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An Early Jurassic (Sinemurian–Toarcian) stratigraphic framework for the occurrence of Organic Matter Preservation Intervals (OMPIs)

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ABSTRACT

Lower Jurassic sedimentary successions in the Atlantic margin basins include several organic-rich intervals, some with source rock potential; time-equivalent units are also identified in on- and offshore areas worldwide. Despite decades of research, it is still unclear which mechanisms lead to the deposition of organic-rich sediments during the Early Jurassic. The objectives of this study are to construct a detailed temporal and geographical framework of Sinemurian–Toarcian organic matter preservation intervals (OMPIs; subdivided into local, regional, and superregional) and roughly constrain the relationship of OMPIs with the Lower Jurassic $\delta^{13}C$ record. This survey combines an in-depth analysis of literature on the distribution of OM in the Sinemurian–Toarcian with new geochemical studies [total organic carbon (TOC) and organic matter pyrolysis] from selected outcrop sections from Portugal, Spain, and Morocco.

Strong local control on OMPIs during most of the Sinemurian is suggested. Regionally widespread organic-rich facies are associated with the most negative $\delta^{13}C$ values of the broad Sinemurian–Pliensbachian negative carbon isotopic trend recorded in organic matter (including the Sinemurian–Pliensbachian Boundary Event). Pliensbachian OMPIs are expressed in the areas bordering the proto-Atlantic Ocean and are often linked with positive $\delta^{13}C$ excursions and short-lived warm intervals, but OMPIs are also defined for the Late Pliensbachian cool interval. Early Toarcian superregional OMPIs are associated with some of the most pronounced $\delta^{13}C$ excursions of the Mesozoic. Toarcian maximum TOC content occurs with the positive $\delta^{13}C$ (recovery) trend following the $\delta^{13}C$ negative shift typically linked with the Early Toarcian Oceanic Anoxic Event (T-OAE), supporting the notion that peak carbon sequestration/ocean anoxia post-dated the main phase of carbon input into the atmosphere, as also suggested by recent modelling efforts. However, additional superregional OMPIs predate and postdate the T-OAE, indicating that conditions favouring preservation of OM (increased productivity and/or enhanced preservation) during the Early Toarcian were not restricted to the T-OAE interval.

The compilation of Sinemurian–Toarcian OMPIs presented in this paper demonstrates that organic-rich intervals of regional and superregional expression in the Lower Jurassic sedimentary record are ubiquitous and may even be more numerous than in the Cretaceous. Considering the association of some of the Sinemurian, Pliensbachian, and Toarcian regional and superregional OMPIs with well-defined carbon isotopic excursions, it is here suggested that these hold the same relevance as the secondary OAEs of the Cretaceous, such as the Hauterivian OAE (Faraoni Event), Late Valanginian OAE (Weissert Event), and Late Aptian–Early Albian OAE (OAE 1b cluster).

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1. Introduction

The deposition and preservation of organic matter-rich sediments during the Early Jurassic has led to the formation of prolific hydrocarbon source-rocks in many locations worldwide (e.g. Powell, 1978; Hallam and Bradshaw, 1979; Fleet et al., 1987; Jenkyns, 1988; Kodina et al., 1988; Baudin et al., 1990; Baudin, 1995; Bessereau et al., 1995; Scotchman, 2001; van de Schootbrugge et al., 2005b; Duarte et al., 2010; Silva et al., 2011, 2017; Suan et al., 2011; Silva and Duarte, 2015; Ruhl et al., 2016; Gómez et al., 2016a; Xu et al., 2017a; Campana et al., 2017). These organic-rich sedimentary rocks are interpreted to be near age equivalent between sedimentary basins and often associated with significant perturbations to the global carbon cycle and changes in regional/global climates and Earth system processes (e.g. Jenkyns and Clayton, 1986, 1997; Jenkyns, 2010; Jenkyns, 1988; van Buchem et al., 1995; Hesselbo et al., 2000a, 2007; van de Schootbrugge et al., 2005a, 2013; Suan et al., 2008, 2010, 2011, 2015; Dera et al., 2010; Korte and Hesselbo, 2011; Silva et al., 2011, 2015, 2017; Gómez and Goy, 2011; Jenkyns and Weedon, 2013; Franceschi et al., 2014; Hermoso et al., 2014; Korte et al., 2015; Silva and Duarte, 2015; Percival et al., 2016; Ruhl et al., 2016; Gómez et al., 2016b; Them et al., 2017; Xu et al., 2017b; Xu et al., 2017a, 2018; Ruebsam et al., 2018, 2020; Danisch et al., 2019; Danise et al., 2019; Slater et al., 2019; Mercuzot et al., 2020; Schöllhorn et al., 2020b; Schöllhorn et al., 2020a; Storm et al., 2020) (Figs. 1 and 2).

Some of the most remarkable Early Jurassic intervals of enhanced preservation of organic matter (OM), such as the Early Toarcian Oceanic Anoxic Event (T-OAE), are being intensively investigated to constrain the mechanisms and feedback processes controlling oceanographic, biological, and climatic changes during important past global environmental perturbations (e.g. Küspert, 1982; Jenkyns and Clayton, 1986; Jenkyns, 1988, 2010; Hesselbo et al., 2000a, 2007; Röhl et al., 2001; Jenkyns et al., 2002; Cohen et al., 2004; Röhl and Schmid-Röhl, 2005; Hermoso et al., 2012, 2013; Hermoso et al., 2009; Al-Suwaidi et al., 2010, 2016; Bodin et al., 2010; Littler et al., 2010; Bodin et al., 2016; Suan et al., 2010, 2011, 2015; Lézin et al., 2013; Sabatino et al., 2013; Kafousia et al., 2014; Kemp and Izumi, 2014; Reolid et al., 2014; Caruthers et al., 2014; Pieńkowski et al., 2016; Martinez et al., 2017; Silva et al., 2017; Them et al., 2017; Xu et al., 2017b, 2018; Ruvalcaba Baroni et al., 2018; Fantasia et al., 2018, 2019b; Ruebsam et al., 2019; Danise et al., 2019; Remírez and Algeo, 2020; Schöllhorn et al., 2020a; Storm et al., 2020; Rodrigues et al., 2021). However, there is increasing evidence that organic carbon burial and preservation, even when associated with widespread carbon sequestration events such as the T-OAE, were modulated by local geography, bathymetry, climate, ocean chemistry, nutrient supply, biological communities and, consequently, the local depositional environment (e.g. Duarte et al., 2010, 2014; Silva et

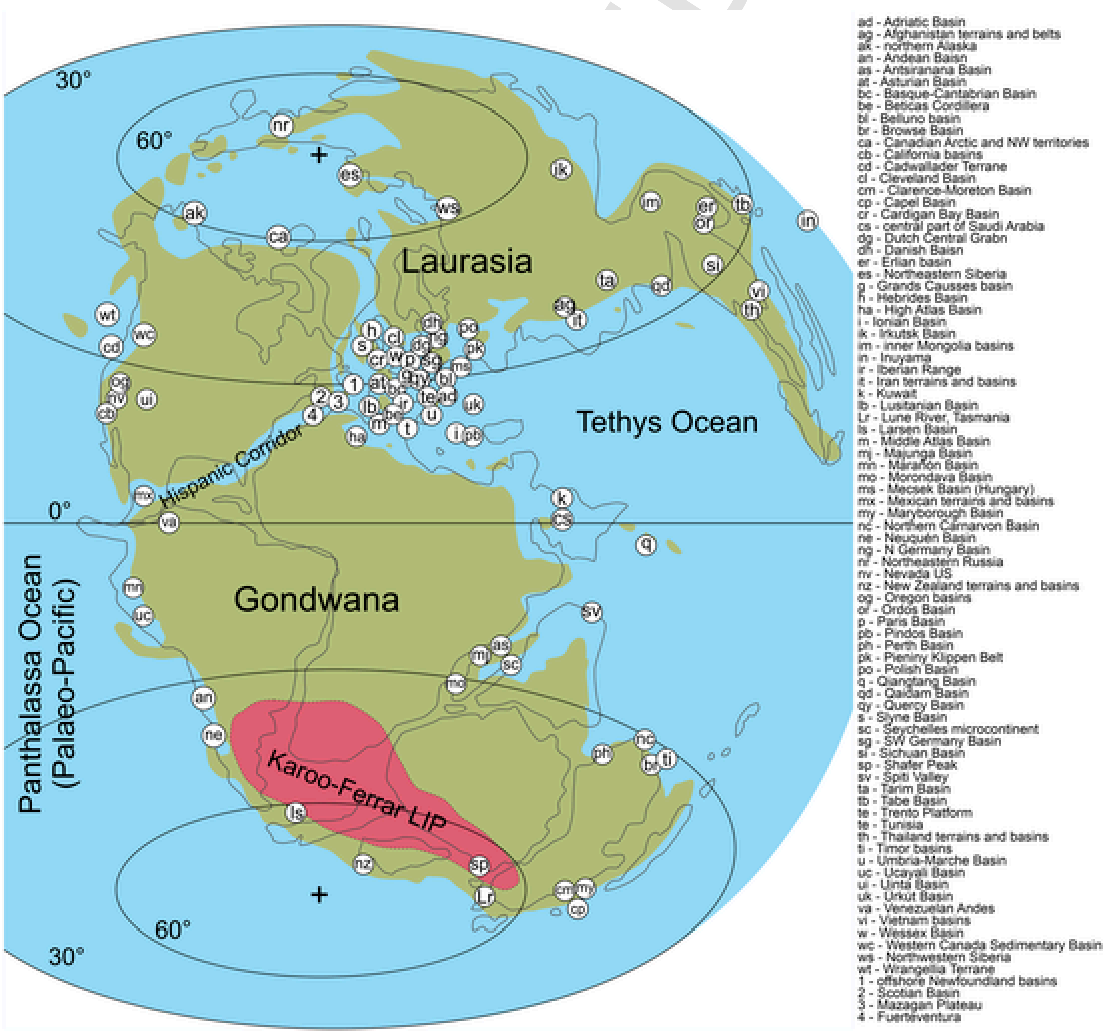


Fig. 1. Base map with Early Jurassic continent positions and locations of basins and localities with known Lower Jurassic sediments (modified from Damborenea (2002) and references therein). The West Netherlands Basin, Southern Beaujolais, Sub-Briançonnais Basin, Northern Calcareous Alps, and Midlands shelf are not represented here.

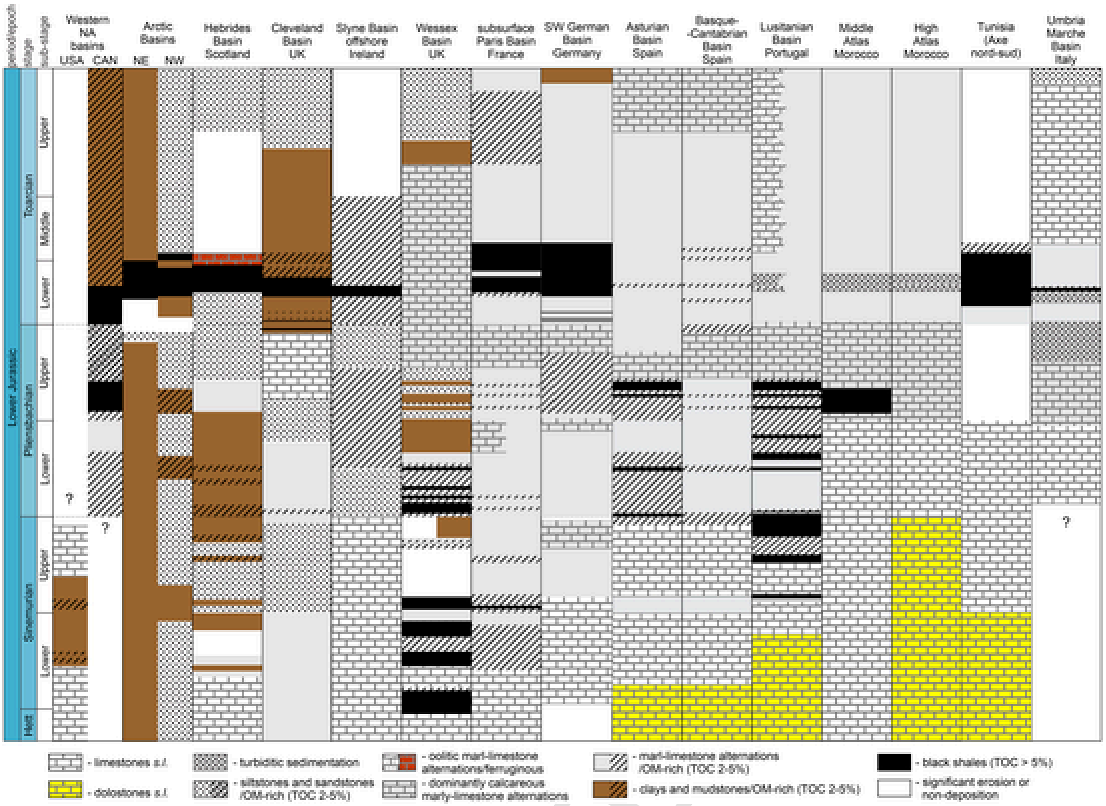


Fig. 2. General temporal and paleogeographical trends of Lower Jurassic organic-rich sedimentary successions. *Western NA basins:* Stratigraphy - (Asgar-Deen et al., 2003); Organic-rich facies - (Caruthers et al., 2011; Porter et al., 2014b; Porter et al., 2014a; Them et al., 2017); *Arctic Basins:* Stratigraphy - (Shurygin et al., 2011); Organic-rich facies – NE (Suan et al., 2011); NW (Kontorovich et al., 1997); *Hebrides Basin (Scotland):* Stratigraphy - (Hesselbo et al., 1998; Hesselbo and Coe, 2000; Morton, 2004); Organic-rich facies – (Thrasher, 1992; Trueblood, 1992; Scotchman, 2001); *Cleveland Basin (UK):* Stratigraphy - (Hesselbo et al., 2000b; Simms et al., 2004; Korte and Hesselbo, 2011); Organic-rich facies – (Jenkyns and Clayton, 1997; Scotchman, 2001; Kemp et al., 2005; Wignall et al., 2005; McArthur et al., 2008; Littler et al., 2010; Caswell and Coe, 2012; Salem, 2013; Porter et al., 2014a; Song et al., 2015); *Slyne Basin (offshore Ireland):* Stratigraphy - (Dancer et al., 2005); Organic-rich facies – (Trueblood, 1992; Scotchman, 2001; Silva et al., 2017); *Wessex Basin (UK)*: Stratigraphy - (Simms, 2004; Gallois, 2008; Jenkyns and Weedon, 2013); Organic-rich facies - (Jenkyns and Clayton, 1997; Weedon and Jenkyns, 1999; Scotchman, 2001; Deconinck et al., 2003; Ruhl et al., 2010; Jenkyns and Weedon, 2013; Price et al., 2016; Xu et al., 2017a); *Paris Basin (France)*: Stratigraphy - (Mouterde et al., 1980); Organic-rich facies - (Bessereau et al., 1995; van Breugel et al., 2006; Hermoso et al., 2012, 2013, 2014; Lézin et al., 2013; Song et al., 2014); *SW Germany*: Stratigraphy - (Menning and Hendrich, 2002; Pieńkowski et al., 2008; Feist-Burkhardt and Pross, 2010); Organic-rich facies - (Röhl et al., 2001; Schmid-Röhl et al., 2002; Röhl and Schmid-Röhl, 2005); data in (McArthur et al., 2008); (Song et al., 2015); *Asturian Basin (Spain)*: Stratigraphy - (Aurell et al., 2003; Comas-Rengifo and Goy, 2010); Organic-rich facies – (Borrego et al., 1996; Gómez et al., 2008, 2016a; Gómez and Goy, 2011); this study; *Basque-Cantabrian Basin (Spain)*: Stratigraphy - (Aurell et al., 2003); Organic-rich facies - (Quesada et al., 1997, 2005; Gómez and Goy, 2011); *Lusitanian Basin (Portugal);* Stratigraphy - (Duarte and Soares, 2002; Duarte et al., 2010, 2014; Silva et al., 2011, 2015); Organic-rich facies - (de Oliveira et al., 2006; Duarte et al., 2010, 2012; Silva et al., 2011, 2012; Poças Ribeiro et al., 2013; Silva, 2015; Silva and Duarte, 2015); this study; *Middle Atlas (Morocco)* Stratigraphy - (Sadki et al., 2014); Organic-rich facies - (Rachidi et al., 2009; Sachse et al., 2012); this study; *High Atlas (Morocco)* Stratigraphy - (Sadki et al., 2014); Organic-rich facies - (Bodin et al., 2010, 2011, 2016); this study; *Tunisia:* Stratigraphy - (Soussi, 2003); Organic-rich facies - (Soua, 2014); *Umbria Marche Basin (Italy):* Stratigraphy - (Monaco et al., 1994; Parisi et al., 1996; Rodríguez-Tovar et al., 2016); Organic-rich facies – (Monaco et al., 1994; Parisi et al., 1996; Bucefalo Palliani et al., 1998; Pancost et al., 2004; Sabatino et al., 2009).

al., 2012, 2013, 2015, 2017, 2021; Reolid et al., 2014; Suan et al., 2015; Rodrigues et al., 2016, 2019, 2020b; Bruneau et al., 2017; Them et al., 2017; Martindale et al., 2017; Rodríguez-Tovar et al., 2017; Ruvalcaba Baroni et al., 2018; Danise et al., 2019) (Fig. 2).

With over 40 years of research on the understanding of the processes leading to Early Jurassic OM enrichment in marine and terrestrial environments, there are still large uncertainties concerning the impact of local versus global factors leading to OM enrichment. Uncertainty arises from the (1) temporal and spatial ambiguity when correlating most Lower Jurassic organic-rich sedimentary successions, (2) limited constraints on the amount, chemical composition, and thermal maturity of OM, and (3) incomplete understanding of environmental and depositional constraints on OM production and preservation at a wide range of geological time intervals. Understanding the link between OM production and preservation/sequestration and climate change is vital to advancing knowledge of processes and feedback mechanisms in the Earth Systems that drive global change at a wide range of time scales and to constrain the relative importance of the many factors (including natural and anthropogenic) contributing to modern-day climate change.

This study focuses on the temporal and spatial occurrence of organic-rich intervals through the Sinemurian–Toarcian interval, comprising most of the Lower Jurassic. The two objectives of this paper are to (1) construct a detailed temporal and geographical framework of Sinemurian–Toarcian organic-rich facies occurrences at a global scale, defined here as organic matter preservation intervals (OMPIs) and (2) approximately constrain the relationship of OMPIs with the Lower Jurassic $\delta^{13}C$ record. This study combines an in-depth analysis of literature on the distribution and characteristics of OM (quantity of carbon and OM type) in the Sinemurian–Toarcian with new geochemical studies from selected outcrop sections from Portugal, Spain, and Morocco.

1.1. Production and preservation of organic matter and its impact on the global carbon cycle

For convenience, the carbon cycle is usually divided into two cycles, the short- and long-term carbon cycles (Berner, 1999). Carbon in the short-term carbon cycle is rapidly exchanged within the surficial reservoirs, consisting of the oceans, atmosphere, biosphere, and soil, whereas carbon in the long-term cycle is slowly exchanged between the geosphere and the surficial systems (Berner, 1999; Ciais et al., 2013). The short-term carbon cycle is the dominant control on atmospheric pCO_2 over millennia, whereas the long-term carbon cycle controls atmospheric CO_2 and O_2 over millions of years (Berner, 1999, 2006).

Equilibrium and kinetic fractionation among OM, CO_2, and HCO_3^- control the carbon isotope ratios of atmospheric CO_2 and oceanic inorganic and organic carbon (e.g. Emerson and Hedges, 2008; Silva et al., 2020a). The main processes that affect the $^{13}C/^{12}C$ ratio of the "superficial" carbon reservoirs at geological time intervals are the (1) equilibrium between dissolved inorganic carbon (DIC) and atmospheric CO_2, (2) fractionation between DIC and carbonate minerals, (3) fixation of CO_2 and production of biomass via photosynthesis, and (4) respiration, remineralisation, preservation, and sedimentary reworking and resuspension of sedimentary OM (Hayes et al., 1999).

There is little isotopic discrimination effect during respiration. However, through photosynthesis in the superficial layers of the ocean, ^{12}C is taken up preferentially over ^{13}C, leaving the superficial DIC pool enriched in ^{13}C. As OM sinks into intermediate and deep waters, "sinking" organic carbon is remineralised, and the deeper ocean DIC pool becomes enriched in ^{12}C. At geological time scales, removal of isotopically "light" OM into the sedimentary OM reservoir generates a relative enrichment in ^{13}C of dissolved CO_2 in seawater and continuation of this process leads to an overall increase in ^{13}C of marine carbonate, atmospheric CO_2, and OM (e.g. Jenkyns and Clayton, 1986; Wefer and Berger, 1991; Marshall, 1992; Patterson and Walter, 1994; Mitchell et al., 1996; Weissert et al., 1998; Holmden et al., 1998; Kump and Arthur, 1999; Hesselbo et al., 2000b; Berner, 2003; Immenhauser et al., 2003; Swart and Eberli, 2005; Panchuk et al., 2006; Newton and Bottrell, 2007; Emerson and Hedges, 2008; Silva et al., 2011, 2020a).

Removal of isotopically light carbon from the oceans through OM sequestration into the geological archive is recorded as an increase in $\delta^{13}C$ of contemporaneous carbonates and marine and terrestrial OM (Berner, 2003). Positive $\delta^{13}C$ excursions are often interpreted to represent enhanced burial and sequestration of OM into the sedimentary realm (e.g. Schlanger and Jenkyns, 1976; Jenkyns, 1988, 2010; Arthur et al., 1990; Kump and Arthur, 1999; Silva et al., 2011, 2020a). In particular cases, such as the case of the T-OAE, the $\delta^{13}C$ record may respond in a more complex way if deposition of OM is accompanied by large-scale carbon input from internal or external sources with different isotopic composition (e.g. Küspert, 1982; Jenkyns and Clayton, 1986; Jenkyns, 1988, 2010; Hesselbo et al., 2000a, 2007; Pálfy and Smith, 2000; Röhl et al., 2001; Jenkyns et al., 2002; Cohen et al., 2004; Röhl and Schmid-Röhl, 2005; Hermoso et al., 2012, 2013; Hermoso et al., 2009; Al-Suwaidi et al., 2010, 2016; Bodin et al., 2010; Littler et al., 2010; Bodin et al., 2016; Suan et al., 2010, 2011, 2015; Lézin et al., 2013; Sabatino et al., 2013; Kafousia et al., 2014; Kemp and Izumi, 2014; Reolid et al., 2014; Caruthers et al., 2014; Silva and Duarte, 2015; Pieńkowski et al., 2016; Martinez et al., 2017; Silva et al., 2017, 2020a; Them et al., 2017; Xu et al., 2017b, 2018; Fantasia et al., 2018, 2019b; Rodrigues et al., 2019, 2020b; Ruebsam et al., 2019; Danise et al., 2019; Storm et al., 2020; Ullmann et al., 2020).

Two primary debates exist today regarding the processes and mechanisms leading to the occurrence and preservation of OM in sediments (Forsman and Hunt, 1958; Huc and Durand, 1974; Pedersen and Calvert, 1990; Largeau and Derenne, 1993; Tyson, 1995; van Buchem et al., 1995; Vandenbroucke and Largeau, 2007; Zonneveld et al., 2010; Jenkyns, 2010; Suárez-Ruiz et al., 2012; Silva et al., 2012, 2013, 2020a; Hemingway et al., 2019; Hülse et al., 2019). The first debate concerns the role of primary productivity and associated carbon flux from surface waters to the sediment-water interface (Pedersen and Calvert, 1990) vs marine anoxia and associated enhanced OM preservation (Demaison and Moore, 1980; Tyson, 1995) in controlling the stratigraphic and geographic occurrence of organic-rich facies; the second relates to the mechanisms leading to the preservation and transformation of OM at geological time scales: degradation/recondensation (Huc and Durand, 1974; Tissot and Welte, 1978, 1984); selective preservation (Philip and Calvin, 1976; Tegelaar et al., 1989; Largeau and Derenne, 1993; Silva et al., 2012); sorptive mineral protection (Bishop et al., 1992; Mayer, 1994; Hedges and Keil, 1995; Kennedy et al., 2002; Burdige, 2007; Kennedy and Wagner, 2011), OM sulfurization (Sinninghe Damsté et al., 1988; Adam et al., 1998), and OM-mineral interaction (van Buchem et al., 1995; Kennedy et al., 2002; Hemingway et al., 2019). It is widely acknowledged that anoxia and enhanced productivity can both drive the occurrence of organic-rich facies. However, organic-rich sedimentary rocks characterized by lipid-rich OM with elevated Hydrogen Index (HI) seems to require low availability of O_2 in bottom waters (e.g. Espitalié et al., 1977, 1985; Demaison and Moore, 1980; Tissot and Welte, 1984; Kohnen et al., 1990, 1992; Goodarzi et al., 1993; Schouten et al., 1994; Sinninghe Damsté et al., 1995; van Buchem et al., 1995; Rodger Harvey et al., 1995; Sun et al., 2002; Farrimond et al., 2003; Lee et al., 2004; Bowden et al., 2006; Zonneveld et al., 2010; Silva et al., 2012).

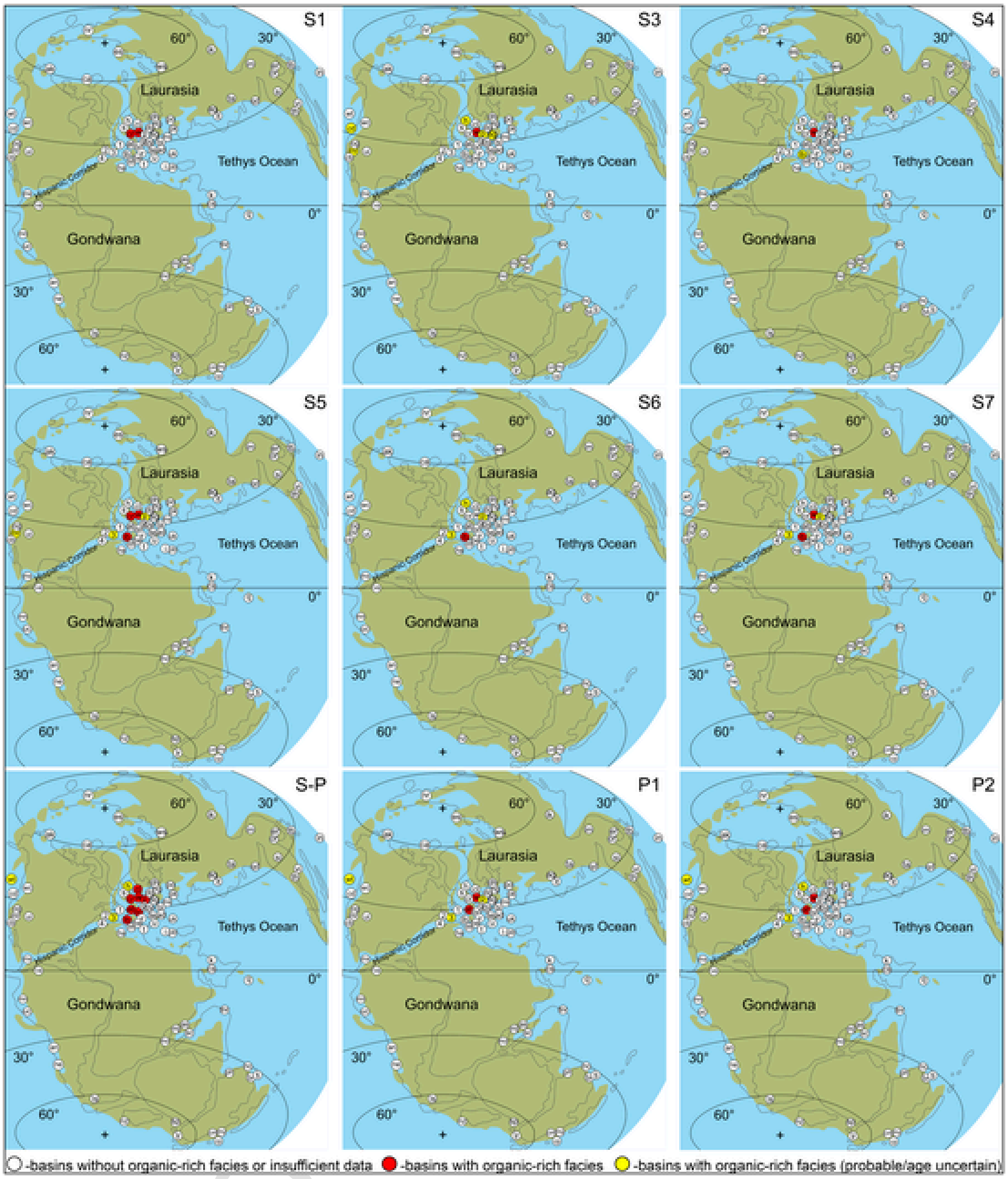
2. Methodology: a stratigraphic framework for Lower Jurassic organic-rich facies

The temporal and spatial occurrence of organic-rich sedimentary successions of Sinemurian–Toarcian age is reviewed by combining published organic geochemical data with new biostratigraphic and geochemical data (total organic carbon, TOC, and organic matter pyrolysis) from Lower Jurassic outcrops from Portugal, Spain, and Morocco (Appendixes 1–4). Some geographic areas are underrepresented in terms of data availability and warrant further studies to better constrain the extent and duration of Early Jurassic organic-rich deposition; a brief review of these locations is presented in Appendix 1 (A1.4. Uncertain data and data underrepresentation). A reference framework for the temporal and spatial occurrence of individual Early Jurassic *Organic Matter Preservation Intervals* (OMPIs) is presented in Figs. 3 and 4 and Appendixes 2–4.

2.1. Organic-rich sedimentary rocks

Organic-rich sedimentary rocks are here defined by present-day TOC values equal or higher than 2%. This cut-off value was chosen because:

a) Sedimentary rocks with TOC > 2% appear slightly darker compared to low-TOC rocks, making them more easily recognisable in the field or in hand samples (see also Hunt, 1995) and, therefore, from literature where geochemical data are not available and rocks are only described as organic-rich or black shales (there are, however, other factors that may also result in dark mudstones, for example, mineralogical composition). Black shale is a common classification in the literature and, in its broadest definition, corresponds to a dark coloured, usually laminated mudstone, calcareous mudstone, or marl with TOC ranging from 1 to 30 wt% (e.g. Swanson, 1961; Weissert, 1981).
b) This value roughly corresponds to the average TOC of "productive" hydrocarbon source rocks (see discussion in Ronov, 1958; Tissot and Welte, 1984; Jones, 1987; Lewan, 1987; Chinn, 1991; Katz, 1995; Law, 1999; Peters et al., 2005).



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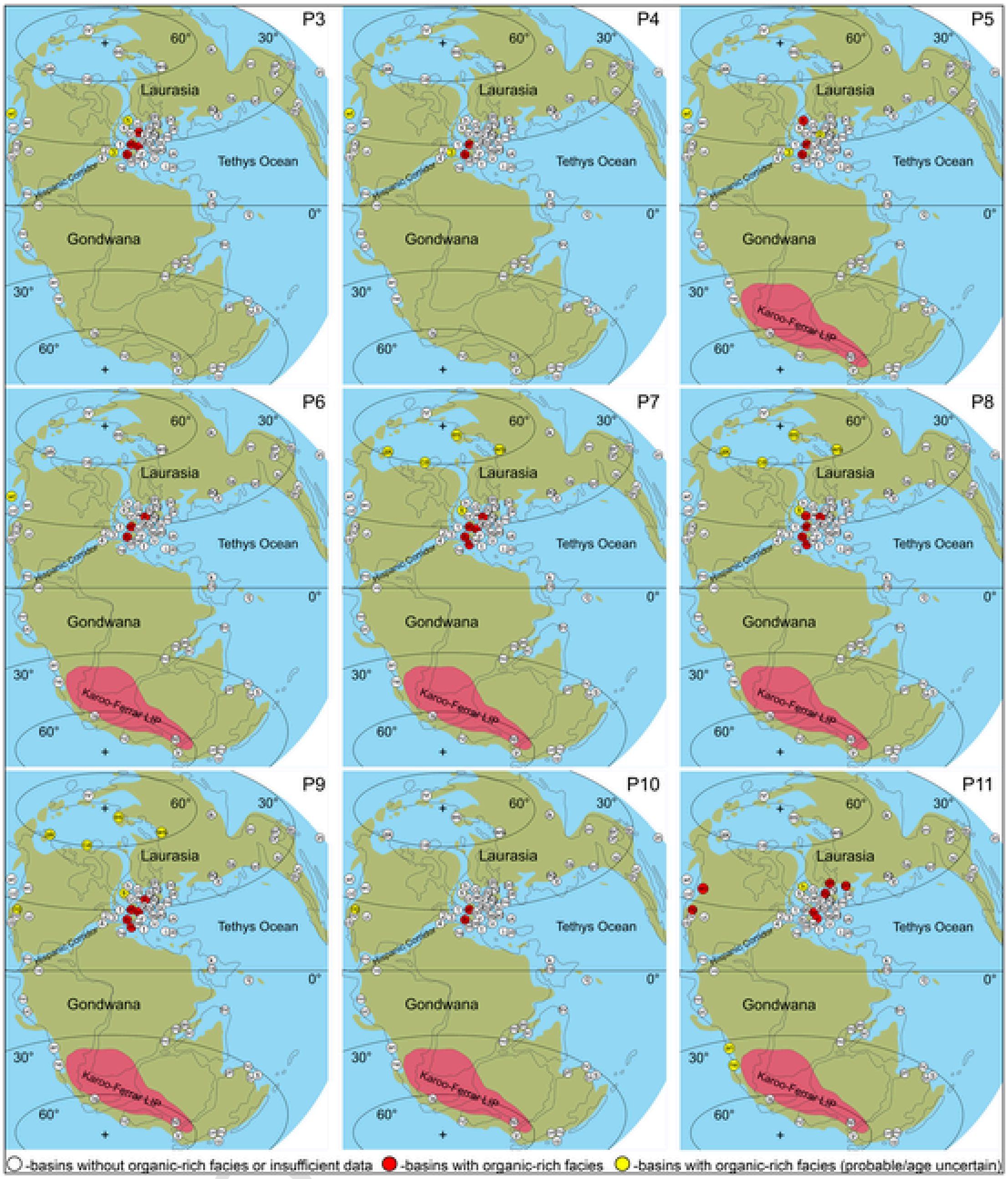

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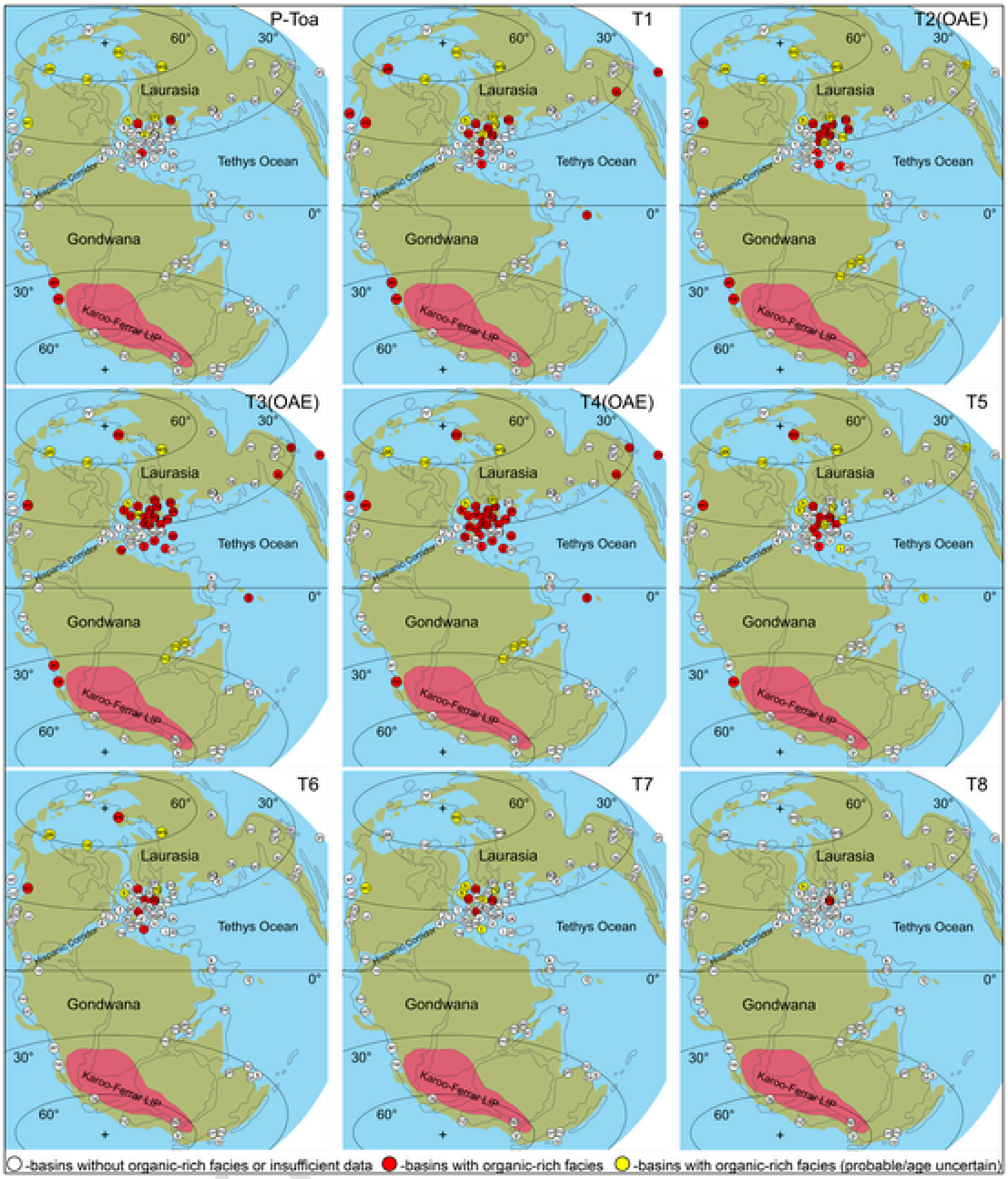

Fig. 3. Geographic distribution of basins with record of organic-rich facies per OMPI. See Fig. 1 for basin identification and Table 1 and Appendix 2 for OMPI identification.

2.2. *Organic Matter Preservation Interval (OMPI)*

An Organic Matter Preservation Interval (OMPI) corresponds to a constrained time interval (at the Subchron level) characterized by the deposition of organic-rich sediments (organic-rich sedimentary rocks with present day TOC > 2%) at local, regional, or superregional scales (as defined in subsection 2.1). OMPIs are defined irrespective of kerogen type and classification, location, and processes of OM production, accumulation, and diagenesis/maturation. They are classified as local, regional, or superregional (Table 1, Figs. 3 and 4 and Appendix 2–4):

a) Local OMPI: Organic-rich sedimentary rocks observed in one section.
b) Regional OMPI: Time-equivalent (at Subchronozone scale) organic-rich sedimentary rocks observed in two or more sections separated by more than ~1000 km (arbitrarily defined) or located in two “non-contiguous” basins.
c) Superregional OMPI: Time-equivalent organic-rich sedimentary rocks in two or more non-contiguous paleogeographical domains. Examples of paleogeographical domains are Northern Europe, Tethys, Mediterranean, Central Atlantic, Arctic, Pacific North America, Pacific South America, Asia.

OMPIs are described by first indicating the Age (Subage) where these are defined, followed by Chron, Subchron, and current understanding of the extent of the associated organic-rich facies, e.g. Early Pliensbachian (*davoei*, *maculatum/capricornus*) regional OMPI (see Table 1 and Appendix 2 for abbreviations for each OMPI). Uncertainty regarding the spatial representation of an OMPI is noted first by indicating the confirmed and then possible spatial classification, e.g., regional (superregional?). In particular cases, we adopt a historical denomina-

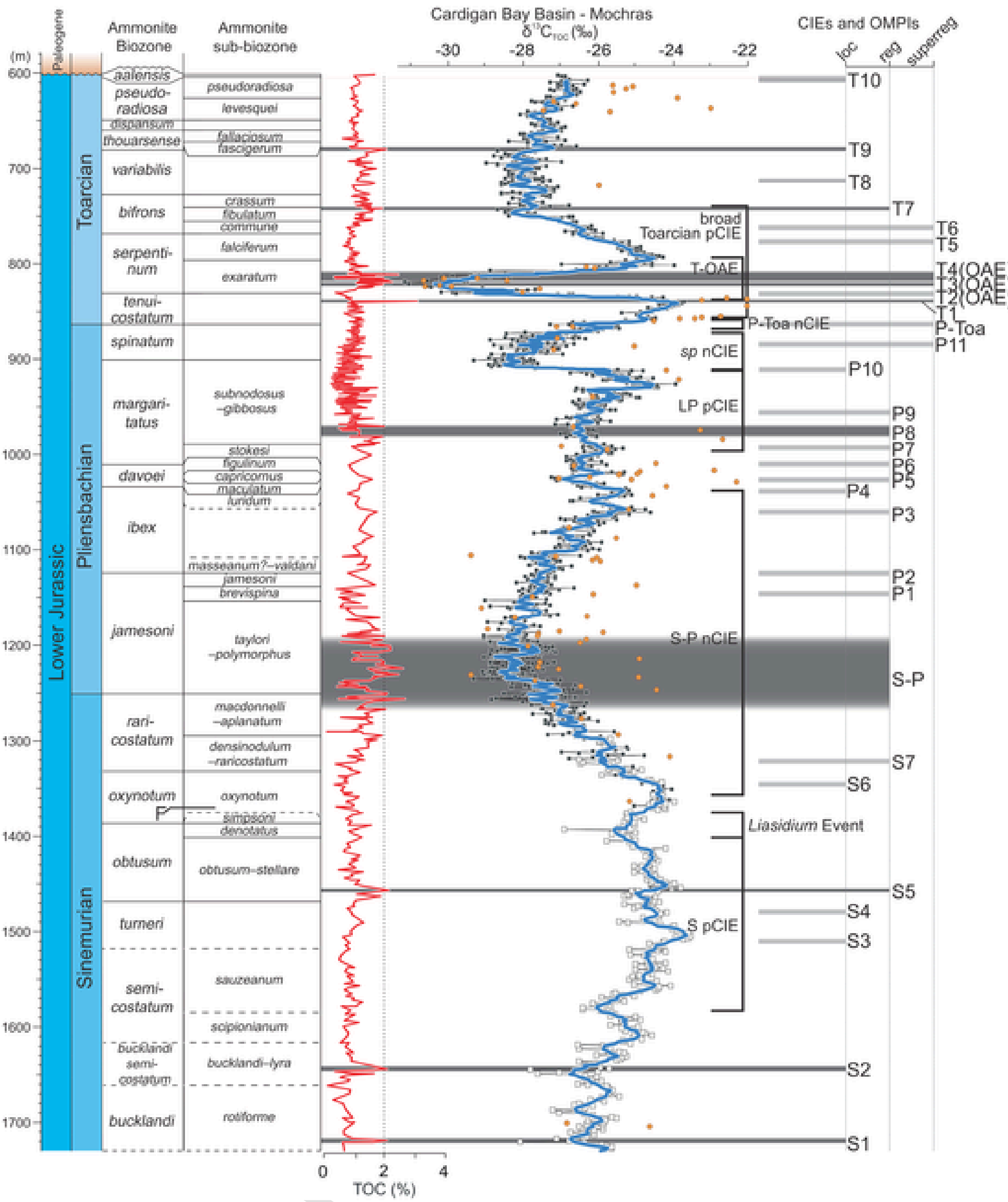

Fig. 4. $\delta^{13}C_{TOC}$, TOC, and projection of OMPIs for the Sinemurian–Toarcian at the reference Llanbedr (Mochras Farm) core, Cardigan Bay Basin (modified from Xu et al., 2018; Storm et al., 2020). The dark grey bars refer to OMPIs observed at Mochras whereas the light grey bars reflect the approximate position of OMPIs against the $\delta^{13}C_{TOC}$ curve. Blue line = seven-point moving average. Black squares = samples taken from core slabs; white squares = samples taken from reserve bags (~1.4-m intervals) of broken core. Orange circles = $\delta^{13}C_{wood}$. Depth of the samples from reserve bags refers to midpoint of the sample interval. Ammonite biostratigraphy after (Ivimey-Cook, 1971; Copestake and Johnson, 2013). Loc – Local; reg – regional; superregional; S – Sinemurian; P – Pliensbachian; LP – Late Pliensbachian; *sp* – *spinatum*; P-TOA – Pliensbachian-Toarcian; T-OAE – Toarcian oceanic anoxic Event; T – Toarcian; pCIE – positive carbon isotopic excursion; nCIE – negative carbon isotopic excursion; OMPI – organic matter preservation interval.

tion of a recognized feature of the geological record intimately associated with the OMPI, such as the Toarcian Oceanic Anoxic Event [e.g., OMPI T2(OAE)]. The minimum requirement to define an OMPI is a single sample, representative of a continuous bed at outcrop scale, with present day TOC > 2% and a confident stratigraphic age assignment (using, for example, bio- or chemostratigraphy) to a Subchronozone. Organic-rich sedimentary rocks that cannot be confidently assigned to a Subchronozone cannot be used to define an OMPI. However, such loosely dated occurrences may support observations from other better-dated sections, as is the case for the Sinemurian successions of Western North American basins.

3. OMPIs: temporal and paleogeographical trends

Because $\delta^{13}C$ records exist for many of the Lower Jurassic sedimentary successions showing organic matter enrichment, below we briefly discuss the stratigraphic occurrence of OMPIs within the context of the Lower Jurassic $\delta^{13}C$ record (Table 1, Figs. 1–4, and Appendixes 1–4). A detailed discussion on processes leading to Early Jurassic short- and long-term global carbon cycle perturbations is beyond the scope of this study.

Table 1
Early Jurassic OMPIs - abbreviations and geographic distribution (see Appendix 2 for references).

Abbreviation	OMPI	Known geographic distribution
OMPI T10	Late Toarcian (*levesquei–aalensis*)	Local (regional?)
OMPI T9	Late Toarcian (*thouarsense*,?)	Local
OMPI T8	Middle Toarcian (*variabilis*,?)	Local (regional?)
OMPI T7	Middle Toarcian (*bifrons*, *fibulatum*)	Regional (superregional?)
OMPI T6	Middle Toarcian (*bifrons*, *commune*)	Superegional
OMPI T5	Early Toarcian (*falciferum*, *elegans–falciferum*)	Superegional
OMPI T4 (OAE)	Early T-OAE + nCIE (positive trend of the T-OAE nCIE)	Superegional
OMPI T3 (OAE)	Early T-OAE -nCIE (negative trend of the TOAE nCIE)	Superegional
OMPI T2 (OAE)	Early T-OAE pre nCIE	Superegional
OMPI T1	Early Toarcian (*tenuicostatum*, *paltum–tenuicostatum*)	Superegional
OMPI P-Toa	Late Pliensbachian-Toarcian Event	Superegional
OMPI P11	Late Pliensbachian (*spinatum*, *hawskerense*)	Superegional
OMPI P10	Late Pliensbachian (*margaritatus*, *gibbosus*)	Local (superregional?)
OMPI P9	Late Pliensbachian (*margaritatus*, *subnodosus/gibbosus*)	Regional (superregional?)
OMPI P8	Late Pliensbachian (*margaritatus*, *subnodosus*)	Regional (superregional?)
OMPI P7	Late Pliensbachian (*margaritatus*, *stokesi*)	Regional (superregional?)
OMPI P6	Early–Late Pliensbachian (*davoei*, *figulinum–margaritatus*, *stokesi*)	regional (superregional?)
OMPI P5	Early Pliensbachian (*davoei*, *maculatum–capricornus*)	Regional (superregional?)
OMPI P4	Early Pliensbachian (*ibex*, *luridum*)	Local (superregional?)
OMPI P3	Early Pliensbachian (*ibex*, *valdani*)	Regional (superregional?)
OMPI P2	Early Pliensbachian (*jamesoni*, *jamesoni–ibex*, *masseanum*)	Regional (superregional?)
OMPI P1	Early Pliensbachian (*jamesoni*, *brevespina*)	Regional (superregional?)
OMPI S-P	Sinemurian–Pliensbachian (*raricostatum*, *raricostatoides–jamesoni*, *polymorphus*) (inc. S-PBE cf. Korte and Hesselbo, 2011)	Regional (superregional?)
OMPI S7	Late Sinemurian (base *raricostatum*, *densinodulum*?)	Regional
OMPI S6	Late Sinemurian (*oxynotum*, *oxynotum*)	Local (regional?)
OMPI S5	Late Sinemurian (*obtusum*, *obtusum–stellare*)	Regional (superregional?)
OMPI S4	Early Sinemurian (*turneri*, *birchii*)	Local (regional?)
OMPI S3	Early Sinemurian (*semicostatum*, *resupinatum–turneri*, *birchii*)	Local (superregional?)
OMPI S2	Early Sinemurian (*bucklandi/semicostatum*)	Local
OMPI S1	Early Sinemurian (*bucklandi*, *rotiforme*)	Local

3.1. Early Sinemurian–Late Pliensbachian

3.1.1. Sinemurian–Early Pliensbachian: OMPIs S1–S7, OMPI S-P and OMPIs P1–P6

The Early Sinemurian to Early Pliensbachian interval is marked by multiple shifts of up to around 5‰ in the carbon isotopic composition of organic and inorganic substrates from marine and continental environments (e.g. Jenkyns et al., 2002; van de Schootbrugge et al., 2005a; Suan et al., 2010; Silva et al., 2011; Jenkyns and Weedon, 2013; Duarte et al., 2014; Price et al., 2016; Ruhl et al., 2016; Xu et al., 2017a; Hesselbo et al., 2020; Schöllhorn et al., 2020b; Storm et al., 2020). Organic matter deposition and preservation in the Sinemurian seems to be strongly dependent on local depositional and tectonic conditions (Chadwick, 1986; Stapel et al., 1996; Rasmussen et al., 1998; Alves et al., 2002; Duarte et al., 2010, 2012, 2014; Poças Ribeiro et al., 2013; Silva et al., 2013), controlling, for example, relative sea-level and organic productivity (Bessereau et al., 1995; Jenkyns and Weedon, 2013; Schöllhorn et al., 2020b). It was suggested that deposition of organic-rich facies and black shales in the Sinemurian appears to be restricted to basins and time-intervals marked by extensional faulting and formations of depocenters with a geometry favourable to water mass stratification and organic matter accumulation and preservation (e.g. the Lusitanian and Wessex basins, Fleet et al., 1987).

A strong cyclicity in TOC distribution in the Wessex Basin is observed in the Lower Pliensbachian (Weedon and Jenkyns, 1990, 1999). It was suggested that these cycles resulted from high-frequency climatic variations that controlled the degree of oxygenation at the seafloor (van Buchem et al., 1995). Changes in the clay mineral assemblages in the Cardigan Bay Basin (Deconinck et al., 2019) reinforce the view that high-frequency climatic variations may have played a pivotal role in the OMPIs associated with the broad Late Sinemurian–Pliensbachian negative carbon isotopic excursion (nCIE). It was recently suggested that the stratigraphic occurrence of positive and negative (0.5–2‰ in TOC) CIEs in Sinemurian and Pliensbachian marine sedimentary archives reflect orbital (astronomical) forcing of the global exogenic carbon cycle, with a ~ 405 kyr eccentricity periodicity (Storm et al., 2020).

3.1.1.1. Sinemurian pCIE. The Sinemurian positive carbon isotopic excursion (pCIE positive and negative trend, ~2.5‰ in TOC) spans from the mid-*semicostatum* to the *obtusum* (*obstusum/stellare* subchronozone) chronozones (Storm et al., 2020) (Fig. 4). Organic-rich facies of this age are geochemically identified by TOC data and temporally constrained by detailed biostratigraphy in the Wessex Basin (Jenkyns and Weedon, 2013; Schöllhorn et al., 2020b), Lusitanian Basin (Poças Ribeiro et al., 2013; Brito et al., 2017; Duarte et al., 2017), and Cardigan Bay Basin (Storm et al., 2020), and additionally observed, for example, in the Cadwallader Terrane (Porter et al., 2014b) (Appendixes 1–3).

In the Wessex Basin, the organic-rich sedimentary rocks of the local (superregional?) OMPI S3 are associated with relatively more negative $\delta^{13}C_{org}$ values (Jenkyns and Weedon, 2013; Schöllhorn et al., 2020b). A similar situation is observed in the Last Creek 2 section from the Cadwallader Terrane (Porter et al., 2014b), where an interval tentatively dated to the top of the *leslei* (= *turneri*) Chronozone presents the same relationship between TOC and $\delta^{13}C_{org}$. However, a recent study from the Wessex Basin seems to indicate that, when accounting for changes in type and source of OM, a corrected $\delta^{13}C$ curve ($\delta^{13}C$-HI index = measured $\delta^{13}C_{TOC}$ – (a * HI), a being the slope of the correlation trend between $\delta^{13}C_{TOC}$ and HI, Schöllhorn et al., 2020b; Suan et al., 2015) shows an association between OMPIs and corrected pCIEs. Further studies are necessary to validate that HI variation truly reflects changes in OM or if, for example, is a reflection of different degradation/diagenetic processes of similar OM (Silva et al., 2013; Suan et al., 2015; Charbonnier et al., 2020; Storm et al., 2020).

3.1.1.2. Sinemurian Liasidium Event. The Early Jurassic Sinemurian *Liasidium* Event, *obtusum chonozone (denotatus* Subchronozone*)* *–oxynotum* chronozone *(oxynotum* Subchronozone*)* (Riding et al., 2013; Hesselbo et al., 2020) is characterized by the acme of the *Liasidium variabile* dinoflagellate and high abundances of *Classopollis classoides,* suggesting an association with high land and sea temperatures (Riding et al., 2013; Hesselbo et al., 2020; Schöllhorn et al., 2020b) (Fig. 4). Masetti et al. (2017) and Franceschi et al. (2019) tentatively correlated the "Arnioceras time Event" nCIE from the southern Alps with the negative shift identified by Riding et al. (2013). So far, this carbon isotopic disturbance appears not to be stratigraphically associated with widespread deposition of OM. However, a poorly age constrained organic-rich interval (currently not defined as an OMPI due to uncertain age assignment) in the Lusitanian Basin (Portugal) may be contemporaneous with the Sinemurian *Liasidium* Event (Appendixes 2 and 3).

3.1.1.3. Sinemurian–Pliensbachian nCIE. The Sinemurian–Pliensbachian nCIE corresponds to a long-term negative and then positive $\delta^{13}C$ trend of about 2–5‰ in organic matter, expressing a significant change in the global carbon cycle across the Late Sinemurian (*oxynotum* Chronozone) to the Early Pliensbachian (*ibex* Chronozone, *luridum* Subchronozone) (Hesselbo et al., 2000b; Jenkyns et al., 2002; Rosales et al., 2004; Silva et al., 2011; Korte and Hesselbo, 2011; Jenkyns and Weedon, 2013; Duarte et al., 2014; Franceschi et al., 2014; Price et al., 2016; Ruhl et al., 2016; Gómez et al., 2016b; Peti et al., 2017; Danisch et al., 2019; Schöllhorn et al., 2020b; Schöllhorn et al., 2020a; Storm et al., 2020; Mercuzot et al., 2020), including the Sinemurian–Pliensbachian Boundary Event (S-PBE, *raricostatum* Chronozone, *aplanatum* Subchronozone–*jamesoni* Chronozone, *taylori* Subchronozone, cf. Korte and Hesselbo, 2011). This long-long term nCIE is marked by numerous episodes of local and regional (superregional?) deposition of OM in several European basins, corresponding to OMPIs S6, S7, S-P, P1, P2, P3, and P4 (Table 1, Fig. 3 and Fig. 4, and Appendixes 2 and 3).

Earth system processes leading to the long ranging Sinemurian–Pliensbachian nCIE and climatic and environmental changes are largely unknown but are speculated to relate to either the opening of the Hispanic Corridor or a late phase of enhanced global continental (silicate) weathering induced by the Central Atlantic Margin Province volcanism (see discussion in Korte and Hesselbo, 2011; Porter et al., 2013; Gómez et al., 2016b; Price et al., 2016; Ruhl et al., 2016; Danisch et al., 2019; Franceschi et al., 2019; Schöllhorn et al., 2020b, Schöllhorn et al., 2020a; Storm et al., 2020). Considering the extremely limited stratigraphic data outside the northern western Tethys realm (see A1.4), significant accumulation of OM (high TOC) during the Sinemurian–Pliensbachian nCIE appears to be most expressive in the Wessex and Lusitanian basins; the exception is the regional (superregional?) OMPI S-P where sedimentary OM enrichment was more widespread (Fig. 3). The regional (superregional?) OMPI S-P is associated with the most negative $\delta^{13}C$ values of the broad Sinemurian–Pliensbachian nCIE [including the Sinemurian–Pliensbachian boundary event (S-PBE) cf. Korte and Hesselbo, 2011], and coincides with the final stages of the Late Sinemurian warming event and transition to subsequent cooling (Korte and Hesselbo, 2011; Gómez et al., 2016b). The Trento platform (Italy) is marked by platform sedimentary deposits with relatively elevated TOC, but below 2%, which are interpreted to be associated with the S-PBE, suggesting relative OM enrichment even in shallow areas at this time (Franceschi et al., 2014, 2019).

3.1.2. Late Pliensbachian pCIE: OMPIs P7–P10

The Late Pliensbachian is marked by a relatively low amplitude long-term pCIE of about 2–4‰ in carbonate, peaking in the *margaritatus* Chronozone (circa *gibbosus* Subchronozone) (Fig. 4 and Appendix 3). The regional (superregional?) OMPIs P7, P8, and P9 are concomitant with superimposed smaller-scale $\delta^{13}C$ positive excursions (~ 0.5–2‰ in carbonate and TOC) observed in multiple basins (see Silva et al., 2011), possibly expressing long (~405 kyr) eccentricity modulation of the global carbon cycle and climatic and environmental conditions at those times (Jenkyns et al., 2002; Silva et al., 2011, 2015; Caruthers et al., 2014; Silva and Duarte, 2015; Mercuzot et al., 2020; Storm et al., 2020).

Despite the recognized gaps in data coverage, organic-rich facies associated with the Late Pliensbachian pCIE are (apparently) concentrated mainly along the areas bordering the proto-Atlantic Ocean and possibly the Boral Sea (Fig. 3). During the Late Pliensbachian, organic productivity and preservation were enhanced during major transgressive episodes (Hallam, 1981; de Oliveira et al., 2006; Duarte et al., 2010; Silva et al., 2011, 2012, 2015; Silva and Duarte, 2015). Rosales et al. (2006) suggested a link between second-order relative sea-level changes in the Basque–Cantabrian Basin and variations in seawater geochemistry during the Early Jurassic, noting a coincidence between transgressions and increasing $\delta^{13}C$, TOC and Hydrogen Index (and vice-versa for regressions). It was also suggested that the Late Pliensbachian black shales were associated with warm temperatures (hyperthermals?) within the relatively cooler Late Pliensbachian (Silva and Duarte, 2015; Peti and Thibault, 2017; Schöllhorn et al., 2020b).

3.1.3. Spinatum nCIE: OMPI P11

The *spinatum* Chronozone is generally regarded as a cool interval (Hinnov and Park, 1999; Price, 1999; Dera et al., 2009b; Dera et al., 2009a; Suan et al., 2010; Korte and Hesselbo, 2011; Silva and Duarte, 2015; Gómez et al., 2016b; Ruebsam et al., 2019; Deconinck et al., 2020; Storm et al., 2020). It has been suggested that geological storage of OM associated with the Late Pliensbachian pCIE resulted in decreased atmospheric CO_2, triggering and/or amplifying cooling in the *spinatum* Chron (e.g. Suan et al., 2010; Silva et al., 2011; Silva and Duarte, 2015; Storm et al., 2020). It was also speculated that a potentially early onset of North Sea doming may have caused the development of regressive facies in the Late Pliensbachian of the North Sea region, which may have promoted the regional shift from a warm to a cooler climate mode (Korte et al., 2015) (Fig. 2). More recently, De Lena et al. (2019) suggested the degassing of volcanic S-species (SO_2) was an unlikely driving mechanism of the cool and dry climate of the late Pliensbachian.

The later stages of the *spinatum* nCIE (cf. Storm et al., 2020) were associated with organic-rich rocks defining the superregional OMPI P11 (Figs. 3 and 4).

3.2. Latest Pliensbachian–Toarcian

The latest Pliensbachian–Toarcian is characterized by large amplitude variations in the $\delta^{13}C$ record. The Early Toarcian is marked by one of the most significant carbon cycle perturbations of the past 200 million years, expressed by a large amplitude nCIE, up to 7‰ in TOC (e.g. Hesselbo et al., 2000a; Röhl et al., 2001; Kemp et al., 2005; Suan et al., 2015), up to 8‰ in fossil wood (e.g. Hesselbo et al., 2007), up to 5‰ in bulk carbonate (e.g. Hermoso et al., 2009; Kafousia et al., 2011), up to 4‰ in macrofossil carbonate (e.g. brachiopods (Suan et al., 2010; Ullmann et al., 2020) superimposed on a broader Early Toarcian pCIE, which encompasses the *tenuicostatum–bifrons* time interval (Fig. 4) (Küspert, 1982; Jenkyns and Clayton, 1986; Jenkyns, 1988, 2010; Hesselbo et al., 2000a, 2007; Röhl et al., 2001; Jenkyns et al., 2002; Cohen et al., 2004; Röhl and Schmid-Röhl, 2005; Woodfine et al., 2008; Hermoso et al., 2009, 2012, 2013; Al-Suwaidi et al., 2010, 2016; Bodin et al., 2010; Littler et al., 2010; Bodin et al., 2016; Suan et al., 2010, 2011, 2015; Lézin et al., 2013; Sabatino et al., 2013; Kafousia et al., 2014; Kemp and Izumi, 2014; Pittet et al., 2014; Reolid et al., 2014; Caruthers et al., 2014; Pieńkowski et al., 2016; Martinez et al., 2017; Silva et al., 2017; Them et al., 2017; Xu et al., 2017b, 2018; Fantasia et al., 2018, 2019b; Ruebsam et al., 2019, 2020; Boulila et al., 2019; Danise et al., 2019; Jin et al., 2020; Ruebsam and Al-Husseini, 2020).

As previously observed, Lower Toarcian organic-rich facies are described in many areas of the northern West Tethys Shelf, Boreal and Panthalassic margins (e.g. Baudin et al., 1990; Röhl et al., 2001; van de Schootbrugge et al., 2005b; Gómez et al., 2008; McArthur et al., 2008; Al-Suwaidi et al., 2010; Jenkyns, 2010; Suan et al., 2011, 2015; Hermoso et al., 2012; Kemp and Izumi, 2014; Pieńkowski et al., 2016; Silva et al., 2017; Them et al., 2017; Ruvalcaba Baroni et al., 2018; Rodrigues et al., 2020c). These are particularly expressive in the central and northern European epicontinental areas, where deposition took place in anoxic basins under dominantly warm and humid conditions and associated with an intensification of the hydrological cycle and fluvial runoff (Jenkyns, 1988, 2010; Rees et al., 1999; Cohen et al., 2004; Pearce et al., 2008; Dera et al., 2009a, 2009b; Brazier et al., 2015; Them et al., 2017; Fantasia et al., 2018; Rodrigues et al., 2021; Rodrigues et al., 2020b). In the southern areas of the West Tethys Shelf, such as western and southern Iberian, and northern Gondwana margin, organic-rich intervals are spatially and temporally restricted (e.g. Hesselbo et al., 2007; Bodin et al., 2010; Gómez and Goy, 2011; Reolid et al., 2012, 2014; Rodrigues et al., 2020b, 2021; Rodrigues et al., 2016, 2019, 2020a; Ait-Itto et al., 2017; Fantasia et al., 2019a; Ruebsam et al., 2020; Silva et al., 2021) and the absence of significant Lower Toarcian OM enrichment is interpreted as a consequence of oxic conditions and low productivity associated with a semi-arid climate and general circulation patterns (e.g. van de Schootbrugge et al., 2005b; Reolid et al., 2014, 2020; Ruvalcaba Baroni et al., 2018; Fantasia et al., 2019b; Rodrigues et al., 2019, 2020a, 2020b; Ruebsam et al., 2020) (Fig. 3 and Appendix 2).

3.2.1. P-Toa nCIE: OMPI P-Toa

A smaller-scale (~2–3‰ in TOC) nCIE is observed at the base of the Toarcian *polymorphum* Chronozone, the Pliensbachian–Toarcian event (P-Toa Event), coincidental with the superregional OMPI P–Toa (Hesselbo et al., 2007; Littler et al., 2010). It was suggested that the P–Toa event resulted from a carbon cycle perturbation analogous to the one marking the T-OAE (Littler et al., 2010), and it is stratigraphically associated with an extinction phase in marine benthos at the base of the Toarcian (e.g. Little and Benton, 1995; Littler et al., 2010; Caruthers et al., 2013) and significant continental weathering and volcanic activity (Percival et al., 2015, 2016).

3.2.2. The broad Lower Toarcian pCIE and the T-OAE nCIE: OMPIs T1, T2 (OAE)–T4(OAE), and T 5–7

The superregional OMPI T1 spans the *paltum–tenuicostatum* subchrons and is associated with the onset of a broad Lower Toarcian pCIE onto which is superimposed the T-OAE nCIE (Fig. 4). The T-OAE is characterized by the geographically widespread deposition of organic-rich sediments (Fig. 4), ocean anoxia - when geographically widespread areas of the seafloor and oceans become oxygen depleted or anoxic (Schlanger and Jenkyns, 1976; Küspert, 1982; Jenkyns and Clayton, 1986; Jenkyns, 1988; Dickson et al., 2017), global warming (Bailey et al., 2003; Dera et al., 2009b; Suan et al., 2010; Danise et al., 2013; Ullmann et al., 2020), ocean acidification (e.g. Suan et al., 2010; Müller et al., 2020), carbonate productivity crisis (e.g. Mattioli et al., 2009; Han et al., 2018), and a 2nd-order mass extinction event in benthic and pelagic groups (e.g. Aberhan and Baumiller, 2003; Arias, 2006; Mattioli et al., 2009; Caswell et al., 2009; Comas-Rengifo et al., 2010; Gómez and Arias, 2010; Gómez and Goy, 2011; García Joral et al., 2011; Fraguas et al., 2012; Danise et al., 2013; Caswell and Frid, 2017; Piazza et al., 2019; Cabral et al., 2020). Several mechanisms have been suggested to have initiated the T-OAE nCIE and carbon cycle perturbation: (1) carbon degassing from the Karoo-Ferrar Large Igneous Province (Pálfy and Smith, 2000, 2) rapid release of methane from gas hydrates from continental-margin and slope sediments (Hesselbo et al., 2000a), (3) thermogenic methane release from subsurface organic-rich shales intruded by dyke and sills associated with Karoo-Ferrar emplacement (McElwain et al., 2005; Svensen et al., 2007, 4) increased CO_2 release from the decomposition and oxidation of terrestrial organic matter associated with increased fungal activity (Pieńkowski et al., 2016, 5) elevated methane release from terrestrial environments (Them et al., 2017), and (6) destabilization of labile cryospheric carbon reservoirs (Silva and Duarte, 2015; Krencker et al., 2019; Ruebsam et al., 2019).

The superregional OMPIs T2(OAE), T3(OAE), and T4(OAE) are associated with widespread deposition of organic-rich facies with high TOC, including black shales. Maximum TOC contents are frequently observed in the superregional OMPI T4(OAE), associated with the positive $\delta^{13}C$ trend of the T-OAE nCIE (the positive carbon isotopic trend of the T-OAE nCIE is here defined as the interval between minimum $\delta^{13}C$ and maximum $\delta^{13}C$ that marks the end of the T-OAE interval) (Fig. 4). The superregional OMPI T4(OAE) is observed even in areas not characterized by significant Lower Toarcian organic-rich deposition, such as the Lusitanian Basin (de Oliveira et al., 2006; Hesselbo et al., 2007; Rodrigues et al., 2016, 2020a).

Carbon sequestration during the T-OAE seems to be broadly associated with a peak in $^{187}Os/^{188}Os$ (Cohen et al., 2004; Percival et al., 2016; Them et al., 2017; van Acken et al., 2019) and a minimum in $\delta^{18}O$ (Suan et al., 2008; Dera et al., 2011; Krencker et al., 2014; Ullmann et al., 2014, 2020; Korte et al., 2015), suggesting that maximum organic carbon sequestration during the T-OAE was concomitant with sustained high temperatures and enhancement of the hydrological cycling and increased weathering rates (Jenkyns, 2010; Xu et al., 2017b; Fantasia et al., 2019a; Rodrigues et al., 2020b; Ullmann et al., 2020). The here demonstrated offset between the T-OAE $\delta^{13}C$ negative trend and the maximum sedimentary TOC values combined with the widespread nature of the superregional OMPI T4(OAE) agree with recent modelling efforts indicating a lag between peak greenhouse gas input and peak anoxia/carbon sequestration (Ullmann et al., 2020).

Several OMPIs are observed after the T-OAE (Fig. 4 and Appendixes 2 and 3). The superregional OMPIs T5 and T6 and the regional OMPI T7 are associated with the decreasing limb of the broad Early Toarcian $\delta^{13}C$ positive excursion initiated at the base of the *tenuicostatum* Chronozone (Fig. 4). Oxygen stressed conditions and carbon sequestration were prevalent from the mid-*semicelatum* Subchron (*tenuicostatum* Chron) to the upper *bifrons* Chron in several locations, as recorded in the German sections (Röhl et al., 2001), Cleveland Basin (Thibault et al., 2018), and Slyne Basin (Silva et al., 2017). Post-OAE black shales in the Paris Basin were likely concomitant with third-order sea-level changes (Hermoso et al., 2013; sensu de Graciansky et al., 1998).

4. Comparison with the Cretaceous OAEs

OAEs were initially defined as brief time intervals during which significant portions of the global oceans were extensively deoxygenated (Schlanger and Jenkyns, 1976; Jenkyns, 1980, 2010; Arthur et al., 1990; Leckie et al., 2002). Although a comprehensive model to explain all OAEs is not yet fully constrained, the consensus is that these events are associated with a sudden influx of greenhouse gasses into the atmosphere and global warming, leading to an acceleration of the hydrological cycle, increased continental weathering, enhanced nutrient delivery to the oceans, intensified oceanic upwelling, and increased productivity in marine and continental environments (Schlanger and Jenkyns, 1976; Weissert, 1989; Arthur et al., 1990; Hesselbo et al., 2000a; Leckie et al., 2002; Bodin et al., 2010; Jenkyns, 2010; Trabucho-Alexandre et al., 2011; Herrle et al., 2015; Montero-Serrano et al., 2015; Xu et al., 2017b; Fantasia et al., 2018; Silva et al., 2020b).

OAEs are characterized by negative or positive CIEs, interpreted to reflect the sudden influx of isotopically light CO_2 into the atmosphere (either from, for example, volcanism, metamorphic degassing, or destabilization of gas hydrates) or burial of vast amounts of organic carbon, respectively (Jenkyns, 2010). However, local basinal responses often override global forcing on the occurrence, onset, and termination of organic-rich deposition associated with OAEs (e.g. Trabucho-Alexandre et al., 2011; Fantasia et al., 2019a; Rodrigues et al., 2019, 2020a) and CIEs (reflecting global carbon cycle change) in combination with evidence for regional to superregional bottom-water deoxygenation have become the accepted characteristic features of an OAE, rather than widespread organic-rich deposition alone (Jenkyns, 2010).

The main Mesozoic OAEs comprise the T-OAE, Early Aptian OAE (OAE 1a), and Cenomanian–Turonian OAE (OAE 2). Secondary OAE events, characterized by less prominently expressed CIE and less severe paleoenvironmental perturbations, are the latest Hauterivian OAE

(Faraoni Event), Late Valanginian OAE (Weissert Event), Late Aptian–Early Albian OAE (OAE 1b cluster), and (more doubtfully) the Late Albian (OAE 1c and OAE 1d) and the Coniacian–Santonian (OAE 3) (Jenkyns, 2010). Main and secondary OAEs are characterized by brief episodes of regional to superregional deposition of organic matter associated with disruption of the global carbon cycle (Jenkyns, 1980, 2010; Arthur et al., 1990; Leckie et al., 2002) (Fig. 5).

The compilation of Sinemurian–Toarcian OMPIs presented in this paper demonstrates that regional and superregional sedimentary sequestration of OM during the Early Jurassic was more common than previously thought (Fig. 5). Because of the association of OMPIs with well-constrained CIEs (Figs. 3 and 4 and Appendices 3 and 4), we propose that the regional (superregional?) OMPI S-P and the superregional OMPI P11 and OMPI P-Toa should now be considered as secondary OAEs, similarly to the secondary Cretaceous OAEs.

The Early Toarcian superregional OMPIs T1, T5, T6, and the regional (superregional?) OMPI T7 occur in association with the Lower Toarcian pCIE, and their spatial spread supports some degree of equivalency to the secondary Cretaceous OAEs. These OMPIs demonstrate that the establishment of (global?) environmental and depositional conditions leading to widespread preservation of OM in the Late Pliensbachian–Early Toarcian was not exclusively associated with the T-OAE.

5. Limitations and future challenges

The methodology presented here for resolving OMPIs applies to any geological interval, is simple and straightforward, and provides a framework for exploratory research on large scale sedimentary organic matter sequestration and its impacts on the global carbon cycle. However, mechanistic understanding of sedimentary OM sequestration at

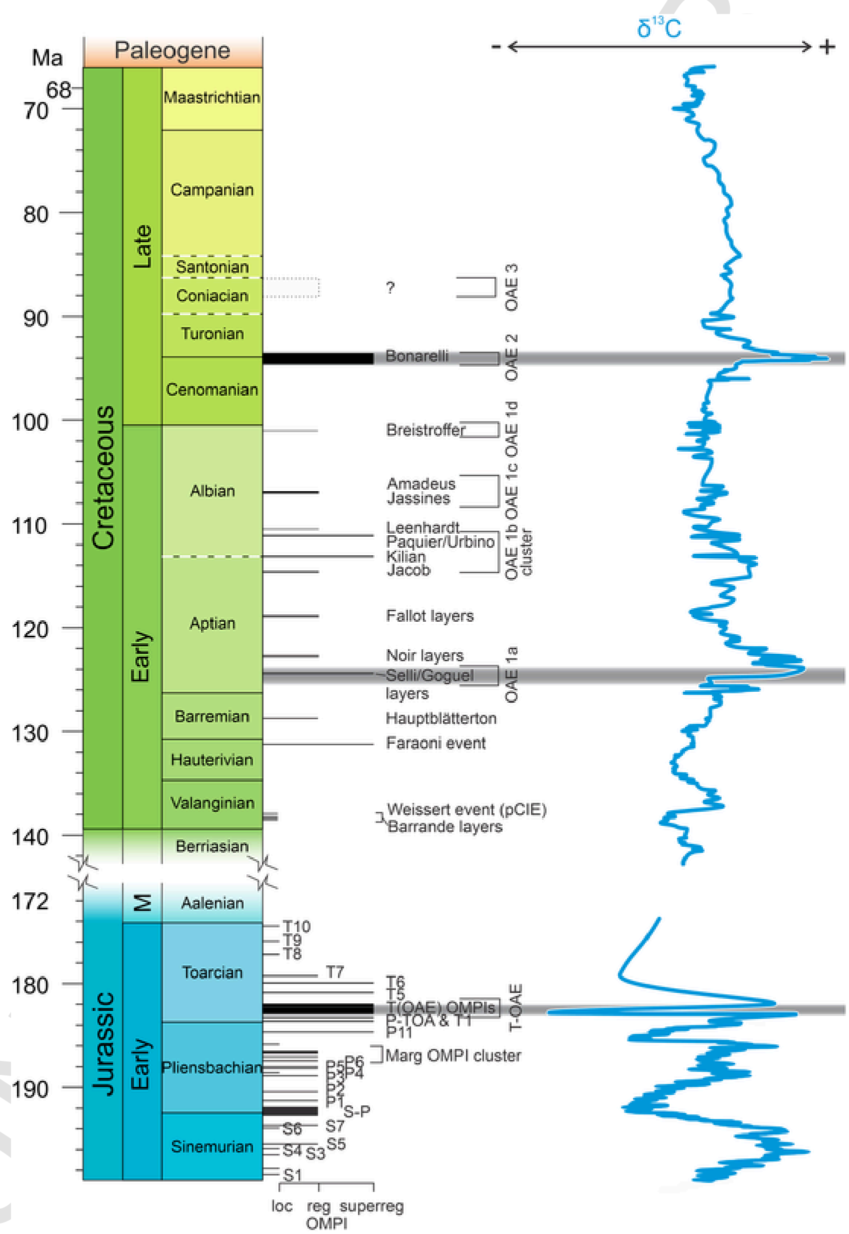

Fig. 5. Mesozoic OAEs and comparison with the Early Jurassic OMPIs record (the grey bars correspond to the main Jurassic–Cretaceous OAEs). The Sinemurian–Pliensbachian record was created using the time scale and $\delta^{13}C$ from the reference Mochras core, Cardigan Bay Basin, UK (Storm et al., 2020). The Toarcian and Cretaceous time scale (Ogg et al., 2016) and $\delta^{13}C$ record (Herrle et al., 2004; Reichelt, 2005; Föllmi et al., 2006; Jarvis et al., 2006; Hesselbo et al., 2007; Voigt et al., 2012; Martinez et al., 2015; Xu et al., 2018) were generated using TSCreator software (PUBLIC7.4_windows10_latest_06September2019). Cretaceous OAEs are from Arthur et al. (1990); Leckie et al. (2002); Bodin et al. (2007); Mutterlose et al. (2009); Jenkyns (2010); Westermann et al. (2010); Föllmi et al. (2012); Herrle et al. (2015).

basinal, intrabasinal, and global scales is often hampered by the lack of spatial data coverage, temporal data uncertainty, and sampling/data resolution for many of the intervals studied (see A1.4). In addition, understanding of sedimentation rates and their impact on true rates of organic carbon accumulation, or carbon burial fluxes, is often limited or non-existent; this is particularly important when using a fixed cut-off value of 2% present-day TOC to define an OMPI or possible source-rocks. For example, sedimentary successions with low TOC/high sedimentation rate may represent the same organic matter accumulation per time unit as high TOC/low sedimentation rate depositional environments. Furthermore, the intrabasinal spatial extent of an OMPI is often unconstrained, even when using the most advanced imaging/modelling tools (e.g. Bruneau et al., 2017). A detailed assessment of sedimentation rates and the translation of present-day TOC values into organic matter accumulation/sequestration rates (or fluxes) for individual OMPIs at the outcrop and basin-scale are however outside the scope of this study.

Thermal maturation may also lead to an underestimation of the original OM accumulation rate, and possibly even the spatial extent of OMPIs, especially if original TOC approximated 2% in now highly mature successions. The magnitude of this effect depends mainly on the level of maturity and the relative proportions of generative and non-generative organic carbon compounds, with type I and type II kerogens having a higher proportion of generative organic carbon which therefore are prone to have depressed original TOC values (Jarvie, 2012). Type III and type IV kerogens are less affected by thermal maturation, and its suppressing effect on TOC, as hydrocarbons generation was limited. Therefore, the loss of carbon via maturation and expulsion may thus have resulted in an underestimation of organic carbon burial throughout geological history and during Early Jurassic OMPIs when using uncorrected present-day TOC values as a proxy.

Although the present work presents a framework for the stratigraphic distribution of OMPIs across the ~20 Myr of the Sinemurian–Toarcian time interval, we emphasize that without a detailed assessment of (1) inter- and intrabasinal bio- and chronostratigraphic correlations, (2) duration and stratigraphic completeness, (3) TOC stratigraphic and basinal variability, and (4) thermal maturation and diagenesis, the study of OMPIs as defined here will generally underestimate the spatial and temporal patterns in geological organic carbon storage and, therefore, result in inaccurate estimates on OM burial and as a sink of atmospheric CO_2. Future research should aim to disentangle (1) the noted complexities in estimating original TOC and organic carbon accumulation rates, (2) temporal and spatial variability in environmental or Earth system feedback mechanisms driving sedimentary carbon sequestration, and (3) their combined impact on the global carbon cycle. Further work is expected to define additional Early Jurassic OMPIs and to improve current understanding of the spatial and temporal distribution of individual OMPIs.

6. Conclusions

The recognition of several Early Jurassic organic matter preservation intervals (OMPIs) of regional and superregional extent gives new insight into the role that OM sequestration played in Earth System process and provides support to current theoretical models for carbon cycle perturbations in the Early Jurassic (e.g., the T-OAE). From this study, we conclude that:

1. Sinemurian regional OMPIs were mainly restricted to a small number of European basins. A relatively more widespread OMPI was associated with the most negative $\delta^{13}C$ interval of the long-term Sinemurian–Pliensbachian $\delta^{13}C$ excursion. The stratigraphic and causal relationship between OMPIs and CIEs (as representative of global carbon cycle perturbations) is not always straightforward.
2. Pliensbachian regional (superregional?) OMPIs occur mostly in the basins bordering the proto-Atlantic Ocean, often associated with positive $\delta^{13}C$ excursions and short-lived warming intervals (hyperthermals?). As suggested previously, it is apparent that significant OM sequestration predates [regional (superregional?) OMPIs 7, 8, and 9] and coincides with the onset of the Late Pliensbachian cool episode and the *spinatum* nCIE (superregional OMPI 11).
3. As extensively discussed in the existing literature, Early Toarcian OMPIs are widespread. The Early Toarcian superregional OMPIs T2 (OAE), T3(OAE) and T4(OAE) are associated with the most pronounced $\delta^{13}C$ negative carbon isotopic excursion in the Mesozoic. Maximum TOC contents [superregional OMPI T4(OAE)] occur in association with the positive $\delta^{13}C$ trend of the Early Toarcian Oceanic Anoxic Event (T-OAE), indicating, and as suggested by recent modelling efforts, that peak OM sequestration and anoxia post-dated the main event of carbon injection into the atmosphere. However, several superregional OMPIs pre- and postdated the TOAE interval. These OMPIs demonstrate that the establishment of (global?) environmental and depositional changes led to widespread preservation of OM in the Late Pliensbachian and Early Toarcian and thus not exclusively during the main Early Toarcian OAE.
4. This study demonstrates that organic-rich intervals of regional and superregional expression in the Early Jurassic are more common than previously thought. It is contended that considering the association of the regional (superregional?) OMPI S-P (*raricostatum*, *raricostatoides–jamesoni*, *polymorphus*) and the superregional OMPI P11 (*spinatum*, *hawskerense*) and OMPI P-Toa with well-characterized CIEs, these should now be considered secondary OAEs, similarly to the secondary Cretaceous OAEs. The superregional Early Toarcian OMPIs T1 (*tenuicostatum*, *paltum–tenuicostatum*), T5 (*falciferum*, *elegans–falciferum*), T6 (*bifrons*, *commune*), and the regional (superregional?) OMPI T7 (*bifrons, fibulatum*) occur in association with the broad lower Toarcian pCIE.

Declaration of Competing Interest

None.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2021.103780.

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