

Evolution of Maastrichtian paleoproductivity along the southern Tethys margin (Jordan): From upwelling to runoff

Maria Ardila-Sanchez^{a,*}, Jihede Haj Messaoud^a, Amir Kalifi^a, Khalil Ibrahim^b,
Muhammad Usman^{a,1}, Israa S. Abu-Mahfouz^c, Samer Aljurf^c, Volker Vahrenkamp^a,
John H. Powell^d, Frans van Buchem^{a,*}

^a King Abdullah University of Science and Technology, Thuwal, Saudi Arabia

^b Department of Earth and Environmental Sciences, Prince El-Hassan Bin Talal Faculty of Natural Resources and Environment, The Hashemite University, Zarqa, Jordan

^c Department of Geosciences, King Fahd University of Petroleum and Minerals, Dhahran, Saudi Arabia

^d British Geological Survey (BGS), Nottingham, United Kingdom

ARTICLE INFO

Editor: Thomas Steuber

Keywords:

Late Cretaceous
Upwelling
Paleoproductivity
OMZ
Tethys
Organic-rich carbonates
Jordan

ABSTRACT

The Upper Cretaceous succession in Jordan records the evolution of an upwelling system characterised by the widespread deposition of phosphorites, siliceous mudstones, and carbonate mudstones with TOC contents of up to 30 wt%. Here, we investigate the Maastrichtian succession of Jordan along a basin-scale SW–NE transect (~61 km) using a multi-proxy approach that integrates sedimentological and geochemical datasets within a high-resolution biostratigraphic framework.

Regional correlations reveal hierarchical depositional cyclicity and a diachronous distribution of organic matter that progressively shifted landward through time. Maximum TOC enrichment occurred under conditions of high paleoproductivity associated with an upwelling system. Trace-element enrichments (Ni, Cu, Zn, Mo, U, and V) further suggest a strong coupling between productivity and the development of an oxygen minimum zone (OMZ), analogous to modern upwelling systems. The organic-rich carbonate interval was deposited from the early Maastrichtian (UC16b) to the late Maastrichtian (UC20a), representing approximately 2.5 myr of sustained high-productivity conditions in central Jordan.

During the late Maastrichtian, a major paleoceanographic reorganization occurred, weakening the upwelling system and causing a split in organic carbon and mineral carbon productivity cells on the Levant shelf. Organic-matter productivity continued into the Paleocene in the nearshore realm (southern Jordan), driven primarily by continental runoff. In contrast, in the more distal, seaward part of the shelf, sediment production shifted toward carbonate-producing planktonic microorganisms.

This study thus provides a dynamic picture of Maastrichtian paleoproductivity in Jordan and presents a model that may apply to other locations along the submerged southern Neo-Tethys Ocean margin shelves.

1. Introduction

During the Cretaceous, the accumulation of organic-rich marine sedimentary rocks is enhanced by a combination of global paleoenvironmental conditions, including greenhouse climates, major transgressive periods, and widespread oceanic anoxia events (Arthur et al., 1990; Wilson and Norris, 2001; Gale et al., 2002; van Buchem et al., 2002; Wilmsen, 2003; Schlager, 2005; Gambacorta et al., 2016; van der Meer et al., 2022). In addition, other authors have emphasized the

important role of primary productivity in the accumulation of organic matter, mostly driven by increases in nutrient supply (e.g., Parrish and Curtis, 1982; Pedersen and Calvert, 1990; Parrish, 1995; Leckie et al., 2002; Katz, 2005; Huc et al., 2005; Hay, 2009; Jenkyns, 2010; Bohacs et al., 2020). Upwelling zones are commonly associated with enhanced nutrient supply and drive the widespread distribution of phosphatic, siliceous, and organic-rich carbonate rocks along Paleozoic and Mesozoic continental shelves (e.g., Parrish and Curtis, 1982; Piper and Link, 2002; Ader et al., 2014; Cañadas et al., 2024). Similar

* Corresponding authors.

E-mail addresses: maria.ardilasanchez@kaust.edu.sa (M. Ardila-Sanchez), frans.vanbuchem@kaust.edu.sa (F. van Buchem).

¹ Present address: Saudi Aramco, Dhahran, 31311, Saudi Arabia.

successions are also observed in modern upwelling systems offshore regions of Namibia, Peru, and the Gulf of California (Shannon and Nelson, 1996; Borchers et al., 2005; Inthorn et al., 2006; Böning et al., 2015; Dale et al., 2015).

During the Late Cretaceous to Eocene, a long-term transgressive phase occurred along the southern margin of the Neo-Tethys Ocean (Flexer and Reymont, 1989; Gvirtzman et al., 1989; Adate et al., 2002; El-Azabi and Farouk, 2011; Powell and Moh'd, 2011). This transgression resulted in the deposition of hemipelagic to shallow marine carbonate rocks, characterised by significant lateral variations in lithology, and locally also in thickness. These successions have been studied in Tunisia, Egypt, Lebanon, and Jordan, for their stratigraphic and paleogeographic significance, as well as their economic importance, including phosphate-rich deposits and hydrocarbon source rocks (Glenn and Arthur, 1990; May, 1991; Tantawy et al., 2001; Soudry et al., 2013; Khoury, 2019).

In Jordan, the Maastrichtian hemipelagic carbonates rocks of the uppermost Belqa Group consist at the base of granular phosphatic marlstone that grades upward into dark-gray, organic-rich marlstone. This interval represents the immature source rock of the Muwaqqar Chalk Marl (MCM) Formation, with total organic carbon (TOC) content values reaching up to 30 wt% (Abu-Mahfouz et al., 2019; Usman et al., 2025). A sharp color change marks the end of the organic-rich interval, followed by interbedded chalky limestone and marlstone with low TOC values (<1 wt%) (Fleurance et al., 2013; Abu-Mahfouz et al., 2019; Bazeen et al., 2020).

The high paleoproductivity recorded in the Levant region during the Late Cretaceous is commonly attributed to a prolonged upwelling system that prevailed along the margins of the Neo-Tethys Ocean for approximately 19 myr (Almogi-Labin et al., 1993; Soudry et al., 2006; Edelman-Furstenberg, 2009; Ashkenazi-Polivoda et al., 2011; Powell and Moh'd, 2011; Abed, 2013; Alsenz et al., 2013; Meilijson et al., 2014). Upwelling currents transport cold, nutrient-rich deep waters into shallow marine environments, enhancing bio-productivity and promoting the accumulation of sediments enriched in phosphorus (P), silica (Si), and organic carbon (C_{org}). In addition, increased continental runoff and weathering rates may also enhance nutrient supply in shallow waters, thereby boosting bioproductivity and organic matter accumulation. The relative influence of these nutrient sources may vary between proximal and distal depositional settings (Leckie et al., 2002; Wilmsen, 2003; Erba, 2004; Batenburg et al., 2016; Poulton et al., 2015; Shmeit et al., 2022).

Another factor controlling the accumulation of organic-rich sediments in marine carbonate settings is the development of anoxic bottom water conditions, commonly associated with restricted or partially restricted environments (Demaision and Moore, 1980; Tissot et al., 1984; Hay, 1995; Tyson, 2005; McArthur et al., 2008; Jenkyns, 2010). A compressional tectonic regime and associated basin segmentation along the northern African and Arabian plates affected accommodation space and the distribution of phosphatic and organic-rich facies of Coniacian-Maastrichtian successions of the Negev, Sinai, and western Egypt (Gvirtzman et al., 1989; Kuss et al., 2000; Guiraud et al., 2005). Similarly, organic-carbon enrichment in the Maastrichtian MCM Formation of Jordan has been attributed to deposition within fault-controlled basins related to the same regional compressional phase (Abed and Amireh, 1999; Abed et al., 2005; Powell and Moh'd, 2011; Fleurance et al., 2013; Ali Hussein et al., 2015; Alqudah et al., 2015; Hakimi et al., 2016; Beik et al., 2017, 2018, 2019; Sokol et al., 2017; Grohmann et al., 2023; Aljurf et al., 2026). A recent tectonostratigraphic study by Kalifi et al. (2025), however, suggests that only minor syndimentary subsidence occurred in Jordan during the Maastrichtian, contrasting with traditional depositional models proposed for the MCM Formation. At the same time, modern upwelling systems demonstrate that intense oxygen minimum zones (OMZs) can also develop in oceanographically open settings, such as along the Namibian and Peruvian continental shelf margins, where they may occur at water depths as shallow as ~50 m (Böning et al., 2004; Böning et al., 2015;

Sweere et al., 2016).

Previous studies of the Maastrichtian of Jordan, largely based on cored wells, have focused on the organic-rich source-rock interval while largely neglecting the overlying chalky marlstone (Ali Hussein et al., 2015; Alqudah et al., 2015; Plado et al., 2016; Beik et al., 2017; Puura et al., 2017; Giraldo-Gómez et al., 2018; Abu-Mahfouz et al., 2019) or have relied on essentially one-dimensional sections (e.g., Usman et al., 2025; Aljurf et al., 2026). This has limited the assessment of spatial and temporal variability and obscured the key controlling factors. Although those studies generally agree that the long-term sea-level rise promoted the development of upwelling systems along the Neo-Tethyan margins, the mechanisms controlling organic-matter accumulation and preservation, as well as the abrupt transition from bituminous mudstone to chalky marlstone, remain poorly constrained and understood.

To address these gaps, this study integrates new shallow-cored wells, outcrop sections, and published datasets from central and southern Jordan (Alqudah et al., 2015; Plado et al., 2016; Puura et al., 2017; Beik et al., 2019; Messaoud et al., 2025). A regional-scale, high-resolution sequence stratigraphic analysis is combined with refined age constraints based on calcareous nannofossil biozones (CNZ). Newly acquired drill cores from central Jordan cover the entire Maastrichtian succession, including the Campanian-Maastrichtian and Cretaceous-Paleogene (K/Pg) boundaries as well as the MCM chalky marlstone, an interval rarely preserved in outcrop due to its clay-rich, soft-weathering characteristics. Recent studies on the biostratigraphy and tectonostratigraphy of the Upper Cretaceous of Jordan (Kalifi et al., 2025; Messaoud et al., 2025) provide the framework to re-evaluate the controls on MCM Formation deposition, focusing on the relative influence of paleoceanographic processes, particularly upwelling dynamics, on the accumulation and preservation of organic matter. The main objectives of this study of the early Maastrichtian-Danian deposits in Jordan are thus to:

- (i) document regional patterns in lateral facies change (sedimentology and geochemistry) within the MCM Formation, along a 61 km long transect, integrating shallow core data and wireline logs, and constrained by a nannofossil-based time framework;
- (ii) characterise the different orders of cyclicity in the studied interval and investigate the role of variations in productivity, bottom-water energy, and oxygenation; and
- (iii) assess the relative contributions of upwelling-driven and continental run-off driven productivity, and their variations in time and space by integrating information from the literature into our dataset and conclude with the proposition of facies maps and a depositional model.

2. Geologic settings

During the Late Cretaceous, the northwestern edge of the Arabian Plate was part of a shallow carbonate platform along the southern margin of the Neo-Tethys Ocean (Fig. 1a) (Ziegler, 2001; Powell and Moh'd, 2011). Regional tectonics and long-term sea-level rise (on second- and third-order scales) influence the sedimentation patterns and depositional architectures of two major lithostratigraphic groups: the Ajlun and Belqa Groups (Fig. 1b) (Powell, 1989; Andrews, 1992; Powell and Moh'd, 2011). The Ajlun Group, consisting of six formations, primarily comprises open marine carbonates in the distal domain and shallow-water carbonates with fluvial-deltaic sandstones in the more proximal domain, near the paleo-shoreline, which was oriented approximately NE-SW (Flexer and Reymont, 1989; Schulze et al., 2003; Powell and Moh'd, 2011; Chirakal et al., 2025). A new high-resolution age model, based on fossil assemblages (ammonites, foraminifera, and calcareous nannofossil zonation), dates the group from the late Albian to the early Coniacian (Messaoud et al., 2025).

The overlying Belqa group was deposited during the late Coniacian to Eocene (Powell and Moh'd, 2011; Messaoud et al., 2025). It comprises

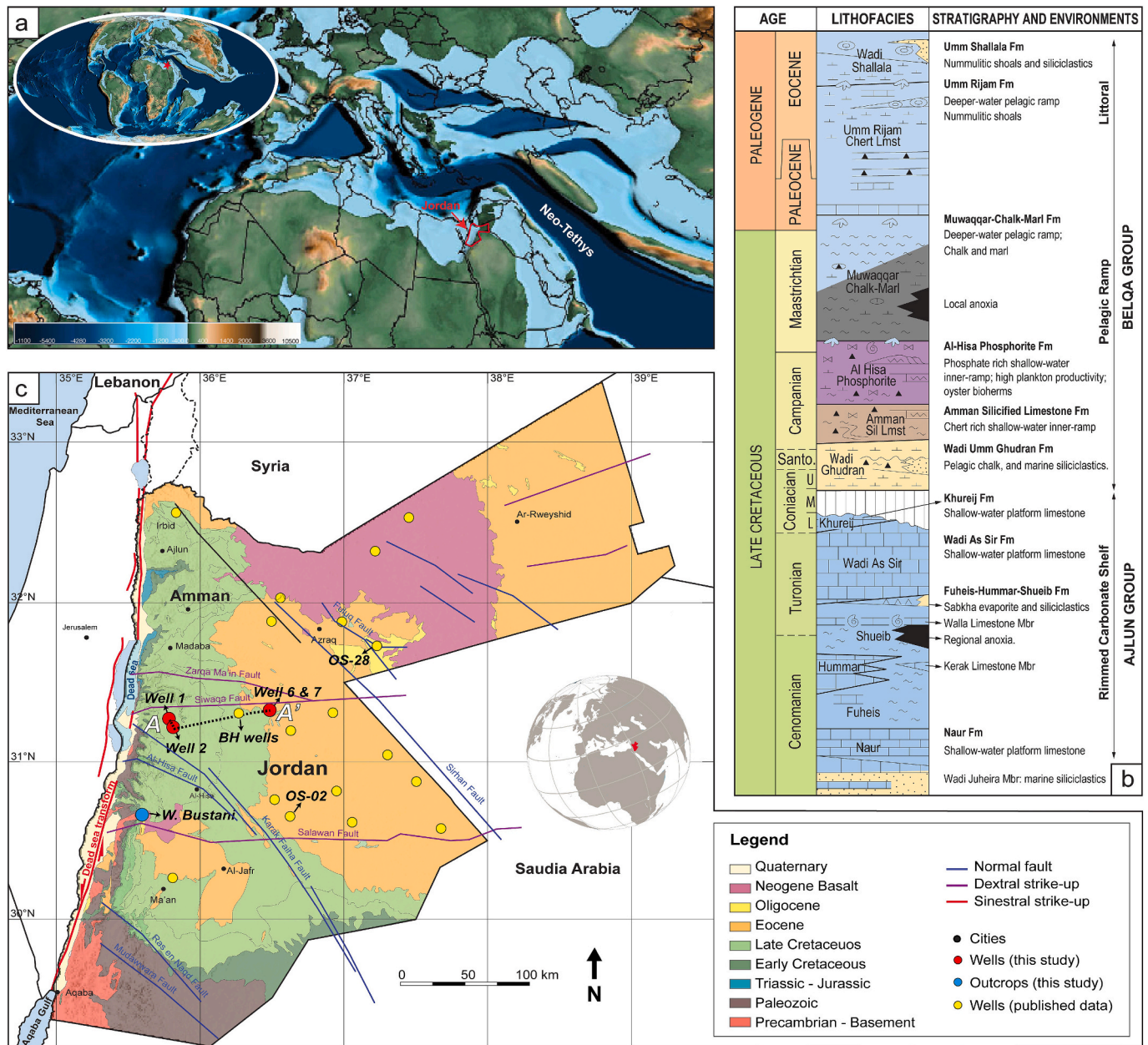


Fig. 1. (a) Paleogeographic reconstruction of the southeastern Neo-Tethys margin during the Late Cretaceous (~70 Ma) (modified from Scotese et al., 2025). (b) Schematic lithostratigraphic column and main depositional sequences of the Ajlun and Belqa groups along the Dead Sea Rift (modified from Powell and Moh'd, 2011). (c) The geological map of Jordan showing the locations of the studied cored wells and outcrop (modified from Kalifi et al., 2025). Published well data (yellow dots) from Alqudah et al. (2015), Plado et al. (2016), and Beik et al. (2019). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

six formations in sequential order (Fig. 1b): (1) Wadi Umm Ghudran (WUG), mainly chalk, chert and sand facies; (2) Amman Silicified Limestone (ASL), rich in cherts and chalky marlstone; (3) Al Hisa Phosphorite (AHP), mainly intercalations of phosphatic marlstone, chert and oyster mounds deposited in shallow water settings; (4) Muwaqqar Chalk Marl (MCM), which is the focus of this study, is generally composed of bituminous chalk and marly facies; (5) Umm Rijam Chert (URC), abundant chert beds intercalated with marly limestone; (6) Wadi Shallalah Chalk (WSC), characterised by chalk and marls interbedding (Andrews, 1992; Powell and Moh'd, 2011). The thickness of the Belqa Group is highly variable across Jordan, ranging from 500 m in the central and northern parts to 100 m in the south (Kalifi et al., 2025).

During the early Maastrichtian, a phase of rapid transgression and high productivity led to the deposition of the organic-rich MCM

Formation, covering a large portion of Jordan (Powell and Moh'd, 2011). The lower part of the MCM Formation is predominantly composed of phosphatic-rich marls and bituminous chalk, while white chalky marls characterise the upper part. The TOC content in the organic-rich beds ranges from 3 to 30 wt%, and is characterised by kerogen Type II- and high sulfur Type II (IIS), sulfur content between 2 and 12 wt%, and Tmax values ranging from 389 to 424 °C (%Ro < 0.6), indicating a thermally immature source rock (Hakimi et al., 2016; Abu-Mahfouz et al., 2019; Usman et al., 2024). The K/Pg boundary hiatus has been identified within the MCM Formation (Messaoud et al., 2025). The top of the formation is marked by high bioturbation (*Thalassinoides*), abundant shallow marine fauna, and the first appearance of the hard, chalky limestone of the URC Formation (Powell and Moh'd, 2011; Beik et al., 2017).

During the Late Cretaceous, the margins of the Arabian Plate became tectonically active due to a major plate tectonic reorganization initiating the closure of the Neo-Tethys Ocean (Ricou, 1994; Vaughan and Searrow, 2003; Agard et al., 2006; Matthews et al., 2012). In intra-plate settings, this phase induced both compressional and extensional tectonic regimes along the northern Arabian Plate (Guiraud and Bosworth, 1997; Litak et al., 1997; Kuss et al., 2000; Guiraud et al., 2005a; Kalifi et al., 2025). Compressional deformation (e.g., Syrian Arc) affected only the northernmost part of Jordan during the Coniacian to early Campanian (Mikbel and Zacher, 1986; Kalifi et al., 2025). In contrast, the extensional/transensional deformation was expressed by the (re)-activation of the NW-SE trending Azraq-Hamza graben system, with tectonic activity reaching a maximum during the early Campanian (ASL Formation) and declining significantly into a sag phase, characterised by subtle differential subsidence during the late Campanian to Eocene (AHP, MCM, and URC formations) (Fig. 2) (Kalifi et al., 2025).

3. Materials and methods

3.1. Materials

The Maastrichtian interval was investigated using four cored wells from central Jordan and one outcrop section in southwest Jordan (Fig. 1c; Table 1). Well 6, located in the Attarat Um Ghudran area, was selected as the reference well as it is a continuous succession that covers almost the entire Maastrichtian interval (only small sections are missing), including both the organic-poor and organic-rich lithofacies of the MCM Formation. The cored interval is approximately 140 m thick and was sampled at high resolution (every 0.5 m) and logged at centimeter scale for texture, fabric, grain types, sedimentary structures, trace fossils, and bioturbation index (BI) (Lazar et al., 2015).

Wells 1 and 2, drilled in the Al-Lajjun area (Fig. 1c), are integrated into the 2-D transect (A-A'). Both well cores include the Campanian-Maastrichtian boundary (Messoud et al., 2025) and were slabbled and analyzed. A detailed geochemical description of Well 2 was previously published by Usman et al. (2025).

The Wadi Bustani outcrop is located along the uplifted escarpment on the eastern flank of the Dead Sea Transform, in southern Jordan (Fig. 1c). The MCM Formation is here about 200 m thick and was logged

as a continuous section at the meter scale.

An overview of the collected data is provided in Table 1 and includes hand and powder samples, thin sections, photographic records, natural gamma-ray (NGAM), as well as spectral gamma-ray (SGR) logs acquired using a handheld device (Radiation Solutions Inc. RS230 BGO Super-Spec).

We also incorporated published data from cored wells located adjacent to the Attarat Um Ghudran, Azraq-Hamza, and Jarf area (Fig. 1c), reported by Alqudah et al. (2015), Plado et al. (2016), and Beik et al. (2019). These wells provide wireline petrophysical logs, biostratigraphic constraints, and geochemical data, which were used to calibrate core depths and establish a robust basin-scale correlation.

3.2. Methods

3.2.1. Petrographic analysis

Thin sections were studied with a LEICA DM 750P polarizing light microscope at different magnifications (2.5×, 4×, and 10×) and digitized with the ZEISS AxioScan 7 slide scanner for enhanced image range acquisition. The thin sections were dyed with an Alizarin Red solution on one half and impregnated with blue epoxy to aid in descriptions, including grain type, lamination, matrix, bioturbation index, fractures, and diagenetic features. Carbonate rocks were classified using the guidelines proposed by Dunham (1962) and with some modifications based on the dominant component, such as grain type or main mineralogy (Flügel and Munnecke, 2010; Lazar et al., 2015).

3.2.2. Biostratigraphic analysis

A total of 90 samples from wells 1, 2, and 6 were analyzed for calcareous nannofossils using the standard smear-slide preparation technique (Bown and Young, 1998). Microscopic examination focused on identifying age-diagnostic taxa to establish biostratigraphic constraints for the uppermost Campanian to the Danian interval and emphasized biozonal assignments to estimate sedimentation rates and locate major stratigraphic boundaries. Cretaceous intervals were assigned using the UC^{TP} (Tethyan Province) biozonation of Burnett (1998), with updates by Thibault (2016), while Danian strata follow the NP zonation of Martini (1970). Taxonomic concepts adhere to Perch-Nielsen (1985) and Bown and Young (1998). Biozonal boundaries were

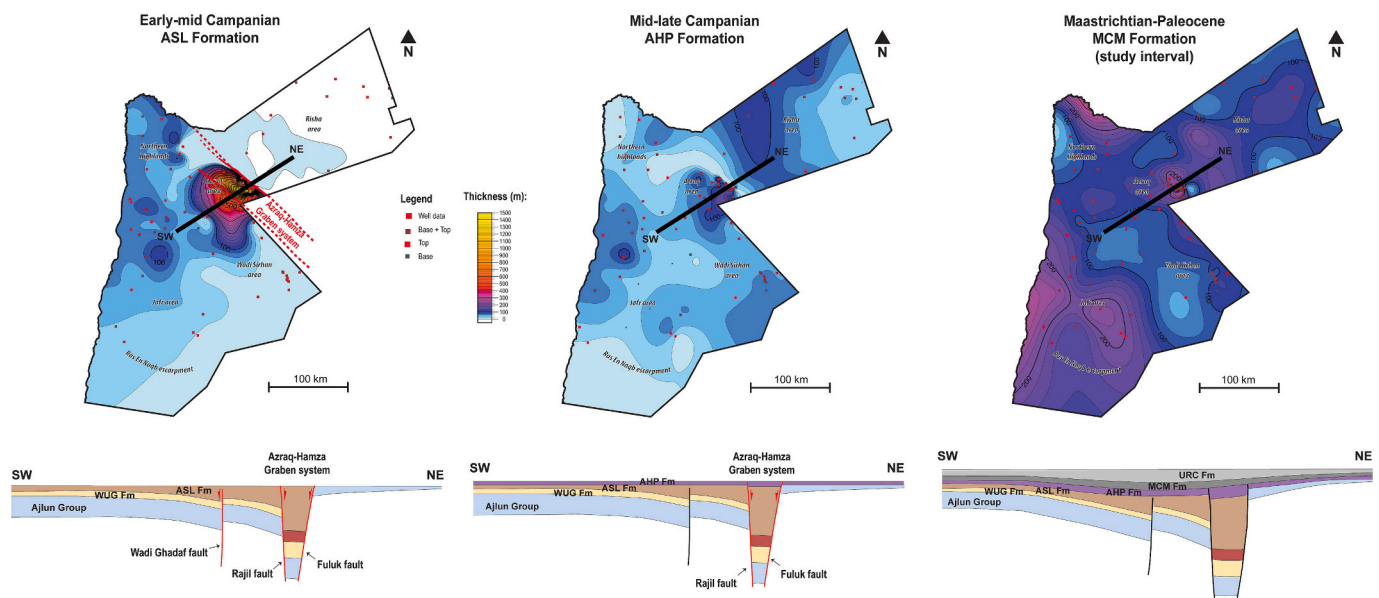


Fig. 2. Thickness maps and schematic cross-sections showing the tectonostratigraphic evolution of central Jordan from the early Campanian to the Paleocene, modified from Kalifi et al. (2025). The Azraq-Hamza Graben system reached maximum activity during deposition of the ASL Formation, followed by a progressive reduction in tectonic activity during deposition of the AHP Formation and transition to a broad sag basin with only minor differential subsidence during deposition of the Maastrichtian-Paleocene MCM Formation.

Table 1

Overview of the data collected in this study.

Data processed in this study								
Name	Thickness (m)	Well logging	Biostratigraphy	Thin section	Geochemical analysis			
					XRF	XRD	ICP-OES	TOC
Well 1	20	SGR	CN biozonation	–	–	33	–	35
Well 2	71	SGR	CN biozonation	15	–	–	–	–
Well 6	139	NGAM	CN biozonation	24	119	–	48	275
Well 7	134	SGR	–	–	–	–	–	56
			CN biozonation					
Wadi Bustani	210	–	(Messaoud et al., 2025)	19	–	–	–	17
Total				58	119	33	48	383
Data compiled from published sources								
Well name	Thickness (m)	Well logging	Additional Data		Reference			
Well 2	71	SGR	TOC, XRD, ICP-OES, Thin section		Usman et al. (2025)			
BH1	117	NGAM	General lithological description, U (ppm), P ₂ O ₅ (%)		Plado et al. (2016) and Puura et al. (2017)			
BH3	104	NGAM	General lithological description, U (ppm), P ₂ O ₅ (%)					
BH5	139	NGAM	General lithological description, U (ppm), P ₂ O ₅ (%)		Alqudah et al. (2015) and Beik et al. (2019)			
OS02	140	–	CN biozonation, TOC %, and CaCO ₃ %					
OS28	90	–	CN biozonation, TOC %, and CaCO ₃ %					

– Not measured.

defined using the First Occurrences (FO) and Last Occurrences (LO) of key taxa.

3.2.3. Geochemical analysis

A total of 505 samples were crushed and ground to a particle size of $\leq 10 \mu\text{m}$ for bulk analysis using an agate mortar to ensure homogenization and minimize inconsistencies (Craigie, 2018). Organic and inorganic geochemical analyses were performed at the Core Lab facilities at King Abdullah University of Science and Technology (KAUST), Saudi Arabia.

Programmed pyrolysis analysis was conducted using the Rock-Eval 7S (Standard Analyzer Unit, Vinci Technologies) to measure its TOC content and evaluate its thermal maturity. TOC values are given in weight percent (wt%). The calcium carbonate content equivalent was calculated from the mineral carbon content parameter (MinC) from the Rock-Eval 7S (Behar et al., 2001; Jiang et al., 2017). Assuming that the inorganic carbon fraction resides as calcium carbonate (CaCO₃), the following stoichiometric formula was applied: $\text{CaCO}_3 = \text{MinC} \times 8.333$.

The mineralogical composition of selected samples from Well 1 were analyzed using the X-ray diffraction (XRD) technique. Powder XRD patterns (2θ ranging from 5° to 70°) were obtained in a Bruker D8 Advance X-ray diffractometer equipped with Cu K α radiation ($\lambda = 1.54 \text{ \AA}$). The mineralogical fractions were semi-quantified in the EVA V4 software, which includes the Powder Diffraction File (PDF-4) database.

X-ray fluorescence (XRF) and Inductively Coupled Plasma Optical Emission Spectroscopy (ICP-OES) analyses were performed to determine the elemental composition. Major and minor element concentrations, expressed in wt%, were measured with a Bruker S8 TIGER Wavelength Dispersive X-Ray Fluorescence (WDXRF) spectrometer equipped with the GEO-QUANT analytical method designed for geological samples.

For trace element analysis, the experiment was conducted in two phases. In the first phase, powdered samples ($\sim 0.2 \text{ g}$ each) were digested using the inverse aqua regia method, which involved a mixture of nitric acid (HNO₃) and hydrochloric acid (HCl) in a 3:1 volumetric ratio (Xu et al., 2012). Acid digestion was performed using an Ultra-WAVE system (Milestone). The resulting solutions were diluted with Milli-Q water to a final volume of 10 mL. The trace element concentrations were measured with an Agilent 5900 ICP-OES instrument and calibrated with two certified reference materials (CRMs). Concentrations are reported in parts per million (ppm). In order to evaluate redox-sensitive elements, the enrichment factor (EF) was calculated using the composition of the Upper Continental Crust (UCC), as follows: $X_{\text{EF}} = (X/\text{Al})_{\text{sample}} / (X/\text{Al})_{\text{UCC}}$ (Taylor and McLennan, 1985).

4. Results

In this section reference Well 6 (Fig. 1c) is presented in detail, including facies descriptions, high-resolution biostratigraphic, and chemostratigraphic analyses (Fig. 3, Supplementary Materials). Based on similar characteristics, a lithostratigraphic subdivision is presented. Subsequently, a regional well log correlation is constructed to evaluate lateral and vertical sedimentological and geochemical facies variations.

4.1. Sedimentary facies classification

Thin-section descriptions follow the classification of the MCM organic-rich interval in Well 2 by Usman et al. (2025), complemented by new observations from this study. Nearly all samples exhibit a fine-grained matrix, with common bioclastic components (foraminifera, ooids, peloids, bivalves, and fish fragments) classified primarily as wackestone to packstone, with localized bioclastic grainstone beds. A total of 17 microfacies, grouped into 10 lithofacies, are distinguished (Table 2). Lithologies are defined as phosphatic bituminous marlstone, bituminous marlstone, dolomitic limestone, chalky marlstone, and chert.

4.2. Lithostratigraphic units

Based on variations in sedimentary features, organic and inorganic components, Well 6 has been subdivided into five lithostratigraphic units (Fig. 3). These are:

4.2.1. Unit 1. Phosphorite (uppermost AHP)

The basal lithostratigraphic unit is 7.65 m (139.45 to 131.8 m depth) thick and characterised by upward coarsening lithofacies with moderate to high organic carbon content (TOC $< 10 \text{ wt\%}$). Although carbonate-dominated, this interval contains elevated phosphate concentrations, reaching up to 14.10 wt% P₂O₅ (Fig. 3), as reflected by the common occurrence of phosphatic-coated grains and phosphatic bioclasts (Fig. 4). The phosphorite Unit 1 is composed of laminated phosphatic marls (LF1) and highly bioturbated phosphorite (LF2) interbedded with decimeter-thick beds of carbonate and chert concretions and siliceous mudstones. The top of the phosphatic-rich unit at 131.8 m marks the lithostratigraphic boundary of the AHP and MCM formations.

4.2.2. Unit 2. Organic-rich carbonate (lower MCM)

This unit is approximately 55 m thick (131.8 to 76.95 m depth). The bulk composition is generally characterised by high CaCO₃ contents (15–93 wt%) and organic carbon (TOC up to 23.2%); however, some

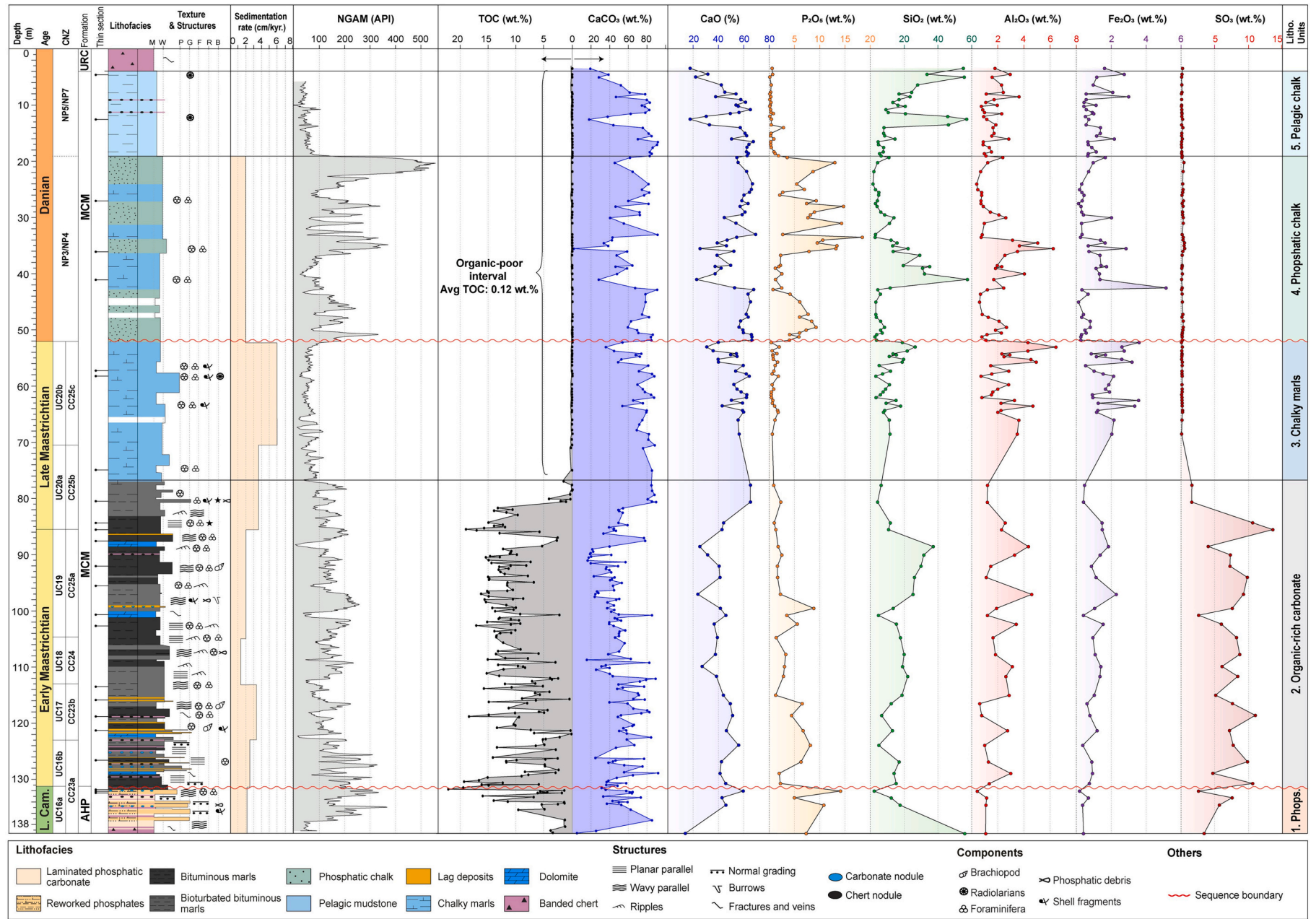


Fig. 3. Detailed lithostratigraphic and biostratigraphic column of Well 6, including gamma-ray log (API), TOC (wt%), calculated CaCO_3 (wt%), selected major oxide (wt%) and calcareous nannofossil biozones (CNZ).

Table 2

Summary of the microfacies and lithofacies classification for reference Well 6.

LF#	Sedimentary Facies	Microfacies	Lithology	Bioclastic components	Non-bioclastic components	Main components and Sedimentary structures	Description
LF1	Laminated phosphatic marls	F1a. Thin laminated peloidal packstone	Phosphatic marls	Large bioclast fragments (>1 mm) and planktonic forams	Phosphate-coated grains (>2 mm), carbonate nodules (5 cm height)	Wavy and lenticular lamination. Normal grading is observed in upward transitions from mudstone to packstone. Bioturbation (BI = 3)—dominance of <i>Thalassinoides</i> (Th)	Peloidal-rich mudstone to packstone embedded in a carbonate-rich matrix. Wavy lamination and normal grading are observed in upward transitions from mudstone to packstone. Strongly bioturbated peloidal phosphates. The main components are peloidal and phosphatic ooids. Chaotic fabric with poorly sorted carbonate and phosphatic-coated grains embedded in a carbonate or silica-rich matrix
LF2	Reworked/ bioturbated phosphates	F2b. Peloidal-oolitic packstone to grainstone	Granular phosphorites	Large pelagic bioclast fragments (>1 mm).	Phosphate-coated grains (>2 mm), phosphatic ooids, peloids, and carbonate nodules (5 cm height)	Sedimentary structures or bedding are not recognized due to the strong bioturbation (BI = 5)—dominance of <i>Thalassinoides</i> (Th) and <i>Chondrites</i> (Ch) burrows	Light to dark brown foraminiferal packstone to grainstone with burrow-mottled texture.
LF3	Bioturbated organic-rich wackestone to packstone	F3a. Organic-rich foraminiferal packstone F3b. Organic-moderate bioclast packstone to grainstone F4a. Thinly laminated OM-rich mudstone F4b. Organic-rich foraminiferal wackestone	Bituminous marlstone	Mainly foraminifers, common fishbone debris, and rare echinoderm spines	Peloids and coated grains	Normal grading, wavy parallel lamination with some local lenticular fabric (e.g., foraminiferal lags)	Dark brown foraminiferal wackestone to packstone. Rare bioturbated.
LF4	Laminated organic-rich wackestone	F4c. Wavy to lenticular matrix OM-rich foraminiferal packstone F4d. Si-rich black laminated mudstone	Bituminous chalk	Mainly foraminifers and rare fishbone debris	Rare peloids	Parallel to wavy laminated with some local lenticular fabric (e.g., foraminiferal lags)	5–15 cm thick bioclastic grainstone beds. The coated grains and bioclasts are densely packed into a crystalline matrix composed of sparite, silica cement, or peloidal cement. Whitish to reddish foraminiferal packstone to grainstone. The matrix is composed of fine carbonate lime
LF5	Lag deposits	F5a. Reworked phosphatic grainstone F5b. Peloidal bioclastic-rich grainstone	Bioclastic limestone	Common mollusk fragments, halimeda fragments, benthic forams, and fishbone debris.	Coated black grains (2–4 mm) and peloids are poorly sorted and poorly rounded	structureless, reworked grains well cemented	Light yellow to whitish color lime mudstone. The fabric is well-sorted. IntraP porosity is common in foraminifer chambers
LF6	Chalky marls	F6a. Foraminiferal packstone to grainstone	Chalky marls	Mainly foraminifers, and large bioclast fragments (>2 mm).	N/A	No clear lamination and heavily packed foraminifera grains	Whitish foraminiferal packstone to grainstone. The matrix is composed of fine carbonate lime and peloids. Fine-crystalline, nonplanar dolomite, and medium-crystalline, planar-s dolomite. The matrix is cloudy, with only a few calcareous fossils identified from the precursor sediment
LF7	Pelagic mudstone	F7a. Planktonic foraminiferal wackestone to packstone F7b. Radiolarian packstone	Chalk	Radiolarian and planktonic foraminifers	N/A	No clear-to-thinly laminated. Chert nodules (3 cm)	Massive and hard black chert, highly fractured, with silica-filled veins that are always distinct
LF8	Phosphatic chalk	F8a. Peloidal/ foraminiferal packstone to grainstone	Chalk	Mainly foraminifers	Well-sorted phosphatic grains	No clear lamination and heavily packed foraminifera grains	
LF9	Dolomitic limestone	F9a. Dolomite nonplanar F9b. Dolomite planar-s	Dolostone	Rare foraminifers	N/A	Fine to medium tight crystalline dolomite	
LF10	Bended chert	F10a. Highly fractured chert	Chert	N/A	N/A	Homogeneous or structureless	

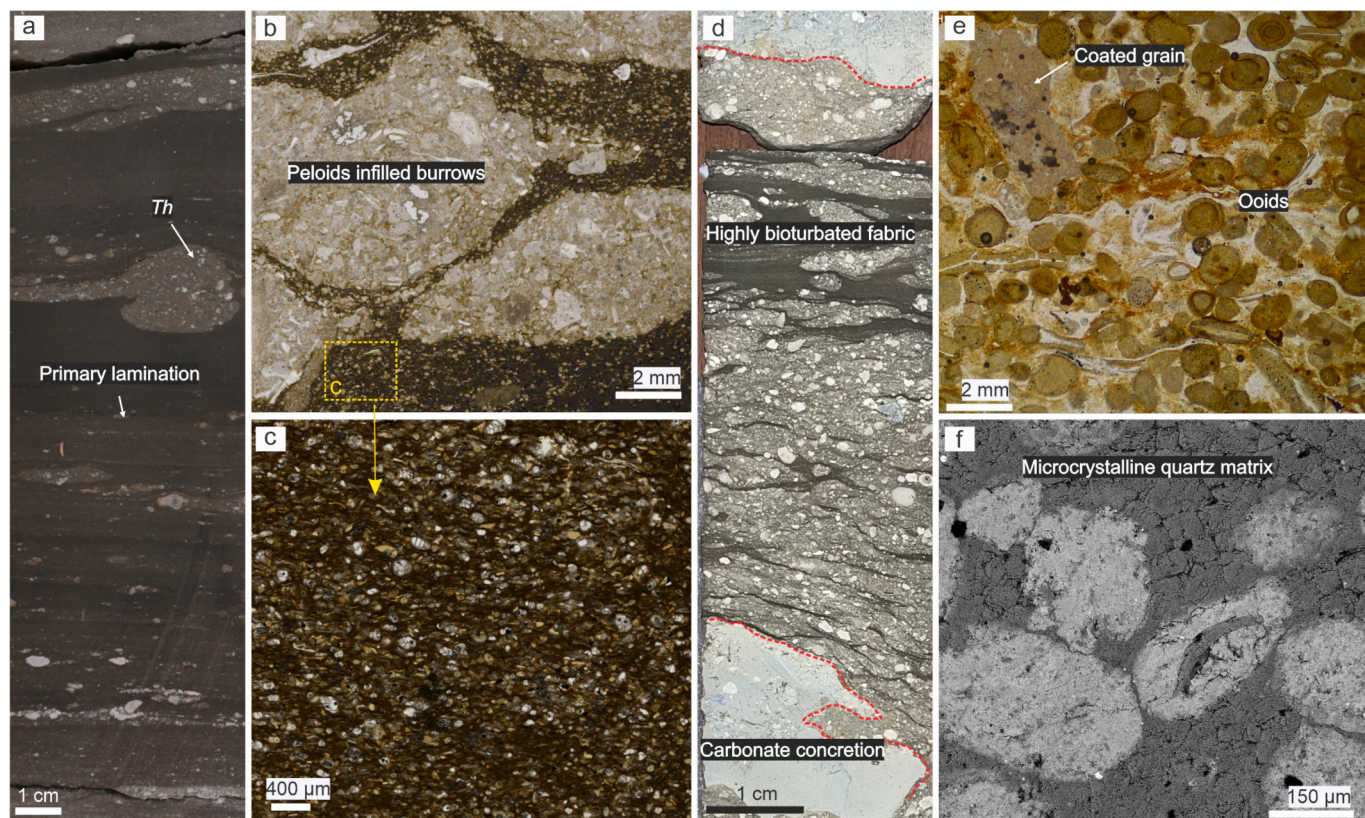


Fig. 4. Key features of the phosphorite Unit 1. (a) Core slab photograph of sub-horizontal burrows common in LF1. Note that primary lamination is partially preserved. Well 2, depth 64.8 m. (b) Photomicrographs in plane-polarized light of the burrow infilled material, which consists of poorly sorted and poorly rounded carbonate and phosphatic grains. (c) Zoom-in of the host sediment described as wavy-laminated phosphatic-rich packstone embedded into a peloidal and argillaceous matrix with lens structures. (d) Core slab photograph showing a burrowed carbonate concretion and the strong bioturbated fabric in LF2. Horizontal and vertical burrows are abundant in the lower part of the lithofacies, becoming less abundant upwards. Well 1, depth 17.5 m. (e) Photomicrographs in plane-polarized light of the burrows infilled material, such as phosphatic ooids and phosphatic coated grains (> 2 mm). Well 2, depth 68.8 m. (f) BSE image of the cemented matrix (darker gray) and phosphatic ooids (lighter gray). Th = *Thalassinoides* burrow.

intervals contain moderately high SiO_2 concentrations (e.g., 97 to 87 m depth), reaching up to 37.20 wt% (Fig. 3). Minor concentrations of element oxides include SO_3 (average 7.35 wt%), P_2O_5 (average 3.41 wt%), and Al_2O_3 (average 2.14 wt%). The dominant lithology is bituminous marlstone with an argillaceous-rich matrix (Fig. 5c and g), and bituminous chalk (Fig. 5f), which constitute nearly 80% of the unit and are classified as either strongly bioturbated (LF3) or weakly bioturbated (Fig. 5a) to laminated (LF4) (Fig. 5d). The bituminous chalky marlstone are interbedded with four dolomitic limestone beds (LF9), each decimeter-thick (Fig. 5i) and lag deposits (LF5) (Fig. 5j and k).

4.2.3. Unit 3. Chalky marls (middle MCM)

The light colored, chalky marls have a thickness of 25 m (76.95 to 52 m depth) and consist primarily of carbonate-rich sediments, with average CaCO_3 contents of 70.5 wt% (Fig. 3). The most common lithofacies in Unit 3 are foraminiferal packstone (LF6) with large bioclast fragments (Fig. 6b) and planktonic foraminiferal mudstone to wackestone (LF7) with a coccolith-rich (calcareous nannofossils) matrix (Fig. 6d). Toward the top of this unit, a subtle decrease in carbonate content coincides with the occurrence of reddish-colored beds (Fig. 6a) and the increase in detrital proxy elements such as SiO_2 and Al_2O_3 , where the concentrations reach maximum values of 26.5 wt%, 6.4 wt%, and shows a strong positive linear correlation (Fig. 7a, $R^2 = 0.91$). The top surface of Unit 3 is interpreted as the K/Pg boundary (see below).

4.2.4. Unit 4. Phosphatic chalk (upper MCM)

This unit is 33 m thick (19 to 52 m depth) with high CaCO_3 concentrations (63.5 wt%). SiO_2 locally reaches a maximum of 57.40 wt%.

In contrast to the chalky marls of Unit 3, the SiO_2 concentrations show a weak correlation with Al_2O_3 (Fig. 7a $R^2 = 0.23$) and are primarily associated with chalcedony precipitation within foraminiferal tests (Fig. 6g). The base of Unit 4 is composed of marly chalks from LF6, which transition upwards into chalks and “sandy” bioturbated phosphatic chalks from LF8. The peloidal-rich facies (Fig. 6f) coincides with elevated P_2O_5 concentrations (locally reaching a maximum of 18.40 wt%) and the highest peaks of gamma-ray signal.

4.2.5. Unit 5. Pelagic chalk (uppermost MCM)

This unit is 14 m thick (19 to 5 m depth) and consists of white chalks (Fig. 6e) intercalated with chert nodules (2–3 cm in length). The dominant lithofacies are foraminiferal mudstone and radiolarian mudstones (LF7) (Fig. 6h). SiO_2 concentrations increase progressively upward, exceeding 50 wt% near the top. Silica occurs as silicified fossils, particularly planktonic foraminifera, as well as microcrystalline quartz overgrowths within the matrix (Fig. 6i). The top of Unit 5 is a 1-m-thick fractured chert bed interpreted as the base of the overlaying URC Formation.

4.3. Biostratigraphy and sedimentation rate

A total of nine calcareous nannofossil biozones (CNZ) are distinguished in Well 6 (Fig. 3), which provide an excellent high-resolution biostratigraphic framework for the lithostratigraphic units:

The phosphorite Unit 1 is assigned to the UC16a nannofossil subzone (late Campanian). The top of UC16a nannofossil subzone is marked by a clear litho- and biostratigraphic transition coinciding with the LO of

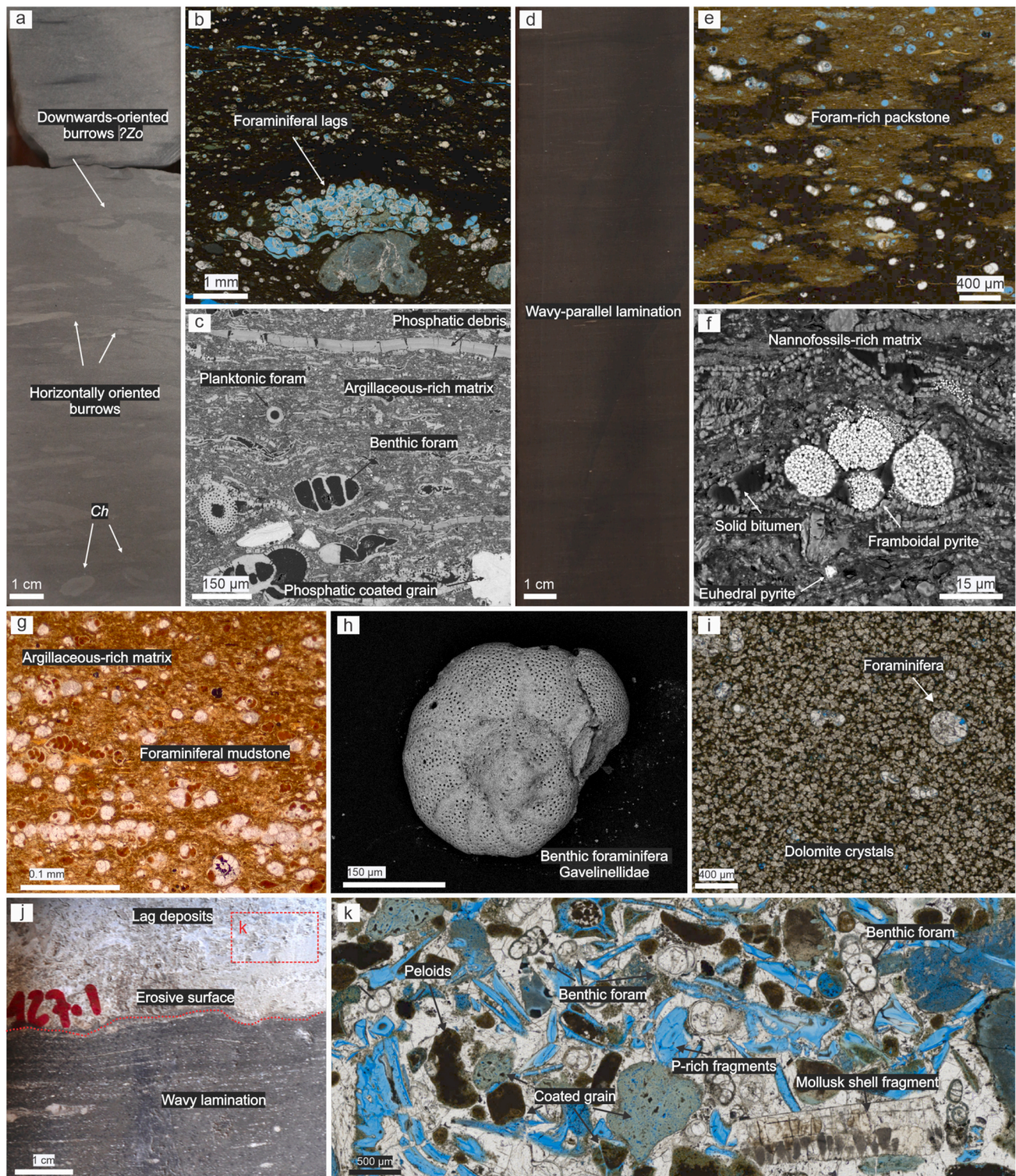


Fig. 5. Key lithofacies features of the organic-rich carbonate Unit 2. (a) Core slab photograph of common fabric in LF3, Well 2, depth 24.8 m (b) Photomicrographs in plane-polarized light of the matrix and lenticular fabric composed mainly of foraminifera, Well 6, depth 126.6 m. (c) BSE image. Common benthic foraminifers and phosphatic components, such as fish bones and coated grains embedded in the argillaceous-rich matrix. (d) Core slab photograph of the parallel to wavy laminated bituminous mudstone, Well 2, depth 52 m. (e) Photomicrographs in plane-polarized light. Planktic forams are the dominant bioclastic component in LF4, common benthic forams, and scarce fish bone detritus. Well 6, depth 131.8 m. (f) BSE image of the fine to medium silty matrix, typically composed of nannofossil-rich fragments, disseminated organic matter, and pyrite. (g) Foraminiferal mudstone with argillaceous matrix in LF3, Well 2, depth 50 m. (h) SEM image of common benthic foraminifera identified in Unit 2, Gavelinellidae?Gavelinella. (i) Tight planar-s dolomite within Unit 2. The fabric is almost completely replaced by cloudy dolomite grains with some preserved sedimentary precursors. Well 6, depth 87.5 m. (j) Core slab photograph of lag deposits (LF5) cutting the underlying sediments with an erosive base, a common feature at the base of Unit 2. Well 6, depth 127.1 m. (k) Photomicrographs in plane-polarized light zoomed in on LF5 with large mollusk fragments (partially silicified), abundant benthic foraminifera, and coated grains embedded in a calcitic crystalline matrix. Ch = *Chondrites* burrow, Zo = *Zoophycos* burrow.

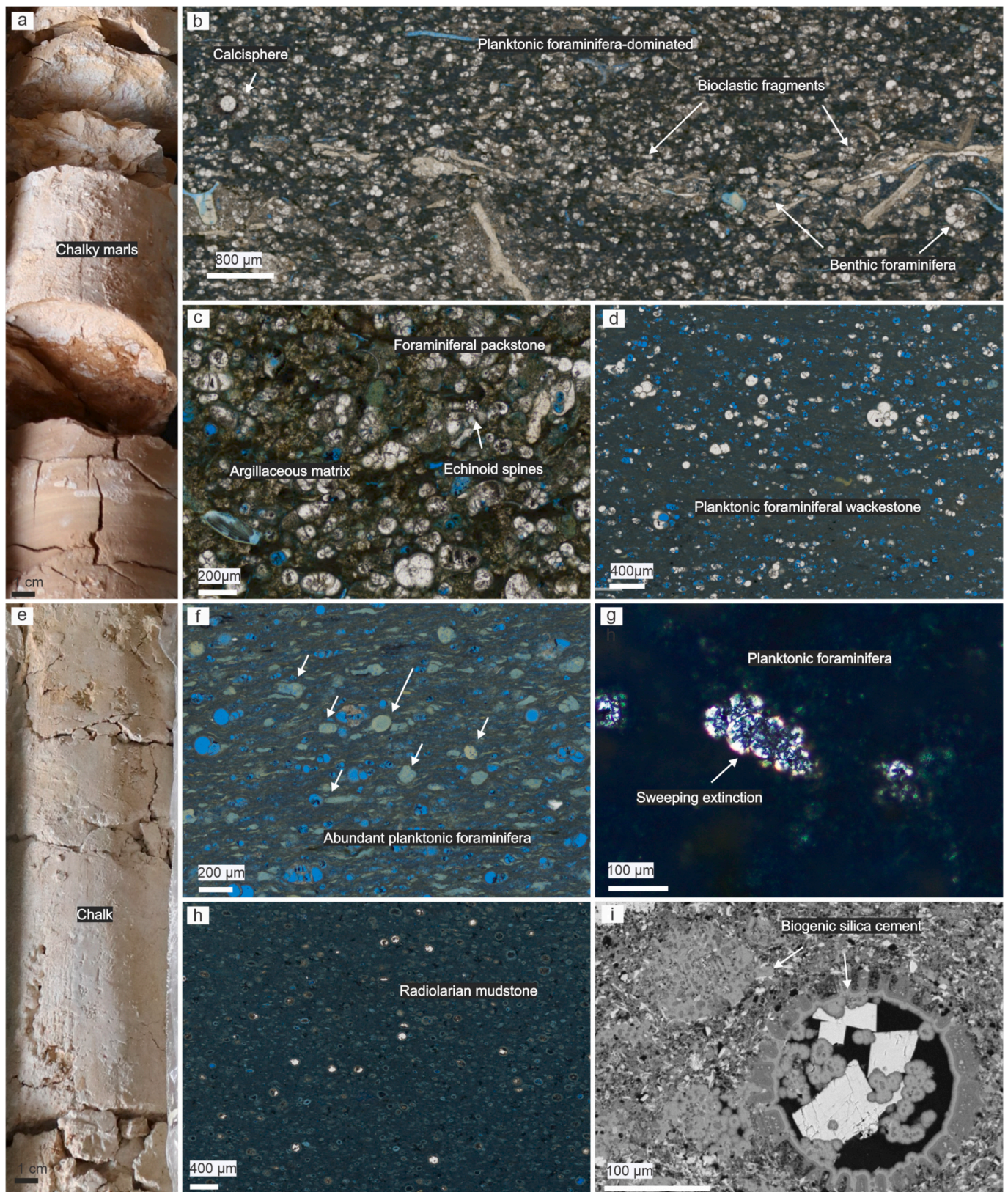


Fig. 6. Key lithofacies features of chalky marls interval (a) Core photograph of the chalky marl Unit 3 near the K/Pg boundary. Well 6, depth 52.5 m. (b) Foraminiferal packstone (LF6). The main components are planktonic foraminifera, which appear uncompacted, replaced by diagenetic calcite cement, and common large bioclastic fragments. Well 6, depth 58.2 m. (c) Densely packed foraminiferal packstone (LF6) within Unit 2. Well 6, depth 80.5 m. (d) Planktonic foraminiferal mudstone to wackestone (LF7) within Unit 3. Well 6, depth 57.3 m. (e) Core photograph of the pelagic chalk Unit 5. Well 6, depth 6.5 m. (f) Peloidal (white arrows) and foraminiferal-rich packstone (LF8) within Unit 4. Well 6, depth 27 m. (g) Cross-polarized light reveals the typical radial character, or “sweeping” extinction of chalcidony. Well 6, depth 41 m. (h) Radiolarian mudstone (LF7) within Unit 5. Well 6, depth 12.5 m (i) Zoom-in SEM image radiolarian mudstone. Radiolarian tests are partially filled with calcite and biogenic silica overgrowth (white arrows).

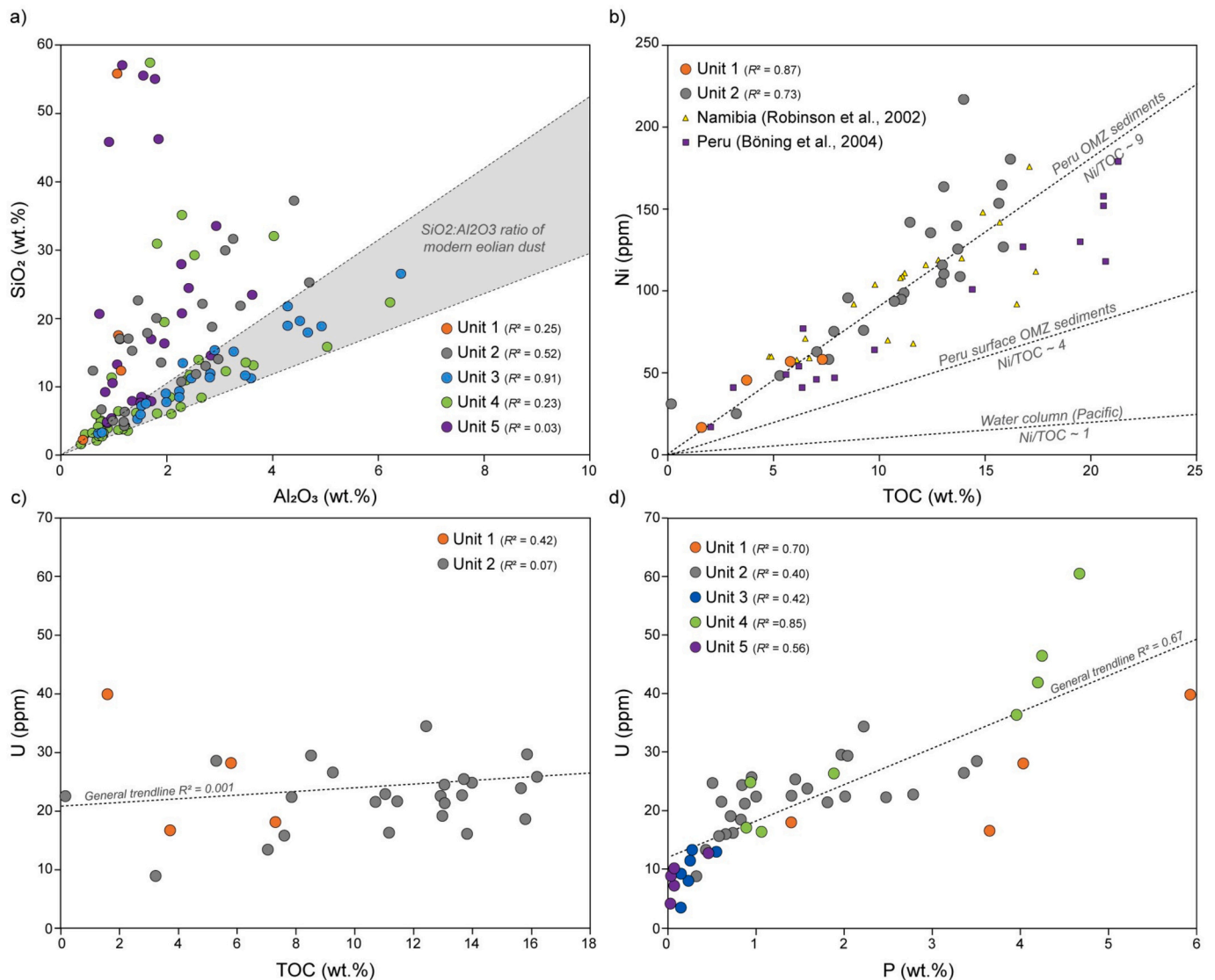


Fig. 7. Geochemical cross-plots for selected samples in Well 6. (a) Relationship between SiO_2 and Al_2O_3 by lithostratigraphic unit. All samples from Unit 3 (chalky marls) fall within the range of ratios characteristic of modern pelagic eolian dust (gray shading) (modified from Ikeda et al., 2017). (b) Bulk Ni concentration versus TOC content showing a strong positive correlation. Ni/TOC ratios from Well 6 are compared with modern sediments from the upwelling settings of Namibia (Robinson et al., 2002) and Peru (Böning et al., 2004). The dashed lines denote the Ni/TOC ratio characteristic of Peru margin sediments (after Böning et al., 2015). (c) A moderate positive correlation between Ni and TOC is observed in the phosphorites of Unit 1. Units 3, 4, and 5 are not plotted due to their very low TOC contents (<0.1 wt.%). (d) Relationship between U and P by lithostratigraphic unit, showing a strong positive correlation in the phosphatic-rich facies of Units 1 and 4.

Uliplanarius trifidus at 131.8 m level and used as a marker of the Campanian-Maastrichtian boundary. Unit 1 spans roughly 200 kyr and has a minimum average sedimentation rate of 2 cm/kyr.

The organic-rich carbonate Unit 2 spans the UC16b to lower UC20a (early Maastrichtian) nanofossil subzone, with a duration of approximately 2.5 myr. Calculated sedimentation rates vary substantially across the unit, ranging from 1.3 cm/kyr in UC18 to 2.5 cm/kyr in UC20a.

The chalky marl Unit 3 is assigned to the middle UC20a-UC20b nanofossil subzones (late Maastrichtian), spanning approximately 800 kyr. The average minimum sedimentation rate of 6 cm/kyr is significantly higher than the organic-rich carbonate interval.

The K/Pg boundary is marked as a stratigraphic hiatus of approximately 2 myr, evidenced by the absence of the interval from the latest nanofossil subzone UC20b to the Paleocene nanofossil Zone NP3 in Well 6.

The phosphatic chalk Unit 4 spans the preserved portion of calcareous nanofossil zones NP3 to NP7 in the studied core, and represents an approximate duration of 1500 kyr. Sedimentation rates decrease from 6

cm/kyr to 2 cm/kyr through this interval. However, the complete NP3–NP7 succession is not preserved, and the missing portion may reflect erosion, a stratigraphic hiatus, condensation, or a combination of these processes. Therefore, the calculated duration refers only to the preserved section recovered in the core. The duration of the overlying pelagic chalk Unit 5 cannot be reliably constrained from the available data and is therefore left open.

In wells 1 and 2 (Fig. 1c), the Maastrichtian MCM interval has been assigned to nanofossil subzones UC16b to lower UC20b based on the LO of *U. trifidus* and FO of *Micula murus*. The Paleocene MCM interval is absent in the Al-Lajjun area.

4.4. Geochemical characterization

4.4.1. Major and trace element concentrations

Selected major and trace elements are plotted against the lithostratigraphic units of Well 6 in Fig. 7 and listed in the supplementary material (Tables S1 and S2). The bulk composition is predominantly

carbonate-rich, with average CaO contents of 50.1 wt% and CaCO₃ contents of 57.5 wt%. However, some intervals contain significant phosphatic and silica-rich fractions, particularly in units 4 and 5 (Fig. 3). Similar mineralogical compositions are observed in Well 1 (Table S3) and in Well 2 reported by Usman et al. (2025). In both wells, calcite is the dominant mineral phase, followed by quartz, apatite, and dolomite. A summary of the XRD results for Well 1 is presented in Table 3.

Enrichment factors (EF) provide an effective assessment of the authigenic enrichment ($X_{EF} > 3$) or depletion of an element (Algeo and Tribouillard, 2009). The elemental concentrations and EF of 48 bulk samples from the Upper Campanian–Danian succession are presented in Table 4 and Fig. 8. Redox-sensitive elements copper (Cu), chromium (Cr), molybdenum (Mo), nickel (Ni), uranium (U), and vanadium (V) in the organic-rich intervals (139 to 76.95 m depth) are enriched by factors from 2.98 to 923.31 relative to the Upper Continental Crust (UCC) (Table 4). Despite its low TOC content, the chalky facies interval (76.9 to 5 m) also exhibits relatively high ratios of trace elements, particularly within the phosphatic chalk Unit 4 (Fig. 8). The Mo/Al ratio reaches up to 172.3, representing a 923-fold enrichment relative to the UCC. The highest Mo values occur in Unit 2, averaging 122.9 ppm (Table 4). In contrast, the organic-poor interval (units 3, 4, and 5) shows lower concentrations, averaging 38 ppm. Cr/Al and Cu/Al show enrichment in all samples with minimum EF of 2.5 and 2.3, respectively, relative to the UCC content. Ni concentrations are also enriched in all measured samples, with a minimum concentration of 16.7 ppm, representing an EF of 14-fold enrichment relative to the UCC (Table 4). Ni is positively correlated with TOC in organic-rich samples, with a coefficient of regression of 0.87 in the phosphorite marls Unit 1 and 0.73 for the organic-rich carbonate Unit 2 (Fig. 7b). Titanium (Ti) concentrations show relatively low variability within the organic-rich facies (149.47 ppm average) but reach a maximum of 364.9 ppm in the chalky marl facies of Unit 3 (Table S2).

The U/Al ratio is enriched in all samples by a factor ranging from 2.87 to 527.10 relative to the UCC, with the highest values occurring in phosphatic-rich facies of units 1 and 4 (Fig. 8). Additionally, U shows weak correlation with TOC ($R^2 = 0.001$, Fig. 7c) but a strong positive correlation with P (Fig. 7d), particularly in the phosphatic-rich facies, including phosphorite marls from Unit 1 ($R^2 = 0.70$) and phosphatic chalk from Unit 4 ($R^2 = 0.85$).

Remarkably, Co/Al and Mn/Al are significantly depleted relative to the UCC in nearly all samples. The Mn/Al ratio in the UCC is 74.62 (Taylor and McLennan, 1985), whereas the highest value recorded in the organic-rich facies is only 37.45. The maximum Co/Al ratio in the same interval is 0.77, considerably lower than the UCC ratio of 2.11 (Fig. 8, Table 4).

4.4.2. Bulk organic geochemistry

A total of 275 samples, covering all lithostratigraphic units in Well 6, were analyzed to characterise their organic matter content (Fig. 3, Table S4). Given the immature nature and the high organic carbon content of these rocks, we categorized TOC content from the Upper Campanian–Danian interval as follows: organic-rich (>10 wt%), organic-moderate (0.5–10 wt%), and organic-poor (<0.5 wt%). Organic matter was observed as bitumen filling foraminiferal chambers and fractures, as well as disseminated throughout the matrix (Fig. 5e and f).

Table 3

A summary of XRD values for the samples from Well 1.

Mineral	Minimum (%)	Maximum (%)	Average (%)
Calcite	48.52	87.86	73.88
Dolomite	0.16	29.29	2.50
Apatite	2.22	21.05	6.17
Clay	2.72	9.73	4.68
Quartz	2.23	25.41	10.04
Pyrite	0.31	4.99	2.59

In the phosphorite Unit 1, TOC values range from 1.44 to 16.80 wt%, with an average of 5.79 wt%, and weak negative linear relationship with CaCO₃ (Fig. 9a). Unit 1 is considered organic-moderate and has good to excellent source-rock quality (Fig. 9b). According to the pseudo-Van-Krevelen diagram (Fig. 9c), samples correspond mostly to kerogen Type II and IIS, which is characteristic of marine environments with sulfur enrichments (Hart and Steen, 2015; Dembicki, 2009).

The organic-rich carbonate Unit 2 represents the Maastrichtian source rock. The interval contains the highest TOC values, reaching up to 23.2 wt%, with an average of 10.5 wt%, and is considered organic-rich. The TOC showed a reasonable inverse linear relationship with CaCO₃ (Fig. 9a). It is an excellent oil-prone source rock, characterised by high Hydrogen Index (HI) values (average 800 mg HC/g TOC) and low Oxygen Index (OI) values (24 mg CO₂/g TOC) (Fig. 9b). The kerogen is classified as Type II and IIS (Fig. 9c), and the calculated vitrinite reflectance (%Ro) reaches up to 0.34%, indicating that this interval is a thermally immature source rock (Tissot and Welte, 1984). While the entire unit displayed elevated TOC values, the interval between 111.6 m and 89 m is particularly enriched in organic carbon “TOC-rich interval” with nearly all samples containing more than 10 wt% TOC (average 12.46 wt%).

The chalky marls, phosphatic chalks, and pelagic chalk units were categorized as an organic-poor interval, with an average content of only 0.12 wt%, and poor hydrocarbon generation potential (Fig. 9b). Rock eval results in Wadi Bustani samples show relatively low TOC values (TOC < 0.21 wt%) (Table S5).

4.5. Well log correlation and orders of cyclicity

In order to investigate the lateral continuity of the observed lithological and geochemical variations, a 61-km-long 2D regional transect (A–A') with a SW–NE orientation has been constructed in central Jordan (Fig. 10). In all wells, the lithostratigraphic units have been identified based on the gamma-ray logs, and where available, facies distribution, TOC content, and the lithological markers (e.g., dolomitic limestone beds). Detailed calcareous nannofossil biostratigraphy obtained from wells 1, 2, and 6 provides essential age-dating constraints.

The regional correlation was flattened at the top of nannofossil subzone UC16a, identified in wells 1, 2, and 6, which corresponds to the Campanian–Maastrichtian boundary. This boundary also marks the top of the phosphorite Unit 1 and is characterised by a strongly bioturbated phosphorite interval with a prominent positive gamma-ray signal making it a reliable and traceable marker across all wells (Fig. 10). Overall, all correlated wells show notable similarities in their organic-matter content and mineralogical composition, yet they display substantial lateral variability in both thickness and stacking patterns.

Usman et al. (2025) identified three depositional cycles within the MCM organic-rich interval in Well 2 (here Unit 2). In this study, the cyclicity, expressed by variations in the gamma-ray logs, lithofacies, and TOC, is traced laterally across multiple wells. The cycles are punctuated by dolomitic limestone beds, some of which have a regional expression, particularly in wells from Attarat Um Ghudran area (Fig. 10). Plado et al. (2016) identified three dolomitic limestone beds in BH-1, BH-3, and BH-5 wells, which were lithostratigraphically correlated with three levels in wells 6 and 7, corroborated by geochemical and petrographic data (Figs. 3 and 5i). The gamma-ray logs exhibit remarkably similar trends in these five longest wells and delineate packages with consistent thickness across the transect. In contrast, wells 1 and 2 are shorter and more condensed and consequently exhibit less clearly expressed log patterns (Fig. 10).

The phosphorite Unit 1 comprises two similar coarsening-upward successions, defined as cycles C₁ and C₂. Each cycle represents a relatively short-term duration of ~100 kyr and a relatively low sedimentation rate (2 cm/kyr). Both cycles begin with fine-grained, laminated phosphorites at the base, with low-diversity trace-fossil assemblages dominated by horizontally oriented burrows (Fig. 4a) associated with a

Table 4Trace element concentrations and enrichment factors (EF) for selected samples from Well 6. The Upper Continental Crust (UCC) normalization is after [Taylor and McLennan \(1985\)](#).

Depth (m)	Co (ppm)	Co EF	Cr (ppm)	Cr EF	Cu (ppm)	Cu EF	Mn (ppm)	Mn EF	Mo (ppm)	Mo EF	Ni (ppm)	Ni EF	U (ppm)	U EF	V (ppm)	V EF
7.7	1.23	0.205	152.67	5.221	57.72	6.553	24.66	0.117	15.35	29.045	111.96	7.222	7.32	7.420	147.75	3.919
10	4.26	0.955	93.83	4.307	29.56	4.505	45.35	0.288	8.84	22.455	81.23	7.034	8.98	12.220	71.44	2.544
12	1.45	0.745	85.12	8.953	16.52	5.769	19.32	0.281	4.23	24.619	38.67	7.673	4.24	13.220	41.23	3.364
15.45	0.72	0.199	84.49	4.777	38.75	7.274	26.7	0.209	13.22	41.363	69.69	7.433	10.29	17.247	75.52	3.312
18.3	6.84	2.716	30.6	2.488	22.64	6.112	158.91	1.788	28.73	129.275	87.46	13.416	12.86	30.999	59.16	3.732
21.8	0.77	0.521	215.98	29.926	25.61	11.781	28.74	0.551	17.5	134.173	57.88	15.128	60.54	248.658	174.21	18.724
27	0.37	0.188	618.07	64.271	107.97	37.275	9.66	0.139	6.59	37.918	108.06	21.197	46.53	143.426	301.35	24.308
30	1.06	0.145	1436.32	40.196	189.37	17.595	13.83	0.054	23.45	36.313	458.82	24.221	36.44	30.230	2282.12	49.541
34.5	1.23	0.082	550.56	7.492	96.79	4.373	11.45	0.022	40.65	30.607	218.24	5.602	41.98	16.933	282.78	2.985
38.5	0.78	0.157	348.39	14.373	58.58	8.024	10.42	0.059	13.27	30.293	129.35	10.066	17.2	21.035	305.02	9.761
42.5	1.66	0.245	466.64	14.078	135.78	13.599	14.38	0.060	187.53	313.043	342.35	19.482	24.94	22.303	1376.02	32.201
47.7	2.45	1.267	69.03	7.314	9.94	3.496	34.47	0.505	9.32	54.638	33.63	6.721	26.46	83.100	21.83	1.794
50.8	0.45	0.244	64.39	7.145	7.46	2.748	16.09	0.247	4.46	27.386	21.21	4.440	16.51	54.309	16.84	1.450
53.7	1.86	0.108	320.15	3.808	58.69	2.318	16.88	0.028	56.76	37.360	129.12	2.897	8.15	2.874	397.66	3.669
55.7	1.8	0.115	409.45	5.352	52.24	2.267	17.12	0.031	70.47	50.974	140.96	3.476	13.1	5.076	332.41	3.371
59.7	1.95	0.221	316.61	7.333	54.54	4.194	22.16	0.071	86	110.216	98.27	4.293	11.58	7.950	297.5	5.345
62.5	1.14	0.107	391	7.497	62.43	3.974	17.49	0.046	98.66	104.670	166.01	6.004	13.41	7.622	563.14	8.375
66	0.28	0.103	111.63	8.421	12.33	3.088	5.73	0.060	20.77	86.697	47.15	6.709	3.57	7.983	92.31	5.402
70.5	0.25	0.079	160.15	10.318	19.08	4.081	8.53	0.076	15.92	56.757	36.47	4.432	9.36	17.877	239.82	11.986
78.6	0.19	0.094	130.37	13.233	9.34	3.148	10.11	0.142	22.23	124.860	31.11	5.957	22.51	67.732	107.03	8.427
81.7	2.02	0.345	222.36	7.785	83.62	9.720	15.24	0.074	247.77	480.020	217.01	14.333	24.81	25.750	445.91	12.111
84.7	0.91	0.284	188.97	12.091	51.07	10.848	8.18	0.072	208.08	736.681	163.75	19.764	24.48	46.429	425.17	21.102
86.2	1.87	0.175	390.18	7.481	70.04	4.458	15.16	0.040	242.15	256.886	139.89	5.059	22.67	12.884	320.31	4.764
89.4	2.4	0.202	256.52	4.413	52.25	2.984	21.91	0.052	263.07	250.423	141.96	4.607	21.64	11.036	651.11	8.689
90.5	2.25	0.180	329.23	5.387	70.16	3.811	18.31	0.041	281.79	255.114	180.55	5.572	25.82	12.523	772.42	9.803
92.2	1.16	0.234	260.22	10.737	62	8.493	13.11	0.075	156.78	357.949	125.7	9.784	25.44	31.116	229.93	7.359
94	1.1	0.250	271.65	12.623	87.36	13.477	11.65	0.075	142.63	366.730	164.76	14.442	18.6	25.620	337.87	12.178
96.3	1.54	0.183	221.13	5.374	46.23	3.730	16.49	0.055	97.73	131.418	98.86	4.532	16.3	11.742	280.57	5.289
97.85	0.62	0.188	161.32	10.017	33.47	6.900	10.01	0.086	70.86	243.468	76.04	8.907	26.56	48.888	121.94	5.873
99.8	1.05	0.185	196.2	7.099	46.73	5.613	10.98	0.055	46.18	92.456	94.94	6.480	22.86	24.518	118.43	3.324
101.45	1.29	0.181	242.34	6.981	53.91	5.156	14.36	0.057	61.51	98.046	108.88	5.917	16.1	13.748	145.78	3.258
103.45	0.68	0.219	191.98	12.675	42.64	9.346	10.17	0.093	57.39	209.659	93.75	11.676	21.54	42.156	142.35	7.290
105.7	0.89	0.225	228.54	11.815	43.45	7.458	10.25	0.073	108.18	309.464	115.9	11.303	19.16	29.362	236.03	9.465
108.7	2.03	0.162	364.8	5.962	69.65	3.779	22.02	0.050	188.25	170.229	127	3.915	29.65	14.363	380.96	4.829
110.44	2.53	0.234	312.58	5.921	52.85	3.324	21.28	0.056	90.9	95.274	105.36	3.765	22.54	12.656	242.21	3.559
112.55	0.56	0.212	104.64	8.099	23.16	5.952	8.11	0.087	19.32	82.747	58.23	8.502	15.78	36.206	63.84	3.833
114.65	0.79	0.226	121.87	7.151	30.13	5.870	9.05	0.073	19.05	61.855	62.99	6.973	13.41	23.326	90.46	4.118
116.45	0.39	0.318	135.14	22.579	44.91	24.912	7.13	0.165	99.87	923.307	95.86	30.213	29.46	145.907	373.74	48.438
118.55	0.87	0.344	242.67	19.624	76.82	20.625	9.29	0.104	156.98	702.445	153.58	23.428	23.87	57.221	355.33	22.290
121.25	1.02	0.196	164.72	6.487	54.65	7.145	11.74	0.064	72.77	158.571	75.42	5.603	22.38	26.126	180.78	5.522
123.85	0.26	0.270	51.92	11.057	18.59	13.143	6.45	0.190	51.95	612.154	48.41	19.447	28.55	180.225	134.7	22.251
126.8	0.89	0.365	204.26	17.163	70.82	19.757	9.05	0.105	123.45	573.978	135.63	21.498	34.46	85.833	556.59	36.278
128.85	0.57	0.184	67.99	4.487	16.35	3.582	18.04	0.165	31.74	115.911	25.18	3.135	8.92	17.451	164.09	8.401
130.65	0.76	0.268	182.76	13.178	69.88	16.729	10.89	0.109	210.88	841.390	110.43	15.021	21.31	45.549	690.4	38.616
132	0.08	0.174	48.04	21.409	22.51	33.305	8.14	0.502	18.45	454.969	16.73	14.064	39.9	527.098	145.9	50.437
133.2	0.47	0.238	133.68	13.893	43.05	14.854	14.05	0.202	89.75	516.113	58.27	11.423	18.11	55.791	276.78	22.313
134.5	0.33	0.167	191.66	19.903	20.61	7.106	13.58	0.195	56.95	327.239	56.96	11.158	28.16	86.684	357.38	28.788
139.55	0.45	0.265	164.98	19.868	20.72	8.284	18.38	0.306	32.21	214.636	45.63	10.366	16.69	59.580	154.53	14.436

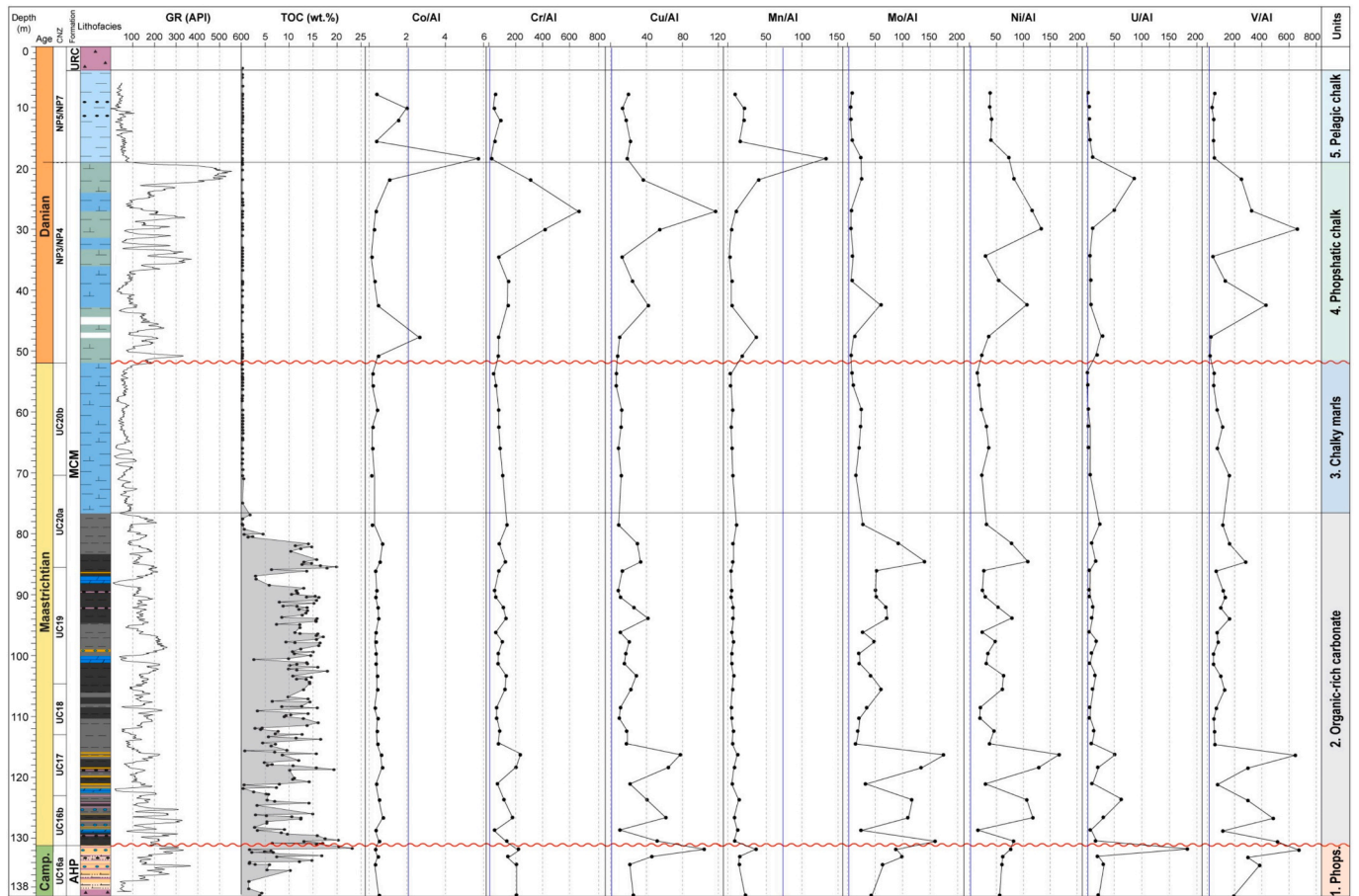


Fig. 8. Biostratigraphic column of Well 6 and selected trace-element proxies normalized to Al. The blue line represents the Upper Continental Crust (UCC) reference ratios for each element (after Taylor and McLennan, 1985). Note that Co/Al and Mn/Al are depleted relative to the UCC reference, whereas the other elements show enrichment. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

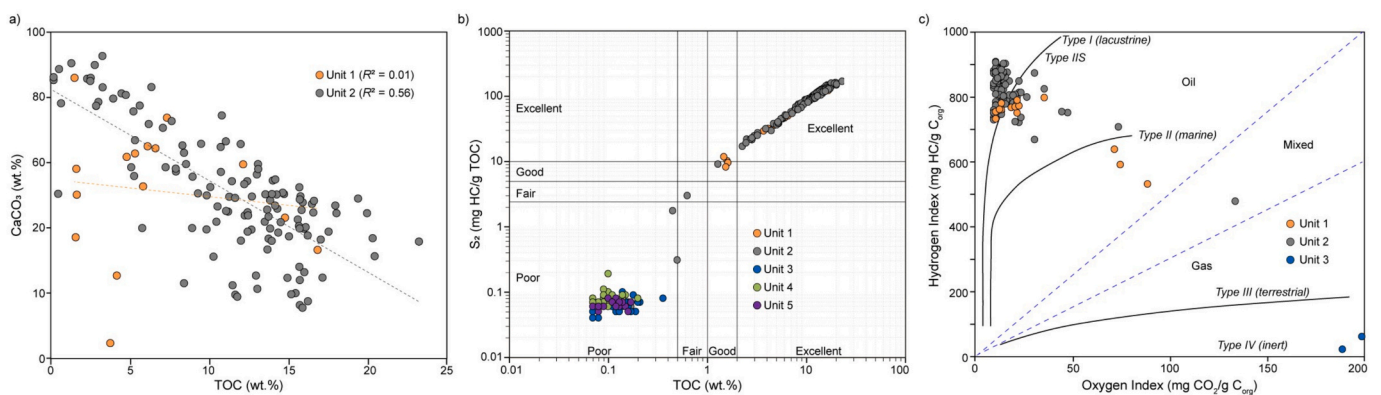


Fig. 9. Organic geochemical characterization of Well 6. (a) Negative linear correlation between calculated CaCO_3 and TOC. The phosphorites (Unit 1) show a very weak relationship, whereas the organic-rich carbonate unit (Unit 2) displays a moderate correlation. (b) Samples from Units 1 and 2 indicate excellent hydrocarbon generation potential, whereas Units 3, 4, and 5 exhibit poor hydrocarbon generation potential. (c) Pseudo-Van Krevelen diagram plotting HI versus OI to determine the original organic matter type. The data suggest a mixture of Type II and Type IIS kerogen. Sulfur-rich samples with high HI and very low OI plot along the Type I trend in the pseudo-Van Krevelen diagram (after Dembicki, 2009).

low gamma-ray signal. Toward the top, both cycles progressively grade into coarser-grained, reworked, and highly bioturbated phosphorites (Fig. 4d). The phosphate content of Unit 1 is high, with apatite ($\text{Ca}_5(\text{PO}_4)_3\text{F}$) concentrations reaching up to 21 wt% in Well 1 (Table S3) and 23 wt% in Well 2 (Usman et al., 2025). Similarly, Plado et al. (2016) reported maximum P_2O_5 concentrations at the same gamma-ray datum in wells BH-1, BH-3, and BH-5 (Fig. 10). High gamma-ray responses in

organic-rich mudstones are commonly attributed to the enrichment of radioactive elements such as uranium (Rider, 2002). However, the U-TOC relationship might be suppressed by the presence of phosphate, which acts as a sorbent, triggering high U precipitation (Baturin and Kochenov, 2001; Lüning and Kolonic, 2003). In Unit 1, U concentrations are closely associated with phosphorous (Fig. 7d), suggesting that the elevated gamma-ray response primarily reflects U enrichment within

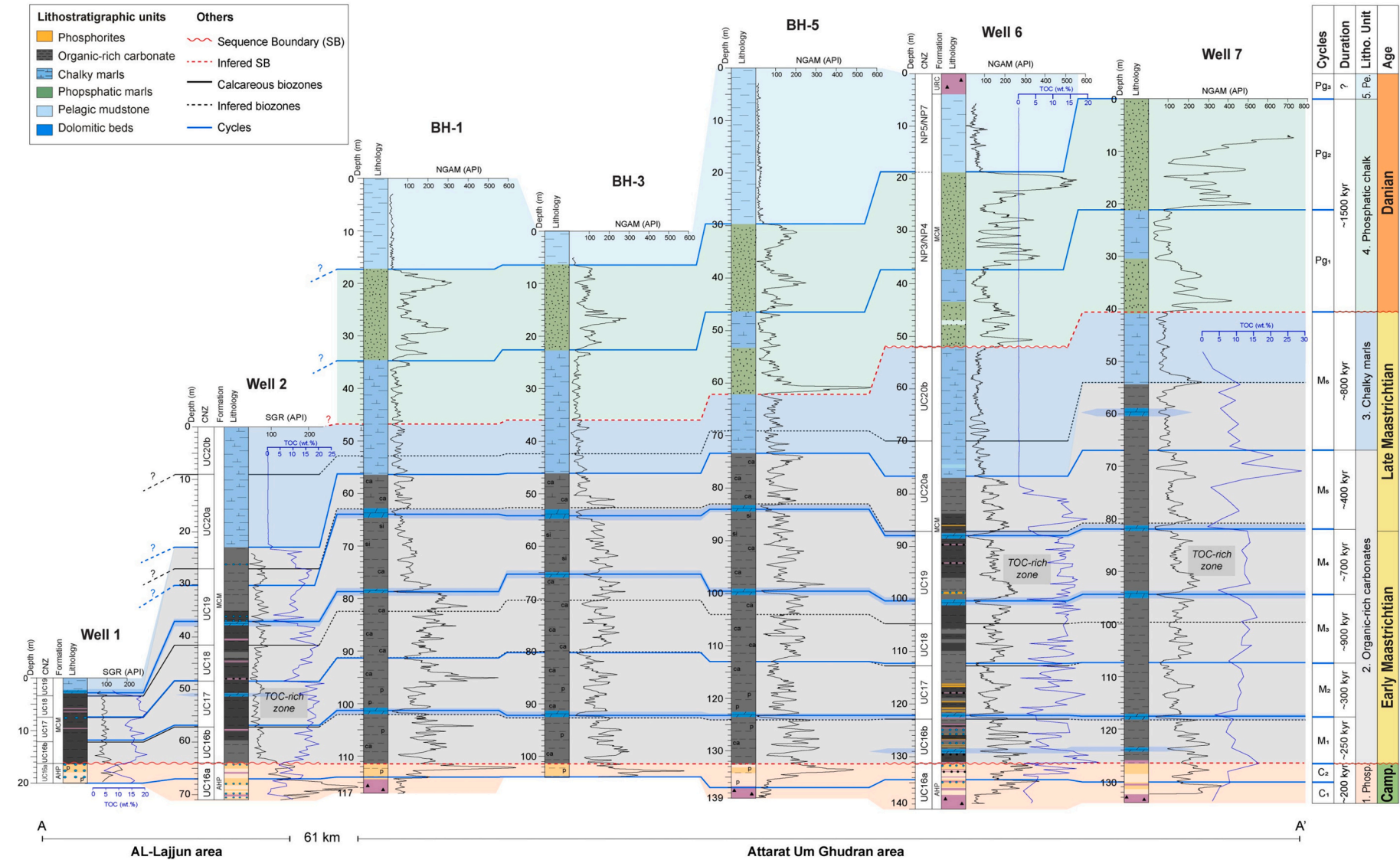


Fig. 10. Regional correlation panel of lithostratigraphic units across key sections in central Jordan. The transect is flattened at the top of UC16a, which coincides with a prominent gamma-ray peak and high phosphorite content. Mineralogical compositions for BH-1, BH-3, and BH-5 are from [Plado et al. \(2016\)](#) and [Puura et al. \(2017\)](#).

phosphatic facies rather than organic-carbon enrichment (Fig. 7c). Phosphorite Unit 1 is clearly identified in cored wells 1, 2, and 6, showing comparable upward-coarsening cycles and trends in bioturbation intensity (Fig. 4).

In contrast, the organic-rich carbonate Unit 2 exhibits much more lateral facies variation along the transect. Its thickness decreases significantly toward the southwest (Al-Lajjun area), reaching only about 20 m in Well 1 compared to the average 50 m observed in the other wells (Fig. 10). Bioturbation and dolomitic limestone beds also vary laterally, being common in Well 6, markedly reduced in Well 2, and very rare in Well 1. Nannofossil age dating was essential for correlating these wells across significant lateral facies changes. Five depositional cycles are identified in this unit:

Cycle M₁ is marked at the base by the transition from the underlying phosphorite Unit 1 (AHP-MCM transition) to a phosphate-lean interval capped by a decimeter-thick dolomitic limestone bed (Fig. 10). This cycle represents a relatively short interval of ~250 kyr, spanning the early Maastrichtian UC16b and part of the UC17 nannofossil zones, with a minimum sedimentation rate of ~2 cm/kyr. The M₁ cycle shows marked variability in facies distribution and gamma-ray response from east to west. Well 6 is dominated by bioturbated mudstone, intercalated by several centimeter-thick lag deposits and carbonate concretions, and shows a relatively high concentration of redox proxies, particularly of Mo, V, and U (Fig. 8), which are also present in the BH wells (Plado et al., 2016). Whereas the condensed wells 1 and 2 display a sharp transition to laminated mudstones and bioclastic wackestone. Although current-related sedimentary structures (e.g., ripples, wavy lamination) are common in wells 1 and 2 (Fig. 5b and d), no lag deposits were identified in those sections. The gamma-ray log shows a subtle upward decrease, punctuated by high peaks related to the phosphatic-rich layers (Fig. 10). The top of cycle M₁ is laterally marked by the first dolomitic limestone bed, recognizable for its low gamma-ray values, which can be followed from Well 7 to BH-1 but not recognized in wells 1 and 2. The TOC content also varies laterally along this cycle. Wells 1 and 2 are characterised by high average TOC contents, 16 wt% and 15 wt%, respectively, whereas in Well 6, the average TOC is 11 wt%.

Cycle M₂ spans the nannofossil Zone UC17 with a duration of ~300 kyr and a sedimentation rate of ~3.3 cm/kyr. The lithofacies distribution and geochemical composition along this cycle are likely similar to those of the underlying Cycle M₁ (Fig. 3). It is characterised by some lag deposits cutting bioturbated mudstones and by a continuous upward decrease in gamma-ray response. The highest TOC values or “TOC-rich zone” in Well 2 occur during this cycle and average 17.6 wt% (Usman et al., 2025). In contrast, Well 6 records lower TOC values for the same interval, averaging 9.8 wt% (Fig. 10). The top of the cycle M₂ in Well 2 coincides with the top of “Cycle 1” in Usman et al. (2025), characterised by a laterally traceable drop in the TOC and gamma-ray response (Fig. 10).

Cycle M₃ spans the calcareous nannofossil zones UC18 and part of UC19, with an estimated duration of ~900 kyr and a low sedimentation rate (~1.33 cm/kyr) (Fig. 3). During deposition of cycle M₃, Well 6 records predominantly laminated to bioturbated marls facies characterised by increase in detrital proxy concentrations (SiO₂, Al₂O₃, and Ti), together with a slight decrease in CaO (Fig. 3). At the same time-equivalent interval of Well 2, Usman et al. (2025) described silica-rich facies mainly dominated by biogenic silica (opal CT). The gamma-ray response shows a subtle upward increase toward the top of the cycle, followed by a sharp decrease at the level of a dolomitic limestone bed along the A–A' transect. This bed is also marked by a distinct low gamma-ray peak and can be traced across all wells (Fig. 10).

Cycle M₄ is marked at the top by another dolomitic limestone bed that can be followed in most wells, except wells 1 and 2. The sequence covers most of the nannofossil Zone UC19, with a duration of ~700 kyr, and an estimated sedimentation rate of ~2 cm/kyr. This sequence is characterised by laminated facies in all cored wells, abundant framboidal pyrite, and a slight decrease in detrital proxies (Fig. 5f). The

gamma-ray log shows a strong response associated with the phosphatic-rich intervals at the base of this cycle observed in wells 6 and 7, which can be followed regionally. It is followed by a slight decreasing upward trend. The TOC content of this sequence shows the highest TOC values, “TOC-rich zone”, in wells 6 and 7 (Fig. 10). In sharp contrast, in Well 1, where this sequence is probably only partly preserved, the facies has changed to chalky marls, with no organic matter (Fig. 10).

Cycle M₅ spans part of the late Maastrichtian nannofossil Zone UC20a, with an estimated duration of 400 kyr and a sedimentation rate of ~2.5 cm/kyr, and dominated by highly bioturbated, carbonate-rich lithofacies in wells 2 and 6 (Fig. 5a). The top of Cycle M₅ is defined by the transition from TOC-rich marls to TOC-lean chalky marls across most of the wells, except in Well 7, where this transition occurs at a higher stratigraphic position. The top of Cycle M₅ coincides with the top of “Cycle 3” defined by Usman et al. (2025) in Well 2. Cycle M₅ is not clearly expressed in the gamma-ray log along the transect; however, TOC contents are markedly high and show stratigraphic variability, with peaks reaching the highest recorded values (e.g., 29.25 wt% in Well 7; Fig. 10).

The overlying cycle M₆ covers the upper part of the UC20a and all the UC20b nannofossil zones, with an estimated duration of 800 kyr, and an average sedimentation rate of 6 cm/kyr. Cycle M₆ is marked at the top by the K/Pg boundary, identified in Well 6, and correlated laterally using a distinctive positive excursion in the gamma-ray signal (Fig. 10). The base of this cycle is marked by an abrupt decline in TOC values along the transect, with concentrations decreasing to an average of 0.17 wt% in Well 6. In contrast, although TOC values also decrease in Well 7, they remain relatively high, averaging ~7.5 wt%, indicating a diachronous transition from organic-rich facies to marlstones during this cycle. The shift to chalky and marly limestones may reflect a major paleoenvironmental transition from a high-productivity, anoxic setting to more oxygenated conditions. This interval is also characterised by a stronger positive correlation between SiO₂ and Al₂O₃ (Fig. 7a), suggesting a closer association between silica and detrital material than in the underlying organic-rich facies. A general decrease in the gamma-ray response is observed throughout this cycle, coinciding with the end of the organic-rich interval and the increase in CaCO₃ content within the chalky marlstone facies (Fig. 3). The high carbonate content likely diluted the concentration of radioactive elements, thereby limiting the gamma-ray response, as carbonate rocks typically contain lower amounts of K, Th, and U compared with clay-rich sediments (Lüning and Kolonic, 2003).

Cycles Pg₁ and Pg₂ are two similar sequences that contain the Paleocene nannofossil zones NP3/NP4, representing a duration of approximately 1500 kyr. The base of cycle Pg₁ corresponds to the K/Pg boundary clearly marked by an abrupt positive shift in gamma-ray response and sharp change in bulk geochemical composition in Well 6 (Fig. 3). Given the organic-poor nature of these cycles, the elevated gamma-ray response is interpreted to be primarily influenced by U associated with phosphatic-rich facies, as described in Unit 4 (Fig. 4). This interpretation is supported by the strong positive correlation between U and P concentrations observed in Unit 4 ($R^2 = 0.85$; Fig. 7d). The top of Pg₂ is placed at the top of a strong gamma-ray response (>400 API) easily traceable laterally in BH wells from Plado et al. (2016).

Cycle Pg₃ comprises the Paleocene nannofossil zones NP5/NP7 and corresponds to the uppermost MCM Formation. The cycle is laterally correlated using the distinctive low gamma-ray response from “clean” chalk facies described in Unit 5 from Well 6 (Fig. 10). The top of the Pg₃ cycle is defined by the first appearance of a hard and fractured chert bed of the URC Formation (Andrews, 1992).

5. Discussion

5.1. Controls on organic carbon enrichment

High concentrations of organic carbon in open marine continental

shelves, particularly within shallow-shelf environments, are rarely controlled by a single process. Instead, TOC enrichment reflects the interplay of multiple processes and conditions, including enhanced carbon burial efficiency often associated with enhanced ocean mixing in upwelling settings, expansion of the oxygen minimum zone (OMZ) during transgressive phases, restricted circulation, and low detrital input (Arthur et al., 1990; Wilson and Norris, 2001; Bohacs et al., 2005; Bjerrum et al., 2006; Inthorn et al., 2006).

Previous geochemical studies of the Maastrichtian MCM Formation have emphasized the important role of upwelling-driven productivity and redox conditions in organic-carbon accumulation and preservation (Fleurance et al., 2013; Usman et al., 2025). Comparable successions characterised by bituminous mudstones, cherts, and phosphorites, have been documented in other Mesozoic intervals, including the Coniacian–Campanian successions of the southern Neo-Tethys margin (e.g., the Tarfaya Basin, Morocco, the Levant region) (El Albani et al., 1999; Soudry et al., 2006) and the Turonian–Santonian successions of the Caribbean Seaway (e.g., the La Luna and Villeta formations, Colombia) (Macellari and De Vries, 1987; Villamil et al., 1999), as well as in modern upwelling systems, such as the Namibia and Peru margins (Robinson et al., 2002; Böning et al., 2004).

In Jordan, the new geochemical data presented in this study provide strong evidence for both elevated paleoproductivity and enhanced preservation. Therefore, productivity-sensitive and redox-sensitive proxies are evaluated to assess their contributions to organic-carbon enrichment within the Maastrichtian MCM Formation. Cu, Ni, and Zn are important micronutrients widely used as proxies for marine productivity (Tribovillard et al., 2006; Sweere et al., 2025). The enrichment of these elements in sediments is largely controlled by the biogenic cycle, through their incorporation into phytoplankton cells, followed by complexation with organic matter and subsequent transport to the seafloor (Tribovillard et al., 2006; Böning et al., 2015). Elevated Cu/Al and Ni/Al ratios, and specifically for Ni, with its abnormal enrichment (Table 4) and good correlation with TOC (Fig. 7b), indicate a substantial authigenic contribution relative to high-nutrient influx, comparable to that observed in Pleistocene sediments deposited in the Benguela current system (Robinson et al., 2002). In addition, low Co/Al and Mn/Al ratios and very low enrichment factors (<0.5) characterise most of the MCM organic-rich interval (Fig. 8). This depletion of Mn and Co is indicative of strong oxygen-deficient bottom waters, comparable with those observed in modern upwelling sediments (Böning et al., 2004; Borchers et al., 2005; Brumsack, 2006; Sweere et al., 2016). Under such conditions, dissolution of Mn oxyhydroxides promotes remobilization of Mn and Co into the water column, resulting in the depletion of both elements in the sediment (Tribovillard et al., 2012).

In addition, high concentrations of Cr, Cu, Mo, Ni, U, V, and Zn in the bituminous marlstone suggest deposition controlled by redox conditions (Table 4) (Brumsack, 2006; Tribovillard et al., 2006; Bennett and Canfield, 2020). The same trace elements have been shown to be typically enriched in modern upwelling settings with strong oxygen depletion (Robinson et al., 2002; Böning et al., 2004; Sweere et al., 2016; Bennett and Canfield, 2020). For instance, the chemical speciation of Mo from its unreactive form, MoO_4^{2-} to thiomolybdate species (MoS_4^{2-}), is typically controlled by the presence of free hydrogen sulfide (H_2S) under euxinic waters and readily absorbed into organic matter (Algeo and Lyons, 2006; Tribovillard et al., 2012). Mo can be incorporated into sediments through terrigenous input, biogenic material, or direct authigenic uptake from seawater; however, because Mo plays only a minor role in biological cycles, its contribution from planktonic detritus is typically very small, particularly in the absence of strong redox gradients (Brumsack, 1986; Robinson et al., 2002; Calvert and Pedersen, 2007).

Although significant remobilization of redox-sensitive elements is generally associated with sediment oxygenation (e.g., bioturbation) or late-burial diagenesis (Algeo and Tribovillard, 2009), the diagenetic redistribution cannot be entirely excluded during deposition. Fleurance

et al. (2013) documented localized U remobilization along calcite veins and fractures within the AHP and MCM formations; however, they concluded that the abnormal enrichment of some trace metals is predominantly authigenic and largely reflects deposition under oxygen-depleted conditions. Within the studied MCM interval, the vertical consistency of Cu, Mo, and V enrichments and their close association with the organic-rich interval support a similar interpretation. Uranium, however, appears to be more closely associated with phosphatic intervals than with TOC (Unit 1 and Unit 3; Fig. 7d), suggesting that its enrichment was further influenced by phosphogenesis and its incorporation into authigenic phosphate phases.

The succession exhibits low detrital content, as indicated by Al_2O_3 concentrations below 6 wt%, whereas high $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios are not correlated with elevated ZrO_2 concentrations (Table S1), further suggesting limited terrigenous input. Consequently, the strong Mo enrichment ($61 < \text{EF} < 923$) observed in these organic-rich facies most likely reflects highly reducing depositional conditions (euxinic conditions), rather than any substantial input from planktonic or terrigenous material.

Sedimentation rates also largely control the accumulation and preservation of organic matter in marine sediments. High rates generally enhance carbon burial by limiting the exposure time for microbial degradation or dilution effects (Demaision and Moore, 1980; Stein, 1986; Arthur et al., 1990). Conversely, low sedimentation rates prolong exposure within the oxic zone, promoting more complete oxidation of organic matter under well-oxygenated bottom-water conditions (Canfield, 1994). However, under anoxic or euxinic bottom-water conditions, the effect of sedimentation rate on organic matter preservation becomes secondary to redox control. In such settings, even low rates can coincide with high TOC contents because the absence of oxygen suppresses microbial degradation and bioturbation, thereby enhancing preservation of organic matter and significant enrichment of authigenic trace elements (Demaision and Moore, 1980; Emerson, 1985; Arthur et al., 1990; Tribovillard et al., 2006). For example, the Atlantic “black shales” deposited during the OAE2 exhibit both high TOC and trace-element enrichments, despite slow sediment accumulation, reflecting strongly reducing conditions that stabilized the organic matter and favored the formation of authigenic phases of elements, especially Mo, U, and V (Arthur et al., 1990). Relatively low sedimentation rates ranging between 1.3 and 3.3 cm/kyr. Were estimated within the organic-rich carbonate Unit 2 (Fig. 2), including in the “TOC-rich zone” which averages ~ 2 cm/kyr and is comparable to those reported by Ashkenazi-Polivoda et al. (2011) in the Negev area of ~ 2.4 cm/kyr. Similarly, in modern upwelling margins such as the Benguela and California systems, even low sedimentation rates can lead to substantial organic-carbon enrichment, since up to $\sim 30\%$ of primary productivity is exported below the euphotic zone (Brumsack, 1986; Algeo and Maynard, 2004; Tribovillard et al., 2012; Algeo and Li, 2020).

Thus, the Maastrichtian organic-rich interval of the MCM Formation reflects the combined influence of enhanced nutrient supply sustained by a prolonged regional upwelling system, while organic-matter preservation was promoted by limited dilution and dysoxic and anoxic conditions at the sediment–water interface despite relatively slow sedimentation rates. Together, these factors allowed the deposition of laterally continuous TOC-rich intervals along central Jordan (Fig. 10). This interpretation contrasts with previously proposed models, in which the organic-matter enrichment was attributed to restricted circulation within fault-controlled basins (Abed and Amireh, 1999; Abed et al., 2005; Powell and Moh'd, 2011; Fleurance et al., 2013; Ali Hussein et al., 2015; Hakimi et al., 2016; Beik et al., 2017, 2018, 2019; Sokol et al., 2017; Grohmann et al., 2023). Although a major graben structure developed in the Sirhan area in northern Jordan during the Coniacian to early Campanian (Fig. 2), only minor subsidence differences occurred during the Maastrichtian MCM Formation, which coincided with the geodynamic sag phase (Kalifi et al., 2025).

5.2. Regional depositional cycles

The decimeter to meter-scale cycles identified in the studied interval exhibit a hierarchical organization comparable to third and fourth-order sequences, and likely influenced by orbital forced (Milankovitch-scale) processes that controlled climatic changes, ocean currents, marine productivity, and weather conditions during the Late Cretaceous (Sageman et al., 1997; Gale et al., 2002; Wilmsen, 2003; Schlager, 2005; Wendler et al., 2014; Gambacorta et al., 2016; Li and Schieber, 2018; Usman et al., 2025; Levell and van Buchem, 2026).

5.2.1. Campanian depositional cycles

The Campanian cycles C_1 and C_2 , which comprise the uppermost part of the AHP Formation, fall within the range of fourth-order cycles (~100 kyr). The cycles reflect depositional conditions that promoted phosphogenesis, bounded by a regionally expressed lateral shift toward carbonate facies belts (Fig. 10). The organic-rich and laminated phosphorites observed in the lower part of each cycle may indicate deposition under stressed bottom-water conditions (Soudry et al., 2006; Fernández-Martínez et al., 2021). In the Negev region, similar stacking patterns have been described in the Phosphate Member of the Mishash Formation, where the laminated facies are referred to as “pristine phosphorites” formed by the authigenic precipitation of carbonate-fluorapatite minerals and deposited under suboxic conditions (Pufahl et al., 2003; Soudry et al., 2006; Levy et al., 2023).

The phosphatic grainstone facies at the top of each Campanian cycle is interpreted as the result of reworking by bottom currents or storm-induced winnowing of earlier laminated pristine phosphorites (Abed and Amireh, 1999; Pufahl et al., 2003). Additionally, the homogenization of sediments was largely driven by intense biogenic reworking by benthic fauna (Fig. 4d), suggesting longer episodes of oxic-suboxic bottom water conditions. Comparable facies have been reported in Devonian phosphatic black shales, where low sediment accumulation rates, coupled with active sediment reworking, promoted high phosphate concentrations (Li and Schieber, 2015).

The rhythmic oxic–anoxic fluctuations recorded in hemipelagic carbonate successions are commonly interpreted as expressions of monsoon-driven climate variability, which is sensitive to precession (~20 kyr) and obliquity (~41 kyr) orbital forcing (Sageman et al., 1997; Prokoph and Thurow, 2000; Wilson and Norris, 2001; Gale et al., 2002). The integration of high-resolution biostratigraphy, geochemical proxies, and bioturbation patterns in this study suggests that the observed oxic–anoxic fluctuations and repetitive reworking and redeposition of phosphate are most likely linked to the short eccentricity (~100 kyr) Milankovitch cycles.

5.2.2. Maastrichtian depositional cycles

The MCM Formation represents a deepening of the depositional environment that occurred during the onset of a transgressive phase in the early Maastrichtian (Powell and Moh'd, 2011; van der Meer et al., 2022; Messaoud et al., 2025). The marked decrease in grain size, (mudstone to packstone facies with abundant planktonic foraminifera; Fig. 5) suggests deposition in a relatively deeper setting compared to the underlying phosphorite Unit 1 (Fig. 3). The internal heterogeneity of the MCM Formation is expressed by two orders of depositional cycles, which are particularly well developed within the organic-rich interval (Unit 2).

Low-oxygen conditions and relatively high-energy bottom currents appear to have predominated during cycles M_1 and M_2 . This interpretation is supported by the presence of low-diversity trace fossils dominated by horizontally oriented burrows (Fig. 5a), the abundance of planktonic foraminifera (Fig. 5e), and the scarcity of benthic foraminifera (e.g., *Gavelinella* spp.) (Fig. 5h), which are tolerant of low-oxygen environments under relatively high food supply (Ashckenazi-Polivoda et al., 2011). The common occurrence of mud lenses (Fig. 5b) and lag deposits (Fig. 5j) within this cycle is interpreted as active bedload transport by bottom currents (Schieber et al., 2010; Emmings et al.,

2020). In this lithofacies, the absence of a micritic matrix indicates high-energy bottom-current conditions, during which finer clay-sized particles were winnowed out, leaving mainly sand-sized grains (Flügel and Munnecke, 2010). The remaining bioclasts, such as mollusks, algae, and fish bones, are highly fragmented (Fig. 5k), suggesting long-distance transport and reworking before being mixed with unbroken benthic fauna that had not been transported.

However, the absence of these features in wells 1 and 2, combined with the presence of “TOC-rich intervals” (Fig. 10), suggests that the southwestern wells (Al-Lajjun area) were located in a more distal position, characterised by lower-energy conditions and strongly anoxic bottom waters, contrasting with the more proximal settings in the Attarat Um Ghudran area, where higher-energy bottom currents prevailed. The lateral variations along the A–A' transect likely reflect medium-order depositional cycles, possibly related to long eccentricity (~405 kyr). The centimeter-scale heterogeneity and erosive surfaces observed in the succession may also record the superposition of higher-frequency cycles. In modern analogs, erosive surfaces within mudstone successions are commonly associated with limited accommodation resulting in active reworking influenced by storms or bottom-water currents (Schieber, 1998; Lazar and Schieber, 2022).

In the M_3 cycle, the increase in detrital proxies, particularly Al_2O_3 and Ti, likely indicates dilution of TOC concentrations in Well 6. In contrast, in Well 2, Usman et al. (2025) reported markedly elevated opal cristobalite-tridymite (CT) attributed primarily to biogenic silica production rather than detrital input. These lateral geochemical variations suggest the presence of at least two distinct primary-producer zones, or paleoproductivity cells, controlling the distribution of carbonate and silica during Cycle M_3 along the A–A' transect (Fig. 10). In modern upwelling systems, similar facies mosaics reflect lateral variations in nutrient supply: diatoms (silica-secreting microorganisms) dominate the central, most active upwelling belts, whereas calcareous nannoplankton and dinoflagellates become more abundant toward the distal margins where nutrient levels are comparatively lower (Robinson et al., 2002; Böning et al., 2004).

During Cycle M_4 , high concentrations of opal CT were reported by Puura et al. (2017) in the BH wells, coinciding with increased SiO_2 concentrations and the occurrence of the “TOC-rich zone” in Well 6. This contrasts with the more distal wells 1 and 2, where the TOC-rich zone developed earlier during cycle M_2 , therefore indicating a diachronous peak in productivity and organic-matter accumulation. This pattern likely reflects a temporal migration of paleoproductivity cells across the basin (Fig. 10).

The last cycle within the organic-rich carbonate Unit 2, Cycle M_5 , represents a short-term depositional cycle with highly bioturbated mudstone identified along the transect, inferring a more oxygenated sediment/water interface, which may have contributed to the decrease in TOC content. However, a distinctive characteristic of this cycle is the occurrence of a pronounced TOC peak in most of the wells (Fig. 10).

The transition from the organic-rich interval to the organic-poor chalky marl facies of Unit 3 occurred during UC20a in Cycle M_6 across all wells in central Jordan, at approximately 69.2 Ma (FO of *Lithraphidites quadratus*) (Fig. 10). The same lithostratigraphic transition has been documented in the PAMA section, located in a more distal position in the Levant region, where it occurs slightly earlier within the planktonic foraminiferal *Pseudoguembelina palpebra* Zone, at approximately 69.85 Ma (Ashckenazi-Polivoda et al., 2011; Meilijson et al., 2014). The strong correlation of detrital elemental proxies from Unit 3 (Fig. 7a), which corresponds to cycle M_6 (Fig. 10), suggests an increase in continental runoff. This trend is consistent with regional lithological and faunal shifts reported in more distal sections of the Levant region, where the abrupt change from organic-rich to organic-poor facies has been attributed to a weakening of the upwelling system and a decline in surface-water fertility under greenhouse warming (Ashckenazi-Polivoda et al., 2011; Alsenz et al., 2013). Consequently, this phase likely reflects a major reorganization of the marine ecosystem, transitioning from

eutrophic, upwelling-dominated conditions to more mesotrophic and oligotrophic, and oxygenated environments during the late Maastrichtian in Jordan. In addition, isotopic data presented by Beik et al. (2018) correlated TOC minimum values with a negative $\delta^{13}\text{C}$ excursion, which has been interpreted as reflecting enhanced runoff and long-eccentricity (~ 405 kyr) forcing during UC20b (time-equivalent to Cycle M₆).

In central Jordan, the K/Pg boundary in outcrop sections is defined by a bioturbated and erosive surface characterised by chert concretions and phosphatic marls (Bazeen et al., 2020; Farouk et al., 2014; Alhejoj et al., 2020). These authors highlight the presence of crushed Cretaceous foraminiferal assemblages and transported plant debris. Similar reworked Late Cretaceous taxa have also been described from calcareous nannofossils in the lower Danian section of the Wadi Bustani outcrop (Messaoud et al., 2025). In the present study, although lateral resolution is limited by the core material, the K/Pg boundary is confidently constrained by high-resolution calcareous nannofossils zone (CNZ), and the sharp geochemical change (high phosphate concentrations) is consistent with the characteristics observed in nearby outcrop sections (Fig. 3). The Danian Pg1 and Pg2 cycles suggest high-energy bottom-water conditions and repeated erosional episodes, as indicated by the abundant reworking of Cretaceous material.

5.3. Late Campanian to early Danian facies evolution in Jordan

In order to better understand the temporal and spatial evolution of the Maastrichtian paleoproductivity, two new transects have been established with two age-dated wells (OS-2 and OS-28; Alqudah et al., 2015; Beik et al., 2019) and the age-dated Wadi Bustani outcrop section (Messaoud et al., 2025) (Fig. 11). The significant lateral variations in TOC distribution observed in these sections highlight both a diachronous component and a geographical component of the organic matter deposition. The age dating of the sections allows us to illustrate the facies evolution in four steps: latest Campanian phosphorites, early Maastrichtian organic-rich carbonate, late Maastrichtian marly chalk, and Danian phosphatic chalk (Fig. 12). These maps, which also include additional control points (Alqudah et al., 2015; Beik et al., 2019), serve as the basis for the proposition of a depositional model (Fig. 13).

5.3.1. Latest Campanian phosphorites (uppermost AHP)

The initiation of the robust upwelling system along the southern margin of the Neo-Tethys Ocean was driven by the interplay between major global oceanographic reorganizations and local tectonic events that began during the late Coniacian (Almogi-Labin et al., 1993; Soudry et al., 2006; Edelman-Furstenberg, 2009; Powell and Moh'd, 2011; Abed, 2013). The westward-flowing Tethyan Circumglobal Current (TCC) is thought to have intensified following an early Santonian cooling event, increasing both its velocity and its capacity to transport nutrient-rich deep waters onto shallow marine platforms (Hein et al., 1993; Puc  at et al., 2005; Soudry et al., 2006; Hay, 2009).

Under this high-productivity system, the phosphorite Unit 1 described here was deposited near the Campanian–Maastrichtian boundary (UC16a), reflecting repeated small-scale cycles superimposed on long-term eustatic sea-level fluctuations (Miller et al., 2005; van der Meer et al., 2022). The same phosphorite interval is reported overlying large oyster mounds (buildups ~ 30 m thick) from central and south Jordan. This interval exhibits very low sedimentation rates, less than 1 cm/kyr, and is characterised by reworked calcareous nannofossils (Messaoud et al., 2025). Some authors assigned this interval as an overall shallowing event up to the Campanian–Maastrichtian boundary, characterised by an intensely *Thalassinoides*-burrowed bed as the top of the AHP Formation (Powell and Moh'd, 2011; Ahmad et al., 2020; Messaoud et al., 2025). However, other authors have interpreted the phosphate-rich deposits as a condensed unit associated with a rise in sea level and deposited during the final phase of a transgressive systems tract (TST) (Pufahl et al., 2003; Abed et al., 2007).

Similar condensed intervals are reported in the surrounding areas. In the Western Desert of Egypt, the Campanian–Maastrichtian boundary near the FO of the planktonic foraminiferal *Rugoglobigerina hexacamerata* Zone (71 Ma) (Tantawy et al., 2001), approximately equivalent to the calcareous nannofossil Zone UC16. Regardless of the exact water depth interpretation of this unit, the top surface in Western Desert of Egypt is interpreted as a regional transgressive surface, characterised by a sedimentation break, reworked phosphate limestone with *Chondrites* burrows and glauconitic sands (El-Azabi and Farouk, 2011; El Ayyat, 2015). Low sedimentation rates (1.4 ± 0.5 cm/kyr) also occurred near the Campanian–Maastrichtian boundary, slightly below the base of the foraminiferal *P. palpebra* Zone in southern Negev (Soudry et al., 2006; Meilijson et al., 2014). The condensed interval is here interpreted as an MFS characterised by interbedded pristine and reworked peloidal phosphates and an unconformity that underlies the early Maastrichtian organic-rich carbonates. The top of our phosphatic Unit 1 is correlated with the top of the Phosphate Member of the Mishash Formation, defined at the base of *P. palpebra* Zone (71.6 Ma) (Ashckenazi-Polivoda et al., 2011; Meilijson et al., 2014).

The different regional interpretations of the Campanian–Maastrichtian sequence boundary might be explained by variations in biostratigraphic approaches, the influence of local tectonics, basin geometry (e.g., the Syrian Arc, Red Sea rifting), or the relative position with respect to the paleoshoreline at the late Campanian. Although the nature of the sequence boundary remains open in the context of our study, it is important to emphasize the regional shallowing of the depositional settings and the occurrence of generalized sediment starvation in Jordanian sections below this surface (Messaoud et al., 2025), conditions under which high-energy bottom currents likely became the dominant sedimentary process for the reworking and redeposition of Campanian pristine phosphorites from the AHP Formation (Fig. 13a).

5.3.2. Early Maastrichtian organic-rich carbonates (lower MCM)

The early Maastrichtian marks a major transgressive phase across the northern Arabian Plate (Miller et al., 2005; Powell and Moh'd, 2011; van der Meer et al., 2022). Sedimentological observations, including reduction in grain size, a drastic decline in bioturbation, and the dominance of planktonic foraminifera, indicate in Jordan progressive deepening of the depositional environment compared to the upper Campanian phosphorite interval below. In parallel, the strong enrichment of redox-sensitive trace elements further supports the establishment of persistently suboxic to anoxic bottom-water conditions during this phase of inner-platform drowning (Fig. 12a).

Biostratigraphic studies from adjacent regions in southern Negev have documented a significant turnover in primary communities (e.g., diatoms, coccolithophores, dinoflagellates, planktic and benthic foraminiferal assemblages), which likely contributed to enhanced organic matter production and redox conditions during the early Maastrichtian (Almogi-Labin et al., 1993; Ashckenazi-Polivoda et al., 2011; Meilijson et al., 2014). These ecological changes may also have directly influenced the mineralogical composition of bulk sediments. Although a detailed microfossil assemblage analysis is beyond the scope of this study, the enrichment of paleoproductivity proxies (Ni, Cu, Si, P, TOC) together with the prevalence of thick siliceous-rich mudstone (e.g., Cycle 2 in Usman et al., 2025) facies supports deposition beneath highly productive, diatom-dominated surface waters, typically associated with nutrient-rich, mesotrophic to eutrophic conditions.

In more distal sections (e.g., PAMA section; Ashckenazi-Polivoda et al., 2011; Meilijson et al., 2014), the “TOC-rich zone” occurs in the lower part of the outcrop succession (44.2–34.4 m), occurring within the FO of the planktonic foraminifera *Pseudoguembelina kempensis* (~ 71.5 Ma), and approximately time-equivalent to the deposition of the “TOC-rich zone” in our Well 2 during UC17 (Fig. 10). In more proximal sections (e.g., OS-02 core; in Beik et al., 2019), this high TOC interval appears later, during the later nannofossil zones UC18/19 (Fig. 11b).

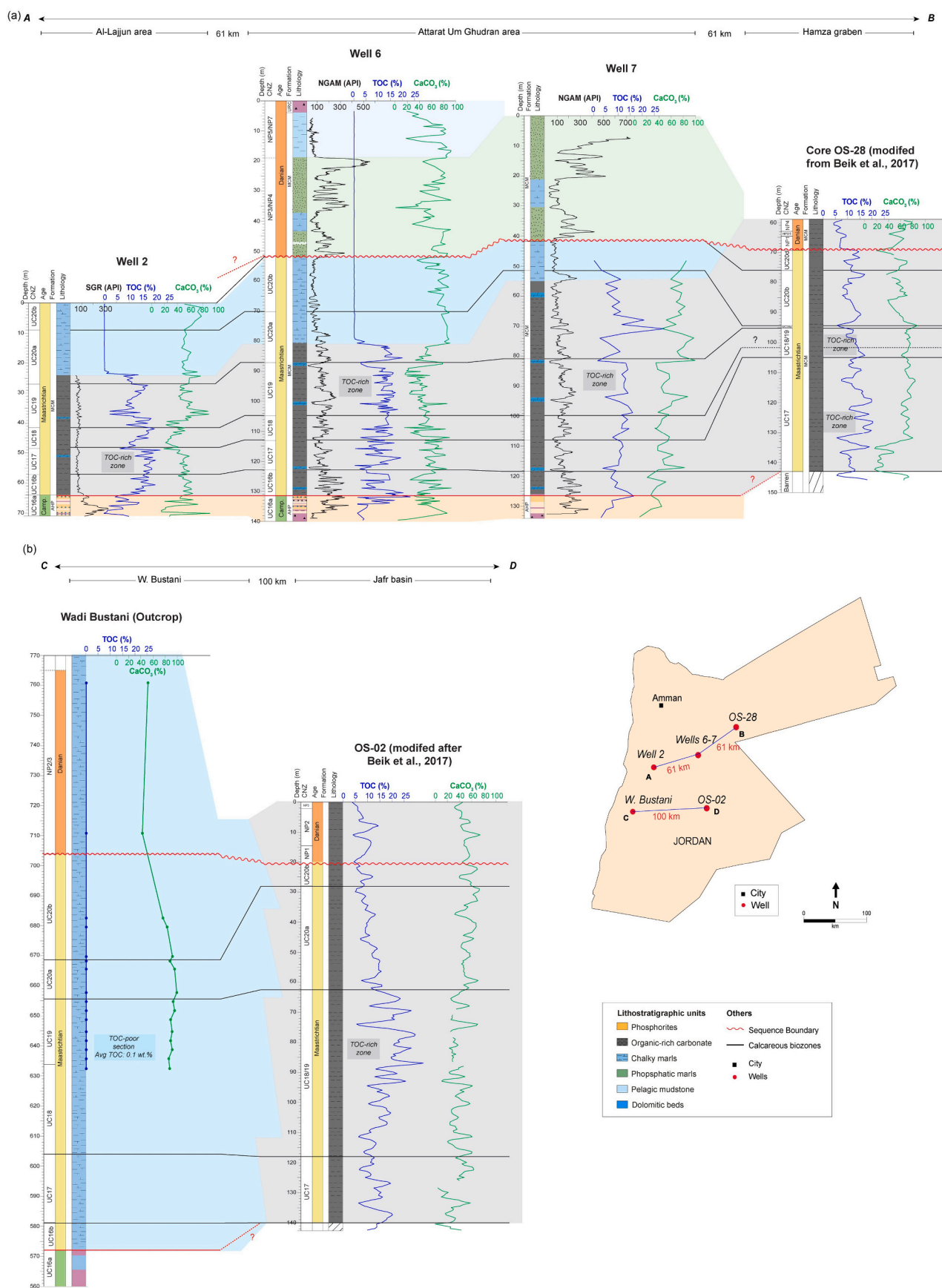


Fig. 11. Regional correlation of key sections of the Late Cretaceous–Paleocene succession in Jordan. (a) A–B transect across central Jordan. (b) C–D transect across southern Jordan. Calcareous nannofossil biozones for the Wadi Bustani section follow [Messaid et al. \(2025\)](#). Biostratigraphic and geochemical data for cores OS-02 and OS-28 are from [Alqudah et al. \(2015\)](#) and [Beik et al. \(2019\)](#).

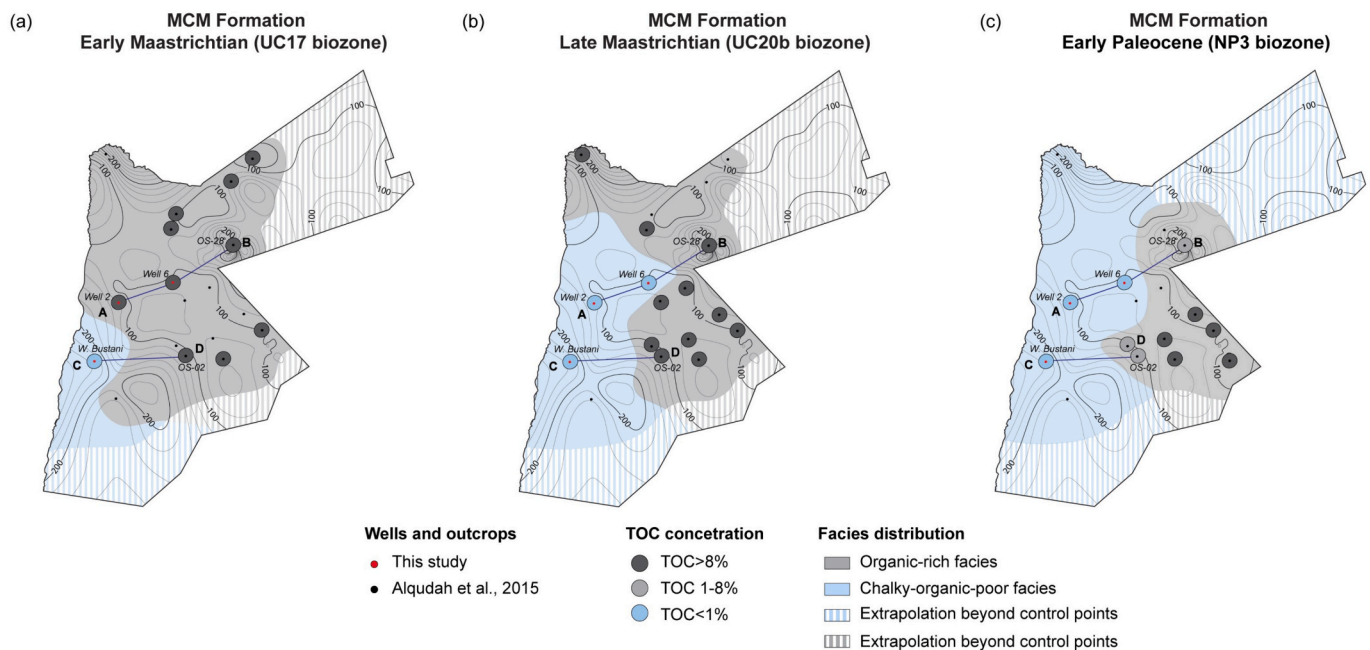


Fig. 12. Spatiotemporal distribution of organic-rich facies during the Late Cretaceous–Paleocene projected on isopach maps for the total thickness of the MCM Formation. The paleoshoreline was located farther south of present-day Jordan, likely in Saudi Arabia (Kalifi et al., 2025). Isopach maps are from Kalifi et al. (2025). Biostratigraphy for the Wadi Bustani section follows Messaoud et al. (2025). Biostratigraphic and geochemical data for cores OS-02 and OS-28 are from Alqudah et al. (2015) and Beik et al. (2019).

Consequently, peak paleoproductivity and the distribution of maximum TOC concentrations developed diachronously. The rapid sea level rise over the margin may have promoted the landward migration of upwelling cells, transporting nutrient-rich waters and expanding the OMZ toward more proximal settings as flooding advanced across the platform (Figs. 12a and 13b).

5.3.3. Late Maastrichtian marly chalk (middle MCM)

The shift from the organic-rich interval to the Marl Member of the Ghareb Formation in southern Negev is described as conformable and occurred at approximately 69.85 Ma (Abramovich et al., 1998; Ashckenazi-Polivoda et al., 2011). In central Jordan, the transition from the bituminous mudstone to the chalky interval within the MCM Formation occurred slightly later, during nannofossil zones UC20a/UC20b (approximately 69.18–69.0 Ma) in wells 2 and 6, likewise without evidence of a depositional hiatus. According to Ashckenazi-Polivoda et al. (2011), the sharp decline in TOC in southern Negev coincides with a shift in foraminiferal assemblages, reflecting a progressive decrease in surface-water productivity and an increase in bottom-water oxygenation.

Consistent with this interpretation, Giraldo-Gómez et al. (2018) and Beik et al. (2019) also suggested a transition to moderately oxygenated waters and increase of humid conditions based on benthic foraminiferal assemblages and clay mineral assemblages in the OS-02 core during UC20b, accompanied by a reduction in TOC concentrations (C–D transect, Fig. 11b). Their results further indicate that organic-matter accumulation persisted diachronously into the Paleocene in eastern and southern Jordan (Figs. 11a and 12c). Consequently, the associated oxygen-depleted zone likely remained active in more landward settings, where nutrient supply may have shifted from upwelling-driven to enhanced continental runoff, promoting continued organic-matter accumulation in these proximal areas (Figs. 12b and 13c).

Shoaling of OMZ boundaries is a characteristic process well

documented in major eastern boundary current systems, such as the California, Humboldt (Peru), and Benguela currents (Keeling et al., 2010; Gilly et al., 2013), as well as in semi-restricted basins such as the Black Sea (Anderson et al., 1994). Off Peru and northern Chile, the OMZ is exceptionally thick and strongly stratified, with its upper boundary shoaling seasonally to depths as shallow as ~50 m (Böning et al., 2004; Fuenzalida et al., 2009; Scholz et al., 2011). This shallow expression contrasts with more distal or open-ocean settings, where the OMZ weakens as a tongue-like shape, and its upper boundary deepens under the influence of subtropical gyre circulation (Fuenzalida et al., 2009). Shoaling of OMZ is often associated with periods of intensified continental weathering and enhanced nutrient delivery to proximal areas (Gilly et al., 2013). Increased runoff promotes water-column stratification and elevated primary productivity, shifting surface waters from oligotrophic “blue-water” conditions to euphotic-to-mesotrophic “green-water” conditions (sensu Wilmsen, 2003). Such nutrient-enriched waters have been shown to trigger widespread anoxia and enhance organic-matter preservation during the Cretaceous (Leckie et al., 2002; Erba, 2004; Batenburg et al., 2016; Poulton et al., 2015; Shmeit et al., 2022).

In the studied interval, the increase in detrital proxies and the decline in TOC observed during UC20b (Cycle M₆) may represent a regional expression of the latest Maastrichtian global oceanographic and climatic perturbations recorded globally by $\delta^{13}\text{C}$ oscillations at middle to high latitudes (Thibault and Gardin, 2007; Voigt et al., 2012; Jung et al., 2013), and locally with the Arabian Plate maximum flooding surface K180 at ~68 Ma (Sharland et al., 2004), which is documented across multiple regions along the Tethyan margin and may have contributed to a reconfiguration of the upwelling system (Philip et al., 2002; Powell and Moh'd, 2011; Beiranvand et al., 2014; Meilijson et al., 2014).

This large-scale reorganization likely enhanced continental runoff and reduced surface-water productivity, shifting toward more oligotrophic surface-water conditions, and favoring nannoplankton

production, as reflected by the abrupt increase in CaCO₃ content forming the bulk sediment and by higher sedimentation rates (Fig. 3). The more oxygenated depositional conditions also favored increased benthic activity, as evidenced by the common occurrence of bioturbation within the chalky units in Well 6. In contrast, suboxic to anoxic conditions persisted in more proximal areas of southern Jordan, enhanced by increased nutrient availability associated with runoff (Fig. 13c).

5.3.4. Danian chalky marls (uppermost MCM)

The deposition of the Upper Maastrichtian succession was interrupted by the global K/Pg boundary event, which is expressed across North Africa and the Arabian Plate as a regional unconformity (Kuss et al., 2000; Tantawy et al., 2001; Adatte et al., 2002; Philip et al., 2002; Farouk et al., 2014). In central Jordan, the overlying Danian interval records evidence of this hiatus, including significant submarine reworking of Cretaceous fauna (Messaoud et al., 2025) and possible subaerial exposure (Alhejoj et al., 2020). The Danian MCM facies reflects the continued flooding and encroachment of the northern Arabian Plate during the Paleocene (van der Meer et al., 2022; Farouk et al., 2014), resulting in the establishment of deeper marine conditions across the region, as evidenced by the occurrence of pelagic chalks, including radiolarian-rich mudstones in Unit 5 (Fig. 12c).

6. Conclusions

An integrated stratigraphic framework (high-resolution biostratigraphy, sedimentology, and geochemical analysis) of the early Maastrichtian-lower Danian succession has been established in seven wells located in central Jordan to investigate the evolution of paleo-productivity across a proximal-to-distal continental shelf margin transect of ~61 km. The main findings are summarized as:

- A stratigraphic time framework combining hierarchy and stacking patterns of depositional cycles, constrained by calcareous nannofossil biozonation, has been defined for the MCM Formation. This cyclicity reflects fluctuations in bottom-water energy, oxygenation, nutrient supply, and productivity, driven by longer term paleoceanographical processes, and shorter-term orbitally forced processes.
- During the latest Campanian, a regional shallowing phase favored the deposition of phosphorites under mesotrophic conditions, accompanied by sediment starvation and reworking of phosphatic deposits in central Jordan. This interpretation is consistent with the regional expression of the Campanian-Maastrichtian boundary across the northern Arabian Plate.
- The organic-rich interval of the MCM Formation in central Jordan was deposited from the early Maastrichtian (UC16b) to the late Maastrichtian (UC20a), spanning approximately 2.5 myr of sustained high productivity. The widespread and laterally continuous bituminous facies documented in this study are more consistent with basin-scale organic-matter enrichment controlled by enhanced paleoproductivity and redox conditions associated with an upwelling system, rather than by localized fault-controlled basins.
- The base of the late Maastrichtian is characterised by the abrupt transition from organic-rich to organic-poor chalky marl facies which is interpreted as the result of a regional reorganization of the paleoceanographic dynamics. The upwelling system weakened in the late Maastrichtian, and the main nutrient supply was replaced by continental run-off-driven proximal productivity (green water conditions) and an organic-poor, coccolith-dominated (blue water conditions) distal productivity.
- The depocenter of organic-rich deposition migrated diachronously eastward (landward) toward nearshore settings, where high productivity persisted into the Paleocene in southern Jordan and adjacent Saudi Arabia.

CRedit authorship contribution statement

Maria Ardila-Sanchez: Writing – review & editing, Writing – original draft, Visualization, Validation, Software, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Jihede Haj Messaoud:** Writing – review & editing, Validation, Methodology, Investigation, Data curation, Conceptualization. **Amir Kalifi:** Writing – review & editing, Validation, Investigation. **Khalil Ibrahim:** Writing – review & editing, Investigation, Conceptualization. **Muhammad Usman:** Validation, Methodology, Investigation, Data curation. **Israa S. Abu-Mahfouz:** Writing – review & editing, Validation. **Samer Aljurf:** Writing – review & editing. **Volker Vahrenkamp:** Writing – review & editing, Validation, Funding acquisition. **John H. Powell:** Writing – review & editing, Validation, Investigation. **Frans van Buchem:** Writing – review & editing, Validation, Supervision, Resources, Project administration, Investigation, Funding acquisition, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

This study was funded by King Abdullah University of Science and Technology (KAUST) through the baseline funding of FVB and VV. We acknowledge the Ministry of Energy and Mineral Resources (MEMR) and Karak International Oil (KIO) for providing the core samples used in this study. The Hashemite University (Az Zarqa, Jordan) is thanked for its logistical support during the fieldwork campaigns. We also extend sincere thanks to Elhadj Marwane Diallo and Mahmoud Mowafi (Earth System Science and Engineering program at KAUST) for their technical assistance with the Rock-Eval analyses, and to Ali Alibrahim for his help with SEM image acquisition. The authors thank the Editor, Prof. Thomas Steuber, and the two reviewers, Mohammad Razmjooei and an anonymous reviewer, for their constructive and insightful comments, which significantly improved the quality and clarity of this manuscript. JHP publishes with the permission of the Director, British Geological Survey (NERC).

Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.palaeo.2026.113986>.

Data availability

The authors confirm that all data necessary for supporting the scientific findings of this paper have been provided.

References

- Abed, A.M., 2013. The eastern Mediterranean phosphorite giants: an interplay between tectonics and upwelling. *GeoArabia* 18, 67–94. <https://doi.org/10.2113/geoarabia180267>.
- Abed, A.M., Amireh, B.S., 1999. Sedimentology, geochemistry, economic potential and palaeogeography of an Upper Cretaceous phosphorite belt in the southeastern desert of Jordan. *Cretac. Res.* 20, 119–133. <https://doi.org/10.1006/cres.1999.0147>.
- Abed, A.M., Arouri, K.R., Boreham, C.J., 2005. Source rock potential of the phosphorite-bituminous chalk-marl sequence in Jordan. *Mar. Pet. Geol.* 22, 413–425. <https://doi.org/10.1016/j.marpetgeo.2004.12.004>.
- Abed, A.M., Sadaqah, R., Al-Jazi, M., 2007. Sequence stratigraphy and evolution of Eshidiyya phosphorite platform, southern Jordan. *Sediment. Geol.* 198, 209–219. <https://doi.org/10.1016/j.sedgeo.2006.12.008>.
- Abramovich, S., Almogi-Labin, A., Benjamini, C., 1998. Decline of the Maastrichtian pelagic ecosystem based on planktic foraminifera assemblage change: Implication for the terminal Cretaceous faunal crisis. *Geology* 26, 63–66. [https://doi.org/10.1130/0091-7613\(1998\)026<0063:DOTMPE>2.3.CO;2](https://doi.org/10.1130/0091-7613(1998)026<0063:DOTMPE>2.3.CO;2).

- Abu-Mahfouz, I.S., Cartwright, J.A., Idiz, E., Hooker, J.N., Robinson, S., van den Boorn, S., 2019. Genesis and role of bitumen in fracture development during early catagenesis. *Pet. Geosci.* 25, 371–388. <https://doi.org/10.1144/petgeo2018-179>.
- Adatte, T., Keller, G., Stinnesbeck, W., 2002. Late cretaceous to early Paleocene climate and sea-level fluctuations: the Tunisian record. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 178, 165–196. [https://doi.org/10.1016/S0031-0182\(01\)00395-9](https://doi.org/10.1016/S0031-0182(01)00395-9).
- Ader, M., Sansjofre, P., Halverson, G.P., Busigny, V., Trindade, R.I., Kunzmann, M., Nogueira, A.C., 2014. Ocean redox structure across the late neoproterozoic oxygenation event: a nitrogen isotope perspective. *Earth Planet. Sci. Lett.* 396, 1–13. <https://doi.org/10.1016/j.epsl.2014.03.042>.
- Agard, P., Monié, P., Gerber, W., Omrani, J., Molinaro, M., Meyer, B., Labrousse, L., Vrielynck, B., Jolivet, L., Yamato, P., 2006. Transient, synobduction exhumation of Zagros blueschists inferred from P-T, deformation, time, and kinematic constraints: implications for Neotethyan wedge dynamics. *J. Geophys. Res. Solid Earth* 111 (B11).
- Ahmad, F., Baoumy, H., Farouk, S., Al-Kahtany, K., El-Sorogy, A., Kirk, J., 2020. Geochemistry and stable isotopes of the upper Campanian–lower Maastrichtian phosphorite-bearing sequence, Central Jordan: Implications for their age, origin, and diagenesis. *Geol. J.* 55, 4453–4468. <https://doi.org/10.1002/gj.3692>.
- Algeo, T.J., Li, C., 2020. Redox classification and calibration of redox thresholds in sedimentary systems. *Geochim. Cosmochim. Acta* 287, 8–26. <https://doi.org/10.1016/j.gca.2020.01.055>.
- Algeo, T.J., Lyons, T.W., 2006. Mo–total organic carbon covariation in modern anoxic marine environments: Implications for analysis of paleoredox and paleohydrographic conditions. *Palaeoceanography* 21. <https://doi.org/10.1029/2004PA001112>.
- Algeo, T.J., Maynard, J.B., 2004. Trace-element behavior and redox facies in core shales of Upper Pennsylvanian Kansas-type cyclothems. *Chem. Geol.* 206, 289–318. <https://doi.org/10.1016/j.chemgeo.2003.12.009>.
- Algeo, T.J., Tribouillard, N., 2009. Environmental analysis of paleoceanographic systems based on molybdenum–uranium covariation. *Chem. Geol.* 268, 211–225. <https://doi.org/10.1016/j.chemgeo.2009.09.001>.
- Alhejoj, I., Farouk, S., Bazeen, Y.S., Ahmad, F., 2020. Depositional sequences and sea-level changes of the upper Maastrichtian–middle Eocene succession in Central Jordan: evidence from foraminiferal biostratigraphy and paleoenvironments. *J. Afr. Earth Sci.* 161, 103663. <https://doi.org/10.1016/j.jafrearsci.2019.103663>.
- Ali Hussein, M., Alqudah, M., Blessenohl, M., Podlaha, O.G., Mutterlose, J., 2015. Depositional environment of late cretaceous to Eocene organic-rich marls from Jordan. *GeoArabia* 20, 191–210. <https://doi.org/10.2113/geoarabia2001191>.
- Aljurf, S., Abu-Mahfouz, I.S., El-Husseiny, A., Hassan, A., Chan, S., Finkbeiner, T., 2026. Vertical variability of geochemical, petrophysical, and rock mechanical properties in upper cretaceous organic-rich carbonate source rocks, Jordan. *J. Asian Earth Sci.* X 15, 100220. <https://doi.org/10.1016/j.jaesx.2026.100220>.
- Almogi-Labin, A., Bein, A., Sass, E., 1993. Late cretaceous upwelling system along the southern Tethys margin (Israel): interrelationship between productivity, bottom water environments, and organic matter preservation. *Palaeoceanography* 8, 671–690. <https://doi.org/10.1029/93PA02197>.
- Alqudah, M., Hussein, M.A., van den Boorn, S., Podlaha, O.G., Mutterlose, J., 2015. Biostratigraphy and depositional setting of Maastrichtian–Eocene oil shales from Jordan. *Mar. Pet. Geol.* 60, 87–104. <https://doi.org/10.1016/j.marpetgeo.2014.07.025>.
- Alsenz, H., Regnery, J., Ashkenazi-Polivoda, S., Meilijson, A., Ron-Yankovich, L., Abramovich, S., Illner, P., Almogi-Labin, A., Feinstein, S., Berner, Z., 2013. Sea surface temperature record of a late cretaceous tropical Southern Tethys upwelling system. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 392, 350–358. <https://doi.org/10.1016/j.palaeo.2013.09.013>.
- Anderson, R., Lyons, T., Cowie, G., 1994. Sedimentary record of a shoaling of the oxic/anoxic interface in the Black Sea. *Mar. Geol.* 116, 373–384. [https://doi.org/10.1016/0025-3227\(94\)90052-3](https://doi.org/10.1016/0025-3227(94)90052-3).
- Andrews, I., 1992. Cretaceous and Paleogene lithostratigraphy in the subsurface of Jordan. *Min. Energy Miner. Resour. C*, 13–52.
- Arthur, M., Brumsack, H.-J., Jenkyns, H., Schlanger, S., 1990. Stratigraphy, Geochemistry, and Paleocyanography of Organic Carbon-Rich Cretaceous Sequences, Cretaceous Resources, Events and Rhythms: Background and Plans for Research. Springer, pp. 75–119. https://doi.org/10.1007/978-94-015-6861-6_6.
- Ashkenazi-Polivoda, S., Abramovich, S., Almogi-Labin, A., Schneider-Mor, A., Feinstein, S., Püttmann, W., Berner, Z., 2011. Paleoenvironments of the latest cretaceous oil shale sequence, Southern Tethys, Israel, as an integral part of the prevailing upwelling system. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 305, 93–108. <https://doi.org/10.1016/j.palaeo.2011.02.018>.
- Batenburg, S.J., De Vleeschouwer, D., Sprovieri, M., Hilgen, F.J., Gale, A.S., Singer, B.S., Montanari, A., 2016. Orbital control on the timing of oceanic anoxia in the late cretaceous. *Clim. Past* 12 (10), 1995–2009. <https://doi.org/10.5194/cp-12-1995-2016>.
- Baturin, G.N., Kochenov, A.V., 2001. Uranium in phosphorites. *Lithol. Miner. Resour.* 36 (4), 303–321. <https://doi.org/10.1023/A:1010406103447>.
- Bazeen, Y.S., Farouk, S., El-Balkiemy, A.F., Ahmad, F., 2020. Late Maastrichtian–early Danian Sea-level changes at Gebel Tawil Shihq in Central Jordan: Implications of chemostratigraphy and foraminiferal biostratigraphy. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 556, 109875. <https://doi.org/10.1016/j.palaeo.2020.109875>.
- Behar, F., Beaumont, V., Penteado, H.D.B., 2001. Rock-Eval 6 technology: performances and developments. *Oil Gas Sci. Technol.* 56, 111–134. <https://doi.org/10.2516/ogst.2001013>.
- Beik, I., Adatte, T., Podlaha, O.G., Mutterlose, J., 2019. Maastrichtian to early Paleocene sea level changes and climatic evolution on the southeastern Tethys margin. *Mar. Pet. Geol.* 100, 285–296. <https://doi.org/10.1016/j.marpetgeo.2018.11.018>.
- Beik, I., Gómez, V.G., Podlaha, O.G., Mutterlose, J., 2017. Microfacies and depositional environment of late cretaceous to early Paleocene oil shales from Jordan. *Arab. J. Geosci.* 10, 1–24. <https://doi.org/10.1007/s12517-017-3118-6>.
- Beik, I., Gómez, V.G., Podlaha, O.G., Mutterlose, J., 2018. The 613C record of Maastrichtian–Paleocene oil shales from Jordan–Stratigraphic and environmental implications for an epicontinental setting. *J. Afr. Earth Sci.* 143, 134–144. <https://doi.org/10.1016/j.jafrearsci.2018.03.022>.
- Beiranvand, B., Ghasemi-Nejad, E., Kamali, M.R., Ahmadi, A., 2014. Sequence stratigraphy of the late Cretaceous–Paleocene Gurpi Formation in southwest Iran. *GeoArabia* 19, 89–102. <https://doi.org/10.2113/geoarabia190289>.
- Bennett, W.W., Canfield, D.E., 2020. Redox-sensitive trace metals as paleoredox proxies: a review and analysis of data from modern sediments. *Earth Sci. Rev.* 204, 103175. <https://doi.org/10.1016/j.earscirev.2020.103175>.
- Bjerrum, C., Bendtsen, J., Legarth, J., 2006. Modeling organic carbon burial during sea level rise with reference to the Cretaceous. *Geochim. Geophys. Geosyst.* 7. <https://doi.org/10.1029/2005GC001032>.
- Bohacs, K.M., Grabowski, G.J., Carroll, A.R., Mankiewicz, P.J., Miskell-Gerhardt, K.J., Schwalbach, J.R., Wegner, M.B., Simo, J.T., 2005. Production, Destruction, and Dilution—The Many Paths to Source-rock Development. <https://doi.org/10.2110/pec.05.82.0061>.
- Bohacs, K.M., Norton, I.O., Neal, J.E., Gilbert, D., 2020. The accumulation of organic—Matter-rich rocks within an earth system's framework: The integrated roles of plate tectonics, atmosphere, ocean, and biota through the Phanerozoic. In: *Regional Geology and Tectonics*. Elsevier, pp. 721–744. <https://doi.org/10.1016/B978-0-444-64134-2.00023-7>.
- Böning, P., Brumsack, H.-J., Böttcher, M.E., Schnetger, B., Kriete, C., Kallmeyer, J., Borchers, S.L., 2004. Geochemistry of Peruvian near-surface sediments. *Geochim. Cosmochim. Acta* 68, 4429–4451. <https://doi.org/10.1016/j.gca.2004.04.027>.
- Böning, P., Shaw, T., Pahnke, K., Brumsack, H.-J., 2015. Nickel as indicator of fresh organic matter in upwelling sediments. *Geochim. Cosmochim. Acta* 162, 99–108. <https://doi.org/10.1016/j.gca.2015.04.027>.
- Borchers, S., Schnetger, B., Böning, P., Brumsack, H.J., 2005. Geochemical signatures of the Namibian diatom belt: perennial upwelling and intermittent anoxia. *Geochim. Geophys. Geosyst.* 6. <https://doi.org/10.1029/2004GC000886>.
- Bown, P., Young, J.R., 1998. Calcareous nannofossil biostratigraphy. In: *British Micropaleontology Society Series*. Chapman & Hall, London, pp. 226–266.
- Brumsack, H.J., 1986. The inorganic geochemistry of cretaceous black shales (DSDP Leg 41) in comparison to modern upwelling sediments from the Gulf of California. *Geol. Soc. Lond. Spec. Publ.* 21, 447–462. <https://doi.org/10.1144/GSL.SP.1986.021.01.30>.
- Brumsack, H.-J., 2006. The trace metal content of recent organic carbon-rich sediments: implications for cretaceous black shale formation. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 232, 344–361. <https://doi.org/10.1016/j.palaeo.2005.05.011>.
- Burnett, J., 1998. Upper cretaceous. In: *Calcareous Nannofossil Biostratigraphy*, pp. 132–199.
- Calvert, S., Pedersen, T., 2007. Chapter fourteen elemental proxies for palaeoclimatic and paleoceanographic variability in marine sediments: interpretation and application. *Dev. Mar. Geol.* 1, 567–644. [https://doi.org/10.1016/S1572-5480\(07\)01019-6](https://doi.org/10.1016/S1572-5480(07)01019-6).
- Canadas, F., Papineau, D., Algeo, T.J., Li, C., 2024. Upwelling-driven high organic production in the late Ediacaran. *Commun. Earth Environ.* 5, 461. <https://doi.org/10.1038/s43247-024-01632-z>.
- Canfield, D.E., 1994. Factors influencing organic carbon preservation in marine sediments. *Chem. Geol.* 114, 315–329. [https://doi.org/10.1016/0009-2541\(94\)90061-2](https://doi.org/10.1016/0009-2541(94)90061-2).
- Chirakal, T., Messaoud, J.H., Alibrahim, A., Ibrahim, K., Grélaud, C., Kalifi, A., Powell, J. H., van Buchem, F., 2025. Integrated carbon isotope stratigraphy and biostratigraphy of Cenomanian to Turonian carbonates from Jordan—an updated age model and sequence stratigraphic correlations with Oman. *Cretac. Res.*, 106207. <https://doi.org/10.1016/j.cretres.2025.106207>.
- Craigie, N., 2018. Principles of Elemental Chemostratigraphy. Springer. <https://doi.org/10.1007/978-3-319-71216-1>.
- Dale, A.W., Sommer, S., Lomnitz, U., Montes, I., Treude, T., Liebetrau, V., Gier, J., Hensen, C., Dengler, M., Stolpovsky, K., 2015. Organic carbon production, mineralisation and preservation on the Peruvian margin. *Biogeosciences* 12, 1537–1559. <https://doi.org/10.5194/bg-12-1537-2015>.
- Dembicki, H., 2009. Three common source rock evaluation errors made by geologists during prospect or play appraisals. *AAPG Bull.* 93 (3), 341–356. <https://doi.org/10.1306/10230808076>.
- Demaion, G.J., Moore, G.T., 1980. Anoxic environments and oil source bed genesis. *Org. Geochem.* 2, 9–31. [https://doi.org/10.1016/0146-6380\(80\)90017-0](https://doi.org/10.1016/0146-6380(80)90017-0).
- Dunham, R.J., 1962. Classification of Carbonate Rocks According to Depositional Texture.
- Edelman-Furstenberg, Y., 2009. Cyclic upwelling facies along the late cretaceous southern Tethys (Israel): taphonomic and ichnofacies evidence of a high-productivity mosaic. *Cretac. Res.* 30, 847–863. <https://doi.org/10.1016/j.cretres.2009.01.005>.
- El Albani, A., Kuhnt, W., Luderer, F., Herbin, J., Caron, M., 1999. Palaeoenvironmental evolution of the Late Cretaceous sequence in the Tarfaya Basin (southwest of Morocco). *Geol. Soc. Lond. Spec. Publ.* 153 (1), 223–240. <https://doi.org/10.1144/GSL.SP.1999.153.01.14>.
- El Ayyat, A.M., 2015. Lithostratigraphy, sedimentology, and cyclicity of the Duwi Formation (late cretaceous) at Abu Tartur plateau, Western Desert of Egypt: evidences for reworking and redeposition. *Arab. J. Geosci.* 8, 99–124. <https://doi.org/10.1007/s12517-013-1169-x>.

- El-Azabi, M.H., Farouk, S., 2011. High-resolution sequence stratigraphy of the Maastrichtian-Ypresian succession along the eastern scarp face of Kharga Oasis, southern Western Desert, Egypt. *Sedimentology* 58, 579–617. <https://doi.org/10.1111/j.1365-3091.2010.01175.x>.
- Emerson, S., 1985. Organic carbon preservation in marine sediments. In: *The Carbon Cycle and Atmospheric CO₂: Natural Variations Archean to Present*, 32, pp. 78–87. <https://doi.org/10.1029/GM032p0078>.
- Emmings, J.F., Davies, S.J., Vane, C.H., Moss-Hayes, V., Stephenson, M.H., 2020. From marine bands to hybrid flows: sedimentology of a Mississippian black shale. *Sedimentology* 67, 261–304. <https://doi.org/10.1111/sed.12642>.
- Erba, E., 2004. Calcareous nannofossils and Mesozoic oceanic anoxic events. *Mar. Micropaleontol.* 52 (1–4), 85–106. <https://doi.org/10.1016/j.marmicro.2004.04.007>.
- Farouk, S., Marzouk, A.M., Ahmad, F., 2014. The Cretaceous/Paleogene boundary in Jordan. *J. Asian Earth Sci.* 94, 113–125. <https://doi.org/10.1016/j.jseas.2014.08.015>.
- Fernández-Martínez, J., Rodríguez-Tovar, F.J., Piñuela, L., Martínez-Ruiz, F., García-Ramos, J.C., 2021. Bottom-and pore-water oxygenation during the early Toarcian Oceanic Anoxic Event (T-OAE) in the Asturian Basin (N Spain): Ichological information to improve facies analysis. *Sediment. Geol.* 419, 105909. <https://doi.org/10.1016/j.sedgeo.2021.105909>.
- Fleurance, S., Cuney, M., Malartre, F., Reyx, J., 2013. Origin of the extreme polymetallic enrichment (Cd, Cr, Mo, Ni, U, V, Zn) of the Late Cretaceous–Early Tertiary Belqa Group, central Jordan. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 369, 201–219. <https://doi.org/10.1016/j.palaeo.2012.10.020>.
- Flexer, A., Reymont, R., 1989. Note on cretaceous transgressive peaks and their relation to geodynamic events for the Arabo-Nubian and the Northern African shields. *J. Afr. Earth Sci.* 8, 65–73. [https://doi.org/10.1016/S0899-5362\(89\)80010-7](https://doi.org/10.1016/S0899-5362(89)80010-7).
- Flügel, E., Munnecke, A., 2010. *Microfacies of Carbonate Rocks: Analysis, Interpretation and Application*. Springer. <https://doi.org/10.1007/978-3-642-03796-2>.
- Fuenzalida, R., Schneider, W., Garcés-Vargas, J., Bravo, L., Lange, C., 2009. Vertical and horizontal extension of the oxygen minimum zone in the eastern South Pacific Ocean. *Deep-Sea Res. II Top. Stud. Oceanogr.* 56, 992–1003. <https://doi.org/10.1016/j.dsr2.2008.11.001>.
- Gale, A.S., Hardenbol, J., Hathway, B., Kennedy, W.J., Young, J.R., Phansalkar, V., 2002. Global correlation of Cenomanian (Upper cretaceous) sequences: evidence for Milankovitch control on sea level. *Geology* 30, 291–294. [https://doi.org/10.1130/0091-7613\(2002\)030<0291:GOCUC>2.0.CO;2](https://doi.org/10.1130/0091-7613(2002)030<0291:GOCUC>2.0.CO;2).
- Gambacorta, G., Bersezio, R., Weissert, H., Erba, E., 2016. Onset and demise of cretaceous oceanic anoxic events: the coupling of surface and bottom oceanic processes in two pelagic basins of the western Tethys. *Palaeogeography* 31, 732–757. <https://doi.org/10.1002/2015PA002922>.
- Gilly, W.F., Beman, J.M., Litvin, S.Y., Robison, B.H., 2013. Oceanographic and biological effects of shoaling of the oxygen minimum zone. *Annu. Rev. Mar. Sci.* 5, 393–420. <https://doi.org/10.1146/annurev-marine-120710-100849>.
- Giraldo-Gómez, V.M., Beik, I., Podlaha, O.G., Mutterlose, J., 2018. A paleoenvironmental analyses of benthic foraminifera from Upper Cretaceous–lower Paleocene oil shales of Jordan. *Cretac. Res.* 91, 1–13. <https://doi.org/10.1016/j.cretres.2018.04.021>.
- Glenn, C.R., Arthur, M.A., 1990. Anatomy and origin of a cretaceous phosphorite-green sand giant, Egypt. *Sedimentology* 37, 123–154. <https://doi.org/10.1111/j.1365-3091.1990.tb01986.x>.
- Grohmann, S., Littke, R., Abu-Mahfouz, I., Gaus, G., Klaver, J., Thüms, N., Schulte, P., Patzek, T., Vahrenkamp, V., 2023. The deposition of type II-S Jordan oil shale in the context of late cretaceous source rock formation in the Eastern Mediterranean realm. Insights from organic and inorganic geochemistry and petrography. *Mar. Pet. Geol.* 148, 106058. <https://doi.org/10.1016/j.marpetgeo.2022.106058>.
- Guiraud, R., Bosworth, W., 1997. Senonian basin inversion and rejuvenation of rifting in Africa and Arabia: synthesis and implications to plate-scale tectonics. *Tectonophysics* 282, 39–82.
- Guiraud, R., Bosworth, W., Thierry, J., Delplanque, A., 2005. Phanerozoic geological evolution of Northern and Central Africa: an overview. *J. Afr. Earth Sci.* 43 (1–3), 83–143. <https://doi.org/10.1016/j.jafrearsci.2005.07.017>.
- Gvirtzman, G., Almogi-Labin, A., Moshkovitz, S., Lewy, Z., Honigstein, A., Reiss, Z., 1989. Upper cretaceous high-resolution multiple stratigraphy, northern margin of the Arabian platform, Central Israel. *Cretac. Res.* 10, 107–135. [https://doi.org/10.1016/0195-6671\(89\)90001-3](https://doi.org/10.1016/0195-6671(89)90001-3).
- Hakimi, M.H., Abdullah, W.H., Alqudah, M., Makeen, Y.M., Mustapha, K.A., 2016. Reducing marine and warm climate conditions during the late cretaceous, and their influence on organic matter enrichment in the oil shale deposits of North Jordan. *Int. J. Coal Geol.* 165, 173–189. <https://doi.org/10.1016/j.coal.2016.08.015>.
- Hart, B., Steen, A., 2015. Programmed pyrolysis (Rock-Eval) data and shale paleoenvironmental analyses: a review. *Interpretation* 3, SH41–SH58. <https://doi.org/10.1190/INT-2014-0168.1>.
- Hay, W.W., 1995. *Palaeogeography of Marine Organic-Carbon-Rich Sediments*. <https://doi.org/10.1130/St40595C2>.
- Hay, W.W., 2009. Cretaceous Oceans and Ocean Modeling. <https://doi.org/10.2110/sepm.091.243>.
- Hein, J.R., Yeh, H.W., Gunn, S.H., Sliter, W.V., Benninger, L.M., Wang, C.H., 1993. Two major Cenozoic episodes of phosphogenesis recorded in equatorial Pacific seamount deposits. *Palaeogeography* 8, 293–311. <https://doi.org/10.1029/93PA00320>.
- Huc, A.Y., van Buchem, F.S.P., Colletta, B., 2005. Stratigraphic Control on Source-Rock Distribution: First and Second Order Scale. <https://doi.org/10.2110/pec.05.82.0225>.
- Ikeda, M., Tada, R., Ozaki, K., 2017. Astronomical pacing of the global silica cycle recorded in Mesozoic bedded cherts. *Nat. Commun.* 8 (1), 15532. <https://doi.org/10.1038/ncomms15532>.
- Inthorn, M., Wagner, T., Scheeder, G., Zabel, M., 2006. Lateral transport controls distribution, quality, and burial of organic matter along continental slopes in high-productivity areas. *Geology* 34, 205–208. <https://doi.org/10.1130/G22153.1>.
- Jenkyns, H.C., 2010. Geochemistry of oceanic anoxic events. *Geochim. Geophys. Geosyst.* 11. <https://doi.org/10.1029/2009GC002788>.
- Jiang, C., Chen, Z., Lavoie, D., Percival, J.B., Kabanov, P., 2017. Mineral carbon MinC (%) from Rock-Eval analysis as a reliable and cost-effective measurement of carbonate contents in shale source and reservoir rocks. *Mar. Pet. Geol.* 83, 184–194. <https://doi.org/10.1016/j.marpetgeo.2017.03.017>.
- Jung, C., Voigt, S., Friedrich, O., Koch, M.C., Frank, M., 2013. Campanian-Maastrichtian Ocean circulation in the tropical Pacific. *Palaeogeography* 28, 562–573. <https://doi.org/10.1002/palo.20051>.
- Kalifi, A., Messaoud, J.H., Ibrahim, K., Powell, J.H., van Buchem, F., 2025. Tectono-stratigraphic evolution of the late Cretaceous-Eocene of Jordan and implications for the Arabian Plate convergent margin phase. *Mar. Pet. Geol.* <https://doi.org/10.1016/j.marpetgeo.2025.107525>, 107525.
- Katz, B.J., 2005. Controlling Factors on Source Rock Development—A Review of Productivity, Preservation, and Sedimentation Rate. <https://doi.org/10.2110/pec.05.82.0007>.
- Keeling, R.F., Körtzinger, A., Gruber, N., 2010. Ocean deoxygenation in a warming world. *Annu. Rev. Mar. Sci.* 2, 199–229. <https://doi.org/10.1146/annurev.marine.010908.163855>.
- Khoury, H.N., 2019. Industrial rocks and minerals of Jordan: a review. *Arab. J. Geosci.* 12, 619. <https://doi.org/10.1007/s12517-019-4750-0>.
- Kuss, J., Westerhold, T., Groß, U., Bauer, J., Lüning, S., 2000. Mapping of late cretaceous stratigraphic sequences along a Syrian Arc Uplift-examples from the Aref el Naqa/Eastern Sinai. *Middle East Res. Cent., Ain Shams Univ., Earth Sci. Serv.* 14, 171–191.
- Lazar, O., Schieber, J., 2022. New Albany Shale, Illinois Basin, USA—Devonian Carbonaceous Mudstone Accumulation in an Epicratonic Sea: Stratigraphic Insights from Outcrop and Subsurface Data. <https://doi.org/10.1306/137123043860>.
- Lazar, O.R., Bohacs, K.M., Schieber, J., Macquaker, J.H.S., Demko, T.M., 2015. Mudstones: nomenclature and description guidelines, mudstone primer: lithofacies variations, diagnostic criteria, and sedimentologic–stratigraphic implications at lamina to bedset scale. *SEPM Society for Sedimentary Geology*. <https://doi.org/10.2110/sepmcsp.12.003>, 0.
- Leckie, R.M., Bralower, T.J., Cashman, R., 2002. Oceanic anoxic events and plankton evolution: biotic response to tectonic forcing during the mid-Cretaceous. *Palaeogeography* 17, 13–11–13–29. <https://doi.org/10.1029/2001PA000623>.
- Levelli, B., van Buchem, F., 2026. The stratigraphic record: Preservation, facies and depositional sequences. In: Levelli, B. (Ed.), *Reading's Sedimentary Environments: Processes, Facies and Stratigraphy*. John Wiley & Sons, pp. 195–260.
- Levy, A., Schneider-Mor, A., Gelman, F., Kamshyly Jr., A., 2023. Petrological and chemical characterizations of pristine and reworked phosphorites from Negev, Israel: insights into industrial usage. *J. Afr. Earth Sci.* 199, 104843. <https://doi.org/10.1016/j.jafrearsci.2023.104843>.
- Li, Y., Schieber, J., 2015. On the origin of a phosphate enriched interval in the Chattanooga Shale (Upper Devonian) of Tennessee—A combined sedimentologic, petrographic, and geochemical study. *Sediment. Geol.* 329, 40–61. <https://doi.org/10.1016/j.sedgeo.2017.12.015>.
- Li, Z., Schieber, J., 2018. Detailed facies analysis of the Upper cretaceous Tununk Shale Member, Henry Mountains Region, Utah: implications for mudstone depositional models in epicontinental seas. *Sediment. Geol.* 364, 141–159. <https://doi.org/10.1016/j.sedgeo.2017.12.015>.
- Litak, R.K., Barazangi, M., Beauchamp, W., Seber, D., Brew, G., Sawaf, T., Al-Youssef, W., 1997. Mesozoic-Cenozoic evolution of the intraplate Euphrates fault system, Syria: implications for regional tectonics. *J. Geol. Soc. Lond.* 154 (4), 653–666.
- Lüning, S., Kolonic, S., 2003. Uranium spectral gamma-ray response as a proxy for organic richness in black shales: applicability and limitations. *J. Pet. Geol.* 26 (2), 153–174. <https://doi.org/10.1111/j.1747-5457.2003.tb00023.x>.
- Macellari, C.E., De Vries, T.J., 1987. Late Cretaceous upwelling and anoxic sedimentation in northwestern South America. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 59, 279–292. [https://doi.org/10.1016/0031-0182\(87\)90086-1](https://doi.org/10.1016/0031-0182(87)90086-1).
- Martini, E., 1970. Standard Palaeogene calcareous nannoplankton zonation. *Nature* 226, 560–561. <https://doi.org/10.1038/226560a0>.
- Matthews, K.J., Seton, M., Müller, R.D., 2012. A global-scale plate reorganization event at 105–100 Ma. *Earth Planet. Sci. Lett.* 355, 283–298.
- May, P.R., 1991. The Eastern Mediterranean Mesozoic Basin: evolution and oil habitat. *AAPG Bull.* 75, 1215–1232. <https://doi.org/10.1306/0C9B2911-1710-11D7-8645000102C1865D>.
- McArthur, J., Algeo, T., Van de Schootbrugge, B., Li, Q., Howarth, R., 2008. Basinal restriction, black shales, Re-Os dating, and the early Toarcian (Jurassic) oceanic anoxic event. *Palaeogeography* 23. <https://doi.org/10.1029/2008PA001607>.
- Meilijson, A., Ashkenazi-Polivoda, S., Ron-Yankovich, L., Illner, P., Alsenz, H., Speijer, R.P., Almogi-Labin, A., Feinstein, S., Berner, Z., Püttmann, W., 2014. Chronostratigraphy of the Upper Cretaceous high productivity sequence of the southern Tethys, Israel. *Cretac. Res.* 50, 187–213. <https://doi.org/10.1016/j.cretres.2014.04.006>.
- Messaoud, J.H., Kalifi, A., Alibrahim, A., Ibrahim, K., Chirakal, T., Ardila-Sanchez, M., Abu Leila, W., Ben Chabaane, N., Grélaud, C., Powell, J.H., van Buchem, F., 2025. Chronostratigraphy of the mixed Upper Cretaceous deposits at the northern margin of the Arabian Plate (Jordan). *News. Stratigr.* <https://doi.org/10.1127/nos/2025/0858>.
- Mikbel, S., Zacher, W., 1986. Fold structures in northern Jordan. *Neues Jahrbuch für Geologie und Paläontologie Monatshefte* 4, 248–256. <https://doi.org/10.1127/njgpm/1986/1986/248>.

- Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., Pekar, S.F., 2005. The Phanerozoic record of global sea-level change. *Science* 310, 1293–1298. <https://doi.org/10.1126/science.1116412>.
- Parrish, J.T., 1995. Paleogeography of corg-rich rocks and the preservation versus production controversy. In: Huc, a.-Y. (Ed.), *Paleogeography, Paleoclimate, and Source Rocks*. American Association of Petroleum Geologists. <https://doi.org/10.1306/St40595C1>, 0.
- Parrish, J., Curtis, R., 1982. Atmospheric circulation, upwelling, and organic-rich rocks in the Mesozoic and Cenozoic eras. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 40, 31–66. [https://doi.org/10.1016/0031-0182\(82\)90084-0](https://doi.org/10.1016/0031-0182(82)90084-0).
- Pedersen, T.F., Calvert, S.E., 1990. Anoxia vs. productivity: what controls the formation of organic-carbon-rich sediments and sedimentary rocks? *AAPG Bull.* 74 (4), 454–466. <https://doi.org/10.1306/0C9B232B-1710-11D7-8645000102C1865D>.
- Perch-Nielsen, K., 1985. Cenozoic calcareous nannofossils. In: *Plankton Stratigraphy*, pp. 427–455.
- Philip, J.M., Roger, J., Vaslet, D., Cecca, F., Gardin, S., Memesh, A.M., 2002. Sequence stratigraphy, biostratigraphy and paleontology of the Maastrichtian-Paleocene Aruma Formation in outcrop in Saudi Arabia. *GeoArabia* 7, 699–718. <https://doi.org/10.2113/geoarabia0704699>.
- Piper, D.Z., Link, P.K., 2002. An upwelling model for the Phosphoria Sea: a Permian, ocean-margin sea in the Northwest United States. *AAPG Bull.* 86, 1217–1235. <https://doi.org/10.1306/61EEDC60-173E-11D7-8645000102C1865D>.
- Plado, J., Ots, S., Puura, V., Aosaar, H., 2016. Interpretation of gamma-ray logs of the stratified oil shale seam in the Attarat Um Ghudran deposit, Central Jordan. *Oil Shale* 33, 340–356. <https://doi.org/10.3176/oil.2016.4.03>.
- Poulton, S.W., Henkel, S., März, C., Urquhart, H., Flögel, S., Kasten, S., Sinninghe Damsté, J.S., Wagner, T., 2015. A continental-weathering control on orbitally driven redox-nutrient cycling during cretaceous Oceanic Anoxic Event 2. *Geology* 43, 963–966. <https://doi.org/10.1130/G36837.1>.
- Powell, J., 1989. Stratigraphy and sedimentation of the Phanerozoic rocks in central and south Jordan. *Bulletin. Geol. Dir. Nat. Resour. Authority. Amman*. <http://pascal-francis.inist.fr/vibad/index.php?action=getRecordDetail&idt=6402231>.
- Powell, J.H., Moh'd, B.K., 2011. Evolution of Cretaceous to Eocene alluvial and carbonate platform sequences in central and South Jordan. *GeoArabia* 16, 29–82. <https://doi.org/10.2113/geoarabia160429>.
- Prokoph, A., Thurow, J., 2000. Diachronous pattern of Milankovitch cyclicity in late Albian pelagic marlstones of the North German Basin. *Sediment. Geol.* 134, 287–303. [https://doi.org/10.1016/S0037-0738\(00\)00050-6](https://doi.org/10.1016/S0037-0738(00)00050-6).
- Pucéat, E., Lécuyer, C., Reisberg, L., 2005. Neodymium isotope evolution of NW Tethyan upper ocean waters throughout the cretaceous. *Earth Planet. Sci. Lett.* 236, 705–720. <https://doi.org/10.1016/j.epsl.2005.03.015>.
- Pufahl, P.K., Grimm, K.A., Abed, A.M., Sadaqah, R.M., 2003. Upper cretaceous (Campanian) phosphorites in Jordan: implications for the formation of a south Tethyan phosphorite giant. *Sediment. Geol.* 161, 175–205. [https://doi.org/10.1016/S0037-0738\(03\)00070-8](https://doi.org/10.1016/S0037-0738(03)00070-8).
- Puura, V., Soesoo, A., Voolma, M., Konsa, M., Aosaar, H., 2017. Petrography and mineralogy of the Attarat Um Ghudran oil shale, central Jordan. *Oil Shale* 34 (2), 110–128.
- Ricou, L.-E., 1994. Tethys reconstructed: plates, continental fragments and their Boundaries since 260 Ma from Central America to South-eastern Asia. *Geodin. Acta* 7 (4), 169–218.
- Rider, M.H., 2002. *The geological interpretation of well logs*, 2nd edn. Rider French Consulting Ltd, Scotland.
- Robinson, R.S., Meyers, P.A., Murray, R.W., 2002. Geochemical evidence for variations in delivery and deposition of sediment in Pleistocene light–dark color cycles under the Benguela Current Upwelling System. *Mar. Geol.* 180, 249–270. [https://doi.org/10.1016/S0025-3227\(01\)00217-1](https://doi.org/10.1016/S0025-3227(01)00217-1).
- Sageman, B.B., Rich, J., Arthur, M.A., Birchfield, G., Dean, W., 1997. Evidence for Milankovitch periodicities in Cenomanian-Turonian lithologic and geochemical cycles, Western Interior USA. *J. Sediment. Res.* 67, 286–302. <https://doi.org/10.1306/D4268554-2B26-11D7-8648000102C1865D>.
- Schieber, J., 1998. Sedimentary features indicating erosion, condensation, and hiatuses in the Chattanooga Shale of Central Tennessee: relevance for sedimentary and stratigraphic evolution. In: *Shales and Mudstones (Volume I, Basin Studies, Sedimentology, and Paleontology)*, pp. 187–215.
- Schieber, J., Southard, J.B., Schimmelmann, A., 2010. Lenticular shale fabrics resulting from intermittent erosion of water-rich muds—interpreting the rock record in the light of recent flume experiments. *J. Sediment. Res.* 80, 119–128. <https://doi.org/10.2110/jsr.2010.005>.
- Schlager, W., 2005. *Carbonate Sedimentology and Sequence Stratigraphy*, No. 8. SEPM Book for Sed Geology.
- Scholz, F., Hensen, C., Noffke, A., Rohde, A., Liebetrau, V., Wallmann, K., 2011. Early diagenesis of redox-sensitive trace metals in the Peru upwelling area—response to ENSO-related oxygen fluctuations in the water column. *Geochim. Cosmochim. Acta* 75, 7257–7276. <https://doi.org/10.1016/j.gca.2011.08.007>.
- Schulze, F., Lewy, Z., Kuss, J., Gharabeh, A., 2003. Cenomanian–Turonian carbonate platform deposits in west central Jordan. *Int. J. Earth Sci.* 92, 641–660. <https://doi.org/10.1007/s00531-003-0339-6>.
- Shannon, L., Nelson, G., 1996. *The Benguela: Large Scale Features and Processes and System Variability, the South Atlantic: Present and Past Circulation*. Springer, pp. 163–210. https://doi.org/10.1007/978-3-642-80353-6_9.
- Sharland, P.R., Casey, D.M., Davies, R.B., Simmons, M.D., Sutcliffe, O.E., 2004. Arabian plate sequence stratigraphy—revisions to SP2. *GeoArabia* 9, 199–214. <https://doi.org/10.2113/geoarabia0901199>.
- Shmeit, M., Giraud, F., Jaillard, E., Reboulet, S., Masrou, M., Spangenberg, J.E., El-Samrani, A., 2022. The Valanginian Weissert Event on the south Tethyan margin: a dynamic paleoceanographic evolution based on the study of calcareous nannofossils. *Mar. Micropaleontol.* 175, 102134. <https://doi.org/10.1016/j.marmicro.2022.102134>.
- Scotese, C.R., Vérard, C., Burgener, L., Elling, R.P., Kocsis, A.T., 2025. The Cretaceous world: plate tectonics, palaeogeography and palaeoclimate. *Geol. Soc. London, Spec. Publ.* 544 (1). <https://doi.org/10.1144/SP544-000>, SP544-2024-28.
- Sokol, E.V., Kozmenko, O.A., Khoury, H.N., Kokh, S.N., Novikova, S.A., Nefedov, A.A., Sokol, I.A., Zaikin, P., 2017. Calcareous sediments of the Muwaqqar Chalk Marl Formation, Jordan: mineralogical and geochemical evidences for Zn and Cd enrichment. *Gondwana Res.* 46, 204–226.
- Soudry, D., Glenn, C., Nathan, Y., Segal, I., Vonderhaar, D., 2006. Evolution of Tethyan phosphogenesis along the northern edges of the Arabian–African shield during the Cretaceous–Eocene as deduced from temporal variations of Ca and Nd isotopes and rates of P accumulation. *Earth Sci. Rev.* 78, 27–57.
- Soudry, D., Nathan, Y., Ehrlich, S., 2013. Geochemical diagenetic trends during phosphorite formation – economic implications: the case of the Negev Campanian phosphorites, Southern Israel. *Sedimentology* 60, 800–819. <https://doi.org/10.1111/j.1365-3091.2012.01361.x>.
- Stein, R., 1986. Organic carbon and sedimentation rate—further evidence for anoxic deep-water conditions in the Cenomanian/Turonian Atlantic Ocean. *Mar. Geol.* 72, 199–209. [https://doi.org/10.1016/0025-3227\(86\)90119-2](https://doi.org/10.1016/0025-3227(86)90119-2).
- Sweere, T., van den Boorn, S., Dickson, A.J., Reichart, G.-J., 2016. Definition of new trace-metal proxies for the controls on organic matter enrichment in marine sediments based on Mn, Co, Mo and Cd concentrations. *Chem. Geol.* 441, 235–245. <https://doi.org/10.1016/j.chemgeo.2016.08.028>.
- Sweere, T., Archer, C., Fleischmann, S., Scholz, F., Kraal, P., Vance, D., 2025. Ni isotope cycling in sediments of highly productive upwelling systems. *Earth Planet. Sci. Lett.* 668, 119549. <https://doi.org/10.1016/j.epsl.2025.119549>.
- Tantawy, A., Keller, G., Adatte, T., Stinnesbeck, W., Kassab, A., Schulte, P., 2001. Maastrichtian to Paleocene depositional environment of the Dakhla Formation, Western Desert, Egypt: sedimentology, mineralogy, and integrated micro- and macrofossil biostratigraphies. *Cretac. Res.* 22, 795–827. <https://doi.org/10.1006/cres.2001.0291>.
- Taylor, S.R., McLennan, S.M., 1985. *The Continental Crust: Its Composition and Evolution*.
- Thibault, N., 2016. Calcareous nannofossil biostratigraphy and turnover dynamics in the late Campanian–Maastrichtian of the tropical South Atlantic. *Rev. Micropaleontol.* 59, 57–69. <https://doi.org/10.1016/j.revmic.2016.01.001>.
- Thibault, N., Gardin, S., 2007. The late Maastrichtian nannofossil record of climate change in the South Atlantic DSDP Hole 525A. *Mar. Micropaleontol.* 65, 163–184. <https://doi.org/10.1016/j.marmicro.2007.07.004>.
- Tissot, B., Demaison, G., Masson, P., Delteil, J.R., Combaz, A., 1984. Paleoenvironment and Petroleum Potential of Middle Cretaceous Black Shales in Atlantic Basins. In: Demaison, G., Murriss, R.J. (Eds.), *Petroleum Geochemistry and Basin Evaluation*. American Association of Petroleum Geologists. <https://doi.org/10.1306/M35439C13>.
- Tribouillard, N., Algeo, T.J., Lyons, T., Riboulleau, A., 2006. Trace metals as paleoredox and paleoproductivity proxies: an update. *Chem. Geol.* 232, 12–32. <https://doi.org/10.1016/j.chemgeo.2006.02.012>.
- Tissot, B.P., Welte, D.H., 1984. *Petroleum formation and occurrence*. Springer Science & Business Media.
- Tribouillard, N., Algeo, T., Baudin, F., Riboulleau, A., 2012. Analysis of marine environmental conditions based on molybdenum–uranium covariation—applications to Mesozoic paleoceanography. *Chem. Geol.* 324, 46–58. <https://doi.org/10.1016/j.chemgeo.2011.09.009>.
- Tyson, R., 2005. The “Productivity Versus Preservation” Controversy: Cause, Flaws, and Resolution. <https://doi.org/10.2110/pec.05.82.0017>.
- Usman, M., Grohmann, S., Abu-Mahfouz, I.S., Vahrenkamp, V., Littke, R., 2024. Effects of geochemical compositional heterogeneities on hydrocarbon expulsion and thermal maturation: an analog study of Maastrichtian source rocks from Jordan. *Int. J. Coal Geol.* 294, 104587. <https://doi.org/10.1016/j.coal.2024.104587>.
- Usman, M., Ardila-Sanchez, M., Idiz, E., Abu-Mahfouz, I.S., van Buchem, F., Vahrenkamp, V., 2025. Organic and inorganic geochemical cyclicity of a Maastrichtian oceanic open-shelf carbonate source rock. *Sci. Rep.* 15, 15993. <https://doi.org/10.1038/s41598-025-99832-w>.
- van Buchem, F.S., Razin, P., Homewood, P.W., Oterdoorn, W.H., Philip, J., 2002. Stratigraphic organization of carbonate ramps and organic-rich intrashelf basins: Natih Formation (middle cretaceous) of northern Oman. *AAPG Bull.* 86 (1), 21–53. <https://doi.org/10.1306/61EEDA30-173E-11D7-8645000102C1865D>.
- van der Meer, D.G., Scotese, C.R., Mills, B.J., Sluijs, A., van de Weg, R.M., 2022. Long-term Phanerozoic global mean sea level: insights from strontium isotope variations and estimates of continental glaciation. *Gondwana Res.* 111, 103–121. <https://doi.org/10.1016/j.gr.2022.07.014>.
- Vaughan, A.P., Searrow, J.H., 2003. Ophiolite obduction pulses as a proxy indicator of superplume events? *Earth Planet. Sci. Lett.* 213 (3–4), 407–416.
- Villamil, T., Arango, C., Hay, W.W., 1999. Plate Tectonic Paleogeographic Hypothesis for Cretaceous Source Rocks and Cherts of Northern South America. <https://doi.org/10.1130/0-8137-2332-9.191>.
- Voigt, S., Gale, A.S., Jung, C., Jenkyns, H.C., 2012. Global correlation of Upper Campanian–Maastrichtian successions using carbon-isotope stratigraphy: development of a new Maastrichtian timescale. *Newsl. Stratigr.* 45, 25. <https://doi.org/10.1127/0078-0421/2012/0016>.

- Wendler, J.E., Meyers, S.R., Wendler, I., Kuss, J., 2014. A million-year-scale astronomical control on late cretaceous sea-level. *News. Stratigr.* 47, 1–19. <https://doi.org/10.1127/0078-0421/2014/0038>.
- Wilmsen, M., 2003. Sequence stratigraphy and palaeoceanography of the Cenomanian Stage in northern Germany. *Cretac. Res.* 24, 525–568. [https://doi.org/10.1016/S0195-6671\(03\)00069-7](https://doi.org/10.1016/S0195-6671(03)00069-7).
- Wilson, P.A., Norris, R.D., 2001. Warm tropical ocean surface and global anoxia during the mid-cretaceous period. *Nature* 412, 425–429. <https://doi.org/10.1038/35086553>.
- Xu, G., Hannah, J.L., Bingen, B., Georgiev, S., Stein, H.J., 2012. Digestion methods for trace element measurements in shales: Paleoredox proxies examined. *Chem. Geol.* 324, 132–147. <https://doi.org/10.1016/j.chemgeo.2012.01.029>.
- Ziegler, M., 2001. Late Permian to Holocene paleofacies evolution of the Arabian Plate and its hydrocarbon occurrences. *GeoArabia* 6, 445–504. <https://doi.org/10.2113/geoarabia0603445>.