

1 Global warming leads to Early Triassic nutrient stress across
2 northern Pangea

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25 **ABSTRACT**

26 The largest extinction in Earth history, in the latest Permian, was followed throughout most of
27 the Early Triassic by a prolonged period of ecologic recovery. What factors delayed biotic
28 recovery are still under debate and partly revolve around impacts of global warming on primary
29 marine productivity. We examined N isotope records from the Festningen section on Spitsbergen
30 to examine changes in nutrient availability through the Early to Middle Triassic along the
31 northern margin of Pangea. Our results show progressive decline in N availability throughout the
32 Griesbachian, leading to severe nutrient limitations through the remainder of the Early Triassic,
33 until returning to a highly productive continental margin in Middle Triassic time. These results
34 are consistent with other studies from northern and western Pangea and thus show regional
35 nutrient limitations occurred in what should have been the main zone of marine primary
36 productivity. Such nutrient limitation likely stressed primary production and consequently
37 contributed to prolonged marine recovery. We suggest this was driven by high ocean
38 temperatures depressing the marine nutricline.

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40

41 **INTRODUCTION**

42 The Early Triassic represents a period of extreme global warming and severely stressed
43 environments (Tribovillard et al., 2006; Chen and Benton, 2012; Joachimski et al., 2012; Sun et
44 al., 2012; Grasby et al., 2013), that followed the Latest Permian Extinction (LPE), the most
45 severe in Earth history (Erwin et al., 2002; Chen and Benton, 2012; Bond and Grasby, 2017).
46 Marine environments were affected by the globally disrupted carbon cycle (Payne et al., 2004;
47 Galfetti et al., 2007; Grasby et al., 2013), and recurrent anoxia (Grasby et al., 2013; Wignall et
48 al., 2016), associated with numerous gaps in the sedimentary record including: absence of
49 biogenic chert and metazoan reefs (Chen and Benton, 2012 and references therein), as well as
50 nitrogen deficiency and absence of phosphorite deposition (Trappe, 1994; Kidder and Worsley,
51 2004; Grasby et al., 2016b). Terrestrial systems were also highly stressed, with a noted gap in
52 coal deposition (Retallack et al., 1996), and evidence of enhanced continental denudation
53 (Sephton et al., 2005; Algeo and Twitchett, 2010; Midwinter et al., 2017). These severe Early
54 Triassic conditions extended for 5–9 Ma after the LPE, until final return in the Middle Triassic to
55 normal marine conditions (Bottjer et al., 2008; Chen and Benton, 2012; Grasby et al., 2013) and
56 sea water temperature (Sun et al., 2012).

57 What prolonged the Early Triassic recovery has been under active debate, one that partly
58 revolves around primary productivity of global oceans at that time. Some researchers have
59 suggested high post LPE bioproductivity (e.g. Meyer et al., 2011; Schobben et al., 2015; Shen et
60 al., 2015) while others have argued for oceans with low primary productivity (e.g. Schoepfer et
61 al., 2013; Song et al., 2013; Winguth et al., 2015; Grasby et al., 2016b). Examination of nitrogen
62 isotope records can help elucidate nutrient availability, and thus constraints on primary
63 productivity in Early Triassic oceans. A detailed nitrogen isotope profile through the Smithian
64 stratotype, in the Sverdrup Basin, Canadian High Arctic, showed evidence for progressively

65 increased nutrient limitation following the LPE, leading to an Early Triassic ‘nutrient gap’ and
66 associated bioproductivity crisis (Grasby et al., 2016b). Whether or not this was local nutrient
67 limitation in the Sverdrup Basin, or a more widespread event, remains uncertain. Here we test the
68 regional extent of N-limited oceans by examining the broader northern Pangea margin. We
69 analysed N isotope data, along with key nutrients (P, N) and bioproductivity proxies (Ba, Ni, and
70 Cu) (Dymond et al., 1992; Steiner et al., 2017) from the Festningen section, Spitsbergen (Fig. 1),
71 that was deposited in an open marine environment (Fig. 2). We show that trends in stable
72 nitrogen isotope values are similar to those observed in the Sverdrup Basin, confirming
73 widespread N-limited conditions across northern Pangea throughout the Early Triassic.

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75 **UPWELLING AND N LIMITATION IN THE PANTHALASSA OCEAN**

76 Upwelling zones of the world oceans, that transport nutrients to the photic zone, form major
77 regions of bioproductivity, mainly along the western margins of continents (western coasts of
78 modern North America, South America, Africa, and Australia) (Capone and Hutchins, 2013).
79 While these upwelling zones represent a minor percentage of ocean area, they are responsible for
80 a major portion of marine primary productivity. Estimates suggest that half the ocean
81 biogeochemical flux of N is derived from continental margins occupying only 20% of world
82 ocean (Walsh, 1991). Nutrient upwelling along these margins is driven by Ekman Transport,
83 which is limited to the upper few hundred meters of the ocean surface (e.g. Huyer, 1983; Currie,
84 1992; Smith, 1995). Typically the ocean’s nutricline is coincident with the thermocline, both of
85 which lie above the base of upwelling, such that upwelling transports these nutrient-rich waters
86 to the photic zone, driving primary productivity. Models and data related to modern climate
87 warming suggest that increasing ocean temperature is depressing the thermocline/nutricline, and

88 there are concerns that if they are depressed below the base of upwelling that this may greatly
89 reduce nutrient transport to the photic zone, and consequently primary productivity
90 (Kamykowski and Zentara, 1986; Behrenfeld et al., 2006; Doney et al., 2012; Moore et al.,
91 2018). In contrast, some studies suggest global warming would intensify upwelling through
92 increased wind shear (Bakun, 1990). While these models suggest a direct link between ocean
93 temperatures and nutrient delivery to the photic zone, the net impact on primary productivity
94 under hothouse Earth conditions remains uncertain.

95 Examination of the nitrogen isotope records provides a means to assess the degree of
96 nutrient stress in ancient marine environments. Heterotrophic denitrification and/or anaerobic
97 ammonia oxidation are the principal mechanism for loss of ocean nitrogen, returning to the
98 atmosphere as N_2 (Ward et al., 2009). Regions of significant denitrification (i.e., the reduction of
99 NO_3/NO_2 to N_2) leave subsurface waters highly enriched in ^{15}N . In contrast, atmospheric
100 nitrogen fixation produces organic matter relatively depleted in ^{15}N , with $\delta^{15}N$ values close to
101 that of the atmosphere (0‰). As such, the $\delta^{15}N$ of organic matter (OM) in modern oceans is
102 highly variable, but relatable to biogeochemical processes of different marine environments
103 (Somes et al., 2010). The rates of N_2 fixation are closely associated with, both geographically
104 and temporally, marine nitrogen removal, implying a close coupling of nitrogen fixation to
105 nitrogen-deficient water in denitrification zones (Deutsch et al., 2007). Thus, in any single
106 location, when N supply from anoxic deep waters to the photic zone is reduced, diazotrophs
107 increase N fixation (Carpenter et al., 1997), producing biomass with significantly reduced values
108 of $\delta^{15}N$ in the particulate OM sediment record. With this, changes in the balance between
109 denitrification and nitrogen fixation through time can be tracked.

110 The geological record of N isotopes can provide key insight into changes in the paleo-
111 marine N cycle. Modern settings demonstrate that there is little diagenetic alternation of $\delta^{15}\text{N}$ in
112 sedimentary OM (Altabet et al., 1999a; Altabet et al., 1999b), and post depositional temperature
113 changes also have limited effect on sedimentary $\delta^{15}\text{N}$ (Ader et al., 1998). Thus, stable isotopes of
114 nitrogen provide an effective tracer of nutrient stress, as the balance between denitrification and
115 nitrogen fixation controls the abundance and isotopic composition of nitrate in local marine
116 environments (Ganeshram et al., 2000).

117 The closure of the Uralian ocean, during final consolidation of Pangea, was associated with
118 development of a nutrient-rich upwelling margin along northern Pangea since at least the
119 Sakmarian (Stemmerik and Worsley, 1995; Beauchamp and Baud, 2002; Reid et al., 2007;
120 Beauchamp and Grasby, 2012; Blomeier et al., 2013). By the Late Permian, coastal upwelling
121 zones along the northern margin of Pangea were major regions of marine primary productivity
122 (Beauchamp and Baud, 2002). However, several studies have shown that immediately following
123 the LPE, upwelling of nutrient-rich waters was greatly reduced across northern and western
124 Pangea, leading to nutrient-limited conditions, enhanced nitrogen fixation, and reduced
125 bioproductivity (Beauchamp and Baud, 2002; Knies et al., 2013; Schoepfer et al., 2013; Grasby
126 et al., 2015). Climate models also suggest reduced mid-latitude upwelling of nutrient-rich waters
127 under Early Triassic hothouse conditions occurred due to decreased wind-stress-driven Ekman
128 transport (Kidder and Worsley, 2004; Winguth et al., 2015). Nutrient trapping due to deepened
129 thermocline/nutricline during Early Triassic hothouse conditions has also been suggested
130 (Grasby et al., 2016b). It is important to recognize that available observations cannot discern
131 between reduced physical upwelling of water, as compared to reduced nutrient transport, in a
132 system where upwelling is maintained or even enhanced. For example, a depressed

133 thermocline/nutricline would limit nutrient transport to the photic zone even if upwelling was
134 intensified, effectively trapping nutrients below the base of upwelling (Grasby et al., 2016b).

135 Along with decreased nutrient delivery to the photic zone, Early Triassic
136 greenhouse/hothouse conditions are thought to increase denitrification related to increased
137 anoxia, that also drives increased N fixation to maintain a Redfield balance (N/P = 14) with
138 phosphorus over longer timescales (Tyrrell, 1999; Kidder and Worsley, 2010), even though N
139 fixation is an overall more energetically costly process in comparison to assimilation of pre-
140 existing fixed N (Brandes et al., 2007). A rise in sea surface temperature would also intensify
141 thermal stratification of the ocean, expand N-limited subtropical gyres (Sarmiento et al., 1998),
142 as well as deepen the thermocline (and nutricline), further limiting nutrient resupply to the photic
143 zone (Kamykowski and Zentara, 1986; Behrenfeld et al., 2006). How this affects marine
144 productivity depends partly on the degree to which diazotrophs offset net N loss (Saltzman,
145 2005). Data from the Sverdrup Basin suggests that initial nutrient limitation at the LPE (Knies et
146 al., 2013) was the start of an even larger trend towards more nutrient restricted conditions
147 throughout the entire Early Triassic that stressed primary production (Grasby et al., 2016b), as
148 marked by a progressive decrease of $\delta^{15}\text{N}$ values through the Griesbachian/Dienerian towards
149 values of atmospheric nitrogen ($\delta^{15}\text{N} \approx 0\%$) and concurrent drops in TOC and bioproductivity
150 proxies.

151 **STUDY AREA AND PALEO-ENVIRONMENTAL SETTING**

152 The Festningen section, located at Kapp Starostin, Spitsbergen (Fig. 1), provides a near
153 continuous exposure of Carboniferous to Cenozoic strata along a low sea-cliff from Kapp
154 Starostin to Festningsdodden, including the thickest known development of Lower Triassic
155 marine stratigraphy on the island. Located in the eastern part of the West Spitsbergen Fold and

156 Thrust Belt, the Festningen section is part of the steeply inclined short-limb of a kilometer-scale
157 east-vergent fold structure (Maher and Craddock, 1988; Dallmann et al., 1993; CASE-Team,
158 2001). An igneous sill (dating from the Cretaceous 124.7 Ma) (Corfu et al., 2013) intrudes the
159 sedimentary section near the basal Triassic contact but has limited and local thermal effect
160 (Grasby et al., 2015).

161 The upper 40 m of the Kapp Starostin Formation is characterized by Lopingian (Upper
162 Permian) spiculitic chert (Blomeier et al., 2013), considered equivalent to the Black Stripe and
163 Lindström formations of the Sverdrup Basin (Beauchamp et al., 2009; Bond et al., 2015). These
164 cherts are in sharp contact with overlying Lower to Middle Triassic shale, siltstone and minor
165 sandstone (Mørk et al., 1982) equivalent to the Blind Fiord Formation of the Sverdrup Basin
166 (Embry, 1989). The contact between the Kapp Starostin and Vardebukta formations coincides
167 with the Latest Permian Extinction, and the basal ~6-7 m of the Vardebukta Formation is latest
168 Permian in age (Wignall et al., 1998; Grasby et al., 2015).

169 The Lower Triassic succession at Festningen is divided into the shale and siltstone
170 dominated Vardebukta (uppermost Changhsingian, Griesbachian and Dienerian) and
171 Tvillingdodden (Smithian/Spathian) formations (Embry, 1989; Wignall et al., 1998; Grasby et
172 al., 2016a; Wignall et al., 2016). The sediments were deposited in a distal shelf setting (Wignall
173 et al., 1998; Stemmerik and Worsley, 2005; Blomeier et al., 2013) at estimated paleolatitudes of
174 ~40 to 45° N (Golonka and Ford, 2000; Scotese, 2004; Hounslow et al., 2007). Early work on the
175 Triassic sequence at Festningen showed onset of ocean anoxia following the LPE (Wignall et al.,
176 1998). Further work demonstrated that this anoxia was the culmination of progressive
177 environmental deterioration, initiated with ocean acidification, loading of toxic metals, and final
178 onset of anoxia (Grasby et al., 2015). Mercury records also show anomalies at the LPE as well as

179 later Smithian extinction events, likely associated with Siberian Trap eruptions (Grasby et al.,
180 2016a). In addition, the Lower Triassic record at Festningen shows changes in bioturbation,
181 paleoecology, pyrite framboid content and trace metal concentrations that demonstrate anoxic
182 phases alternated with intervals of better ventilation (Wignall et al., 2016). Through Early
183 Triassic time, only the Dienerian and early Smithian had oxygenation sufficient for supporting a
184 diverse benthic community. Anisian strata (Middle Triassic) are represented by phosphatic
185 organic-rich shales of the Botneheia Formation (Krajewski, 2008).

186 The Paleozoic–Mesozoic transition along northwest Pangea was marked by a change from
187 very low rates of biosiliceous sediment accumulation during the Late Permian to rapid rates of
188 clastic sedimentation throughout the Early and Middle Triassic (Embry and Beauchamp, 2008;
189 Beauchamp and Grasby, 2012). Deteriorating environmental conditions during the Late Permian,
190 interpreted as resulting from the shoaling of the calcite lysocline and establishment of
191 increasingly acidic conditions in response to global warming, inhibited the production and
192 preservation of biogenic benthic carbonates in a vast area from Western Canada (Schoepfer et
193 al., 2013) to the Canadian (Beauchamp et al., 2009) and Norwegian (Bond et al., 2015) Arctic.
194 Slowly-producing siliceous sponges were the only organisms that could thrive in such a hostile
195 environment, resulting in significant condensation and preservation of their spiculitic remains as
196 biogenic chert (Beauchamp and Grasby, 2012). The subsequent encroachment of anoxic waters
197 onto the shallow shelves further contributed to the near-complete eradication of carbonate-
198 secreting benthic organisms (Grasby and Beauchamp, 2009). However, organic productivity in
199 the upper part of the water column remained high as shown by increasing TOC up to the LPE
200 horizon, which reflects enhanced preservation of organic matter in uppermost Permian sediments
201 due to the establishment of anoxic conditions on the sea floor, not an increase in organic

202 productivity near the sea surface (Grasby and Beauchamp, 2009). A major shift in sedimentary
203 regime occurred across the LPE as large volumes of terrigenous clastic material were shed onto
204 northwestern Pangea shelves and basins throughout the Early and Middle Triassic. While the
205 change in sedimentation rates and style across the LPE might have affected the absolute
206 concentration of certain elements, elemental and isotopic ratios should have remained unaffected
207 by these changes.

208 **METHODS**

209 Samples were collected at the Festningen Section, located at N78°5.72'; E13°49.424' (relative to
210 NAD83). Field sampling was conducted relative to the Kapp Starostin/Vardebukta formation
211 contact that represents the LPE horizon, whereby samples are recorded in metres above
212 (positive) and below (negative) the top of the last chert bed that defines the top of the Kapp
213 Starostin Formation. Sample spacing varied, from 20 cm within 1 to 2 m of the formation contact
214 to higher spacing of 50 cm for the rest of the section sampled. Weathered surfaces were removed
215 and then samples were collected from an isolated layer no greater than 2 cm thick. These same
216 samples have been subject of previous studies (Bond et al., 2015; Grasby et al., 2015; Grasby et
217 al., 2016a; Wignall et al., 2016), and previously published data are indicated where relevant. We
218 used the previously published organic carbon isotope record to sub-sample for N isotope
219 analyses, such that the selected sample subset would accurately reproduce the full carbon isotope
220 trend through time. From this, 98 samples were selected for N isotope analyses (approximately
221 every 3rd sample). Analytical results are provided in **Table 1**.

222 In the laboratory, any remaining weathered surfaces were removed and fresh samples were
223 powdered by agate mortar and pestle. Total N and $\delta^{15}\text{N}$ were analysed by using an elemental
224 analyser connected to an isotope ratio mass spectrometer (EA-IRMS). Approximately 20% of the

225 samples were analyzed in duplicate with a mean standard deviation of 0.16‰. Results are
226 reported as standard δ -values in per mil (‰ vs. air). The reference material used during analysis
227 of the samples (IA-R001 wheat flour, Iso-Analytical Ltd.) had a $\delta^{15}\text{N}$ value of 2.55‰ versus air
228 and contained 1.88% N (w/w). Control samples were analysed to check the accuracy of the
229 measurements, with precision better than $\pm 0.2\text{‰}$.

230 **RESULTS**

231 The Festningen section records significant changes in global biogeochemical cycles that
232 occurred through Late Permian to Middle Triassic time as expressed by variation in $\delta^{13}\text{C}_{\text{org}}$. We
233 use this record, to place into context new data presented here; including nutrient (N and P), stable
234 isotope ($\delta^{15}\text{N}$), and element enrichment factors for bioproductivity proxies for (Ba, Ni, Cu).
235 Previous work has shown that the organic carbon isotope record at Festningen can be readily
236 correlated with that of the Smithian stratotype (Fig. 3) and other sections in the Sverdrup Basin,
237 as well as inorganic carbon isotope records from the Tethys Ocean (Grasby et al., 2015; Grasby
238 et al., 2016a; Wignall et al., 2016). The Festningen $\delta^{13}\text{C}_{\text{org}}$ record (Fig. 4a) shows a significant
239 10‰ negative shift at the LPE boundary, followed by progressive recovery through
240 Griesbachian/Dienerian time, returning to close to pre-extinction values in the lower Smithian.
241 There is then a progressive drop through the Smithian to lows of $\sim -33\text{‰}$, below the
242 Smithian/Spathian boundary, coincident with a significant late Smithian anoxic event (Grasby et
243 al., 2013; Wignall et al., 2015). Above this $\delta^{13}\text{C}_{\text{org}}$ values fluctuate through the Spathian and
244 finally stabilize in the Anisian.

245 The total organic carbon (TOC) values prior to the Capitanian Crises are generally low ($<$
246 $\sim 0.5\%$), then increase in the uppermost Permian (Fig. 4b). TOC values drop at the LPE and then
247 briefly increase in the basal Griesbachian up to values of 1.0%, before dropping to values $< 0.1\%$

248 through to the end of the Smithian. The TOC values then increase up to 1.0% in the Spathian
249 before declining again. At the onset of the Anisian the TOC values show a significant
250 progressive increase to values up to 3% in the Middle Triassic Botneheia Formation.

251 $\delta^{15}\text{N}$ data

252
253 The nitrogen isotope record at Festningen shows initial high $\delta^{15}\text{N}$ values ($> 8\text{‰}$) prior to the
254 Capitanian Crises of Bond et al. (2015), and then a drop to values of $\sim 6\text{‰}$ through the latest
255 Permian (Fig. 4c). At the LPE event $\delta^{15}\text{N}$ values fall again to $\sim 3\text{‰}$ in the lower Griesbachian.
256 Through the remainder of the Lower Triassic $\delta^{15}\text{N}$ values continued to decline down to values of
257 $\sim 1\text{‰}$ near the top of the Spathian. The $\delta^{15}\text{N}$ values then showed a positive shift, increasing to
258 $\sim 2\text{‰}$ in the Anisian.

259 N and P data

260 The concentrations of N and P in sediments vary from 0.015 to 0.18% and 0.001 to 0.53%
261 respectively. For N, concentrations are low at the base of the section and show an increasing
262 trend about the level corresponding with the Capitanian Crises through to the lower Griesbachian
263 to $\sim 0.12\%$ (Fig. 4d). Values are then low ($< 0.50\%$) throughout the remainder of the Lower
264 Triassic but increase in the Middle Triassic to the highest values observed in the section (0.18%).
265 Through this time period P shows different trends, remaining low through the Permian (0.03%)
266 and then showing a general increasing trend through the Griesbachian/Dienerian. The P values
267 then remained relatively stable at $\sim 0.1\%$ through the rest of the Lower Triassic with the
268 exception of a brief increase to $> 0.2\%$ in the Smithian. P values are highest in the Anisian
269 Botneheia Formation ($> 0.25\%$) which is characterized by abundance of phosphate nodules.
270 While our data in Figure 4e shows relatively high P in the Middle Triassic, this represents only
271 values measured in shales and does not account for the presence of these P nodules. Estimates

272 from equivalent units indicate that bulk rock P content (P in shales and nodules) can be up to 10
273 x these concentrations (Grasby et al., 2016b). The N/P ratio is highest in the Upper Permian (2 to
274 5) and then drops significantly through the Griesbachian to values < 1 (Fig. 4f). The N/P ratio
275 stays low throughout the Lower Triassic and increases to ~ 1 in the Middle Triassic, but could be
276 up to an order of magnitude lower when accounting for P in phosphate nodules. In general,
277 trends in the N/P closely corresponds to changes in TOC through the Lower Triassic but this
278 relationship breaks down in the Middle Triassic.

279 **Paleoproductivity proxies**

280
281 Trace metals that act as micronutrients can be used as proxies for paleoproductivity (Tribovillard
282 et al., 2006). To counter potential dilution affects related to changes in sedimentation rates, these
283 are best examined as element enrichment factors (EF), defined as deviations of Al normalised
284 samples (to account for potential terrestrial input) from average shale values, taken here as Post
285 Archean Average Shale values (PAAS) (Wedepohl, 1995). Whereby $EF > 1$, calculated as $EF =$
286 $(X_{\text{sample}}/Al_{\text{sample}})/(X_{\text{PAAS}}/Al_{\text{PAAS}})$, are thought to indicate high primary productivity. As no single
287 productivity proxy is completely reliable, we calculated these for Ba, Ni and Cu as plotted in
288 Figure 4g. Enrichment factors for Ba and Ni are high (>1) through the Permian and then all show
289 a significant drop to values $\ll 1$ immediately following the LPE, with a brief recovery followed
290 by a shift to values near 1 through most of the Early Triassic. In contrast, Cu EFs remain low
291 though most of the section. There is a brief return to high EF in the Spathian consistent with
292 higher TOC values at that time. Following this EF return to values close to 1 and then shift to
293 higher values (>1) in the Anisian. These trends are also illustrated by comparing average EF
294 values for Ba and Ni during the Late Permian (1.17, 1.46 respectively), 50 m above the
295 extinction horizon (0.58, 0.99) and the remainder of the Early Triassic (0.88, 1.12).

296 **DISCUSSION**

297 **Early Triassic Nitrogen limitation**

298 The Festningen section records significant fluctuations in $\delta^{13}\text{C}_{\text{org}}$ that are consistent with
299 inorganic records from Tethyan sections as well as the $\delta^{13}\text{C}_{\text{org}}$ record from the Smithian
300 stratotype in the Sverdrup Basin (Fig. 3). This demonstrates that Festningen records open marine
301 conditions, with geochemical records that reflect global signals of perturbations to the carbon
302 pool. The $>8\text{‰}$ Late Permian $\delta^{15}\text{N}$ values at Festningen are consistent with significant water-
303 column denitrification occurring in continental margin upwelling zones that leaves subsurface
304 waters highly enriched in ^{15}N (Knies et al., 2013). These results are similar to those of other
305 studies showing that the broader margin of northern and western Pangea was a regional zone of
306 upwelling in Late Permian time (Beauchamp and Baud, 2002; Kidder and Worsley, 2004; Knies
307 et al., 2013; Schoepfer et al., 2013; Grasby et al., 2016b) (Fig. 2a). The $\delta^{15}\text{N}$ record of
308 Festningen then shows a $\sim 3\text{‰}$ decline in $\delta^{15}\text{N}$ values across the LPE, that we interpret to reflect
309 a decline of nutrient delivery to the photic zone, and as a consequence greater reliance on
310 atmospheric N fixation to support primary productivity. This is consistent with broader regional
311 drops in $\delta^{15}\text{N}$ values observed at the LPE across northern and western Pangea, including the
312 Sverdrup and Western Canada basins (Knies et al., 2013; Schoepfer et al., 2013; Grasby et al.,
313 2016b). In contrast to this regional trend, the East Greenland Fiskegrav section had lower overall
314 $\delta^{15}\text{N}$ values prior to the LPE, and does not show any significant change across the extinction
315 event (Mettam et al., 2017). This likely reflects the highly restricted nature of the East Greenland
316 Basin, which was separated from wider Panthalassa ocean circulation during Late Permian time
317 (Mettam et al., 2017; Roberts et al., 2018). East Greenland was thus not affected by regional
318 upwelling prior to the LPE, nor changes to N delivery after the extinction. Instead it more likely
319 reflects a depositional environment that was isolated from overall changes in the global ocean.

320 This is also seen as a distinctly different carbon isotope record in East Greenland (Sansón-
321 Barrera et al., 2015), as compared to both northern Pangea and other global records.

322 On a more global scale, a drop in $\delta^{15}\text{N}$ across the LPE occurs in most studied sections, and
323 can be even more significant elsewhere than northern Pangea (Fig. 5). Numerous Tethyan
324 sections from China show a significant drop in $\delta^{15}\text{N}$ values to zero or less right after the LPE
325 (Cao et al., 2009; Luo et al., 2011; Saitoh et al., 2014; Xiang et al., 2016). A lesser, but still
326 noticeable, $\delta^{15}\text{N}$ drop across the LPE is also recorded in the Peri-gondwanan sections at Guryul
327 (Algeo et al., 2007) and the western Tethyan Bulla section (Jia et al., 2012). It is instructive to
328 compare these sections, as NW Pangea shows the most enriched $\delta^{15}\text{N}$ values prior to the
329 extinction, consistent with previous research that indicated this region of the Panthalassa Ocean
330 was a dominant upwelling zone with strong nutrient recycling (Beauchamp and Baud, 2002;
331 Knies et al., 2013; Schoepfer et al., 2013; Grasby et al., 2015). To do this we used the two time
332 markers of the LPE and Permian/Triassic boundary to stretch different records in order to allow
333 comparison of N isotope trends across Pangea as seen in Figure 5. The very rapid and significant
334 negative shifts across the LPE in the Tethys, to $\delta^{15}\text{N}$ values consistent with a nutrient-limited
335 system supported by N_2 fixation, suggest more severe nutrient limitations in the Tethys than
336 northern Pangea. This can be explained by model results that show the Tethys Sea was strongly
337 stratified at the LPE due to limited connection with the larger Panthalassa Ocean (Kiehl and
338 Shields, 2005). In contrast, the only $\sim 3\%$ drop in $\delta^{15}\text{N}$ values across the LPE in sections from
339 northern Pangea suggests that while stressed, the region still maintained relatively high nutrient
340 levels in the immediate aftermath of the LPE.

341 The growing prevalence of atmospherically fixed nitrogen immediately following the LPE
342 is supported by the widespread evidence for cyanobacteria blooms, including diazotrophs, during

343 the Early Triassic, as documented by biomarker data in sections globally (Grice et al., 2005; Xie
344 et al., 2005; Hays et al., 2007; Xie et al., 2007; Hays et al., 2012). This is consistent with
345 eukaryotic algae being more vulnerable to extinction in a N-limited ocean than prokaryotic
346 photoautotrophs, whereby primary producers capable of nitrogen fixation would be favoured
347 (Anbar and Knoll, 2004; Knoll et al., 2007). Thermal stress of the Early Triassic ocean would
348 also decrease microbial diversity (Sharp et al., 2014) that might have further limited primary
349 productivity.

350 Most previous studies of the $\delta^{15}\text{N}$ record have just focused on the LPE itself, making it
351 difficult to examine the Early Triassic record on a global basis. However, for northern Pangea we
352 show that the Festningen $\delta^{15}\text{N}$ record is very similar to the Smith Creek record of the Sverdrup
353 Basin (Grasby et al., 2016b). Namely, the initial negative shift in $\delta^{15}\text{N}$ values at the LPE
354 boundary at Festningen was just the start of a longer term progressive decline. At Festningen The
355 negative $\delta^{15}\text{N}$ shift was characterised by a further $\sim 4\%$ decline in $\delta^{15}\text{N}$ through the
356 Griesbachian. The $\delta^{15}\text{N}$ values then remained low ($\sim 1\text{-}2\%$) through the Smithian and Spathian.
357 We interpret this to indicate ongoing development of nutrient stress with a N cycle dominated by
358 diazotrophs fixating atmospheric N_2 through the entire Early Triassic. In contrast to the Tethys
359 region, where nutrient limitation developed very rapidly, the disruption of the N cycle along
360 northern Pangea appears to be characterised by longer-term progressive development of nutrient
361 limitation through the Griesbachian/Dienerian.

362 Early Triassic N-limited conditions at Festningen appear to have also directly affected
363 primary productivity, as observed by the very low TOC levels despite recurrent anoxia (Grasby
364 et al., 2013; Wignall et al., 2016) that would have otherwise enhanced preservation of organic
365 matter in a productive margin. Through the Early Triassic TOC also closely tracks the N/P ratio,

366 with lower TOC values corresponding to drop in N/P (Fig. 4), further supporting that N-
367 limitation directly affected primary productivity. Drawdown of bio-essential micronutrients
368 related to Early Triassic anoxia may have placed even further stress on productivity at this time
369 (Grasby and Beauchamp, 2009). The EFs for primary productivity proxies Ba and Ni track each
370 other and show values that are consistent with a productive upwelling setting prior to the LPE,
371 that can shift to values <1 (indicating low productivity) through the Early Triassic. The EFs for
372 Cu are low throughout the section and may not reflect changes in bioproductivity. Overall, we
373 interpret these results to suggest that stable isotope evidence for Early Triassic N-limited
374 conditions is also manifest as reduced primary production along the northern margin of Pangea,
375 characterised by the overall organic-lean shales. This is also seen in the organic-lean Lower
376 Triassic Vega-Phroso Member shales and subsurface equivalents in NE British Columbia, that
377 have apparently high TOC due to secondary oil migration (Riediger et al., 1990; Riediger, 1997).

378 **Middle Triassic return to productive margins**

379 Following the Early Triassic accumulation of organic lean shales at Festningen, there was a
380 significant switch in the Middle Triassic to deposition of black organic-rich phosphatic shales
381 that are indicative of a change to a highly productive margin. This switch from organic lean to
382 organic-rich conditions is coincident with the end of the Early Triassic hothouse and return to
383 normal marine temperatures (Sun et al., 2012). This was a significant event as the black shale
384 deposited at Festningen is represented by the Botneheia Formation which forms a key source
385 rock of the Barents Sea oil fields (Krajewski, 2008). This is also correlative with the
386 development of other major black shale source rocks of Middle Triassic age, including the
387 Murray Harbour Formation of the Sverdrup Basin (Grasby et al., 2016b), the Shublik Formation
388 of Alaska (Parrish et al., 2001) and Doig Phosphate zone of western Canada (Riediger, 1997).
389 Similar Middle Triassic black shales are also found in the western Tethys (Bernasconi and Riva,

390 1993), suggesting that cooling oceans led to an increase of primary productivity and associated
391 drawdown and sequestration of atmospheric CO₂.

392 Similar to the Sverdrup Basin, the $\delta^{15}\text{N}$ values of Festningen still remained low through
393 the Middle Triassic. While we interpret the low $\delta^{15}\text{N}$ values of the Early Triassic to reflect
394 nutrient-stressed conditions leading to a productivity crisis, these N isotope values can only be
395 interpreted in the broader context of the shale geochemistry. In contrast to the organic lean Early
396 Triassic shales, the widespread formation of Middle Triassic source rocks requires a different
397 interpretation of the N isotope data. We argue that the low $\delta^{15}\text{N}$ values still represent N-limited
398 conditions, but in the sense that it reflects a shift to low N/P waters (as implied by abundant P
399 nodules) at that time. Thus, similar to Grasby et al. (2016b), the formation of Middle Triassic
400 organic rich source rocks is related to renewed upwelling of P-rich waters that had been trapped
401 by a deepened thermocline below the zone of upwelling through Early Triassic time – setting off
402 a phosphate bomb that drove high productivity and in turn draw down of atmospheric CO₂.
403 Although N₂ fixation has a high energy demand, when nutrient N concentrations are low relative
404 to nutrient P (i.e. N/P \ll Redfield ratio) diazotrophs can outcompete non-N₂ fixing algae
405 (Tyrrell, 1999). Thus a large flux of low N/P waters to the photic zone would have fueled
406 atmospheric nitrogen fixation to provide sufficient nutrient N supply to drive the observed
407 increased primary productivity (Grasby et al., 2016b), while still maintaining low $\delta^{15}\text{N}$ values
408 due to N-limited conditions (relative to P). This interpretation is similar to Cretaceous Ocean
409 Anoxic Events (OAEs), whereby atmospheric N₂ fixation was the dominant source of N, along
410 with remineralised P, driving high bioproductivity and organic carbon burial (Kuypers et al.,
411 2004).

412 **Global Implications**

413 Various indirect lines of argument have been used to suggest significantly reduced primary
414 productivity following the LPE, including significantly reduced fossil abundance (Twitchett et
415 al., 2001; Payne, 2005; Twitchett, 2007), reductions in body size in many fossil groups,
416 including conodonts (Luo et al., 2008), sponges (Liu et al., 2013), brachiopods (He et al., 2007;
417 He et al., 2010) and molluscs (Twitchett, 2007). In addition, the negative excursion in marine
418 carbonate $\delta^{13}\text{C}$ profiles (Rampino and Caldeira, 2005) has been used to argue for reduced
419 primary production, although other models such as volcanic emissions can also explain this
420 (Payne and Kump, 2007). Shen et al. (2014) also show proxies for primary productivity in south
421 China are consistent with decreased productivity across the LPE. Retallack (2004) also made the
422 interesting observation that most Lower Triassic sections show much lower TOC levels than
423 underlying Late Permian or overlying Middle Triassic rocks in the same sequence. Similarly, the
424 scarcity of organic-rich shale from Lower Triassic shelf sections has been noted by the global
425 search for petroleum source rocks (Tissot, 1979; Klemme and Ulmishek, 1991). Given evidence
426 for global anoxia, that would favour organic preservation, the scarcity of organic-rich shales
427 during the Early Triassic is consistent with overall reduced global primary productivity during
428 that time.

429 In contrast to the above, increased primary productivity after the LPE has been favoured by
430 several authors. Carbon isotope data, such as positive excursions in marine carbonate $\delta^{13}\text{C}$
431 profiles, have also been used to argue for increased productivity in at least parts of the Early
432 Triassic (Suzuki et al., 1998; Horacek et al., 2007a). As well, enhanced carbon-isotope depth
433 gradients in Lower Triassic limestone of the Tethys have been suggested to reflect increased
434 primary productivity (Meyer et al., 2011). However, Song et al. (2013) used similar data to make
435 the opposite argument, for reduced productivity in the immediate aftermath of the LPE.

436 Development of organic-rich mudstones in deep water Panthalassa sections (Kakuwa, 1996;
437 Suzuki et al., 1998; Sashida et al., 2000; Takemura et al., 2004) immediately following the LPE
438 suggests potential for enhanced mid-Panthalassa productivity, at least during the Griesbachian;
439 after which deposition switches back to organic lean conditions. Shen et al. (2015) also argued
440 for increased primary productivity from the late Changhsingian to the Griesbachian in most
441 regions of the globe, including northern Pangea, based on geochemical proxies. Their study,
442 however, is largely restricted to the absolute latest Permian (post LPE) or earliest Triassic
443 (Griesbachian) and does not provide insight into overall trends in the Early Triassic.

444 Most of the debate though has been based on observations of Tethyan sections. More
445 relevant to study of changes of marine primary productivity in the aftermath of the LPE is to
446 focus on the marine records of those areas affected by upwelling, i.e. the northern and western
447 margins of Pangea. To obtain a broader northern Pangea perspective, we compare here the
448 geochemical record from Spitsbergen and the Sverdrup Basin, with data plotted as a function of
449 time rather than stratigraphic thickness (Fig. 6). For Figure 6 we have excluded the basal portion
450 of the Festningen section as correlative data is not available. This figure shows consistent trends
451 from Late Permian through to Middle Triassic along the northern margin of Pangea. The region
452 was characterised by significant upwelling in the Late Permian (Beauchamp and Baud, 2002;
453 Knies et al., 2013; Schoepfer et al., 2013; Grasby et al., 2015; Grasby et al., 2016b), with a shift
454 to nutrient reduced conditions at the LPE. There is then an overall trend to greater nutrient stress
455 through the Early Triassic, albeit with some diachroneity. For instance, the shift to lower $\delta^{15}\text{N}$
456 values was earlier at Festningen (initiated in the Griesbachian) as compared to Smith Creek that
457 did not reach lower values until the early Dienerian. As well, minimum $\delta^{15}\text{N}$ values at
458 Festningen are $\sim 2\text{‰}$ higher than that recorded in the Smith Creek section of the Sverdrup Basin,

459 suggesting nutrient limitation may have been slightly greater in the Sverdrup Basin. Overall
460 though, our data demonstrate a significant shift occurred in northern Pangea, from N supply
461 related to water-column denitrification occurring in continental margin upwelling zones prior to
462 the LPE, to N supply related to atmospheric N fixation in the aftermath. This change is
463 coincident with the shift from Late Permian greenhouse to Early Triassic hothouse conditions
464 demonstrated by conodont $\delta^{18}\text{O}$ data (Sun et al., 2012).

465 Our observations of nutrient limitations in upwelling zones is consistent with the HEAT
466 model of Kidder and Worsley (2010), whereby global warming drives increased ocean anoxia,
467 that in turn leads to enhanced denitrification and anaerobic ammonium oxidation, favouring
468 atmospheric N_2 fixation by diazotrophs. As this is a more energy intensive process as well as Fe
469 limited, overall net primary productivity would be reduced. Our results also directly support
470 predictions of weakened trade winds in response to Early Triassic global warming, and a
471 coincident decline in Ekman transport of nutrients to the photic zone and primary productivity
472 (Winguth et al., 2015). As such, our data for northern Pangea are consistent with hothouse Earth
473 conditions creating extremely stressed marine environments that limited biodiversity and overall
474 bioproductivity. Our model is similar to concerns expressed for modern global warming leading
475 to deep ocean nutrient trapping and drastic decline in shallow water biological productivity of the
476 ocean (Kamykowski and Zentara, 1986; Bopp et al., 2001; Sarmiento et al., 2004; Behrenfeld et
477 al., 2006; Cermeño et al., 2008; Doney et al., 2012; Moore et al., 2018).

478 During the time of the single super continent Pangea, eastern boundary currents would
479 have been limited to the Panthalassa Ocean (Fig. 2a), suggesting that upwelling zones and
480 associated primary productivity would have been restricted to an even smaller extent of world
481 oceans than today – largely along the western and northern margin of Pangea, but also possibly

482 in the Tethys as zones of equatorial upwelling (Kidder and Worsley, 2004; Grasby et al., 2016b).
483 While decline in nutrient upwelling in northern Pangea appears to have placed direct stress on
484 local marine ecosystems, significantly reducing primary productivity across northern Pangea,
485 broader global trends remain to be determined. Similar detailed studies of nutrient stress through
486 the Early Triassic in Tethyan sections are required. However, given the disproportionate role that
487 upwelling zones have on net biologic productivity, reduced upwelling along northern Pangea
488 would likely have had global impacts, and contributed to overall marine stress and prolonged
489 biotic recovery during the Early Triassic hothouse. A significant decline in marine productivity
490 could also have had larger global effects, as shown by Winguth et al. (2015) who demonstrated
491 that it could alter global climate (Andreae, 2007) and sustain hothouse conditions.

492 Intriguingly, N₂ fixation is also suggested to have provided sufficient nutrient N to drive
493 the carbon pump to effectively reduce atmospheric CO₂ levels during the mid-Cretaceous
494 greenhouse (Kuypers et al., 1999; Kuypers et al., 2004). The Anisian is marked by a similar
495 widespread deposition of black shales driven by atmospheric N₂ fixation marking the end of the
496 Early Triassic Hothouse. We suggest then that renewed upwelling of P-rich waters along with
497 atmospheric N₂ fixation may have played a similar major role in driving the carbon pump to
498 reduce the detrimental effects of Early Triassic global warming.

499 Our results can also provide some insight into global carbon budgets. Despite numerous
500 attempts to explain significant shifts in carbonate isotope values through the Early Triassic, there
501 is no consistent view on overall drivers. Our work is similar to that of Grasby et al. (2013) who
502 showed organic carbon isotope records of northern Pangea are consistent with the inorganic
503 records of the Neo-Tethys and Tethys regions (Fig. 7), supporting an external driver for negative
504 carbon excursions such as eruption of the Siberian Traps (e.g. Payne and Kump, 2007). More

505 relevant is that our results from Festningen, as well as the Sverdrup Basin (Grasby et al., 2016b),
506 indicate that the large carbon isotope excursions through the Early Triassic are not related to
507 changes in nutrient levels as shown by the Nitrogen isotope records (Figs. 6, 7). These results
508 suggest some degree of decoupling of the carbon cycle from nutrient levels, whereby low
509 productivity and carbon burial in the Early Triassic occurred as nutrients were trapped in the
510 deep ocean by a depressed thermocline (Grasby et al., 2016b), and primary productivity was
511 limited by low rates of N-fixation. If correct, this lends further support to an external driver such
512 as volcanism for the large carbon isotope excursions observed through the Early Triassic.

513 **CONCLUSIONS**

514 Geochemical records from the Festningen section on Spitsbergen provide new insight into the
515 evolution of the ocean nutrient levels and bioproductivity from Late Permian through to Middle
516 Triassic time. Our results show that under Early Triassic Hothouse conditions there was a
517 progressive decrease in nutrient levels, whereby nutrient-N became restricted and dominantly
518 sourced from atmospheric N₂ fixation. This nutrient stress appears to have also reduced primary
519 productivity, yet shows no relationship to major swings in the global C isotope record through
520 Early Triassic time. These patterns are observed across the margin of northern Pangea, which at
521 that time in Earth history should have been a major zone of bioproductivity in world oceans,
522 such that nutrient stress in this region would have had global impact, and may have played a
523 significant role in the delayed recovery of life following the Latest Permian mass extinction. The
524 apparent decoupling of decreased N-nutrient levels along with bioproductivity from the carbon
525 isotope record supports an external driver such as volcanic emissions to explain major swings in
526 $\delta^{13}\text{C}$ through the Early Triassic. These conditions were ameliorated in the Middle Triassic as

527 marked by return of highly productive margins across northern Pangea, which were coincident
528 with final cooling of world oceans, radiation of life, and return of normal marine ecosystems.

529 Our findings may also provide insight into concerns over the impact of modern climate
530 warming as both increased wind shear, driving increased upwelling, as well as deepening of the
531 nutricline are predicted. These two impacts would have opposing effects on marine productivity.
532 Our study suggests that in the Early Triassic the deepening of the nutricline offset any increased
533 physical upwelling such that warming oceans saw a net decline in primary productivity. Our
534 results are consistent with predicted impacts of global warming on modern marine
535 bioproductivity (e.g. Moore et al., 2018), although while those models suggest millennial
536 timescale impacts, our results indicate productivity was impacted for millions of years following
537 the LPE.

538

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REFERENCES

- 547 Ader, M., Boudou, J.-P., Javoy, M., Goffe, B., and Daniels, E., 1998, Isotope study on organic nitrogen of
548 Westphalian anthracites from the Western Middle field of Pennsylvania (U.S.A.) and from the Bramsche
549 Massif (Germany): *Organic Geochemistry*, v. 29, no. 1–3, p. 315-323.
- 550 Algeo, T. J., Hannigan, R., Rowe, H., Brookfield, M., Baud, A., Krystyn, L., and Ellwood, B. B., 2007, Sequencing
551 events across the Permian–Triassic boundary, Guryul Ravine (Kashmir, India): *Palaeogeography,*
552 *Palaeoclimatology, Palaeoecology*, v. 252, no. 1, p. 328-346.
- 553 Algeo, T. J., Meyers, P. A., Robinson, R. S., Rowe, H., and Jiang, G. Q., 2014, Icehouse–greenhouse variations in
554 marine denitrification *Biogeosciences*, v. 11, p. 1273-1295.
- 555 Algeo, T. J., and Twitchett, R. J., 2010, Anomalous Early Triassic sediment fluxes due to elevated weathering rates
556 and their biological consequences: *Geology*, v. 38, no. 11, p. 1023-1026.
- 557 Altabet, M. A., Murray, D. W., and Prell, W. L., 1999a, Climatically linked oscillations in Arabian Sea
558 denitrification over the past 1 m.y.: Implications for the marine N cycle: *Paleoceanography*, v. 14, no. 6, p.
559 732-743.

560 Altabet, M. A., Pilskaln, C., Thunell, R., Pride, C., Sigman, D., Chavez, F., and Francois, R., 1999b, The nitrogen
561 isotope biogeochemistry of sinking particles from the margin of the Eastern North Pacific: Deep Sea
562 Research Part I: Oceanographic Research Papers, v. 46, no. 4, p. 655-679.

563 Anbar, A. D., and Knoll, A. H., 2004, Proterozoic ocean chemistry and evolution: a bioinorganic bridge?: Science,
564 v. 297, p. 1137-1142.

565 Andreae, M. O., 2007, Aerosols Before Pollution: Science, v. 315, no. 5808, p. 50-51.

566 Bakun, A., 1990, Global Climate Change and Intensification of Coastal Ocean Upwelling: Science, v. 247, no. 4939,
567 p. 198-201.

568 Beauchamp, B., and Baud, A., 2002, Growth and demise of Permian biogenic chert along Northwest Pangea;
569 evidence for end-Permian collapse of thermohaline circulation Palaeogeography, Palaeoclimatology,
570 Palaeoecology, v. 184, no. 1-2, p. 37-63.

571 Beauchamp, B., and Grasby, S. E., 2012, Permian lysocline shoaling and ocean acidification along NW Pangea led
572 to carbonate eradication and chert expansion: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 350-
573 352, p. 73-90.

574 Beauchamp, B., Henderson, C. M. B., Grasby, S. E., Gates, L., Beatty, T., Utting, J., and James, N. P., 2009, Late
575 Permian sedimentation in the Sverdrup Basin, Canadian Arctic: the Lindström and Black Stripe formations:
576 Canadian Society of Petroleum Geology Bulletin, v. 57, p. 167-191.

577 Behrenfeld, M. J., O'Malley, R. T., Siegel, D. A., McClain, C. R., Sarmiento, J. L., Feldman, G. C., Milligan, A. J.,
578 Falkowski, P. G., Letelier, R. M., and Boss, E. S., 2006, Climate-driven trends in contemporary ocean
579 productivity: Nature, v. 444, no. 7120, p. 752-755.

580 Bernasconi, S., and Riva, A., 1993, Organic Geochemistry and Depositional Environment of a Hydrocarbon Source
581 Rock: The Middle Triassic Grenzbitumenzone Formation, Southern Alps, Italy/Switzerland, in Spencer, A.
582 M., ed., Generation, Accumulation and Production of Europe's Hydrocarbons III: Special Publication of
583 the European Association of Petroleum Geoscientists No. 3: Berlin, Heidelberg, Springer Berlin
584 Heidelberg, p. 179-190.

585 Blomeier, D., Dustira, A. M., Forke, H., and Scheibner, C., 2013, Facies analysis and depositional environments of a
586 storm-dominated, temperate to cold, mixed siliceous-carbonate ramp: the Permian Kapp Starostin
587 Formation in NE Svalbard: Norwegian Journal of Geology, v. 93, p. 75-98.

588 Bond, D. P. G., and Grasby, S. E., 2017, On the causes of mass extinctions: Palaeogeography, Palaeoclimatology,
589 Palaeoecology, v. 478, no. Supplement C, p. 3-29.

590 Bond, D. P. G., Wignall, P. B., Joachimski, M. M., Sun, Y., Savov, I., Grasby, S. E., Beauchamp, B., and Blomeier,
591 D. P. G., 2015, An abrupt extinction in the Middle Permian (Capitanian) of the Boreal Realm (Spitsbergen)
592 and its link to anoxia and acidification: Geological Society of America Bulletin.

593 Bopp, L., Monfray, P., Aumont, O., Dufresne, J.-L., Le Treut, H., Madec, G., Terray, L., and Orr, J. C., 2001,
594 Potential impact of climate change on marine export production: Global Biogeochemical Cycles, v. 15, no.
595 1, p. 81-99.

596 Bottjer, D. J., Clapham, M. E., Fraiser, M. L., and Powers, C. M., 2008, Understanding mechanisms for the end-
597 Permian mass extinction and the protracted Early Triassic aftermath and recovery: GSA Today, v. 18, no.
598 9, p. 4-10.

599 Brandes, J. A., Devol, A. H., and Deutsch, C., 2007, New Developments in the Marine Nitrogen Cycle: Chemical
600 Reviews, v. 107, no. 2, p. 577-589.

601 Cao, C., Love, G. D., Hays, L. E., Wang, W., Shen, S., and Summons, R. E., 2009, Biogeochemical evidence for
602 euxinic oceans and ecological disturbance presaging the end-Permian mass extinction event: Earth and
603 Planetary Science Letters, v. 281, no. 3, p. 188-201.

604 Capone, D. G., and Hutchins, D. A., 2013, Microbial biogeochemistry of coastal upwelling regimes in a changing
605 ocean: Nature Geosci, v. 6, no. 9, p. 711-717.

606 Carpenter, E. J., Harvey, H. R., Fry, B., and Capone, D. G., 1997, Biogeochemical tracers of the marine
607 cyanobacterium Trichodesmium: Deep Sea Research Part I: Oceanographic Research Papers, v. 44, no. 1,
608 p. 27-38.

609 CASE-Team, 2001, The evolution of the West Spitsbergen Fold-and-Thrust Belt: Geologisches Jahrbuch, v. B91, p.
610 733-773.

611 Cermeño, P., Dutkiewicz, S., Harris, R. P., Follows, M., Schofield, O., and Falkowski, P. G., 2008, The role of
612 nutricline depth in regulating the ocean carbon cycle: Proceedings of the National Academy of Sciences, v.
613 105, no. 51, p. 20344-20349.

614 Chen, Z.-Q., and Benton, M. J., 2012, The timing and pattern of biotic recovery following the end-Permian mass
615 extinction: Nature Geosci, v. 5, no. 6, p. 375-383.

- 616 Corfu, F., Polteau, S., Planke, S., Faleide, J. I., Svensen, H., Zayoncheck, A., and Stolbov, N., 2013, U–Pb
617 geochronology of Cretaceous magmatism on Svalbard and Franz Josef Land, Barents Sea Large Igneous
618 Province: *Geological Magazine*, v. 150, p. 1127-1135.
- 619 Currie, R. I., 1992, Circulation and upwelling off the coast of south-east arabia: *Oceanologica Acta*, v. 15, p. 43-60.
- 620 Dallmann, W. K., Andresen, A., S.G., B., Maher, H. D., and Ohta, Y., 1993, Tertiary fold-and-thrust belt of
621 Spitsbergen, Svalbard.: *Norsk Polarinstitute Meddelelser*, v. 128, p. 1-46.
- 622 Deutsch, C., Sarmiento, J. L., Sigman, D. M., Gruber, N., and Dunne, J. P., 2007, Spatial coupling of nitrogen inputs
623 and losses in the ocean: *Nature*, v. 445, no. 7124, p. 163-167.
- 624 Doney, S. C., Ruckelshaus, M., Duffy, J. E., Barry, J. P., Chan, F., English, C. A., Galindo, H. M., Grebmeier, J. M.,
625 Hollowed, A. B., Knowlton, N., Polovina, J., Rabalais, N. N., Sydeman, W. J., and Talley, L. D., 2012,
626 Climate Change Impacts on Marine Ecosystems: *Annual Review of Marine Science*, v. 4, no. 1, p. 11-37.
- 627 Dymond, J., Suess, E., and Lyle, M., 1992, Barium in Deep-Sea Sediment: A Geochemical Proxy for
628 Paleoproductivity: *Paleoceanography*, v. 7, no. 2, p. 163-181.
- 629 Embry, A. F., 1989, Correlation of Upper Palaeozoic and Mesozoic sequences between Svalbard, Canadian Arctic
630 Archipelago, and northern Alaska, *Correlation in Hydrocarbon Exploration*, Springer Netherlands, p. 89-
631 98.
- 632 Embry, A. F., and Beauchamp, B., 2008, Sverdrup Basin, *in* Miall, A. D., ed., *The Sedimentary Basins of Unites*
633 *States and Canada*: Amsterdam, Elsevier, p. 451-472.
- 634 Erwin, D. H., Bowring, S. A., and Yugan, J., 2002, End-Permian mass extinctions: A review, *in* Koeberl, C., and
635 MacLeod, K. G., eds., *Catastrophic events and mass extinctions: Impacts and beyond*, Volume Geological
636 Society of America Special Paper 356, Geological Society of America, p. 363-383.
- 637 Galfetti, T., Bucher, H., Ovtcharova, M., Schaltegger, U., Brayard, A., Brühwiler, T., Goudemand, N., Weissert, H.,
638 Hochuli, P. A., Cordey, F., and Guodun, K., 2007, Timing of the Early Triassic carbon cycle perturbations
639 inferred from new U-Pb ages and ammonoid biochronozones: *Earth and Planetary Science Letters*, v. 258,
640 no. 3-4, p. 593-604.
- 641 Ganeshram, R. S., Pedersen, T. F., Calvert, S. E., McNeill, G. W., and Fontugne, M. R., 2000, Glacial-interglacial
642 variability in denitrification in the World's Oceans: Causes and consequences: *Paleoceanography*, v. 15, no.
643 4, p. 361-376.
- 644 Golonka, J., and Ford, D., 2000, Pangean (Late Carboniferous-Middle Jurassic) paleoenvironment and lithofacies:
645 *Palaeogeography, Palaeoclimatology, Palaeoecology* v. 161, p. 1-34.
- 646 Grasby, S. E., and Beauchamp, B., 2009, Latest Permian to Early Triassic basin-to-shelf anoxia in the Sverdrup
647 Basin, Arctic Canada *Chemical Geology*, v. 264, p. 232-246.
- 648 Grasby, S. E., Beauchamp, B., Bond, D. P. G., Wignall, P. B., and Sanei, H., 2016a, Mercury anomalies associated
649 with three extinction events (Capitanian Crisis, Latest Permian Extinction and the Smithian/Spathian
650 Extinction) in NW Pangea: *Geological Magazine*, v. 153, no. 2, p. 285-297.
- 651 Grasby, S. E., Beauchamp, B., Bond, D. P. G., Wignall, P. B., Talavera, C., Galloway, J. M., Piepjohn, K.,
652 Reinhardt, L., and Blomeier, D., 2015, Progressive environmental deterioration in NW Pangea leading to
653 the Latest Permian Extinction: *Geological Society of America Bulletin*, v. 127, no. 9/10, p. 1331-1347.
- 654 Grasby, S. E., Beauchamp, B., Embry, A. F., and Sanei, H., 2013, Recurrent Early Triassic ocean anoxia: *Geology*,
655 v. 41, p. 175-178.
- 656 Grasby, S. E., Beauchamp, B., and Knies, J., 2016b, Early Triassic productivity crises delayed recovery from
657 world's worst mass extinction: *Geology*, v. 44, no. 9, p. 779-782.
- 658 Grice, K., Cao, C., Love, G. D., Böttcher, M. E., Twitchett, R. J., Grosjean, E., Summons, R. E., Turgeon, S. C.,
659 Dunning, W., and Jin, Y., 2005, Photic zone euxinia during the Permian-Triassic superanoxic event:
660 *Science*, v. 307, p. 706-709.
- 661 Hays, L., Beatty, T., Henderson, C. M. B., Love, G. D., and Summons, R. E., 2007, Evidence for photic zone
662 euxinia through the end-Permian mass extinction in the Panthalassic Ocean (Peace River Basin, Western
663 Canada): *Palaeoworld*, v. 16, p. 39-50.
- 664 Hays, L. E., Grice, K., Foster, C. B., and Summons, R. E., 2012, Biomarker and isotopic trends in a Permian–
665 Triassic sedimentary section at Kap Stosch, Greenland: *Organic Geochemistry*, v. 43, no. Supplement C, p.
666 67-82.
- 667 He, W., Shi, G. R., Feng, Q., Campi, M. J., Gu, S., Bu, J., Peng, Y., and Meng, Y., 2007, Brachiopod
668 miniaturization and its possible causes during the Permian–Triassic crisis in deep water environments,
669 South China: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 252, no. 1, p. 145-163.
- 670 He, W. H., Twitchett, R. J., Zhang, Y., Shi, G. R., Feng, Q. L., Yu, J. X., Wu, S. B., and Peng, X. F., 2010, Controls
671 on body size during the Late Permian mass extinction event: *Geobiology*, v. 8, no. 5, p. 391-402.

- 672 Horacek, M., Brandner, R., and Abart, R., 2007a, Carbon isotope record of the P/T boundary and the Lower Triassic
673 in the Southern Alps: Evidence for rapid changes in storage of organic carbon: *Palaeogeography,*
674 *Palaeoclimatology, Palaeoecology*, v. 252, no. 1-2, p. 347-354.
- 675 Horacek, M., Richoz, S., Brandner, R., Krystyn, L., and Spötl, C., 2007b, Evidence for recurrent changes in Lower
676 Triassic oceanic circulation of the Tethys: The $\delta^{13}\text{C}$ record from marine sections in Iran:
677 *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 252, no. 1-2, p. 355-369.
- 678 Hounslow, M. W., Hu, M., Mørk, A., Vigran, J. O., Weitschat, W., and Orchard, M. J., 2007, Magneto-
679 biostratigraphy of the Middle to Upper Triassic transition, central Spitsbergen, arctic Norway: *Journal of*
680 *the Geological Society*, v. 164, no. 3, p. 581-597.
- 681 Huyer, A., 1983, Coastal upwelling in the California current system: *Progress in Oceanography*, v. 12, no. 3, p. 259-
682 284.
- 683 Jia, C., Huang, J., Kershaw, S., Luo, G., Farabegoli, E., Perri, M. C., Chen, L., Bai, X., and Xie, S., 2012, Microbial
684 response to limited nutrients in shallow water immediately after the end-Permian mass extinction:
685 *Geobiology*, v. 10, no. 1, p. 60-71.
- 686 Joachimski, M. M., Lai, X., Shen, S., Jiang, H., Luo, G., Chen, B., Chen, J., and Sun, Y., 2012, Climate warming in
687 the latest Permian and the Permian–Triassic mass extinction: *Geology*, v. 40, no. 3, p. 195-198.
- 688 Kakuwa, Y., 1996, Permian-Triassic mass extinction event recorded in bedded chert sequence in southwest Japan:
689 *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 121, no. 1, p. 35-51.
- 690 Kamykowski, D., and Zentara, S.-J., 1986, Predicting plant nutrient concentrations from temperature and sigma-t in
691 the upper kilometer of the world ocean: *Deep Sea Research Part A. Oceanographic Research Papers*, v. 33,
692 no. 1, p. 89-105.
- 693 Kidder, D. L., and Worsley, T. R., 2004, Causes and consequences of extreme Permo-Triassic warming to globally
694 equable climate and relation to the Permo-Triassic extinction and recovery: *Palaeogeography,*
695 *Palaeoclimatology, Palaeoecology*, v. 203, p. 207-237.
- 696 -, 2010, Phanerozoic Large Igneous Provinces (LIPs), HEATT (Haline Euxinic Acidic Thermal Transgression)
697 episodes, and mass extinctions: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 295, no. 1-2, p.
698 162-191.
- 699 Klemme, H. D., and Ulmishek, G. F., 1991, Effective Petroleum Source Rocks of the World: Stratigraphic
700 Distribution and Controlling Depositional Factors (1): *AAPG Bulletin*, v. 75, no. 12, p. 1809-1851.
- 701 Knies, J., Grasby, S. E., Beauchamp, B., and Schubert, C., 2013, Water mass denitrification during the Latest
702 Permian Extinction in the Sverdrup Basin, Arctic Canada: *Geology* v. 41, p. 167-170.
- 703 Knoll, A. H., Bambach, R. K., Payne, J. L., Pruss, S., and Fischer, W. W., 2007, Paleophysiology and end-Permian
704 mass extinction: *Earth and Planetary Science Letters*, v. 256, p. 295-313.
- 705 Krajewski, K. P., 2008, The Botneheia Formation [Middle Triassic] in Edgeoya and Barentsoya, Svalbard:
706 lithostratigraphy, facies, phosphogenesis, paleoenvironment *Polish Polar Research* v. 29, no. 4, p. 319-364.
- 707 Kuypers, M. M. M., Pancost, R. D., and Damsté, J. S. S., 1999, A large and abrupt fall in atmospheric CO₂
708 concentration during Cretaceous times: *Nature*, v. 399, p. 342.
- 709 Kuypers, M. M. M., van Breugel, Y., Schouten, S., and Erba, E., 2004, N₂-fixing cyanobacteria supplied nutrient N
710 for Cretaceous oceanic anoxic events: *Geology*, v. 32, no. 10, p. 853-856.
- 711 Liu, G., Feng, Q., Shen, J. U. N., Yu, J., He, W., and Algeo, T. J., 2013, Decline of siliceous sponges and spicule
712 miniaturization induced by marine productivity collapse and expanding anoxia during the Permian-Triassic
713 crises in south China: *PALAIOS*, v. 28, no. 8, p. 664-679.
- 714 Luo, G., Lai, X., Shi, G. R., Jiang, H., Yin, H., Xie, S., Tong, J., Zhang, K., He, W., and Wignall, P. B., 2008, Size
715 variation of conodont elements of the Hindeodus–Isarcicella clade during the Permian–Triassic transition in
716 South China and its implication for mass extinction: *Palaeogeography, Palaeoclimatology, Palaeoecology*,
717 v. 264, no. 1, p. 176-187.
- 718 Luo, G., Wang, Y., Algeo, T. J., Kump, L. R., Bai, X., Yang, H., Yao, L., and Xie, S., 2011, Enhanced nitrogen
719 fixation in the immediate aftermath of the latest Permian marine mass extinction: *Geology*, v. 39, no. 7, p.
720 647-650.
- 721 Maher, H. D., and Craddock, C., 1988, Decoupling as an alternate model for transgression during the initial opening
722 of the Norwegian Greenland Sea: *Polar Research*, v. 6, p. 137-140.
- 723 Mettam, C., Zerkle, A. L., Claire, M. W., Izon, G., Junium, C. J., and Twitchett, R. J., 2017, High-frequency
724 fluctuations in redox conditions during the latest Permian mass extinction: *Palaeogeography,*
725 *Palaeoclimatology, Palaeoecology*, v. 485, no. Supplement C, p. 210-223.

726 Meyer, K. M., Yu, M., Jost, A. B., Kelley, B. M., and Payne, J. L., 2011, $\delta^{13}\text{C}$ evidence that high primary
727 productivity delayed recovery from end-Permian mass extinction: *Earth and Planetary Science Letters*, v.
728 302, no. 3–4, p. 378-384.

729 Midwinter, D., Hadlari, T., and Dewing, K., 2017, Lower Triassic river-dominated deltaic successions from the
730 Sverdrup Basin, Canadian Arctic: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 476, p. 55-67.

731 Moore, J. K., Fu, W., Primeau, F., Britten, G. L., Lindsay, K., Long, M., Doney, S. C., Mahowald, N., Hoffman, F.,
732 and Randerson, J. T., 2018, Sustained climate warming drives declining marine biological productivity:
733 *Science*, v. 359, no. 6380, p. 1139-1143.

734 Mørk, A., Knarud, R., and Worsley, D., 1982, Depositional and diagenetic environments of the Triassic and Lower
735 Jurassic succession of Svalbard, *in* Embry, A. F., and Balkwill, H. R., eds., *Arctic geology and geophysics:*
736 *proceedings of the Third International Symposium on Arctic Geology: Calgary, Canadian Society of*
737 *Petroleum Geologists*, p. 371-398.

738 Parrish, J. T., Droser, M. L., and Bottjer, D. J., 2001, A Triassic Upwelling Zone: The Shublik Formation, Arctic
739 Alaska, U.S.A: *Journal of Sedimentary Research*, v. 71, no. 2, p. 272-285.

740 Payne, J. L., 2005, Evolutionary dynamics of gastropod size across the end-Permian extinction and through the
741 Triassic recovery interval: *Paleobiology*, v. 31, no. 2, p. 269-290.

742 Payne, J. L., and Kump, L. R., 2007, Evidence for recurrent Early Triassic massive volcanism from quantitative
743 interpretation of carbon isotope fluctuations: *Earth and Planetary Science Letters*, v. 256, p. 264-277.

744 Payne, J. L., Lehrmann, D., J., Wei, J., Orchard, M. J., Schrag, D. P., and Knoll, A. H., 2004, Large perturbations of
745 the carbon cycle during recovery from the End-Permian extinction: *Science*, v. 305, p. 506-509.

746 Rampino, M. R., and Caldeira, K., 2005, Major perturbation of ocean chemistry and a ‘Strangelove Ocean’ after the
747 end-Permian mass extinction: *Terra Nova*, v. 17, no. 6, p. 554-559.

748 Reid, C. M., James, N. P., Beauchamp, B., and Kyser, T. K., 2007, Faunal turnover and changing oceanography:
749 Late Palaeozoic warm-to-cool water carbonates, Sverdrup Basin, Canadian Arctic Archipelago:
750 *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 249, p. 128-159.

751 Retallack, G. J., 2004, Comment—Contrasting Deep-water Records from the Upper Permian and Lower Triassic of
752 South Tibet and British Columbia: Evidence for a Diachronous Mass Extinction (Wignall and Newton,
753 2003): *PALAIOS*, v. 19, no. 1, p. 101-102.

754 Retallack, G. J., Veevers, J. J., and Morante, R., 1996, Global coal gap between Permian-Triassic extinction and
755 Middle Triassic recovery of peat-forming plants: *Geological Society of America Bulletin*, v. 108, p. 195-
756 207.

757 Riediger, C. L., 1997, Geochemistry of Potential Hydrocarbon Source Rocks of Triassic Age in the Rocky Mountain
758 Foothills of Northeastern British Columbia and West-Central Alberta: *Bulletin of Canadian Petroleum*
759 *Geology*, v. 45, no. 4, p. 719-741.

760 Riediger, C. L., Brooks, P. W., M.G., F., and Snowdon, L. R., 1990, Lower and Middle Triassic source rocks,
761 thermal maturation, and oil-source rock correlations in the Peace River Embayment area, Alberta and
762 British Columbia: *Bulletin of Canadian Petroleum Geology*, v. 38A, no. 1, p. 218-235.

763 Roberts, J., Turchyn, A. V., Wignall, P. B., Newton, R. J., and Vane, C. H., 2018, Disentangling diagenesis from the
764 rock record: an example from the Permo-Triassic Wordie Creek Formation, East Greenland: *Geochemistry,*
765 *Geophysics, Geosystems*.

766 Saitoh, M., Ueno, Y., Nishizawa, M., Isozaki, Y., Takai, K., Yao, J., and Ji, Z., 2014, Nitrogen isotope
767 chemostratigraphy across the Permian–Triassic boundary at Chaotian, Sichuan, South China: *Journal of*
768 *Asian Earth Sciences*, v. 93, no. Supplement C, p. 113-128.

769 Saltzman, M. R., 2005, Phosphorus, nitrogen, and the redox evolution of the Paleozoic oceans: *Geology*, v. 33, no.
770 7, p. 573-576.

771 Sanson-Barrera, A., Hochuli, P. A., Bucher, H., Schneebeli-Hermann, E., Weissert, H., Adatte, T., and Bernasconi,
772 S. M., 2015, Late Permian–earliest Triassic high-resolution organic carbon isotope and palynofacies
773 records from Kap Stosch (East Greenland): *Global and Planetary Change*, v. 133, no. Supplement C, p.
774 149-166.

775 Sarmiento, J. L., Slater, R., Barber, R., Bopp, L., Doney, S. C., Hirst, A. C., Kleypas, J., Matear, R., Mikolajewicz,
776 U., Monfray, P., Soldatov, V., Spall, S. A., and Stouffer, R., 2004, Response of ocean ecosystems to
777 climate warming: *Global Biogeochemical Cycles*, v. 18, no. 3.

778 Sashida, K., Igo, H., Adachi, S., Ueno, K., Kajiwara, Y., Nakornsri, N., and Sardud, A., 2000, Late Permian to
779 Middle Triassic Radiolarian Faunas from Northern Thailand: *Journal of Paleontology*, v. 74, no. 5, p. 789-
780 811.

781 Schobben, M., Stebbins, A., Ghaderi, A., Strauss, H., Korn, D., and Korte, C., 2015, Flourishing ocean drives the
782 end-Permian marine mass extinction: *Proceedings of the National Academy of Sciences of the United*
783 *States of America*, v. 112, no. 33, p. 10298-10303.

784 Schoepfer, S. D., Henderson, C. M., Garrison, G. H., Foriel, J., Ward, P. D., Selby, D., Hower, J. C., Algeo, T. J.,
785 and Shen, Y., 2013, Termination of a continent-margin upwelling system at the Permian–Triassic boundary
786 (Opal Creek, Alberta, Canada): *Global and Planetary Change*, v. 105, no. 0, p. 21-35.

787 Scotese, C. R., 2004, A continental drift flipbook: *Journal of Geology*, v. 112, p. 729-741.

788 Sephton, M. A., Looy, C. V., Brinkhuis, H., Wignall, P. B., de Leeuw, J. W., and Visscher, H., 2005, Catastrophic
789 soil erosion during the end-Permian biotic crisis *Geology*, v. 33, no. 12, p. 941-944.

790 Sharp, C. E., Brady, A. L., Sharp, G. H., Grasby, S. E., Stott, M. B., and Dunfield, P. F., 2014, Humboldt's spa:
791 microbial diversity is controlled by temperature in geothermal environments: *ISME J*, v. 8, no. 6, p. 1166-
792 1174.

793 Shen, J., Schoepfer, S. D., Feng, Q., Zhou, L., Yu, J., Song, H., Wei, H., and Algeo, T. J., 2015, Marine productivity
794 changes during the end-Permian crisis and Early Triassic recovery: *Earth-Science Reviews*, v. 149, p. 136-
795 162.

796 Shen, J., Zhou, L., Feng, Q., Zhang, M., Lei, Y., Zhang, N., Yu, J., and Gu, S., 2014, Paleo-productivity evolution
797 across the Permian-Triassic boundary and quantitative calculation of primary productivity of black rock
798 series from the Dalong Formation, South China: *Science China Earth Sciences*, v. 57, no. 7, p. 1583-1594.

799 Smith, R. L., 1995, The physical processes of coastal ocean upwelling systems, *in* Summerhayes, C. P., Emeis, K.-
800 C., Angel, M. V., Smith, R. L., and Zeitzschel, B., eds., *Upwelling in the ocean: modern processes and*
801 *ancient records*, Volume 18, Wiley, p. 39–64.

802 Somes, C. J., Schmittner, A., Galbraith, E. D., Lehmann, M. F., Altabet, M. A., Montoya, J. P., Letelier, R. M., Mix,
803 A. C., Bourbonnais, A., and Eby, M., 2010, Simulating the global distribution of nitrogen isotopes in the
804 ocean: *Global Biogeochemical Cycles*, v. 24, no. 4.

805 Song, H., Tong, J., Algeo, T. J., Horacek, M., Qiu, H., Song, H., Tian, L., and Chen, Z.-Q., 2013, Large vertical
806 $\delta^{13}\text{C}_{\text{DIC}}$ gradients in Early Triassic seas of the South China craton: Implications for oceanographic
807 changes related to Siberian Traps volcanism: *Global and Planetary Change*, v. 105, p. 7-20.

808 Steiner, Z., Lazar, B., Torfstein, A., and Erez, J., 2017, Testing the utility of geochemical proxies for
809 paleoproductivity in oxic sedimentary marine settings of the Gulf of Aqaba, Red Sea: *Chemical Geology*,
810 v. 473, no. Supplement C, p. 40-49.

811 Stemmerik, L., and Worsley, D., 1995, Permian History of the Barents Shelf Area, *in* Scholle, P., Peryt, T., and
812 Ulmer-Scholle, D., eds., *The Permian of Northern Pangea*, Springer Berlin Heidelberg, p. 81-97.

813 Stemmerik, L., and Worsley, D., 2005, 30 years on - Arctic Upper Palaeozoic stratigraphy, depositional evolution
814 and hydrocarbon prospectivity: *Norsk Geologisk Tidsskrift* v. 85, p. 151-168.

815 Sun, Y., Joachimski, M. M., Wignall, P. B., Yan, C., Chen, Y., Jiang, H., Wang, L., and Lai, X., 2012, Lethally Hot
816 Temperatures During the Early Triassic Greenhouse: *Science*, v. 338, no. 6105, p. 366-370.

817 Suzuki, N., Ishida, K., Shinomiya, Y., and Ishiga, H., 1998, High productivity in the earliest Triassic ocean: black
818 shales, Southwest Japan: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 141, no. 1, p. 53-65.

819 Takemura, S., Sakamoto, S., Takemura, A., Nishimura, T., Aita, Y., Yamakita, S., Kamata, Y., Spörl, K. B.,
820 Campbell, H. J., Sakai, T., Suzuki, N., Hori, R., Sakakibara, M., Ogane, K., Kodama, K., and Nakamura,
821 Y., 2004, Lithofacies of Middle to Late Permian pelagic sedimentary rocks at Arrow Rocks, North Island,
822 New Zealand: *News Osaka Micropaleontologists (NOM)*, v. 13, p. 21-28.

823 Tissot, B., 1979, Effects on prolific petroleum source rocks and major coal deposits caused by sea-level changes:
824 *Nature*, v. 277, no. 5696, p. 463-465.

825 Trappe, J., 1994, Pangean Phosphorites - Ordinary Phosphorite Genesis in an Extraordinary World?, *in* Embry, A.
826 F., Beauchamp, B., and Glass, D. J., eds., *Pangea: Global Environments and Resources*: Calgary, Alberta,
827 Canadian Society of Petroleum Geologists, p. 469-478.

828 Tribouillard, N., Algeo, T. J., Lyons, T., and Riboulleau, A., 2006, Trace metals as paleoredox and
829 paleoproductivity proxies: An update: *Chemical Geology*, v. 232, no. 1–2, p. 12-32.

830 Twitchett, R. J., 2007, The Lilliput effect in the aftermath of the end-Permian extinction event: *Palaeogeography,*
831 *Palaeoclimatology, Palaeoecology*, v. 252, no. 1, p. 132-144.

832 Twitchett, R. J., Looy, C. V., Morante, R., Visscher, H., and Wignall, P. B., 2001, Rapid and synchronous collapse of
833 marine and terrestrial ecosystems during the end-Permian biotic crisis: *Geology*, v. 29, no. 4, p. 351-354.

834 Tyrrell, T., 1999, The relative influences of nitrogen and phosphorus on oceanic primary production: *Nature*, v. 400,
835 no. 6744, p. 525-531.

836 Walsh, J. J., 1991, Importance of continental margins in the marine biogeochemical cycling of carbon and nitrogen:
837 Nature, v. 350, no. 6313, p. 53-55.

838 Ward, B. B., Devol, A. H., Rich, J. J., Chang, B. X., Bulow, S. E., Naik, H., Pratihary, A., and Jayakumar, A., 2009,
839 Denitrification as the dominant nitrogen loss process in the Arabian Sea: Nature, v. 461, p. 78.

840 Wedepohl, K. H., 1995, The composition of the continental crust: Geochimica et Cosmochimica Acta, v. 59, no. 7,
841 p. 1217-1232.

842 Wignall, P. B., Bond, D. P. G., Sun, Y., Grasby, S. E., Beauchamp, B., Joachimski, M. M., and Blomeier, D. P. G.,
843 2015, Ultra-shallow-marine anoxia in an Early Triassic shallow-marine clastic ramp (Spitsbergen) and the
844 suppression of benthic radiation: Geological Magazine, v. 153, no. 2, p. 316-331.

845 -, 2016, Ultra-shallow-marine anoxia in an Early Triassic shallow-marine clastic ramp (Spitsbergen) and the
846 suppression of benthic radiation: Geological Magazine, v. 153, no. 2, p. 316-331.

847 Wignall, P. B., Morante, R., and Newton, R., 1998, The Permo-Triassic transition in Spitsbergen: $\delta^{13}\text{C}_{\text{org}}$
848 chemostratigraphy, Fe and S geochemistry, facies, fauna and trace fossils: Geologic Magazine, v. 135, p.
849 47-62.

850 Winguth, A. M. E., Shields, C. A., and Winguth, C., 2015, Transition into a Hothouse World at the Permian–
851 Triassic boundary—A model study: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 440, p. 316-
852 327.

853 Xiang, L., Schoepfer, S. D., Zhang, H., Yuan, D.-x., Cao, C.-q., Zheng, Q.-f., Henderson, C. M., and Shen, S.-z.,
854 2016, Oceanic redox evolution across the end-Permian mass extinction at Shangsi, South China:
855 Palaeogeography, Palaeoclimatology, Palaeoecology, v. 448, no. Supplement C, p. 59-71.

856 Xie, S.-C., Pancost, R. D., Huang, X.-Y., Jiao, D., Lu, L.-Q., Huang, J.-H., Yang, F.-Q., and Evershed, R. P., 2007,
857 Molecular and isotopic evidence for episodic environmental change across the Permo/Triassic boundary at
858 Meishan in south China: Global and Planetary Change, v. 55, no. 56-65.

859 Xie, S., Pancost, R. D., Yin, H., Wang, H., and Evershed, R. P., 2005, Two episodes of microbial change coupled
860 with Permo/Triassic faunal mass extinction: Nature, v. 434, p. 494-497.

861

862 **FIGURE CAPTIONS**

863

864 Figure 1. Map showing Svalbard and the location of the Festningen section near the entrance of
865 Isfjorden on the island of Spitsbergen.

866

867 Figure 2. Palaeogeographic maps showing the study area in Late Permian time, including (A)
868 reconstruction of Pangea (after Scotese, 2004) showing likely patterns of Coriolis affect driven
869 ocean circulation, forming eastern boundary currents in the Panthalassa Ocean along the western
870 margins of Pangea. (B) Paleogeographic map showing relative locations of sections from
871 Spitsbergen and the Sverdrup Basin during Early Triassic time, as well as location relative to
872 Pangea (inset box).

873

874 Figure 3. Chemostratigraphic correlation of organic carbon records from Late Permian through
875 to Middle Triassic time of the Festningen section (this study) with that from sections in the
876 Sverdrup Basin, Arctic Canada (Grasby et al., 2016a). LPE=Latest Permian Extinction level,
877 Grie. = Griesbachian.

878

879 Figure 4. Plots of geochemical data for Festningen. (A) $\delta^{13}\text{C}$ of organic carbon, red dots indicate
880 sub-samples selected for $\delta^{15}\text{N}$ measurement. (B) Total organic carbon (TOC). (C) $\delta^{15}\text{N}$ of N in
881 organics, (D) total N in sediments. (E) total P in sediments. (F) ratio of N/P in sediments. (G)
882 plots of element enrichment factors (EF) for Ba, Ni, and Cu. LPE = Latest Permian Extinction
883 level, Tri. = Triassic, Bot. = Botneheia, Fm = Formation.

884

885 Figure 5. Comparison of $\delta^{15}\text{N}$ records from global sections, after Saitoh et al. (2014). P/T =
886 Permian Triassic Boundary, LPE = Latest Permian Extinction. Data are derived from Festningen
887 (this study), Arctic Canada (Knies et al., 2013), Opal Creek (Schoepfer et al., 2013), Guryul
888 Ravine (Algeo et al., 2007), Bulla (Jia et al., 2012), Zuodeng and Taiping (Luo et al., 2011)
889 Meishan (Cao et al., 2009), Chaotian (Saitoh et al., 2014). Temporal control to compare data was
890 provided by stretching individual sections such that the two time points, P/T = Permian Triassic
891 Boundary and LPE = Latest Permian Extinction level are at the same vertical scale.

892

893 Figure 6. Comparison of Early Triassic trends across NW Pangea from sections at Smith Creek
894 (Grasby et al., 2016b) and Festningen (this study). The basal portion of the Festningen section
895 has been excluded as correlative data is not available. Data are shown for $\delta^{15}\text{N}$, $\delta^{13}\text{C}$ of organic
896 carbon, and total organic carbon (TOC). Plots are shown relative to time by stretching data plots

897 of individual sub-stage boundaries. Sedimentary gaps, changes in biodiversity, and ocean
898 temperature are from Chen and Benton (2012), Grasby et al. (2016b), and Sun et al. (2012). Pr. =
899 Permian, Ch = Changhsingian, Gr. = Griesbachian, Di.= Dienerian, Sm. = Smithian, TOC= total
900 organic carbon.

901 Figure 7. Plot of inorganic carbon isotope records through the Early Triassic from the Neotethys
902 Zal section (Horacek et al., 2007b) and Tethys Nanpanjiang section (Sun et al., 2012), as well as
903 the organic carbon Boreal records from Smith Creek (Grasby et al., 2013), and Festningen
904 (Grasby et al. 2016a) showing similarity of global trends in carbon isotope from both inorganic
905 and organic carbon pools. The trends in N isotope values from Smith Creek (Grasby et al. 2016b)
906 and Festningen (this study) are largely decoupled from changes to the global carbon pool. Gries.
907 = Griesbachian, Dien. = Dienerian.

908