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Phosphorites, glass ramps and carbonate factories: The evolution of an epicontinental sea and a Late Palaeozoic upwelling system (Phosphoria Rock Complex)

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ABSTRACT

The Permian Phosphoria Rock Complex of the western USA contains an enigmatic assemblage of bioelemental rocks (i.e. phosphorites and cherts) that accumulated in a depositional system with no modern analogue. This study utilizes detailed sedimentological, stratigraphic and petrographic examination to evaluate the genetic relations of phosphorites, spiculitic chert and carbonates of the Ervay cycle (depositional sequence) and propose a unified oceanographic model for their deposition.

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The Ervay cycle contains three marine and one terrestrial facies association, each of which composes the bulk of a single lithostratigraphic unit. The marine facies associations include: (i) granular phosphorites (Retort Member); (ii) spiculitic cherty dolostones (Tosi Member); and (iii) marine to peritidal carbonates (Ervay Member). Red beds and intercalated gypsum (Goose Egg Formation) accumulated in the vast desert adjacent to the sea. The three marine members are chronostratigraphically distinct, successive and conformably stacked. They are not coeval facies belts. They reflect the progressive evolution of the epicontinental sea from the location of: (i) authigenic phosphogenesis (lowstand to transgression); to (ii) a glass ramp with biosiliceous (sponge) deposition (transgression); to (iii) a carbonate ramp (regression). This succession of switching biochemical sediment factories records the evolution of sea-level, nutrient supply, upwelling, oxygenation and dissolved Si. Intense upwelling, potentially coupled with aeolian input, led to sedimentary condensation and phosphogenesis. Decreased upwelling intensity during transgression increased oxygenation sufficiently for a siliceous sponge benthos. Sponges were favoured over biocalcifiers due to elevated dissolved silica and a low carbonate saturation state. The cessation of sponge dominance and transition to a carbonate ramp occurred due to decreasing upwelling intensity, Si drawdown and an increased carbonate saturation state. These results provide insight into the role of Si loading in faunal turnover on glass ramps and highlight how differences in dissolved Si utilizers in pre-Cretaceous versus post-Cretaceous upwelling systems influence the resultant deposits.

Keywords Ervay Member, glass ramp, palaeoceanography, Permian, Phosphoria, phosphorites, silica cycle, upwelling.

INTRODUCTION

The Phosphoria Rock Complex (PRC) of the western United States is a diverse group of sedimentary rocks that accumulated in a Permian subtropical epicontinental sea along western Pangaea (the ‘Phosphoria Sea’; Fig. 1). Its well-known phosphorites comprise one of a small number of phosphorite giants, having formed through phosphogenesis in environments that extended over hundreds of kilometres and have no modern analogue (Glenn *et al.*, 1994b; Maughan, 1994; Hiatt & Budd, 2003; Hein *et al.*, 2004; Pufahl, 2010). As such, the PRC has been the focus of significant attention aimed at reconstructing the potentially unique oceanographic conditions and any analogy to modern shelf-edge phosphogenic environments (e.g. Hiatt & Budd, 2001; Piper *et al.*, 2007). Although most hypotheses invoke upwelling, consensus about the detailed oceanographic conditions remains elusive (Hein *et al.*, 2004).

Recent attention has increasingly focused on the origin of chert in the Phosphoria Rock Complex, which composes a significant percentage of the succession (Fig. 2; McKelvey *et al.*, 1959). Long recognized as the product of silica remobilized from sponge spicules, the chert-rich deposits were traditionally interpreted as accumulating outboard of shallow-marine sediments (Keller, 1941; Cressman, 1955; Peterson, 1984). Recent work, however, has re-interpreted many as glass-ramp deposits (Wistort *et al.*, 2017; Hood *et al.*, 2018; Ritterbush, 2019; Matheson & Frank, 2020).

Glass ramps (*sensu* Gates *et al.*, 2004) are depositional systems in which siliceous sponge meadows with subordinate to non-existent calcareous invertebrates extend from offshore environments to shallow-marine and coastal settings. Glass ramps represent a mode of neritic sedimentation distinct from siliciclastics and carbonates (Ritterbush, 2019). One of the most prominent areas of glass-ramp deposition was the north-western Pangaeian margin during the Middle to Late Permian (orange, Fig. 1A; Ritterbush, 2019). During this interval of time, referred to as the Permian Chert Event, radiolarites and spiculites accumulated offshore in a belt stretching from the Phosphoria Sea northward. At the same time, glass ramps were a prominent, although episodic, component of shallow seas where they alternated with carbonates and phosphorites (e.g. Murchey & Jones, 1992; Murchey, 2004; Blomeier *et al.*, 2013; Ritterbush, 2019). As with the phosphorites, no modern analogue exists for widespread neritic sponge meadows and hypotheses for the conditions

that would favour the dominance of sponges over biocalcifiers are wide ranging (Reid *et al.*, 2008; Beauchamp & Grasby, 2012; Ritterbush, 2019).

The PRC phosphorites and glass-ramp deposits have largely been studied separately. Comprehensive studies incorporating both are undoubtedly required to produce a unified model regarding the biological and oceanographic controls on sedimentation. To this end, the upper PRC (the Ervay cycle) contains a conformable succession of three disparate lithologies that indicate that the eastern Phosphoria Sea successively evolved from: (i) a location of authigenic phosphogenesis; to (ii) an area of biosiliceous deposition on a glass ramp; to (iii) a carbonate ramp (Inden & Coalson, 1996; Matheson & Frank, 2020).

The purpose of this study is to document the sedimentology and stratigraphy of the Ervay cycle in the Bighorn Basin, where it is best exposed. The results are utilized to test hypotheses regarding the formation of the phosphorites and glass-ramp deposits and to propose an oceanographic model that encompasses both. Addressing the conditions that led to the transitions between sediment factories provides a novel dataset that differs from studies that utilize individual lithological members. This dataset provides insight into the late Palaeozoic oceanographic conditions along western Pangaea and the enigmatic controls on Phanerozoic glass-ramp formation.

GEOLOGICAL SETTING

The Phosphoria Sea sat landward of the Havallah-Slide Mountain backarc basin and was surrounded by a widespread desert characterized by windblown silt (loess), mudflats and salinas (Fig. 1A and B; Walker, 1967; Benison & Goldstein, 2000; Sweet *et al.*, 2013; Domeier & Torsvik, 2014).

The ‘Phosphoria Rock Complex’ (PRC; *sensu* Yochelson, 1968) is a catch-all term for all sedimentary rocks that accumulated in or around the Phosphoria Sea. The informal group contains four formations. Its namesake, the Phosphoria Formation, contains phosphorites, cherts, and organic-rich mudstones that comprise the bulk of the PRC in the centre of the region covered by the sea (south Idaho, Fig. 1C; Claypool *et al.*, 1978; Cook, 1984; Hiatt & Budd, 2001). The Shedhorn Formation to the north contains shallow-marine siliciclastics shed off the Milk River Uplift (Cressman & Swanson, 1964; Thornburg, 1990). Carbonates of the Park City Formation accumulated around the remainder of

the sea (Inden & Coalson, 1996; Whalen, 1996). In the Bighorn Basin, deposits from the desert surrounding the sea are referred to as the Goose Egg Formation (Burk & Thomas, 1956; Lane, 1973). Each formation composes the bulk of the PRC succession in a specific portion of the sea (Fig. 1C). However, tongues of each formation also extend significant distances (Fig. 2A; McKelvey *et al.*, 1959). For example, in the western Bighorn Basin, the PRC is predominantly carbonates of the Park City Formation but also contains tongues of the Phosphoria Formation that extend from its depocentre to the west (Figs 1C and 2B). In the southern extension of the Phosphoria Sea, the confusingly named Park City Group contains a variety of lithologies similar to those in the PRC (Fig. 1C).

The PRC contains three retrogradationally stacked depositional sequences (cycles): the Grandeur, Franson and Ervay (Fig. 2; Peterson, 1984; Hiatt, 1997). This expansion of the sea occurred despite falling eustatic sea-level (Ross & Ross, 1988; Haq & Schutter, 2008), which is attributed to regional tectonic subsidence, potentially related to shifts in plate boundary dynamics (Maughan, 1984; Peterson, 1984; Trexler *et al.*, 2004; Henderson *et al.*, 2012; Beauchamp, 2017; Henderson, 2017; Pommer & Sarg, 2019). Progressive shifts in the PRC depocentre northward and eastward (i.e. towards the Bighorn Basin) controlled first-order stratigraphic trends and imparted regional variability between the expression of the Franson and Ervay cycles. These two cycles each contain phosphorite, chert and carbonate, although their stacking varies (Fig. 2A). The Franson cycle has received the bulk of attention due to the economic exploitation of Meade Peak phosphorites in Idaho (Hein *et al.*, 2004).

The Bighorn Basin of northern Wyoming contains all three cycles. The Grandeur and Franson record partial inundation producing interstratified marine and terrestrial deposits (Fig. 2B; Campbell, 1962; Inden & Coalson, 1996). The basin sat landward of Meade Peak phosphorite deposition (Maughan, 1984). Increased tectonic subsidence led to greater inundation during the Ervay cycle, the focus of this study (Peterson, 1984). Ervay cycle marine deposits are phosphorite (Retort Member) and chert (Tosi Member) tongues of the Phosphoria Formation overlain by carbonates of the Ervay Member (Fig. 2B; Inden & Coalson, 1996; Matheson & Frank, 2020).

Age of the Phosphoria Rock Complex

Two published age models exist for the PRC (Fig. 3). The first, deposition from the latest Kungurian

through to the end Wordian, was developed using conodont and brachiopod biostratigraphy (Yochelson, 1968; Behnken *et al.*, 1986; Wardlaw, 1979, 2015; Wardlaw & Collinson, 1984, 1986). Conodonts and brachiopods are utilized because fusulinids, corals and trilobites are largely or completely absent despite being common in other Permian deposits, potentially the result of cool, upwelled water (Duncan, 1961; Yochelson, 1968; Wardlaw, 2015). This age model is supported by unpublished lime-micrite U-Pb ages and $^{87}\text{Sr}/^{86}\text{Sr}$ values from the Ervay Member in western Wyoming (Fig. 3A; Pommer, 2017, 2018) and agrees with the age of underlying strata and earlier work using cephalopods, fish fauna and bivalves (Miller & Cline, 1934; Branson & Branson, 1941; Miller *et al.*, 1957; Ciriacks, 1963; Morgan, 1980; Mytton *et al.*, 1983). In this scheme, the Retort Member accumulated over 1.3 Myr in the Wordian and the Tosi and Ervay members were deposited within the following *ca* 1 Myr (Fig. 3A; Wardlaw, 2015; Pommer, 2018).

Comparatively, Davydov *et al.* (2018b) combined a U-Pb date from the first known and datable ash in the PRC (Meade Peak Member, south Idaho) with a revised biostratigraphic interpretation to propose that PRC sedimentation continued up until the Permo–Triassic boundary (Fig. 3B). Continuous Late Permian sedimentation was also suggested by Saltzman & Sedlacek (2013) for the Park City Group to the south (Fig. 1C), although correlation between the two is difficult (Maughan, 1984; Wardlaw & Collinson, 1986). In the Davydov *et al.* (2018b) study area, the Franson, Retort, Tosi and Ervay are absent (far left side of Fig. 2A; McKelvey *et al.*, 1959; Mytton *et al.*, 1983; Maughan, 1984). As a result, regional correlation of the Meade Peak Member is used to suggest a late Wuchiapingian to Changhsingian age for the Ervay cycle (Fig. 3B).

The model of Davydov *et al.* (2018b) suggests a relatively conformable contact between the PRC and overlying Triassic Dinwoody Formation (Clark & Carr, 1984; Paull & Paull, 1987; Hoffman *et al.*, 2013). Description and interpretation of the PRC–Dinwoody contact are ambiguous in supporting one model over the other. The contact is generally poorly exposed, sharp, featureless and contains only minor local evidence for a disconformity (small-scale relief, reworked clasts, infilled fractures and minor pedogenesis), leading many to question the magnitude of any disconformity (Moritz, 1951; Kummel, 1954; Cressman, 1955; Cressman & Swanson, 1964; Boyd & Maughan, 1972; Schock *et al.*, 1981; Peterson, 1984; Paull & Paull, 1986a,b; Paull *et al.*, 1989). However, it has also been interpreted as recording significant regional post-PRC, pre-Dinwoody (pre-Triassic) erosion that led

to absence of the Retort, Tosi and Ervay members in Idaho [the study area of Davydov *et al.* (2018b); Fig. 2A] (McKee & McKee, 1967; Schock *et al.*, 1981; Maughan, 1984; Peterson, 1984; Paull & Paull, 1986b; Paull *et al.*, 1989; Matheson, 2019). Alternatively, lateral gradation from the Tosi to the Retort or Cherty Shale members, seaward pinch out, or some combination thereof might explain the absence of the Ervay cycle westward (Sheldon, 1963; Schock *et al.*, 1981; Peterson, 1984).

The Ervay cycle is thus broadly constrained to within a 14 Myr interval in the Middle to Late Permian and its relation to major biotic, climatic and oceanographic events remains poorly constrained. Future work is needed to elucidate whether issues with biostratigraphic zonation schemes, widespread diachroneity, incorrect regional lithostratigraphic correlation, geographic variability in the duration of sedimentation, analytical error, or other uninvestigated explanations are the cause of the disagreement and whether one model is solely correct, or both have elements of validity. Neither model is definitively favoured in this study as the exact age of the PRC does not strongly affect its oceanographic interpretation.

METHODOLOGY

Twenty-eight outcrops, 31 cores and wireline logs from 178 non-cored wells were used to constrain the PRC in the Bighorn Basin (Fig. 4; Appendix S1). Cores were accessed at the USGS Core Research Center in Denver, CO and wireline petrophysical logs were retrieved from the Wyoming Oil & Gas Conservation Commission (2018).

Lithology, sedimentary structures, fossil content, bioturbation index (BI; Taylor & Goldring, 1993) and stratigraphic architecture were recorded with over 600 samples collected and *ca* 500 thin sections examined, primarily from the Franson Member and Ervay cycle. Normal petrographic techniques on a Nikon Eclipse E400 POL microscope (Nikon Corporation, Tokyo, Japan) included volumetric estimations via visual comparison, white card petrography (Folk, 1987) and carbonate staining (Dickinson, 1965, 1966). Cold-cathode cathodoluminescence (CL) and energy dispersive X-ray spectrometry were carried out using a CITL 8200 Mk 5 (Cambridge Image Technology Limited, Hatfield, UK) mounted on a polarized light microscope using an acceleration voltage of approximately 15 keV and a beam current of 280 μ A (Vortisch *et al.*, 2003).

Bulk elemental compositions for all samples were determined using a Bruker TRACER 5i portable X-ray fluorescence spectrometer (Bruker Corporation, Billerica, MA, USA). Limestone samples were analyzed with no filter at 15 kV and 15 μ A and a count time of 30 s. Mudrocks and bioelemental sediments were analyzed under low-energy and under high-energy conditions with different filters to target a series of overlapping $K\alpha$ ranges and enable analysis of relatively light and heavy elements. The elements Al to Fe on the periodic table were analyzed under low energy conditions as described above for limestones. High-energy analyses were undertaken at 50 kV and a count time of 45 s, with a Cu 75 μ m:Ti 25 μ m:Al 200 μ m filter. In general, the raw count rate/s (rcps) signals lay between 20,000 and 100,000 rcps.

X-ray powder diffraction (XRD) analysis of 15 select samples (eight from the Ervay cycle) were performed to confirm mineralogical compositions (Appendix S2). Clay analyses on fine-grained samples were carried out after acidification to remove carbonates and glycolation. Samples were analyzed on a Rigaku Ultima III diffractometer (Bragg-Brentano geometry) (Rigaku Corporation, Tokyo, Japan) at Texas Tech University using a 0.02° 2 θ step size at 6 seconds per step, 45 kV and 40 mA. Results were used in conjunction with 64 XRD results from 25 cores, available from the USGS Core Research Center (2018).

FACIES ASSOCIATIONS AND INTERPRETATIONS

This manuscript focuses solely on the Ervay cycle (Retort, Tosi and Ervay members). The uppermost portion of the underlying Franson cycle is included to discuss the transition between the cycles.

Within this interval, 12 lithofacies in four facies associations are recognized based on macrofacies and microfacies analyses, consistent with previous studies (Table 1; Maughan, 1984; Peterson, 1984; Inden & Coalson, 1996; Pommer & Sarg, 2019; Matheson & Frank, 2020).

Facies Association 1 – Phosphorites

Facies Association 1 contains granular phosphorites and related sediments (Fig. 5A to C). FA1-1 is true phosphorite (i.e. >18 wt. % P_2O_5 ; Glenn *et al.*, 1994b), ranging from phosclast packstone with

dolomicritic matrix to phosclast grainstone (*sensu* Trappe, 2001) with subordinate sandy inarticulate (chitinophosphate) brachiopod pack/grainstone. Phosclasts (phosphatic grains) are: (i) peloids and coated grains; (ii) subangular to well-rounded intraclasts containing siliceous sponge spicules, detrital quartz silt/sand and other phosclasts; and (iii) inarticulate brachiopod shells (Fig. 5D and E). Phosphatic grains contain disseminated irregular to cubic micron-scale pyrite and trace to >20% detrital quartz sand/silt (Fig. 5). The only calcitic skeletal grains are minor (<5%) heavily fragmented to articulated brachiopod shells (Fig. 5B, D and E). FA1-2 is phosphatic dolowackestone (Fig. 5C). Both facies are poorly sorted and structureless with minor normal grading.

Interpretation.

Sedimentary phosphate (i.e. francolite) peloids, coated grains, nodules and intraclasts are the product of authigenic phosphogenesis from phosphorous-enriched pore fluids in the shallow-marine subsurface (Glenn *et al.*, 1994b; Pufahl & Grimm, 2003). Granular phosphorites, such as those in FA1, are the product of reworking of such authigenic phosphate, expanded on below (Föllmi, 1996).

Facies Association 2 – Cherty facies

Facies Association 2 is composed of high-density black, grey or purple chert nodules within spiculitic host-rock (Fig. 6); the term ‘host rock’ refers to the rock surrounding nodules. These facies are described in greater detail in Matheson & Frank (2020).

Facies 2-1 – Spiculitic black chert

FA2-1 contains black chert nodules in dolomicrite or lime-micrite, or sandstone. Nodules range from oblate spheroids in pyritic laminated host rock (Fig. 6A and B) to irregular bulbous nodules replacing burrow networks, predominantly *Thalassinoides*, in structureless host rock (Fig. 6C and D; Matheson & Frank, 2020). Both definitively silicified trace fossils and trace-like chert nodules are present (*sensu* Wistort *et al.*, 2019). Nodules and host rock contain abundant monaxon siliceous sponge spicules (Fig. 6E). Calcareous skeletal grains are largely absent. In a few instances, brachiopods, bryozoans and pelmatozoans comprise <25% of the rock.

Interpretation.

The abundant sponge spicules and chert nodules (remobilized silica) and lack of calcareous skeletal grains strongly suggest accumulation in offshore siliceous sponge meadows (Matheson & Frank, 2020). Evidence suggests environments with variable sediment oxygenation, ranging from anoxic, which led to laminated pyritic deposits, to more oxygenated leading to the bioturbated, burrow-mottled facies.

Facies 2-2 – Grey and purple spiculitic and silicified-evaporite-nodule-bearing cherty deposits

FA2-2 contains grey and purple chert nodules (nodule density is lower than FA2-1) with siliceous sponge spicules. Host rock is predominantly dolomite, with lesser dolomitic shale containing iron oxides and detrital micas in eastern exposures. Bioturbation ranges from BI = 5 with chaotic nodules (Fig. 6F) to BI = 2 with tubular chert nodules (Fig. 6G). In the terminology of Wistort *et al.* (2019), these nodules are both silicified trace fossils (based on spreiten) and trace-like chert nodules (Matheson & Frank, 2020). Sedimentary structures include flat and microbial laminations as well as contorted laminations around (now silicified) enterolithic anhydrite. Rare <10 cm dolomite beds with chert nodules preserve spicule-free silicified molluscan–oolitic grainstone. Apart from these beds, calcareous skeletal grains are non-existent.

Interpretation.

FA2-2 accumulated in shallow-marine to peritidal environments. Bioturbated spiculitic dolomicrites accumulated in well-oxygenated coastal sponge meadows equivalent to those in FA2-1; this is supported by lithostratigraphic relations. Tidal flats bounding the subtidal meadows contained intertidal microbial mats and supratidal nodular anhydrite. Quartz and clays from terrestrial red beds (FA4) mixed with marine sediment in supratidal to terrestrial dolomitic shale. Sponge spicules were likely washed from subtidal meadows onto tidal flats by tides or storms, resulting in spicules and chert in all peritidal deposits (Matheson & Frank, 2020).

Facies 2-3 – Spiculitic fossiliferous to barren carbonates

FA2-3 also contains abundant spiculitic black, grey and purple chert nodules (Fig. 6H to L). The primary difference from FA2-1/2 is the host rock, which are brachiopod–bryozoan facies with authigenic glauconite and phosphate (cf. FA3-1) to microbial and peloidal dolomudstones (cf. FA3-4). Calcareous grains are abundant within chert nodules where also present in the host rock; this is not the case in FA2-1/2.

Interpretation.

This facies accumulated in sponge meadows with a more equal mix of calcareous and siliceous benthos. This is consistent with their stratigraphic occurrence, which is discussed below.

Facies Association 3 – Carbonate facies

FA3-1 – Brachiopod–bryozoan–pelmatozoan carbonates with authigenic minerals

FA3-1 is brachiopod–bryozoan–pelmatozoan floatstone, packstone and wackestone with lesser rudstone (Fig. 7A to C). Subordinate components include siliceous sponge spicules, small phosphatic intraclasts and subangular quartz silt/sand (Table 1; Fig. 7C). Authigenic phosphate, glauconite, silica and iron oxides are pervasive within intraskeletal porosity, having replaced skeletal elements, or both (Fig. 7A to C). Chert nodules replacing carbonates (i.e. containing calcareous grains) are minor but widespread (Fig. 7C), making FA3-1 transitional with FA2-3 (cf. Fig. 6H and I). FA3-1 beds are pervasively burrow mottled and lack sedimentary structures.

Interpretation.

The lack of photozoan skeletal components, prevalence of filter feeders and authigenic phosphate indicate a heterozoan carbonate benthos on a mesotrophic, well-oxygenated seafloor with subordinate siliceous sponges. The bioturbation and minor fragmentation of bioclasts suggest deposition below fair-weather and potentially storm-wave base.

The widespread authigenic phosphate and glauconite indicate variable redox conditions, at least locally (Glenn & Arthur, 1988; O'Brien *et al.*, 1990). Glauconite forms from suboxic pore water during the partial microbial reduction of iron-(oxyhydr)oxides (Harder, 1980; Odin & Fullagar, 1988) and phosphate indicates elevated pore water P, either due to organic matter degradation or desorption from iron-(oxyhydr)oxides. That authigenic cements are largely restricted to bioclasts is likely the product of microbially mediated organic-matter decomposition within skeletons that produced micro-reducing areas facilitating the precipitation of the redox-sensitive minerals (Harder, 1980; Odin & Fullagar, 1988).

FA3-2 – Brachiopod–bryozoan–molluscan carbonates lacking authigenic minerals

FA3-2 is transitional between FA3-1 and FA3-3 (molluscan carbonates) with a mixture of bryozoans, brachiopods, pelmatozoans and molluscs (Fig. 7D to F). Cementation by authigenic phosphate and glauconite is less common and limited to the smallest skeletal grains.

Interpretation.

This facies accumulated in more nutrient-poor water landward of FA3-1. That the skeletal assemblage contains a combination of the fauna found in FA3-1 and FA3-3 indicates a transitional environment between the two.

FA3-3 – Molluscan carbonates

FA3-3 is molluscan-peloid grainstone, rudstone and packstone with generally diminutive (<2 cm) bivalves; lesser gastropods, scaphopods and cephalopods (Fig. 7G and H); and a mixture of minor allochems including rip-up clasts of FA3-4 (Table 1; Fig. 7H). Micrite envelopes, cortoids and isopachous cement are common. Beds are often burrow mottled although some contain planar tabular or trough cross-bedding and heavily abraded bioclasts.

Interpretation.

FA3-4 records a molluscan carbonate factory with aragonitic skeletons that underwent dissolution and/or alteration, at least some of which (for example, micritization) occurred in the marine realm. Pisolitic intraclasts and cross-bedding indicate shallow, high-energy environments, interpreted as shorefaces on island belts nucleated on antecedent topography (Simmons & Scholle, 1990). Structureless beds suggest accumulation extended into sheltered or sub-fair-weather wave-base settings.

The absence of authigenic glauconite and phosphate cementation and presence of aragonitic bioclasts suggest warmer and higher pH oligotrophic water with a greater carbonate saturation state. The small size of molluscs is interpreted as an indicator of elevated salinity, as is supported by palaeoecological studies of these facies (Wilkerson, 1967; Yochelson, 1968; Roberts, 1982).

FA3-4 – Pisoidal, peloidal and microbial carbonates

FA3-4 contains non-skeletal carbonates (i.e. microbial and abiogenic) with ubiquitous fabric-retentive microcrystalline dolomitization (Herrod, 1980; Coalson & Inden, 1990; Inden & Coalson, 1996). Most common is peloid and coated grain dolograinstone (Table 1). Beds contain abundant fenestral porosity (Fig. 7I and J; Tebbutt *et al.*, 1965) and lesser endolithic borings, cavities with geopetal sediment, bottom growth marine cements, teepee structures, mud-filled sheet fractures, angular rip-up clasts and near-vertical protrusions on internal erosion surfaces. Coated grains are 0.25 mm to 2.0 cm in size and despite not all being larger than 2 mm (Peryt, 1983) are referred to as pisoids as they lack identifiable nuclei and have irregular cortical laminations and shapes. Broken, recoated and composite grains are common (Fig. 7K). Beds are structureless and poorly sorted, contain rare reverse grading, and often alternate with microbial bindstone with minor birds-eye voids. Subordinate depositional textures include solution collapse breccias and barren dolomudstones with wispy incipient microbial lamination.

Interpretation.

The presence of birds-eyes, microbial lamination, laminoid fenestral fabric, small spired gastropods and teepee structures are all indicative of restricted peritidal deposition (Scholle & Kinsman, 1974;

Esteban & Pray, 1983; Shinn, 1983). Wispy laminated dolomudstones accumulated in shallow restricted subtidal or lagoonal environments, whereas pisoids formed autochthonously in intertidal to supratidal settings with elevated salinity. Borings, geopetal sediment, rip-up clasts and related features are suggestive of early (i.e. marine and meteoric) cementation. Broken pisoids, sheet fractures, mudcracks, teepees, internal erosion surfaces and rip-up clasts are evidence for widespread and repeated exposure, desiccation and reworking. The fabric-retentive dolomitization is consistent with evaporative reflux in peritidal environments (Warren, 2000).

Facies Association 4 – Terrestrial deposits

Facies Association 4 contains intercalated red beds (FA4-1), gypsum/anhydrite (FA4-2) and gypsiferous dolostone (FA4-3) of the Goose Egg Formation (Fig. 8). Red beds (iron-rich quartz siltstone and claystone; FA4-1) are mostly structureless, with minor pedogenic alteration and flat to contorted laminations and anhydrite nodules (Fig. 8A to D; Matheson, 2019). Gypsum (FA4-2; Fig. 8E) and dolostone (FA4-3) beds contain chickenwire fabrics, boxworks, solution collapse breccias and microbial laminations.

Interpretation.

The red beds are interpreted as forming from aeolian dust (loess) in the western Pangaeian desert (Fig. 1). Periods of low sea-level produced thick red bed successions, whereas interbedded gypsum and red beds formed during high sea-level, with gypsum and dolostone beds (FA4-2/3) accumulating in hypersaline lagoons and broad, landward supratidal to terrestrial coastal salinas (Inden & Coalson, 1996; Matheson, 2019). This interpretation of a semi-arid to arid landscape of windblown dust dotted with salinas is consistent with similar Carboniferous–Triassic deposits that stretch across the US midcontinent (Figs 1, 2 and 4; e.g. Walker, 1967; Benison & Goldstein, 2000; Sweet *et al.*, 2013; Zambito & Benison, 2013).

FACIES DISTRIBUTIONS – WALTHERIAN VERSUS NON-WALTHERIAN DEPOSITION

Reconstructing the oceanographic controls on the genesis of the Ervay cycle phosphorites (FA1) and cherts (FA2) requires understanding of the spatiotemporal relationships between the facies associations. Reference to the facies associations can be simplified because each of the three marine lithostratigraphic members is almost exclusively composed of one of the three marine facies associations (Table 2): the Retort Member is mostly FA1 (phosphorites); the Tosi Member is mostly composed of FA2 (cherty deposits); and the Ervay Member is predominantly FA3 (carbonates). Terrestrial facies (FA4) form landward equivalents to all three (Fig. 9).

The exact nature of the chronostratigraphic relationships of the marine members is less obvious because biostratigraphic constraints are coarse and absolute age dates are lacking (e.g. Wardlaw & Collinson, 1986). Two end-member options are possible: (i) the Retort, Tosi and Ervay interfinger and thus accumulated adjacent to one another on the seafloor ('Waltherian' deposition); or (ii) the three units are vertically stacked and accumulated successively on the seafloor (i.e. each represents a time-specific stratigraphic unit).

Among studies that have focused on the PRC in the Bighorn Basin, there is a consensus that interfingering of the Retort, Tosi and Ervay members is limited and that the three accumulated largely in a 'layer-cake' package that indicates successive deposition where each member occupied the entirety of the flooded portion of the Bighorn Basin for a discrete period of time (McCue, 1953; Campbell, 1962; Cole, 1970; Simmons & Scholle, 1990; Ulmer-Scholle & Scholle, 1994; Inden & Coalson, 1996; Matheson & Frank, 2020). The interpretation of successive deposition is supported by stratigraphic architecture, basin-wide correlations, coast trajectories and marker-bed correlation presented herein (Figs 9 to 12).

Each of the three members is readily identified in outcrop, core and on wireline logs due to their unique compositions. The Retort, a prominent peak on gamma-ray logs, is the most aerally restricted unit (Fig. 10A). Both the Tosi and Ervay are found across the extent of the Bighorn Basin with the contact between the two well-exposed (Figs 9, 10B, 10C, 11 and 12A to E). Everywhere they are found, the Ervay overlies the Tosi with little intergradation at the contact (for example, Fig. 12D and E), with the exception of in the south-west (for example, Owl Creek; Fig. 12B). Tongues of both the Ervay and Tosi also extend into terrestrial red beds to the east (Fig. 9). These relationships strongly indicate that the Tosi accumulated across the entire Bighorn Basin with no coeval landward Ervay

Member carbonates (Figs 9 and 10B; Beatty, 1957; Lane, 1973; Simmons & Scholle, 1990; Ulmer-Scholle & Scholle, 1994; Inden & Coalson, 1996; Matheson & Frank, 2020).

The successive nature of the units is further supported by regionally correlatable lower-order sequences and parasequences within carbonates of both the Franson and Ervay members (Coalson & Inden, 1990; Clark, 1994; Inden & Coalson, 1996; this study). These cycles are associated with shifts from glauconite-rich, phosphate-rich and chert-rich carbonates atop flooding surfaces (FA2-3 and FA3-1) upward into chert-poor carbonates (FA3-2/3/4; Inden & Coalson, 1996). The basal portion of each cycle is readily recognizable on wireline logs and in core and outcrop because each has a distinct appearance. As such, these are robust marker beds. Two marker beds in the Ervay and one in the Franson were correlated across the Bighorn Basin in this study (green dashed lines, Fig. 9). They are easily correlated in marine strata and are less prominent but sometimes identifiable in peritidal facies (Fig. 11). Their distinctness makes them useful chronostratigraphic surfaces (i.e. timelines). The fact that these timelines are found wholly within the carbonate members with relatively little stratigraphic variation in the thickness of the cycles, especially the lowermost cycle in the Ervay Member (Fig. 9), is strong supporting evidence for successive deposition of the three members, particularly the Tosi and Ervay.

To summarize, stratigraphic relationships indicate that the Retort, Tosi and Ervay accumulated as chronostratigraphically distinct (i.e. conformably stacked) units across the Bighorn Basin. As such, their distributions do not follow Walther's Law, which is not unheard of in biochemical sediments (e.g. Brett *et al.*, 2012). This interpretation that each member is chronostratigraphically distinct is not meant to be absolute. There is likely some minor diachroneity both along the contacts between (Fig. 12B; discussed below) and within the members (Figs 11 and 12C), but this is superimposed on the first-order trend of successive modes of sediment production. Additionally, the magnitude of any diachroneity is interpreted to be subordinate to the duration of time over which each member occupied the entire seafloor.

The interpretation herein is not meant to extend beyond the Ervay cycle in the Bighorn Basin. Despite the widespread agreement within the Bighorn Basin, some workers, especially those working at the regional scale, have suggested that the Tosi Member cherts accumulated offshore from coeval Ervay Member carbonates (e.g. Maughan, 1984; Peterson, 1984; Wardlaw, 2015). This differing

interpretation stems from four issues. First, the Ervay cycle stratigraphic relationships might differ outside of the Bighorn Basin where there is contention regarding any interfingering (for example, in the Wind River Basin; cf. Ahlstrand & Peterson, 1978; Rath, 1981). Second, the majority of work in the PRC has focused on the underlying Franson cycle where chert (Rex Chert Member) and carbonate facies (Franson Member) do interfinger, having accumulated laterally adjacent to one another (Fig. 2; e.g. Peterson, 1984; Hendrix & Byers, 2000). Some studies have overlooked the differences between the two cycles. Third, the scale of the area being evaluated is important. Whereas successive stacking is present within the Bighorn Basin (*ca* 150+ km of depositional strike), the same might not be true at the scale of the entire Phosphoria Sea (>500 km), although the lithostratigraphic relations further west remain contentious (cf. Fig. 2A; Sheldon, 1963; Maughan, 1984; Peterson, 1984). Fourth, the Tosi is relatively poorly exposed in the eastern Bighorn Basin. This has largely masked the extent and range of biosiliceous facies that strongly support the layer-cake model (Fig. 9; Matheson & Frank, 2020).

With the understanding that the marine members of the Ervay cycle in the Bighorn Basin represent successive units, a number of details of the stratigraphic stacking are relevant (Table 2). First, the Retort Member is underlain by the Franson Member, which is composed of fossiliferous to peritidal carbonates of FA3 with minor chert-rich carbonates of FA2-3. The Retort Member is a thin (<1.5 m) tongue that is predominantly phosphorite (FA1) with minor chert of FA2-1. The Retort has a sharp base with minimal evidence for significant erosion (Fig. 12A and C). The contact between the Retort and overlying Tosi is either sharp or gradational over a few decimetres, and everywhere appears conformable. The Tosi directly overlies the Franson at a sharp contact beyond the Retort pinch out (Figs 9, 10 and 12E). Atop this contact, the Tosi is cherty deposits of FA2-1 with minor phosphorites in the west and FA2-2 in the east (Figs 9 and 12). In the south-west (basinward), the top of the Tosi is transitional to interbedded chert and carbonates of the Ervay Member (Fig. 12B), whereas at landward sites the Tosi–Ervay contact is sharp and abrupt (Fig. 12D and E). The Ervay is fossiliferous to peritidal carbonates (FA3) with a few chert-rich carbonate intervals (FA2-3). In the south-east corner of the study area, the Tosi is overlain by FA4 red beds and gypsum (Fig. 9; Matheson & Frank, 2020).

DEPOSITIONAL MODELS

Each marine member is predominantly composed of a single facies association and each member accumulated successively. This means that three separate depositional models are required to explain the nature of sedimentation at different points in the Ervay cycle (Fig. 13).

Facies Association 1 – Authigenic phosphogenesis – Retort Member

The Retort Member formed via the two-step process common to all Phanerozoic granular phosphorites (Glenn *et al.*, 1994a; Pufahl & Groat, 2017). First, pristine phosphorites formed during a period of sediment starvation in which elevated primary productivity led to aerobic respiration of organic matter in the water column depleting dissolved oxygen (Glenn *et al.*, 1994a,b; Filippelli, 2008). Suboxic to anoxic bottom water led to sedimentary condensation and organic matter accumulation (Diaz & Rosenberg, 1995; Föllmi, 2016). Bacterial decomposition of organic matter in reducing conditions within a few decimetres of the seafloor liberated phosphorous, elevated pore water P, and led to the precipitation of authigenic phosphate in laminated (pristine) phosphorites or as nodules and authigenic grains (Froelich *et al.*, 1988; Glenn & Arthur, 1988; Jarvis *et al.*, 1994; Pufahl & Grimm, 2003; Crosby & Bailey, 2012). The pristine phosphorites were subsequently eroded and reworked, producing granular phosphorites (Föllmi, 1996). The pyrite, phosphatic intraclasts and minimal bioclasts support this sequence of events (Fig. 5). Intraclasts indicate multiple generations of phosphogenesis and reworking during this period, likely by storms, although this is masked by the beds' structureless nature (cf. Hiatt & Budd, 2003). True phosphorites (FA1-1) in the western basin are likely autochthonous (Fig. 10A; cf. Hendrix & Byers, 2000; Hiatt & Budd, 2001). Some phosphatic clasts were transported landward from this area forming dolowackestones (FA1-2). The conditions that led to the eutrophication, sediment starvation and dysoxia or anoxia necessary for phosphogenesis are discussed below.

Facies Association 2 – Glass ramp – Tosi Member

The prevalence of spicules over calcareous skeletal fragments in the Tosi Member indicates areas of high-density siliceous sponges, variably termed sponge meadows, grounds or gardens (Hogg *et al.*, 2010; Ritterbush, 2019). Remobilization of spicule silica during burial, potentially beginning within a

few metres of the seafloor, led to the precipitation of chert nodules (Fig. 6; Ulmer-Scholle & Scholle, 1994). Nodule shape is a product of heterogeneity caused by bioturbation, whereas colour (black, grey or purple) is a product of redox conditions during chertification (Matheson & Frank, 2020).

Sponge meadows occurred across a variety of environments ranging from low to moderate-energy offshore to low-energy coastal settings rimmed by evaporative tidal flats (Fig. 13). These deposits thus formed across a broad glass ramp (i.e. a neritic biosiliceous factory), albeit one with a low-gradient and non-‘typical’ (in the sense of traditional carbonate ramps) hydrodynamic subdivisions (i.e. low-energy nearshore) due to its epicontinental location. As a result of these differences, Matheson & Frank (2020) refer to the Tosi as an epeiric glass ramp, although this nomenclature is simply used to denote variance from other more ‘traditional’ glass ramps (for example, Permian examples in Arctic Canada and the Barents Sea; Gates *et al.*, 2004; Beauchamp *et al.*, 2009; Blomeier *et al.*, 2013; Ritterbush, 2019).

In deeper areas, mesotrophic conditions and moderate organic matter rainout generated anoxic pore water which inhibited infaunal colonization, led to sulphate reduction, the precipitation of framboidal pyrite, and the preservation of laminated deposits with oblate spheroid black chert nodules (Figs 6A, 6B and 13). Offshore areas that were better ventilated, and thus more pervasively bioturbated (for example, Fig. 6C and D), likely related to impingement of waves. The attenuation of wave energy across the nearly >150 km width of the epicontinental sea that included the Bighorn Basin clearly led to the lower energy nearshore environments that were better oxygenated and pervasively burrow mottled (for example, Fig. 6F; Matheson & Frank, 2020). Despite the prevalence of coastal sponge meadows, carbonates accumulated in patchy rare and short-lived mutually exclusive areas at the same time, as indicated by their lack of spicules. These molluscan, and lesser peloidal/oolitic, carbonate grainstones might record locally elevated tidal currents, or periods of increased storm activity and might have formed in small shoals limited to the most landward portions of the epeiric glass ramp (Matheson & Frank, 2020). The available palaeoecological dataset (i.e. spicules) does not allow easy interpretation of water temperature. Waters along the coast were clearly warm as evidenced by evaporite nodules and the patchy aragonitic carbonate fauna, potentially indicating that the Tosi glass ramp had a thermocline as is evident in the overlying Ervay Member carbonates.

Facies Association 3 – Carbonate ramp – Franson and Ervay members

Facies Association 3 records a carbonate ramp (Fig. 13). Offshore, a heterozoan filter-feeding biota in cool, mesotrophic waters resulted in the precipitation of authigenic glauconite and phosphate cements (FA3-1). Landward, lower nutrient levels coupled with the influence of the hot and arid continental climate led to elevated water temperature, salinity and carbonate saturation state resulting in both aragonitic molluscan communities in shallow subtidal and shoreface settings (FA3-3) and pisoidal and microbial tidal flats (FA3-4; Simmons & Scholle, 1990; Inden & Coalson, 1996). These relationships indicate that the Ervay accumulated on a temperature stratified ramp with a thermocline below shallow warm waters heated by the surrounding desert (cf. James, 1997). The cooler water temperatures below this thermocline, potentially coupled with elevated trophic resources from upwelling, explains the prevalence of a heterozoan assemblage (Yochelson, 1968; James, 1997). The low diversity and predominance of molluscs nearshore (i.e. in warmer water) likely reflects restricted conditions controlled by elevated evaporation and lagoonal circulation that generated higher salinity waters above the thermocline (Campbell, 1962; Hite, 1978) as is indicated by palaeoecological studies of the molluscan fauna (Wilkerson, 1967; Yochelson, 1968; Roberts, 1982).

SEQUENCE STRATIGRAPHY

The stratigraphic architecture of the Ervay cycle (Figs 9 and 10) coupled with facies trends at individual sites (Figs 11 and 12) provide constraints on multiple orders of relative sea-level change through deposition. The Retort Member rests on what is widely interpreted as a sequence boundary (Peterson, 1984; Inden & Coalson, 1996; Hiatt, 1997), although the aerial extent of the Retort as compared to the Franson suggests the magnitude of relative sea-level fall was relatively low (Fig. 9). The arid climate prevented significant alteration of any subaerially exposed carbonates, potentially masking the duration and magnitude of sea-level fall.

The Retort then formed through phosphogenesis and reworking during lowstand and early transgression. This timing can be constrained because the conformably overlying Tosi contains the

most landward-extending marine deposits in the southern study area and therefore must mark the maximum transgression (Fig. 9; Lane, 1973; Matheson & Frank, 2020). Phosphorite formation and amalgamation were thus closely spaced in time in a single sea-level cycle (Pufahl *et al.*, 2003; cf. Baturin, 1971).

The Tosi then records deposition during transgression (Fig. 14). The Tosi–Ervay contact is defined by a switch of the benthic biota from biosiliceous (Tosi) to calcareous (Ervay) as part of the highest-order transition from transgression to regression in the Ervay cycle (Figs 9 and 14). This transition was more abrupt in shallow as compared to offshore areas as indicated by the sharp versus gradational contacts, respectively. The Ervay Member records carbonate sedimentation during the subsequent regression that terminated PRC deposition (Fig. 9). The Ervay Member is thickest along the eastern flank of the basin, where it is predominantly peritidal deposits (FA3-4; Fig. 10C; Inden & Coalson, 1996). This is likely due to greater sedimentation rates in peritidal deposits and is consistent with accumulation during a period of progradation, as opposed to the Tosi thickness trends, which indicate transgression (Fig. 10). Regression also explains the progressive shallowing of carbonate facies upward (Figs 11 and 14).

The interpretation of lowstand to transgressive phosphogenesis, chert accumulation during transgression and carbonates forming during regression is consistent with previous work (Peterson, 1984; Simmons & Scholle, 1990; Ulmer-Scholle & Scholle, 1994; Inden & Coalson, 1996). It must be noted that this interpretation is an over-simplification in that the tectonic controls on subsidence that led to the northward movement of the PRC depocentre between the Franson and Ervay cycles also created along-strike variability in stacking patterns that are not described here (cf. Madof *et al.*, 2016).

There is also substantial evidence within the Ervay cycle of a number of lower-order (i.e. higher-frequency, lower-amplitude) relative sea-level fluctuations (Fig. 11). The most important of these occurred at the time of the Tosi–Ervay contact as deposition of the Tosi ceased during a relative sea-level fall (Fig. 9). Regression created stranded salinas and evaporitic lagoons that led to strengthened lagoonal circulation, facilitating a period during which downwelling brines were associated with reflux dolomitization and the precipitation of anhydrite nodules in the Tosi (Hite, 1978; Ulmer-Scholle & Scholle, 1994; Matheson & Frank, 2020). The absence of identifiable lower-order Tosi highstand systems tract deposits is likely the result of low biosiliceous accumulation rates and

extensive, pervasive bioturbation (Matheson & Frank, 2020).

Intertonguing marine-terrestrial facies and offshore facies cyclicity in the Ervay Member are also suggestive of parasequences (or lower-order sequences) formed by high-frequency relative sea-level fluctuations (Figs 9 and 11; Clark, 1994; Inden & Coalson, 1996). These are the cycles that can be readily correlated as timelines across the basin since transgressive deposits overlying many flooding surfaces are cherty carbonates of FA2-3 (Fig. 6H to L; green dashed lines, Fig. 9; Inden & Coalson, 1996). As some cycles contain only subtle facies variations that might be autogenic in origin, the number of cycles identified by various authors and in individual outcrops/cores differs (Fig. 11). Similar cyclicity likely also occurred during deposition of the Retort and Tosi but is less identifiable due to sedimentary condensation and poor preservation, respectively (Matheson & Frank, 2020).

DISCUSSION

As evidenced by their stratigraphic relations, the Retort–Tosi–Ervay succession records a unique, well-exposed and well-cored, siliciclastic-sediment-starved ramp in which three disparate modes of biochemical marine sediment production occurred successively and conformably across the study area (Figs 9, 13 and 14; Inden & Coalson, 1996). The Franson–Retort contact is a sequence boundary and thus records a break in sedimentation following deposition on a carbonate ramp (Franson). Following sea-level rise, the carbonate benthos did not immediately return. Instead, the area became the locus of sediment starvation and phosphogenesis (Retort), then transitioned to an area of widespread siliceous sponge meadows (Tosi) and finally returned to being a widespread carbonate factory (Ervay). With the exception of the Franson–Retort contact, this all occurred without any significant breaks in sedimentation.

Understanding the biological and oceanographic controls on deposition of the Ervay cycle biochemical sediments is thus reliant on understanding both the oceanographic conditions under which each individual sediment type was deposited and what parameters changed so as to cause the evolution between the three sediment factories.

Oceanographic controls on deposition

Formation of the Retort Member phosphorites

Various oceanographic models have been proposed to explain PRC phosphogenesis. As previously described, the nature of phosphogenesis suggests that the most likely explanation involves oxygen limitation leading to the exclusion of a macrobenthos, sedimentary condensation, and the precipitation of authigenic phosphate minerals. The upper Franson carbonates lack significant authigenic glauconite and francolite, suggesting an increase in nutrient loading between the Franson and overlying Retort, a well-established cause of oxygen depletion (Föllmi *et al.*, 1994). As a result, enhanced primary productivity due to upwelling of nutrient-rich water has long been viewed as the most logical explanation, although eutrophication due to aeolian dust has also been suggested. Alternatively, it has been proposed that phosphogenesis was not related to elevated primary productivity (at least compared to other PRC sediments) but was simply the product of relative sediment starvation.

Upwelling versus aeolian nutrient input.

Initial oceanographic models inferred an upwelling system along western Pangaea analogous to that along the modern Peruvian shelf margin (McKelvey *et al.*, 1959; Sheldon, 1963; Brittenham, 1976; Hein *et al.*, 2004). This analogy has since been questioned because PRC phosphorites occupied a much broader range of water depths than on modern shelves and accumulated on a ramp, not a shelf (Peterson, 1984; Wardlaw & Collinson, 1986; Whalen, 1996; Hendrix & Byers, 2000; Hiatt & Budd, 2001; Hein *et al.*, 2004; Pufahl, 2010), although some authors still argue for a shelf geometry (Stephens & Carroll, 1999; Piper *et al.*, 2007).

As a result, many subsequent oceanographic models for the PRC have evolved to invoke upwelling through mechanisms that differ from the well-known modern zones of upwelling. To this end, differences among models reflect disagreement surrounding the origin of the upwelled water (Carroll *et al.*, 1998; Hiatt & Budd, 2001; Piper *et al.*, 2007), the potential role of salinity or temperature stratification (Dahl *et al.*, 1993; Stephens & Carroll, 1999; Piper & Link, 2002) and the

mechanisms controlling transport of nutrient-rich water into the shallow portions of the sea (Hite, 1978; Carroll *et al.*, 1998; Hiatt & Budd, 2003; Pufahl, 2010).

As an alternative to upwelling, Cecil *et al.* (2018) invoked aeolian nutrient input to explain phosphogenesis in the PRC. There is some support for aeolian input into the sea in angular quartz silt spread throughout marine facies (for example, Fig. 7C; Hiatt, 1997; Carroll *et al.*, 1998; Wailing, 2000, 2010; Piper, 2001; Matheson & Frank, 2020) and in modelling that indicates seasonally offshore winds (Gibbs *et al.*, 2002; Peyser & Poulsen, 2008). It is possible that nutrient fertilization from windblown sediment could have played a role in eutrophication as Si, Fe and P can all be supplied by wind. Silicon input would come from the dissolution of clays and potentially amorphous surfaces layers on aeolian quartz dust (Cecil, 2015; Cecil *et al.*, 2018). Iron would have been supplied by clays and/or (oxyhydr)oxides (Jickells *et al.*, 2005). Phosphorous could have been absorbed to iron-(oxyhydr)oxides and clays (Filippelli, 2008; Wang *et al.*, 2012; Drummond *et al.*, 2015). Some of these nutrients would have been bioavailable and promoted primary productivity as they do in the modern oceans (Benitez-Nelson, 2000; Bressac *et al.*, 2019).

However, while aeolian input undoubtedly occurred, there is also evidence arguing against it being the predominant or sole nutrient source. Upwelling-derived phosphorites are commonly associated with organic-rich fine-grained rocks and pyrite, as they require sufficient organic-matter flux to produce sediment anoxia and sulphate reduction (O'Brien & Veeh, 1981; Glenn *et al.*, 1994b; Föllmi, 1996). Both are present in the Retort (Fig. 5D and E; Claypool *et al.*, 1978; Hiatt, 1997). It is difficult to envision similar conditions occurring solely due to aeolian nutrient input as they would require a significant volume of aeolian input considering that only a small fraction of the input material would have been bioavailable.

Aeolian silt in marine deposits represents a crude proxy for aeolian input (Carroll *et al.*, 1998). Silt is not overly abundant in many of the Retort phosclasts (Fig. 5E) that presumably took extended periods of time to form and theoretically would have represented the most eutrophic conditions and therefore greatest aeolian input in this model. This argues against vast aeolian nutrient input as the sole mechanism of nutrient delivery. Relatedly, Retort organic-rich shales are thicker and contain up to 19 wt. % organic carbon in Montana (Claypool *et al.*, 1978; Maughan, 1984). Similarly, the concentration of chert nodules (i.e. the total Si) in the Tosi decreases in a shoreward direction (Figs 6

and 12; Matheson & Frank, 2020). Likewise, phosphate and glauconite cements in carbonates (FA3-1) are limited to offshore facies, despite the presence of angular silt across all coeval carbonate facies. These trends all suggest a lack of direct correlation between aeolian input (silt) and elevated nutrients and support introduction of nutrients from the west through upwelling. Furthermore, aeolian silt is present in all of the phosphorites, cherty deposits and carbonates and there is no clear systematic change between the three, as would be expected if temporally variable aeolian nutrient input controlled the nature of sedimentation. This is especially true given that the Ervay Member carbonates (which accumulated under the lowest nutrient levels, see below) contain notable silt (for example, Fig. 7C) despite sedimentation rates that were presumably greater than both the phosphorites and cherts.

Aeolian input is therefore difficult to reconcile as the sole contributor to eutrophication and phosphogenesis. This should, however, not completely rule it out; it likely played a subordinate role as upwelling and aeolian input are promoted by similar climatic conditions (Diester-Haass, 1981; Carroll *et al.*, 1998; Hiatt & Budd, 2003).

Sedimentary condensation versus eutrophication.

Although their work focuses on the Park City Group and adjacent rocks further south, Ketner (2009) and Ritterbush (2019) posited that the PRC phosphorites (and glass ramps) reflect condensation without upwelling (cf. Hein *et al.*, 2004). This interpretation is partly based on the presumption that the wide Phosphoria Sea (Fig. 1) would have led to the attenuation of the nutrient content of upwelled water, and thus upwelling could not explain the epicontinental phosphorites. However, some upwelling models have argued that nutrient attenuation might not have been an issue, either due to seasonal wind-driven Ekman transport leading to upwelling from an intermediate water mass (Hiatt & Budd, 2003) or lagoonal circulation with onshore-directed, nutrient-rich surface currents created by downwelling coastal brines (Hite, 1978; Pufahl, 2010; Pufahl & Groat, 2017; see also O'Brien & Veeh, 1981; Glenn *et al.*, 1994a,b).

Regardless, if the entire PRC succession in the Bighorn Basin suggested sediment starvation (for example, was organic-rich muds, phosphorites and hardgrounds), a story related to sedimentary

condensation might be tenable for explaining sediment factory switching in the Ervay cycle. However, the Retort Member accumulated during a period of sea-level rise inhospitable to most benthos but is encased by deposits (i.e. the Franson and Tosi members) formed when conditions supported benthic macrofauna (Fig. 14). Invoking sediment starvation in isolation (i.e. without a cause) during rising sea-level is not a sufficient explanation for these differences. Since sedimentation was predominantly biogenic (i.e. *in situ*), sediment starvation would have required an oceanographic change that explains why the Franson calcareous biota did not recolonize the seafloor during sea-level rise (Föllmi, 2016). Clearly, the onset of Retort phosphogenesis was the result of a change in seawater chemistry that led to or was associated with sediment starvation.

Summary of phosphogenesis.

In summary, upwelling, even if by different physical oceanographic mechanisms than observed today, provides the most parsimonious explanation for the oxygen limitation that led to sedimentary condensation and phosphogenesis. The duration, intensity and origin of upwelled water remain difficult to constrain, especially in landward settings such as the Bighorn Basin. However, to use this as an argument against upwelling belies the lack of consensus on Phosphoria Sea water depths (Hein *et al.*, 2004), that the water masses responsible for modern upwelling systems are complex, and that intermediate water masses can serve as a source of upwelled nutrients from only tens of metres water depth (Hiatt & Budd, 2003; Rae, 2005). Furthermore, it ignores the potential for nutrient recycling, which may have been significant in the epicontinental and partially restricted Phosphoria Sea, partially negating nutrient attenuation.

Tosi glass ramp development

It logically follows that the switch from phosphogenesis (Retort Member) to biogenic sedimentation (Tosi Member) reflected a diminishment of the impact of upwelling (or aeolian input) and the related oxygen limitation (Hiatt & Budd, 2003). The next question is what then was the control on the siliceous sponge (Tosi) versus calcareous (Ervay) biota?

Global lysocline shoaling?

In part due to their relatively recent recognition, there is little consensus regarding the controls on glass-ramp formation. Beauchamp & Grasby (2012) studied Permian glass ramps in the Sverdrup Basin, which lay in the northern reaches of the western Pangaeian chert belt. These authors postulated that increased atmospheric CO₂ coupled with muted silicate weathering led to shoaling of the global lysocline, which was amplified by upwelling (cf. Hönisch *et al.*, 2012). Progressive lysocline shoaling and landward migration of low-pH waters led to a diminishment of the calcareous benthos and the progressive expansion of sponge meadows into coastal seas. This occurred from the Carboniferous through the end Permian with a temporary reversal during the Wordian (Fig. 3C; Beauchamp & Desrochers, 1997; Beauchamp & Grasby, 2012).

The Tosi glass ramp is not consistent with the lysocline shoaling model, regardless of which PRC age model is used. If it is Wordian, the Tosi glass ramp formed at the same time as sponges retreated to deeper environments in the Sverdrup (Fig. 3A, C and D) and atmospheric CO₂ decreased (Birgenheier *et al.*, 2010). Comparatively, if it is Late Permian in age (for example, Changhsingian as in Davydov *et al.*, 2018b), the overlying Ervay Member, which records a clear shift back to (still Permian) carbonates, would record a clear trend from biosiliceous to biocalcareous deposition towards the Permo–Triassic Boundary (PTB), which is at odds with the lysocline model of progressive eradication of shallow carbonates up to the PTB (Fig. 3B, C and E). Formation of the Tosi in the Bighorn Basin thus records less of a secular trend than the glass ramps in the Sverdrup Basin and more of an episodic or discrete biosiliceous ‘event’. While shoaling of the global lysocline might have set the stage for this event, a separate explanation is required for the Tosi, one which relies on a non-secular change in the Phosphoria Sea.

Dissolved silica?

Dissolved silica (silicic acid; DSi) availability in the oceans is an important limiting parameter in modern siliceous sponge growth and is elevated in many modern sponge-dominated habitats

(Maldonado *et al.*, 1999; Conway *et al.*, 2005; Hogg *et al.*, 2010; López-Acosta *et al.*, 2018a,b). As a result, elevated DSi might favour (or be required for?) glass-ramp development (Maldonado *et al.*, 2011; Ritterbush, 2019). Diatoms have played a large part in the DSi deficiency since the Cretaceous. As such, Palaeozoic oceans likely had higher DSi, which might have created conditions more favourable for siliceous sponges (Maldonado *et al.*, 1999; Racki & Cordey, 2000; Planavsky *et al.*, 2010; López-Acosta *et al.*, 2018a). However, DSi alone cannot explain the uneven distribution of pre-Cretaceous glass ramps (Ritterbush, 2019), nor does it provide a mechanism to explain the limited temporal extent of the Tosi.

Locally (Phosphoria Sea) or regionally (north-west Pangaea) elevated DSi concentrations provide a more episodic and viable explanation for the Tosi glass ramp. Such a phenomenon is conceptually plausible as upwelled water often contains elevated DSi (Nelson *et al.*, 1995; Tréguer & De La Rocha, 2013). Post-Mesozoic upwelling systems are associated with abundant pelagic DSi utilization (i.e. diatom blooms) for this reason (Suess & Thiede, 1981; Behl & Garrison, 1994). On top of this, the peri-Laurentian arcs outboard of the Phosphoria Sea may have prevented significant outflow of deep water from the sea (cf. Fig. 1; Hein *et al.*, 2004; Ketner, 2009). As such, the Phosphoria Sea may have been a major transient reservoir and/or sink in the Permian Si cycle (Tréguer & De La Rocha, 2013; Ritterbush, 2019).

Ervay cycle oceanographic model

The aforementioned discussion and constraints imposed by the facies support an oceanographic model for the Ervay cycle in which progressively decreasing upwelling intensity and related oceanographic parameters (including sea-level) caused transitions between the three modes of sediment production (Figs 14 and 15). Despite the fact that upwelling is favoured as the nutrient delivery mechanism in this study, the following oceanographic model is equally viable if the nutrients were instead mostly supplied by aeolian input.

Intense upwelling and phosphogenesis

Upwelling, potentially supplemented by aeolian input, provided elevated nutrients as sea-level rose at the onset of the Ervay cycle. Elevated primary productivity led to the development of an oxygen minimum zone, the exclusion of a calcareous benthos as had been present during the Franson, and authigenic phosphogenesis in a narrow subsurface zone stabilized by sediment starvation (Glenn & Arthur, 1988; Föllmi, 1996; Pufahl & Groat, 2017). Granular phosphorites were produced by episodic reworking of these deposits. Whether upwelling eutrophic waters were cool, as is typical of upwelled deep water, or were warmer (Stephens & Carroll, 1999; Hiatt & Budd, 2001, 2003) is unclear based on the results of this study alone but is relatively unimportant in the oceanographic model proposed here.

Decreasing upwelling intensity and glass ramps

Upwelling intensity is interpreted to have gradually diminished during transgression. Potential causes for this include: (i) an increasing area over which nutrient-rich waters were dispersed; (ii) climatic variability associated with sea-level change (if eustatic); or (iii) bathymetric (tectonic) controls on the source of upwelled water. Decreased organic loading caused a diminishment of the oxygen deficiency that is recorded in the transition from phosphorites (Retort) to biogenic deposition (Tosi). The sponge biota and bioturbation within the Tosi imply increased oxygenation, but the presence of laminated, pyritic deposits implies mesotrophy, and potentially episodic dysoxia. Suboxic conditions could have played a role in the persistence of siliceous sponges as they are capable of tolerating low oxygen levels (Carballo *et al.*, 2004; Mills *et al.*, 2014; Mentel *et al.*, 2014). However, oxygenation could not have been the sole control on sponge dominance as bioturbated, and thus oxygenated, shallow facies in the Tosi (FA2-2) are still largely sponge dominated. What, then, led to the dominance of siliceous sponges?

Firstly, whereas upwelling intensity had decreased by Tosi deposition, aerobic respiration of organic matter in the water column would have still led to increased CO₂, decreased pH and a lower carbonate saturation state, factors that would have provided a further competitive advantage to sponges over biocalcifiers. Secondly, the lack of radiolarians in the Retort and Tosi suggests that DSi was recycled in the water column or unutilized by pelagic organisms. Resultantly, upwelled DSi may

have become increasingly concentrated through lowstand and transgression. Thus, as soon as upwelling intensity decreased to the point where a marine benthos could recolonize, the elevated DSi conditions, lower carbonate saturation state and variable oxygenation were ideal for sponges (Fig. 15). That the calcareous benthos was more common than is preserved in the Tosi cannot be ruled out as calcitic skeletons might have been dissolved in the shallow subsurface due to bacterial degradation of organic matter (cf. Ku *et al.*, 1999; James *et al.*, 2005). It is likely that once established, the glass ramp would have persisted through progressive diminishment of upwelling intensity and conditions increasingly less ideal to their colonization (i.e. hysteresis; Ritterbush, 2019), as elaborated on below.

Cessation of glass-ramp deposition

The transition from the Tosi biosiliceous biota to the Ervay calcareous biota occurred during, but was not necessarily due to, the continual decrease in upwelling intensity. Decreased intensity would have led to well-oxygenated bottom water, as is evidenced by the bioturbated offshore carbonates (FA3-1) versus laminated offshore Tosi facies (FA2-1). At the same time, pH and carbonate saturation would have risen, increasing the favourability of a calcareous biota. However, authigenic cements in FA3-1 indicate the sea was still mesotrophic and likely cool, as suggested by the sub-thermocline heterozoan biota (Fig. 13).

Spicules are increasingly recognized as an important silica sink, in part because they are less susceptible to marine dissolution than siliceous plankton (Maldonado *et al.*, 1999, 2011; Ritterbush, 2019). That the Tosi glass ramp must have caused a sufficient drawdown of DSi is unquestionable because it occupied at least 75,000 to 100,000 km² (Matheson & Frank, 2020). Calculations by Ritterbush (2019) indicate that glass-ramp DSi sinks could have significantly reduced the Permian DSi reservoir. It is thus possible that local DSi depletion by the Tosi sponges coupled with decreasing DSi inputs due to diminished upwelling limited the availability of DSi such as to negatively affect the sponge biota, effectively acting as a long-term negative feedback cycle to the conditions that favoured glass-ramp initiation (Ritterbush, 2019).

Once sponges took over, even as upwelling intensity decreased and DSi was depleted, the biota might have been resilient and able to remain dominant until some threshold was met in the

aforementioned conditions at which point sponges no longer outcompeted the biocalcifiers (Ritterbush, 2019). Thus, by the Tosi–Ervay transition, the sponge benthos, especially in shallow waters, was likely a relic of a ‘past’ seawater chemistry (i.e. anachronistic).

The transition from transgression (Tosi) to regression (Ervay), which included a minor lower-order sea-level fall, likely also favoured intensification of lagoonal circulation of downwelling brines (Figs 9 and 13; Campbell, 1962; Hite, 1978; Matheson & Frank, 2020). Downwelling brines might have decreased the impact of upwelling and contributed to mortality in the sponge benthos (Hogg *et al.*, 2010). That the impact of downwelling brines would have decreased offshore might partially explain why the Tosi–Ervay transition is sharp (occurred quickly) in onshore areas (for example, Fig. 12D and E) whereas the transition was more gradual offshore, resulting in interbedding at the transition (for example, Fig. 12B; Table 2).

This model well explains the observed trends within the Ervay cycle of the Bighorn Basin. However, mention should be made of facies relationships further west. The trend in the most seaward (westward) portion of the Bighorn Basin is towards increasing gradation at the Tosi–Ervay contact (Fig. 12B). This might indicate that the Tosi–Ervay contact becomes increasingly diachronous westward (i.e. outside the Bighorn Basin) and that, at least to some degree in the lower Ervay Member, the two interfinger (e.g. Wardlaw, 2015; Pommer & Sarg, 2019). This does not negate the successive nature of the units across the vast majority of 150 km wide Bighorn Basin. Nor does it argue the oceanographic model presented herein. It is expected that upwelled water and nutrient content would be more intense further offshore. Thus, just as temporal variation in upwelling intensity through time led to the phosphorite–chert–carbonate vertical succession in the Bighorn Basin, geographic variation in upwelling intensity might have also led to a lateral chert–carbonate gradation but at a scale of hundreds of kilometres. In this sense, the succession would at a vast scale follow Walther’s Law, even though it does not within most of the Bighorn Basin. This is not unexpected across what was a broad, low-gradient epeiric platform.

Biosiliceous deposition in the Ervay Member

Chert-rich transgressive deposits at the base of Ervay parasequences (FA2-3; Figs 9 and 11) indicate

that siliceous sponges episodically became more common again during deposition of the Ervay Member, although not at the complete eradication of the calcareous biota. This supports an at least indirect relationship between sea-level and sponge colonization, perhaps due to temporarily increased upwelling intensity or slowed lagoonal circulation.

Application of Ervay cycle concepts to the remainder of the Phosphoria Rock Complex

This oceanographic model must be considered with respect to the underlying Franson cycle, which also contains a phosphorite (Meade Peak), chert (Rex) and carbonate (Franson) member, whose stacking differs from the Ervay cycle (Fig. 2A; Peterson, 1984; Whalen, 1996). Unlike in the Ervay cycle, the Rex Chert was solely limited to areas outboard of Franson carbonates (Fig. 2A; Hendrix & Byers, 2000). One explanation for this difference is the progressive increase in landward tectonic subsidence that led to the retrogradational stacking of the cycles.

However, Hiatt (1997) also used various chemical proxies to suggest the oceanographic conditions leading to phosphogenesis were also more extreme during the Retort than the Meade Peak. More intense upwelling might partially explain the landward extent of Retort phosphorites and the ability of sponges to colonize into more coastal areas in the Tosi compared to the Rex (Fig. 2A). Progressive evolution of the global ocean, perhaps due to rising atmospheric CO₂ and strengthening of ocean currents or shoaling of the global lysocline, as proposed by Beauchamp & Grasby (2012) for the Sverdrup Basin, might have also contributed to this apparent intensification, despite it not explaining the trends within each cycle.

GLOBAL IMPLICATIONS

The dominant DSi utilizers in the Palaeozoic (i.e. sponges and radiolarians) differed from those in the silica-depleted, post-Mesozoic ocean (i.e. diatoms; e.g. Suess & Thiede, 1981; Glenn *et al.*, 1994b). Diatoms in the Cenozoic oceans are a major consumer of nutrients in upwelling systems (Bruland *et al.*, 2001; Tréguer & De La Rocha, 2013). There is thus a distinct difference in how upwelled DSi would have been utilized and stored in the Palaeozoic. The Ervay cycle provides a prominent example

in which this difference is manifest in the rock record. In modern settings, biosiliceous (diatomaceous) sediments in upwelling zones are found inboard of phosphorites, in part because phosphogenesis is often concentrated along the shelf edge (Pufahl, 2010). The Ervay cycle indicates that Palaeozoic systems may have differed in that biosiliceous sediments (spiculites and radiolarites) accumulated outboard of or completely replacing phosphorites, especially in epicontinental settings.

In addition, siliceous sponges are more resistant to dissolution and are a more permanent Si sink than diatoms (Maldonado *et al.*, 2011). The role of this difference on the Palaeozoic Si cycle remains unclear (Tréguer & De La Rocha, 2013; Ritterbush, 2019). However, results from this study clearly suggest that silica loading and DSi depletion might have been just as dominant a control on faunal turnover in pre-Cenozoic settings as phosphorous loading (Föllmi *et al.*, 1994; Glenn *et al.*, 1994b; Reid *et al.*, 2007). Clearly, these differences need further investigation and integration into reconstructions of Palaeozoic bioelemental deposition.

CONCLUSIONS

The Ervay cycle of the Phosphoria Rock Complex contains a unique succession of three conformable members that each record a disparate mode of sediment production. The Retort Member records authigenic phosphogenesis and reworking into granular phosphorites during lowstand and transgression. The Tosi Member records an epeiric glass ramp with widespread offshore to coastal siliceous sponge meadows that formed during transgression. The Ervay Member records a temperature-stratified carbonate ramp with offshore glauconite and phosphate-rich heterozoan facies and onshore molluscan, microbial and pisoidal peritidal facies that formed during the subsequent regression.

Whereas distinct oceanographic conditions were clearly responsible for the deposition of each member, conformable contacts and gradational transitions in offshore settings suggest that these conditions existed along a spectrum. These deposits support upwelling as a prominent feature of the Phosphoria Sea during deposition of all three sediment types, although potentially with subordinate aeolian nutrient input.

The proposed oceanographic model for the Ervay cycle involves progressive variation in upwelling intensity, nutrient levels, dissolved silica and oxygenation leading to the transitions

between the three sediment factories. Intense upwelling led to Retort Member phosphogenesis. The transitions from phosphogenesis to glass ramp (Tosi) to carbonate ramp (Ervay) are interpreted as the result of progressively decreasing upwelling intensity. Sponge colonization was favoured during the Tosi glass ramp due to: (i) increased, but potentially still below normal, oxygenation; (ii) elevated dissolved silica; and (iii) decreased carbonate saturation state. Cessation of sponge dominance and the return of a calcareous benthos records the depletion of DSi coupled with an increased carbonate saturation state. These results add an intriguing case study to the list of well-documented glass ramps and further supports elevated DSi and decreased carbonate saturation state (regardless of mechanism) as potentially unifying controls in their development.

The recognition of the related causes of the phosphorites and glass-ramp deposits in this study sheds new light on potential differences in biosiliceous deposition in pre-Cretaceous upwelling systems where sponges were a prominent Si sink as compared to diatoms in Cretaceous and younger systems. This interpretation may prove key in future investigation of other Palaeozoic bioelemental deposits.

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DATA AVAILABILITY

The list of outcrops, cores and wireline logs utilized in this study are included in the supplementary materials for this publication. The XRD analyses for this study are also available in the supplementary data file. Various petrophysical and petrographic reports, thin section and core images, and XRD data for the studied cores are available from the USGS Core Research Center's online core library.

Wireline logs used in this study are publicly available through the Wyoming Oil and Gas Conservation Commission. Any other data not provided are available from the corresponding author upon request.

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FIGURE CAPTIONS

Fig. 1. (A) Middle–Late Permian global palaeogeographic setting based on the 260 Ma time slice of Blakey (2017). The extent of (B) and (C) is shown by the red box. Orange shading is the Permian spiculite belt, modified after Murchey & Jones (1992) and Ritterbush (2019). Pink shading is the approximate extent of the western Laurentian desert (cf. Fig. 4 inset). CPM = Central Pangaeian Mountains, SMH = Slide Mountain-Havallah Basin. (B) Palaeogeographic map of the Phosphoria Sea at its maximum extent (the Ervay cycle). Water shading ignores bathymetric complexities (cf. Peterson, 1984). Base image modified from 270 Ma time slice of Blakey (2013) based on data from this study as well as Sheldon (1963), Ahlstrand & Peterson (1978), Peterson (1984), Whalen (1996) and Hendrix & Byers (2000). (C) The depocentres for each formation in the Phosphoria Rock Complex, modified after Sheldon (1963) and Whalen (1996). Evaporitic basins in Wyoming and surrounding states (for example, ‘Goose Egg Basin’; cf. Wardlaw, 2015) are not shown due to poor constraints on their relation to the Phosphoria Sea. Modified after Matheson & Frank (2020).

Fig. 2. (A) East–west lithostratigraphy of the Phosphoria Rock Complex (PRC). See Fig. 1B for location. Modified after Maughan (1984) based on data from Peterson (1984). (B) Stratigraphy of the PRC in the southern Bighorn Basin. Based on research conducted by the authors combined with frameworks from Peterson (1984), Paull & Paull (1990), Simmons & Scholle (1990) and Inden & Coalson (1996). Neither cross-section represents north–south variability. No absolute chronostratigraphic ages are implied, see Fig. 3. Modified after Matheson & Frank (2020). Minn = Minnekahta Member.

Fig. 3. A comparison of published age models for the Phosphoria Rock Complex. (A) The age model of Wardlaw (2015) based on conodont and brachiopod biostratigraphic constraints, which is supported by unpublished U-Pb ages and $^{87}\text{Sr}/^{86}\text{Sr}$ values from lime micrite and a brachiopod shell in the Ervay Member from Pommer (2018). Isochrons for the aforementioned U-Pb ages have considerable uncertainty (± 3 Ma and 12 Ma; Pommer, 2018); only the uppermost (i.e. youngest) extent of these is shown. The stratigraphic relationships of the Retort, Tosi and Ervay members as presented by Wardlaw (2015) have been modified based on the stratigraphic work presented herein;

see Fig. 2 and the text for discussion. **(B)** The model of Davydov *et al.* (2018b) based on a U-Pb age from an ash bed in the Meade Peak Member of southern Idaho combined with a modified biostratigraphic zonation scheme. Dinwoody age constraints are based on Paull & Paull (1994) and references therein. **(C)** The relative proportion of the Sverdrup Basin covered by siliceous sponge meadows (i.e. glass-ramp deposits) and carbonate benthos as modified from Beauchamp & Grasby (2012); one data point is provided per stage. **(D)** and **(E)** The inferred stratigraphic placement of the Tosi glass ramp using the two age models. Geological time scale based on Gradstein *et al.* (2012) and Ogg *et al.* (2016). Changing the Kungurian–Roadian age as proposed by Davydov *et al.* (2018a) would not largely change the interpretations in this study. *The Blacktail or Blacktail Creek member was first proposed by Schock *et al.* (1981) as a member of the Park City Formation in southern Montana that is equivalent to the Ervay Member. It is composed of cherty dolomitic mudstone, siltstone and silty carbonate rocks. Its usage has not been widely adopted, in part because of a paucity of subsequent research on the PRC in the area.

Fig. 4. The study area for this project, the Bighorn Basin. Outcrop belts modified after Love & Christiansen (2014), Lopez (2000, 2001) and Vuke *et al.* (2000). Inset shows the location of the basin (red box) and the outlines of the marine and marginal marine Phosphoria Rock Complex (PRC) facies (blue) and equivalent terrestrial deposits (red) (modified from Walker, 1967; Peterson, 1984; Thornburg, 1990; Ulmer-Scholle & Scholle, 1994; Whalen, 1996). Modified after Matheson & Frank (2020). Site names corresponding to locality abbreviations are included in Appendix S1. Cross-sections A–A' and B–B' are shown in Fig. 9.

Fig. 5. Facies Association 1 phosphorites from the Retort Member. Red arrows in all panels are select francolite grains. Yellow arrows are calcitic brachiopods. Green arrows are detrital quartz grains. **(A)** Granular phosphorite (FA1-1) in outcrop at the South Fork of Owl Creek (OC). **(B)** Granular phosphorite (FA1-1) from T41X-21G RP Gooseberry (T41XG; *ca* 1851.5 m; 6075 ft). **(C)** Phosphatic dolowackestone (FA1-2) from 93 Cottonwood Creek (93CC; *ca* 2878 m; 9443 ft). **(D)** and **(E)** Photomicrographs of silica-cemented sandy phosclast (peloid-coated grain-intraclast) grainstone (FA1-1) from the core piece in (B). Photomicrographs are of samples under plane polarized light.

Grains include: CB = calcitic brachiopods, IB = inarticulate brachiopods, PCG = phosphate coated grains, PI = phosphatic intraclasts, PP = phosphate peloids and QS = quartz sand.

Fig. 6. Facies Association 2 spiculitic cherty deposits from FA2-1 (A) to (E), FA2-2 (F) and (G), and FA2-3 (H) to (L). FA2-1 and 2 (A) to (G) are from the Tosi Member and FA2-3 (H) to (L) is from the Ervay Member. (A) Ovoid black chert nodules in a laminated dolomicrite at the top of the Tosi Member in Wind River Canyon (WR). (B) Horizontally elongate black chert nodules (yellow arrows) with calcified anhydrite nodules (red arrow) in 28-1 Hunt Rawhide (281HR; depth of *ca* 1866 m; 6123 ft). (C) Bedding plane exposure of irregular black chert nodules replacing *Thalassinoides* burrows at Shoshone River (SR). (D) Vertical exposure of irregular black chert nodules replacing burrows at Owl Creek (OC). (E) Photomicrograph of well-preserved spicules within a black chert nodule at Anchor Dam (AD) under plane polarized light. The nodules are similar to those in (D). (F) Irregular patchy grey chert nodules (silicification) at Spence Dome (SD). In the terminology of Wistort *et al.* (2019), the chert nodules in (C) are silicified trace fossils, those in (D) are trace-like chert nodules, and (F) contains both (see description in Matheson & Frank, 2020). (G) A purple chert nodule in a dolomicritic host rock in the Tosi Member of 2-16 ANR (ANR; *ca* 1255.5 m, 4119 ft). (H) A portion of a *ca* 1.5 m interval with abundant chert nodules (yellow arrows) in a skeletal packstone to grainstone host rock in 82 Pitchfork (82P; *ca* 1137.5 m; 3732 ft). The host rock in this instance contains abundant authigenic minerals including glauconite and phosphate (cf. Fig. 7). (I) Photomicrograph in plane polarized light from immediately below the base of the core in (H) showing silicification (yellow arrow) around abundant bryozoans with phosphate cement in zooecia (red arrow) and peloidal glauconite. (J) Plane polarized light photomicrograph of patchy silicification (yellow arrow) in a barren dolomicritic matrix from the red box in (K). (K) Core of a *ca* 2.5 m interval of patchy (light tan) silicification within the Ervay Member of 1 Crooked Draw (1CD; *ca* 3944.75 to 3946.25 m, 12942 to 12947 ft). (L) Purple chert nodules within an otherwise peritidal Ervay Member parasequence at Mahogany Butte (MB). Depths marked on cores are in feet. Scale bar increments in (C), (D) and (L) are 2 cm and those on (F) are 1 cm.

Fig. 7. Facies Association 3 carbonates from the Ervay and Franson members. (A) to (C) FA3-1, (D)

to (F) FA3-2, (G) and (H) FA3-3, (I) to (K) FA3-4. Stratigraphic location noted on all panels: bte = beyond Tosi extent (cf. Fig 9), mat = metres above Tosi, mbr = metres below Retort. (A) Bryozoan floatstone with fossiliferous (brachiopod, pelmatozoan and bryozoan) wackestone matrix and authigenic replacement and cementation by phosphate (brown/purple), glauconite (green), iron oxides (orange), and silica (white) in 28-1 Hunt Rawhide. (B) Phosphatic and glauconitic sandy skeletal packstone in 87 Pitchfork. (C) Pelmatozoan–bryozoan wackestone to packstone with silty lime mud matrix and abundant authigenic cementation and replacement at Anchor Dam. (D) and (E) Brachiopod and bryozoan floatstone with skeletal wackestone matrix in 28-1 Hunt Rawhide (D) and T41X-21G RP Gooseberry (E). Black specks are small fragmented phosphate replaced/cemented bioclasts. Base of Tosi/Retort not cored in (D) and so the possible depth range is given. (F) Skeletal grainstone at Wind River Canyon with a mixture of heterozoan grains and micrite envelopes after molluscs. (G) Mollusc dolofloatstone to rudstone with mollusc-peloid dolopackstone to wackestone matrix at Double Rainbow Ranch. (H) Scaphopod floatstone with peloid-mollusc packstone to wackestone matrix at Five Springs. Matrix mud and peloids are replaced by very finely crystalline to micritic dolomite. This unit contains angular and highly irregular rip-up clasts of underlying pisolitic grainstone (PI). (I) Fenestral fabric in a peloidal grainstone replaced by fabric-retentive mimetic dolomite in 4 Medicine Wheel. The purple arrow is a thin band of cement that is either bottom-growth marine cement or meteoric fracture fill, both are present within these deposits. (J) Partially occluded fenestral fabric in a peloidal dolograins at Shoshone River. (K) Pisoidal grainstone with micritic dolomite cement (yellow arrow) and later crystalline dolomite occluding intergranular porosity at Cottonwood Canyon. Photomicrographs in (C), (F), (G), (J) and (K) are all in plane polarized light. Br = brachiopod, By = bryozoan, C = chert cement, CN = chert nodule, FF = fenestral fabric, G = glauconite cement, I = iron oxide cement, M = mollusc mould (likely pelecypods), ME = micrite envelope after mollusc shell, G = gastropod mould, P = pelmatozoan, Pe = peloid, Ph = phosphate (francolite) cement, PhS = phosphate replacing skeleton, PI = pisolitic intraclasts [similar texture to (K)], S = silicified skeleton, Sc = scaphopod.

Fig. 8. Facies Association 4. (A) Poorly exposed interbedded red beds and gypsum (purple arrows) of the Ervay Member of the Goose Egg Formation in the western Powder River Basin [Mayoworth

(MA)]. The top of the Ervay Member is marked by the blue arrow, with the overlying 23.5 m thick Freezout Member (Goose Egg Formation) red beds seen in the background (yellow arrow). The Ervay–Freezout contact is generally accepted as the Permian–Triassic boundary (Boyd & Maughan, 1972). **(B)** to **(E)** Better exposed examples of Facies Association 4 in the PRC underlying the Ervay cycle. **(B)** Red beds of the Opeche Member (base) and carbonates and gypsum of the overlying Minnekahta Member (above yellow arrows) at Red Gulch. **(C)** and **(D)** Red beds with enterolithic gypsum nodules in the Difficulty Member of the 2-16 ANR core (**C**; depth *ca* 1258 m; 4128 ft) and the very uppermost Ervay Member in 11-10 Federal (**D**). **(E)** Laminated gypsum in the Difficulty Member at Little Mountain; laminations are defined by thin seams of red siltstone. Jacob staff subdivisions are 10 cm.

Fig. 9. West to east (top) and south to north (bottom) stratigraphic cross-sections across the Bighorn Basin hung on the base of the Retort/Tosi members. Cross-section locations shown on Fig. 4. Only the sequence stratigraphic interpretation of the highest-order relative sea-level fluctuations is shown on B–B'. See Matheson & Frank (2020) for discussion of the chert colouration within the Tosi. FSST = falling-stage systems tract, GR = gamma ray, LST = lowstand systems tract, MFZ = maximum flooding zone, SB = sequence boundary, TST = transgressive systems tract. Modified after Matheson & Frank (2020).

Fig. 10. **(A)** Extent of the Retort Member. **(B)** Isochore map of the Tosi Member after Matheson and Frank (2020). If the Tosi is present and its thickness is measurable at an outcrop/core, the colours used are the same as on Fig. 4; otherwise colouration is shown in the legend. **(C)** Isochore map of the Ervay Member with the initial Ervay coastline shown (see Fig. 9). See Inden & Coalson (1996) for a higher-resolution map of Cottonwood Creek Field peritidal deposits (FA3-4). Note the differing depocentres between the Tosi and Ervay members; the Tosi is thicker offshore and the Ervay is thicker onshore (peritidal deposits, i.e. FA3-3 and FA3-4).

Fig. 11. Representative stratigraphic sections of the Ervay cycle across the study area showing first-order trends, including regression through deposition of Ervay Member carbonates, as well as the

variably identifiable parasequences within the Ervay and Tosi members (yellow triangles, as on Fig. 12B). Chert marker beds highlighted are those that are widely correlatable and utilized as timelines, see Fig. 9 and the text for discussion. For more partial and complete Ervay cycle sections, see Fig. 12, Coalson & Inden (1990) and Inden & Coalson (1996) for examples from the eastern side of the basin where the Ervay Member is predominantly peritidal facies (FA3-4), and Matheson & Frank (2020); fig. 10. See Table 1 for more thorough description of each facies association. All sections contain underlying strata in the Franson cycle that are not shown, and all three sections were chosen because at each the contact with the overlying Dinwoody Formation was exposed. Depositional texture abbreviations: fs = floatstone, gs = grainstone, ms = mudstone, ps = packstone, rs = rudstone, ws = wackestone. Sequence stratigraphic abbreviations: FS = flooding surface; FSST = falling-stage systems tract; HST = highstand systems tract; LST = lowstand systems tract; MFZ = maximum flooding zone; SB = sequence boundary; TST = transgressive systems tract. Other abbreviations: Fran = Franson Member; Ret = Retort Member.

Fig. 12. First-order transitions between lithostratigraphic members from the west (A) to (C), south-central (D), and north-east (E) portions of the basin. (A) Gradational Retort to Tosi transition at the South Fork of Owl Creek (OC). Silicified brachiopods shown by yellow arrows. (B) The Tosi to Ervay transition including interbedding of Facies Associations 2 and 3 at the South Fork of Owl Creek. Parasequences from deeper-water to shallower-water carbonates in the Ervay Member are shown by yellow triangles; see text for discussion. (C) Franson (F), Retort (R) and Tosi (T) in the T41X-21G RP Gooseberry core (T41XG) showing interbedded chert (FA2-1) and phosphorite (FA1-1). Images brightened and contrast increased to aid viewing of chert nodules (yellow arrows). (D) A sharper Tosi-Ervay transition at Wind River Canyon (WR). Minor chert nodule horizons in the fossiliferous Ervay facies are shown by yellow arrows. (E) The Franson through the top of the PRC at Spence Dome (SD). Chert nodules preserving mollusc moulds in the upper Franson shown by yellow arrows. Note the absence of the Retort Member and sharp contacts between all units. See figs 4 and 10 in Matheson & Frank (2020) for other exposures of the Franson through Ervay succession.

Fig. 13. Depositional models for each of the three members of the Ervay cycle. See text for details.

The model for deposition of the Retort Member shows the very end of phosphorite deposition (i.e. transition to the biosiliceous Tosi Member). FSST = falling-stage systems tract, HST = highstand systems tract, LST = lowstand systems tract, MFZ = maximum flooding zone, TST = transgressive systems tract.

Fig. 14. Simplified stratigraphic succession of the Ervay cycle showing first-order trends. Exact parasequences within the Franson and Ervay members (Inden and Coalson, 1996) are not shown for simplicity (see Figs 9 and 11). Cherty carbonates (green dashed lines) represent idealized parasequences (cf. Fig. 9). Not to scale. FSST = falling-stage systems tract, HST = highstand systems tract, LST = lowstand systems tract, MFZ = maximum flooding zone, TST = transgressive systems tract.

Fig. 15 – The proposed mechanisms by which temporally variable nutrient input, in this case interpreted to be the by-product of systematic changes in the intensity of upwelling, led to the bioelemental sediments of the Ervay cycle. References to oxygenation refer only to bottom water.

*See text for discussion of possible mechanisms by which upwelling intensity might have varied. Dark grey boxes are those interpreted as the direct controls on the nature of the benthic fauna that produced the PRC sediments. Black boxes are the depositional system that produced each member of the Ervay cycle. Dashed lines are knock-on effects that played a role in dictating ‘future’ sedimentation types. DS_i = dissolved silica.

Table 1. Description and interpretation of facies within the Franson and Ervay cycles (Phosphoria Rock Complex) in the Bighorn Basin.

Table 2. Marine facies distributions within the Franson and Ervay cycles in the Bighorn Basin.

Appendix S1. List of all outcrops, cores and wireline logs used in this study.

Appendix S2. X-ray diffraction data.

Table 1 – Description and interpretation of facies within the Franson and Ervay cycles (Phosphoria Rock Complex) in the Bighorn Basin.

Facies	Lithology*	Sedimentary structures	Allochems	Benthos skeletal mineralogy	Early diagenesis	Depositional environment	Stratigraphic distribution (cf. Table 2)
<i>Facies Association 1: Phosphorites (Fig. 5)</i>							
FA1-1: Granular phosphorites (Figs 5A, 5B, 5D, 5E, 12A and 12C)	Phosclast packstones with dolomicritic matrix, phosclast grainstones with equigranular megaquartz cement, inarticulate brachiopod packstone/grainstone	Predominantly structureless, rare grading	Common: phosclasts (coated grains, peloids, intraclasts, inarticulate brachiopods, fish bones), quartz silt and sand. Rare: calcitic brachiopods, siliceous spicules	Limited benthos, chitin-phosphate and minor calcite and biosilica	Authigenic phosphogenesis and pyrite precipitation	Authigenic phosphogenesis and reworking in offshore, eutrophic, anoxic, low-energy settings with occasional storms	Bulk of Retort, minor component in Tosi
FA1-2: Phosphatic dolostone (Fig. 5C)	Phosclast dolowackestone	Structureless	Phosclasts (as above)	N/A	Authigenic phosphogenesis	Phosphogenesis in FA1-1 environments with phosclasts transported onshore	Retort, similar to variants of FA3-1 in Ervay
<i>Facies Association 2: Cherty deposits (Fig. 6)</i>							
FA2-1: Black chert (Figs 6A to 6E; 12A to 12D)	Spiculitic black chert in dolomicrite, lime micrite, sandstone	Variably flat laminated (with pyrite) to structureless (BI: 1 to 4), discrete	Common: monaxon siliceous spicules. Rare:	Predominantly siliceous	Silica remobilization and chertification**	Sponge meadows in offshore, mesotrophic, low to moderate-energy settings with variable oxygen-limitation and anoxic sediments	Bulk of Tosi (basinward occurrences), minor component in

		<i>Thalassinoides</i>	brachiopods, bryozoans, pelmatozoans				Retort
FA2-2: Grey and purple chert (Figs 6F, 6G and 12E)	Spiculitic and silicified-evaporite-nodule-bearing grey and purple chert in dolomicrite and dolomitic shale	Structureless (BI: 3 to 5), microbial laminations, flat laminations, contorted laminations around silicified enterolithic anhydrite nodules	Common: monaxon siliceous spicules. Minor: beds of molluscs and ooids/peloids. Rare: brachiopods, bryozoans, pelmatozoans	Predominantly siliceous	Silica remobilization and chertification**	Well oxygenated, low-energy shallow-marine sponge meadows to evaporitic tidal flats	Tosi (landward occurrences)
FA2-3: Cherty fossiliferous to barren carbonates (Figs 6H to 6L and 12B)	Spiculitic black, grey, and purple chert in fossiliferous to barren subtidal/peritidal carbonates (cf. FA3)	Predominantly bioturbated (BI: 5 to 6) and structureless, discrete <i>Thalassinoides</i>	Common: monaxon siliceous spicules, brachiopods, bryozoans, pelmatozoans Minor: see FA3-1	Mixed calcareous and siliceous	Authigenic cementation (see FA3-1), silica remobilization and chertification**	Mixed calcareous and sponge benthos in offshore to coastal/peritidal environments	Basal TST of parasequences in the Franson and Ervay members
<i>Facies Association 3: Carbonates (Fig. 7)</i>							
FA3-1: Fossiliferous carbonates with authigenic	Fossiliferous floatstone, packstone, wackestone and	Structureless (BI: 5 to 6)	Common: brachiopods, bryozoans, pelmatozoans.	Predominantly calcitic	Authigenic cementation and replacement by phosphate,	Offshore, nutrient-rich (mesotrophic), well-oxygenated, sub-thermocline water with variable reducing conditions in	Franson and Ervay

phosphate and glauconite (Figs 7A to 7C, 12B and 12D)	minor rudstone in silty lime micrite, dolomicrite, or very finely crystalline anhedral dolomite matrix		Minor: trilobites, ostracods, subrounded phosclasts, siliceous spicules, fish bones, glauconite peloids		glauconite, iron oxides and quartz/chert**	shallow seafloor (see text)	
FA3-2: Brach-bryo-molluscan carbonates lacking authigenic minerals (Figs 7D to 7F and 12B)	As above	Structureless (BI: 4 to 6)	Common: As above plus molluscs (cf. FA3-3)	Mixed calcitic and aragonitic	Minor authigenic phosphate/glauconite and replacement (or reworking?) and minor marine carbonate diagenesis (cf. FA3-3)	Offshore, transitional between FA3-1 and FA3-3	Franson and Ervay
FA3-3: Molluscan carbonates (Fig. 7G and 7H)	Molluscan grainstone, rudstone and packstone	Structureless (BI: 3 to 5) or planar tabular trough cross-bedded	Common: bivalves, peloids, gastropods, scaphopods, cephalopods. Minor: green algae, solitary corals, miliolid foraminifera, pisoids, pisolitic intraclasts (FA3-4)	Predominantly aragonitic	Marine carbonate diagenesis (micritization, marine isopachous cement growth, etc.)	Shallow subtidal (shoreface to sub-fair-weather wave-base) area with oligotrophic, warm water and potentially elevated salinity	Franson and Ervay

FA3-4: Pisoidal, peloidal, and microbial carbonates (Figs 7I to 7K and 12E)	Non-skeletal pisoidal, peloidal, microbial carbonates (bindstone, grainstone, solution-collapse breccia, barren dolomudstone)	Structureless and poorly sorted, to reverse graded, abundant fenestral fabric, microbial laminations (including stromatolites), teepee structures. Minor: desiccation cracks	Common: Coated grains (pisoids, see text), peloids, oncoliths. Minor: gastropods, miliolid foraminifera, ostracods, mollusc debris, <i>Thartharella</i> sp.	Microbes dominate, limited biota	Fabric-retentive reflux dolomitization and synsedimentary lithification (geopetal sediments, fractures, endolithic borings)	Sabkha, ranging from mesosaline to hypersaline subtidal (incipient microbial laminations in barren dolomudstone) to supratidal, transitional to FA4	Franson and Ervay
<i>Facies Association 4: Terrestrial deposits (Fig. 8)</i>							
FA4-1: Red beds (Fig. 8A, 8B, 8D and 8E)	Iron-rich claystone, siltstone, mudstone, minor sandstone	Structureless (outcrop), or weakly laminated, sometimes contorted, with anhydrite nodules, chickenwire fabric and minor pedogenic alteration	None	NA	Gypsum cementation	Windblown silt (loess) with minor pedogenic modification	Goose Egg Formation
FA4-2: Bedded gypsum/anhydrite (Fig. 8A and 8C)	Gypsum & anhydrite (20 cm to 3 m) with thin (<30 cm)	Flat to microbial (?) laminations, chickenwire	None	NA	NA	Hypersaline lagoons and coastal salinas	Goose Egg Formation

	sucrosic dolostone beds at base	fabric					
FA4-3: Gypsiferous dolostone (not figured)	Gypsiferous sucrosic dolostones	Microbial laminations, gypmouldic porosity, chickenwire fabric, boxworks, solution-collapse breccias	None	NA	Reflux dolomitization	Hypersaline lagoons and coastal salinas (transitional to FA3-4)	Goose Egg Formation

*Facies are variably sandier in the north-west Bighorn Basin due to tongues of sand extending from the Shedhorn Formation, but are otherwise identical to their non-sandy counterparts

(Peterson, 1984; Inden & Coalson, 1996; Matheson & Frank, 2020).

**See Ulmer-Scholle & Scholle (1994) and Matheson & Frank (2020) for discussion of silica diagenesis.

Table 2 – Marine facies distributions within the Franson and Ervay cycles in the Bighorn Basin.

Member	Dominant facies	Subordinate facies	Thickness (Figs 9 and 10)	Upper contact – west (offshore)	Upper contact – east (onshore)
Ervay	FA3 (all) – carbonate facies	FA2-3 (moderate) – cherty carbonates	<40 m*	Dinwoody Formation – sharp (conformable or disconformity, see text)	Freezout Member (red beds; Goose Egg Formation) – sharp (Fig. 12E)
Tosi	FA2-1 & FA2-2 – cherty facies	FA2-1 (rare) – phosphorites	2 to 12 m	Ervay – sharp but conformable, sometimes with interbedding (Fig. 12B)	Ervay carbonates (FA3-3 and FA3-4) or terrestrial facies (FA4) – sharp but conformable
Retort	FA1 (all) - phosphorites	FA2-1 (rare) – cherty facies	Often <0.75 m, rarely up to 1.5 m	Tosi – gradational but abrupt (Fig. 12A, C)	Tosi – gradational (?) from dolomicrite with phosclasts to dolomicrite with chert
Franson	FA3 (all) – carbonate facies	FA2-3 (minor) – cherty carbonates	<25 m across the basin	Retort – sharp with minor erosion	Tosi (Retort pinched out) – sharp

* See Allison (1984), Coalson & Inden (1990) and Inden & Coalson (1996) for discussion of the thickening in the peritidal deposits in the eastern Bighorn Basin.

Undifferentiated Permian and Triassic	Triassic (Dinwoody and Chugwater fms)
Permian (Phosphoria Rock Complex)	Pennsylvanian (Tensleep and Amsden fms)
Undifferentiated late Palaeozoic-Mesozoic	

- Measured outcrops and cores**
Outcrop section location	Logged by other authors
Logged core location	Wells with wireline logs

Pennsylvanian Tensleep Formation & Early Permian Nowood Complex

- Retort Member present as FA1-1 (granular phosphorites) or high gamma spike
- Retort Member present as FA1-2 or moderate/equivocal gamma spike
- Retort Member not present or no gamma spike
- White outline = core/outcrop with "Retort interval" exposed

FA1 (Phosphorites)

- FA1-1 (Granular phosphorite)
- FA1-2 (Phosphatic dolowackestone)

FA2 (Cherty dolomite)

- FA2-1 (Black chert)
- FA2-2 (Grey chert)
- FA2-3 (Carbonates w. chert)

FA3 (Carbonates)

- FA3-1 (Heterozoan carbs w. glauc/phosphate)
- FA3-2 (Heterozoan carbs w. molluscs)
- FA3-3 (Molluscan carbonates)
- FA3-4 (Peritidal carbonates)

- ★ Collected & slabbled sample
- ★ Thin sectioned sample

Bryozoa

Brachiopods

Crinoids

Sponge spicules

Phosphatic peloids

Phosphatic coated grains

Phosphatic intraclasts

P Thin (<30 cm) phosphate bed

P Phosphate cement

G Glaucoplastic cement

Chert nodules

Thalassanoides

Gastropods

Bivalves

Scaphopods

Unidentified shell debris

Pisoids

Peloids

Dessication cracks

Fenestral fabric

Vertical borings

Microbial laminations

Chickenwire fabric

Calcite geodes

Horizontal laminations

Low-angle laminations

Wavy bedding

Time 3 - Ervay Mbr - Carbonate Ramp

- Regression (HST & FSST)
- Weak upwelling
- Oxygenated water & sediment

Time 2 - Tosi Mbr - Glass Ramp

- TST up to MFZ
- Decreasing upwelling
- Variable oxygenation

Time 1 - Retort Mbr - Authigenic Phosphogenesis

- LST & early TST
- Strong upwelling - eutrophic
- Water column anoxia

Phosphorite & phosphatic dolostone - FA1	Siliceous sponges	Bivalves	Microbial laminations
Spiculitic black, grey & purple chert - FA2	Sponge spicules	Gastropods	Teepee structures
Marine & peritidal carbonates - FA3	Brachiopods	Scaphopods	Evaporite nodules
Red beds & gypsum - terrestrial facies - FA4	Bryozoa	Pisoids	Salinas
Phosphatic peloids and coated grains	Crinoids	Peloids	Upwelling water
Phosphatic intraclasts	Burrows	Milliolid forams	Downwelling brines

Decreasing Upwelling Intensity (See Figs 14 and 15)

Upwelling intensity (nutrient content)*