The Late Pennsylvanian Midcontinent Sea (LPMS) of North America reached its greatest extent (~2.1×10⁶ km²) 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.
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 δ34S-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...
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 cyclothem 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

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Fig. 3. Organic fraction of cyclothem 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 locations). 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”).
southwest, 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 (Arbenn, 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 Ancstral 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 Ancstral 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; 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). The LPMS received an estimated discharge volume of ~800–1500 km³ yr⁻¹, equivalent to that of modern Hudson Bay (~975 km³ yr⁻¹) and the Baltic Sea (~485 km³ yr⁻¹) when normalized to drainage area (Algeo et al., 2005). 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 km³ 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 subpolar coastal 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 δ¹⁸O values across the Appalachian and Ouachita basins, suggesting extensive reworking and transport of sediment westward.
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).
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. 6).

![Diagram](NE-SW.png)

**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).

![Diagram](Spatial-Patterns.png)

**Fig. 5.** Spatial patterns within LPMS: (A) Neodymium isotopes ($\varepsilon_{\text{Nd}}$); (B) watermass circulation patterns; (C) palaeoredox conditions. In A, no label = sedimentary rocks, V = volcanic rocks, and X = crystalline rocks. Sediment $\varepsilon_{\text{Nd}}$ 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, $\varepsilon_{\text{Nd}}$ 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 volcanlastic 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; Jeeckel, 1994, 1999). The amplitude of Late Pennsylvaniaian glacio-eustatic fluctuations was ~60–150 m (median estimate ~100 m) based on analyses of contemporaneous ice-sheet mass (Crowley and Baum, 1991; Soreghan and Giles, 1999) and condont δ18O variation (Joachimski et al., 2006). These amplitudes are equivalent to maximum hightstand 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; Chester, 1990). This calculation yields a deep-to-shallow density differential (ΔσN) of ~11±4, of which ~70% is attributable to salinity variation, with a halocline, and ~30% to temperature variation, consistent with a thermocline. This ΔσN 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; Hobs, 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 synchronicity 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 subpycnocline 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 northeasterly 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. Coveny 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 Pennsylvania–early Permian uplift of the Wyoming Arch (Maughan, 1983; 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 northwestern margin of the Ouachita–Marathon Front, at a paleolatitude of ~5° 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*10^3 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 hypersaline watermasses (~20% salinity), as in many modern estuaries and the Baltic Sea (Friedrich, 1969; Hälfors 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-nektonic biota consisting of conodonts, ammonoids, fish, and orbiculoiid 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, echiuroiderms, 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 hypersaline 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 subpycnocline 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 (Hirsh, 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., Emely 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 physicochemical 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 subpynocline 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 10% 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 grainule 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
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 hydrogenuously sourced trace metals became highly concentrated owing to limited siliclastic 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 cyclothem 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$34S = 40 ± 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 ~0.3×10$^6$ 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). Cyclotherm model at left from Heckel (1977); core-core shale stratigraphy at right modified from Algeo et al. (2004).
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 associated with preconditioned, oxygen-deficient eastern tropical Panthalassic surface waters and oxygen demand in intermediate waters of the Panthalassic OMZ shallowed to the maximum degree (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-pycnocline 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
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, unrestricted deepwater exchange, and (3) lateral advection of deepwaters advected through the Hudson Strait, which are preconditioned "estuarine circulation" systems versus (B) silled basins and (C) continent-margin upwelling zones. From Algeo et al. (2008).

![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).](image-url)
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 upon 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 (\(>0.3 \times 10^6 \text{ km}^2\)) 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 subpycnocline 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).

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