

1 The early Toarcian Oceanic Anoxic Event (Jenkyns Event) in the
2 Alpine-Mediterranean Tethys, north African margin, and north
3 European epicontinental seaway

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12 **ABSTRACT**

13 The early Toarcian Oceanic Anoxic Event (Jenkyns Event) was associated with major
14 world-wide climatic changes with profound effects on the global carbon cycle. This review revisits
15 the available literature covering the Jenkyns Event applying an updated common stratigraphic
16 definition, allowing illustration of the development and evolution of anoxia in the Alpine-
17 Mediterranean Tethys, north African margin, and North European epicontinental basins within a
18 high-resolution temporal framework. The survey combines geographic and stratigraphic
19 distribution of black shale, organic-matter properties (total organic carbon content and
20 composition), variations in benthic fauna, distribution of euhedral and framboidal pyrite, and redox
21 conditions reconstructed on the basis of both inorganic and organic geochemical data. The
22 compilation demonstrates that bottom waters were generally well oxygenated prior to the negative

23 carbon-isotope excursion of the Toarcian Oceanic Anoxic Event whose onset was marked by the
24 synchronous deterioration in bottom-water oxygen conditions at supra-regional scale. Persistent
25 euxinia was dominantly confined to the north European epicontinental basins and sub-basins,
26 paralleled by a supraregional decline in oxygen content at the seafloor also in the Alpine-
27 mediterranean Tethys area. In the interval of time represented by the core of the negative carbon-
28 isotope excursion the most extreme redox conditions were reached along with intense euxinia
29 extending periodically into the photic zone accompanied by deposition of black shales whose
30 organic-matter content reached maximum values. Recovery to better oxygenated conditions was a
31 diachronous process that started, in most places, at a time immediately following the end of the
32 negative carbon-isotope excursion. The Alpine-Mediterranean Tethys became well oxygenated,
33 while north European epicontinental areas experienced anoxia with less intense and intermittent
34 sulphidic conditions interspersed with brief periods of oxygenation. $\Delta^{18}\text{O}$ variations reflect a
35 progressive increase in fresh-water input to the northern European epicontinental basins and sub-
36 basins that reached its acme in correspondence with the lowest values of the negative carbon-
37 isotope anomaly. In these areas, the proximity to sources of fresh-water input and the local
38 physiography or geographic restriction limited water exchange with the Tethys Ocean, favouring
39 the onset of anoxia/euxinia and organic-matter preservation. These basins and sub-basins, due to
40 their relatively closed physiography and redox conditions, acted as pools of dissolved divalent
41 manganese associated with accumulation of iron sulphides. Part of the soluble manganese spilled
42 out of these basins in oxygen minimum zones, being deposited/precipitated at the edge of the more
43 oxygenated Tethys Ocean, and thereby leading to the formation of local manganese-rich
44 carbonates deposited during the Jenkyns Event.

45

46 **Keywords:** anoxia, black shales, chemostratigraphy, Jurassic, T-OAE, Toarcian

47

48 **1. INTRODUCTION**

49 The Toarcian Oceanic Anoxic Event (T-OAE) was originally identified as a time of
50 globally developed anoxia as indicated by the widespread deposition of lower Toarcian black
51 shales associated with a positive carbon-isotope excursion (Jenkyns, 1985, 1988). Later
52 investigations showed that the broad positive carbon-isotope excursion extending over much of
53 the lower Toarcian is interrupted in its central portion by an abrupt negative carbonate and organic-
54 carbon isotope anomaly (Jenkyns and Clayton, 1986, 1997). This negative carbon-isotope
55 excursion (negative CIE) has now been globally observed both in shallow- and deep-marine
56 records as well as in continental archives (Jenkyns and Clayton, 1986; Hesselbo et al., 2000, 2007;
57 Schouten et al., 2000; Röhl et al., 2001; Jenkyns et al., 2001; Jenkyns et al., 2002; McElwain et
58 al., 2005; Kemp et al., 2005; Emmanuel et al., 2006; van Breugel et al., 2006; Sabatino et al., 2009;
59 Al-Suwaidi et al., 2010; Caruthers et al., 2011; Gröcke et al., 2011; Hesselbo and Pieńkowski,
60 2011; Izumi et al., 2012; Trabucho-Alexandre et al., 2012; Kafousia et al., 2011; Kafousia et al.,
61 2014; Ikeda and Hori, 2014; Kemp and Izumi, 2014; Reolid, 2014; Xu et al., 2017; Them et al.,
62 2017a; Fantasia et al., 2018; Ikeda et al., 2018; Filatova et al., 2020; Reolid et al. 2020b; Ruebsam
63 and Al-Husseini, 2020; Ramirez and Algeo, 2020a, b; Hougård et al., 2021; Erba et al., 2022; Liu
64 et al., 2022; Kemp et al., 2022a; Kunert and Kendall, 2023), and is thus a real phenomenon of the
65 global carbon cycle.

66 The T-OAE was associated with a phase of extreme global climate change (Jenkyns, 2003;
67 Jenkyns, 2010; Cohen et al., 2004; Ullmann et al., 2013; Percival et al., 2016; Them et al., 2017b;
68 Jenkyns and Macfarlane, 2021) characterized by extraordinary warmth (Dera et al., 2011; Korte

69 and Hesselbo, 2011; Gómez et al., 2016; Ruebsam et al., 2020d), accelerated weathering (Jenkyns,
70 2003; Jenkyns, 2010; Cohen et al., 2004; Ullmann et al., 2013; Percival et al., 2016; Them et al.,
71 2017b; Jenkyns and Macfarlane, 2021; Liu et al., 2022), a major marine transgression (Hallam,
72 1981; Haq et al., 1987; Hesselbo and Jenkyns, 1998; Hardenbol et al., 1998; Hesselbo, 2008), and
73 ocean acidification (Erba, 2004; Trecalli et al., 2012; Casellato and Erba, 2015; Posenato et al.,
74 2018; Müller et al., 2020; Ettinger et al., 2021). Widespread accumulation of black shales observed
75 in the north European epicontinental seaway seems to have been associated with relatively low
76 depositional rates of accompanying clastic and carbonate, leading to a degree of enrichment in
77 organic matter even if its flux to the sea floor was not in any way enhanced. In this case, improved
78 preservation of organic matter was largely due to the frequent and prolonged presence of
79 anoxia/euxinia extending upwards from the sea floor (Mattioli et al., 2004, 2009; Reolid et al.,
80 2021a, Kemp et al., 2022b; Ruebsam et al., 2022b). These extreme palaeoenvironmental conditions
81 impacted marine primary producers affecting the biocalcification of calcareous nannoplankton and
82 induced palaeoecological shifts in the dinoflagellate cyst community (Bucefalo Palliani et al. 2002;
83 Erba, 2004; Mattioli et al., 2004, 2008, 2009; Tremolada et al., 2005; Casellato and Erba, 2015;
84 Erba et al., 2019; Reolid et al., 2020b).

85 Venting of large quantities of greenhouse gases related to the degassing of the Karoo–
86 Ferrar Large Igneous Province (Percival et al., 2015; Heimdal et al., 2021; Ruhl et al., 2022),
87 dissociation of gas hydrates along continental margins and/or terrestrial environments (Hesselbo
88 et al., 2000; Pálffy and Smith, 2000; McElawin et al., 2005; Svensen et al., 2007; Percival et al.,
89 2015; Them et al., 2017b; Ruebsam et al., 2019), and/or input of thermogenic methane related to
90 metamorphism of organic-rich sediments (Jenkyns, 2010; Fantasia et al., 2018; Xu et al., 2018)
91 are identified as the main potential triggers of the extreme climatic conditions that existed during

92 the negative carbon-isotope excursion of the T-OAE. The overarching broad positive carbon-
93 isotope excursion is, instead, attributed to accelerated global marine and lacustrine carbon burial,
94 for which an abundant globally distributed sedimentary record exists (Jenkyns, 1988; Jenkyns,
95 2010; Fantasia et al., 2018; Xu et al., 2018; Silva et al., 2021a).

96 This study presents an in-depth analysis of literature data available for the distribution of
97 black shales and the stratigraphic variation in redox conditions inferred from both inorganic and
98 organic materials as well as from benthic fauna (i.e., bioturbation, micro- and macro-benthos),
99 during the T-OAE in the Alpine-Mediterranean Tethys, north African margin, and north European
100 epicontinental seaway. Over the years, the lack of a consistent definition of the T-OAE has resulted
101 in the identification of different stratigraphic intervals as diagnostic of the event. As a consequence,
102 many palaeoenvironmental changes and depositional processes described as having occurred
103 within the same time interval are, in fact, diachronous. It is, therefore, necessary to standardize
104 definitions and subdivisions of the T-OAE and establish a unique stratigraphic framework for the
105 analysis and correlation of the available datasets. The aim of this review is to gain a comprehensive
106 picture of temporal changes within a solid stratigraphic framework, allowing the reconstruction of
107 the surface- and bottom-water dynamics and evolution in synchronous time windows.

108

109 **2. DEFINITION AND SUBDIVISION OF THE TOARCIAN OCEANIC ANOXIC EVENT**

110 After the original identification of the T-OAE on the basis of the globally distributed and
111 apparently coeval organic-rich black shales and an accompanying broad positive carbon-isotope
112 excursion interrupted by a negative trough (Jenkyns, 1988; Jenkyns and Clayton, 1997), Boulila
113 et al. (2014) and Boulila and Hinnov (2017) identified the T-OAE as the negative CIE disrupting
114 the long-term $\delta^{13}\text{C}$ variation, and divided this interval into two segments: a lower decreasing part

115 and an upper increasing part. However, the identification of these two segments is not always
116 consistent. In fact, only the Sancerre-Couy section shows a wedge-shaped $\delta^{13}\text{C}_{\text{carb}}$ curve, while all
117 the other sections exhibit an intervening valley floor in the $\delta^{13}\text{C}_{\text{carb}}$ and/or $\delta^{13}\text{C}_{\text{org}}$ curves.
118 Therefore, the two-fold subdivision seems inappropriate. Muller et al. (2017) suggested using the
119 term Jenkyns Event as a synonym for the T-OAE and, consequently, the label Jenkyns Event was
120 applied to the entire positive carbon-isotope excursion. Muller et al. (2017) subdivided the Jenkyns
121 Event into three intervals, namely Interval 1, covering the positive plateau preceding the base of
122 the negative CIE, Interval 2 covering the decreasing part and the valley floor of the negative CIE,
123 and Interval 3 matching the rising limb of the negative CIE together with the lower part of the
124 upper plateau in various sections (Fig. 1). Thibault et al. (2018) only considered the negative CIE
125 subdivided in three intervals A, B, and C corresponding to the positive plateau below the negative
126 CIE and a part of the decreasing limb, the valley floor, and the increasing limb, respectively.
127 However, the application of this approach varies significantly from section to section and is thus
128 inconsistent. In particular, the three-fold subdivision is not appropriate for the Sancerre-Couy
129 section where the $\delta^{13}\text{C}_{\text{carb}}$ negative excursion is characterized by a wedge-shaped pattern (Fig. 1).
130 Ruebsam et al. (2019) used the term ‘Toa-CIE’ for the negative $\delta^{13}\text{C}$ excursion and identified three
131 stages: the Toa-CIE stage A being the decline in $\delta^{13}\text{C}$ values, the Toa-CIE stage B corresponding
132 to the valley floor, and the Toa-CIE stage C representing the increase to pre-excursion values.
133 Boulila et al. (2019) suggested that the T-OAE coincides with the negative CIE, although they
134 placed the base as varying from the uppermost part of the lower plateau to the base of the
135 decreasing limb (Fig. 1). Ruebsam and Al-Husseini (2020) did not specify the position of the T-
136 OAE but named the negative CIE the T-CIE and subdivided it into a falling limb, a valley and a
137 rising limb, preceded and followed by a lower and an upper plateau, respectively. Reolid et al.

138 (2020) used both the terms T-OAE and the ‘Jenkyns Event’ specifying that the former term applies
139 to marine successions with evidence of oxygen-depleted conditions while the latter has to be used
140 for the global changes that occurred during the Early Toarcian, including anoxia, enhanced
141 organic-matter burial, biotic crises in marine and terrestrial ecosystems, warming and sea-level
142 rise. However, precise definitions of the beginning and end of the T-OAE and the Jenkyns Event
143 were not provided, preventing their unambiguous identification and correlation on a regional,
144 supra-regional and global scale. Erba et al. (2022) adopted the definition given by Jenkyns (1988)
145 and used the term T-OAE for the entire overarching C isotopic positive excursion, inclusive of the
146 negative CIE in its central part (Jenkyns, 2010). The term Jenkyns Event was solely applied to the
147 negative CIE, which correlates with the ammonite uppermost *tenuicostatum* Zone–*exaratum*
148 Subzone interval (Xu et al., 2018; Storm et al., 2020) and falling within the NJT6 nannofossil Zone
149 (Ferreira et al., 2019, Visentin and Erba, 2021). Erba et al. (2022) subdivided the Jenkyns Event
150 in two isotopic segments, namely the J1 from the base of the falling limb up to the top of the valley
151 floor, and the J2 for the rising limb (Fig. 1).

152 In this paper, we adopt the definition of Jenkyns (1988) to identify the T-OAE
153 corresponding to the broad $\delta^{13}\text{C}$ positive excursion, inclusive of the negative CIE in its central
154 portion, and consider the term Jenkyns Event as a synonym for T-OAE, as initially proposed by
155 Muller et al. (2017) (Fig. 1 and Fig. S1 in the Supplementary Material). The literature survey
156 conducted on all the analysed sections shows that only three C isotopic records (Sancerre-Couy,
157 FR-210-078 Core, and Rietheim) exhibit a wedge-shaped profile of the $\delta^{13}\text{C}_{\text{carb}}$ negative CIE but
158 not of the associated $\delta^{13}\text{C}_{\text{org}}$ anomaly (Hermoso et al., 2009a; Ruebsam et al., 2014; Montero-
159 Serrano et al., 2015), possibly as a result of diagenetic imprints on the carbonate record. These
160 $\delta^{13}\text{C}_{\text{carb}}$ profiles, therefore, represent an exception while the typical negative CIE consists of an

161 initial decrease, a valley, and a final increase. Consequently, in this review we adopt the three-fold
162 subdivision of the negative CIE proposed by Ruebsam and Al-Husseini (2020) (Fig. 1). In
163 particular, stratigraphically from bottom to top, we identify within the idealized T-OAE carbon-
164 isotope profile six isotopic segments: a pre-plateau positive excursion and a pre-negative CIE
165 plateau that predates the negative CIE, a falling limb, a valley floor, and a rising limb forming the
166 negative CIE, and a post-excursion plateau. These segments are separated by inflection points that
167 mark the transition from one segment to the other. In particular, from bottom to top, we identify:
168 an onset point of the pre-plateau positive excursion coincident with the onset of the T-OAE (onset
169 of the pre-plateau positive excursion), an inflection point of the pre-negative CIE plateau (onset of
170 the pre-negative CIE plateau interval); an onset point of the negative carbon-isotope anomaly
171 (onset of the falling limb); an inflection point that marks the transition from the progressively
172 decreasing carbon-isotope values to the valley interval characterized by preferentially light $\delta^{13}\text{C}$
173 values (base of the falling limb); a point that marks the end of the valley floor and the beginning
174 of the progressive increase back to higher carbon-isotope values (onset of the rising limb); an onset
175 point of the post-excursion plateau with relatively stable $\delta^{13}\text{C}$ values (top of the rising limb); and
176 finally an inflection point that marks the end of the post-excursion plateau and the end of the T-
177 OAE (top of the post-negative CIE plateau interval).

178

179 **3. METHODOLOGY**

180 In the present study we considered only those sections with an available carbon-isotope
181 record, either inorganic or organic, and with a resolution sufficient for identifying the negative
182 carbon-isotope anomaly of the T-OAE. A total of 89 sections offer the above-mentioned
183 characteristics and were used for this review (Fig. 2 and Table 1).

184 Stratigraphically from bottom to top, the Jenkyns Event was analysed in five time intervals
185 (Fig. 1): a) a ‘pre-negative CIE plateau interval’ corresponding to the interval directly preceding
186 the onset point of the negative carbon-isotope anomaly of the T-OAE; b) an interval right above
187 the onset point of the negative CIE (‘falling limb onset’) identified by the basal part of the negative
188 CIE; c) an interval called ‘valley floor’ associated with the most negative values reached by the
189 negative carbon-isotope anomaly prior to the steady increase back to heavier $\delta^{13}\text{C}$ values, i.e. close
190 to the onset of the rising limb; d) an upper part that features the final interval of the gradual increase
191 back to pre-anomaly values (‘top rising limb’); and finally e) the interval immediately following
192 the end of the negative CIE, called ‘post-negative CIE plateau interval’.

193 For each time interval, a series of different parameters were considered in order to map the
194 regional distribution of specific properties within the Alpine-Mediterranean Tethys and in
195 basins/sub-basins of epicontinental northern Europe. The initial review of the stratigraphy
196 available for individual sections resulted in the identification of hiatuses in the interval preceding
197 the negative CIE of the T-OAE based on sedimentological, biostratigraphic and
198 chemostratigraphic criteria.

199 Based on the lithological description, we mapped the presence or absence of black shales
200 or dark grey shales within each interval, either covering part or the entire interval of interest.
201 However, it should be noted that, for the sake of readability, in those cases where both black shales
202 and dark grey shales were recorded, we report on the maps exclusively the presence of black shales.
203 Moreover, for each interval we reported, where present, the occurrence of dominant terrestrial
204 organic matter (i.e., woody fragments, coaly horizons, etc.). As far as the average total organic
205 carbon (TOC) is concerned, we distinguished six different discrete TOC classes ($\text{TOC} \leq 0.5\%$, 0.5
206 $< \text{TOC} \leq 1.0\%$, $1.0 < \text{TOC} \leq 2.5\%$, $2.5 < \text{TOC} \leq 5.0\%$, $5.0 < \text{TOC} \leq 10.0\%$, and $\text{TOC} > 10.0\%$).

207 We mapped the relative stratigraphic variation in oxygen-isotope ratios from the onset of
208 the falling limb up to the inflection point at the onset of the rising limb. Most oxygen-isotope data
209 utilized were measured on bulk carbonate, with the exception of the Portuguese Peniche section
210 (Site 9), the Spanish Barranco de la Cañada (Site 22), and the Yorkshire record in UK (Site 85)
211 where, in the absence of bulk carbonate data, measurements on belemnites, brachiopods and
212 bivalves were used. For the sake of completeness, we highlighted those records in which a
213 moderate to severe diagenetic overprint was recorded in the oxygen-isotope values. The $\Delta^{18}\text{O}$ were
214 estimated for each interval by subtracting the $\delta^{18}\text{O}$ value at the onset of the rising limb from the
215 $\delta^{18}\text{O}$ value at the onset of the falling limb, the latter used as a background value against which the
216 shift is measured. $\Delta^{18}\text{O}$ is computed as follows:

217

$$218 \quad \Delta^{18}\text{O} = \delta^{18}\text{O}_{\text{rising limb onset}} - \delta^{18}\text{O}_{\text{falling limb onset}}$$

219

220 Consequently, negative $\Delta^{18}\text{O}$ values indicate a shift to lower oxygen-isotope ratios, while positive
221 $\Delta^{18}\text{O}$ values represent a rise to higher oxygen-isotope ratios.

222 The presence or absence of benthic fauna (i.e., bioturbation, micro- and macro-benthos)
223 was divided into three major classes: a) intervals with ‘no benthic fauna’, i.e. intervals completely
224 devoid of benthic fossils and/or bioturbation, in some cases associated with the presence of well-
225 defined laminations; b) intervals with a ‘limited benthic fauna’, i.e. where benthic fossils are rare
226 to very rare, and/or bioturbation is uncommon; c) intervals where a ‘benthic fauna’ is present with
227 the persistent presence of benthic fossils and/or intense bioturbation. Moreover, for each site the
228 presence/absence of pyrite either euhedral (nodules, laminae, etc.) or in the form of framboids was

229 plotted. Where available, we also report the average dimension of pyrite framboids, separating
230 three classes: $\leq 5 \mu\text{m}$, between 5 and $10 \mu\text{m}$, and $> 10 \mu\text{m}$.

231 The redox conditions were reconstructed based both on inorganic and organic data,
232 adopting the classification of redox facies proposed by Tyson and Pearson (1991), with oxic
233 conditions associated with oxygen levels $> 2.0 \text{ ml/l}$, dysoxic conditions characterized by values
234 between 2.0 and 0.2 ml/l , suboxic conditions between 0.2 and $\sim 0 \text{ ml/l}$, anoxic when oxygen content
235 is equal to 0 ml/l , and euxinic when the complete lack of oxygen is accompanied by the presence
236 of free H_2S . As regards inorganic data, the redox conditions were distinguished into four main
237 classes based on bulk elemental data and isotopic data (e.g., Mo isotopes). In particular, based on
238 inorganic data, we distinguished: a) intervals characterized by fully oxic conditions; b) intervals
239 that reached suboxic conditions; c) intervals that experienced phases of fully anoxic conditions;
240 and d) intervals of anoxia alternated with either short or prolonged interludes of euxinia. Redox
241 conditions were classified also by using CNS data (total organic carbon (TOC), nitrogen and
242 sulphur) and, more precisely, on the basis of molecular biomarkers. Using these data we
243 distinguished oxic–suboxic intervals from those recording anoxic or anoxic–euxinic conditions.
244 The documentation of lipids derived from anaerobic photoautotrophic bacteria (e.g. chlorobactane,
245 okenone, isorenieratane) was used to map the presence of photic-zone euxinia (Summons and
246 Powell, 1986; Sinninghe Damsté et al., 1993; Ruebsam et al., 2018). For each site plotted on the
247 maps the most intense conditions experienced during the specific stratigraphic interval are
248 reported. This approach means that, for example, an interval characterized by pulses of suboxia
249 alternating with oxic conditions was classified as ‘suboxic’, and in cases of fully anoxic or even
250 euxinic conditions alternating with suboxic or oxic conditions the interval was classified as
251 ‘anoxic–euxinic’.

252 Average Mn and Fe concentrations, where available from elemental data, were reported
253 for each interval. In particular, for Mn we distinguished four classes: a) concentrations \leq 500 ppm;
254 b) between 500 and 1000 ppm; c) between 1000 and 2000 ppm; and d) concentrations $>$ 2000 ppm.
255 Moreover, the presence of Mn-nodules, Mn-hardgrounds, and general descriptions of Mn-rich
256 carbonates were plotted. As regards to Fe, seven classes of average concentrations were defined:
257 a) concentrations \leq 1%; b) between 1 and 2%; c) between 2 and 3%; d) between 3 and 4%; e)
258 between 4 and 5%; f) between 5 and 6%; and g) average concentrations $>$ 6%.

259 Different proxies may either record processes that operated at different time-scales or have
260 a different sensitivity to changes in redox conditions. For example, short-lived intervals of pore
261 water re-oxygenation could have been recorded by benthic fauna without being resolved by
262 geochemical data. Analogously, minor changes in oxygen availability could have been recorded
263 by variations in the benthic fauna without inducing any significant change in the geochemical
264 record. In order to bypass these limitations, redox conditions were estimated combining data on
265 benthic fauna and bioturbation with inorganic and organic geochemical data. Different numerical
266 values were attributed for the various discrete classes as follows: a) for benthic fauna (i.e.
267 bioturbation, micro- and macro-benthos) a value of 0 was assigned when present, 1 for limited
268 benthic fauna, and 2 for absence of benthic fauna; b) a combined redox indicator based on
269 inorganic and/or organic geochemical data where a value of 0 was assigned for oxic conditions
270 (i.e., oxic conditions on inorganic data and/or oxic–suboxic conditions on organic data), 1 for
271 suboxic conditions (i.e., suboxia on inorganic data), 2 for anoxic, and 3 for euxinic. A ‘Redox
272 Index’ (RI) was subsequently estimated combining the two separate numbers above for those sites
273 where both data on benthic fauna and inorganic and/or organic geochemical data were available.

274 Such indices can vary from 0 to 5, with 0 representing completely oxygenated conditions and 5
275 correspondig to euxinic conditions with total absence of benthic fauna.

276

277 **4. SEDIMENTOLOGICAL AND GEOCHEMICAL VARIATIONS ACROSS THE** 278 **TOARCIAN OCEANIC ANOXIC EVENT**

279 **4.1. Palaeobathymetry**

280 During the early Toarcian, the Alpine-Mediterranean Tethys was separated from the proto-
281 Atlantic Ocean by a deeper marine area branching from the Hispanic Corridor. To the north a
282 relatively shallow water epicontinental sea, connected to the Arctic Ocean by the Viking Corridor,
283 extended from about 30°N to about 50°N (Figure 2). This shelfal region was characterized by the
284 presence of relatively deeper epicontinental basins and sub-basins, such as the Cleveland Basin,
285 Paris Basin, and the Northwestern and Southwestern German basins. Sedimentological evidence,
286 such as current-bedded clastic sediments as well as the local presence of silt-grade sedimentary
287 structures in black shales, suggest generally shallow-water depths (e.g., Trabucho-Alexandre et
288 al., 2012). Water palaeodepths of about 15 to 30 m were estimated for the lower Toarcian black
289 shales in Yorkshire (Site 84), north-east England (Hallam, 1967). Frimmel et al. (2004) and Röhl
290 et al. (2001) reconstructed water palaeodepths of about 50 m and 100–150 m, respectively, for
291 coeval facies in south-west Germany. A paralic environment, ranging from fluvial to marginal
292 marine was associated with the deposition of the Bornholm record (Site 81) in the Eastern Danish
293 Basin (Hesselbo et al., 2000; Percival et al., 2015). Moving south, lower than 30°N, a series of
294 tropical carbonate systems developed both attached to continental landmasses and as isolated
295 platforms (e.g., Winterer, 1998; Woodfine et al., 2008; Sabatino et al., 2009; Ettinger et al., 2021).
296 Water depth estimates for pelagic and hemipelagic settings in the southern portion are scarce.

297 Bjerrum et al. (2001) suggested a water palaeodepth of about 200 m for the shallow westerly
298 dipping homoclinal ramp of the Lusitanian Basin at the margin of the proto-Atlantic Ocean (Site
299 9). Erba et al. (2022), on the basis of previous paleobathymetric reconstructions for the Jurassic
300 troughs in present-day Northern Italy (Bernoulli and Jenkyns, 1974; Bosellini and Winterer, 1975;
301 Bernoulli et al., 1979), estimated a palaeobathymetry of about 1000 and 1500 m water depth for
302 the Tethyan pelagic records of Gajum (Site 41) and Sogno (Site 40) in the Lombardy Basin
303 (northern Italy), respectively. These values are in agreement with estimates given by Winterer
304 (1998) for the East Sebino Basin in the easternmost part of the Lombardy Basin and the adjacent
305 Belluno Basin in the Southern Alps. Relatively great palaeodepths are implied also for the so-
306 called Kastelli Pelites (Site 53) in the Pindos Zone (northern Peloponnese, Greece), based on the
307 occurrence of radiolarites stratigraphically associated with organic-rich T-OAE shales (Kafousia
308 et al., 2011).

309

310 **4.2. Occurrence of a pre-negative carbon-isotope anomaly hiatus**

311 Evidence of a hiatus in the *tenuicostatum* Zone immediately preceding the interval of the
312 negative CIE of the T-OAE was documented in many sections (Figure 3) deposited at shallow
313 depths, with a pattern of sedimentation ranging from shallow marine to hemipelagic. In particular,
314 a major erosion surface was observed in the interval directly preceding the negative CIE in the
315 Dades Valley section (Site 4) in the Moroccan Central High Atlas (Krencker et al., 2015, 2019).
316 The shallow-marine record of the Roqueredonde section (Site 25) in the Grands Causses Basin in
317 France (Bomou et al., 2021) also shows the presence of a significant hiatus at the onset level of
318 the negative CIE, as confirmed by the absence of ammonites characteristic of the lower part of the
319 *tenuicostatum* Zone. The occurrence of a packstone unit with tempestites associated with an

320 erosional surface, overlain by sediments with chaotically organized fossils and capped by low-
321 angle ripples, marks the base of the negative CIE in the Andra Core HTM 102 (Site 32) in the Paris
322 Basin (van Breugel et al., 2006). In the Penne Château-Granier section (Site 30) (Quercy Basin,
323 France) the upper Pliensbachian to basal Toarcian (lower *tenuicostatum* Zone) bioclastic
324 carbonates are separated by a hardground from the lower Toarcian (upper *tenuicostatum* Zone)
325 “Schistes Carton” (Emmanuel et al., 2006). At the Monte Mangart section (site 35) in the Julian
326 Alps (Sabatino et al., 2009), a hardground consisting of ferromanganese oxyhydroxide nodules is
327 present at the base of the negative CIE interval. A hiatus at the base of the negative CIE has been
328 proposed by Ruebsam et al. (2020d) for the Valdorbja record (Site 42) in the Umbria-Marche
329 Basin, even if here the actual occurrence of a gap is somewhat ambiguous. A gap at the same
330 stratigraphic position has been identified also in the shelfal successions of Gipf (Site 66) and
331 Riniken (Site 67) (Fantasia et al., 2018) and at Riethem (Site 68) (Montero-Serrano et al., 2015)
332 in the Swiss Tabular Jura based on ammonite biostratigraphy (Wiedenmayer, 1980). Nannofossil
333 biostratigraphy highlighted a hiatus within the *tenuicostatum* Zone at the base of the negative CIE
334 interval in the L1 (Site 76) and Schandelah (Site 77) cores in the North German Basin (van de
335 Schootbrugge et al., 2019; Visentin et al., 2021). The hemipelagic succession of the Cuers section
336 (Site 29) in the Dauphinois Basin (France) is characterized within the *tenuicostatum* Zone by a
337 hardground at the top of a breccia overlain by marly limestones and marls within the *tenuicostatum*
338 Zone (Leonide et al., 2012). On the basis of carbon-isotope data, Pittet et al. (2014) identified a
339 hiatus in correspondence with the interval immediately below the onset level of the negative CIE
340 of the T-OAE (interval 2 in their work) in the Rabaçal (Site 11) and Ribeiro (Site 12) sections in
341 the Lusitanian Basin (Portugal). Also a few sections deposited at greater water depths exhibit a
342 hiatus: for example, the Foug Tillicht section (Site 7) in the deeper water domain of the Central

343 High Atlas Basin (Morocco) (Bodin et al., 2016), and the pelagic record of the Gajum Core (Site
344 41) in the Lombardy Basin (northwestern Italy) deposited at about 1000 m of water depth (Visentin
345 and Erba, 2021; Erba et al., 2022).

346 As suggested by Pittet et al. (2014), in offshore localities of the Lusitanian Basin, a
347 combination of a major drop of sea level and a subsequent rapid transgression probably explains
348 the observed marine erosion in the ‘pre-negative CIE plateau interval’. A rapid sea-level fall
349 preceding the negative CIE was also suggested by Krencker et al. (2019), based on the observation
350 of coeval deeply incised valleys in marine sections cropping out in Greenland and Morocco. The
351 presence of gaps before and after the negative CIE of the T-OAE has also been explored in the
352 study by Ruebsam and Al-Husseini (2020). Their research, along with investigations conducted
353 by Röhl et al. (2001), Hermoso et al. (2013), Pittet et al. (2014), Krencker et al. (2019), Ruebsam
354 et al. (2019), and Ruebsam et al. (2020c), has provided evidence indicating that these gaps coincide
355 with periods of sea-level falls and lowstands. Glacio-eustasy has been proposed as a potential
356 driver of sea-level fluctuations, with lowstands aligned with colder climatic conditions (e.g., Suan
357 et al., 2010; Korte and Hesselbo, 2011; Krencker et al., 2019; Ruebsam et al., 2019, 2020e).
358 However, while sea-level variations of few tens of metres during the Toarcian (Haq, 2017) can be
359 invoked for shallow-water settings, they cannot explain the occurrence of hiatuses in deeper water
360 settings. A supra-regional early Toarcian tectonic phase associated with the opening of the Alpine
361 and Ligurian Tethys starting from about 185 Ma (Schettino and Turco, 2011) was probably
362 responsible for gaps and massive re-sedimentation observed during the interval preceding the
363 negative CIE of the T-OAE.

364

365 **4.3. Black shales and TOC content**

366 The distribution of black shales/dark grey shales and their TOC content for the 89 studied
367 sites are reported in Figures 4 and 5, respectively. A detailed review of TOC data was recently
368 presented by Kemp et al. (2022b). Our examination of TOC data differs from theirs by illustrating
369 the greater number of sections considered in our compilation. Furthermore, while Kemp et al.
370 (2022b) presented the average TOC distribution within the negative CIE of the T-OAE, in our
371 study we captured the evolution of the TOCs before, during and after this isotopic phenomenon.

372 Available data indicate that, except for very few localities, deposition of black shales
373 during the ‘pre-negative CIE plateau interval’ did not take place (Fig. 4a). Only very few sections,
374 probably reflecting very local conditions, were characterized by the deposition of black shales over
375 this interval of time. Very dark grey to black mudstones and silty mudstones were observed in the
376 L05-04 well in the Dutch Central Graben (Site 80) (Trabucho-Alexandre et al., 2012), and black
377 shales were described by Suan et al. (2018) for the interval directly below the negative CIE of the
378 Jenkyns Event at Zázrivá in Slovakia (Site 54) at middle shelf depths. From a lithological point of
379 view, black shales associated with the ‘pre-negative CIE plateau interval’ are reported at the
380 Petousi section in the Ionian Zone in Greece (Site 51) (Kafousia et al., 2013, 2014). These shales
381 are indeed characterized by generally very low TOC values from about 0.1 to 1.8 % (Fig. 5a).
382 Relatively higher values, from about 3 to 4.5 %, were observed at Châabet El Attaris in Tunisia
383 (Site 3) (Ruebsam et al., 2022a; Reolid et al., 2023), Castrovido section in Spain (Site 20) (Danise
384 et al., 2019, and the L05-04 well in the Dutch Central Graben (Site 80) (Trabucho-Alexandre et
385 al., 2012). Relatively high average TOC values of about 3% documented in the Spanish Sierra
386 Palomera (Rambla del Salto) section (Site 21) are related to dispersed continental organic matter
387 rather than the presence of black shales (Gomez and Arias, 2010; Gomez and Goy, 2011; Danise
388 et al., 2019). Discrete black-shale intervals, i.e., Seegrasschiefer and Tafelfleins horizons, with

389 high TOC content of about 6% and 9%, respectively, are also known from the Dotternhausen
390 section (Site 72) (Schouten et al., 2000; Frimmel et al., 2004; Schwark and Frimmel, 2004;
391 Dickson et al., 2017; Wang Y. et al., 2021).

392 At the onset of the negative CIE ('falling limb onset interval'), deposition of black shales
393 became widespread from shallow- to deep-water settings throughout epicontinental northern
394 Europe and the Alpine-Mediterranean Tethys (Fig. 4b). In this stratigraphic interval, average TOC
395 values range from less than 1 % to about 10.5 % (Fig. 5b), with the highest average values from 7
396 to 10.5 % observed in the SW German Basin at Dotternhausen (Site 72) (Schouten et al., 2000;
397 Frimmel et al., 2004; Röhl and Schmid-Röhl, 2005; Dickson et al., 2017; Wang Y. et al., 2020,
398 2021), Denkingen borehole (BEB 1012) (Site 73) (Suan et al., 2015), Dormettingen (Site 74)
399 (Ajuaba et al., 2022), and Aubach section (Site 75) (Hougård et al., 2021). At Gipf (Site 66) in the
400 Swiss Tabular Jura the average value is 5.5 % (Fantasia et al., 2018, 2019b), and the FR-2010-078
401 core (Site 34) in the Paris Basin shows average values of 4.5 % and 5 %, respectively (Ruebsam
402 et al., 2022b). An average TOC value of 5.9 % occurs at Bornholm (Site 81) in the Eastern Danish
403 Basin associated with the presence of coaly horizons (McElwain et al., 2005).

404 The deposition of black shales reached its maximum extent in correspondence with the
405 lowest part of the $\delta^{13}\text{C}$ negative trough ('valley floor interval') in most basins and sub-basins from
406 northern Europe to the Alpine-Mediterranean Tethys (Fig. 4c). Generally, the highest TOC values
407 are observed in correspondence with this interval, with average values ranging from less than 0.5
408 % up to 17.5 % (Fig. 5c). The highest average values of 17.5 % were observed in the Paris Basin
409 at Bascharge (Site 33) (Hermoso et al., 2014). Very high values were measured also in the L05-04
410 borehole (Site 80) in the Dutch Central Graben with average values of 15.5 % (Trabucho-
411 Alexandre et al., 2012), in the Rijswijk-1 Core (Site 78) in the Netherlands with average values of

412 13.5 % (Dickson et al., 2017; Houben et al., 2021), and in the Yorkshire composite section (Site
413 85) in UK with average values of about 12 % (Saelen et al., 1996, 2000; McArthur et al., 2008;
414 Kemp et al., 2011; Percival et al., 2015; Dickson et al., 2017). The Tethyan section of Bächental
415 in Austria (Site 70), unlike other coeval Tethyan sections, exhibits high TOC average values of
416 about 6 % and peaks up to about 9.5 % (Neumeister et al., 2015).

417 In the highest part of the negative CIE interval ('top rising limb interval') a progressive
418 decrease in the black shale distribution is observed at a regional scale (Fig. 4d). In this interval,
419 apart from a few sections, average TOC contents decrease (Fig. 5d). Noteworthy exceptions are
420 the Andra HTM 102 record (Site 32) in the Paris Basin with a rise from an average TOC value of
421 3 % (in the lowest part of the negative carbon-isotope trough) to average TOC values of 16 % (van
422 Breugel et al., 2006) in the 'top rising limb interval', and the Hungarian Réka Valley section (Site
423 61) with a change from 5 to 7 % (Ruebsam et al., 2018). Excluding the Andra HTM 102 record,
424 characterized by very high TOC average values in the 'top rising limb interval', TOC values range
425 from less than 0.5 % up to 10.5 %. High average TOC values were measured at Dormettingen (Site
426 74) in the South German Basin with average values of 10.5 % (Ajuaba et al., 2022), in the
427 Shandelah Core (Site 77) in the North German Basin with an average TOC of 10 % (Baroni et al.,
428 2018), in the Rijswijk-1 Core (Site 78) in the West Netherlands Basin with average values of 9.5
429 % (Dickson et al., 2017; Houben et al., 2021), and in the L05-04 Core (Site 80) from the Dutch
430 Central Graben with average values of 9 % (Trabucho-Alexandre et al., 2012).

431 In correspondence with the 'post-negative CIE plateau interval', deposition of black shales
432 occurred on a local to regional scale and organic-rich facies persisted mainly in the northern
433 European epicontinental basins and sub-basins (Fig. 4e). In particular, black-shale deposition
434 continued in the Paris Basin at Sancerre (Site 31) (Hermoso et al., 2009a), Andra HTM 102 (Site

435 32) (van Breugel et al., 2006), Bascharge (Site 33) (Hermoso et al., 2014), in the SW German
436 Basin at Dotternhausen (Site 72) (Schouten et al., 2000; Frimmel et al., 2004; Schwark and
437 Frimmel, 2004; Dickson et al., 2017), Denkingen borehole (Site 73) (Suan et al., 2015), in the
438 North German Basin in the L1 (Site 76) and Schandelah (Site 77) cores (Visentin et al., 2021),
439 both characterized by medium brown shales rather than black shales *sensu stricto*, and in the
440 Cleveland Basin in Yorkshire (Site 85) (Saelen et al., 1996, 2000; Jenkyns and Clayton, 1997;
441 Hesselbo et al., 2000; Kemp et al., 2005; Dickson et al., 2017). Black-shale deposition persisted
442 also in Tethyan areas in the Petousi section in Greece (Site 51) (Kafousia et al., 2013, 2014).
443 Average TOC values during the ‘post-negative CIE plateau interval’ range from < 0.5 % to about
444 16 % (Fig. 5e). Highest values occur in the Paris Basin, in particular at the Andra HTM 102 site
445 (Site 32) with average values of 16 % (van Breugel et al., 2006), and at Bascharge (Site 33) with
446 average values of about 7.5 % (Hermoso et al., 2014). TOC values in the SW and North German
447 Basin are characterized by values of about 5 %, rather lower than the 6–10 % observed in the ‘top
448 rising limb Interval’, as for example at Dotternhausen (Site 72) (Schouten et al., 2000; Frimmel et
449 al., 2004; Röhl and Schmid-Röhl, 2005; Dickson et al., 2017; Wang Y. et al., 2020, 2021),
450 Denkingen borehole (Site 73) (Suan et al., 2015), Aubach (Site 75) (Hougaard et al., 2021), and
451 Schandelah Core (Site 77) (Baroni et al., 2018), with the the exception of the Dormettingen record
452 (Site 74) characterized by average values of 8 % (Ajuaba et al., 2022). The Yorkshire section of
453 the Cleveland Basin (Site 85) shows the same average values of about 5.5 % in both intervals
454 (Saelen et al., 1996, 2000; McArthur et al., 2008; Kemp et al., 2011; Percival et al., 2015; Dickson
455 et al., 2017). High TOC values were measured also in the Netherlands in the Rijswijk-1 Core (Site
456 78) with average values of 9 % (Dickson et al., 2017; Houben et al., 2021), and in the English East
457 Midlands Shelf at the Holwell Quarries (Site 86) where black shales are present exclusively

458 following the negative CIE with an average TOC value of 8.5 % (Caswell and Coe, 2012). In
459 contrast to other coeval Tethyan records, high TOC values were observed also in the Sachrang
460 section in Bavaria, Germany (Site 71) with average TOC values of 7.6 % (Ebli et al., 1998).

461 Overall, through the whole of the T-OAE interval, a contribution from land-derived organic
462 matter is significant in all those records from sites proximal to palaeocoastlines (Fig. 4). In
463 particular, the palynofacies composition of the sedimentary organic matter at the Boumardoul
464 n'Imazighn composite section (Site 5) and Foug Tillicht record (Site 7) in the Moroccan Atlas
465 (Bodin et al., 2016) is dominated by continent-derived particles including opaque and translucent
466 phytoclasts as well as sporomorphs. In the Portuguese section of Peniche (Site 9) organic matter
467 is dominated by small phytoclasts with the occurrence in some intervals of large wood fragments
468 (Fantasia et al., 2019a; Rodrigues et al., 2020a). Dominance of terrestrially derived palynomorphs
469 is observed also in the Spanish records of Fonte Coberta/Rabaçal (Site 11) (Rodrigues et al.,
470 2020a), La Cerradura (Site 15) and Fuente Vidriera (Site 16) (Rodrigues et al., 2019), Sierra
471 Palomera (Rambla del Salto) (Site 21) (Barrón et al., 1999), and Es Cosconar (section 4) (Site 24)
472 (Rosales et al., 2018). The organic content of the Mechowo IG 1 (Site 56), Gorzów Wielkopolski
473 IG 1 (Site 57), Suliszowice 38 BN (Site 58), Brody-Lubienia BL 1 (Site 59), Parkoszowice (Site
474 60) records, deposited in the Polish Basin at shallow water depths in an environment proximal to
475 the palaeocoastline, is made up of material that originates almost entirely from the terrestrial
476 environment (wood, cuticles, and spores), with a negligible content of marine organic matter
477 (Hesselbo and Pienkowski, 2011; Pienkowski et al., 2016). The Bornholm record (Site 81) in the
478 Eastern Danish Basin is characterized by coaly intervals deposited in a fluvial to marginal marine
479 environment (Hesselbo et al., 2000; McElwain et al., 2005). In the Mochras Farm Borehole (Site
480 82) in the Cardigan Bay Basin, UK terrestrial material with abundant macrofossil wood accounts

481 for a relatively elevated percentage of the total organic matter (van de Schootbrugge et al., 2005;
482 Xu et al., 2018). Finally, weakly laminated sediments with preserved wood were observed in the
483 English Seavington St Michael record in the Wessex Basin (Site 84) (Boomer et al., 2021).

484

485 **4.4. Variation in oxygen-isotope data**

486 The variation in the oxygen-isotope ratios ($\Delta^{18}\text{O}$) calculated from the onset of the falling
487 limb to the onset of the rising limb is reported in Figure 6. Sections characterized by evidence of
488 moderate to significant diagenetic overprint of the $\delta^{18}\text{O}$ signal are highlighted with a pinkish circle
489 on Figure 6 and reported in detail on tables in the Appendix A. Indeed, some of the analysed
490 micrites show evidence of diagenetic alteration of the $\delta^{18}\text{O}$ signal (e.g., Gomez and Goy, 2011;
491 Suan et al., 2015; Jenkyns and Macfarlane, 2021). However, the fact that long time series of
492 oxygen-isotope data from different localities show parallel trends suggest that - although absolute
493 values are untrustworthy - the trends captured by $\Delta^{18}\text{O}$ values are likely meaningful in terms of
494 palaeotemperature and/or salinity variation.

495 The most negative $\Delta^{18}\text{O}$ values are observed in the French sections of Roqueredonde (Site
496 25) and Bascharge (Site 33) with values of -13 ‰ and -11 ‰, respectively (Hermoso et al., 2014;
497 Bomou et al., 2021). In some sections, an anomalous rise in the oxygen-isotope ratios is marked
498 by positive $\Delta^{18}\text{O}$ values. The highest value is recorded in the Tunisian Châabet El Attaris section
499 (Site 3) with values of 3 ‰ (Ruebsam et al., 2022a). Positive $\Delta^{18}\text{O}$ values, from about 1 to ~ 0 ‰,
500 were observed also in the Algerian Mellala section (Site 2) (Reolid et al., 2014b), and in the
501 Portuguese and Spanish records of Figueira da Foz (Site 10), Fonte Coberta/Rabaçal (Site 11), La
502 Cerradura (Site 15), and Arroyo Mingarrón (Site 17) (Duarte et al., 2007; Reolid et al., 2014a;
503 Reolid et al., 2020a). These results probably reflect some local diagenetic trends. Apart from these

504 sections, $\Delta^{18}\text{O}$ values range from about -4.5 ‰ to about -0.5 ‰. Although diagenetic overprints
505 are to be expected in sediments of this age, common patterns at supra-regional scale seem
506 indicative of at least a partial primary signal. In particular, $\Delta^{18}\text{O}$ values ≥ -2 ‰ are mainly located
507 in the southernmost part of the studied area at latitudes lower than 30°N. On the contrary, the most
508 negative $\Delta^{18}\text{O}$ values, indicative of a greater relative decrease in $\delta^{18}\text{O}$ ratios, are mainly confined
509 to higher latitudes, in the northern European continental shelf.

510

511 **4.5. Spatio-temporal redox variations**

512 ‘Redox Indexes’ estimated for the considered sites are reported in Appendix B and plotted
513 on palaeogeographic maps in Figure 7. It should be noted that redox conditions estimated from
514 inorganic and organic geochemical data, when both present, are always in agreement. A
515 speculative distribution of redox conditions far from the locations of the investigated records is
516 also tentatively illustrated on the maps. The variation in benthic fauna, distribution of euhedral and
517 framboidal pyrite, inorganic and organic geochemical data within the various stratigraphic
518 intervals, here used to estimate the various ‘Redox Indexes’, are discussed in detail in the
519 Supplementary Material.

520 During the ‘pre-negative CIE plateau interval’ (Fig. 7a) estimated RI mainly range from 0
521 to 2, with most of the sites being characterized by values equal to 0, indicating stable well-
522 oxygenated conditions. An exception is represented by the Tunisian Châabet El Attaris record (Site
523 3) and the German Dotternhausen record (Site 72), both with a value of 4, and the L05-04 Core
524 (Site 80) with a value of 3. In particular, at Châabet El Attaris, anoxic conditions are paralleled by
525 the absence of macro-benthos and benthic foraminifera and by the occurrence of plane-parallel
526 sedimentary structures (Reolid et al., 2021b; Ruebsam et al., 2022a). At Dotternhausen, the

527 discrete TOC-rich intercalations (i.e., Seegrasschiefer and Tafelfleins horizons) lack benthic
528 communities (e.g., Röhl et al., 2001) and, as captured by organic geochemical data (Frimmel et
529 al., 2004; Schwark and Frimmel, 2004), reflect time intervals characterized by suboxic–anoxic
530 conditions. In the L05-04 Core, however, anoxia as reconstructed by elemental data, was probably
531 interrupted by brief episodes of re-oxygenation, as illustrated by common traces of bioturbation
532 within the thin beds (Trabucho-Alexandre et al., 2012).

533 During the ‘falling limb onset interval’ (Fig. 7b) most of the records from the northern
534 epicontinental basins and sub-basins are characterized by RI values of 5, indicative of euxinic
535 conditions with the complete lack of benthic fauna. Based on available data, the Dutch Central
536 Graben experienced peculiar conditions, as recorded in the F11-01 (Site 79) and L05-04 (Site 80)
537 cores. Here, even though geochemical data point to predominantly anoxic bottom waters (Houben
538 et al., 2021), the observation of reworked sediment by strong currents and bioturbation suggests
539 that oxic conditions prevailed (Trabucho-Alexandre et al., 2012). It is thus possible that, during
540 this stratigraphic interval, anoxia in the Dutch Graben was frequently interrupted by storm activity
541 and geostrophic flows capable of mixing and re-oxygenating the water-column. RI values of 5 can
542 be observed also at Valdorbia (Site 42) in the Umbria-Marche Basin, and at Zázrivá in Slovakia
543 (Site 54), thereby indicating that relatively stable euxinia was established locally away from the
544 northern shelfal area. Anoxia with lack of benthic fauna (Bomou et al., 2021) was established in
545 the French Roqueredonde record (Site 25) and continued to persist also in the Tunisian Châabet
546 El Attaris section (Site 3) (Reolid et al., 2021b; Ruebsam et al., 2022a). Less severe or less
547 continuous anoxia can be inferred for records such as the Swiss Gipf (Site 66) and Riniken (Site
548 67) successions where, even if geochemical data suggest the occurrence of anoxia, some limited
549 bioturbation is recorded (Fantasia et al., 2018, 2019b). In particular, in the Algerian Raknet El

550 Kahla (Site 1) and Mellala (Site 2) records, and at La Cerradura (Site 15) and West Rodiles (Site
551 18) in Spain the complete lack of bioturbation (Fig. S1 in the Supplementary Material) supports
552 the presence of suboxia, as suggested by geochemical data (Figs. S3 and S4 in the Supplementary
553 Material).

554 In correspondence with the ‘valley floor interval’ (Fig. 7c) RI values equal to 5 continued
555 to persist in most of the northern epicontinental basins and sub-basins and at Valdorbia in central
556 Italy (Site 42) and Zázrivá in Slovakia (Site 54), extending also to the Châabet El Attaris section
557 in Tunisia (Site 3) and the Hungarian Réka Valley (Site 64). Anoxic conditions with no benthic
558 fauna (RI equal to 4) were recorded at Roqueredonde (Site 25), Bächental (Site 70), and the
559 Mochras Farm Borehole (Site 82), indicating the extension of stable anoxic conditions to these
560 areas. On the contrary, records such as the Swiss Gipf (Site 66) and Riniken (Site 67) sections
561 were probably characterized by less continuous anoxic conditions, as suggested by the occurrence
562 of minor bioturbation (Fantasia et al., 2018, 2019b). Sedimentological data indicate that the Dutch
563 Central Graben continued to be subjected to frequent reworking by bottom currents that probably
564 fostered some re-oxygenation of bottom waters (Trabucho-Alexandre et al., 2012).

565 In correspondence with the ‘top rising limb interval’ estimated RI do not exhibit major
566 changes with respect to the ‘valley floor interval’. However, variations can be observed in some
567 records. In particular, at La Cerradura (Site 15) the RI falls from 3 to 2, associated with the
568 establishment of suboxic conditions (Reolid et al., 2014a; Rodrigues et al., 2019; Silva et al.,
569 2021b) and evidence for some limited recovery of benthic fauna (Reolid et al., 2014a; Baeza-
570 Carratalà et al., 2017; Reolid et al., 2018; Rodriguez-Tovar, 2021; Simo and Reolid, 2021). The
571 Sogno Core (Site 40) in the Lombardy Basin records a shift of RI from 3 to 0, indicating that,
572 during the ‘top rising limb interval’, conditions in this part of the basin were already fully

573 oxygenated. Similarly, the Bachental section (Site 70) records a shift of RI from 4 to 2, associated
574 with a change from anoxic conditions with no benthic fauna to dominant suboxia and the
575 occurrence of some limited bioturbation.

576 In the 'post-negative CIE interval', RI values of 5 persist exclusively in some sections
577 located in the northern epicontinental area and at Châabet El Attaris (Site 3). In the latter locality,
578 organic geochemical data (Ruebsam et al., 2022a) suggest that PZE was apparently more intense
579 at times following the negative CIE than during it. By contrast, most of the records outside of the
580 northern epicontinental shelfal area exhibit values of 0, indicating fully oxygenated conditions. An
581 exception is the La Cerradura section in Spain (Site 15), with an RI value of 1 during this
582 stratigraphic interval, likely deposited under feeble and discontinuous dysoxic conditions, as
583 suggested by minor enrichments in redox-sensitive elements (Reolid et al., 2014a; Silva et al.,
584 2021b) and evidence of bioturbation (Reolid et al., 2014a; Baeza-Carratalà et al., 2017; Reolid et
585 al., 2018; Rodriguez-Tovar, 2021; Simo and Reolid, 2021). A change to lower RI values can be
586 observed also at Zázrivá (Site 54) from 5 to 2, indicating a shift to suboxic conditions witnessed
587 by a recovery of benthic foraminifera in a depositional setting characterized by improved
588 oxygenation with transient episodes of lower oxygenation (Suan et al., 2018). A clear change to
589 lower RI values can be observed also within the northern epicontinental area, as in the Swiss
590 Tabular Jura at Rietheim (Site 68) and in the SW German Basin at Dotternhausen (Site 72), both
591 with a change in the RI from 5 to 4. In both these records, the anoxic to euxinic conditions indicated
592 by geochemical data are accompanied by the occurrence of several thin bioturbated layers with
593 moderately diverse benthic fauna (e.g., bivalves) showing a progressive moderate increase in
594 abundance and diversity (Schouten et al., 2000; Röhl et al., 2001; Bailey et al., 2003; Schwark and
595 Frimmel, 2004; Röhl and Schmid-Röhl, 2005; Montero-Serrano et al., 2015; Dickson et al., 2017;

596 Baroni et al., 2018; ; Fantasia et al., 2018; Fantasia et al., 2018; Wang Y. et al., 2020). Such
597 evidence suggests that during the ‘post-negative CIE interval’ anoxia–euxinia was less persistent
598 and frequently interrupted by phases of re-oxygenation.

599

600 **4.6. Variations in manganese and iron concentrations**

601 Manganese and iron contents within sediments are strongly dependent on variations in
602 bottom-water redox conditions (Calvert and Pedersen, 1996; Canfield et al., 1996; Wijsman et al.,
603 2001; Lyons and Severmann, 2006; Tribovillard et al., 2006; Raiswell and Canfield, 2012).
604 Changes in Mn and Fe concentrations in the investigated records across the negative CIE of the T-
605 OAE are reported in Figures 8 and 9. Additionally, in Figure 9 the presence of Mn nodules and
606 hardgrounds (i.e., Mn-oxyhydroxides), indicative of low sedimentation rates and oxidizing
607 conditions (Jenkyns, 1970, 1971), and Mn-rich carbonates, associated with reducing pore waters
608 (Calvert and Pedersen, 1993, 1996), are reported.

609 In the ‘pre-negative CIE plateau interval’ many successions are characterized by a rise of
610 Mn concentration compared to background levels (Figure 8a). A clear rise is observed in the
611 French Sancerre-Couy succession (Site 31) (Hermoso et al., 2009b, 2013), the Sogno Core in the
612 Lombardy Basin (Site 40) (Gambacorta et al., 2023), the Monte Sorgenza succession in the
613 Campania-Lucania Platform (Site 46) (Lu et al., 2010), the Croatian Adriatic Platform in the
614 Velebit-A record (Site 49) (Sabatino et al., 2013), and the German Dotternhausen record (Site 72)
615 (Bailey et al., 2003; Dickson et al., 2017; Baroni et al., 2018; Wang et al., 2020). The average Mn
616 concentrations range from < 100 ppm to > 3000 ppm (Fig. 8b), with elevated values principally
617 observed in the northern epicontinental seaway at Sancerre-Couy (Site 31) (Hermoso et al., 2009b,
618 2013), the German Sachrang (Site 71) (Ebli et al., 1998) and Dotternhausen (Site 72) (Bailey et

619 al., 2003; Dickson et al., 2017; Baroni et al., 2018; Wang et al., 2020) records, and in the
620 Schandelah Core (Site 77) in the North German Basin (Baroni et al., 2018). High average
621 concentrations of about 3300 ppm were observed also in the Tethyan Sogno Core (Site 40)
622 (Gambacorta et al., 2023). Fe-Mn hardgrounds (i.e., Mn-oxyhydroxides) are documented in the
623 ‘pre-negative CIE interval’ of the Monte Mangart succession (Site 35) (Sabatino et al., 2009,
624 2011), and at Kovk in the Slovenian Adriatic Platform (Ettinger et al., 2021). Manganese nodules
625 (i.e., Mn-oxyhydroxides) are described in the ‘pre-negative CIE interval’ also for the Tölgyhát
626 succession (Site 63) in the Hungarian Gerecse Basin (Muller et al., 2021). As for Mn, increased
627 Fe concentrations with respect to background values are reported in Figure 9a. In particular, a rise
628 is observed in the Sancerre-Couy record (Site 31) (Hermoso et al., 2009b), at Dotternhausen (Site
629 72) (Baroni et al., 2018; Wang et al., 2020), in the German Schandelah Core (Site 77) (Baroni et
630 al., 2018), and in the Dutch L05-04 Core (Site 80) (Trabucho-Alexandre et al., 2012). Fe average
631 concentrations range from less than 1 % to more than 6 %, with higher values clustered in the
632 northern epicontinental area, as in the L05-04 Core (Trabucho-Alexandre et al., 2012) and in the
633 Mochras Farm (Site 82) (Xu et al., 2018) (Fig. 9b).

634 With the onset of the negative CIE (‘falling limb onset interval’) (Fig. 8c), and across the
635 valley floor (Figs. 8d and 8e), the average Mn concentrations in the northern epicontinental basins
636 and sub-basins decrease to lower values, whereas higher concentrations are observed at sites
637 located either further south towards the open ocean, as in the Sogno Core (Site 40) (Gambacorta
638 et al., 2023), at Monte Mangart (Site 35) (Sabatino et al., 2011), and in the extremely Mn-enriched
639 succession of the Hungarian Úrkút section (Site 62) (Vetö et al., 1997; Suan et al., 2016), or close
640 to the proto-Atlantic in the Raknet El Kahla (Site 1) and Peniche (Site 9) sections (Hermoso et al.,
641 2009b; Silva and Duarte, 2015; Reolid et al., 2012; Ruebsam et al., 2020b). However, it should be

642 noted that in the ‘top rising limb interval’ some sections of the north European epicontinental sea
643 show a limited increase in Mn concentrations (Fig. 8e), such as at Sancerre-Couy (Site 31)
644 (Hermoso et al., 2009b, 2013), in the Schandelah Core (Site 77) (Baroni et al., 2018), and in
645 Yorkshire (Site 85) (Bailey et al., 2003; McArthur et al., 2008; Pearce et al., 2008; Dickson et al.,
646 2017; Baroni et al., 2018; Thibault et al., 2018; McArthur, 2019; Ramirez and Algeo, 2020a). The
647 negative CIE interval of many successions is characterized by the presence of Mn-rich carbonates,
648 such as the Dogna Core (Site 36) drilled next to the Longarone section in the Belluno Basin in
649 Northwestern Italy (Jenkyns et al., 1985; Bellanca et al., 1999), the Zázrivá record in the basin
650 within the Oravicum crustal block and the Bohemian Massif (Site 54) where common manganese-
651 rich carbonates are documented (Suan et al., 2018), and in the Austrian Bächental (Site 70) site
652 (Kodina et al., 1988; Neumeister et al. 2015, 2016; Suan et al., 2016). Fe-Mn crusts (hardground,
653 i.e., Mn-oxyhydroxides) are described from Kovk in Slovenia (Site 47) (Ettinger et al., 2021),
654 while in the Tölgyhát section in the Gerecse Basin (Site 63) the occurrence of manganese nodules
655 dispersed in a thin-layered clayey marl is reported (Muller et al., 2021). Differently from Mn, an
656 increase in Fe average concentrations is reported across the negative CIE interval in the north
657 European epicontinental seaway, while relatively low Fe values still persist outside of this area
658 (Figs. 9c, 9d, and 9e). High Fe average concentrations were observed in the north European
659 epicontinental seaway in the French record of Roqueredonde (Site 25) (Bomou et al., 2021), in the
660 Dutch Central Graben in the F11-01 (Site 79) and L05-04 (Site 80) cores (Trabucho-Alexandre et
661 al., 2012), the Cardigan Basin in the Mochras Farm Borehole (Site 82) (Xu et al., 2018), and the
662 Cleveland Basin in Yorkshire (Site 85) (McArthur et al., 2008; Baroni et al., 2018; Thibault et al.,
663 2018). Relatively high concentrations occur also in the Swiss Tabular Jura at Rietheim (Site 68)
664 (Montero-Serrano et al., 2015), the SW German Basin at Dotternhausen (Site 72) (Baroni et al.,

665 2018; Wang et al., 2020), and the North German Basin at Aubach (Site 75) (Hougaard et al., 2021).
666 Relatively high average Fe concentrations were observed also in the Alpine-Mediterranean Tethys
667 in the Austrian Bächental section (Site 70) and Sachrang site (Site 71) (Ebli et al., 1998;
668 Neumeister et al., 2016),

669 In the ‘post-negative CIE plateau interval’, with the exception of very few records, a
670 general decrease in Mn average concentrations is documented throughout the investigated area
671 (Fig. 8f). With the exception of the Sachrang record (Site 71), where only two data points suggest
672 concentrations as high as about 5900 ppm (Ebli et al., 1998), all other records exhibit average
673 concentrations lower than 2000 ppm. In particular, Mn average concentrations between 1000 and
674 2000 ppm were observed at Sancerre-Couy (Site 31) with average values of 1500 ppm (Hermoso
675 et al., 2009b, 2013), and the Sogno Core (Site 40) with average values of 1700 ppm (Gambacorta
676 et al., 2023). Similarly, a general decrease in Fe average concentrations is documented in the
677 analysed records during the ‘post-negative CIE plateau interval’ (Fig. 9f). Relatively higher values
678 continued to persist in the northern European epicontinental basins and sub-basins as opposed to
679 the Alpine-mediterranean area, although with lower values compared to those obtained for the
680 negative CIE interval. Only the Mochras Farm Borehole (Site 82) and the Yorkshire succession
681 (Site 85) exhibit relatively high average values of about 6 % (Baroni et al., 2018; Thibault et al.,
682 2018; Xu et al., 2018), while all the other records show Fe average concentrations lower than 4 %.
683

684 **5. PALAEOCEANOGRAPHIC CHANGES ACROSS THE TOARCIAN OCEANIC** 685 **ANOXIC EVENT**

686 **5.1. Evolution of redox conditions**

687 The stratigraphic variation in estimated ‘Redox Indexes’ integrated with the other data
688 presented herein shed light on the evolution of watermass conditions across the Jenkyns Event in
689 the Alpine-Mediterranean Tethys and north European epicontinental seaway. Available data
690 clearly indicate that just prior to the onset of the negative CIE (‘pre-negative CIE interval’) most
691 of the considered records were characterized by well-oxygenated bottom waters (Fig. 7a) as
692 indicated by widespread sediments hosting benthic fauna (e.g., trace fossils, bivalves, brachiopods,
693 etc.), rare deposition of black shales, and lack of small pyrite framboids (<5 μm). Anoxia was
694 confined to very local areas, such as some deeper parts of the Dutch Central Graben (RI = 3), in a
695 limited part of the SW German Basin (RI = 4), and the Tunisian Atlas (RI = 4), as illustrated by
696 black shales containing either continental or degraded marine organic matter, as suggested by
697 average values of TOC < 5 %. The common occurrence of euhedral pyrite and framboids larger
698 than 5 μm has been attributed to early to late phases of diagenesis (Love and Amstutz, 1966; Wilkin
699 and Barnes, 1996; Wilkin et al., 1996).

700 At the onset of the negative CIE (‘falling limb onset interval’) a general deterioration of
701 bottom-water oxygen conditions occurred at supra-regional scale in all the investigated sites (Fig.
702 7b). The deposition of black shales and dark grey shales became widespread, typically associated
703 with an increase in the average TOC content and HI values, although redox conditions were
704 variable at different locations. Inorganic and organic geochemical proxies and the absence of
705 benthic fauna (RI = 5) indicate that euxinia, and at times PZE, were confined to the north European
706 epicontinental basins and sub-basins (Cleveland Basin, Paris Basin, North German and SW
707 German basins). Local conditions might have hindered the establishment of stable anoxic/euxinic
708 conditions, as in the Dutch Graben, where storms and geostrophic flows frequently re-oxygenated
709 the water-column during the interval represented by the entire negative CIE (Trabucho-Alexandre

710 et al., 2012). Indeed, the widespread drastic transition to facies without bioturbation and benthic
711 fossils in those areas where anoxia was never reached during the ‘falling limb onset interval’
712 indicate that the sediments were at least affected by lowered oxygen availability in pore waters,
713 thus paralleling, albeit with lower intensity, what is observed on a supra-regional scale.

714 The most extreme intense anoxia/euxinia occurred in correspondence with the lowest point
715 in the carbon-isotope trough (‘valley floor interval’) (Fig. 7c). The geographical distribution of
716 black shales reached its maximum extent at different latitudes and palaeowater depths. In this
717 stratigraphic interval, some sections present the only black shale occurrence of the entire
718 succession, such as in the Lusitanian Basin at Peniche (Hesselbo et al., 2007). The organic matter
719 content reached its maximum in most of the records with highest values up to more than 20 %.
720 Pyrite framboids reached their smallest sizes (<5 µm), further confirming a shift to the most
721 extreme euxinic conditions in the water column, locally and temporarily extending into the photic
722 zone as indicated by biomarkers for green sulfur bacteria. Based on available data, euxinia (RI =
723 5), although mainly confined to northern European areas, occurred locally also in other basins such
724 as the Tunisian Atlas (Ruebsam et al., 2022b), the Slovakian Basin within the Oravicum crustal
725 block and the Bohemian Massif (Suan et al., 2018), and the Mecsek Basin in Hungary (Ruebsam
726 et al., 2018).

727 The latest phase of the negative CIE (‘top rising limb interval’) was characterized by
728 persistent reducing conditions (Fig. 7d). Locally, the recovery from hypoxia started during the
729 rising limb isotopic segment. In some sections, the deposition was characterized either by the
730 absence of black shales, such as in the Lombardy Basin in the pelagic Sogno and Gajum records
731 (Gambacorta et al., 2023), or a shift to lighter coloured shales as in the L1 and Schandelah cores
732 in the North German Basin (Visentin et al., 2021). Some local changes to suboxic conditions

733 occurred also in the euxinic north epicontinental basins, as in the Sancerre-Couy section (RI = 3)
734 (Hermoso et al., 2009b, 2013) of the Paris Basin. TOC values decrease both in epicontinental areas
735 and in the Alpine-Mediterranean Tethys. However, available data indicate that the recovery from
736 anoxia was not synchronous with relatively milder conditions (RI equal to 1 or 2) or even a return
737 to fully oxygenated conditions (RI = 0) occurring first in the Alpine-Mediterranean Tethys, north
738 African margin, and proto-Atlantic areas and later in the north European epicontinental basins and
739 sub-basins. As observed also by Silva et al. (2021b), the stratigraphic interval corresponding to the
740 rising limb isotopic segment (partly covered by their organic-matter preservation interval T4
741 (OAE)) is characterized by widespread deposition of organic-rich facies, thus indicating that
742 organic-carbon sequestration continued to occur also with the positive $\delta^{13}\text{C}$ trend of the Jenkyns
743 Event.

744 While the progressive recovery from extreme redox conditions in the latest part of the
745 negative CIE was in most cases subtle, a more drastic and widespread shift to more oxygenated
746 bottom waters is documented for the ‘post-negative CIE plateau interval’ (Fig. 7e). All available
747 data suggest that after the negative CIE anoxia/euxinia persisted exclusively in north European
748 epicontinental areas. With the exceptions of the Petousi section in the Ionian Zone (Kafousia et
749 al., 2013, 2014) and the Châabet El Attaris record in Tunisia (Reolid et al., 2021b), black shales
750 and dark grey shales remained confined to north European basins and sub-basins with relatively
751 high average TOC and HI values. An additional exception is represented by the Alpine-
752 Mediterranean section of Sachrang in Bavaria (Germany), where the deposition of organic-rich
753 marlstone persisted after the end of the negative CIE (Ebli et al., 1998). Apart from these areas,
754 benthic fauna fully recovered as illustrated by RI equal to 0 in most of the sections with common
755 bioturbation and high ichnodiversity (e.g., Fernandez-Martinez et al., 2021), paralleled by the

756 general absence of euhedral pyrite. Euxinia was likely confined to relatively smaller marine areas
757 of the northern epicontinental basins and sub-basins. Overall, over the ‘post-negative CIE plateau
758 interval’, sulphidic conditions became less intense and more intermittent with brief periods of
759 oxygenation that illustrate a gradual increase of oxygen availability and less stable and more
760 seasonal PZE (Schwark and Frimmel, 2004; McArthur et al., 2008; Dickson et al., 2017; Baroni
761 et al. 2018; Thibault et al., 2018; McArthur, 2019; Houben et al., 2021; Ajuaba et al., 2022). In
762 summary, as shown also by Silva et al. (2021b), enhanced organic-matter preservation, even if
763 confined to a relatively smaller region, continued to take place also during the stratigraphic interval
764 associated with the broad positive carbon-isotope excursion following the negative carbon-isotope
765 anomaly.

766 Distribution of anoxia played a major role in the preservation of organic matter during the
767 T-OAE. The variation in TOC versus the ‘Redox Indexes’ estimated for each stratigraphic interval
768 are reported in Figure 10. Lower TOC values are recorded in correspondence with the generally
769 well-oxygenated ‘pre-negative CIE interval’, while values rise in parallel with the negative CIE
770 with maximum values at the level of the ‘valley floor interval’, confirming that, as observed also
771 by Ramirez and Algeo (2020b), local controls on carbon burial were amplified during the early
772 Toarcian event. As expected, the rise in TOC content parallels the increase in the ‘Redox Index’,
773 because of the more intense reducing conditions that favoured an enhanced preservation and also
774 by the decrease in or lack of sediment reworking by benthic fauna.

775

776 **5.2. Palaeoceanographic changes across the Toarcian Oceanic Anoxic Event in the Alpine-** 777 **Mediterranean Tethys, north African margin, and north European epicontinental seaway**

778 The causes of the observed widespread deoxygenation associated with the T-OAE, and, in
779 particular, the mechanisms behind the establishment of intense anoxia/euxinia confined to north
780 European epicontinental basins and sub-basins have been a matter of debate. An intensification of
781 fresh-water runoff that favoured the establishment of a stronger pycnocline that prevented an
782 efficient circulation has been suggested to explain the formation of anoxic/euxinic conditions in
783 northern epicontinental basins (e.g., Saelen et al., 1996; Röhl et al., 2001; McArthur et al., 2008;
784 Dera and Donnadieu, 2012; French et al., 2014; Dickson et al., 2017; Ruebsam et al., 2018;
785 McArthur, 2019; Ramirez and Algeo, 2020a).

786 The significant rise in $p\text{CO}_2$ concentrations reconstructed for the early Toarcian time
787 interval (McElwain et al., 2005; Ruebsam et al., 2020a) presumably favoured an increase of
788 average global temperatures (Bailey et al., 2003; Dera et al., 2011; Korte and Hesselbo, 2011;
789 Gómez et al., 2016; Ruebsam et al., 2020d; Erba et al., 2022) and the acceleration of the
790 hydrological cycle with increased weathering and runoff (Brumsack, 1991; Jones and Jenkyns,
791 1991; Bjerrum et al., 2001; Cohen et al., 2004; Dera et al., 2009a; Hermoso and Pellenard, 2014;
792 Brazier et al., 2015; Montero-Serrano et al., 2015; Percival et al., 2015; Fantasia et al., 2018;
793 Jenkyns and Macfarlane, 2021). Using a coupled ocean–atmosphere model (Fast Ocean
794 Atmosphere Model – FOAM) Dera and Donnadieu (2012) estimated for the early Toarcian an
795 average increase of 9 cm/yr in global precipitation rates and a 3.5 cm/yr rise in mean annual
796 continental runoff. Higher volumes of fresh-water could then have entered particularly the northern
797 epicontinental basins, perhaps extending as far south as the Austro-Alpine region (Sachrang and
798 Bächental sections), thereby impacting the surface-water salinity and circulation.

799 A progressive warming of about 7–10 °C from the onset of the negative CIE to the interval
800 characterized by the most negative carbon-isotope values at the inflection point marking the onset

801 of the rising limb is well supported by TEX₈₆-based sea-surface temperatures from the Tethyan
802 region (Ruebsam et al. 2020d). By considering the typical $\Delta^{18}\text{O}_{\text{carb}}$ gradients between 0.2–0.3 ‰
803 per °C for marine carbonates (Leng and Marshall, 2004; Maslin and Dickson, 2015), a decrease in
804 $\delta^{18}\text{O}$ of about 2 ‰ is thus well explained by the shift to warmer conditions. Because oxygen
805 isotopes can be greatly affected by diagenesis, their use as a reliable proxy for temperature
806 reconstruction can be hindered (e.g., Marshall, 1992, Blanchet et al., 2012). However, the large
807 number of records considered in this review allows a supra-regional smoothing of local distortions.
808 The compilation of $\Delta^{18}\text{O}$ data presented in Figure 6 offers some interesting considerations.
809 Previous works (e.g., Saelen et al., 1996; Röhl et al., 2001; Dera and Donnadieu, 2012)
810 demonstrated a systematic offset in the oxygen-isotope composition between records from
811 northern European epicontinental basins and sub-basins and those from the Alpine-Mediterranean
812 Tethys. Records located at higher latitudes, thus presumably cooler, exhibit counterintuitively
813 lower $\Delta^{18}\text{O}_{\text{carb}}$ values than coeval records from lower latitudes (see tables in the Appendix A). In
814 particular, many records show a decrease in $\delta^{18}\text{O}$ larger than the hypothetical 2 ‰. Based on these
815 observations, local variations in salinity might have influenced the lower $\Delta^{18}\text{O}_{\text{carb}}$ record in
816 addition to temperature. In agreement with suggestions by previous authors, based both on bulk
817 carbonate and belemnite oxygen-isotope data (e.g., Saelen et al., 1996; Röhl et al., 2001; Rosales
818 et al., 2004 BEL; Dera and Donnadieu, 2012; Harazim et al., 2013), we speculate that on top of
819 the global rise in temperature, the observed lower $\Delta^{18}\text{O}_{\text{carb}}$ values reflect a strong north–south
820 gradient in salinity stratification, characterized by a progressive increase in runoff and fresh-water
821 input to the northern European epicontinental basins and sub-basins that probably reached its acme
822 in correspondence of the inflection point at the onset of the rising limb of the carbon-isotope curve
823 (Fig. 11). Following Wei et al. (2018) and Wei and Algeo (2020), a recent study by Ramirez and

824 Algeo (2020a) used the B/Ga and S/TOC ratios to reconstruct watermass salinity in the Cleveland
825 Basin (UK) and inferred a change from initially weakly brackish to strongly brackish conditions
826 during the Jenkyns Event. Although some freshening did take place, Hesselbo et al. (2020) argued
827 that the suggested degree of freshening (i.e., brackish or even freshwater conditions) proposed by
828 Ramirez and Algeo (2020a) is unrealistic. In particular, according to Hesselbo et al. (2020),
829 reconstructed palaeo-salinities are not in agreement with the presence of organisms such as
830 ammonites, benthic crinoids, belemnites, gastropods and bivalves typical of normal marine
831 conditions. However, reconstruction of the possible sources of fresh-water input is difficult.
832 Ruebsam et al. (2020c), based on organic geochemical and palynological data, suggested a massive
833 increase in sediment delivery to the north European epicontinental basins from the hinterland. The
834 presence of rivers located on emerged lands bounding the Causses and Quercy basins in France
835 and SW German Basin capable of delivering fresh waters was suggested, based on oxygen-isotope
836 data (Röhl et al., 2001; Emmanuel et al., 2006; Mailliot et al., 2009). A coeval Toarcian–Bajocian
837 subsurface succession from the North German Basin was interpreted as a river-dominated deltaic
838 system with typical delta plain (distributary deltaic channel belts and associated sheetsands) and
839 delta-front (distributary mouth bar complexes) deposits (Zimmermann et al., 2017). Abundant
840 phytoclasts in various basins were attributed to relatively proximal riverine inputs, as in the
841 Mecsek Basin in Hungary (e.g., Baranyi et al., 2016; Suan et al., 2018; Rodrigues et al., 2020a, b,
842 2021). As suggested by Baroni et al. (2018), continental runoff could have been paralleled by a
843 southward flow through the Viking Corridor of brackish waters derived from the Arctic (Bjerrum
844 et al., 2001) that could have reached northern epicontinental areas (Dera et al., 2009b; Korte et al.,
845 2015) (Fig. 11).

846 Lowered surface-water salinities would have favoured the formation of a stable pycnocline
847 that reduced the circulation efficiency, thereby inducing bottom-water deoxygenation. However,
848 even if a common deterioration in bottom-water oxygenation can be observed at supra-regional
849 scale, estimated redox conditions reported in Figure 7 demonstrate that anoxia/euxina was not
850 equally distributed. The physiography of the different basins and sub-basins might have played a
851 major role in controlling the extent and intensity of bottom-water anoxia. The relatively open
852 Alpine-Mediterranean Tethys and the regions close to the proto-Atlantic, with the exception of
853 some local settings, were generally affected by relatively less severe redox conditions. The Austro-
854 Alpine localities (i.e., Bächental, in particular, and Sachrang sections) are indeed anomalous
855 compared with other Alpine-Mediterranean facies. Unlike other coeval Tethyan sections, these
856 records are unusual in having relatively high TOC content, as well as Mn-rich carbonates (Ebli et
857 al., 1998; Neumeister et al., 2015, 2016; Suan et al., 2016). The establishment of the peculiar redox
858 conditions developed in this area were probably controlled by the relative proximity of this
859 region to the epicontinental margin, which, further enhanced by local physiographic conditions,
860 favored the influence of a possible fresh-water input from the north. Unlike the Alpine-
861 Mediterranean Tethys and north African margin, the articulated submarine physiography of the
862 northern European area with shallow-water depths of about 15–150 m (Hallam, 1967; Röhl et al.,
863 2001; Frimmel et al., 2004) and poor communication with the rest of the Tethys Ocean was likely
864 favourable to the establishment of a relatively sluggish circulation (Remirez and Algeo, 2020b).
865 Geochemical data are consistent with this interpretation. Changes in the Mo/TOC ratio in records
866 from northern epicontinental basins, a proxy for reconstructing palaeohydrographic conditions
867 (Algeo and Lyons, 2006; Algeo and Rowe, 2012), indicate aqueous Mo drawdown connected to
868 severe watermass stagnation (e.g., McArthur et al., 2008; Pearce et al., 2008; Hermoso et al., 2013;

869 Dickson et al., 2017; McArthur, 2019; Ramirez and Algeo, 2020a, b; Chen et al, 2021; Houben et
870 al., 2021; Wang et al., 2021). The distribution of gammacerane, an organic geochemical proxy
871 used to infer changes in water-column stratification, commonly related to hypersalinity (Sinnighe
872 Damsté et al., 1995), is in agreement with a model of watermass restriction mainly confined to
873 northern epicontinental basins and sub-basins. Evidence of gammacerane is reported for only a
874 limited number of sites and, with the exception of the peculiar Austro-Alpine region (Neumeister
875 et al., 2015), never thus far described in sections from the Alpine-Mediterranean Tethys and north
876 African area. On the contrary, gammacerane is reported for the north European epicontinental
877 shelf, with major concentrations reached in the ‘valley floor interval’ (Farrimond et al., 1989;
878 French et al., 2014; Ruebsam et al., 2018; Xu et al., 2018; Ajuaba et al., 2022). According to Baroni
879 et al. (2018), the onset of bottom-water anoxia in the north European basins and sub-basins was
880 amplified by large-scale ocean dynamics. Based on the results of FOAM models, they suggested
881 the presence of a clockwise gyre over the Tethys Ocean capable of bringing oxygenated waters
882 from the Equator. According to their models, due to the particular articulated physiography of the
883 northern epicontinental seaway the northward limb of the gyre was significantly weakened,
884 thereby making this region highly sensitive to continental runoff and stratification. Indeed,
885 available data suggest that deposition of organic matter during the Jenkyns Event was mainly
886 controlled by enhanced preservation favoured by persistent anoxia/euxinia with relatively modest
887 flux rates of organic carbon but low supply of accompanying clastic and carbonate materials,
888 thereby leading to stratigraphic condensation (Kemp et al., 2022b; Ruebsam et al., 2022b).

889 The recovery from anoxia occurred diachronously, with an earlier progressive re-
890 oxygenation in the Alpine-Mediterranean Tethys and in the north African margin than in the north
891 European epicontinental seaway. The establishment of less intense reducing conditions started

892 during the rising limb isotopic segment. This timing is in agreement with global ocean redox
893 trends, estimated by means of rhenium and molybdenum mass balance models, which depict a
894 contraction of anoxia–euxinia before the end of the negative CIE (Kunert and Kendall, 2023).
895 Geochemical data indicate that, with the end of the negative CIE, salinities moved progressively
896 from strongly brackish back to weakly brackish (Remirez and Algeo, 2020a). The gradual cooling
897 that accompanied the end of the negative CIE (Gambacorta et al., 2023) could have reduced
898 weathering and runoff, thereby favouring the weakening of the pycnocline and the re-
899 establishment of a more efficient circulation (Gambacorta et al., 2023) and a progressive
900 termination of the event. van de Schootbrugge et al. (2020) interpreted the change in the
901 dinoflagellate association from offshore Norway and the Yorkshire Coast (UK) as evidence of an
902 enhanced connectivity between the Arctic and the Tethys through the Viking Corridor as the result
903 of a relative sea-level rise. According to their interpretation, the arrival of low-salinity and
904 oxygenated Arctic waters favoured the shift and establishment of dysoxic or even oxic conditions.
905 However, this interpretation is not in agreement with the available sedimentologic and
906 geochemical evidence, and a greater fresh-water input from the Arctic would have more likely
907 favoured an intensification of watermass stratification rather than the re-establishment of more
908 oxygenated bottom-water conditions.

909

910 **5.3. Variations in manganese and iron pools during the Toarcian Oceanic Anoxic Event**

911 Lower Toarcian manganese-rich carbonates, with Mn contents reaching in some places
912 very high concentrations of up to more than 20 %, associated with organic-rich deposits have been
913 widely described in the literature (e.g., Jenkyns et al., 1985, 1991; Jenkyns, 1988; Neumeister et
914 al., 2016; Suan et al., 2016). However, their origin is still a matter of debate in part due to the

915 stratigraphic control that may not be sufficiently precise to understand the possible causal link
916 between the T-OAE and the observed enrichments. Indeed, most Mn-rich carbonates occur in the
917 Alpine-Mediterranean Tethys, in particular in the Italian Lombardy Basin (e.g., Sogno Core)
918 (Gambacorta et al., 2023), Belluno Basin (Dogna Core) and associated outcrop (Bellanca et al.,
919 1999; Jenkyns et al., 2001), in the Julian Alps (Monte Mangart section) (Sabatino et al., 2009),
920 and in the Austro-Alpine area (Bächental and Sachrang sections) (Germann, 1973; Ebli et al.,
921 1998; Neumeister et al., 2015, 2016; Suan et al., 2016) and the Bakony Mountains of western
922 Hungary (e.g., Jenkyns et al., 1991; Polgári et al., 1991).

923 The major sources of manganese to the ocean are riverine inputs and, to a lesser extent,
924 hydrothermal activity (Corbin et al., 2000; Tribovillard et al., 2006; van Hulten et al., 2017). The
925 sedimentary geochemistry of manganese is strongly affected by changes in bottom-water redox
926 conditions (Calvert and Pedersen, 1996; Tribovillard et al., 2006) (Fig. 12). Manganese, present in
927 low concentrations in the dissolved phase, is scavenged as Mn(IV) in particulate Mn-
928 oxyhydroxides (Calvert and Pedersen, 1993, 1996). Under reducing conditions, however,
929 oxyhydroxides are dissolved and Mn reduced to soluble Mn(II) that can diffuse either upward or
930 downward within the sediments (Brumsack, 1986, 1989; Rajendran et al., 1992; Tribovillard et
931 al., 2006). Upward diffusion of Mn(II) can lead either to the accumulation of Mn-oxyhydroxides
932 in oxic pore waters or to the escape back into the water column if anoxic pore waters extend up to
933 the sediment-water interface (Calvert and Pedersen, 1996; Cruse and Lyons, 2004). By contrast,
934 the downward diffusion of Mn(II) can lead to MnCO₃ precipitation, a process that, with the
935 exception of silled anoxic basins, becomes efficient under reducing pore waters but oxic bottom
936 waters (Pedersen and Price, 1982; Calvert and Pedersen, 1993, 1996; Morford et al., 2001;
937 Tribovillard et al., 2006). Generally, in the case of a poorly efficient fixation of Mn in carbonates,

938 sediments are generally depleted in Mn under reducing conditions (Hild and Brumsack, 1998;
939 Tribovillard et al., 2006). So, even if Mn-oxyhydroxides and Mn-carbonates are precipitated under
940 opposite redox conditions in pore waters, their presence in open-marine settings suggests in both
941 cases the occurrence of a surplus in Mn associated with oxic bottom-water conditions.

942 Higher fluxes of hydrothermally derived Mn could have been behind these enrichments in
943 those areas close to the ridges of the Tethys Ocean (Jenkyns et al., 1991). Local hydrothermal
944 sources of Mn were probably important in the Hungarian Úrkút deposit (Polgári et al., 1991, 2012),
945 otherwise redox conditions were likely the governing factor responsible for the observed Mn
946 enrichments. Such an origin is consistent with the spatial distribution of high Mn concentrations
947 in the Lower Toarcian (Fig. 8). However, additional contributions from rivers should be taken into
948 consideration. Fresh water associated with the increased runoff probably sourced additional
949 amounts of Mn to shelfal and basinal areas (Gambacorta et al., 2023). However, as there is little
950 evidence of rivers in the Alpine-Mediterranean Tethys, probably most of the Mn that reached this
951 area was exported from epicontinental northern Europe. On top of these contributions to the Mn
952 budget, an additional input from Arctic waters from the Viking Corridor cannot be excluded. In
953 fact, due to the specific hydrographic configuration, the watermasses of the modern Arctic are
954 enriched in Mn derived from rivers or coastal erosion (März et al., 2011; Macdonald and Gobeil,
955 2012; Meinhardt et al., 2016a, 2016b). A similar process could have occurred also in Toarcian
956 times, thereby transporting Mn-enriched waters from the Viking Corridor to northern
957 epicontinental areas. Manganese delivered to the anoxic/euxinic north European epicontinental
958 basins and sub-basins would both have remained in solution and, when not fixed into Mn-
959 carbonates, diffused from the sediment-water interface back to the water column. The
960 epicontinental basins and, to a minor extent, the other minor anoxic/euxinic basins, likely acted as

961 pools of dissolved manganese. Some of the dissolved Mn may have been reprecipitated in the Mn
962 pool as Mn-carbonates or may have been transported from epicontinental basins into the Tethys
963 Ocean where, upon encountering oxygenated conditions, it would have been deposited as Mn-
964 oxyhydroxides or Mn-carbonates, according to the prevailing redox conditions (Fig. 13).
965 Theoretically, as proposed in the idealized ‘bathtub ring’ model proposed by Force and Cannon
966 (1988), Mn-rich sediments would then have formed all along the edge of the Mn-pool adjacent to
967 the more oxygenated Tethys Ocean and in the northern part of the Adria microplate. However,
968 available data, even if scattered, indicate that observed Mn-rich deposits do not form a continuous
969 ring around the anoxic/euxinic north European epicontinental basins and sub-basins, but are
970 limited to the southeast margin (i.e., Belluno Basin, Lombardy Basin, Julian Basin, Austro-Alpine
971 nappes, Hungary) of the Alpine-Mediterranean Tethys and to the northern part of the Adria
972 microplate. Some hypotheses can be advanced to explain the observed distribution. First, it is
973 possible that truly oxic conditions were encountered only at the edge with the oxygenated Tethys
974 Ocean while, in the rest of the surrounding areas, redox conditions were not favorable for the
975 deposition of Mn. Otherwise, a major control by basin physiography could lie at the heart of the
976 observed trends. In particular, the outflow of large amounts of dissolved Mn could have been
977 favoured only at the edges of the anoxic/euxinic basins where sills were sufficiently deep to allow
978 the Mn-rich waters to pass through and exported elsewhere to be ultimately deposited, depending
979 on the diagenetic pathway, as Mn-oxyhydroxides or Mn-carbonates. In present-day oceans,
980 oxygen minimum zones are enriched in Mn (Calvert and Pedersen, 1993, 1996; Brumsack, 2006).
981 As shown in some models proposed (Jenkyns, 1985, 1988; Jenkyns et al., 1991), oxygen-minima
982 could have extended southwards during the T-OAE acting as a conveyor belt of divalent Mn from
983 the north European epicontinental seaway to the Alpine-Mediterranean Tethys.

984 Interesting considerations also arise from reconstructed stratigraphic and palaeogeographic
985 variations in Fe concentrations. Iron is delivered to the oceans mainly from rivers, aeolian dust and
986 hydrothermal fluxes (Raiswell and Canfield, 2012). Under oxic conditions, Fe(III) is the
987 thermodynamically stable geochemical species that tends to form mainly nanoparticulate Fe-
988 oxyhydroxides. Conversely, Fe(II) is stable under anoxic conditions and either diffuses from
989 reduced pore waters or forms FeCO_3 and Fe-sulfides (Raiswell and Canfield, 2012). Iron is
990 recycled from shallow shelf sediments, either transported as Fe oxyhydroxides with oxic seawater
991 towards the basin or diffused from pore waters, and laterally transported to adjacent marine basins
992 (Fe shuttling) (Canfield et al., 1996; Wijsman et al., 2001; Lyons and Severmann, 2006; Dellwig
993 et al., 2010; Raiswell and Canfield, 2012; Lenstra et al., 2019). In the presence of sulfidic (euxinic)
994 waters mobilized Fe can be precipitated and trapped in the form of Fe sulfides (Lyons and Berner,
995 1992; Canfield et al., 1996; Lyons and Severmann, 2006; Raiswell and Canfield, 2012) (Fig. 9).
996 Consequently, in an opposite way to Mn, sediments deposited under reducing conditions are
997 commonly enriched in Fe.

998 Relatively high Fe concentrations are observed in the north European epicontinental basins
999 and sub-basins (Fig. 9). Greater amounts of Fe were likely delivered to shelfal areas by the
1000 enhanced riverine supply, and then trapped into these basins as Fe sulphides. Hydrothermal fluxes
1001 from spreading ridges likely constituted an additional Fe source to Tethyan areas, although the
1002 generally mild redox conditions and open physiography did not allow accumulation of iron in great
1003 quantities.

1004 Even if Mn and Fe data at high-resolution are still relatively limited, some common patterns
1005 can be outlined and a hypothetical evolution of Mn and Fe sketched out for the early Toarcian
1006 (Fig. 14). In most records, the ‘pre-negative CIE interval’ is characterized by a rise in Mn and Fe

1007 content (Figs. 8a, 9a) with respect to background values (Fig. 14b). In this stratigraphic interval,
1008 due to the prevailing oxic conditions, Mn diffusion from sediments was very limited and Mn and
1009 Fe accumulated mainly as oxyhydroxides. With the onset of anoxia/euxinia Mn transported to the
1010 epincontinental basins was rapidly remobilized, leaving Mn-depleted deposits (Fig. 8c) as a
1011 consequence of the dominance of Mn diffusion back into the water column. Part of this Mn could
1012 potentially have flowed out of the basins and reprecipitated forming a rim of Mn-oxyhydroxides
1013 and Mn-carbonates at the contact with the better oxygenated waters of the Tethys Ocean, thus
1014 representing an additional contribution to the Mn sourced from hydrothermal ridges. By contrast,
1015 Fe was trapped as Fe-sulphide inside the anoxic/euxinic restricted basins (Figs. 9c, 14c). On a
1016 smaller scale, a similar process probably operated locally in some basins in the Alpine-
1017 Mediterranean Tethys (Figs. 8 and 9) as highlighted by the enrichments in Mn and Fe in these
1018 areas. Similarly, local suboxic to anoxic conditions and higher amounts of riverine-derived Mn
1019 and Fe favoured the fixation of Fe-sulphides and the remobilization of Mn and its reprecipitation
1020 downcurrent under oxic conditions. Such processes persisted throughout the interval of the
1021 negative CIE (Figs. 14d and 14e). With the end of the negative CIE ('post-negative CIE plateau
1022 interval'), the relative weakening of runoff could have resulted in lowered amounts of riverine-
1023 supplied Mn and Fe, and the consequent decrease in Mn and Fe concentrations (Figs. 8f, 8f, and
1024 14f). The establishment of less intense and intermittent sulphidic conditions in northern
1025 epicontinental basins probably decreased the fixation of Fe into sulphide, although the
1026 environment remained unfavourable for the fixation of Mn in oxyhydroxides. However, it should
1027 be noted that some sites record a rise in Mn concentration right after the negative CIE, such as at
1028 Sancerre-Couy in the Paris Basin (Hermoso et al., 2009b), in the Schandelah Core in the North
1029 German Basin (Baroni et al., 2018), and in the Cleveland Basin, Yorkshire (UK) (Baroni et al.,

1030 2018). Probably these local rises in Mn concentrations represent either intervals of transient oxic
1031 conditions favourable to the massive fixation of Mn-oxyhydroxides and Mn-carbonates, or a
1032 reduction in the outflow of Mn with the prevalent fixation into Mn-carbonates inside the North
1033 European epicontinental Mn-rich pool.

1034

1035 **6. CONCLUSIONS**

1036 With this review we provide a detailed temporal and palaeogeographical characterization
1037 of the T-OAE in the Alpine-Mediterranean Tethys, north Africa and north European epicontinental
1038 basins. From this study, we conclude that:

- 1039 1. In agreement with suggestions made by other authors in the past, with the exclusion of only
1040 three $\delta^{13}\text{C}_{\text{carb}}$ isotopic records characterized by a wedge-shaped profile, it is always
1041 possible to apply a three-fold subdivision of the negative CIE. In particular, from bottom
1042 to top, six isotopic segments can be recognized within the T-OAE isotopic excursion: a
1043 pre-plateau positive excursion and a pre-negative CIE plateau that predates the negative
1044 CIE, a falling limb, a valley floor, and a rising limb forming the negative CIE, and a post-
1045 negative CIE plateau.
- 1046 2. In the stratigraphic interval immediately preceding the negative CIE of the T-OAE bottom
1047 waters were generally well-oxygenated, with anoxia confined to very local areas, such as
1048 some deeper parts of the Dutch Central Graben, in a limited part of the SW German Basin,
1049 and locally in the Tunisian Atlas and Moroccan High Atlas.
- 1050 3. A synchronous shift to poorly oxygenated bottom waters correlates with the onset of the
1051 negative CIE and remained prevalent during this interval, with widespread deposition of
1052 organic-rich black shales on a supra-regional scale. Euxinia was mainly confined to the

1053 north European epicontinental basins and sub-basins, while a general decrease in oxygen
1054 availability occurred in certain basins of the Alpine-Mediterranean Tethys.

1055 4. The most extreme palaeoceanographic conditions were reached in the core of the negative
1056 CIE when organic-matter preservation, and probably stratigraphic condensation, reached
1057 their maximum extent.

1058 5. The widespread recovery to better oxygenated conditions started immediately after the end
1059 of the negative CIE, although the onset of improved bottom-water oxygenation is locally
1060 documented just before its end. With the end of the negative CIE, deposition of black shales
1061 and dark grey shales remained confined to north European basins and sub-basins. Apart
1062 from this area, excluding very few local exceptions, well-oxygenated conditions were fully
1063 re-established in the Alpine-Mediterranean Tethys and North African margin, as
1064 documented by the presence of benthic fauna, common bioturbation and high
1065 ichnodiversity, in agreement with geochemical data. At this time, the north European
1066 epicontinental seaway experienced brief periods of oxygenation with anoxia likely
1067 confined to more limited areas characterized by less intense and intermittent sulphidic
1068 conditions associated with more seasonal photic-zone euxinia. Evidently, while the onset
1069 of anoxia was a synchronous event, its termination was earlier in the Alpine-Mediterranean
1070 Tethys than in the north European epicontinental basins.

1071 6. The presented compilation of $\Delta^{18}\text{O}_{\text{carb}}$ data clearly demonstrates, as observed by other
1072 authors in the past, a salinity-dependent systematic offset between the northern European
1073 epicontinental basins and sub-basins and Alpine-Mediterranean Tethys. Oxygen-isotope
1074 data show that the shift to lower ratios from the onset of the falling limb to the onset of the
1075 rising limb is different from section to section. In particular, higher $\Delta^{18}\text{O}_{\text{carb}}$ amplitudes are

1076 recorded at sites located in the northern epicontinental area, indicating that temperature
1077 cannot be the primary control behind the observed signal. We interpret these data as the
1078 evidence of a progressive increase in runoff and/or in Arctic fresh-water input to the
1079 northern European epicontinental basins and sub-basins that reached its acme in the core
1080 of the negative CIE. The increase in fresh-water flow coupled with the closed physiography
1081 of the north European epicontinental basins and sub-basins promoted the formation of a
1082 stronger pycnocline. The limited water exchange with the Tethys Ocean favoured the onset
1083 of anoxia/euxinia into these confined basins and the enhanced preservation of organic
1084 matter.

1085 7. The limited exchange with surrounding areas, the proximity to the input of riverine-sourced
1086 Mn and Fe, and the peculiar redox conditions of the northern epicontinental basins and sub-
1087 basins favoured their role as pools for dissolved manganese that diffused into the water
1088 column, and the intense accumulation of iron into iron sulphides. We speculate that a part
1089 of the large amount of dissolved manganese present in epicontinental basins and sub-basins
1090 was exported and deposited at the interface with the more oxygenated waters of the Tethys
1091 Ocean. This input of manganese could have represented an additional contribution to the
1092 formation of the manganese-rich carbonates observed in correspondence with the T-OAE
1093 in the Alpine-Mediterranean region.

1094

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1101

1102 **APPENDICES**

1103 Data relative to variations in benthic fauna, distribution of euhedral and framboidal pyrite,
1104 inorganic and organic geochemical data are presented and discussed in detail in the Supplementary
1105 Material. Appendix A consists of spreadsheets reporting data and respective bibliographic sources
1106 used to produce the maps presented in this work. Separate spreadsheets are available for the
1107 intervals deposited below the negative CIE (pre-negative CIE plateau), onset of the negative CIE
1108 (falling limb onset), around the lowest point in the trough of the negative carbon-isotope anomaly
1109 (valley floor), the final interval of the gradual increase back to pre-anomaly values (top rising
1110 limb), and right above the negative CIE (post-negative CIE plateau). Appendix B consists of a
1111 spreadsheet summarizing all the available data and computed ‘Redox Indexes’ for each
1112 stratigraphic interval.

1113

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1872 deltaic system in the North German Basin: Subsurface mapping of ancient deltas-morphology,
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1874 **FIGURE AND TABLE CAPTIONS**

1875 Figure 1 – Definitions and subdivisions proposed for the Toarcian Oceanic Anoxic Event
1876 (T-OAE). Schematic $\delta^{13}\text{C}$ reference curve for the latest Pliensbachian–Toarcian time interval is
1877 modified after Ruebsam and Al-Husseini, 2020. Ammonite biozones are after Elmi et al. (1997),
1878 Macchioni (2002), Page (2003), and Nikitenko et al. (2008). Calcareous nannofossil biozones are
1879 after Ferreira et al. (2019) and Visentin and Erba (2021). The vertical dashed lines represent the
1880 variability in the extent of the subdivisions. Definitions and subdivisions adopted in this study, and
1881 relative stratigraphic position of intervals analysed are reported in the bottom right of the figure.
1882 Inflection points, from bottom to top: 1. onset of the pre-plateau positive excursion; 2. onset of the
1883 pre-negative CIE plateau interval; 3. onset of the falling limb; 4. base of the falling limb; 5. onset
1884 of the rising limb; 6. top of the rising limb; 7. top of the post-negative CIE plateau interval.
1885 Negative carbon-isotope excursion: negative CIE; DP-CIE: decreasing part of the negative carbon-
1886 isotope excursion; IP-CIE: increasing part of the negative-carbon isotope excursion.

1887 Figure 2 – Palaeogeographic map of Alpine-Mediterranean Tethys and north European
1888 epicontinental basins and sub-basins during the Toarcian (modified after Ruebsam et al., 2018)
1889 and relative position of locations with known record of the negative carbon-isotope excursion of
1890 the T-OAE. CB: Cleveland Basin; PB: Paris Basin; NWGB: Northwestern German Basin; SWGB:
1891 Southwestern German Basin.

1892 Table 1 – List of sites considered in the study with the relative numerical code used in the
1893 text and figures for their identification.

1894 Figure 3 – Palaeogeographic position of sites with the evidence of a hiatus in the interval
1895 directly preceding the negative carbon-isotope excursion of the T-OAE ('pre-negative CIE plateau
1896 interval'). See Appendix A for data used to produce this map.

1897 Figure 4 – Palaeogeographic distribution of black shales and dark grey shales across the T-
1898 OAE. Sites dominated by terrestrial organic matter (woody fragments, coaly horizons, etc.) are
1899 highlighted with a yellow circle. See Figure 1 for identification of different stratigraphic intervals
1900 and Appendix A for data used to produce this map.

1901 Figure 5 – Palaeogeographic distribution of average total organic carbon (TOC) content
1902 across the T-OAE. Locations without TOC data are not reported on maps. See Figure 1 for
1903 identification of different stratigraphic intervals and Appendix A for data used to produce this map.

1904 Figure 6 – Palaeogeographic distribution of $\Delta^{18}\text{O}$ values, computed as the difference in the
1905 oxygen-isotope composition at the onset of the rising limb and at the onset of the falling limb of
1906 the negative carbon-isotope excursion of the T-OAE. Locations without $\delta^{18}\text{O}$ data are not reported
1907 on maps. See Appendix A for data used to produce this map. Isotopic records characterized by a
1908 moderate to significant diagenetic overprint are highlighted with a pinkish circle.

1909 Figure 7 – Palaeogeographic distribution of prevailing dysoxic–anoxic conditions and
1910 euxinic conditions in the Alpine-Mediterranean Tethys and north European epicontinental basins
1911 and sub-basins across the T-OAE. Redox conditions at each site are estimated on the basis of
1912 Redox Indexes computed by combining data on benthic fauna and bioturbation with inorganic and
1913 organic geochemical data. The areal extent of redox conditions far from available records is
1914 speculative. See text for details.

1915 Figure 8 – Palaeogeographic distribution of Mn nodules, Mn-hardgrounds, Mn-rich
1916 carbonates (indicated with different colour borders), and Mn elemental concentrations (reported
1917 using different colour fillings) across the T-OAE. Locations without Mn data are not reported on
1918 maps. See Figure 1 for identification of different stratigraphic intervals and Appendix A for data
1919 used to produce this map.

1920 Figure 9 – Palaeogeographic distribution of Fe elemental concentrations across the T-OAE.
1921 Locations without Fe data are not reported on maps. See Figure 1 for identification of different
1922 stratigraphic intervals and Appendix A for data used to produce this map.

1923 Figure 10 – Cross-plots of average total organic carbon (TOC) versus estimated redox
1924 conditions. Sites dominated by terrestrial organic matter (woody fragments, coaly horizons, etc.)
1925 are reported in red. For site identification and palaeogeographic location refer to Table 1 and Figure
1926 1. See text for details.

1927 Figure 11 – Hypothetical fresh-water input of Arctic water masses and continental runoff
1928 into the epicontinental basins and sub-basins. See text for details.

1929 Figure 12 – Manganese and Fe redox cycling. On the left: oxic bottom and pore waters
1930 favour the oxidation of Mn and Fe with the deposition of Mn and Fe oxides and hydroxides.
1931 Reducing pore waters conditions occurring within the sediments below the chemocline favour
1932 instead the fixation of Mn in carbonates and of Fe in both carbonates and sulphides. On the right:
1933 under poorly-oxygenated bottom water conditions (i.e., with the chemocline located in the water
1934 column above the sediment-water interface) reduced Mn and Fe are either fixed within the
1935 sediments in the form of Mn-carbonates, Fe-carbonates, and Fe-sulphides, or diffused back to the
1936 water column as divalent Mn and Fe.

1937 Figure 13 – Manganese cycling in the Alpine-Mediterranean Tethys and north European
1938 epicontinental basins and sub-basins during the T-OAE. See text for details.

1939 Figure 14 – Scheme representing the temporal and palaeogeographic variation in Mn and
1940 Fe in north European epicontinental basins and sub-basins across the T-OAE. See text for details.

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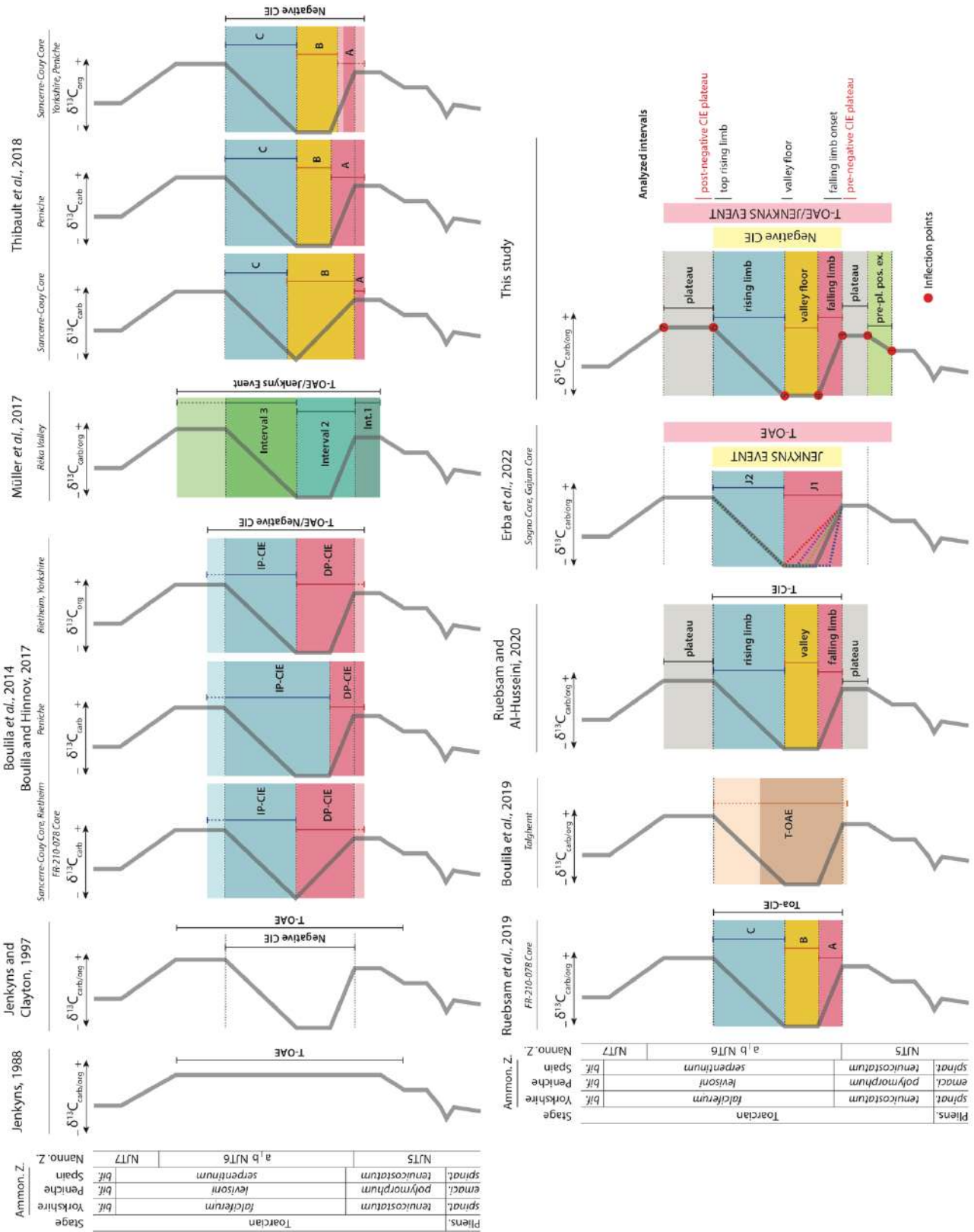
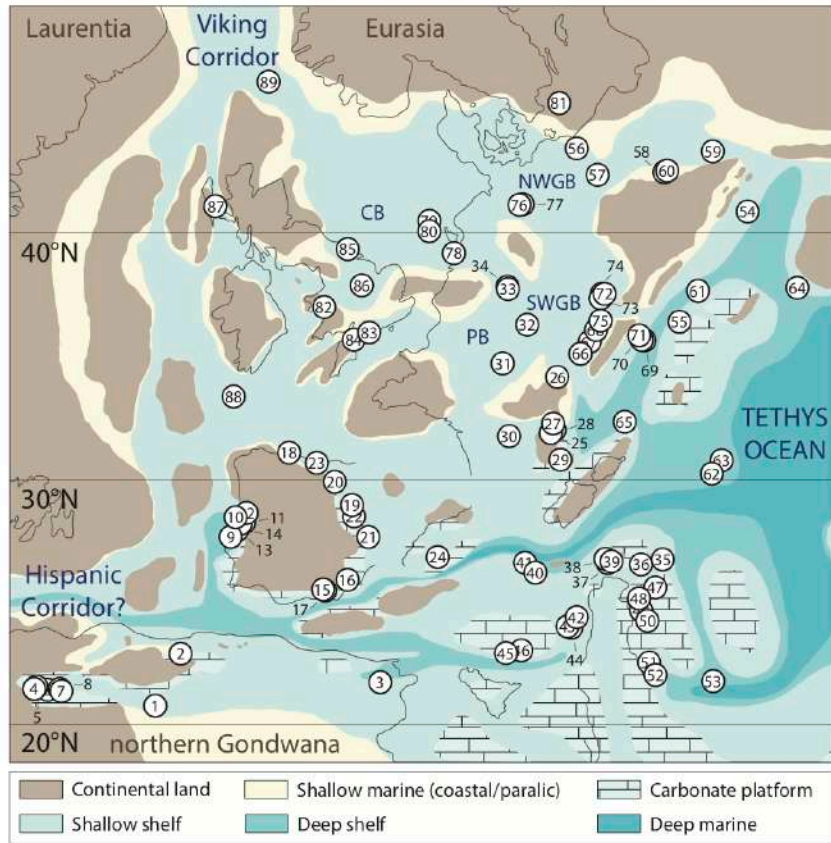


Figure 1



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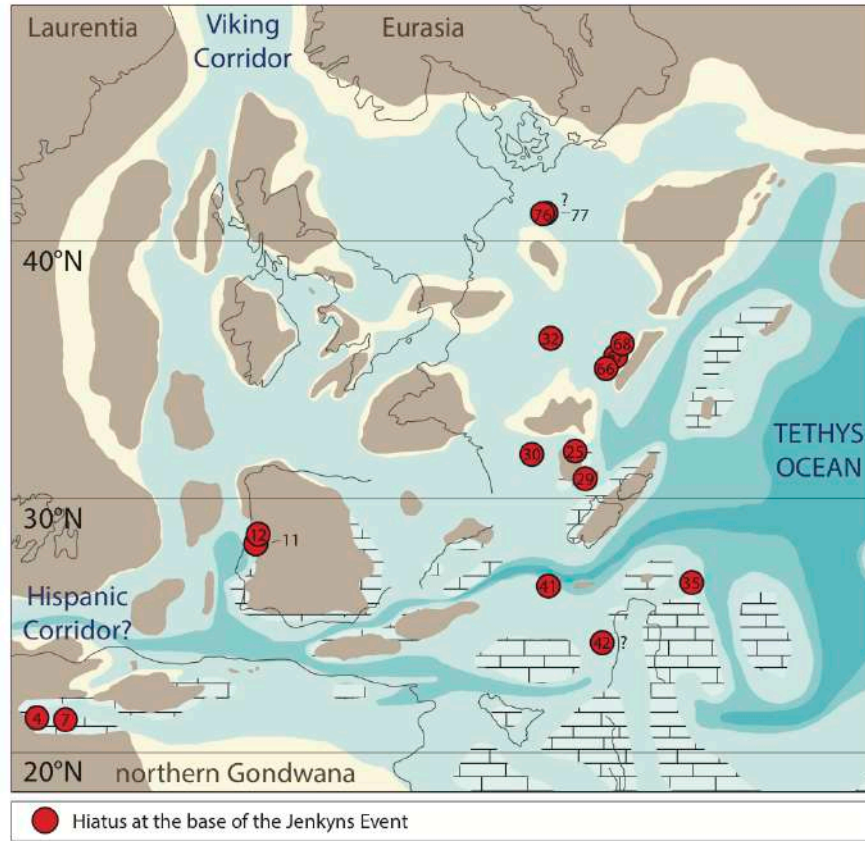
Figure 2

1	Raknet El Kahla	31	Sancerre-Couy Core	61	Réka Valley
2	Mellala	32	Andra HTM 102	62	Úrkút
3	Châabet El Attaris	33	Bascharage	63	Tölgyhát
4	Dades Valley	34	FR-210-078 Core	64	Dobravitsa-1
5	Boumardoul n'Imazighn composite section	35	Monte Mangart	65	Creux de l'Ours
6	Amellago	36	Dogna Core	66	Gipf
7	Foum Tillicht	37	Madonna della Corona	67	Riniken
8	Talghemt	38	Colma di Malcesine	68	Rietheim
9	Peniche	39	Sega d'Ala	69	Scheibelberg
10	Figueira da Foz	40	Sogno Core	70	Bächental
11	Fonte Coberta/Rabaçal	41	Gajum Core	71	Sachrang
12	Ribeiro (Coimbra composite section)	42	Valdorbia	72	Dotternhausen
13	Porto de Mós	43	Pozzale	73	Denkingen borehole (BEB 1012)
14	María Pares	44	Somma	74	Dormettingen
15	La Cerradura	45	Mercato San Severino	75	Aubach
16	Fuente Vidriera	46	Monte Sorigenza	76	Core L1
17	Arroyo Mingarrón	47	Kovk	77	Schandelah Core
18	West Rodiles-Santa Mera	48	Gornje Jelenje	78	Rijswijk-1 Core
19	La Almunia-Ricla	49	Velebit-A	79	F11-01
20	Castrovido	50	Velebit-B	80	L05-04
21	Sierra Palomera (Rambia del Salto)	51	Petousi	81	Bornholm
22	Barranco de la Cañada	52	Toka	82	Mochras Farm Borehole
23	Castillo de Pedroso	53	Kastelli	83	Winterborne (Kingstone)
24	Es Cosconar (section 4)	54	Zázrivá	84	Seavington St Michael
25	Roqueredonde	55	Skladana Skala	85	Yorkshire composite section*
26	Beaujolais (Lafarge quarry)	56	Mechowo IG 1	86	Holwell Quarries
27	Truc de Balduc	57	Gorzów Wielkopolski IG 1	87	Raasay
28	Fontaneilles	58	Suliszowice 38 BN	88	Well 18/25-1
29	Cuers	59	Brody-Lubienia BL 1	89	Gulfaks 34/10-35 Core
30	Penne Château-Granier section	60	Parkosovice		

1946 *Hawskers Bottoms, Port Mulgrave, Runswick Bay, Kettleless, Saltwick Bay, Old Nabb, Ravenscar, Staithes, Robin Hood's Bay, Blea Wyke (Peak), Saltwick Nab, Dove's Nest borehole, Whitby

1947 **Table 1**

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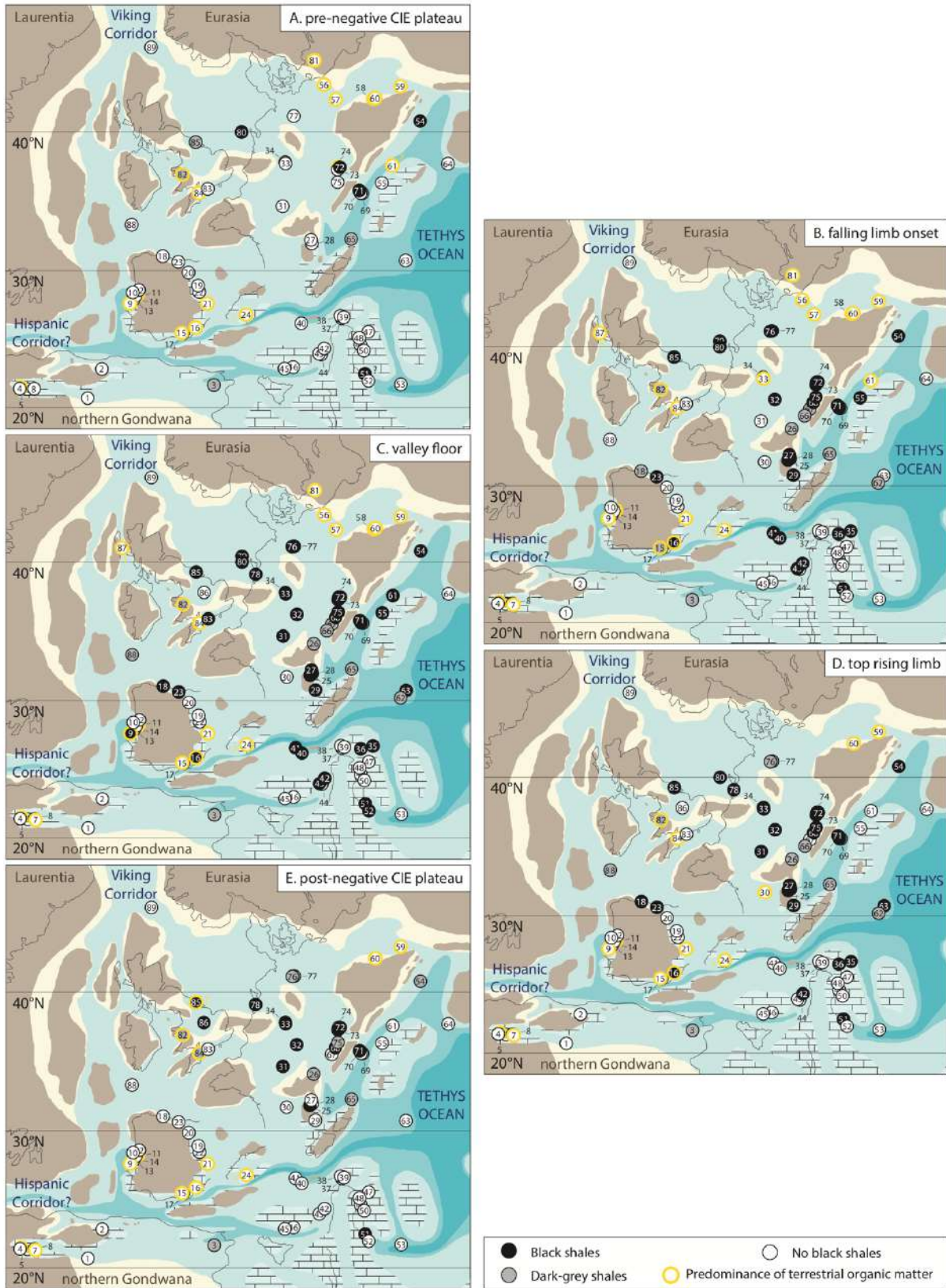


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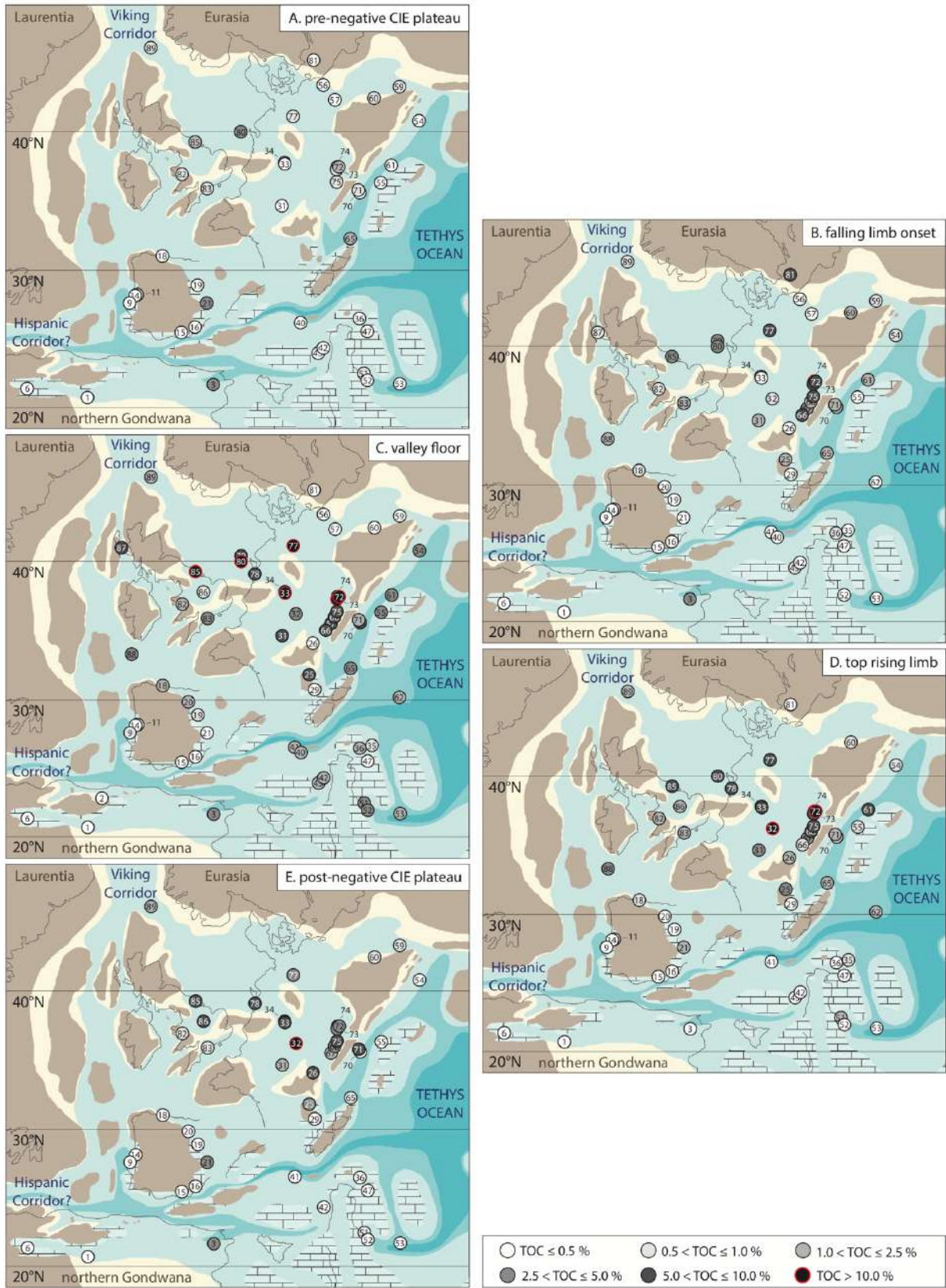
Figure 3



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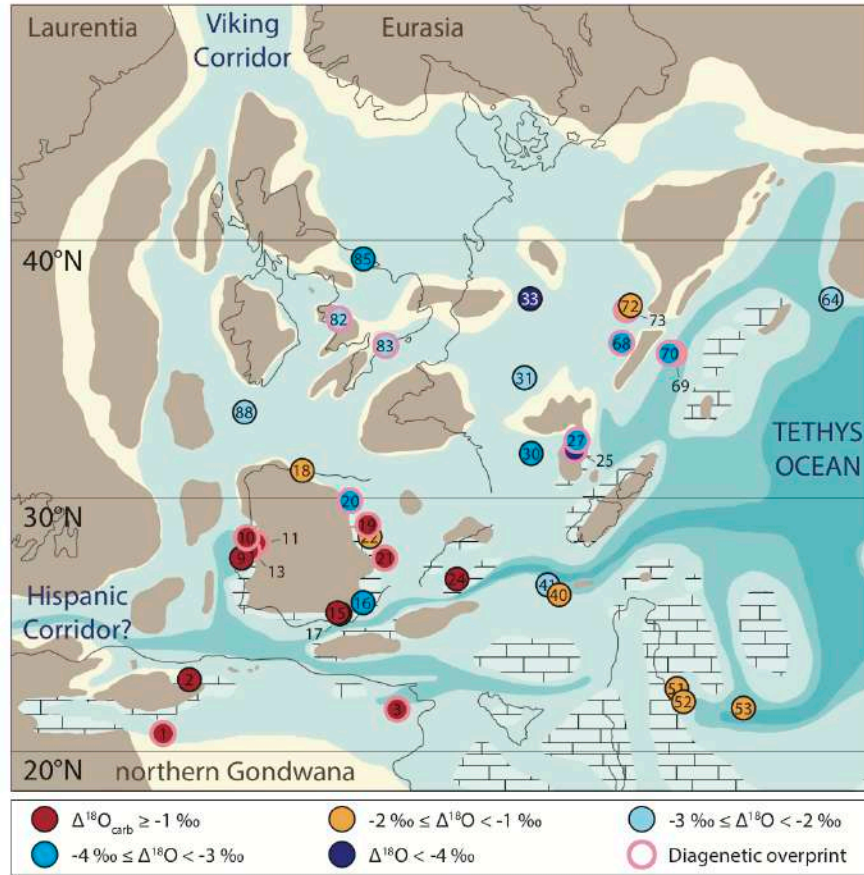
Figure 4



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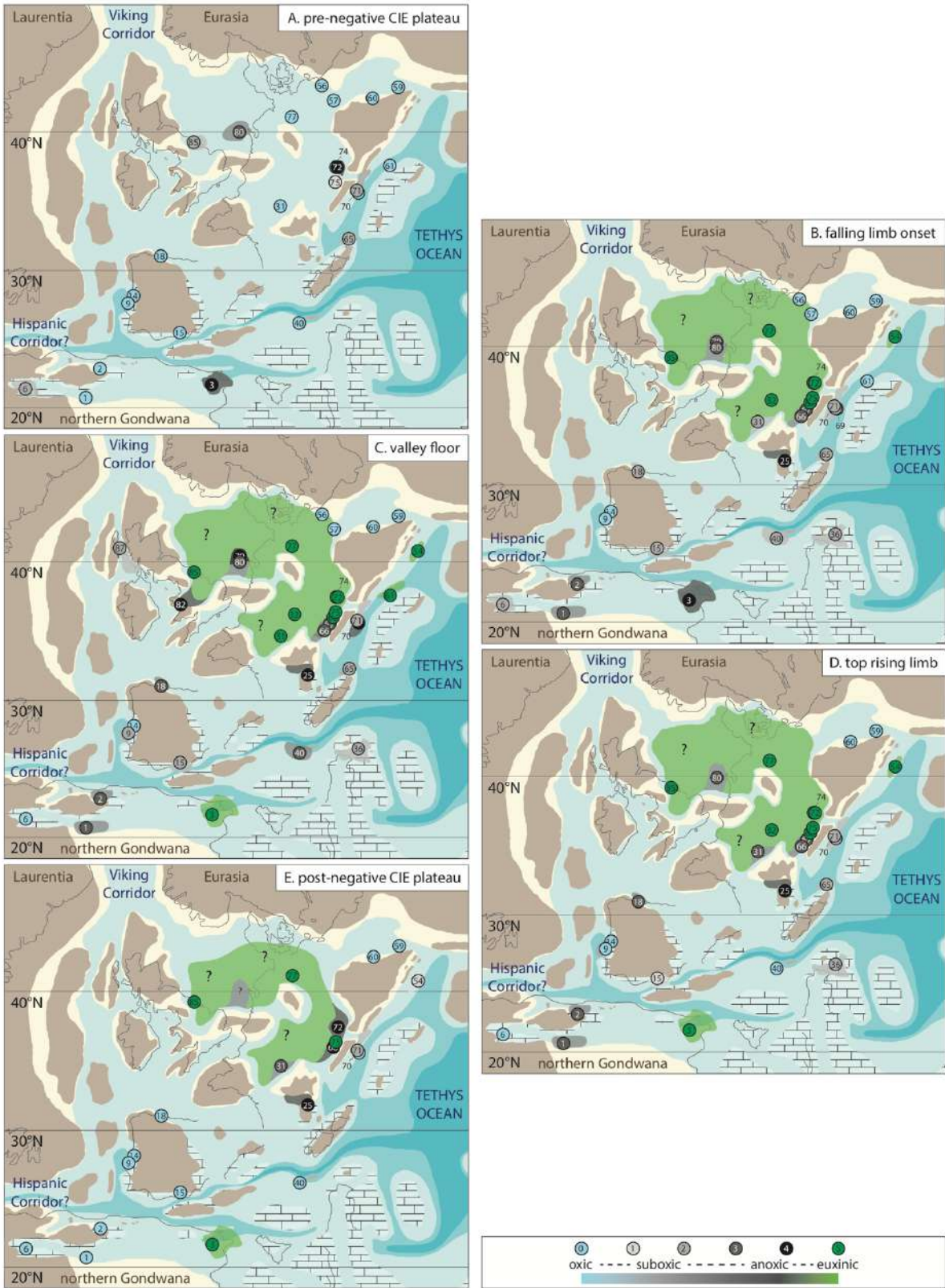
Figure 5



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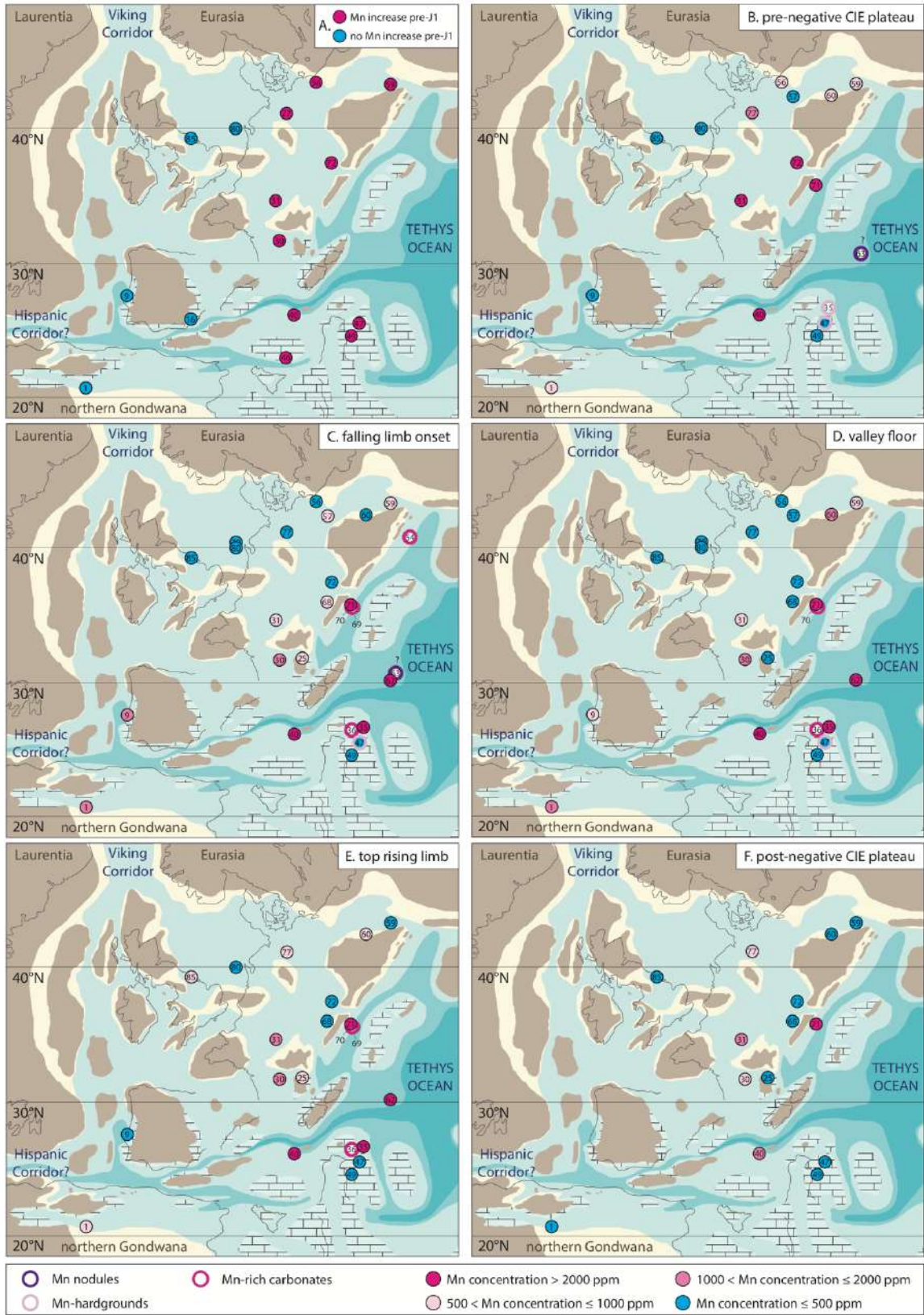
Figure 6



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Figure 7

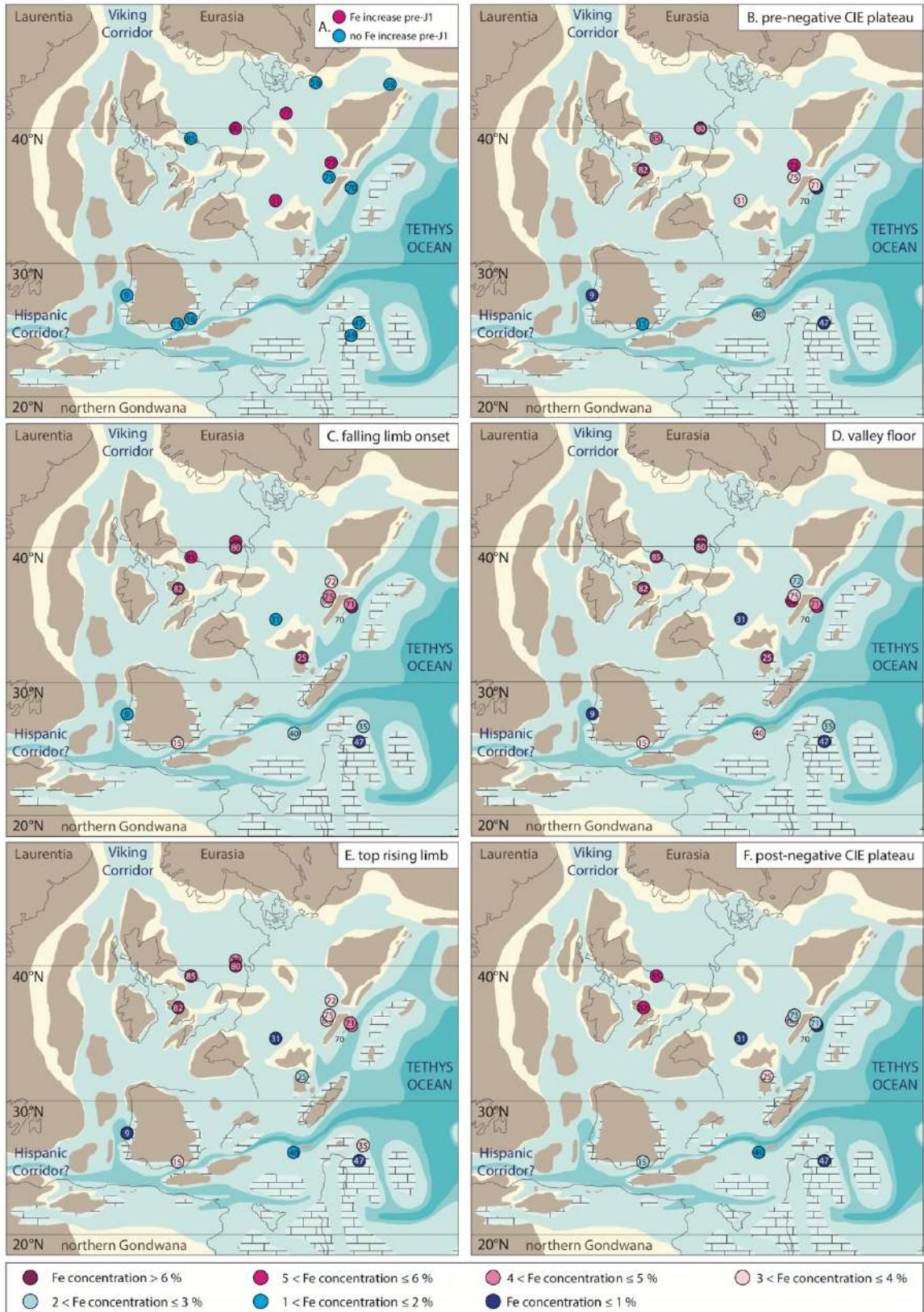


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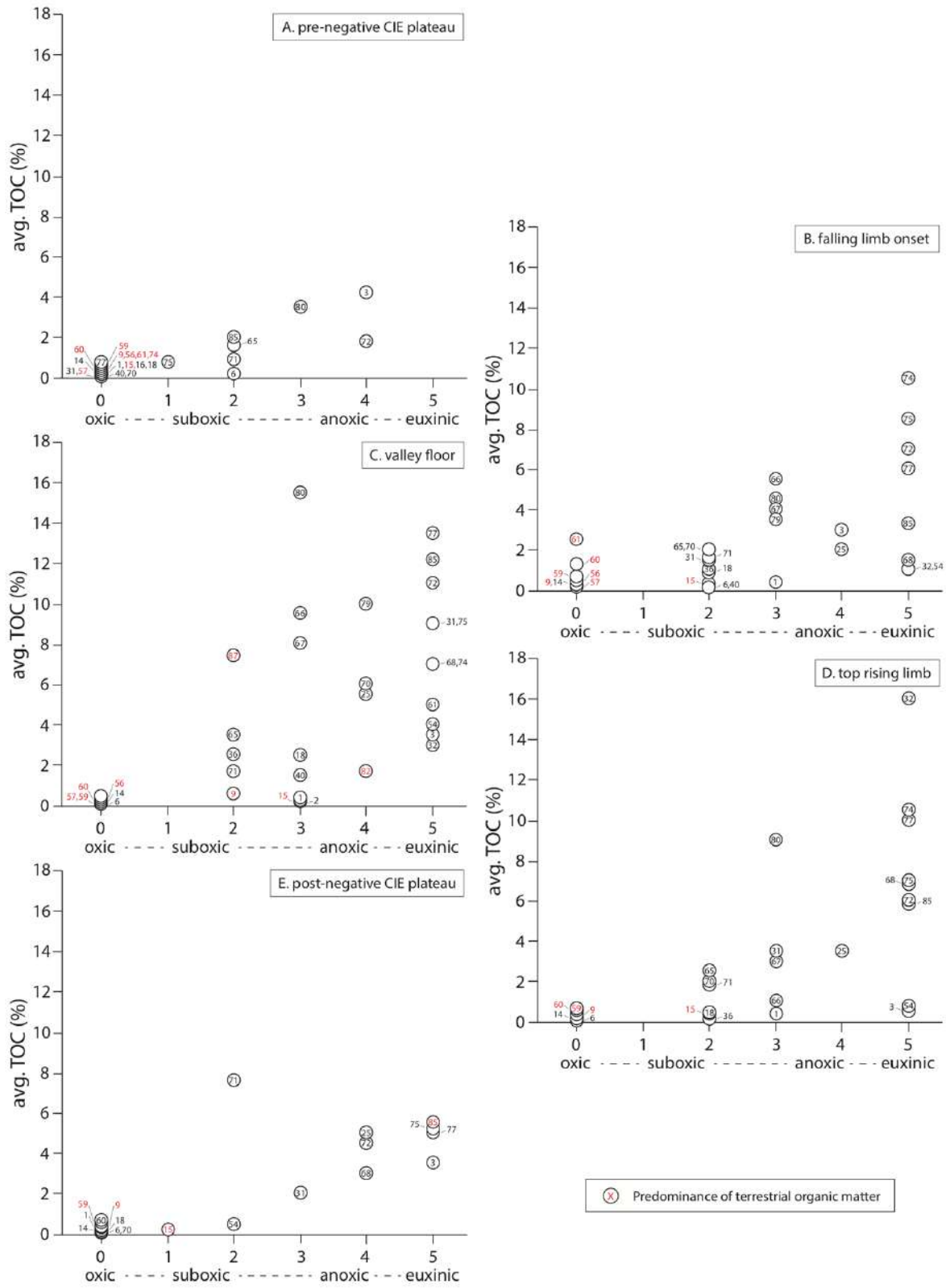
Figure 8

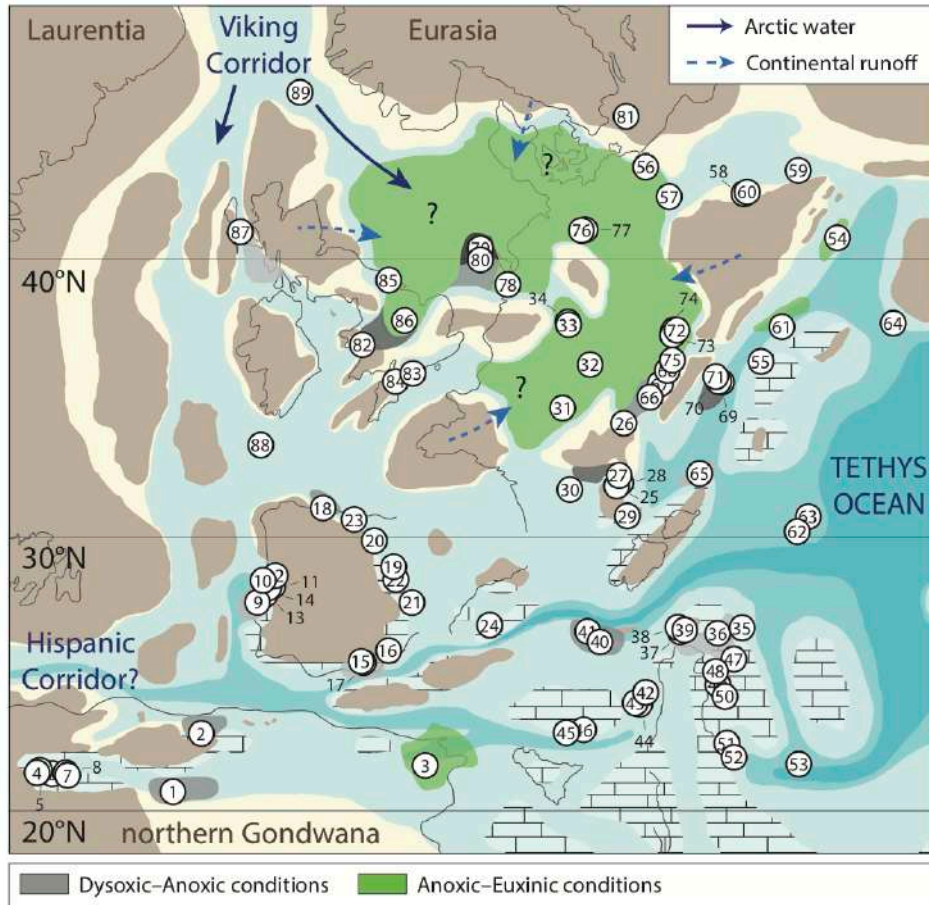


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Figure 9

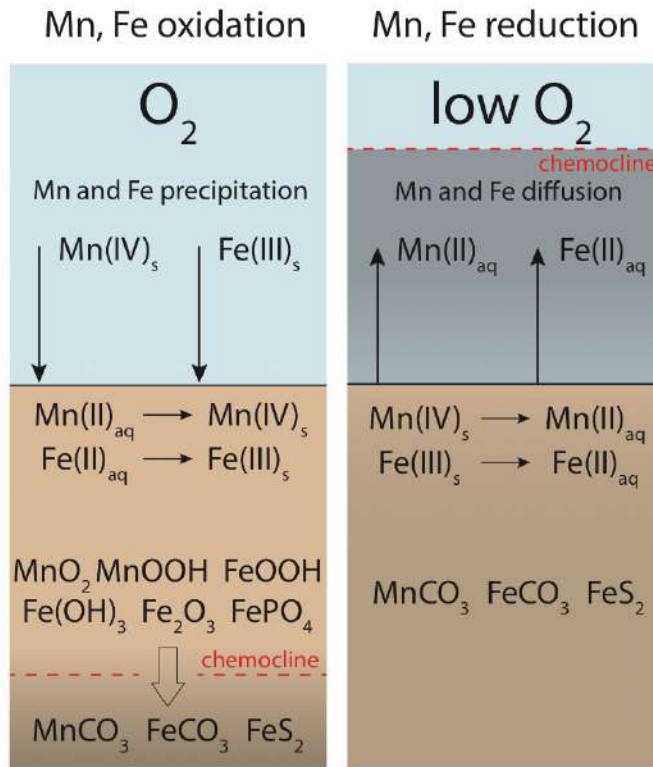




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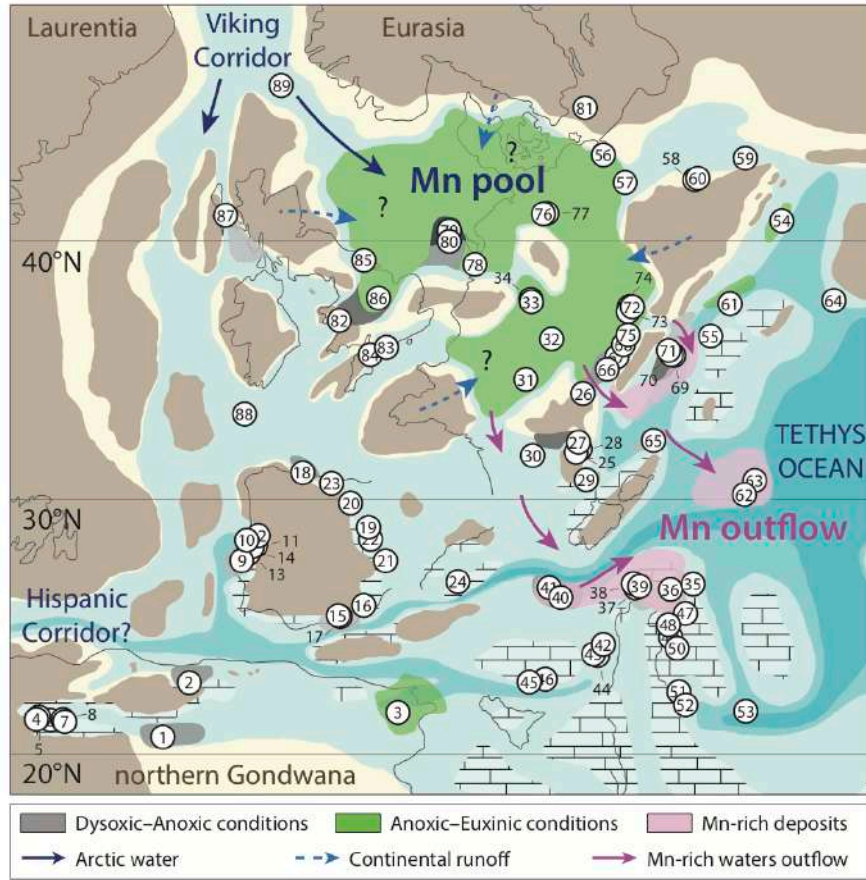
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Figure 11



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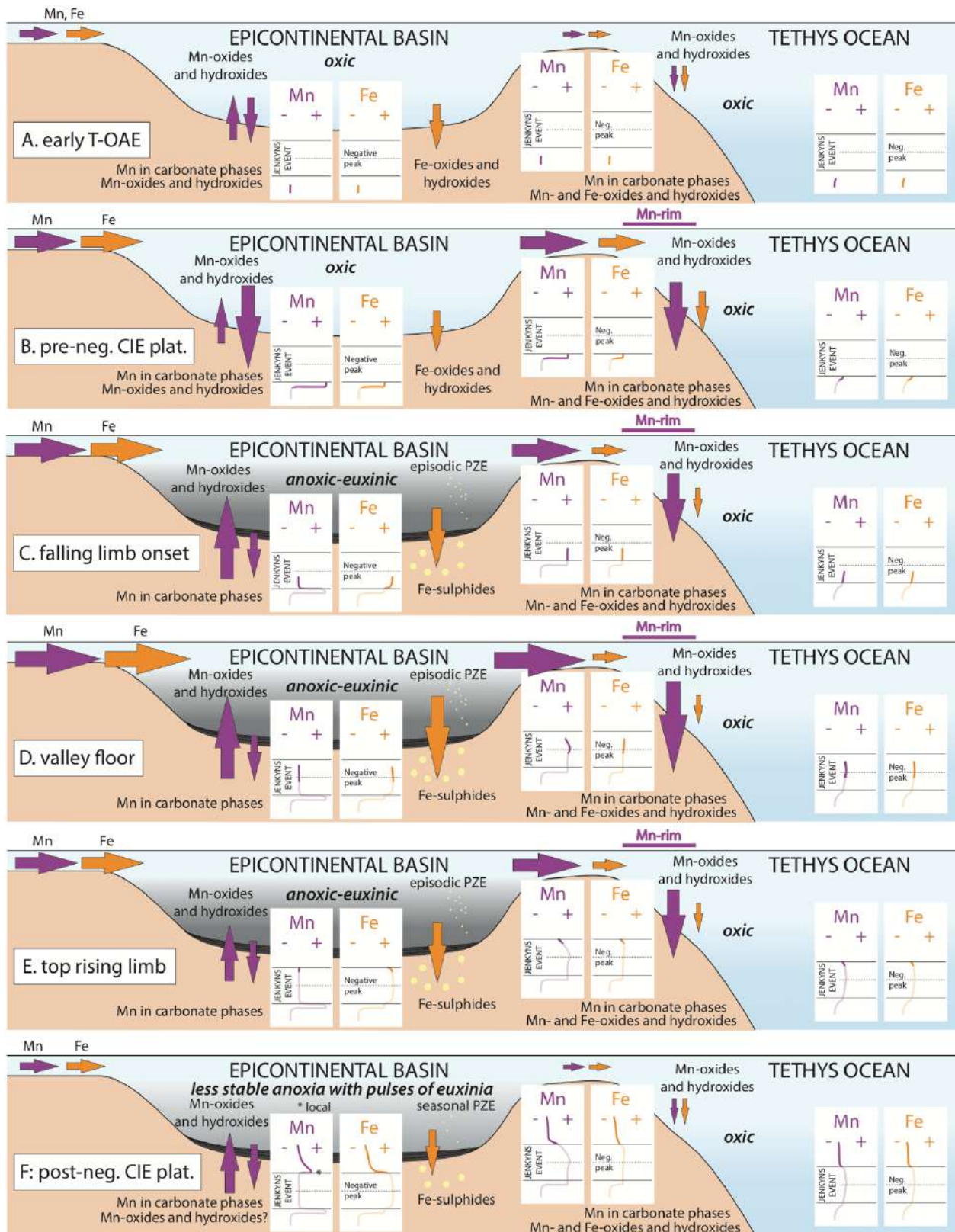
Figure 12



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Figure 13



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Figure 14