Environmental changes during the onset of the Late Pliensbachian Event (Early Jurassic) in the Cardigan Bay Basin, Wales.

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Abstract. The Late Pliensbachian Event (LPE), in the Early Jurassic, is associated with a perturbation in the global carbon cycle (positive carbon isotope excursion (CIE) of ~ 2 ‰), cooling of ~5°C, and the deposition of widespread regressive facies. Cooling during the Late Pliensbachian has been linked to enhanced organic matter burial and/or disruption of thermohaline ocean circulation due to sea level lowstand of at least regional extent . Orbital forcing had a strong influence on the Pliensbachian environments and recent studies show that the terrestrial realm and the marine realm in and around the Cardigan Bay Basin, UK, were strongly influenced by orbital climate forcing. In the present study we build on the previously published data for long eccentricity cycle E459 ± 1 and extend the palaeoenvironmental record to include E458 ± 1. We explore the environmental and depositional changes on orbital time scales for the Llanbedr (Mochras Farm) core during the onset of the LPE. Clay mineralogy, XRF elemental analysis, isotope ratio mass spectrometry, and palynology are combined to resolve systematic changes in erosion, weathering, fire, grain size and riverine influx. Our results indicate distinctively different environments before and after the onset of the LPE positive CIE, and show increased physical erosion relative to chemical weathering. We also identify five swings in the climate, in tandem with the 405 kyr eccentricity minima and maxima. Eccentricity maxima are linked to precessionally repeated occurrences of a semi-arid monsoonal climate with high fire activity and relatively coarser sediment from terrestrial runoff. In contrast, 405 kyr minima in the Mochras core are linked to a more persistent, annually wet climate, low fire activity, and relatively finer grained deposits across multiple precession cycles. The onset of the LPE +ve CIE did not impact the expression of the 405 kyr cycle in the proxy records; however, during the second pulse of heavier carbon (^{13}C) enrichment, the clay minerals record a change from dominant chemical weathering to dominant physical erosion.

1 Introduction

The Early Jurassic is a period marked by large climatic fluctuations and associated carbon-isotope excursions (CIE's) in an overall warmer than present and high pCO_2 world (McElwain *et al*., 2005; Korte and Hesselbo, 2011; Steinthorsdottir and Vajda, 2015; Korte *et al.*, 2015; Robinson *et al.*, 2016). A series of small and medium sized CIE's have recently been documented for the Sinemurian and Pliensbachian, which have mainly been from European, North African and North American records (Korte and Hesselbo, 2011; Franceschi *et al*., 2014; Korte *et al*., 2015; Price *et al*., 2016; De Lena *et al*., 2019; Hesselbo *et al*., 2020a; Mercuzot *et al*., 2020; Storm *et al*., 2020; Silva *et al*., 2021; Cifer *et al*., 2022; Bodin *et al*., 2023). Notable is the pronounced positive CIE in the Late Pliensbachian, which has been called the Late Pliensbachian Event (LPE) and is linked to climatic cooling (Hesselbo and Korte, 2011; Korte *et al*., 2015) and a supra-regional/global sea level low stand (Hallam, 1981; de Graciansky *et al*., 1998; Hesselbo and Jenkyns, 1998; Hesselbo, 2008). The LPE has been recognized by a positive shift in benthic marine oxygen-isotopes (~1.5–2 per mil) (Bailey *et al*., 2003; Rosales *et al*., 2004,2006; Suan *et al*., 2010; Dera *et al*., 2011a; Korte and Hesselbo, 2011; Gómez *et al*., 2016; Alberti *et al*., 2019, 2021), coeval with a positive shift in marine and terrestrial carbon isotopes (~2 per mil) (Jenkyns and Clayton, 1986; McArthur *et al*., 2000; Morettini *et al*., 2002; Quesada *et al*., 2005; Rosales *et al*., 2006; Suan *et al*., 2010; Korte and Hesselbo, 2011; Silva *et al*., 2011; Gómez *et al*., 2016; De Lena *et al*., 2019).

A cooler Late Pliensbachian climate has been suggested based on low pCO_2 values inferred by leaf stomatal index data from eastern Australia (Steinthorsdottir and Vajda, 2015), the presence of glendonites in northern Siberia (Kaplan, 1978; Price, 1999; Rogov and Zakharov, 2010), vegetation shifts from a diverse flora of different plant groups to one mainly dominated by bryophytes in Siberia (Ilyina, 1985; Zakharov *et al*., 2006), and possible ice rafted debris found in Siberia (Price, 1999; Suan *et al*., 2011). Whilst the presence of ice sheets is strongly debated, a general cooling period (~5°C lower; Korte *et al*., 2015; Gómez *et al*., 2016) is evident from several temperature reconstructions from NW Europe. A cooling is hypothesized via enhanced carbon burial in the marine sediments, leading to lower pCO_2 values and initiating cooler climatic conditions (Jenkyns and Clayton, 1986; Suan *et al*., 2010; Silva *et al*., 2011; Storm *et al*., 2020). Direct evidence of large-scale carbon burial in Upper Pliensbachian marine deposits has not yet been documented (Silva *et al*., 2021). Alternatively, cooling has been suggested to be caused by a lower sea level which would have disrupted ocean circulation in the Laurasian Seaway, reducing poleward heat transport from the tropics (Korte *et al*., 2015). In the UK region, a dome structure in the North Sea has been linked to shedding of sediments during sea level low stands from the Late Toarcian and possibly before (Underhill and Partington, 1993; Korte *et al*., 2015; Archer et al. 2019). Disruption of the ocean circulation between the western Tethys and the Boreal realm is supported by marine migration patterns (Schweigert, 2005; Zakharov *et al*., 2006; Bourillot *et al*., 2008; Nikitenko, 2008; Dera *et al*., 2011b; van de Schootbrugge *et al*., 2019) and numerical models (Bjerrum *et al*., 2001; Dera and Donnadieu, 2012; Ruvalcaba Baroni *et al*. 2018); however, the net direction of the flows remain debated.

An additional factor to be considered is that a strong orbital control exists on the Pliensbachian sedimentary successions (Weedon and Jenkyns, 1990; Ruhl *et al*., 2016; Hinnov *et al*., 2018; Storm *et al*., 2020; Hollaar *et al*., 2021). Previous studies have indicated that sea level changes, possibly coupled to glacio-eustatic rise and fall, occurred during the LPE on a 100 kyr (short eccentricity) time scale (Korte and Hesselbo, 2011). A high-resolution record of charcoal, clay mineralogy, bulk-organic carbon-isotopes, TOC and $CaCO_3$ encompassing approximately one 405 kyr cycle from the Llanbedr (Mochras Farm) borehole, Cardigan Bay Basin, NW Wales, UK, suggested that the long-eccentricity orbital cycle had a significant effect on background climatic and environmental change during the late Pliensbachian, particularly affecting the hydrological regime of the region (Hollaar *et al*., 2021). This previous research focussed on orbital forcing of environmental change for a time lacking any large excursion in $\delta^{13}C_{org}$, and so unaffected by perturbations to the global carbon cycle. Here, we expand on the record of Hollaar *et al*. (2021) to cover two long eccentricity cycles (which we identify as spanning from cycle E459 ±1 to the start of E457 ±1 of Laskar *et al*. 2011 and Laskar 2020), where the final parts of E458 and the start of E457 are interrupted by onset of the Late Pliensbachian Event (Fig. 1). This longer record allows us to more robustly examine the influence of the long eccentricity cycle and the potential impact of a global carbon cycle perturbation on the palaeoclimate and depositional environment. We find that the long eccentricity forcing continued to dictate the precise timing of major environmental changes in the Cardigan Bay Basin, including the initial step of the positive carbon isotope excursion.

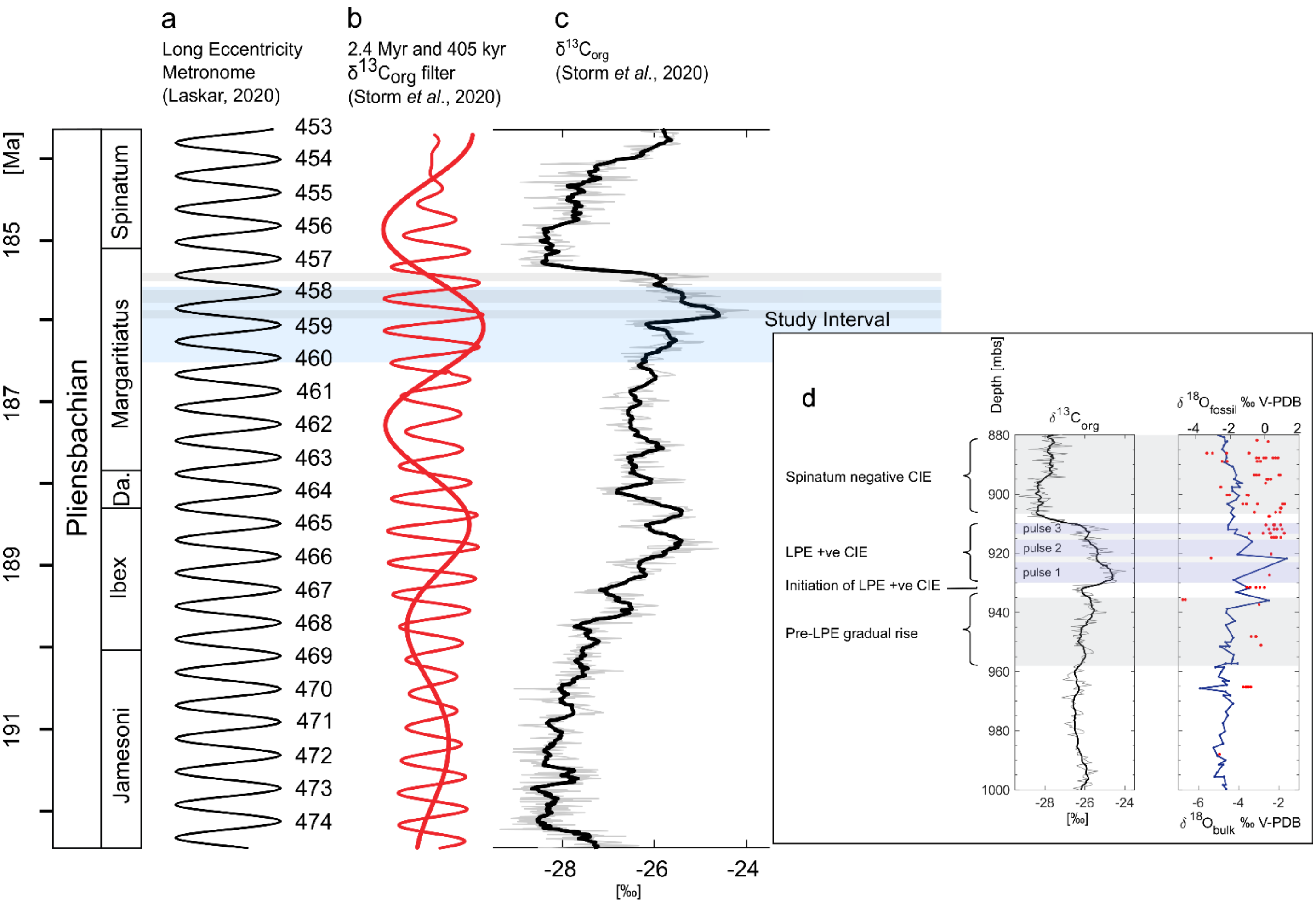

Figure 1: Stratigraphic framework of the Mochras borehole. (a) 405 kyr metronome (Laskar, 2020) which shows that this study spans E459 ±1 to E457 ±1. (b) 2.4 Myr and 405 kyr filter derived from the $\delta^{13}C_{org}$ record from Storm *et al*. (2020). A slight offset in pacing is observed in the 405 kyr metronome based on an assumed fixed 405 kyr cycle length (a), versus filtering of the 405 kyr signal from the orbital solution (b). (c) $\delta^{13}C_{org}$ curve from the Mochras borehole (Storm *et al*., 2020), showing the ~1.8 ‰ +CIE that marks the LPE. High resolution data are visualized in light grey and a 10-step moving average in black. The blue bar marks interval in the Mochras borehole considered in this study. The three grey shaded bars represent the three pulses in the +CIE of the LPE. (d) Close-up of the $\delta^{13}C_{org}$ (Storm *et al*., 2020) and $\delta^{18}O_{bulk}$ and $\delta^{18}O_{fossil}$ (Ullmann *et al*., 2022) from the Late Pliensbachian of the Mochras core. A pre-LPE gradual rise is recorded in the $\delta^{13}C_{org}$, followed by the initiation of the LPE +ve CIE, which consists of three pulses. After the LPE +ve CIE, $\delta^{13}C_{org}$ values drop

recorded starting at ~910 mbs, and the Spinatum negative CIE is recorded. The $\delta^{18}O_{bulk}$ of the Mochras core (blue) is diagenetically altered and unlikely to preserve a palaeoclimatic imprint (Ullmann *et al*., 2022). Also, shown are $\delta^{18}O_{fossil}$ values (red).

2 Material

2.1 Palaeo-location and setting

Associated with the break-up of Pangea, connections between oceans via epicontinental seaways were established during the Early Jurassic, such as the Hispanic Corridor, which connected the north-western Tethys and the eastern Panthalassa, and the Viking Corridor which linked the north-western Tethys Ocean to the Boreal Sea (Sellwood and Jenkyns, 1975; Smith *et al.*, 1983; Bjerrum *et al*., 2001; Damborenea *et al*., 2012). The linking passage of the NW Tethys Ocean and the Boreal Sea (south of the Viking Corridor) is the palaeogeographical location of the Llanbedr (Mochras Farm) borehole, Cardigan Bay Basin, NW Wales, UK (Fig. 2) – referred to hereafter as Mochras. Due to the location of the Mochras succession during the Late Pliensbachian, it was subject to both polar and equatorial influences allowing study of variations in the circulation in the N-S Laurasian Seaway (including the Viking Corridor) prior to and across the LPE. Mochras was located at a mid-palaeolatitude of ~35° N (cf. Torsvik and Cocks, 2017).

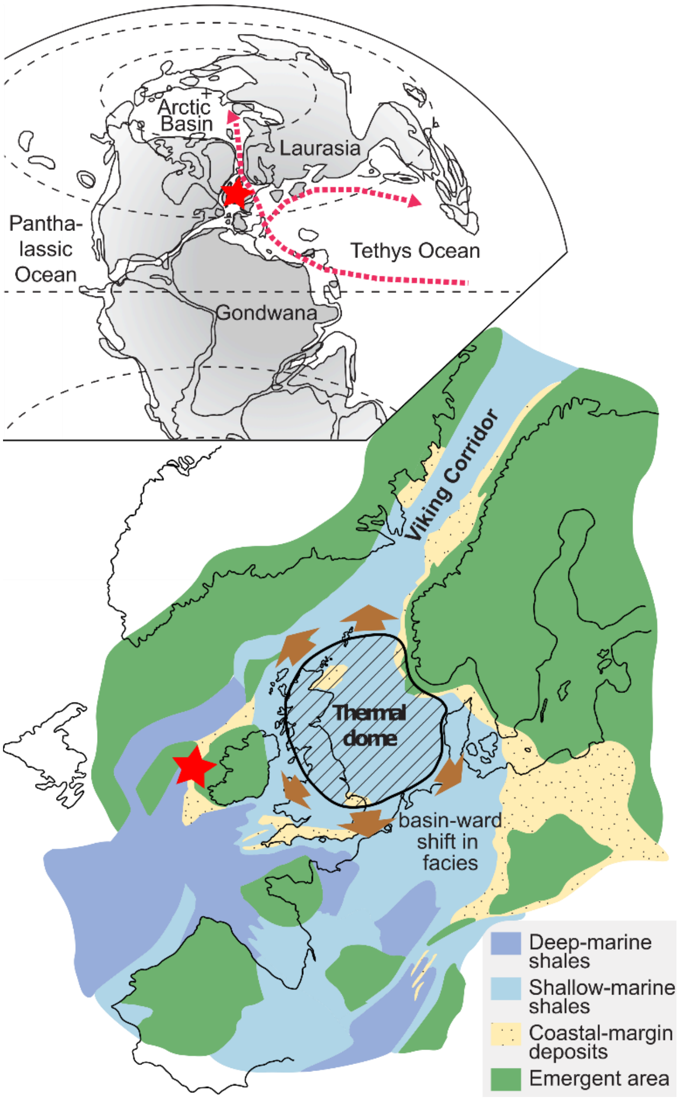

Figure 2: Palaeolocation of the Mochras borehole in the context of potential North Sea doming. Figure reprinted and adapted from Korte *et al*. (2015), which is open access (https://creativecommons.org/licenses/by/4.0/). The Mochras borehole was located at a paleolatitude of ~35° N in the Cardigan Bay Basin (Torsvik and Cocks, 2017). Circulation in the Tethys Ocean and between there and the Boreal region influenced the depositional environment of the Mochras core (Pieńkowski *et al*., 2021). Late Pliensbachian sea level fall potentially resulted in occlusion of the Viking Corridor as the topography of the North Sea dome structure disrupted circulation in the seaway (Korte *et al*., 2015).

The depositional environment of Mochras is likely characterized by a rift setting, which is reflected by the relatively open and deep marine facies and the evidence for below storm wave-base and contourite deposition (Pieńkowski *et al*., 2021), but always with a strong terrestrial influence (van de Schootbrugge *et al*., 2005; Riding *et al*., 2013) from the nearby landmasses (Dobson and Whittington, 1987). The Cardigan Bay Basin fill was downthrown against the Early Paleozoic Welsh Massif by a major normal fault system, probably comprising the Bala, Mochras and Tonfanau faults at the eastern and south-eastern margins of the basin in Late Paleozoic–Early Mesozoic time (Woodland, 1971; Tappin *et al*., 1994). The main source of detrital material is understood to be the Caledonian Welsh Massif, followed by the Irish and Scottish landmasses (Deconinck *et al*., 2019). Other massifs that could have influenced the provenance are the London-Brabant Massif to the south east, and Cornubia to the south (van de Schootbrugge *et al*., 2005), depending on the marine circulation and sediment transport at the time.

2.2 Core location and material

The Llanbedr (Mochras Farm) Borehole was drilled onshore in the Cardigan Bay Basin (52 48' 32" N, 4 08' 44" W) in 1967–1969, North Wales (Woodland, 1971; Hesselbo *et al*., 2013). The borehole recovered a 1300 m thick Early Jurassic sequence (601.83–1906.78 metres below surface (mbs)), yielding the most complete and extended Early Jurassic succession in the UK, being double the thickness of same age strata in other UK cores and outcrops (Hesselbo *et al*., 2013; Ruhl *et al*., 2016). The Lower Jurassic is biostratigraphically complete at the zonal level (Ivimey-Cook, 1971; Copestake and Johnson, 2014), with the top truncated and unconformably overlain by Cenozoic strata (Woodland, 1971; Dobson and Whittington, 1987; Tappin *et al*., 1994; Hesselbo *et al*., 2013). The lithology is dominated by argillaceous sediments, with alternating muddy limestone, marl and mudstone (Woodland, 1971; Sellwood and Jenkyns, 1975).

The Pliensbachian Stage in the Mochras borehole occurs between ~1250 and ~865 mbs, with the Margaritatus Zone between ~1013 and 909 mbs (Page in Copestake and Johnson, 2014). The Pliensbachian interval comprises alternations of mudstone (with a moderate total organic carbon [TOC]) and organic-poor limestones, with a pronounced cyclicity at ~1 ± 0.5 m wavelength (Ruhl *et al*., 2016). The Upper Pliensbachian contains intervals that are silty and locally sandy, whilst levels of relative organic enrichment also occur through the Pliensbachian (Ruhl *et al*., 2016). Overall, the Upper Pliensbachian is relatively rich in carbonate (Ruhl *et al*., 2016; Ullmann *et al*., 2022).

3 Methods

For this study, samples were taken at a ~30 cm resolution from slabbed core from 934–918 mbs for XRD and mass spectrometry, as well as palynofacies and microcharcoal analysis. XRF analyses were made at a 1 cm resolution from 934–918 mbs (complete dataset deposited as Damaschke *et al*., 2021). These new samples complement samples and data at 10 cm resolution from 951–934 mbs published in Hollaar *et al*. (2021).

3.1 TOC, $CaCO_3$ and bulk organic carbon isotope mass spectrometry

TOC and $\delta^{13}C_{org}$ were measured to track the changes in the total organic fraction and the bulk organic carbon isotope ratios in relation to the other palaeoenvironmental proxy data.

Powdered bulk-rock samples (~ 2 g) were decarbonated in 50 ml of 3.3% HCl. After this, the samples were transferred to a hot bath of 79 °C for 1 h to remove siderite and dolomite. Subsequently, the samples were centrifuged and the liquid decanted. The samples were rinsed repeatedly with distilled water to reach neutral *p*H. After this, the samples were oven-dried at 40 °C, re-powdered and weighed into tin capsules for mass spectrometry using the Sercon Integra 2 stable isotope analyser at the University of Exeter Environment andand Sustainability Institute (ESI), stable isotope facility on the Penryn Campus, Cornwall. Samples were run alongside in-house reference material (bovine liver; $\delta^{13}C$ -28.61 and Alanine; $\delta^{13}C$ -19.62) which was used to correct for instrument drift and to determine the $\delta^{13}C$ values of the samples. The $\delta^{13}C_{org}$ values are reported relative to V-PDB following a within-run laboratory standard calibration. Total organic carbon was determined using the CO_2 beam area relative to the bovine liver standard (%C 47.24). Replicate analysis of the in-house standards gave a precision of ± <0.1 ‰ (2 SD).

The carbonate content was measured by the dry weight sample loss before and after decarbonation. The %C content derived from the mass spectrometer was corrected for carbonate loss to derive TOC.

3.2 X-Ray Diffraction (XRD) to determine clay mineralogy

Clay mineral analysis was performed to gain insight into the relative importance of physical erosion versus chemical weathering and related changes in the hydrological cycle.

About 2–3 g of gently powdered bulk-rock was decarbonated with a 0.2 M HCl solution. The clay sized fraction (< 2 µm) was extracted with a syringe after decantation of the suspension after 95 minutes following Stokes' law. The extracted fraction was centrifuged and oriented on glass slides for X-ray diffraction analysis (XRD) using a Bruker D4 Endeavour diffractometer (Bruker, Billerica, MA, USA) with Cu Kα radiations, LynxEye detector and Ni filter under 40 kV voltage and 25 mA intensity (Biogéosciences Laboratory, Université Bourgogne/Franche-Comté, Dijon). Following Moore and Reynolds (1997), the clay phases were discriminated in three runs per sample: (1) air-drying at room temperature; (2) ethylene-glycol solvation for 24 h under vacuum; (3) heating at 490 °C for 2 h.

Identification of the clay minerals was based on their main diffraction peaks and by comparison of the three diffractograms obtained. The proportion of each clay mineral on glycolated diffractograms was measured using the MACDIFF 4.2.5. software (Petschick, 2000). Identification of the clay minerals follows the methods in Deconinck *et al*. (2019) and Moore and Reynolds (1997).

3.3 Palynofacies and microcharcoal

Palynofacies were examined to explore shifts in the terrestrial versus marine origins of the particulate organic matter. Each ~ 20 g bulk rock sample was split into 0.5 cm^3 fragments, minimizing breakage of charcoal and other particles, to optimize the surface area for extraction of organic matter using a palynological acid maceration technique. The samples were first treated with cold hydrochloric acid (10% and 37% HCl) to remove carbonates. Following this, hydrofluoric acid (40% HF) was added to the samples to remove silicates. Carbonate precipitation was prevented, by adding cold concentrated HCl (37%) after 48 h. The samples were neutralized via multiple dilution-settling-decanting cycles using DI water, after which 5 droplets of the mixed residue were taken for the analysis of palynofacies prior to sieving. The remaining residue was sieved through a 125 µm and 10 µm mesh to extract the microcharcoal fraction.

A known quantity (111 µl) out of a known volume of liquid containing the 10–125 µm sieved residue was mounted onto a palynological slide using glycerine jelly. This fraction, containing the microscopic charcoal, was analysed and the charcoal particles counted using an Olympus (BX53) transmitted light microscope (40 x10 magnification). For each palynological slide four transects (two transects in the middle and one on the left and right side of the coverslip) were followed and the number of charcoal particles determined. Charcoal particles were identified with the following criteria: opaque and black, often elongated lath-like shape with sharp edges, original anatomy preserved, brittle appearance with a lustrous shine (Scott, 2010). These data were then scaled up to the known quantity of the sample according the method of Belcher *et al*. (2005).

Palynofacies were grouped broadly according to Oboh-Ikuenobe *et al*. (2005): sporomorphs, fungal remains, freshwater algae, marine palynomorphs, structured phytoclasts, unstructured phytoclasts, black debris, amorphous organic matter (AOM), and charcoal (further described in Hollaar *et al*., 2021). The palynofacies were quantified on a palynological slide using the optical light microscope (40 x10 magnification) and counting a minimum of 300 particles per slide. Because the samples are AOM-dominated, counting was continued until a minimum of 100 non-AOM particles were observed. We used the percentage of terrestrial phytoclasts, which includes sporomorphs, and structured and unstructured phytoclasts, to examine changes in terrestrial organic particle content.

3.4 X-Ray Fluorescence (XRF) to determine detrital elements

Detrital elemental ratios were examined to analyse changes in relative terrestrial influx and the type of material transported from the land to the marine realm. The slabbed archive halves of the Mochras borehole were scanned via automated X-ray fluorescence (XRF) at a 1 cm resolution for the interval 951 –918 mbs, with the ITRAX MC at the British Geological Survey Core Scanning Facility (CSF), Keyworth, UK (Damaschke *et al*., 2021). The measurement window was 10 s and long-term drift in the measurement values was counteracted by regular internal calibration with a glass reference (NIST-610). Duplicate measurements were taken every 5 m for a 50 cm interval to additionally verify the measured results.

3.5 Statistical analysis

Principal component analysis (PCA) was performed to examine a potential change in the proxy data before and after the +vie CIE. This was executed in the software PAST on the normalized dataset including microcharcoal, TOC, $CaCO_3$, $\delta^{13}C_{org}$, S/I, K/I, primary clay mineralogy, Si/Al, Zr/Rb. The samples before the +ve CIE (951.0–930.4 mbs) and the samples after the +ve CIE (930.3–918.0 mbs) are grouped to examine a potential difference in the sedimentary composition before and after the +ve CIE. A Pearson's correlation was executed in Matlab R2017b. The *p* value tests the hypothesis of no correlation against the alternative hypothesis of a positive or negative correlation (significance level at $p \leq 0.05$).

4 Results

4.1 TOC, $CaCO_3$ and bulk organic carbon isotope ratio mass spectrometry

Alternating TOC-enhanced and Ca-rich lithological couplets occur on a metre scale through the studied interval with TOC and $CaCO_3$ having a strong negative correlation (r = -0.64, $p = 0.001$). TOC content fluctuates in the range 0.17–1.72 wt% (mean 0.8 wt%) and the highest fluctuations of TOC content are found from 939–930 mbs. The $CaCO_3$ content fluctuates in opposition to TOC and varies between 14 and 89 %. The studied interval is generally high in $CaCO_3$ (mean 58 %) (Fig. 3). The $\delta^{13}C_{org}$ displays a minor (~0.5 ‰) shift towards more positive values at ~944 mbs (as reported in Storm *et al*., 2020; Hollaar *et al*., 2021). At ~ 930 mbs an abrupt shift of ~1.8 ‰ (Figs. 1, 4; Storm *et al*., 2020) indicates the onset of the Late Pliensbachian Event (LPE) in the Mochras core. In agreement with this, the results of the present study show a shift from ~ minus 27 per mil to ~ minus 25.15 per mil between 930.8 and 930.4 mbs (Fig. 4). The $\delta^{13}C_{org}$ data presented here have been divided into three phases: the pre-LPE gradual rise, followed by the +ve CIE, which is subdivided into pulses 1, 2 and 3

(Fig. 1). After the onset of the positive $\delta^{13}C_{org}$ excursion, the TOC content drops to the lowest values (from 0.85 % before and 0.6 % after the +ve CIE on average), but the 1 metre fluctuations continue (Figs. 3, 4). No overall change in the $CaCO_3$ content is observed through the positive carbon-isotope excursion (Fig. 3).

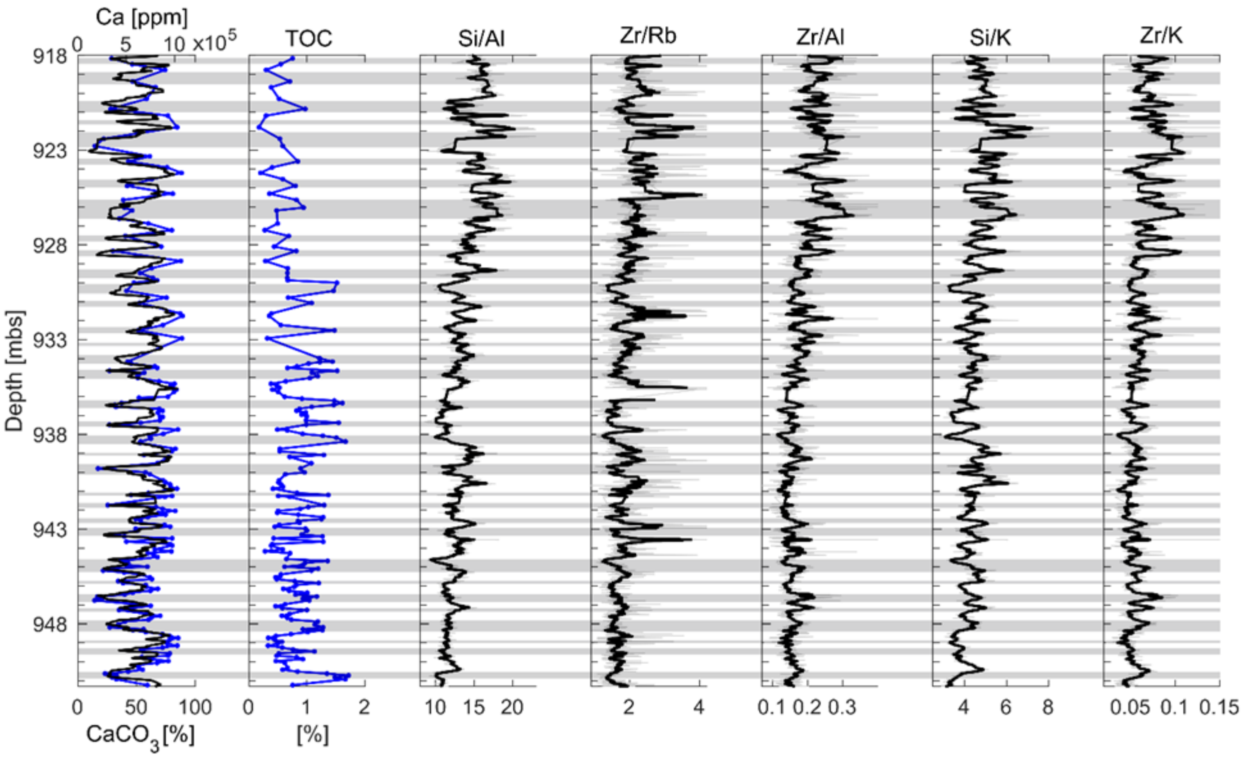

Figure 3: Detrital ratios over the Ca-rich and TOC-enhanced lithological couplets for the studied interval. Overview of Ca (black, derived from Ruhl *et al*. 2016), $CaCO_3$ (blue), and TOC content of the studied interval 951–918 mbs. The grey shading represents the TOC-enhanced beds and the unshaded bands mark the Ca-rich (limestone) beds. The detrital ratios reflect the silt to fine sand fraction (Si, Zr) versus the clay fraction (Rb, Al, K). Two increasing upward cycles are observed in the Si/Al and Zr/Rb ratios. The pattern observed in all detrital ratios (except the Ti/Al) is similar and likley reflects overall upwards coarsening.

4.2 Clay minerals

XRD analysis shows that the main clay types found in this interval are illite, random illite-smectite mixed-layers (I-S R0) [hereafter referred to as smectite], and kaolinite. Illite and kaolinite co-fluctuate in the interval studied here, and are directly out of phase with smectite abundance. Chlorite and R1 I-S are present in minor proportions, but reach sporadically higher relative abundance (> 10 %) from ~ 932 mbs upwards, with sustained >10% abundance at ~925–918 mbs (Fig. 4 and SI Fig. 1). The relative abundances of smectite and illite and of kaolinite and illite are expressed by the ratio S/I and K/I respectively. These ratios were calculated according to the intensity of the main diffraction peak of each mineral.

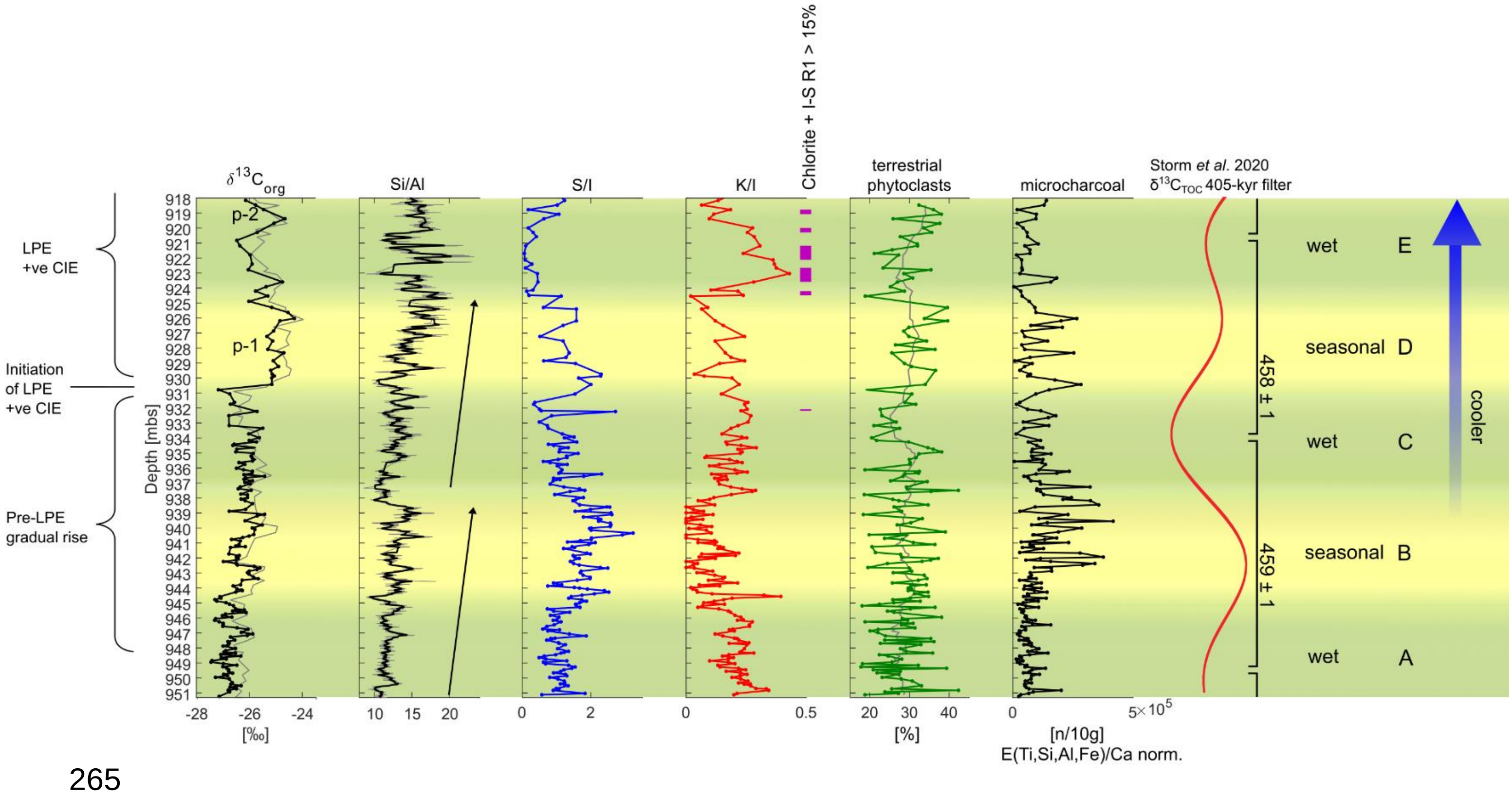

Figure 4: **Synthesis diagram showing the climatic swings observed in tandem with the long eccentricity cycle.** The studied interval (Upper Pliensbachian Margaritatus Zone) comprises part of the pre-LPE gradual rise, the initiation of the LPE +ve CIE and pulse 1 and 2 ($\delta^{13}C_{org}$ data in black from this study and in light grey from Storm *et al*. (2020)). Five climatic phases (A–E) are interpreted from the Si/Al, smectite/illite, kaolinite/illite, chlorite and I-S R1 abundance and the microcharcoal abundance. In tandem with the 405 kyr cycle (Storm *et al*., 2020) climatic state of a year-round wet climate, low fire activity and fine-grained sediments across multiple precession cycles (phase A and C) alternates with a climatic state that includes repeated precessionally driven states that are semi-arid, with high fire activity and coarser sediments (phase B and D). The top of the record (phase E) indicates increased physical erosion (chlorite + I-S R1, kaolinite) relative to chemical weathering. In terrestrial phytoclast column grey line = 10-step moving average.

4.3 Organic matter

The type of particulate organic matter, and more specifically the abundance in the marine versus terrestrial origin of the particles, fluctuates on a metre scale from 18–42 % (Fig. 4, SI Fig. 2). Palynofacies indicate that the type of organic matter does not change in relation to the metre-scale lithological facies cycles (no correlation between percentage terrestrial phytoclasts and TOC or $CaCO_3$). No large and abrupt changes are recorded in the terrestrial/marine proportions, but the proportion of terrestrial phytoclasts has 4 high phases: between 944.6 and

942.0 mbs, 937.5 and 934.9 mbs, 930.4 and 925.4 mbs, and 920.3 and 918.0 mbs (SI Fig. 2). The first and second high phase falls within the + 0.5 ‰ positive swing in the $\delta^{13}C_{org}$ whilst the latter two high phases correspond to pulse 1 and pulse 2 in the +ve CIE. Amorphous organic matter (AOM) is very abundant, followed by unstructured phytoclasts, with lower proportions of structured phytoclasts and charcoal (SI Fig. 3). Charcoal particles make up a relatively large proportion of the terrestrial particulate organic matter (~10 % on average) and ~3.5 % on average of the total particulate organic matter fraction (SI Fig. 3). Only sparse marine and terrestrial palynomorphs were observed (SI Fig. 3).

To assess the character of the observed fluctuations in microcharcoal abundance, whether changes in microcharcoal can be related to enhanced runoff from the land and/or organic preservation, or if the microcharcoal signifies changes in fire activity on land, the charcoal record was corrected for detrital influx. We adjust the charcoal particle abundances using the XRF elemental record, normalizing to the total terrigenous influx following Daniau *et al*. (2013) and Hollaar *et al*. (2021). The stratigraphic trends in the normalized microcharcoal for E_{ter}/Ca, Si/Al, Ti/Al and Fe/Al remain the same (SI Fig. 4). The absolute number of microcharcoal particles decreases, with raw mean charcoal particles 1.06 x 10^5 per 10 g and E_{ter}/Ca normalized mean 9.7 x 10^4 n/10g, Ti/Al normalized 6.4 x 10^4 n/10g, Si/Al normalized 7.7 x 10^4 n/10g, Fe/Al normalized 9.8x10^4 mean number of microcharcoal particles per 10 g (SI Fig. 4). The number of microcharcoal particles per 10 g processed rock decreases when correcting for terrestrial run-off changes, hinting that perhaps part of the 'background' microcharcoal is related to terrestrial influx; the normalisation also shows that the observed patterns in microcharcoal abundances are not influenced by changes in terrestrial runoff and taphonomy. Hence, the highs and lows in the microcharcoal record can be interpreted to represent changes in the fire regime on land. The microcharcoal abundance fluctuates strongly in the record presented here; however, no clear difference in microcharcoal content has been observed before and after the onset of the +ve CIE.

4.4 Detrital elemental ratios (XRF)

Strong similarities are observed between the fluctuating ratios of Si/Al, Si/K, Zr/Rb, Zr/Al and Zr/K (Fig. 3). The elements Al, Rb and K sit principally in the clay fraction (e.g. Calvert and Pederson, 2007), whereas Si and Zr are often found in greater abundance in the coarser fraction related to silt and sand grade quartz and heavy minerals (Calvert and Pederson, 2007). The ratios all show clear metre-scale fluctuations, and these are superimposed on two increasing-upward trends observed in both the Si/Al and the Zr/Rb, followed by a drop and rise to peak values in the latest part of phase D and phase E above the onset of the +ve CIE (Figs. 3, 4). A parallel trend is observed between the clay ratios (XRD) and elemental ratios Si/Al and Zr/Rb (Fig. 3). Phases of high S/I correspond to the peaks in the two coarsening upward sequences, whereas phases of high K/I correspond to the low phases in the two coarsening upward sequences. After the +ve CIE onset (in phase E) this relationship turns around, and an enrichment in the kaolinite/illite ratio corresponds to the elemental ratios, where highest kaolinite relative abundance is observed in parallel with elemental ratios suggesting maximum coarse fraction.

4.5 PCA analysis

The proxy datasets ($\delta^{13}C_{org}$, TOC, percentage terrestrial phytoclasts, microcharcoal, smectite/illite, kaolinite/illite, abundance of chlorite and R1 I-S, Si/Al, Zr/Rb, Zr/Al) were normalized between 0–1 and run for PCA analysis in PAST. Sixty-four percent of the variance is explained by the first three axes (PCA-1 27.7 %, PCA-2 19.7 %, PCA-3 15.3 %) inside the 95 % confidence interval.

PC-1 mainly explains the anti-correlation of TOC and $CaCO_3$. PC-2 shows the anti-correlation of K/I and S/I. Positive loadings were observed for S/I, microcharcoal, macrocharcoal and $CaCO_3$. For PC-2, negative loadings were observed for K/I, abundance of chlorite + I-S R1. PC-3 shows strong positive loadings (> 0.3) for $\delta^{13}C_{org}$, Si/Al and Zr/Al.

Plotting PC-1 (y-axis) over PC-3 (x-axis) shows that the samples after the onset of the +ve CIE are grouped to the top of the y-axis (more associated with S/I compared to K/I) and to the right of the x-axis (more associated with primary minerals, phytoclasts, and higher Si/Al, Zr/Rb and Zr/Al) (Fig. 5).

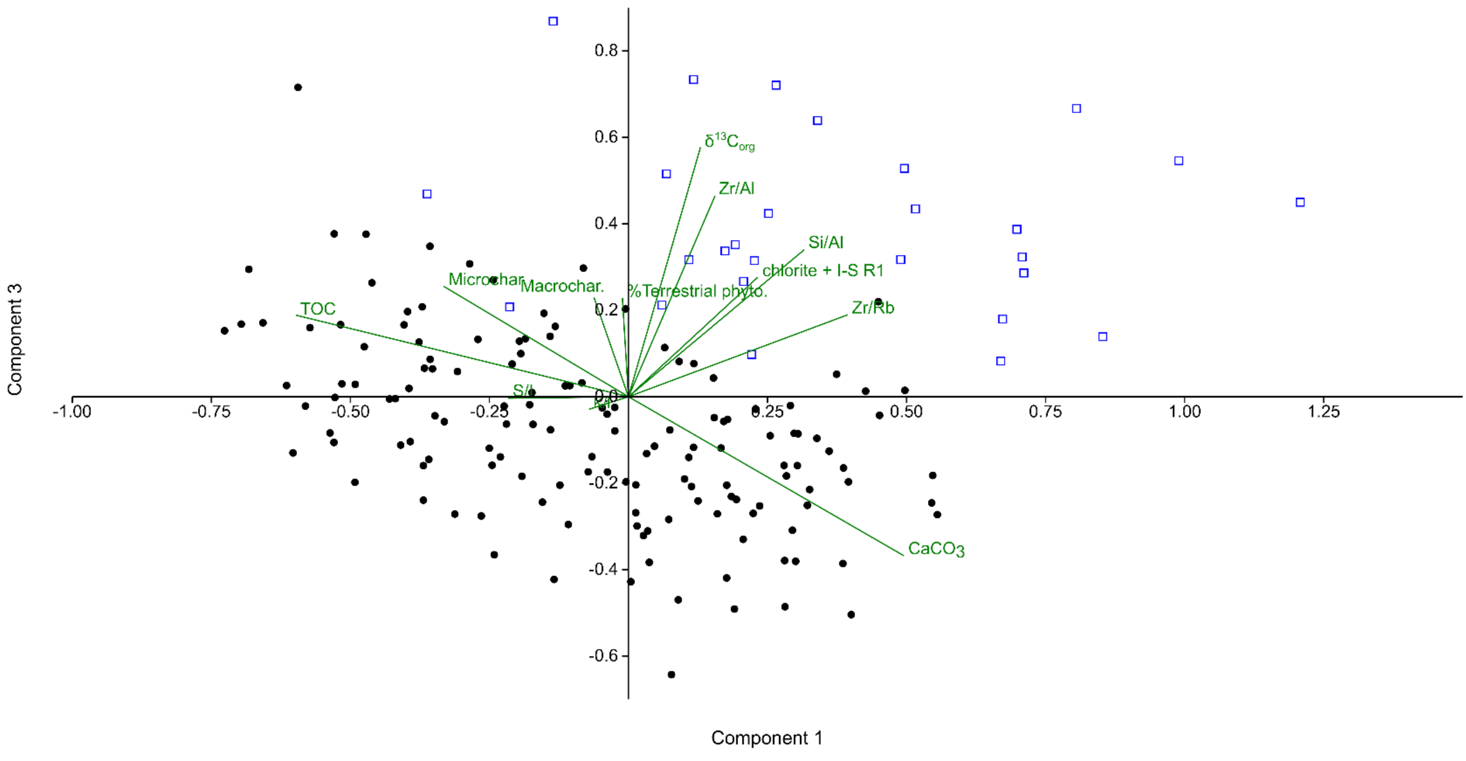

Figure 5: PCA-analysis shows a distinctly different depositional signature before and after the onset of the LPE +ve CIE in the Mochras core. PCA plot of PC-1 and PC-3: all samples before the onset of the LPE +ve CIE are marked in black closed circles and the samples after the onset of the LPE +ve CIE are marked in blue open squares.

5 Discussion

Figure 1 provides the context for the LPE 'cooling event' at Mochras set within the background record. Shifts in bulk $\delta^{18}O_{carb}$ are coeval to the $\delta^{13}C_{org}$ change to heavier isotopic values (~930 mbs) and reach a maximum in the Margaritatus Zone (>1 ‰) (Ullmann *et al*., 2022). The bulk oxygen-isotope excursions of Mochras are affected by diagenesis and are deemed unlikely to reflect environmental conditions (Ullmann *et al*., 2022). However, oxygen isotope data from marine benthic and nektobenthic molluscs and brachiopods show heavier values during the late Margaritatus Zone concurrent with a positive shift in $\delta^{13}C_{org}$, indicating cooling during the LPE in the nearby Cleveland Basin (Robin Hood's Bay and Staithes) (Korte and Hesselbo, 2011) and this trend is also observed in several European sections (e.g. Korte *et al*., 2015). The duration of the +ve CIE has been estimated as ~0.4–0.6 Myr in the Cardigan Bay Basin (Ruhl *et al*., 2016; Storm *et al*., 2020; Pieńkowski *et al*., 2021).

5.1 Background sedimentological and environmental variations

The Mochras succession shows metre-scale alternating TOC-enhanced and Ca-rich lithological couplets (mudstone/limestone; Fig. 3). Previous assessments of the palaeoenvironmental signature of these TOC-enhanced and Ca-rich couplets indicate strongly that the different depositional modes are driven by orbital precession (Ruhl *et al*., 2016; Hinnov *et al*., 2018; Storm *et al*., 2020; Hollaar *et al*., 2021; Pieńkowski *et al*., 2021). Precession driven changes in monsoonal strength have been suggested to influence the deposition and preservation of TOC and carbonate in the Cardigan Bay Basin (Ruhl *et al*., 2016), although the impact may have been expressed, at least partially, by changes in strength of bottom currents in the seaway as a whole (Pieńkowski *et al*., 2021).

The preservation of primary carbonate is poor in the Mochras borehole, making it complex to determine in detail the relative importance of carbonate producers for the bulk carbonate content (Ullmann *et al*., 2022). However, Early Jurassic, pelagic settings in the Tethys region often received abiotic fine-grained carbonate from shallow marine carbonate platforms (Weedon, 1986; Cobianchi and Picotti, 2001; Krencker *et al*., 2020) and partly via carbonate producing organisms (such as coccolithophores in zooplankton pellets) (Weedon, 1986; van de Schootbrugge *et al*., 2005, e.g. Weedon *et al*., 2019; Slater *et al*., 2022). Coccolithophores are often poorly preserved and recrystallized (Weedon, 1986; Weedon *et al*., 2019; Slater *et al*., 2022). The organic matter found in the studied section of the Mochras borehole varies between 18 and 42% terrestrial phytoclasts (Fig. 4). Phytoclasts are common, but palynomorphs are relatively sparse and poorly preserved. Marine amorphous organic matter is the main constituent in the present study of particulate organic matter in unsieved macerated samples, in the interval studied here (951–918 mbs). Examination of variations in the terrestrial/marine proportions of organic matter, shows no correspondence between the type of organic matter and the TOC-enhanced or Ca-rich lithological alternations. However, previous research has indicated that the percentage of terrestrial phytoclasts show precession forcing independent of the lithological couplets (so out of phase with precession scale changes in Ca-TOC content) between 951–934 mbs in the Mochras core (Hollaar *et al*., 2021). Such orbital forcing of the terrestrial vs marine proportions of organic matter were also found in Early Jurassic sediments of Dorset, and were similarly independent of the lithological facies (Waterhouse, 1999). Terrestrial phytoclast content show a weak expression of long-eccentricity driven variations in the section studied (Fig. 4).

Fossil charcoal makes up a substantial proportion of the organic fractions (11% of the terrestrial fraction) and has previously been shown to vary considerably over long-eccentricity cycle 459 ± 1 peaking in abundance during the phase of maximum eccentricity (Hollaar *et al*., 2021). Microcharcoal also appears to be most abundant during the maximum phase of the subsequent long eccentricity cycle 458 ± 1 (Fig. 4). Additionally, K/I and S/I clay mineral ratios appear to alternate in response to long-eccentricity drivers (Fig. 4) up to 931 mbs where the clay mineral signature changes. Detrital clays form in soil weathering profiles and/or physical weathering of bedrock. Chemical weathering is enhanced in a high humidity environment with relatively high temperatures and rainfall, when clays are formed in the first stages of soil development. In the modern day, kaolinite is primarily formed in tropical soils, under year-round rainfall and high temperatures (Thiry, 2000). Smectite also occurs in the tropics, but is more common in the subtropical to Mediterranean regions, where humidity is still high, but periods of drought also occur (Thiry, 2000). Hence, smectite forms predominantly in soil profiles under a warm and seasonally dry climate (Chamley, 1989; Raucsik and Varga, 2008), and kaolinite in a year-round humid climate (Chamley, 1989; Ruffell *et al*., 2002). Similarly, alternating intervals of kaolinite and smectite dominance were observed for the Late Sinemurian (Munier *et al*., 2021) and the Pliensbachian of Mochras (Deconinck *et al*., 2019).

The predominantly detrital character of these clay minerals has been confirmed by TEM scanning of Pliensbachian smectite minerals, which revealed the fleecy morphology and lack of overgrowth (Deconinck *et al*., 2019). Therefore, the alternations of smectite and kaolinite are interpreted to reflect palaeoclimatic signatures of a changing hydrological cycle, with a year-round wet climate evidenced by high K/I ratios, and a more monsoon-like climate with seasonal rainfall with high S/I (Deconinck *et al*., 2019; Hollaar *et al*., 2021; Munier *et al*., 2021) (See Figs. 3,and 6). The intervals with a signal for weaker seasons appears to correspond to phases of low eccentricity in the 405 kyr cycle, and signals of greater seasonality with periods of high more pronounced eccentricity (Fig. 4) in the 405 kyr cycle. Between 951 and 930 mbs high K/I occurs during phases of low long eccentricity suggesting an enhanced hydrological cycle (Hollaar *et al*., 2021) with more intense weathering, and enhanced fine grained terrestrial runoff to the marine record (Deconinck *et al*., 2019). In contrast, phases of maximum long-eccentricity appear to be smectite-rich, indicating seasonal rainfall, enhanced fire (Hollaar *et al*., 2021) and thus periods of droughts, and lower terrestrial runoff and subsequent lower dilution (Deconinck *et al*., 2019).

Detrital elemental ratios increase accordingly during the smectite-rich phases, and are lower during kaolinite-rich phases between 951 and 930 mbs. Detrital elemental ratios can be used to explore changes in sediment composition (e.g. Thibault *et al*., 2018; Hesselbo *et al*., 2020b) and the similarity of the long-term trend in Zr/Rb and Si/Al (Fig. 3) indicates that these elemental ratios reflect grainsize. The clay fraction (hosting Al, and Rb (Chen *et al*., 1999)), diminishes upwards, whereas the coarser silt to sand fraction (associated with Si (Hesselbo *et al*., 2020b) and Zr (Chen *et al*., 2006)), increases upward (Figs. 3, 4). The grainsize changes inferred here reflect two overall coarsening upwards sequences (Figs. 3, 4). These sequences may reflect changes in clastic transport due to changes in the proximity to the shore/siliciclastic source, changes in runoff due to a changing hydrological cycle, changes in the intensity of weathering of the bedrock, or accelerated bottom currents with greater carrying capacity of coarser sediments.

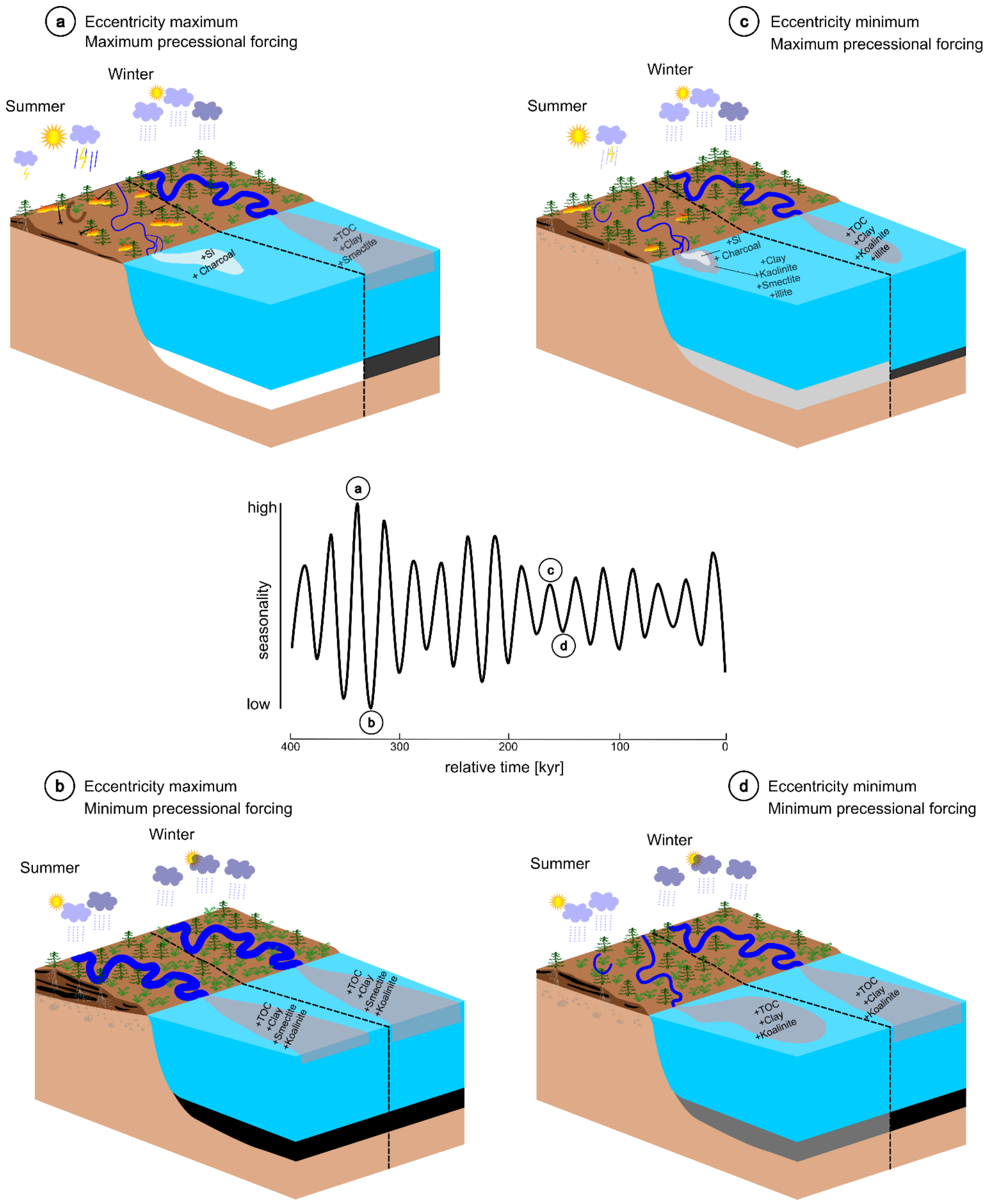

Figure 6: Scheme of four environmental scenarios under the influence of eccentricity on a precessional time scale. (a) Most extreme seasonal contrast in the northern hemisphere occurs during maximum precessional forcing (i.e. low precession index) and maximum amplitude modulation by eccentricity. Seasonal contrast leads to a wet season that allows biomass to build up, high terrestrial runoff, and relatively enhanced organic burial in marine settings. During the dry season, fuel moisture levels are lower and fires are rapidly ignited and spread. Intensified monsoonal rains may lead to enhanced coarse-grained terrestrial runoff. Overall, less terrestrial runoff during this dry season results in less dilution of carbonate production, and/or less primary productivity of

organic plankton. (b) Minimum precessional forcing and maximum amplitude modulation of eccentricity leads to the least seasonal contrast. Chemical weathering on land is more intense during this year-round humid climate. And although biomass is abundant, fire is suppressed due to the high moisture status. Both seasons are humid and have considerable terrestrial runoff, resulting in marine organic burial. (c) Moderate seasonality occurs during maximum precessional forcing and minimum amplitude modulation of the eccentricity cycle. During the wet season biomass grows, and during the dry season fires can occur due to drier fuel conditions. However, due to a lesser seasonal contrast the dry conditions are less pronounced and fire is not widespread. Runoff includes coarse- and fine-grained sediments, and charcoal during the dry-season. (d) Seasonal contrast is low during minimum precessional forcing and minimum amplitude modulation of the eccentricity cycle. Both seasons were humid and experienced runoff of fine-grained sediments and organic burial in marine settings. Moderately thick soil profiles could develop under this humid climate (figure developed from Martinez and Dera, 2015).

5.2 Depositional and environmental changes before and after the LPE +ve CIE

5.2.1 Climate forcing of the hydrological cycle

The LPE +ve CIE begins around 930 mbs in the Mochras core and encompasses the remaining part of the studied section (Fig. 4). We contrasted all the pre-CIE sediment signatures with those of the +ve CIE signatures using principal components analysis which indicates distinctly different sedimentary composition and environmental signature before and after the onset of the +ve CIE in Mochras (Fig. 5).

Before the +ve CIE onset, the clay mineral assemblage shows alternating phases of smectite and kaolinite, indicating pedogenic weathering. The relative abundance of the detrital clay types observed in the studied interval have the potential to hold important palaeoclimatic information regarding the hydrological cycle and the relative proportion of chemical weathering and physical erosion. The hydrological cycle was forced by the 405 kyr eccentricity before the +CIE, with alternating eccentricity maxima linked to enhanced seasonality (smectite) and eccentricity minima to an equitable wet climate (kaolinite) (Figs. 3, 6). and Higher frequency cycles are not observed in the clay mineral ratios, with no precession or obliquity forcing detected in the high-resolution part of the study 951–934 mbs (Hollaar *et al*., 2021) and no expression of the 100 kyr cycle in the record presented here. The formation of developed kaolinite-rich, and to a lesser extent smectite-rich soil profiles, requires a steady landscape for many tens of thousands of years, although the ~1 Myr timescale of Thiry (2000) seems excessive in our case given the clear expression of clay mineral changes through long-eccentricity cycles. Also, the transportation and deposition of continental clays will occur after soil formation and add further time between formation and final deposition (Chamley, 1989; Thiry, 2000). Thus, there is likely to be a lag of the climatic signal observed in the marine sediments (Chamley, 1989; Thiry, 2000). However, we note that high frequency climatic swings have been recorded in the clay mineral record in some instances, such as in the Lower Cretaceous in SE Spain (Moiroud *et al*., 2012). The limestone-marl alternations there are enhanced in smectite versus kaolinite and illite, respectively, reflecting precession scale swings from a semi-arid to a tropical humid climate (Moiroud *et al*., 2012). Precession and higher frequency shifts in the clay record are likely caused by fluctuations in runoff conditions rather than the formation of soils with a different clay fraction.

Directly after the initial +ve CIE shift from 930–924 mbs (Phase 1 of Fig. 4) little seems to change, and the system evidently continued to respond as before to the long eccentricity forcing, despite the predicted cooling (Korte and Hesselbo, 2011; Korte *et al*., 2015; Gómez *et al*., 2016). However, from around 924 mbs up to the top of the studied section (Phase 2 of Fig. 4) the clay mineral assemblage displays a distinctly different composition, with kaolinite dominating especially the early part of phase 2 of the LPE (Fig. 4). At the same time there is an enhancement of the primary minerals illite and chlorite, and I-S R1 (Fig. 4 and SI Fig. 1). Although an enhancement in detrital kaolinite indicates an acceleration of the hydrological cycle, detrital kaolinite is dual in origin and can also be derived from reworking of the primary source material (Deconinck *et al*., 2019). If the climate is cooler, chemical weathering becomes less dominant and physical erosion of the bedrock becomes the main detrital source of clay minerals. In the Cardigan Bay Basin, the bedrock of the surrounding Variscan massifs (such as the Scottish, Welsh and Irish massifs) were a likely source of these clays. In the Early Jurassic of the NW Tethys region, Lower Paleozoic mudrocks bearing mica-illite and chlorite were emergent (Merriman, 2006; Deconinck *et al*., 2019), hence the enhancement of illite and chlorite likely indicates physical erosion in the region surrounding the study site. Finally, authigenic clay particles could have been formed during burial diagenesis. At temperatures between 60-70 °C smectite illitization occurs and I-S R1 is formed; however, the high abundance of smectite in Mochras indicates limited burial diagenesis at that location (Deconinck *et al*., 2019). Weak-moderate thermal diagenesis is confirmed for the Pliensbachian of Mochras, with T_{max} from pyrolysis analysis between 421 °C and 434 °C (van de Schootbrugge *et al*., 2005; Storm *et al*., 2020). Therefore, I-S R1 in Mochras is interpreted to be derived from chemical weathering of illite (Deconinck *et al*., 2019). The coeval increase of these primary clay minerals, I-S R1 and kaolinite, indicate that during this period physical erosion dominated over soil chemical weathering (Deconinck *et al*., 2019; Munier *et al*., 2021). This is similar to what has been observed for the latest Pliensbachian in Mochras previously (Deconinck *et al*., 2019).

Erosion of weathering profiles transports clay minerals (including kaolinite and smectite) to the marine realm. In the ocean, the differential settling of kaolinite (near shore) and smectite (more distal) could occur based on the morphology and size of clay particles (Thiry, 2000). However, comparison of long-term inferred regional sea level changes from surrounding UK basins (Hesselbo, 2008) suggests that the relative proportions of smectite and kaolinite are not influenced by changes in relative sea level in the Pliensbachian of Mochras (Deconinck *et al*., 2019). On the assumption that the coarsening upward sequences at Mochras are indicative of relative sea level change, it can also be argued that the proximity to shore did not impact the proportions of smectite and kaolinite. Instead, we observe enhanced smectite during ‘proximal’ deposition and enhanced kaolinite at times of more ‘distal’ deposition, the opposite of what might be expected (Fig. 4).

We suggest that the first phase of the LPE (Fig. 4, phase 1) was characterised by repeated periods of rainfall in a seasonal climate forced by precession in which chemical weathering (smectite formation) dominated the sedimentary signatures. This corresponds to maximum long-eccentricity and shows the same climatic signature as during maximum eccentricity phases before the +ve CIE. This is then followed by a second phase (Fig. 4, phase 2) where the climate is generally cooler, overall potentially more arid, but with rainfall throughout the year over multiple precession cycles. This appears to have favoured deep physical erosion, owing to the abundance of primary clay minerals, kaolinite and I-S R1. This interval corresponds to a minimum phase in the

405 kyr eccentricity based on Storm *et al*. (2020). This interpretation is further supported by decreasing and then low microcharcoal abundance, pointing to suppression of fire activity at this time.

5.2.2 Climate forcing of sedimentary changes

Two coarsening upward cycles that predate the onset of the +ve CIE and continue for a few metres after its initiation, are present in the detrital elemental ratios (best expressed in Si/Al and Zr/Rb records) (Figs. 3, 4), and indicate a changing sediment influx over the studied interval. Previous study of the lithofacies of the Mochras borehole has also shown the coarsening-upward sequences of 0.5–3 m thickness, which are observed to be followed upwards by a thinner fining-upward succession (Pieńkowski *et al*., 2021). This reported fining-upward part is not reflected in the elemental ratios of the two sequences shown in this study. Furthermore, the coarsest phases of these sequences are approximately coeval with decreasing trends in the K/I ratio and increasing trends in the S/I. This could indicate that periods of a strong monsoonal/seasonal climate (indicated by S/I) brought coarser grained material to the basin, whereas periods of year-round humidity (K/I) are associated with higher chemical weathering (low Si/Al). Therefore, these two coarsening upwards cycles appear to link to increasing long-eccentricity. A similar mechanism has been inferred for the northern South China Sea region in the Miocene, where coarser grained material is found during periods of a strong summer monsoon and relatively lower chemical weathering (Clift *et al*., 2014). Present day studies show that bedrock erosion and associated sediment transport is greater in areas with high seasonal contrast (Molnar, 2001; Molnar, 2004). Hence, the Si/Al record also appears to reflect weathering and erosion conditions on land (Clift *et al*., 2014, 2020), driven by long-eccentricity modulated climate (SI Fig.5). However, other scenarios that would influence the grain size on this time scale cannot be dismissed and include changes in proximity to siliciclastic source, or changes in sediment transport via bottom water currents.

Changes in bottom water current strength and direction likely affected the depositional site of the Mochras core (Pieńkowski *et al*., 2021) although there is as yet no consensus on the processes that likely controlled these palaeoceanographic parameters. In the UK region, the North Sea tectonic dome structure may have disrupted the circulation in the N-S Laurasian Seaway (including the Viking Corridor) in the Late Pliensbachian when global sea-levels are suggested to have been low (Haq 2018) and therefore diminished the connectivity between western Tethys and the Boreal realm, hypothetically reducing poleward heat transport from the tropics (Korte *et al*., 2015). This mechanism has also been argued to explain the later cooling observed in NW Europe during the transition of the warmer Toarcian to the cooler Aalenian and Bajocian (Korte *et al*., 2015). Late Pliensbachian occlusion of the Viking Corridor is supported by the provincialism of marine faunas at this time, showing a distinct Euro-Boreal province and a Mediterranean province (Dera *et al*., 2011b). During the Toarcian, a northward expansion of invertebrate faunal species has been found (Schweigert, 2005; Zakharov *et al*., 2006; Bourillot et al., 2008; Nikitenko, 2008), indicating a northward (warmer) flow through the Viking corridor (Korte *et al*., 2015). More recently, a southward expansion of Arctic dinoflagellates into the Viking Corridor was suggested for the termination of the T-OAE (van de Schootbrugge *et al*., 2019), which is in agreement with a N to S flow through the Viking Corridor suggested by numerical models (Bjerrum *et al*., 2001; Dera and Donnadieu, 2012; Ruvalcaba Baroni *et al*. 2018) and sparse Nd-isotopes (Dera *et al*., 2009).

Over the European Epicontinental Shelf (EES), and the Tethys as a whole, a clockwise circular gyre likely brought oxygenated warm Tethyan waters to the southwest shelf, with a progressively weaker north and eastward flow due to rough bathymetry and substantial islands palaeogeography (Ruvalcaba Baroni *et al*., 2018). This predominantly surface flow is modelled to have extended to shelfal sea floor depths. Only episodically might nutrient-rich Boreal waters have penetrated south onto the EES in these coupled ocean-atmosphere GCM model scenarios (Dera and Donnadieu, 2012). The modelling also suggests – counter-intuitively – that the clockwise surface gyre of the Tethys extended further northwards and impacted the EES more effectively when the Hispanic corridor was more open. The timing of the opening of the Hispanic corridor is debated and varies from the Hettangian to Pliensbachian (Aberhan, 2001; Porter *et al*., 2013; Sha, 2019).

An alternative bottom current configuration was discussed for Mochras specifically wherein changes in north-to-south current strength (cf. Bjerrum *et al*., 2001) are proposed for the changes in grain size and silt or sand versus clay content via contour currents (Pieńkowski *et al*., 2021). A strong flow from the cooler and shallow boreal waters is hypothesized to have brought a coarser grainsize fraction in suspension and as bedload, which was then deposited in the Cardigan Bay Basin while flowing to the deeper and warmer waters of the peri-Tethys (Pieńkowski *et al*., 2021). Times of a strong north to south current are proposed to be associated with more oxygenated bottom waters (Pieńkowski *et al*., 2021). In contrast, when the north to south current became weaker, less coarse material will have been carried in suspension and as bedload, and a relatively higher clay proportion will have been deposited in the Cardigan Bay Basin (Pieńkowski *et al*., 2021). In this scenario, times of sluggish currents are associated with low bottom water oxygenation (Pieńkowski *et al*., 2021) and thus climate forcing of current strength could explain the deposition of alternating coarser and finer fractions in the Mochras borehole (Pieńkowski *et al*., 2021).

Our research suggests that orbital cycles both before and during the onset of the +ve CIE have a significant influence on seasonality and hydrology, affecting both fire regimes and sediment depositional character. Further research is required to consider how long-eccentricity and obliquity cycles might interact with north-south flow in the Cardigan Bay Basin and circulation processes. What is clear is that orbital cycles have impact on terrestrial processes in the terrestrial sediment source areas (Hollaar *et al*., 2021) and led to differences in deposition within the marine sediments in the Mochras core (Ruhl *et al*., 2016; Pieńkowski *et al*., 2021). Our data indicate that periods of coarser sediment deposition correspond to periods that include more seasonal climates before the onset of the +ve CIE (low kaolinite), which is in line with the hypothesized grainsize changes caused by contour currents (Pieńkowski *et al*., 2021). However, after the onset of the +ve CIE, although we suggest that the chemical weathering rate decreased, enhanced runoff and physical erosion are indicated by a peak in primary clay minerals and K/I. Enhanced runoff could be expected to impact the thermohaline contour currents (Dera and Donnadieu, 2012). Simultaneously, an increasingly cold climate (as indicated by enhanced physical erosion over chemical weathering) indicates a boreal influence. It remains to be determined to what extent orbital cycles might have the power to influence ocean circulation in the basin.

Relatively coarse sediments in the Late Pliensbachian have also been related to shallower sediment deposition in UK basins (Hesselbo and Jenkyns, 1998; Hesselbo, 2008; Korte and Hesselbo, 2011). Around the UK area, these regressive facies are plausibly related to enhanced sediment shedding from the North Sea dome structure during sea level low stand across the region (Korte and Hesselbo, 2011). Sequence stratigraphy of the Lower

Jurassic of the Wessex, Cleveland and Hebrides basins (Hesselbo and Jenkyns, 1998; Hesselbo, 2008; Archer et al., 2019) shows relative sea level changes and sand influxes in the late Margaritatus Zone in the studied basins. Noteworthy in the Mochras borehole are phases of low $\delta^{18}O$ of macrofossils which seem to correspond to high phases of macrofossil wood concomitant with low sea level, suggesting a possible control of relative sea level on the oxygen-isotope record and the source of detrital material (Ullmann *et al*., 2022). The broad spatial distribution of these basins suggests that associated regression and/or sediment influx is of at least regional scale (Hesselbo, 2008). The results presented here fall within this phase of regression (Hesselbo andand Jenkyns, 1998; Hesselbo, 2008).

In the context of the North Sea topographic dome structure (occlusion of the Viking Corridor in regional ocean flow) as a possible cause of the Late Pliensbachian cooling, these facies can be interpreted to represent shallowing upward in relatively shallow water, or the supply of coarser sediment into a deep-water system. The doming is hypothesized to have minimized or prohibited southward flow of cooler waters from the Boreal and northward flow from warmer waters from the Mediterranean area (Korte *et al*., 2015). The Mochras borehole is situated on the southwestern flank of the dome and would have been cut-off from the northern parts of the Laurasian Seaway, including the Hebrides Basin and Cleveland Basin (Korte *et al*., 2015). This change in seaway circulation could have impacted the source area of the detrital sediments in the Mochras borehole.

Superimposed on these larger-scale factors affecting grain size, orbital forcing clearly also had a strong impact. The Cardigan Bay Basin (Mochras) is positioned about 290 km to the SW of the Cleveland Basin and at a similar latitude, but to the W of the Wessex Basin (Ziegler, 1990; Torsvik andand Cocks, 2017), and is therefore expected to be impacted by the same regional changes in sea level and/or sediment flux. In the Late Pliensbachian of the Cleveland Basin, the detrital ratios of Si/Al, Zr/Al and Zr/Rb show similar coarsening upward sequences, which have been interpreted to reflect changes in riverine transport of siliciclastic grains and grainsize (Thibault *et al*., 2018). The inferred changes in sea-level in the Cleveland Basin occur at a 100 kyr pacing (Huang andand Hesselbo, 2014; Hesselbo *et al*., 2020b), potentially linking the regression cycles to short eccentricity (Huang *et al*., 2010 and refs therein) and long-eccentricity (Thibault *et al*., 2018). This would mean that eccentricity driven changes in inferred sea level change could be linked to glacioeustatic cycles during these times (Brandt, 1986; Suan *et al*., 2010; Korte andand Hesselbo, 2011; Krencker *et al*., 2019; Ruebsam *et al*., 2019, 2020b; Ruebsam and Schwark, 2021; Ruebsam and Al-Husseini, 2021). Glacioeustatic sea level changes are discussed for the Early Jurassic and Middle Jurassic (Krencker *et al*., 2019; Bodin *et al*., 2020; Ruebsam and Schwark, 2021; Nordt *et al*., 2022). A recent study on the rapid transgression observed at the Pliensbachian–Toarcian boundary, ruled out other mechanisms that could force sea level at this time scale, such as aquifer-eustacy, and emphasise that glacioeustatic changes in sea level are a likely possibility at times in the Early Jurassic (Krencker *et al*., 2019). Therefore, our findings overall are compatible with the episodic occurrence of continental ice at the poles (Brandt, 1986; Price, 1999; Suan *et al*., 2010; Korte and Hesselbo, 2011; Korte *et al*., 2015; Bougeault *et al*., 2017; Krencker *et al*., 2019; Ruebsam *et al*., 2019, 2020a, 2020b; Ruebsam and Schwark, 2021; Ruebsam and Al-Husseini, 2021).

6 Conclusions

The terrestrial environment adjacent to the Cardigan Bay Basin was strongly influenced by orbitally driven climate forcings (particularly precession and eccentricity) and colder climate linked to the Late Pliensbachian Event (LPE).Long-eccentricity forcing remained strong both prior to and during the LPE. Our results identify five swings in the climate in the study interval in tandem with the 405 kyr eccentricity minima and maxima. Eccentricity maxima are linked to precessionally repeated occurrences of a semi-arid monsoonal climate with high fire activity and relatively coarser sediment from terrestrial runoff. In contrast, 405 kyr minima in the Mochras core are linked to a more persistent, annually wet climate, low fire activity, and relatively finer grained deposits across multiple precession cycles. Although the 405 kyr cycle in the proxy records persists through the onset of the LPE +ve CIE, the expression in the clay mineralogical record changes to indicate year-round relatively cool and wet climate extended over multiple precession cycles driving significant erosion of bedrock. Therefore, both the Milankovitch forcings and larger climatic shifts operate in tandem to govern changes in the terrestrial environment.

Data availability: Supplementary data is available at the National Geoscience Data Centre at Keyworth (NGDC) at (doi to be added) for the interval 934–918 mbs. All data presented for the interval 951–934 mbs is available at the National Geoscience Data Centre at Keyworth (NGDC) at https://doi.org/10.5285/d6b7c567-49f0-44c7-a94c-e82fa17ff98e (Hollaar *et al*., 2021). The full Mochras XRF dataset is in Damaschke *et al*. (2021).

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