadian Appalachians: geochemical and isotopic constraints on the Horton Group in the St. Marys basin, Nova Scotia. Geol. Soc. Am. Bull. 112, 997–1011.
Murray, J.W., 1991. Hydrographic variability in the Black Sea. In: Izdar, E., Murray, J.W. (Eds.), Black Sea Oceanography. Kluwer, Dordrecht, pp. 1–16.
Parrish, J.T., 1993. Climate of the supercontinent Pangea. J. Geol. 101, 215–233.
Parrish, J.T., Peterson, F., 1988. Wind directions predicted from global circulation models and wind directions determined from eolian sandstones of the western United States—a comparison. Sediment. Geol. 56, 261–282.
Patchett, P.J., Gehrels, G.E., 1998. Continental influence on Canadian Cordilleran terranes from Nd isotopic study, and significance for crustal growth processes. J. Geol. 106, 269–280.
Patchett, P.J., Ross, G.M., Gleason, J.D., 1999. Continental drainage in North America during the Phanerozoic from Nd Isotopes. Science 283, 671–673.
Pelletier, B.R., 1986. Seafloor morphology and sediments. In: Martini, I.P. (Ed.), Canadian Inland Seas. Oceanogr. Ser., vol. 44. Elsevier, Amsterdam, pp. 143–162.
Pelletier, B.R., Wagner, F.J.E., Grant, A.C., 1968. Marine geology. In: Beals, C.S. (Ed.), Science, History and Hudson Bay, 2. Canad. Dept. Energy, Mines and Resources, Ottawa, Canada, pp. 557–613.

Peterson, J.A., MacCary, L.M., 1987. Regional stratigraphy and general petroleum geology of the U.S. portion of the Williston Basin and adjacent areas. In: Longman, M.W. (Ed.), Williston Basin: Anatomy of a Cratonic Oil Province. Rocky Mountain Assoc. Geol., Denver, Colorado, pp. 9–43.

Pett, R.J., Roff, J.C., 1982. Some observations and deductions concerning the deep waters of Hudson Bay. Nat. Can. 109, 767–774.

Poulsen, C.J., Pollard, D., Montañez, I.P., Rowley, D., 2007. Late Paleozoic tropical climate response to Gondwanan deglaciation. Geology 35, 771–774.

Pride, C., Thunell, R., Sigman, D., Keigwin, L., Altabet, M., Tappa, E., 1999. Nitrogen isotopic variations in the Gulf of California since the last deglaciation: response to global climate change. Paleoceanography 14, 397–409.

Prinsenberg, S.J., 1986. Salinity and temperature distributions of Hudson Bay and James Bay. In: Martini, I.P. (Ed.), Canadian Inland Seas. Oceanogr. Ser., vol. 44. Elsevier, Amsterdam, pp. 163–186.

Raffaelli, D.G., 1996. Intertidal Ecology. Chapman & Hall, New York. 356 pp.

Reimers, C.E., Suess, E., 1983. Spatial and temporal patterns of organic matter accumulation on the Peru continental margin. In: Thiede, J., Suess, E. (Eds.), Coastal Upwelling: Its Sediment Record, pt. B: Sedimentary Records of Ancient Coastal Upwelling. Plenum, New York, pp. 311–345.

Roden, G.I., 1964. Oceanographic aspects of Gulf of California. In: van Andel, T.H., Shor Jr., G.G. (Eds.), Marine Geology of the Gulf of California. Am. Assoc. Petrol. Geol. Mem., vol. 3, pp. 30–58.

Roff, J.C., Legendre, L., 1986. Physico-chemical and biological oceanography of Hudson Bay. In: Martini, I.P. (Ed.), Canadian Inland Seas. Oceanogr. Ser., vol. 44. Elsevier, Amsterdam, pp. 265–291.

Samson, S.D., Coler, D.G., Speer, J.A., 1995. Geochemical and Nd–Sr–Pb isotopic composition of Alleghenian granites of the southern Appalachians: origin, tectonic setting, and source characterization. Earth Planet. Sci. Lett. 134, 359–376.

Schatzel, S.J., Stewart, B.W., 2000. Coal mineral matter origin and provenance; a neodymium isotope study of the lower Kittanning Coal bed, western Pennsylvania. Geol. Soc. Am. Prog. Abstr. 32 (7), 82.

Schultz, R.B., Coveney Jr., R.M., 1992. Time-dependent changes for Midcontinent Pennsylvanian black shales. Chem. Geol. 99, 83–100.

Schutter, S.R., Heckel, P.H., 1985. Missourian (Early Late Pennsylvanian) climate in Midcontinent North America. Int. J. Coal Geol. 5, 111–140.

Schwartz, J.J., Frost, C.D., Snoke, A.W., 2005. Contrasting Nd isotopic characteristics from the Baker Terrane of northeastern Oregon; implications for tectonic evolution. Geol. Soc. Am. Prog. Abstr. 37 (7), 19.

Scotese, C.R., 1998. Quicktime Computer Animations, PALEOMAP Project. Dept. Geology, University of Texas at Arlington, Arlington, Texas. CD-ROM.

Scranton, M.I., Sayles, F.L., Bacon, M.P., Brewer, P.G., 1987. Temporal changes in the hydrography and chemistry of the Cariaco Trench. Deep-Sea Res. 34, 945–963.

Shimkus, K.M., Trimonis, E.S., 1974. Modern sedimentation in Black Sea. In: Degens, E.T., Ross, D.A. (Eds.), Black Sea—Geology, Chemistry, and Biology. Am. Assoc. Petrol. Geol. Mem., vol. 20, pp. 249–278.

Simard, R.-L., Dostal, J., Roots, C.F., 2003. Development of late Paleozoic volcanic arcs in the Canadian Cordillera: an example from the Klinit Group, northern British Columbia and southern Yukon. Can. J. Earth Sci. 40, 907–924.

Smith, R.L., 1992. Coastal upwelling in the modern ocean. In: Summerhayes, C.P., Prell, W.L., Emeis, K.-C. (Eds.), Upwelling Systems: Evolution Since the Early Miocene. Geol. Soc. London Spec. Publ., vol. 64, pp. 9–28.

Smith, A.D., Lambert, R. St.J., 1995. Nd, Sr, and Pb isotopic evidence for contrasting origins of late Paleozoic volcanic rocks from the Slide Mountain and Cache Creek terranes, south-central British Columbia. Can. J. Earth Sci. 32, 447–459.

Somers, I.F., Long, B.G., 1994. Note on the sediments and hydrology of the Gulf of Carpentaria, Australia. Aust. J. Mar. Freshw. Res. 45, 283–291.

Soreghan, G.S., Giles, K.A., 1999. Amplitudes of Late Pennsylvanian glacioeustasy. Geology 27, 255–258.

Soreghan, G.S., Elmore, R.D., Lewchuk, M.T., 2002. Sedimentologic-magnetic record of western Pangean climate in upper Paleozoic loessite (lower Cutler beds, Utah). Geol. Soc. Am. Bull. 114, 1019–1035.

Speed, R.C., Sharp, W.D., Foland, K.A., 1997. Late Paleozoic granitoid gneisses of northeastern Venezuela and the North America–Gondwana collision zone. J. Geol. 105, 457–470.

Stamm, R.G., Wardlaw, B.R., 2003. Conodont faunas of the late Middle Pennsylvanian (Desmoinesian) lower Kittanning cyclothem, U.S.A. In: Cecil, C.B. (Ed.), Climate Controls on Stratigraphy. SEPM-SSG Spec. Publ., vol. 77, pp. 95–121.

Summerhayes, C.P., Emeis, K.-C., Angel, M.V., Smith, R.L., Zeitzschel, B., 1995. Upwelling in the ocean: modern processes and ancient records. In: Summerhayes, C.P., et al. (Ed.), Upwelling in the Ocean: Modern Processes and Ancient Records. Wiley, Chichester, United Kingdom, pp. 1–37.

Tabor, N.J., 2007. Permo-Pennsylvanian palaeotemperatures from Fe-oxide and phyllosilicate $\delta^{18}O$ values. Earth Planet. Sci. Lett. 253, 159–171.

Tabor, N.J., Montañez, I.P., 2002. Shifts in late Paleozoic atmospheric circulation over western equatorial Pangea: insights from pedogenic mineral $\delta^{18}O$ compositions. Geology 30, 1127–1130.

Thunell, R.C., Kepple, A.B., 2004. Glacial–Holocene $\delta_{15}N$ record from the Gulf of Tehuantepec, Mexico: implications for denitrification in the eastern equatorial Pacific and changes in atmospheric N_2O. Glob. Biogeochem. Cycles 18, 1–12 GB1001.

Trexler, J.H., Snyder, W.S., Cashman, P.H., Gallegos, D.M., Spinosa, C., 1991. Mississippian through Permian orogenesis in eastern Nevada: post-Antler pre-Sonoma tectonics of the western Cordillera. In: Cooper, J.D., Stevens, C.H. (Eds.), Paleozoic Paleogeography of the Western United States, II. SEPM-Pacific Section, vol. 67, pp. 317–330.

Walker, D.A., Golonka, J., Ried, A., Reid, S., 1995. The effects of paleolatitude and paleogeography on carbonate sedimentation in the Late Paleozoic. In: Huc, A.-Y. (Ed.), Paleogeography, Paleoclimate, and Source Rocks. Am. Assoc. Petrol. Geol., Studies in Geology, vol. 40, pp. 133–155.

Wallin, E.T., Noto, R.C., Gehrels, G.E., 2000. Provenance of the Antelope Mountain Quartzite, Yreka terrane, California: evidence for large-scale late Paleozoic sinistral displacement along the North American Cordilleran margin and implications for the mid-Paleozoic fringing-arc model. In: Soreghan, M.J., Gehrels, G.E. (Eds.), Paleozoic and Triassic Paleogeography and Tectonics of Western Nevada and Northern California. Geol. Soc. Am. Spec. Pap., vol. 347, pp. 119–131.

Wanless, H.R., Wright, C.R., 1978. Paleoenvironmental maps of Pennsylvanian rocks, Illinois Basin and northern midcontinent region. Geol. Soc. Am., Map and Chart Series MC-23. 32 pp.

Watney, W.L., French, J.A., Doveton, J.H., Youle, J.C., Guy, W.J., 1995. Cycle hierarchy and genetic stratigraphy of Middle and Upper Pennsylvanian strata in the Upper Midcontinent. In: Hyne, N.J. (Ed.), Sequence Stratigraphy of the Mid-Continent. Tulsa Geol. Soc. Spec. Publ., vol. 4, pp. 141–192.

Wells, M.R., Allison, P.A., Hampson, G.J., Piggott, M.D., Pain, C.C., 2005a. Modeling ancient tides: the Upper Carboniferous epi-continental seaway of Northwest Europe. Sedimentology 52, 715–735.

Wells, M.R., Allison, P.A., Piggott, M.D., Pain, C.C., Hampson, G.J., De Oliveira, C.R.E., 2005b. Large sea, small tides: the Late Carboniferous seaway of NW Europe. J. Geol. Soc. Lond. 162, 417–420.

Wignall, P.B., 1994. Black Shales. Clarendon, Oxford. 127 pp.

Witzke, B.J., 1987. Models for circulation patterns in epicontinental seas applied to Paleozoic facies of North American craton. Paleoceanography 2, 229–248.

Wyrtki, K., 1967. Circulation and water masses in the eastern equatorial Pacific Ocean. Int. J. Oceanogr. Limnol. 1, 117–147.

Yang, K.-M., Dorobek, S.L., 1995. The Permian Basin of West Texas and New Mexico: tectonic history of a “composite” foreland basin and its effects on stratigraphic development. In: Dorobek, S.L., Ross, G.M. (Eds.), Stratigraphic Evolution of Foreland Basins. SEPM-Soc. Sedim. Geol. Spec. Publ., vol. 52, pp. 149–174.

Ye, H., Royden, L., Burchfiel, C., Schuepbach, M., 1996. Late Paleozoic deformation of interior North America: the greater Ancestral Rocky Mountains. Am. Assoc. Petrol. Geol. Bull. 80, 1397–1432.

Youle, J.C., Watney, W.L., Lambert, L.L., 1994. Stratal hierarchy and sequence stratigraphy—Middle Pennsylvanian, southwestern Kansas, U.S.A. In: Klein, G.D. (Ed.), Pangea: Paleoclimate, Tectonics, and Sedimentation During Accretion, Zenith, and Breakup of a Supercontinent. Geol. Soc. Am. Spec. Pap., vol. 288, pp. 267–285.