Origin, sequence stratigraphy and depositional environment of an upper Ordovician (Hirnantian) deglacial black shale, Jordan

Howard A. Armstrong\textsuperscript{a,\!*}, Brian R. Turner\textsuperscript{a}, Issa M. Makhlouf\textsuperscript{b}, Graham P. Weedon\textsuperscript{c}, Mark Williams\textsuperscript{d}, Ahmad Al Smadi\textsuperscript{e}, Abdulfattah Abu Salah\textsuperscript{e}

\textsuperscript{a}Department of Earth Sciences, University of Durham, Science Laboratories, Durham, U.K.
\textsuperscript{b}Department of Earth and Environmental Sciences, Hashemite University, Zarqa, Jordan
\textsuperscript{c}Department of Geography, University of Wales, Swansea, Singleton Park, Swansea SA2 8PP, U.K.
\textsuperscript{d}British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB3 0ET, U.K.
\textsuperscript{e}Natural Resources Authority, P. O. Box 7, Amman 11118, Jordan

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Abstract

The upper Ordovician succession of Jordan was located \textasciitilde60\degree S, less than 100 km from the Hirnantian ice sheet margin. New graptolite dates indicate glaciation ended in Jordan in the late Hirnantian (\textit{persculptus} Biozone). The succession records two glacial advances within the Ammar Formation and the subsequent deglaciations. Organic-rich black shales (Batra Formation) form part of the final deglacial transgressive succession that in-filled an existing low stand glacial continental shelf topography. The base of the black shale is coincident with the maximum flooding surface. During transgression, interfluves and sub-basin margins were breached and black shale deposition expanded rapidly across the region. The top of the black shales coincides with peak highstand. The \textquotedblleft expanding puddle model\textquotedblright \ (sensu Wignall) for black shale deposition, adapted for the peri-glacial setting, provides the best explanation for this sequence of events.

We propose a hypothesis in which anoxic conditions were initiated beneath the halocline in a salinity stratified water column; a fresher surface layer resulted from ice meltwater generated during early deglaciation. During the initial stages of marine incursion, nutrients in the monimolimnion were isolated from the euphotic zone by the halocline. Increasing total organic carbon (TOC) and $\delta^{13}$C\textsubscript{org} up section indicates the organic carbon content of the shales was controlled mainly by increasing bioproductivity in the mixolimnion (the Strakhov model). Mixolimnion nutrient levels were sustained by a continual and increasing supply of meltwater-derived nutrients, modulated by obliquity changes in high latitude insolation. Anoxia was sustained over tens to hundreds of thousands of years. The formation of black shales on the north Gondwana shelf was little different to those observed in modern black shale environments, suggesting that it was the nature of the Ordovician seas that pre-disposed them to anoxia.

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\textsuperscript{*} Corresponding author.
\textit{E-mail address:} h.a.armstrong@durham.ac.uk (H.A. Armstrong).

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1. Introduction

Early post-glacial, transgressive, organic-rich black shales were a ubiquitous feature of the northern Gondwana continental shelf during the late Ordovician to early Silurian. These rocks, which contain up to 10% total organic carbon, have an elevated gamma log response (up to 200 API, Andrews, 1991) due to uranium-enrichment, and constitute one of the world’s major hydrocarbon source rocks. The origin of these black shales is uncertain, and resolution of this problem, the aim of this paper, could provide considerable palaeoenvironmental, palaeoclimatic and palaeooceanographic information.

Elucidation of the origin of these rocks is hampered by the fact that modern open shelf settings are only subject to seasonal anoxia (Tyson, 1996; Tyson and Pearson, 1991) and black shales are not currently forming on modern shelves, despite anthropogenic inputs, making present day shelf seas the most nutrient-rich in Earth history. Further, authigenic U-enrichment (Anderson et al., 1989) and the widespread preservation of non-bioturbated, millimetre-scale laminations in these rocks would be unsustainable with seasonal oxia (Wignall, 1991). This led Wignall (1991) to ask the question: if black shales are not forming on continental shelves at the present day then why should they have formed in the past?

2. Existing models for anoxia

Studies of modern oxygen-poor environments have shown that oxygen deficiency may be due to decreased oxygen supply or increased oxygen demand beneath areas of high productivity (Demaison and Moore, 1980; Wignall, 1991) and black shales are not currently forming on modern shelves, despite anthropogenic inputs, making present day shelf seas the most nutrient-rich in Earth history. Further, authigenic U-enrichment (Anderson et al., 1989) and the widespread preservation of non-bioturbated, millimetre-scale laminations in these rocks would be unsustainable with seasonal oxia (Wignall, 1991). This led Wignall (1991) to ask the question: if black shales are not forming on continental shelves at the present day then why should they have formed in the past?

2.1. Upwelling model

Wind-driven coastal upwelling zones are the most important, and the vast majority of the world’s dysaerobic sediment forms in these areas. At the present day, anoxic and dysoxic conditions only occur in coastal upwelling zones at the intersection of an intensified oxygen minimum zone (OMZ) and the upper slope or continental shelf break, in water depths typically >50 m (Wignall, 1994). Typically, radiolarian and/or diatoms are abundant and the sediments are silica-enriched. The upper and lower boundaries of the OMZ are marked by phosphate enrichment (Demaison and Moore, 1980; Loughman, 1984). Organic-rich shale accumulation in upwelling zones is due to elevated productivity levels in a regime of vigorous circulation and high organic loading. Oxygen deficiency aids the preservation of organic carbon in the sediment (Parrish, 1982; Finney and Berry, 1997). For example, despite the very high productivity around the margins of Antarctica, the exceptional oxygen-richness of the water prevents high organic carbon values in the sediments (Demaison and Moore, 1980). Typically, OMZs associated with upwelling zones are only stable on a decade time-scale (Wignall, 1990), whilst oxygenation events from the influx of turbidites tend to be once-in-a-hundred-year events (Souter et al., 1981).

Analogy with modern systems predicts that ancient black shales forming in association with coastal upwelling should be palaeogeographically restricted to low latitudes. However, given the right climatic circumstances, upwelling zones could form parallel to coastlines, in which case they would have a long, narrow sheet-like geometry. The interval of black shale formation should coincide with evidence for increased current activity. Black shales should be intimately associated with chert (diatom in high latitudes and radiolarian at low latitudes) and phosphate-rich facies. These facies associations are rarely observed in early Palaeozoic high latitude settings as diatoms had not appeared by this time and mechanisms exist for rapid phosphorous recycling into the upper water column where it is utilized by the biosphere (van Cappellen, 1993).

Black shales, indicative of late Ordovician–early Silurian anoxia, have been attributed to oceanic upwelling. At this time, much of the northern margin of Gondwana (including Jordan) lay in high palaeolatitudes within a zone of westerly winds, which according to computer simulations generated intense and in some areas seasonally restricted upwelling (Parrish, 1982; Moore et al., 1993). The same simulations, however, indicate that upwelling contin-
ued into the Wenlock, long after deposition of black shale had ceased; hence, more complex processes must have been operating. However, the black shales in Jordan fail the low latitude restriction and facies geometry predicted by the upwelling model.

2.2. Silled basin model

The modern Black Sea, which is the largest anoxic basin in the world, has been identified as a good analogue for ancient black shale (Pompeckj, 1909; Tyson et al., 1979; Wignall, 1994). The basin is almost landlocked and its link with the Mediterranean Sea through the Straits of Bosporus is silled, and locally <30 m deep. The Black Sea receives large volumes of fresh water from rivers draining its northern hinterland as well as marine Mediterranean water. Due to the density contrast between these water types, there is a well-developed halocline at ~200 m, which severely restricts advection of oxygen-rich surface waters to depth (Wignall, 1994). Lateral advection is also restricted due to the small amount of water entering the basin, and circulation beneath the halocline is sluggish, taking ~1000 years for complete mixing of the water beneath the halocline (Brumsack, 1989). The combination of poor circulation and moderate oxygen demand from decaying and descending organic matter results in an anoxic monimolimnion that contains free hydrogen sulphide.

The sediments accumulating in the deepest parts of the Black Sea basin consist of fine-grained siliciclastic turbidites derived from the basin margins (Lyons and Berner, 1992) and organic-rich, microlaminated, coccolith oozes and sapropels. Two conflicting models are available for the formation of the organic-rich sapropels, both of which rely on the influx of marine water (Fig. 1). In the Deuser model (Deuser, 1974), marine water produces a puddle in the deeper parts of the basin which gradually expands upwards. Organic carbon accumulation is the result of enhanced preservation beneath the anoxic water column. Changes in surface productivity are not considered important in this model and progressive restriction of the size of the mixolimnion would gradually restrict nutrient availability and thus could lead to a decline in productivity and the resulting sedimentary organic carbon. In contrast, the Strakhov model (Calvert, 1990; Calvert and Fontugne, 1987; Calvert et al., 1987; Strakhov, 1971) invokes elevated productivity. During initial stages of marine incursion, nutrients in the monimolimnion are envisaged to have remained isolated from the euphotic zone by the halocline. The gradually rising monimolimnion then releases nutrients to the mixolimnion (“van Cappellen effect”) and triggers a high productivity episode, which is sustained by organic matter production in the euphotic zone. Thus, the elevated sedimentary organic carbon values record a high flux of organic matter to the sediment at a time when much of the water column was fully oxygenated. This model does not take into account the initial concentration of black shale deposition in the deepest parts of the basin and must be allied with the presence of weak bottom currents to concentrate sediment and organic material (Huc, 1988).

Whilst the Black Sea is not an ideal analogue for a peri-glacial shelf, it does provide diagnostic criteria for the recognition of black shales that formed in a
silled basin setting. Most organic-rich shales initially accumulate in topographic lows, due to the concentration of low density organic particles by weak currents (Huc, 1988), and would have a restricted distribution pattern. In the Strakhov model, black shales formed beneath the halocline and are interbedded with normally oxygenated facies. This results from climate-driven changes in water advection (Wignall, 1994) or, with increasing sea level, a breach of the sill, which allows large influxes of oxygenated marine waters (Tyson and Pearson, 1991). The Deuser model predicts declining euphotic zone bioproductivity, in contrast to the Strakhov model which predicts sedimentary organic matter will record patterns of euphotic zone bioproductivity. During deglaciation, in ice margin settings, melting of the ice sheet and hence nutrient flux and euphotic zone bioproductivity would be modulated by changes in high-latitude seasonal insolation.

High latitude seasonal insolation is controlled to a large degree by ~21 kyr precession cycles, and low-latitude seasonal extremes are entirely dominated by precession (Berger, 1978; Budziak and al, 2000; Molfino and McIntyre, 1990; Perks et al., 2002; Rostek et al., 1997). However, over the annual cycle, precessional insolation anomalies sum to zero at all latitudes (Liu and Herbert, 2004). In contrast, obliquity variations (~41 kyr periodicity) change the annual insolation received as a function of latitude (Ruddiman and McIntyre, 1984). A high obliquity configuration increases insolation at latitudes above 45° in both hemispheres and decreases the solar energy received in the tropics and subtropics. Therefore, obliquity changes the amplitude of the seasonal cycle and alters the equator-to-pole insolation gradient. Milankovitch theory predicts the Hirnantian ice sheet, nutrient flux, bioproductivity and preserved sedimentary organic matter (TOC) would primarily record obliquity cycles.

3. Sequence stratigraphical context for black shale formation

Wignall (1994) and Wignall and Maynard (1994) identified two distinct types of black shale within the transgressive sedimentary record: basal transgressive (BT) and maximum flooding (MF) black shales (see also van Wagoner et al., 1990), depending on whether they rested on the transgressive surface (ts) or the maximum flooding surface (mfs). Basal transgressive black shales are further characterised by the fact that: (1) they only develop in topographic hollows and basin-centre locations at times of non-deposition on the basin margin; (2) they have no time equivalent deeper water facies (Grabowski and Glaser, 1990); (3) they exhibit no lateral facies variation, but are interbedded with shallow-water facies, perhaps implying pelagic depositional processes; and (4) they are probably shallow water facies as they occur at the base of the transgressive systems tract (TST). In areas where sediment influx is low, this surface would be diachronous towards the basin margin. In such cases, where the TST is poorly developed, black shale may closely overlie transgressive surfaces, or the sequence boundary, transgressive surface and maximum flooding surface may all amalgamate into a single erosive or non-depositional surface (Wignall, 1994; Wignall and Maynard, 1994). On the other hand, MF black shales that rest conformably on the maximum flooding surface can extend a considerable distance on to the shelf and pass proximally into thicker sections of shallower, marginal marine facies. They represent a deep-water, sediment starved condensed section.

Wignall (1991) proposed the “expanding pudding model” to explain the evolution of black shale basins. In the pudding, dyoxic/anoxic bottom waters, in basin-floor lows, are inferred beneath a stratified water column. Stratification can be vertical gradients of either temperature (Heckel, 1971; Tyson et al., 1979) or salinity/density (Baird et al., 1987; Fleet et al., 1987). For example, Rossignol-Strick et al. (1982) explained the development of anoxic conditions and sapropel formation in the Mediterranean Sea by the presence of a cap of low-density freshwater, which stopped the sinking of higher density oxygen-bearing waters. As sea level rises, incised valleys and existing topography on the basin floor are gradually filled to a point where the interfluve areas or sub-basin margins are drowned (Fig. 2). This results in rapid expansion of accommodation space. If the sedimentation rate is constant or reduced, the accommodation space cannot be filled, the basin becomes starved (Wignall, 1991), and an extensive ravinement surface develops. This model predicts black shales rapidly expanding outwards from the area of greatest basin depth (controlled
by existing basin floor topography) and the area of deposition is at its greatest extent during highstand.

4. Stratigraphical and sedimentological context of the upper Ordovician of Jordan

Ordovician sedimentation occurred on the margins of the more extensive North African (Gondwana) shallow-marine shelf, within terrestrial, subtidal, marginal marine and shelf environments (Amireh et al., 2001; Makhlfouf, 1995; Powell et al., 1994), and by upper Ordovician times the shelf was influenced by Hirnantian glaciation (Brenchley et al., 1994; Long, 1993; Marshall et al., 1997; Qing and Veizer, 1994) and the presence of a temperate “Laurentide-scale” ice sheet in the continental interior (Turner et al., 2002) (Fig. 3). The current upper Ordovician lithostratigraphical framework in Jordan does not adequately reflect the changing facies associated with this glaciation or allow for surface and subsurface correlation of the stratigraphy. We have therefore adopted an informal scheme prior to submission to the Geology Directorate of the Natural Resources Authority of Jordan for formal acceptance (Fig. 3).

4.1. Dubaydib Formation

The Llanvirn(?)-Ashgill Dubaydib Formation, which comprises sandstones dominated by hummocky cross-stratification, is divided into three members (Fig. 4). The lower member, which attains a thickness of 25 m, is defined by the first occurrence of the trace fossil Sabellarifex and comprises quartz-rich sandstones that grade upward into siltstone (Makhlouf, 1992). The unit is pervasively burrowed by a variety of vertical, uniserial or U-shaped burrows with circular cross-sections, attributable to the Skolithos Ichnofacies characteristic of high-energy, shallow water conditions associated with moving sands. The middle member comprises 54 m of channelised, fine-
grained, cross-bedded sandstones, with subordinate very fine, silty sandstone. The upper member is 76 m thick and contains six coarsening-upward cycles of greenish siltstone to grey fine-grained sandstone (Makhlouf, 1992). Trace fossils include *Cruziana*, *Chondrites* and *Planolites* assigned to the *Cruziana* Ichnofacies, which is best developed below normal fair-weather wave base in well-sorted silts and sands in relatively quiet water conditions.

### 4.2. Tubeiliyat Formation

The Ashgillian Tubeiliyat Formation is up to 105 m thick and conformably overlies the Dubaydib Formation (Fig. 4). It comprises greenish, silty shale and hummocky cross-stratified sandstone (Makhlouf, 1992), organised into coarsening-upward and less commonly fining-upward packages, which decrease in thickness from ~25 m to 10 m upwards. Trace fossils of the *Cruziana* Ichnofacies are common in the greenish silty shales.

### 4.3. Ammar Formation

Abed et al. (1993) and Amireh et al. (2001) included all the late Ordovician (Ashgillian) glacial sediments in southern Jordan within the Ammar Formation (Fig. 5). They divided it into two units, each underlain by a glacial erosion surface. The upper unit of the Ammar Formation comprises erosively based, palaeovalley channel-fill sandstones which consist of: (1) coarse-grained, proximal, braided outwash plain sandstones containing glacially faceted and striated clasts, typically concentrated in the lower 1–3 m, which Abed et al. (1993) and Powell et al. (1994, Fig. 5) interpreted as tillite; and (2) coarse-grained, proximal, braided outwash plain sandstones, similar to those in (1), except that they grade up into thin marine shoreface sandstones containing trace fossils and brachiopods (Abed et al., 1993; Amireh et al., 2001) similar to the Tubeiliyat shoreface sediments beneath the channel-fill sandstones.

Abed et al. (1993, Fig. 6) considered the glacially incised lower unit of the Ammar Formation to comprise a sandy channel lag conglomerate up to 2 m thick, containing glacially faceted and striated clasts. This is overlain by a 30 m thick, greenish-grey, massive, well-sorted sandy siltstone, which is extensively deformed and thought to be completely void of primary sedimentary structures and fossils (Fig. 5). A re-examination of the channel-fills in the Jebel Ammar area shows that they exceed 90 m in

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**Fig. 4.** Currently accepted lithostratigraphy and chronostratigraphy of the late Ordovician of southern Jordan, and its correlation with Saudi Arabia. The scheme used herein is a modification of this and raises the Tubeiliyat and Batra members to formation status.
thickness locally and that, despite tectonic and soft-sediment deformation of the sandstones, they preserve rare primary ripple cross-stratification, burrows and brachiopod moulds.

Erosively based channels, which differ from the upper Ammar glaciofluvial channels, are locally incised into the top of the Tubeiliyat formation (Fig. 5). These are filled by up to 50 m of massive, pale grey, fine-grained, fawn-weathering sandstone. The sandstone is uniformly fine-grained, micaceous, undeformed and contains small-scale, trough cross-beds and ripple cross-lamination. The sandstone along the sides and floor of the channel contains brachiopods and numerous trace fossils, including Harlania, Cruziana and brachiopod resting traces. Thus, two glacial incision events are recognized in the late Ordovician of southern Jordan. The first occurs beneath the disturbed and undisturbed channel sandstones locally incised into the top of the Tubeiliyat Formation, with the former included in the lower part of the Ammar Formation by Abed et al. (1993). The second incision occurs at the base of the upper Ammar Formation glaciofluvial to marine channels. These are incised into the deformed and undeformed lower channels, as well as the underlying non-channelised Tubeiliyat sediments.

At least two episodes of glacial and peri-glacial deposition are present in Saudi Arabia termed the Zarqa and Sarah Formations (McClure, 1978; McClure, 1988; Vaslet, 1990; Young, 1981). Both formations have glacial unconformities at their bases and contain glacial, glacio-fluvial and glacio-marine facies (Vaslet, 1990). We consider the lower and upper members of the Ammar Formation to correlate with the Zarqa and Sarah Formations (Fig. 5).

4.4. Batra Formation

The Batra Formation is ~40–120 m thick and, in the Southern Desert region, it conformably overlies
the Ammar Formation or disconformably overlies the Tubeiliyat Formation (Fig. 4). In the type area of Wadi Batn el Ghul, the formation comprises 3 members (Andrews, 1991). The lower member is found in shallow wells, apparently restricted to fault-bounded graben structures. Only the middle member outcrops and spans the Ordovician–Silurian boundary (Powell, 1989).

The Batra Formation is variably graptolitic. Masri (1988) identified abundant graptolites, sparse thin-shelled bivalves and rare trilobites in the mudstone. Andrews (1991) reviewed the biostratigraphy of this formation from surface exposures and exploration wells and concluded that the formation ranged in age from Ashgill (*persculptus* Biozone age) to mid-Wenlock. New collections from black shales from core in the type area contain abundant *Normalograptus parvulus* (Fig. 6), which is likely co-specific with *Normaolgraptus persculptus* and ranges from the *persculptus* to *acuminatus* biozones. In the Al Jafr area (well JF-1, 3494 ft depth; Fig. 3), side wall cores contain *Glyptograptus persculptus*, whereas in the Wadi Sirhan area (well WS-6, 1399.8 m depth; Fig. 3) the lower member contains an *acuminatus* Biozone fauna. *Monoclimacis flumendosae* and *Pristograptus dubius* were reported at the top of the formation in the Risha area (Paleoservices, 1987). Close to the border with Saudi Arabia, the Batra Formation is overlain by the Ratiya Formation; the basal 10–15 m of which consists of mudstone yielding a *sedgwickii* Biozone fauna of the uppermost Aeronian (Powell, 1989). These data indicate that the base and the top of the Batra Formation young northwards (Fig. 4).

In the type area, the lower member (*persculptus* Biozone age) comprises 17.44 m of black shales (borehole BG14; locality N2930504 E3557410). They comprise laminated, black siltstone to dark grey homogeneous claystone couplets. The parallel laminated siltstones are organic-rich, and grade upwards into mudstone, with irregular patches of siltstone (*?*starved ripples) and homogeneous claystones (Fig. 7). The absence of bioturbation indicates anoxic bottom water during deposition. The siltstone laminites, characteristic of the whole formation, contain pyrite and marcasite concretions throughout, confirming an anoxic depositional environment.

The middle member (*persculptus–early acuminatus* Biozone age) similarly consists of grey and iron-oxide-rich maroon, parallel-laminated siltstones and grey homogeneous mudstones that become more sandy upwards; all units are variably graptolitic (Powell, 1989). These are petrographically identical to the turbidites of the lower member, except that the beds thicken upwards and they have been intensively oxidised following lithification. The couplets are interpreted as the Td and Te microfacies of distal turbidites (Fig. 7).
The Batra Formation is synchronous with black shale deposits around the Gondwana margin (Sutcliffe et al., 2000; Sutcliffe et al., 2001). In Saudi Arabia, the subsurface equivalent of the Batra Formation is the Qusaiba Shale, dated as Ashgill-Ludlow (McClure, 1988) (Fig. 4). The comparable basal “hot” black shale facies contains persculptus Biozone graptolites and the overlying Qusaiba Shale yields Rhudanian, gregarious Biozone and convolutus Biozone graptolites.
5. Tests

A critical assessment of depositional models, and an explanation of the processes of black shale formation (the Deuser and Strakhov models) in the peri-glacial setting of Jordan, depend on three tests: (1) placing these rocks in their correct sequence stratigraphical context, (2) demonstrating an increase in TOC and $\delta^{13}$Corg (proxies for bioproductivity) up section, and (3) detecting obliquity forcing in these proxy data.

Data are drawn from the upper Ordovician to lower Silurian siliciclastic succession of the Southern Desert region of Jordan, as well as published and unpublished well data from the Al Jafir, Wadi Sirhan and Risha areas (Fig. 3). Detailed sedimentary logs and biostratigraphical analysis provide the framework for sequence stratigraphical analysis of the succession, including Fischer plots, following the methodology of Sadler et al. (1993).

$^{13}$C/$^{12}$C analyses were performed on de-carbonated material by combustion in a Carlo Erba 1500 on-line to a VG TripleTrap and Optima dual-inlet mass spectrometer, with $\delta^{13}$C values calculated to the VPDB scale using a within-run laboratory standard (cellulose, Sigma Chemical prod. no. C-6413) calibrated against NBS-19 and NBS-22. Replicate analysis of well-mixed samples indicated a precision of $\pm 0.1\%$ (1 S.D.). The carbon isotopic relationships associated with the production and burial of organic carbon have been reviewed by Wang et al. (1997).

Total organic carbon was measured on de-carbonated samples using a Leco CS244 carbon analyser. Typical repeatabilities (relative standard deviations) were better than 0.7% of the measured carbon values of analytical standards (0.816% carbon) and 3.8% of the measured carbon values of the samples. Spectral analysis of %TOC and $\delta^{13}$Corg was conducted using an algorithm for irregularly spaced data (Press et al., 1992).

5.1. Test 1—sequence stratigraphy

A relative sea level curve summarizes our sequence stratigraphical interpretation, which allows correlation with the better known sections in Baltica and Laurentia through matching relative sea level events (Fig. 8, rsl a–e). It is convenient to divide the discussion into pre-, glacial and post-glacial periods.

5.1.1. Pre-glacial

The Dubaydib and Tubeiliyat Formations are characterized by wave and storm-dominated shoreface environments. Deposition was interrupted in the lower part of the Dubaydib Formation (middle member) by the development of a regionally extensive, erosively based channel system (Figs. 4 and 8, rsl a). The channels have been variously interpreted as rip-current channels (Makhlouf, 1995) or storm channels (Powell, 1989), but their precise origin and relationship to base-level change have yet to be determined. Similar channels, interpreted as subtidal channels, occur at

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Fig. 8. Relative sea level curves from Baltica (after Nielsen, 2004), Laurentia (after Ross and Ross, 1992, 1996) and Jordan. Upper Ordovician–lower Silurian biostratigraphy and chronostratigraphy from (Fortey et al., 2000). Jordan lithostratigraphy as proposed in this paper. Letter notation on the Jordan sea level curve identifies relative sea level events described that can be used to correlate the curve (see text for details). Abbreviations: L, lower; M, middle; U, upper. Graptolite biozones are: acumin., acuminatus; persc., persculptus; extra., extraordinarius; comp., complanatus; linear., linearis.
about the same lithostratigraphical level in Libya where biostratigraphical evidence suggests they are lower Ashgill in age (Turner et al., 2002). Newly collected brachiopods from shell lags at the base of the channel-fill succession in southern Jordan support this age assignment (Prof. D.A.T. Harper, 2002, pers. comm.).

The Fischer plot for the upper member of the Dubaydib and Tubeiliyat Formations (cycle thickness data from Makhlouf, 1992) (Fig. 9a) exhibits two orders of cyclicity, the individual cycles and a long term, higher order of cyclicity. The latter defines two transgressive–regressive cycles within the upper member of the Dubaydib Formation and the lower part of the Tubeiliyat Formation (Fig. 8, rsl b; Fig. 9a, peak HST at cycle 3; Fig. 8, rsl c; Fig. 9a, peak HST at cycle 8) followed by a general increase in cycle thickness to the top of cycle 18, close to the top of the Tubeiliyat Formation.

The Dubaydib and Tubeiliyat Formations span the Ashgill (~447–443 Ma, sensu Fortey, 2000) and these orders of cyclicity are considered third- and second-order. The increase in accommodation space within the latter Ashgill most likely marks the onset of glacial isostatic subsidence and increased sediment supply, and we suggest that much of the deposition in the Tubeiliyat Formation is glacially forced.

5.1.2. Glacial

The top of the shoreface sediments (Tubeiliyat Formation) is truncated and incised by two episodes of channel incision (Fig. 8, rsl b and c). The first glacial incision and deposition of thin (<2 m), conglomeratic outwash plain sediments was quickly followed by marine transgression and deposition of marine shoreface sandstones which occupy the greater part of the fill of the lower channels, incised into the top of the Tubeiliyat Formation.

The second incision (Fig. 8, rsl c) formed a set of much larger, erosively based channels, now interpreted as sub-glacial tunnel valleys (Turner et al., 2002). The channels have been previously interpreted as marking the onset of the Hirnantian glaciation, coincident with a major eustatic fall in global sea-level of between 50 and 100 m (Woodcock and Smallwood, 1988; Ross and Ross, 1988; Eyles, 1993; Heredia and Beresi, 1995). The data presented here indicates that two phases of ice advance and retreat occurred during the glaciation, within the Ammar Formation. This led to the development of two relatively short-lived, third-order glacially induced lowstands. The interpretation of the stratigraphically equivalent glacial lowstand elsewhere in North Africa as second order (Luning et al., 2000) is incompatible with the short-lived nature of the Hirnantian glaciation (Sutcliffe et al., 2000).

5.1.3. Post-glacial

The onset of post-glacial transgression (Fig. 8, rsl d) occurs in the persculptus Biozone. The channel-fill of the upper member of the Ammar Formation comprises basal conglomerates and fluvial cross-stratified sandstones containing flood-generated fining-upwards cycles, deposited within perennial braided channels characterised by fluctuating discharge (Abed et al., 1993). Fluvial sandstones fine
upwards into marine sediments with the transgressive surface (ts) placed at the base of the marine sediments in the channel-fill succession (Fig. 5).

The base of the Batra Formation therefore represents the maximum flooding surface external to the depocentre. This would be equivalent to the ravinement surface that developed when interflues within the existing basin floor topography were breached, with the resulting rapid, lateral expansion of basinal anoxic conditions. This is seen in the diachronous onset of black shales northwards, towards the glacial fore-bulge, between the Hirnantian and early Rhuddanian. The black shales of the lower Batra Formation, therefore, formed between the maximum flooding surface and the base of the falling stage systems tract (FSST), and we interpret them as maximum flooding (MF) black shales that developed within an expanding puddle.

The Fischer plot for the lower part of the Batra Formation records an increase–decrease–increase in accommodation space (Fig. 9b). The initial increase reflects the continuing transgression. The decline in accommodation space (“regression”) from the peak of highstand (cycle 11) through into the organic-poor shales of the middle member could be interpreted in two ways in a four system model. This decline could be the result of relative sea level fall, the FSST, or alternately, it could represent increasing sediment starvation as the transgression continued. The main sedimentological change occurs at the base of the middle member of the Batra Formation with the appearance of grey, organic-lean shales. These are still microlaminated (i.e. no bioturbation) indicating oxidation was not syn-depositional. A second major sedimentological change occurs within the middle member where turbidites increase in thickness and grain-size (Fig. 9b, cycle 115; Fig. 8, rsl e). These changes are interpreted as recording increased sediment input to the basin reflecting relative sea level fall at the start of the FSST. Global eustatic sea-level curves (Ross and Ross, 1996) indicate that this second-order sea-level rise continued until the late Rhuddanian or early Aeronian. However, Melchin et al. (1998) highlighted a possible regression during the late acuminatus Biozone and this could correlate with rsl e (Fig. 8).

Fig. 10. Temporal and cyclic variation in TOC and organic carbon isotopic values for the lower Batra Formation. (a) $\delta^{13}$C$_{org}$ plotted against height above the base of the Batra Mudstone Formation. (b) TOC plotted against height above the base of the Batra Mudstone Formation. (c, d) Spectral analysis of %TOC and $\delta^{13}$C$_{org}$ using an algorithm for irregularly spaced data (Press et al., 1992). 95% CL = one sided 95% level confidence levels (Mann and Lees, 1996), bw = band width. Abbreviations: OM, organic matter; TOC, total organic carbon.
5.2. Test 2—temporal variation in TOC and $\delta^{13}C_{\text{org}}$

$\delta^{13}C_{\text{org}}$ values from the lower member of the Batra Formation (well BG14) show metre-scale oscillations and a rising trend of $-31\%$ to $-29.5\%$ (Fig. 10a). TOC values, taken as a proxy for bioproductivity, increase cyclicly from 2% to 8% (Fig. 10b). TOC increases in three stages up section. This corresponds to 2.7 cycles through the TOC time series.

5.3. Test 3—obliquity forcing of TOC and $\delta^{13}C_{\text{org}}$

Spectral analysis indicates regular cyclicity in the TOC data, with a wavelength of 6.38 m (Fig. 10c), but shows no significant cyclicity in the $\delta^{13}C_{\text{org}}$ data (Fig. 10d). However, overall, there is a general trend to more positive $\delta^{13}C_{\text{org}}$ values and increasing TOC. Milankovitch theory predicts the cycles in the TOC data will reflect orbital obliquity. The Ordovician orbital-obliquity cycle period was ~36,000 years (Berger et al., 1989) and the time series has a minimum duration of ~97,000 years.

This result can be corroborated using stratigraphical thickness data from the international boundary stratotype at Dob’s Linn. Here the ratio of the stratigraphical thickness of the first post-glacial highstand (negative shift in $\delta^{13}C_{\text{org}}$ during the persculpus Biozone) to strata representing the glacial maximum (~extraordinarius Biozone) is 0.72 m (thickness data from Underwood et al., 1997). Assuming sedimentation rate was constant and the extraordinarius Biozone represents two eccentricity cycles, as envisaged by Sutcliffe et al. (2000), then the duration of the HST is 144 kyr.

6. Model for the formation of black shales during deglaciation

Luning et al. (2000) reviewed the occurrence of anoxic, black shale facies throughout North Africa and the Arabian Peninsula offering a three-system, sequence stratigraphical model characterised by the succession in Libya. They placed the transgressive surface at the base of the black shales. However, the interpretation of the early post-glacial black shale as BT shales (Luning et al., 2000) is inconsistent with the field evidence and stacking patterns seen in the equivalent Jordan succession, where the black shale lies on the mfs. Initially organic-rich black shale was deposited in glacially incised shelf sub-basins. As the deglacial transgression progressed, sub-basin margins and interfluves were eventually breached producing a rapid increase in accommodation space. Low sedimentation rates meant that this space could not be filled and the basin became starved. The spread of black shale away from the depocentres within glacially incised shelf depressions or sub-basins was diachronous. We therefore contend that black shales are MF shales (sensu Wignall, 1991) and that they

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Fig. 11. Conceptual model for the upper Ordovician organic-rich “hot” shale in Jordan. (a) Initiation of the anoxic puddle. (b) Expansion of the puddle during the HST. Abbreviations as in previous figures.
fulfil the diagnostic criteria of the expanding puddle model (Fig. 11).

Anoxic conditions could have been initiated in two ways: (1) through the upward expansion of a warm salinity dense oceanic monimolimnion that had remained anoxic throughout the glaciation or (2) formation beneath a stratified water column within a sub-basin on the shelf. In the second scenario, it is likely that, during initial stages of marine incursion, nutrients in the monimolimnion remained isolated from the euphotic zone by the halocline, the gradual rising monimolimnion released nutrients to the mixolimnion (van Cappellen effect), and triggered a high productivity episode which initiated anoxia. By analogy with the present Antarctic margin and the Black Sea, it would seem more likely that stratification probably resulted from a salinity gradient with more saline waters at depth. The fresher surface mixolimnion would have resulted from ice meltwater generated during the earliest stages of deglaciation, coincident with a reduction in density driven thermohaline circulation as the ice margin retreated.

Increasing TOC and $\delta^{13}C_{org}$ up section indicate that once euxinic conditions were established, photic zone bioproductivity was the primary cause of sustained anoxia supporting the Strakhov hypothesis. The supply of meltwater-derived nutrients was directly related to the seasonal melting dynamics of the ice sheet and was forced by the predicted changes in orbital obliquity. Anoxia was sustained in this way over tens–hundreds of thousands of years.

7. Conclusions

The onset of Hirnantian glaciation in Jordan is marked by the Ammar Formation that lies in the proximal part of the glacial depositional system. In Jordan, the second of two glacial incision events resulted in steep, U-shaped tunnel valleys up to 4 km wide and 70 km long, trending NNE–SSW, parallel to palaeo-iceflow. These were to become the initial depocentres for black shale deposition. The outcrops of the Ammar Formation delimit the proximity of the ice margin, at or near the grounding line.

Sequence stratigraphical analysis indicates that the succeeding organic-rich black shale of the Batra Formation are MF shales that formed beneath a salinity stratified water column. Initially anoxia was developed beneath the halocline as a consequence of a cap of low density glacial meltwater, declining thermohaline circulation and the ingress of normal marine waters at depth. Increasing values of TOC and $\delta^{13}C_{org}$ upwards indicates that once anoxia was established it was sustained by mixolimnion bioproductivity which itself was sustained by continued and increasing levels of meltwater-derived nutrients. Ice sheet melting was modulated by orbital obliquity. As the deglacial transgression continued, sub-basin margins were eventually breached resulting in the rapid lateral expansion of the black shale facies. This scenario is best encapsulated in the expanding puddle model for black shale deposition.

Modern day processes, operating in silled basins, can be used to explain the development of organic-rich black shales on the peri-glacial Gondwana shelf in Jordan. Peri-glacial settings are fully oxygenated at the present day. What distinguishes the Ordovician black shales in Jordan is the initiation anoxia of early in the deglacial transgression. Either, the shelfal basins in which these sediments formed were highly restricted silled basins, comparable to the Black Sea at the present day and low levels of oxygen in the atmosphere–ocean pre-disposed Ordovician shelf seas to anoxia or, the Hirnantian glaciation was insufficient to develop an intense thermohaline circulation that could ventilate the world’s oceans.

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