Non-marine carbon-isotope stratigraphy of the Triassic-Jurassic transition in the Polish Basin and its relationships to organic carbon preservation, pCO₂ and palaeotemperature

Grzegorz Pieńkowski, Stephen P. Hesselbo, Maria Barbacka, Melanie J. Leng

A R T I C L E   I N F O

Keywords:
Chemostratigraphy
Correlation
Paralic facies
Palaeoclimate
Carbon isotopes

A B S T R A C T

New carbon-isotope data obtained from homogenous organic material (separated microfossil wood; δ¹³Cwood) from the upper Rhaetian and entire Lower Jurassic permit chemostratigraphic correlation of these marginal/non-marine deposits with the biostratigraphically well-constrained Llanbedr (Mochras Farm) core in N Wales and other marine profiles, supported by sequence stratigraphic correlation and biostratigraphical markers. Statistically significant (Rs = 0.61) positive exponential correlation between δ¹³Cwood values and continental TOC (TOCcont) concentrations occurs and can be defined empirically by equation. Changes of δ¹³Cwood observed in C₃ plants depends on δ¹³CO₂ of atmosphere and can be modulated by other factors such as pCO₂ causing fractionation (enrichment in ¹²C) of C isotopes in source C₃ plants and, to lesser extent, by soil moisture content. Floral remains occurring in the relatively stable palaeolatitude and climatic zone of the Polish Basin in the time interval studied lend no support for significant precipitation impact on the δ¹³C fractionation, although enhanced precipitation could have had a limited impact during the Toarcian Oceanic Anoxic Event (T-OAE). We argue that the observed relation between δ¹³Cwood values and TOCcont reflects the global carbon cycle forcing. Such correlations may develop because fluxes of ¹²C-enriched methane, mobilized from near-surface carbon sources, lead to global warming, decreased δ¹³Cwood and enhanced (usually fungally-mediated) decomposition of the terrestrial carbon pool, while subsequent massive burial of organic carbon results in higher δ¹³C values in all carbon cycle reservoirs, and the attendant drawdown of atmospheric CO₂ leads to global cooling and promotes sequestration of soil organic matter. In turn, this relation can be used as an indirect indicator of atmospheric temperature trends, although organic carbon isotope records are potentially subject to many different influences. Based on the δ¹³Cwood /TOCcont relationship, an approximate qualitative estimation of general trends in air temperature is suggested for c. 40°N paleolatitude and the warm temperate climatic zone. The observed hypothetical trends in temperature are generally in concordance with pCO₂ trends calculated from stomatal index. A weak δ¹³Cwood and TOCcont correlation in Rhaetian deposits is explained by local environmental factors (TOC concentration dependent on a more localized fluvial plain settings), while mostly deltaic – coastal deposits contain more representative, averaged material delivered from a large catchment area.

1. Introduction

Correlation of biostratigraphically-constrained marine sections with marginal-marine or continental deposits with poor biostratigraphical control is regarded as one of most complex and challenging, yet very important, problems in stratigraphy. As shown for example by Hesselbo and Pieńkowski (2011), carbon-isotope (δ¹³C) chemostratigraphy provides a means of age correlation for strata deposited in diverse marine and non-marine depositional environments. However, such correlations are reliable only when appropriate profiles are accessible. In particular, long, cored borehole sections offer the opportunity to reconstruct Earth system evolution in the geological past and contain, for example, the...
records of tectonic, carbon cycle, climatic and biological cycles. Expanded core sections, providing a continuous record of the Lower Jurassic series – such as the Llanbedr (Mochras Farm) core recovered in Wales, UK (Woodland, 1971; Katz et al., 2005; Hesselbo et al., 2013; Ruhl et al., 2016; Xu et al., 2018; Storm et al., 2020), hereafter referred to as Mochras, and Kaszewy borehole drilled in central Poland (Figs. 1, 2), are exceptionally valuable. The isotope ratios of atmospheric CO₂ (δ¹³CO₂) to a certain degree reflect perturbations in atmospheric C fluxes (e.g. Pagani et al., 2006 for the PETM case), and therefore estimates of δ¹³CO₂ are central to estimating past climate sensitivity to changes in atmospheric CO₂ pressure (pCO₂). While inputs of isotopically light carbon to the atmosphere-ocean system result in decreasing δ¹³CO₂ and increasing pCO₂ recorded as a negative carbon isotope excursion in both inorganic and organic carbon enriched burial of organic carbon and enhanced silicate weathering leads to a fall in atmospheric pCO₂ and a positive excursion (Gröcke et al., 1999; Kump and Arthur, 1999). The pCO₂ and related temperature is a basic climate variable, yet its estimates in deep geological past are difficult and often controversial as the carbon cycle involves multiple sources and sinks in a dynamic rather than a static mass-balance equilibrium (Kump and Arthur, 1999). The near surface carbon cycling, such as release of methane clathrates, permafrost destruction, decomposition of soil organic matter (acting as a source of δ¹³C-depleted carbon) and burial of organic matter and carbonate or silicate weathering (acting as a sink – Cohen et al., 2004; Them III et al., 2017b; Kump, 2018; Ullmann et al., 2020) played a major role in modulating atmospheric CO₂ concentrations. Nevertheless, values of δ¹³C (or δ¹³C TOC) which come from sedimentary rocks (specifically, the values obtained from fossil plants) are widely used to estimate δ¹³C of ancient plants and infer changes in δ¹³CO₂. Additionally, δ¹³C depends on pCO₂ - the higher pCO₂, the stronger fractionation of δ¹³C by plants towards light values. Thus the currently debated effect of CO₂ concentration on δ¹³C ratios (δ¹³C) is crucial to reconstructing ancient environments and quantifying the carbon cycle (Hare et al., 2018; Ruebsam et al., 2020).

Currently most of the temperature estimates for this time interval have been based principally on oxygen isotopes (δ¹⁸O) from calcarceous macrofossils, such as oysters, brachiopods and belemnites, or calcareous nannofossils (Suan et al., 2008, 2010; Korte and Hesselbo, 2011; Dera et al., 2011; Silva and Duarte, 2015; Gómez et al., 2016; Price et al., 2016; Peti and Thibault, 2017 – for recent comprehensive summary see Ruebsam et al., 2020) from European marine sections. However, temperature interpretation based on δ¹⁸O can be distorted by changes in the local source(s) of water and local/regional evaporation precipitation dynamics as well as diagenesis, and therefore choosing the sites and samples unaffected by these processes is essential. Estimating palaeotemperatures based on δ¹⁸O is also prone to problems such as the vital effects of the organism precipitating the carbonate (Price, 2017). Furthermore, the thermal perturbations are likely to have been controlled also by the silicate weathering flux, in the case of the Lower Jurassic extending these times of warmth by limitation of the global silicate weathering flux (e.g. Ullmann et al., 2020). It should be noted that Os-isotope proxies from eastern and western Panthalassa show a modest decline in weathering intensity after the peak of the negative carbon-isotope excursion of the T-OAE (Them III et al., 2017b, Kemp et al., 2020). Nevertheless, a broadly synchronous coupling between massive carbon release and enhanced global continental crust weathering during T-OAE is not controversial (Remirez and Algeo, 2020b). According to Remirez and Algeo (2020a,b), following previous authors cited therein, the exceptional expression of the T-OAE in the NW European Basin (including paleotemperature interpretation) point to strong regional oceanographic factors in the development of watermass stratification, deepwater anoxia, and enhanced organic matter accumulation – although an idea of very low-salinity conditions in the NW European Basin, caused by and its isolation and enhanced freshwater runoff during the T-OAE (Remirez and Algeo, 2020a) remains controversial (Hesselbo et al., 2020c). Herein, we discuss δ¹³C values and their relation to TOC_cont in context of the global carbon cycle and resulting interpretation of climate changes - including temperature trends – with particular reference to the Kaszewy core.

2. Geological setting

The Polish Basin is a large, now tectonically inverted, sedimentary basin covering most of modern-day Poland, and representing the eastern arm of the NE European epicontinental basin (Dadlez et al., 1998; Dadlez, 2003; Fig. 1). The upper Rhaetian strata and the whole Lower Jurassic series belong to the Kamienna Group composed of sili-ciclastic deposits (mudstones, sandstones) with subordinate conglomerates, coaly and siderite beds. Throughout the Mesozoic, regional subsidence patterns in the Polish Basin were dominated by the development of the NW–SE-trending Mid-Polish Trough, which was filled with up to 8000 m of Permian–Mesozoic sediments, of which c. 1200–1300 m belongs to the Kamienna Group (Pieńkowski, 2004). Intermittently, this long-term thermal subsidence was influenced by local modifications imposed mainly by lateral salt movements in its central and NW segments, creating salt pillows. The Kaszewy borehole is situated at a slope of one of these broad and flat pillows, called Wojszyce structure (Dadlez, 1998). In the latest Triassic – Early Jurassic times the Polish basin was located at a latitude about 40°N, within a warm temperate/winter-wet climate belt (e.g. Rees et al., 2006; Dera et al., 2009). Sedimentology, lithofacies, ichnofacies, biofacies as well as sequence stratigraphy and palaeogeography of the NE European basin have been studied for decades and published in number of papers (Dadlez and Kopik, 1972; Pieńkowski, 2004, 2014; Pieńkowski, 2014; Pieńkowski et al., 2008; Pieńkowski and Waksmundzka, 2009; Hesselbo and Pieńkowski, 2011; Pieńkowski et al., 2012, 2014, 2016; Barth et al., 2018a, 2018b).

The latest Rhaetian and Early Jurassic were times of extreme
Fig. 2. Stratigraphic correlation between Kaszewy (δ\textsuperscript{13}C\textsubscript{wood} - black) and Mochras borehole (blue – δ\textsuperscript{13}C\textsubscript{TOC}, dashed red – δ\textsuperscript{13}C\textsubscript{carbonate} and black dots - δ\textsuperscript{13}C\textsubscript{wood}), using carbon isotope chemostratigraphy, sequence stratigraphy and biostratigraphical proxies. Reference C isotope curves of the profile Mochras (biostratigraphically calibrated) are based on Storm et al., 2020 (δ\textsuperscript{13}C\textsubscript{TOC}, δ\textsuperscript{13}C\textsubscript{wood}) and Katz et al., 2005 (δ\textsuperscript{13}C\textsubscript{carbonate}). For the Rhaetian section, the reference curve is based on Ruhl et al. (2010) (black). TOC content in Kaszewy is shown – marine (blue line) and continental (red line). Note two options of chemostratigraphic correlation of the upper Sinemurian – lower Pliensbachian transition. The first option (green bars) implies condensation and incomplete δ\textsuperscript{13}C\textsubscript{wood} dataset in Kaszewy in the lowermost Pliensbachian, while the second correlation (pink bars) implies strong erosion at the sequence boundaries V and VI, resulting in absence of most of the lower Pliensbachian and atypical (for Polish basin) sedimentary development of both upper Sinemurian and lower Pliensbachian. Considering the above reasons, δ\textsuperscript{13}C\textsubscript{wood} correlation and correlations with other profiles (Fig. 3), the first option was adopted as more likely, despite more similar general isotope trends when adopting the second option. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
environmental change. Through this period there are well-documented examples of enhanced volcanic activity, rapid transitions from cold, or even glacial climates, through to super-greenhouse events, multiple large-magnitude isotopic anomalies, global sea-level changes, and mass extinctions. These events not only reflect changes in the global climate system but are also thought to have had significant influence on the evolution of Jurassic marine and terrestrial biota. However, our knowledge to date is fragmentary, largely because of the fragmentary and discontinuous nature of the existing datasets – particularly, with respect to atmospheric/terrestrial realm. Studies hitherto have largely focused on two events, connected to the development of Large Igneous Provinces (LIPs): Central Atlantic Magmatic Province (CAMP) at the Rhaetian-Hettangian transition, and early Toarcian Karoo-Ferrar Large Igneous Province, connected with the Toarcian Oceanic Anoxic Event (T-OAE – Jenkyns, 1988). While marine strata have been studied more extensively in terms of oceanographic, climatic and biological processes (Korte and Hesselbo, 2011; Dera et al., 2011; Lindström et al., 2012, 2017; Harazim et al., 2013; Suan et al., 2015; Ruhl et al., 2016; Them et al., 2018; Van de Schootbrugge et al., 2013, 2019; Storm et al., 2020), reports from continental or marginal-marine materials remain relatively sparse (e.g. Pieńkowski, 2004, 2014; McElwain et al., 2005; Hesselbo and Pieńkowski, 2011; Pieńkowski et al., 2012, 2014, 2016; Them et al. 2017 a,b; Baker et al., 2017; Xu et al., 2017; Ruesbam et al., 2019; Liu et al., 2020). Records in epicontinental seaways may potentially show a pronounced regional overprint on carbon isotope composition of sedimentary organic matter (McArthur et al., 2008; Ruhl et al., 2010; Remirez and Algeo, 2020b), usually resulting from mixing of different organic components (Suan et al., 2015), but results obtained from homogenous woody material (δ13Cwood) are more reliable in interpreting global changes more directly, as they reflect atmospheric (i.e. global) processes (Grocke, 2002) and can therefore be used more confidently for chronostratigraphic correlations (see for Lower Jurassic examples: Hesselbo and Pieńkowski, 2011; Pieńkowski et al., 2016; Them III et al., 2017b; Ruesbam et al., 2020; Storm et al., 2020).

3. Materials and methods

3.1. Lithology and stratigraphy

A new and unique geological archive (2050.6 m through the Norian - Rhaetian- Lower Jurassic strata) comes from the Kaszewy 1 borehole (Fig. 1), drilled in central Poland (52° 12′ 00.06″ N; 19° 29′ 35.38″ E) by PGE (Polish Energy Group) in order to characterize the potential for carbon capture and storage (CCS) in that area. The entire core has been logged and sampled. Samples for geochemical analyses have been taken from similar lithologies (mudstones) in order to avoid lithological bias. The length and continuity of the Kaszewy core (98% core recovery) is focused on two events, connected to the development of Large Igneous Provinces (LIPs): Central Atlantic Magmatic Province (CAMP) at the Rhaetian-Hettangian transition, and early Toarcian Karoo-Ferrar Large Igneous Province, connected with the Toarcian Oceanic Anoxic Event (T-OAE – Jenkyns, 1988). While marine strata have been studied more extensively in terms of oceanographic, climatic and biological processes (Korte and Hesselbo, 2011; Dera et al., 2011; Lindström et al., 2012, 2017; Harazim et al., 2013; Suan et al., 2015; Ruhl et al., 2016; Them et al., 2018; Van de Schootbrugge et al., 2013, 2019; Storm et al., 2020), reports from continental or marginal-marine materials remain relatively sparse (e.g. Pieńkowski, 2004, 2014; McElwain et al., 2005; Hesselbo and Pieńkowski, 2011; Pieńkowski et al., 2012, 2014, 2016; Them et al. 2017 a,b; Baker et al., 2017; Xu et al., 2017; Ruesbam et al., 2019; Liu et al., 2020). Records in epicontinental seaways may potentially show a pronounced regional overprint on carbon isotope composition of sedimentary organic matter (McArthur et al., 2008; Ruhl et al., 2010; Remirez and Algeo, 2020b), usually resulting from mixing of different organic components (Suan et al., 2015), but results obtained from homogenous woody material (δ13Cwood) are more reliable in interpreting global changes more directly, as they reflect atmospheric (i.e. global) processes (Grocke, 2002) and can therefore be used more confidently for chronostratigraphic correlations (see for Lower Jurassic examples: Hesselbo and Pieńkowski, 2011; Pieńkowski et al., 2016; Them III et al., 2017b; Ruesbam et al., 2020; Storm et al., 2020).

3.2. Carbon isotopes

Phytoclast samples for analysis were manually picked from HF palynomacerals and subsequently dried, weighed and sealed in tin capsules. Phytoclast samples were run at two laboratories, depending on their mass (Supplementary data 1). At the Research Laboratory for Archaeology and History of Art (RLAHA), University of Oxford, UK, larger samples were run on a Sercon Europa EA-GSL sample converter connected to a Sercon 20–22 stable isotope gas-ratio mass spectrometer running in continuous flow mode with a helium carrier gas with a flow rate of 70 ml per min. Carbon isotope ratios were measured against an internal alanine standard (δ13Calanine = −26.9 ± 0.2‰ V-PDB) using a single point calibration. The in-house RLAHA alanine standard is checked weekly against USGS40, USGS41 and IAEA-CH6 international reference materials. At the at the British Geological Survey, Nottingham, UK, smaller samples were analysed by combustion in a Costech Elemental Analyser (EA) on-line to a VG TripleTrap and Optima dual-inlet mass spectrometer, with δ13Corg values reported relative to V-PDB following a within-run laboratory standard calibration, with NBS-18, NBS-19 and NBS-22. Replicate analysis of well-mixed samples show a reproducibility of ± < 0.1‰ (1 SD).

3.3. Total Organic Carbon, Rock Eval 6 pyrolysis

Total Organic Carbon (TOC) analyses of 300 samples were performed using the chromatographic, coulometric method (procedure PB – 23) using an automated LECO analyser (Supplementary data 1). A total of 52 naturally carbonate-free mudstone samples (mostly those representing mixed environments – in order to evaluate which ones contain more marine kerogen) were analysed in the Rock Eval pyrolysis apparatus - model 6 Turbo (Vinci Technologies), in the laboratory of the Polish Geological Institute (Supplementary data 2). Crushed bulk rock material was thermally decomposed in a helium or nitrogen atmosphere. Every sample was heated to 650 °C. The amount of free hydrocarbon (S1) thermally liberated from a rock sample at 300 °C is measured using a Flame Ionization Detector (FID). Volatile components released during pyrolysis are separated into two streams. One of them passes through the FID and is registered as peak S2. Another volatile component released during pyrolysis, i.e., carbon dioxide generated from kerogen, is recorded by a thermal conductivity detector as the S3 peak. Repeat analyses of the parameters S1, S2, S3 and TOC agree within ± 0.05. The Oxygen Index (OI, in mg CO2/g TOC) is calculated according to the formula (S3*/100)/TOC. Hydrogen index (HI) = S2 (mg/g)/TOCx100. Only mudstone was sampled, therefore some parts of profile are sampled more sparsely due to the high sandstone content and sample spacing is heterogeneous. Coals and coaly mudstones were avoided as non-representative. Parts containing marine kerogen were sampled more sparsely. Only samples paired with δ13Cwood data and of clearly continental origin (dominating kerogen type III and IV, about 80% of all samples) were used for δ13Cwood, TOCcont and temperature interpretation (Fig. 2; Supplementary data 1, 2).

3.4. Stomatial index (SI)

Fifty-one specimens containing fossil plant remains from the Kaszewy core interval 1936.5–1296 m were studied. Only few levels contained well preserved remains with preserved cuticle, belonging mostly to genera Baiera Braun emend Florin and Czekanowskia Herr, on which the SI counting has been performed. Their determination was possible on a genus level only, but that is not a limitation for stomatal index study. Baiera and Czekanowskia are ginkgophytes, which means that all studied remains have the same nearest living equivalent species (NLEs), making ginkgophytes especially suitable for SI calculation (Beerling et al., 1998; Xie et al., 2006). The cuticles were prepared with standard methods in Schulze solution followed by 3% KOH. Only parts containing marine kerogen were sampled more sparsely. Only samples paired with δ13Cwood data and of clearly continental origin (dominating kerogen type III and IV, about 80% of all samples) were used for δ13Cwood, TOCcont and temperature interpretation (Fig. 2; Supplementary data 1, 2).
ED – number of epidermal cells/area (Jones and Rowe, 1999). SI was counted according to Kürschner (1996), using 40 microphotographs for Baiera and 13 microphotographs for Czekanowskia (totally for 21 leaf fragments). For those specimens, 8–19 measurements were carried out, depending on the state of preservation. The atmospheric CO2 level was estimated as 600 ppm × SR where SR is the Stomatal Ratio, i.e. SI of the nearest living equivalent divided by SI of fossil plant: SR = SI of NLEs/ SI fossil. Standardization was applied after Berner and Kothavala (2001) and SI for the extant Ginkgo biloba was used according to Beierling and Royer (2002a), as 9.1 at 350 ppm. Based on this relationship, palaeoatmospheric CO2 pressure (pCO2) was estimated, using stomatal parameters of fossil plants – stomatal density (SD) and stomatal index (SI) in relation with parameters of their NLEs, defined as the stomatal ratio (SR) (Beierling and Royer, 2002a; Beierling and Kothavala, 2002b; Royer et al., 2001). Plant reaction appears in leaf structure, especially in stomatal density, which is closely associated with fluctuations of atmospheric pCO2 and displays an inverse correlation to pCO2 (Woodward, 1987; Beierling, 1999; Poole and Kürschner, 1999; Mosbrugger, 1999; Royer et al., 2001; Haworth and Raschi, 2014). SI may be influenced by many factors that affect plants during their life, but, as has been discussed by many authors (Chen et al., 2001; Erdei et al., 2002b; Korte and Hesselbo, 2011), other factors such as kinetic isotope fractionations should be also considered as additional factors. The general pattern of the curve shows more negative values (mostly below −27‰, with minimum of −28.3‰) in the strata assigned (based also on palynomorphs) to the Rhaetian (1905–1979 m) with three more prominent negative CIEs. Upwards from the Triassic/Jurassic boundary the δ13Cwood values show a general, long-term trend toward heavier values, punctuated by five conspicuous negative CIEs (Figs. 1, 2). This trend is reversed at 1445.9 m and then again continuing upwards toward heavier values, with the heaviest value (−22.9‰) recorded in the upper Pliensbachian at 1266 m. A series of five prominent, step-wise CIEs (down to −30.6‰ at 1237.5 m, starting at 1247.3 m and ending at 1188.7 m) is assigned to the Toarcian Oceanic Anoxic Event/T-OAE, also called the Jenkins Event) and can be clearly correlated with reference profiles of Mochras, Yorkshire, Peniche (Thibault et al., 2017), and the composite reference profile by Ruebsam and Al-Husseini (2020) – Fig. 4. The major negative excursion of the T-OAE is unambiguous and has already been recognized (with further biostratigraphic constraint) in many locations across the same basin (Hesselbo et al., 2000; Hesselbo and Pieńkowski, 2011). From 1188.7 m upwards the δ13Cwood values stabilize at a level with a median c. −24‰. The section above the T-OAE is a special case because it has already been demonstrated that the lower Toarcian is commonly truncated in the Polish Basin beneath younger Toarcian strata (Pieńkowski, 2004; Hesselbo and Pieńkowski, 2011).

All conspicuous negative CIEs can be correlated with the δ13C curves from the Mochras borehole (Katz et al., 2005; Storm et al., 2020) and from the St Audrie’s Bay section for the Rhaetian (Korte et al., 2009; Ruhl et al., 2010; Fig. 2). δ13Cwood values from Kaszewy are generally heavier by about 1‰ than the corresponding values from δ13Corg values from Mochras, likely due to contributions from marine organic matter to the Mochras curve (Storm et al., 2020), but what is important for chemostratigraphic correlation is concordance in general trends. This average δ13C contrast disappears considering the results of δ13Cwood from Mochras. The correlation presented shows also a general concordance with other biostratigraphically constrained δ13C curves from marine sections in Europe and North America, e.g. almost complete Lower Jurassic section from Schandelah in Germany (Van de Schootbrugge et al., 2019), Rhaetian-Hettangian in Danish Basin (Lindström et al., 2012) and Upper Triassic in North America (Whiteside et al., 2010; Bartolini et al., 2012), Sinemurian-Pliensbachian in Yorkshire, UK (Korte and Hesselbo, 2011), Hettangian to Pliensbachian in Dorset, UK (Schöllhorn et al., 2020), Sinemurian-Pliensbachian in Spain (Gómez et al., 2016), Sinemurian in the UK (Riding et al., 2013; Hesselbo et al., 2020a), the Pliensbachian in the western US (De Lena et al., 2019), Pliensbachian and Toarcian in Portugal (Fantasia et al., 2019) and France (Hermoso and Pellenard, 2014) – Fig. 3; Supplementary data 3. In Mochras (Storm et al., 2020) and Dorset (Schöllhorn et al., 2020),
Fig. 3. Chemostratigraphic correlation of Kaszewy profile with other C isotope curves and major negative CIEs observed in reference sections and recognized in Kaszewy (see Supplementary data 3). All curves are based on δ13Corganic, except for Basque-Cantabrian Basin (Quesada et al., 2005) where the δ13C values are based on belemnites. The Toarcian Oceanic Anoxic Event (=Jenkyns Event) section is presented in detail in Fig. 4.

Fig. 4. Chemostratigraphic correlation of the lower Toarcian section and the Toarcian Oceanic Anoxic Event (=Jenkyns Event) of Kaszewy profile with selected C isotope curves, showing correlation of C isotope steps T. 1–5 (defined in other profiles in the Polish basin - Hesselbo and Pieńkowski, 2011) with other C isotope curves and composite picture of T-OAE. Biostratigraphy: P.sp. – Pleuroceras spinatum zone (Pliensbachian), D.Tenuicostaum – Dactylioceras tenuicostatum zone (Toarcian), H.falciferum – Harpoceras falciferum zone (Toarcian). Note that positions of Dactylioceras tenuicostatum/Harpoceras falciferum boundary (sub-Boreal realm) and the Dactylioceras polymorphum/Harpoceras levisoni boundary (Mediterranean Realm) may slightly differ (Ruebsam and Al-Husseini, 2020). δ13Corg trends: r.l. – rising limb, pl. – plateau, vl – valley, f.l. – falling limb; A-O – δ13Corg steps (after Ruebsam and Al-Husseini, 2020). Kaszewy profile is much more expanded than coeval marine sections, while most of the section correspond to the falling limb of T-OAE (5 steps – T. 1–5) and rising limb is less complete, likely with condensations or hiatuses in the upper part (see Hesselbo and Pieńkowski, 2011).
there is a uncertainty with the placement of the upper Sinemurian CIE, labelled S2 in Figs. 2, 3. In other reports, the conspicuous negative CIE occurs in the obtusum zone, while in Mochras and Dorset the negative CIE occurs within the obtusum zone. The answer could be missing section (particularly in Dorset) or imprecise biostratigraphy. In the meantime, the stratigraphical position of this CIE in Mochras and Kaszeway remains open – it can be either of obtusum or oxynotum age. It should be noted, that some parts of the profile at Kaszeway represent reduced or condensed sections, particularly at the sequence boundaries (e.g. the Pliensbachian section). There are some reductions in thickness of Pliensbachian section observed between Kaszeway (located at the side of salt pillow) and Krośniewice borehole (located between salt structures 20 km to the NW – Dadlez, 1973). Likely, these reductions can be explained somewhat by the reduction in the rate of marine sedimentation (e.g. the Pliensbachian section). There are some reductions in thickness of Pliensbachian section observed between Kaszeway (located at the side of salt pillow) and Krośniewice borehole (located between salt structures 20 km to the NW – Dadlez, 1973). Likely, these reductions can be explained somewhat by the reduction in the rate of marine sedimentation.

The cross-plot of the δ13Cwood to TOCcont ratio (224 Rhaetian and Lower Jurassic samples: Fig. 5) shows significant positive correlation between the two variables with Pearson correlation coefficient \( r = 0.53 \) \((R^2 = 0.28)\) for linear fit and \( r = 0.59 \) and \( R^2 = 0.59 \) for exponential fit. The p-value is \(< 0.0001\). The result is statistically significant at \( p < .05\). For the Rhaetian (66 samples) \( r = 0.16\) for the Lower Jurassic (159 samples) \( r = 0.57\); Spearman’s Rank Correlation Coefficient (in 224 samples): \( R_s = 0.61\), \( p\) val \(= 6.44^{-24}\); the result is significant the significance threshold 0.001 (see Supplementary 4). The correlation is interpreted as a result of differentiated decomposition of terrestrial organic matter related to climate changes, but also other factors could have influenced the observed relationship.

4.2. Total organic carbon (TOC), kerogen analyses

In the Rhaetian strata, TOC is usually below 1%, except for some darker mudstones interpreted to have been deposited in local ponds and lakes, where TOC content can reach 4% and more. The lowermost part of the Rhaetian strata is represented by red beds yielding very few or no palynomacerals (Supplementary data 1) and these strata are excluded from further considerations concerning TOC and δ13Cwood as non-representative. In the Hettangian section TOC is in places abundant (Fig. 2), but most of these intervals contain marine kerogen (Supplementary data 2) occurring mostly in strata coeval with major transgressions of the planorbis and iaxisus zones (Pieńkowski, 2004; Barth et al., 2018a, 2018b). In the upper Sinemurian – lower Pliensbachian section the studied samples usually contain more than 1% (1.14% in average). Moreover, a few beds have TOC up to 3–5%, and the sample from the lowest part of Pliensbachian (depth of 1462.3 m) contains organic carbon as high as 15.3%. These samples contain marine kerogen associated with the jamesoni-ibex transgression and maximum flooding surface (Pieńkowski, 2004; Barth et al., 2018a, 2018b). In the upper Pliensbachian – Toarcian section TOC is usually in the range of 0 to 3% (1.0% on average). In contrast, most of the lower Toarcian is characterized by a low content of TOC (usually below 0.5%). In order to check the character of the kerogen, 52 mudstone samples from selected sections (based on sedimentological studies) were screened using RockEval 6 pyrolysis and interpreted using the Van Krevelen plot (Supplementary data 2). Kerogen was classified in classical types II, III and IV (type I has not been found), ranged in the order of H/C ratio obtained from the RockEval6 analyses (Vandenbroucke and Largeau, 2007). Type II originated in a shallow marine environment with phytoplanktonic input as the primary source. Type III and IV are found in deltaic/fluvial and coastal settings and derive from higher plant debris, commonly highly reworked and often degraded by fungi (Fig. 6). Type IV is considered as heavily altered Type III with very low H/C or Hydrogen Index and containing only the most resistant chemical constituents of the terrestrial OM (inertinite). Type II is usually mixed with kerogen type III. Samples containing only kerogen type III and IV (in total 224 samples) were taken for δ13Cwood to TOCcont interpretation (Figs. 2, 5; Supplementary data 1, 2).

4.3. Carbon isotope – total organic carbon ratio

The cross-plot of the δ13Cwood to TOCcont ratio (224 Rhaetian and Lower Jurassic samples: Fig. 5) shows significant positive correlation between the two variables with Pearson correlation coefficient \( r = 0.53 \) \((R^2 = 0.28)\) for linear fit and \( r = 0.59 \) and \( R^2 = 0.59 \) for exponential fit. In both cases the p-value is \(< 0.5\). The result is statistically significant at \( p < .05\). The Spearman’s Rank Correlation Coefficient (Rs) has been also applied, as it is more appropriate in the case of exponential correlation: \( R_s = 0.61\), \( p\) val \(= 6.44^{-24}\) (for all 224 samples), the result is significant with the significance threshold 0.001 (Supplementary data 4). Samples are shown with symbols representing their stratigraphical position (Fig. 5). \( R_s = 0.61\) for all samples containing continental
kerogen is satisfactory for demonstrating a statistically significant trend/correlation, classified (depending on literature) as a moderate or strong (Evans, 1996), while the Rhaetian section (66 samples) is characterized by a weak linear positive correlation of \( r = 0.16 \), with a noticeable number of outliers. The correlation coefficients are highest when applying exponential law fits, and slightly degrade when using linear fits (with exception of the Rhaetian, where linear fit is slightly higher).

### 4.4. Stomatal index and pCO\(_2\)

The SI and pCO\(_2\) could be calculated only in six levels, but four of them occurred in the relatively narrow upper Hettangian interval, and as they suggest marked fluctuations of pCO\(_2\), they could be compared with results obtained from the corresponding \( \delta^{13}\text{C}_{\text{wood}} \) results. SI and pCO\(_2\) values were calculated from ginkgophytes: *Czekanowskia* sp. 1 - upper Hettangian, 1694.2 m (1400 ± 200 ppm CO\(_2\)); *Baiera* sp. 2 - upper Hettangian, 1688.9 m (1500 ± 400 ppm CO\(_2\)); 1686.9 m (1200 ppm ± 150 ppm CO\(_2\)); *Czekanowskia* sp. 1 - upper Hettangian, 1677.2 m (1300 ± 100 ppm CO\(_2\)); *Baiera* sp. 1 (lower Sinemurian, 1624.9 m – 1500 ± 100 ppm CO\(_2\)) and *Baiera* sp. 4 (upper Pliensbachian, 1276 m – 1200 ± 150 ppm CO\(_2\)). The varied measurement error depends on the size and preservation quality of the measured cuticle surfaces. The difference between the two last samples shows a marked decrease of pCO\(_2\) from a high value in the early Sinemurian to much lower value in the Late Pliensbachian (Fig. 7). For the Hettangian, our results fit the other results obtained in other studies, where pCO\(_2\) varied from 1000 to 1500 ppm (Barbacka, 2011), following high 2000–2500 ppm at the T/J boundary (Steinhordsdottir et al., 2011). Similar values are given in the compilation of Foster et al. (2017), where CO\(_2\) remains above 1000 ppm for Hettangian and Sinemurian. For the Late Pliensbachian, the single obtained value of c. 1200 pCO\(_2\) from Kaszewy is the highest hitherto reported. For this age there were values given by Xie et al. (2006) of c. 940 ppm or c. 900 ppm, but these values were calculated from a different conifer species, which is discussed by Steinhordsdottir and Vajda (2015). More generally, the concentration of c. 960 ppm pCO\(_2\) for the Early Jurassic was interpreted also from Swedish (Beering et al., 1998), Danish (McElwain et al., 2005 – of note is a wide range between 500 and 2500 ppm for lower Toarcian), Chinese (c. 800 ppm – Chen et al., 2001) and Australian (900 ppm based on araucarian conifers) material (Steinhordsdottir and Vajda, 2015) - see Fig. 7 and discussion by Barbacka (2011). Generally, the differences in SI and pCO\(_2\) calculations are not outstanding: pCO\(_2\) values oscillate within certain range and average values are given. Nevertheless, some inaccuracies are possible, for example pCO\(_2\) can be also influenced by the Carboniferous standardization of NLE, with different authors making different assumptions. The SI and SD of the NLE are used as reference for counting these values for fossil plants. Calibration usually is made empirically in controlled values of CO\(_2\) concentration. Different authors have calibrated *Ginkgo biloba* under different conditions from 300 to 560 ppm which results in fluctuations of calculated pCO\(_2\) values ranging from c. 960 to c. 1500 ppm respectively, which is discussed by Barbacka (2011). Such deviations should appear if differently calibrated NLE are used in the pCO\(_2\) calculation. The results from the Late Hettangian of Kaszewy, calculated from *Baiera* sp. 2, fit within general climatic trends given by Kürschner (2001). We observe that *Czekanowskia* and *Baiera* stomatal density (sensitive to pCO\(_2\) and consequently temperature differences) is generally compatible with \( \delta^{13}\text{C}_{\text{wood}} \) while these genera are not themselves specifically related to a strict range of temperature. Their deciduous character is connected with seasonal temperature and precipitation variations (Fig. 7). Both genera commonly occur together among dominant elements of Siberian-Canadian provinces and are known as mesotemperate plants, characteristic of flood plain/delta peat-forming assemblages (Rees et al., 2000). *Czekanowskia* is one of the commonest plants of coal bearing deposits of Siberia, forming wetland forests (Krassilov, 2003). Commonly regarded as an indicator of temperate climate, in Poland it was found (apart from Kaszewy) in the northern margin of the Holy Cross Mountains (Barbacka et al., 2014) together with other Siberian element *Pseudotorellia* Florin (Kürschner, 2001). The genus *Baiera* in Poland has been reported from the Holy Cross Mountains and the Lublin Coal Basin (Barbacka et al., 2014). In Mesozoic times, *Baiera*, together with other ginkgophytes of similar type (*Ginkgoites, Sphenobaiera*) were widespread and were connected with rather stable, warm to temperate climate and humid conditions (Barbacka, 2011; Pacyna, 2013; Barbacka et al., 2014, 2015, 2016; Zhou, 2009).

### 5. Carbon cycle and temperature interpretation

The interpretation of the relationship between \( \delta^{13}\text{C}_{\text{wood}} \) and TOC\(_{\text{cont}} \) must rely on some baseline assumptions. The carbon cycle includes many simultaneous processes with different time scales, involving very complex processes of carbon mass and isotopic fluxes to and out of the ocean–atmosphere system. Concerning the carbon cycle, in particular the fluxes into the atmospheric system, we focus here on processes registered in plant matter (Gröcke, 2002). Experimental data from plant growth chambers obtained by Schubert and Jahren (2012), further developed by Cui and Schubert (2016) allowed formulation of a simple model (identified as the C3 proxy) in which plant carbon-isotope composition depends on changes in only two atmospheric variables, i.e. source isotopic composition of carbon dioxide (\( \delta^{13}\text{C}_{\text{CO}_2} \)) and the pCO\(_2\). However, for \( \delta^{13}\text{C} \) to be used as an accurate and precise method to reconstruct pCO\(_2\), the major requirement is to demonstrate that changes in pCO2 are the main driver of changes in \( \delta^{13}\text{C} \). It should be emphasized that only those experiments which are based on C3 plants can be taken in account for the current study, as C4 plants (absent in Triassic - Early Jurassic times) show different fractionation pattern (Diefendorf et al., 2010). Source isotopic composition of \( \delta^{13}\text{C}_{\text{CO}_2} \) during the Early Jurassic epoch is a complex issue, as carbon released to the atmosphere can have multiple different sources with different \( \delta^{13}\text{C}_{\text{CO}_2} \) values, anywhere from −70‰ (gas hydrate) through thermogenic methane (−50 to −7‰), oxic degradation of organic matter (−20 to −30‰) (Them III et al., 2017a) and volcanic CO\(_2\) (−6 ± 2‰) (Beering and Brettinall, 2007; Ruebsam et al., 2019, 2020; Gales et al., 2020). Thus, observed changes to \( \delta^{13}\text{C} \) of the atmosphere (on the influx side) can be caused by widely varying flux changes of carbon depending on the source, which would have widely varying effects on changes in pCO\(_2\) for the same change in \( \delta^{13}\text{C} \) - and resulting temperature (one should bear in mind that changes to \( \delta^{13}\text{C} \) can be caused by changes in carbon sink, such as decreasing organic carbon burial). However, negative carbon isotope excursions of an amplitude of about 3–4‰ observed in the latest Triassic and Early Jurassic can only be produced by emissions of \( 13\text{C} \)-depleted carbon such as gas hydrates, thermogenic methane or oxic organic matter degradation (Them III et al., 2017a; Pienkowski et al., 2016; Ruebsam et al., 2020) or a mixture of these sources (Pagani et al., 2006), because unrealistically high amounts of volcanogenic CO\(_2\) (released in short time – short carbon cycle) would be required to sufficiently affect the isotopic composition of the exchangeable carbon reservoir (Beering and Brettinall, 2007; Ruebsam et al., 2020). For example, Heimdal et al. (2020) postulate for the latest Rhaetian-earliest Hettangian case that thermogenic carbon generated from the contact aureoles around CAMP sills represented a credible source for the negative CIEs. Concerning short-term fluxes of carbon into the atmosphere-ocean system, volcanogenic emissions (usually prolonged over long periods of time) acted generally rather as a trigger for a cascade of surface carbon cycling processes, mobilizing methane and CO\(_2\) characterized by a more negative \( \delta^{13}\text{C}_{\text{CO}_2} \) signature. Thus, fluxes from the shallow sedimentary sources were a major contributor to the growth of pCO\(_2\) with negative \( 13\text{C} \) signature (e.g. Hesselbo et al., 2000; Them III et al., 2017a; Ruebsam et al., 2020). Additionally, increasing pCO\(_2\) causes \( \delta^{13}\text{C} \) negative fractionation by C3 plants and both caused the
negative isotopic signal registered in fossil plants (Hare et al., 2018). Similarly, enhanced decomposition of continental carbon pool caused by rising temperature has a short-term, strong feedback effect (Pierikowski et al., 2016). The rising pCO₂ was a major factor in raising temperature, and temperature growth led to depletion of TOCcont.

On the other hand, increasing pCO₂ climate warming and enhanced hydrological cycle (causing enhanced delivery of nutrients to the oceans) lead to increased ocean productivity and increased rate of burial of organic carbon, resulting in drawdown and fall in atmospheric pCO₂, and positive excursions in both organic and inorganic carbon (Kump and Arthur, 1999). The drawdown of atmospheric CO₂ led to global cooling and higher accumulation of TOCcont.

Significant, moderate or strong (Evans, 1996) correlation between δ¹³Cwood and TOCcont \( r = 0.53–0.59; R_s = 0.61; \) Fig. 5 \) could likely develop in response to the global carbon cycle (cf. Gröcke et al., 1999). While the fluxes of CO₂ with negative δ¹³C signature led to rising pCO₂, negative C isotopic signal registered in fossil plants and global warming, massive burial of δ¹³C-depleted organic carbon at a global scale resulted in higher δ¹³C values in all carbon cycle reservoirs, and the attendant drawdown of atmospheric CO₂ led to global cooling. As noted by Pierikowski et al. (2016), the variability in TOCcont contents in the Polish late Pliensbachian to early Toarcian successions is interpreted primarily as a reflection of the efficiency of terrestrial biodegradation, which depended particularly on temperature, creating favourable conditions for fungi (Fig. 6) that efficiently degraded the most common and otherwise resistant to degradation land palynomaceral, i.e. wood, composed mainly of lignin (López-Mondevjar et al., 2018). Increased fungal activity was noted in other periods characterized by elevated temperature, such as Palaeocene-Eocene thermal maximum (PETM) – Kender et al. (2012). In contrast, lower temperatures slow down fungal-mediated decomposition rates (Feng et al., 2008), enhancing terrestrial carbon accumulation. Similar processes connected with accumulation and destruction of soil organic matter are observed in recent experiments involving soil warming leading to decomposition of lignin in soil (Feng et al., 2008). It should be noted that bacteria can play some role in wood degradation as well (Robert Blanchette et al., 1990). However, bacterial degradation occurs at a slow rate compared with fungal decay and detritivore fungi are singled out as the only organisms which were able to degrade lignin tissues (Richardson et al., 2012). The current study extends the δ¹³Cwood and TOCcont dataset for Poland to the whole Lower Jurassic series and the late Raetian (Figs. 2, 5, 7). If it is assumed that the temperature is the main factor of TOCcont reduction, then we can link the observed relationship between δ¹³Cwood and TOCcont to the temperature warming-cooling trends, associated with global carbon cycle. An important point is that this correlation does not demonstrate δ¹³Cwood – temperature direct causation. In this case, the δ¹³Cwood, temperature and TOCcont would be responding to another forcing (CO₂ fluxes and Corg burial, i.e. carbon cycle – cf. Gröcke et al., 1999). Additionally, possible influence of sedimentary and diagenetic factors, reflected also as palynofacies inversions (Pierikowski and Waksundzka, 2009) could alter the TOCcont content.

On the other hand, extremely enhanced hydrological cycle at the peaks of T-OAE could have had some impact on the observed δ¹³Cwood – TOCcont relation, in that organic matter would have been rapidly removed and delivered to the receiving basin before the decomposition processes had fully taken effect. This could be responsible in slightly reversed δ¹³Cwood – TOCcont trend observed in four Toarcian samples with most negative δ¹³Cwood values (Fig. 5).

Weaker correlation in the entirely continental Raetian section can be explained by much stronger impact of local environmental factors, influencing strongly differentiated TOCcont sequestration that is facies specific in space and time, for example fluvial plain/lacustrine shifts of environments. In contrast, most of the Lower Jurassic strata were deposited in marginal-marine/deltaic environments, where the TOCcont content was averaged in sedimentary processes. Thus, results obtained from these settings show fewer outliers.

Inferred latest Triassic-Early Jurassic temperature trends for the c. 40°N paleolatitude are shown in stratigraphical order in the Fig. 7. As shown by floral remains, the temperatures were generally temperate to warm, with a slight decreasing trend over time, punctuated by two marked hotter periods – in the Late Raetian and in particular in the early Toarcian, during the Toarcian Oceanic Anoxic Event Carbon Isotope Excursion (T-OAE CIE), when temperatures peaked. Several less marked periods of temperature rise were indicated in the Hettangian (plantarbis, lissicus), Sinemurian (turnet – confirmed also by Schöllhorn et al., 2020 and Hesselbo et al., 2020a), oxynotum (see also Riding et al., 2013) and Pliensbachian (davoei); the latter reported by several authors, e.g. Dera et al. (2011); Gomez et al. (2015); Price et al. (2016); Peti and Thibault (2017). Lower temperatures occurred in the latest Raetian/earliest Hettangian, late Hettangian (periodically), early Pliensbachian (jamesoni-brevispina) and particularly in the Late Pliensbachian (stokesi, gibbosus, spinatum), which seems to be the coolest period in the Early Jurassic, confirmed in many papers (Price, 1999; Korte and Hesselbo, 2011; Dera et al., 2009, 2011). This cooler interval was punctuated by a short-lived warming episode at the margaritatus/spinatum boundary. Of note are relatively frequent inferred temperature shifts in the late Raetian and the late Hettangian, which can be attributed to the methane releases from clathrates or wetlands, triggered by increased decomposition of continental carbon pool caused by rising temperature.

Fig. 6. Four photographs from the lower Toarcian samples, showing wood strongly decomposed by fungi and fungal mycelia and spores: A – sample 1235.6 m: fungal mycelia; B – sample 1201.1 m: fungal mycelia and spores; C – sample 1201.1 m: fungal mycelia and spores; D – sample 1246.9 m: segmented fungal mycelia; E – sample 1201.1 m: fungal spore. Scale for A–E = 20 μ.
by peaks of volcanic activity of the Central Atlantic Magmatic Province (Pálfy et al., 2001; Hesselbo et al., 2002; Ruhl and Kürschner, 2011; Blackburn et al., 2013) or by thermogenic carbon generated from the contact aureoles around CAMP sills (e.g., Heimdal et al., 2020). Rapid climate shifts (inferred from clay mineral composition) in the late Rhaetian and possibly earliest Hettangian with marked cool and dry episodes was postulated as one of the possible scenarios of the end-Triassic mass extinction on continents (Pieńkowski et al., 2014). Similar contrasting temperature shifts are observed in the early Toarcian, when the T-OAE CIE and coeval greenhouse period might have been
The δ13Cwood of plant matter in the geological record would be generally dependent on the δ13C of the palaeoeatmosphere (Hasegawa, 1997; Hesselbo and Pieńkowski, 2011; Schubert and Jahren, 2012; Cui and Schubert, 2016; Pieńkowski et al., 2016; Them III et al., 2017b; Hare et al., 2018; Ruebsam et al., 2020). Moreover, a parallel stratigraphic evolution between terrestrial and marine organic matter-δ13C (e.g. Hesselbo and Pieńkowski, 2011; Them III et al., 2017b; Storm et al., 2020) strengthens the hypothesis that changes in the carbon isotope curve are a genuine feature reflecting the global carbon cycle and change in δ13C of atmospheric CO2. From an ecophysiological standpoint changes in δ13C are also linked to changes in water use efficiency of the plants, therefore Lomax et al. (2019) regarded that the model of Schubert and Jahren (2012) and Cui and Schubert (2016) is only operable over a limited climate space. Indeed, mean annual precipitation (MAP) can affect carbon isotope fractionation in plant materials, whereby δ13C values are negatively correlated with MAP in extant and fossil C3 plant (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018). However, differences in fractionation caused by MAP are significant (over 2‰) only when the most contrasting plant habitats are compared - dry (< 600 mm/year) and very humid (> 2000 mm/year) MAP (Diefendorf et al., 2010; Kohn, 2010). Moreover, Hare et al. (2018) showed that the contributions from changing MAP are insignificant when compared to the effect of pCO2 for C3 plants. Influence of mean annual precipitation (MAP) within a relatively stable climatic belt generally fall into the ‘complacent’ field, suggesting little interannual perturbation and mostly uniform growing conditions, with weak to moderate seasonality. Our material comes from the same latitudinal zone of changing humidity, but not drastically enough to cause any significant δ13C fractionation in C3 plants (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018).

Humidity-related fractionation effects could thus only slightly modulate the observed C isotopes trends (for example during T-OAE, when floristic and palynological proxies indicate increased humidity; Pieńkowski et al., 2016), rather than being the significant factor, or change systematically in conjunction with pCO2, thus indirectly amplifying rather than destroying the observed relationship. It should be noted that based on spore-pollen assemblages from marine strata of Yorkshire, UK (Morgans, 1999) generally fall into the ‘complacent’ field, suggesting little interannual perturbation and mostly uniform growing conditions, with weak to moderate seasonality. Our material comes from the same latitudinal zone of changing humidity, but not drastically enough to cause any significant δ13C fractionation in C3 plants (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018).

6. Variables influencing δ13Cwood and TOCcont - discussion

The δ13Cwood of plant matter in the geological record would be generally dependent on the δ13C of the palaeoeatmosphere (Hasegawa, 1997; Hesselbo and Pieńkowski, 2011; Schubert and Jahren, 2012; Cui and Schubert, 2016; Pieńkowski et al., 2016; Them III et al., 2017b; Storm et al., 2020) strengthens the hypothesis that changes in the carbon isotope curve are a genuine feature reflecting the global carbon cycle and change in δ13C of atmospheric CO2. From an ecophysiological standpoint changes in δ13C are also linked to changes in water use efficiency of the plants, therefore Lomax et al. (2019) regarded that the model of Schubert and Jahren (2012) and Cui and Schubert (2016) is only operable over a limited climate space. Indeed, mean annual precipitation (MAP) can affect carbon isotope fractionation in plant materials, whereby δ13C values are negatively correlated with MAP in extant and fossil C3 plant (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018). However, differences in fractionation caused by MAP are significant (over 2‰) only when the most contrasting plant habitats are compared - dry (< 600 mm/year) and very humid (> 2000 mm/year) MAP (Diefendorf et al., 2010; Kohn, 2010). Moreover, Hare et al. (2018) showed that the contributions from changing MAP are insignificant when compared to the effect of pCO2 for C3 plants. Influence of mean annual precipitation (MAP) within a relatively stable climatic belt generally fall into the ‘complacent’ field, suggesting little interannual perturbation and mostly uniform growing conditions, with weak to moderate seasonality. Our material comes from the same latitudinal zone of changing humidity, but not drastically enough to cause any significant δ13C fractionation in C3 plants (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018).

Humidity-related fractionation effects could thus only slightly modulate the observed C isotopes trends (for example during T-OAE, when floristic and palynological proxies indicate increased humidity; Pieńkowski et al., 2016), rather than being the significant factor, or change systematically in conjunction with pCO2, thus indirectly amplifying rather than destroying the observed relationship. It should be noted that based on spore-pollen assemblages from marine strata of Yorkshire, UK (Morgans, 1999) generally fall into the ‘complacent’ field, suggesting little interannual perturbation and mostly uniform growing conditions, with weak to moderate seasonality. Our material comes from the same latitudinal zone of changing humidity, but not drastically enough to cause any significant δ13C fractionation in C3 plants (Diefendorf et al., 2010; Kohn, 2010; Hare et al., 2018).
material by taphonomic processes. In our opinion, Slater et al. (2019) interpretation of extremely wet/dry seasons should be relaxed in favor of more humid conditions enabling the survival of extremely hydrophilic plants.

Plant chamber experiments have also revealed relationships between carbon isotope discrimination and changing pO2 (Porter et al., 2017), but this variable in the geological record is interpreted from reconstructions which vary widely, particularly for the Mesozoic and early Cenozoic eras (Glasspool and Scott, 2010). In respect to the geological time interval studied herein, these low-resolution models are often controversial, although they confirm a general rule that high rates of organic carbon burial results in subsequent oxygen production (Krause et al., 2018).

The described herein δ13Cwood/TOCcont relationship is surprising. However, if the correlation is so significant, it follows that there is a natural reason for it. There is also a relationship between terrestrial TOC and temperature from actualistic experimental work. As described above, decomposition of soil labile carbon is highly sensitive to temperature variation and elevated pCO2 resulting in higher temperature could be conducive for enhanced soil organic matter decomposition (Fang et al., 2005; Feng et al., 2008; Pieńkowski et al., 2016; López-Mondéjar et al., 2018). The continental kerogen studied herein had likely been oxidized on land or in rivers and before delivery to the receiving basin – remineralization of TOCcont in a marginal-marine basin was possible, but among different continental types of organic matter, the wood and charcoal was least affected by degradation in the basin (Pieńkowski and Waksmundzka, 2009). Even the most critical approaches (Davidson and Janssens, 2006) admit that despite controversies, the observational data are converging to demonstrate that irrespective of labile or recalcitrant character, the soil carbon pool decomposes with apparent detectable temperature sensitivity (Fang et al., 2005; Feng et al., 2008). Wood is known to react more to higher temperature changes (kinetic theory), which can explain why the population of samples with more diversified δ13Cwood and TOCcont values (e.g. Toarcian) shows relatively higher coefficient in δ13Cwood/TOCcont function (Fig. 5).

Considering the above arguments, it should be noted that some emerging reports from Lower Jurassic marine sediments (Hesselbo et al., 2020b; Ullmann et al., 2020) demonstrate a lack of correlation between δ13Corg and δ18O (reflecting in general, although still debatably, sea-water temperature) in some sections, for example the lower Sinemurian in UK or the Toarcian in Iberia. However, Schöllhorn et al. (2020) show good convergence between these two variables in Sinemurian of Dorset (UK), while there are divergences in the lower Pliensbachian. Of note is also the fact that Hesselbo et al. (2020a, 2020b) point to divergence in the lower Sinemurian, while the Hettangian part shows rather good convergence between the two variables. Similarly, Price et al. (2016) compared δ13C and δ18O for the Sinemurian and Pliensbachian section in Dorset and there is a generally good convergence for most of the section, but in some parts of the profile one can observe some divergences as well. There is good convergence of δ13C and δ18O both in the Rhaetian-Hettangian sections in SW UK (Korte et al., 2009) and in the uppermost Pliensbachian – lower Toarcian section in the Peniche GSSP profile in Portugal (Suan et al., 2008; Fantasia et al., 2019). Nevertheless, existing δ13C and δ18O divergences tend to cast some doubts, at least on universality of δ13Corg and δ18O/temperature relations because one should expect even stronger relationship between these parameters in marine sediments than there is between δ13Cwood and marine temperature, as the relationship is more direct. However, there are some inherent uncertainties regarding δ13C and δ18O results from marine deposits and their interpretation. It is now well established that bulk organic C-isotope records need to be regarded with caution due to the mixing effects of different types of carbon, each with their own δ13C signature (Van de Schootbrugge et al., 2008; Suan et al., 2015; Schöllhorn et al., 2020). In Kaszewy these problems are avoided, because the δ13Cwood values comes from homogenous material. The other question is the possible influence of oceanographic processes, acting (at least partly) independently from global atmospheric pCO2 and temperature changes. Opening of the Hispanic corridor in Sinemurian and its widening in Pliensbachian (Porter et al., 2013) impacted oceanic circulation, marine faunal exchange pattern and, very probably, also isotopic and temperature pattern in the Jurassic seaways of the European area.

This research supports and extends similar results obtained from the late Pliensbachian-early Toarcian deposits of the Polish basin (Pieńkowski et al., 2016) and this is a first attempt to infer a continuous record of air temperature trends through such a long period of geological time. Independently obtained trends of interpreted palaeotemperatures are juxtaposed with data from stomatal index which are thought to represent changes in pCO2. It seems that the proposed relationship works as a temperature signal only for a warm/winter-wet climatic zone. The δ13Cwood/TOCcont trend observed in the semi-arid climate belt seems to be different, at least during the Late Pliensbachian-early Toarcian interval in the Southern Tethyan and Iberian margin (Rodrigues et al., 2019). It is possible that during the T-OAE the Southern Tethyan and Iberian margin was influenced by the adjacent tropical climatic zone, which caused the observed differences. Of note is also modeling of major carbon cycle perturbations around the Triassic-Jurassic boundary, showing coincidence between δ13C and pCO2 changes (Heimdal et al., 2020).

7. Conclusions

1. There is an observed empirical, highly significant relationship between δ13Cwood and TOCcont in the Rhaetian/Lower Jurassic from Kaszewy that can be defined by an equation. The premise is that in mid-latitudes, a controlling factor on the relation between δ13Cwood and preservation of continental total organic carbon (TOCcont) is the efficiency of terrestrial biodegradation which is pCO2 and temperature dependent, although there are several factors that control both δ13Cwood and TOCcont.

2. The δ13Cof the Rhaetian and Early Jurassic was largely controlled by δ13C-depleted fluxes in and out of the ocean-atmosphere system. Likely, there is no non-causal correlations between δ13Cwood/TOCcont and temperature trends. Such correlations may develop because massive fluxes into the ocean-atmosphere system and subsequent burial of 13C-depleted organic carbon (flux out of the ocean-atmosphere system), at a global scale results in global warming and cooling episodes, registered in the continental carbon pool as δ13Cwood and TOCcont fluctuations.

3. The relationship between δ13Cwood and TOCcont can hypothetically be useful, even if non-causal, for ~25 Myr long latest Rhaetian and Early Jurassic air temperature interpretation at c. 40°N paleolatitude. However, its utility depends on how strong the carbon cycle signal is, and how many other influences have operated on the system.

4. Independently obtained trends of interpreted palaeotemperatures were juxtaposed with data from stomatal index which are thought to represent changes in pCO2, and results obtained from stomatal index calculations are compatible with the interpreted temperature trends.

5. In our material, deposits of coastal/deltaic environments are most suitable for studying the relationship between carbon-isotopes, continental TOC, and temperature, because these facies contain more representative, averaged material delivered from a large catchment area – in contrast to alluvial plains where TOC is dependent on a more localized fluvial setting.

6. Unique, expanded and continuous cores from Kaszewy and Mochras allowed reliable δ13C chemostratigraphic correlation of marine and marginal/non-marine Lower Jurassic deposits.

7. While this study suggests an overall implication for Earth system studies and offers potentially independent means (registered in the continental carbon pool) to estimate latest Triassic to Early Jurassic
atmospheric temperatures, the relationship between $\text{TOC}_{\text{org}}$, $\delta^{13}$C wood and temperature should be further tested and treated as still hypothetical because in general organic carbon isotope records are potentially subject to many different influences.

Declaration of Competing Interest

The authors declare no conflict of interest.

Acknowledgements

We thank Przemysław Karcz and Marcin Janas for performing RockEval 6 pyrolytic analyses. We thank Dr Marta Hodbod and Dr Marta Wąskmundzka for palynocarval separation and identification of palynomorphs. Peter Ditchfield (Oxford) is thanked for C-isotope analysis. We are grateful to the PGE Company for granting access to the core and agreeing to take samples. We are grateful to Professor Thomas Algeo, guest editor of this volume, and two anonymous reviewers for constructive comments which improved the paper and to managing editors Professors Shane Schoepfer and Alessandra Negri for editorial handling. This paper is financed (G.P., M.B.) from resources of the National Science Centre (Poland), granted on the basis of decisions no. DEC-2012/06/M/ST1/00478, no. 2017/25/B/ST1/02235 and no. 2017/25/B/ST1/01273. This is a contribution to the ICDP and NERC project JET (grant number NE/N018508/1) and IGCP project 632 “Continental Crises of the Jurassic”.

Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2020.103383.

References

Baker, S.J., Hesselbo, S.P., Lenton, T.M., Duarte, L.V., Belcher, C.M., 2017. Charcoal evidence that rising atmospheric oxygen terminated Early Jurassic ocean anoxia. Nat. Commun. 8, 15018. https://doi.org/10.1038/ncomms15018.

Barbacka, M., 2011. Biodiversity and the reconstruction of Early Jurassic flora from the Meckes Mountains (S. Hungary). Acta Palaeobot. 51, 127–179.

Barbacka, M., Pacyna, G., Feldman Olszewska, A., Ziaja, J., Boder, E., 2014. Triassic–Jurassic flora of Poland; floristical support of climatic changes. Acta Geol. Pol. 64, 281–308.

Barbacka, M., Puspoki, Z., Boder, E., Forgazé, Z., Hámor-Vidi, M., Pacyna, G., McNichol, R.W., 2015. Palaeotopography related plant succession stages in a coal forming deltaic succession in early Jurassic in Hungary. Palaeogeogr. Palaeoclimatol. Palaeoecol. 440, 579–593.

Barbacka, M., Popa, M.E., Mída, J., Boder, E., Puspoki, Z., McNichol, R.W., 2016. A quantitative approach for identifying plant ecogroups in the Romanian Early Jurassic terrestrial vegetation. Palaeogeogr. Palaeocl. Palaeoecol. 446, 44–54.

Barth, G., Franz, M., Heinrich, C., Ernst, W., Zimmermann, J., Wolfram, M., 2018a. Marine and terrestrial sedimentation across the T–J transition in the North German Basin. Palaeogeogr. Palaeoclimatol. Palaeoecol. 489, 74–94.

Barth, G., Piekowski, G., Zimmermann, J., Franz, M., Kuhnlm, G., 2018b. Palaeogeographical evolution of the Lower Jurassic: high-resolution biostratigraphy and sequence stratigraphy in the Central European Basin. In: Kilhams, B., Kukla, P.A., Mazur, S., McKee, T., Mijnlieff, H.F., Van Oijk, Y. (Eds.), Mesozoic Resource Potential in the Southern Permian Basin. Geological Society, London, Sp. Publ., pp. 469. https://doi.org/10.1144/SP469.8.

Bartolini, A., Guex, J., Spangenberg, J.E., Schoene, B., Taylor, D.G., Schaltegger, U., Harazim, D., van de Schootbrugge, B., Sorichter, K., Fiebig, J., Duarte, L.V., Nullens, K.B., 2013. Zircon U–Pb geochronology links the end-Triassic extinction with the Central Atlantic Magmatic Province. Science 340, 941–945.

Beerling, D.J., 2009. Influence of the palaeocontinent and greenhouse effect on Hettangian clay mineral assemblages (Holy Cross Mts. area, Polish Basin). Geol. Quart. 53, 163–185.

Beerling, D.J., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.

Chen, L.Q., Cheng-Sen, Li, Chaloner, W.G., Beerling, D.J., Sun, Q.G., Collinson, M.E., Mitchell, P.L., 2001. Assessing the potential for the formatted characters of extant and fossil Ginkgo leaves to signal atmospheric CO2 change. Am. J. Bot. 88, 1309–1315.

Collinson, M.E., Royer, D.L., Lunt, D.J., 2007. Evidence of the role of atmospheric $\text{CO}_2$ in continental weathering. Geology 35, 231–234.

Dera, G., Pellenard, P., Pejčin, J.F., Pácuét, E., Domergue, M., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.

Dera, G., Pellenard, P., Pejčin, J.F., Pácuét, E., Domergue, M., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.

Dera, G., Pellenard, P., Pejčin, J.F., Pácuét, E., Domergue, M., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.

Dera, G., Pellenard, P., Pejčin, J.F., Pácuét, E., Domergue, M., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.

Dera, G., Pellenard, P., Pejčin, J.F., Pácuét, E., Domergue, M., 2012. The mineralogical record of the Early Toarcian stepwise climate changes and other environmental variations (Ciechocinek Formation, Polish Basin). Vol. Jurassic 10, 1–24.
