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1 **Progressive environmental deterioration in NW Pangea leading**
2 **to the Latest Permian Extinction**

3

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25 **ABSTRACT**

26 Stratigraphic records from northwestern Pangea provide unique insight into global processes
27 that occurred during the Latest Permian Extinction (LPE). We examined a detailed geochemical
28 record of the Festningen Section, Spitsbergen. A stepwise extinction is noted: 1) loss of
29 carbonate shelly macrofauna, 2) loss of siliceous sponges in conjunction with an abrupt change
30 in ichnofabrics as well as dramatic change in the terrestrial environment, and 3) final loss of all
31 trace fossils. We interpret loss of carbonate producers as related to higher-latitude shoaling of
32 the lysocline in relationship to building atmospheric CO₂ in higher latitudes. The loss of siliceous
33 sponges is coincident the global LPE event and is related to onset of high loading rates of toxic
34 metals (Hg, As, Co) that we suggest are derived from Siberian Trap eruptions. The final
35 extinction stage is coincident with redox sensitive trace metal and other proxy data which
36 suggest onset of anoxia, after the other extinction events. These results show a remarkable
37 record of progressive environmental deterioration in NW Pangea during the extinction crises.

38 **1.0 INTRODUCTION**

39 The Latest Permian Extinction (LPE) represents a period of dramatic climate change associated
40 with disruption of global biogeochemical cycles and the worst mass extinction event in Earth
41 history. Over 90% of marine species and 70% of terrestrial vertebrates went extinct at this time
42 (Erwin, 2006). While numerous extinction mechanisms have been proposed, growing evidence
43 supports environmental effects associated with massive eruption of the Siberian Traps
44 (Campbell et al., 1992; Grasby et al., 2011; Renne et al., 1995; Saunders and Reichow, 2009;
45 Shen et al., 2011; Wignall, 2001). The original volume of the Siberian Traps and West Siberian
46 rift system is difficult to estimate, but upper-end figures of 3 - 4 x 10⁶ km³ (Courtillot et al., 1999;
47 Fedorenko et al., 2000) make this mega-scale eruption one of the largest in earth history.

48 Magma intruded through the Tunguska Basin, and was associated with combustion of organic
49 rich sediments (Grasby et al., 2011; Reichow et al., 2009; Retallack and Jahren, 2008; Retallack
50 and Krull, 2006; Svensen et al., 2009), along with release of large volumes of CO₂ (White and
51 Saunders, 2005; Wignall, 2001), deleterious atmospheric gases (Beerling et al., 2007; Black et al.,
52 2012; Black et al., 2014; Kaiho and Koga, 2013; Svensen et al., 2009), and toxic elements (Grasby
53 et al., 2013a; Grasby et al., 2011; Sanei et al., 2012). Oxygen isotope records suggest that rapid
54 global warming and extremely high ocean temperatures developed at this time (Romano et al.,
55 2013; Sun et al., 2012), invoking a hothouse scenario (Kidder and Worsley, 2010; Retallack,
56 1999; Song et al., 2014). Acid ocean conditions may also have developed at this time
57 (Beauchamp and Grasby, 2012; Heydari and Hassanzadeh, 2003; Kidder and Worsley, 2004,
58 2010; Liang, 2002; Payne et al., 2007). Global anoxia has long been suggested to be an important
59 environmental stress associated with the LPE (Isozaki, 1997; Knoll et al., 1996; Wignall and
60 Hallam, 1992; Wignall and Twitchett, 1996). While some regions show evidence of photic zone
61 euxinia in the Tethys and Panthalassa (Grice et al., 2005; Hays et al., 2007; Kump et al., 2005; Xie
62 et al., 2007), the extinction event has also been suggested to occur under at least locally oxic
63 conditions in NW Pangea (Algeo et al., 2010; Knies et al., 2013; Proemse et al., 2013) and in
64 Neotethys (Korte et al., 2004; Loope et al., 2013; Richoz et al., 2010) (Fig. 1a).

65 Given the above, the relative timing of various environmental stresses becomes critical to
66 understanding the role they played during the mass extinction. To address this question we
67 examined the Festningen section in Spitsbergen (Wignall et al., 1998), a shelf sea location on
68 northern Pangean margin during Late Permian time (Figs. 1b,c). The Festningen section was one
69 of the earliest locations where development of anoxia in association with the mass extinction
70 event was demonstrated (Wignall et al., 1998). However, this study was based on a low sample
71 density for carbon isotope data that did not provide clarity as to detailed biogeochemical events

72 occurring during the extinction period. Subsequent work at other sites in Spitsbergen has
73 pointed to the gradual development of anoxia across the LPE event (Dustira et al., 2013), as well
74 as in correlative strata in the Sverdrup Basin (Grasby and Beauchamp, 2009). To elucidate the
75 relative timing of various environmental stressors we have undertaken detailed analyses of the
76 Festningen based on high resolution sampling through the LPE.

77

78 **2.0 STUDY AREA**

79 The Festningen section is located at Kapp Starostin, west of the mouth of Grønfjorden where it
80 enters Isfjorden on Nordenskiöld Land, Spitsbergen (Fig. 1b). In Permian time the area formed
81 part of a broad epicontinental shelf on the northwestern margin of Pangea (Fig. 1c), along with
82 correlative strata from the Wandel Sea (North Greenland), the Sverdrup Basin (Canadian High
83 Arctic), and the Barents Sea and Timan-Pechora Basin (Russia) (Stemmerik and Worsley, 2005).
84 Spitsbergen was at a paleolatitude of $\sim 40\text{-}45^\circ$ N during the Middle to Late Permian (Golonka and
85 Ford, 2000; Scotese, 2004).

86

87 **2.1 Festningen Section**

88 The Festningen section occurs as $\sim 45^\circ$ eastward dipping beds (Fig. 2) forming a ~ 7 km coastal
89 section exposed in a low sea-cliff, including near continuous exposure of Carboniferous to
90 Cenozoic strata, from Kapp Starostin to Festningsdodden. The section is located in the eastern
91 part of the West Spitsbergen Fold and Thrust Belt, an intra-continental fold and thrust belt
92 ranging over more than 300 km along the west coast from the Brøgger Peninsula in the North to
93 the Sørkapp in the very South (CASE-Team, 2001; Dallmann et al., 1993; Maher and Craddock,
94 1988). The intense crustal shortening is a result of the northward directed movement of

95 Greenland against the Barents shelf during the Eocene, before Spitsbergen was finally separated
96 from Greenland. The Festningen section is part of the steeply inclined short-limb of a kilometer-
97 scale east-vergent fold structure. A sill cuts through the series (dating from the Cretaceous 124.7
98 Ma) (Corfu et al., 2013). Festningen was located in the central Spitsbergen region where Upper
99 Permian sediments, deposited in a distal shelf setting, are thickest (Wignall et al., 1998)
100 (Blomeier et al., in press). Festningen represents the type-section for both the Kapp Starostin
101 and Vardebukta formations which are examined here.

102 The Kapp Starostin Formation is a Middle to Late Permian unit that was deposited at a time
103 of tectonic quiescence and passive subsidence following a major relative sea level drop
104 coinciding with the Lower/Middle Permian boundary (Blomeier et al., 2013). An initial Roadian
105 transgression led to the deposition of a widespread heterozoan carbonate (Vøringen Member),
106 which was followed by a series of regressions and transgressions that led to the progradation of
107 heterozoan carbonates and cherts over much of the Barents Shelf and Svalbard (Blomeier et al.,
108 2013), as well as in the paleogeographically adjoining Sverdrup Basin (van Hauen, Degerbøls and
109 Troid Fiord formations; Beauchamp et al., 2009). The uppermost fossiliferous carbonate unit in
110 the Kapp Starostin Formation occurs ~40 m below the contact with the overlying uppermost
111 Permian-Lower Triassic Vardebukta Formation. The topmost part of the Kapp Starostin
112 Formation is dominated by spiculitic chert, an interval that is in part Late Permian in age
113 (Blomeier et al., 2013) and considered equivalent to the Black Stripe and Lindström formations
114 of the Sverdrup Basin (Beauchamp et al., 2009).

115 The Vardebukta Formation is a unit of shale, siltstone and minor sandstone that is devoid of
116 carbonate and chert. The formation is mostly Early Triassic (Griesbachian–Dienerian) in age as
117 shown by ammonoid and conodont fauna (Mørk et al., 1982; Nakrem et al., 2008; Tozer and

118 Parker, 1968). While the contact between the Kapp Starostin and Vardebukta formations was
119 for many years considered the Permian-Triassic Boundary (PTB) (e.g.(Mørk, 1982
120 #662;Mangerud, 1993 #555), it is now widely accepted that the basal beds of the Vardebukta
121 Formation are latest Permian (Changhsingian) in age. While *Hindeodus parvus* – the globally
122 recognized fossil for the base Triassic as documented at the PTB GSSP at Meishan, China (Yin et
123 al., 2001) – has yet to be recovered in the basal Vardebukta Formation at Festningen,
124 chemostratigraphic considerations have led Wignall et al. (1998) to place the P-T boundary ~6 m
125 above the Kapp Starostin-Vardebukta contact based on the stratigraphic position of the globally-
126 recognized $d^{13}C$ minimum, a practice since followed by others (e.g. Dustira et al., 2013).

127 Wignall et al. (1998) show a stepwise loss of fauna at Festningen as summarized here. The
128 majority of carbonate secreting taxa was lost ~12 m below the top of the Kapp Starostin
129 Formation. The brachiopod fauna present in the upper-most beds may represent an early to late
130 Lopingian age (Nakamura et al., 1987). After loss of carbonate fauna, siliceous sponges were the
131 only taxa that remained abundant to the top of the formation. However, ichnofabrics indicating
132 the persistent presence of soft-bodied fauna are also abundant. The top of the Kapp Starostin
133 Formation coincides with the loss of siliceous sponges and an abrupt change in ichnofabrics,
134 marked by disappearance of *Zoophycus* and *Chondrites*. The basal 5 m of the Vardebukta
135 Formation is characterised by *Planolites* and pyritized small burrows, above which sediments
136 become finely laminated and lacking in trace fossils.

137 Change is also observed in the terrestrial environment as indicated by palynological
138 assemblages at Festningen (Mangerud and Konieczny, 1993). The upper-most Kapp Starostin
139 Formation is dominated by a variety of pollen and spores from gymnosperms (conifers,
140 pteridosperms, and rare cordaites) and pteridophytes. Basal rocks of the overlying Vardebukta

141 Formation contain an overall lower diversity of palynomorphs than observed in the Kapp
142 Starostin Formation with the exception that spores of lycopodiophytes and bryophytes are
143 present with greater diversity than observed in underlying strata. Pollen of gymnosperms are
144 represented by *Lunatisporites* spp., a spore with pteridosperm affinity, and the first appearance
145 of *Tympanicysta stoschiana* occurs in the basal Vardebukta Formation. Acritarchs (*Veryhachium*
146 spp. and *Micrhystridium* spp.) then recover and become abundant in the Vardebukta Formation
147 (Mangerud and Konieczny, 1993).

148 Acritarchs may have constituted the pioneering taxa of the planktonic oceanic realm
149 following marine perturbation associated with the latest Permian event. Oceanic conditions may
150 have been favourable for the development of widespread acritarch and prasinophyte blooms
151 due to stratified ocean waters and elevated atmospheric carbon dioxide concentrations
152 associated with volcanic activity and/or extreme oligotrophy in the mixed layer due to slow
153 oceanic circulation (Martin, 1996; Payne and van de Schootbrugge, 2007).

154

155 **3.0 METHODS**

156 **3.1 Sample Collection**

157 Sampling was conducted at 50 cm spacing, from 20 to 4 m below the top of the Kapp Starostin
158 Formation, and then across the LPE interval sample spacing was reduced to 20 cm, from 4 m
159 below to 18 m above the formation contact. Sample spacing is reported in metres above
160 (positive) and below (negative) the last chert bed that defines the top of the Kapp Starostin
161 Formation. A total of 93 samples are included in this study.

162 In the field, weathered surfaces were removed and samples were collected from a narrow
163 defined zone no greater than 2 cm thick. In the laboratory any remaining weathered surfaces
164 were removed and fresh samples were powdered using an agate mortar and pestle and split
165 into sub-samples for subsequent analyses.

166

167 **3.2 Geochemistry**

168 Total organic carbon (TOC) was measured using Rock-Eval 6[®], with $\pm 5\%$ analytical error of
169 reported value, based on repeats and reproducibility of standards run after every 5th sample
170 (Lafargue et al., 1998). Total sulphur (TS) and total carbon (TC) was measured on a LECO 444
171 analyzer, with the average of 3 repeat measurements reported, with $\pm 2\%$ analytical error. Total
172 inorganic carbon (TIC) was calculated from (TIC=TC-TOC). Elemental determinations were
173 conducted on powdered samples digested in a 2:2:1:1 acid solution of H₂O-HF-HClO₄-HNO₃, and
174 subsequently analyzed using a PerkinElmer Elan 9000 mass spectrometer, with $\pm 2\%$ analytical
175 error. Hg was measured at GSC-Atlantic by LECO[®] AMA254 mercury analyzer (Hall and Pelchat,
176 1997) ($\pm 10\%$).

177

178 **3.3 Stable Isotope analyses**

179 Stable isotope measurements were conducted at the Isotope Science Laboratory, University of
180 Calgary. For determination of $\delta^{13}\text{C}_{\text{org}}$, samples were washed with hydrochloric acid, and rinsed
181 with hot distilled water to remove any carbonate before determination of $\delta^{13}\text{C}$ of organic
182 carbon. $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ were measured using Continuous Flow-Elemental Analysis-Isotope Ratio
183 Mass Spectrometry, with a Finnigan Mat Delta+XL mass spectrometer interfaced with a Costech

184 4010 elemental analyzer. Standards were run every 5th sample. Combined analytical and
185 sampling error for $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{15}\text{N}_{\text{org}}$ is $\pm 0.2\text{‰}$.

186

187 **3.4 Absolute age dating**

188 Zircons were separated from bentonite layers using conventional heavy liquid and magnetic
189 techniques at Curtin University, Perth (Australia). Zircon grains were handpicked under a
190 binocular microscope. Together with standards BR266 (Stern, 2001) and OGC-1 (Stern et al.,
191 2009) and NIST NBS610 glass, these zircons were mounted in 25 mm diameter epoxy disc and
192 then polished and coated with gold.

193 Zircons were imaged using Cathodoluminescence (CL) techniques on a Zeiss 1555 VP-
194 FESEM in the Centre for Microscopy, Characterisation and Analysis of the University of Western
195 Australia. Zircon analyses were performed on the SHRIMP II at the John de Laeter Centre for
196 Isotope Research, Curtin University, and followed standard operation procedures (Compston et
197 al., 1984; Williams, 1998). The primary (O_2^-) ion beam was 0.7 nA on a 15 μm spot. The data
198 were processed using the SQUID and ISOPLOT program (Ludwig, 2003; Ludwig, 2009). Common
199 Pb was subtracted from the measured compositions using the measured ^{204}Pb and a common Pb
200 composition from the model of (Stacey and Kramers, 1975) at the appropriate stage of each
201 analysis.

202

203 **4.0 RESULTS**

204 **4.1 Absolute Age Dating**

205 Two previously unreported bentonite layers ~2 cm thick were found in the basal Vardebukta
206 Formation, 2.6 and 13 m above the top of the Kapp Starostin Formation (hereafter referred to as
207 ash layer). The layers were isolated and collected in the field. Zircons were only recovered from
208 the lower layer at 2.6 m above the formation boundary. The zircon grains are inclusion-free
209 bipyramidal prisms that are sometimes slightly rounded. These grains range in length from 60
210 μm to 100 μm , and are light brown and occasionally light pink. The CL imaging shows uniform
211 zircons with typical oscillatory zoning and composite zircons with cores overgrown by thin rims
212 (Fig. 3a).

213 Twenty one analyses were performed on thirteen zircons. Age data are presented in Table
214 1 with 1σ precisions. Six of the 21 analyses were rejected due to the high common lead, and
215 sixteen analyses yielded concordant or nearly concordant ages ranging from 244 Ma to 2685 Ma
216 (Fig. 3b). Ten concordant or nearly concordant analyses plot in one single population with a
217 weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 252 ± 3 Ma (MSWD = 0.92) (Fig. 3b). There are also six older
218 ages: two Late Silurian-Early Devonian (412 ± 8 , 428 ± 8 Ma), and four Neoproterozoic (2645 ± 6 ,
219 2663 ± 17 , 2642 ± 12 and 2685 ± 4 Ma).

220

221 **4.2 Carbon isotope records**

222 Given the lack of carbonates, the organic carbon isotope record was examined at the Festningen
223 section. At 15 m above the Kapp Starostin Formation contact, shales show visible signs of
224 thermal alteration from an overlying Cretaceous sill which starts at ~19 m. Thermal effects can
225 also be observed in the geochemical record close to the sill itself, although there is no apparent
226 impact on the key part of the section in the basal 15 m of the Vardebukta Formation (Fig. 4).
227 Rock-Eval 6[®] results in the basal 15 m provide an average Tmax value of 453 °C, indicating that

228 organic matter in the shales are not thermally affected by the overlying sill, and that away from
229 the localised thermal affects, the Festningen section has never been heated past the upper end
230 of the oil window (note Tmax values reflect relative, not actual, burial temperatures). At the
231 equivalent maximum burial temperatures, the stable isotope values of organic carbon are not
232 altered (Hayes et al., 1983). The Oxygen Index derived from RockEval analyses has an average
233 value of 28, consistent with well-preserved organic matter.

234 The $\delta^{13}\text{C}$ record of the organic carbon shows two initial minor negative shifts of 1 to 2‰
235 at -12 m (where calcareous shelly macrofauna are lost), and then again at 3 m below the top of
236 the Kapp Starostin Formation (Fig. 4a, Table 2). There is a brief positive shift in $\delta^{13}\text{C}$ -values just
237 below the Kapp Starostin/Vardebukta contact. The top of the Kapp Starostin Formation is
238 marked by onset of a progressive ~8‰ negative shift in $\delta^{13}\text{C}$, over the basal 5 m of the
239 Vardebukta Formation, to a low of -33‰. This $\delta^{13}\text{C}$ low is coincident with the level where
240 bioturbation disappears (Wignall, 1998). The $\delta^{13}\text{C}$ values then remain relatively stable for the
241 next 10 m, after which there are thermal affects due to the overlying sill (Fig. 4).

242 Overall organic carbon content is low, with values less than 0.8% TOC throughout the
243 studied interval (Fig. 4b). In the interval from 20 to 12 m below the top of the Kapp Starostin
244 Formation, TOC values vary around 0.5 to 0.6%. TOC values then drop at the level associated
245 with loss of calcareous macrofauna (~12 m below the top of the Kapp Starostin Formation) to
246 values around 0.4%. The TOC remains at these values for the remaining 12 m of the Kapp
247 Starostin Formation. Above the Kapp Starostin Formation, there is a progressive drop in TOC
248 associated with the drop in $\delta^{13}\text{C}_{\text{org}}$ values until the first ash layer at 2.6 m, where there is an
249 abrupt increase in TOC to values of 0.5 % above this level. TOC then fluctuates for the rest of the
250 section, with peak values of 1.03% at 8.2 m above the formation contact.

251 The TIC record is also plotted in Figure 4c. Even in the basal part of the section, where
252 carbonate fossils are observed, TIC is still low (~0.5 %). Above the loss of shelly macrofauna TIC
253 values drop to <0.1% for the remainder of the Kapp Starostin Formation. Above the formation
254 contact, there is a progressive increase in TIC over the zone where $\delta^{13}\text{C}_{\text{org}}$ values drop, to values
255 of ~1%. The TIC values then remain relatively constant until the zone of thermal influence where
256 they drop again. The one exception is peak values >2.5% around 8 m above the formation
257 contact, coincident with a zone of peak TOC values (Fig. 4).

258

259 **4.3 Redox Proxies**

260 Several trace elements have been shown to be useful proxies for marine redox state (Mo, U, V)
261 in addition to pyrite associated Fe (Fepy) (Scott and Lyons, 2012; Tribovillard et al., 2006). The
262 variation of these proxy elements are plotted in Figure 5. Fepy values are consistently low (< 0.5
263 %) in the upper Kapp Starostin Formation, through the zone of the last carbonate producers,
264 and across the formation boundary as marked by the loss of sponges (Fig. 5a). Pyrite is rare until
265 above the first ash layer at 2.6 m above the formation contact, after which there is a significant
266 increase in Fepy, to values above 1%. This increase in pyrite is seen also in a plot of TS versus
267 TOC (Fig. 6a). Here samples below the ash bed from both the Kapp Starostin and Vardebukta
268 formations plot close to the oxic/sub oxic boundary as defined for ancient sediments (Raiswell
269 and Berner, 1985). Samples above the ash layer show significantly higher Fepy values, with peak
270 levels in the finely laminated black shale at ~ 8 m above the Formation contact (Fig. 5a). In
271 general, Fepy values show inverse trends to $\delta^{13}\text{C}_{\text{org}}$ (Fig. 6b) across the boundary.

272 The Mo concentrations in the Kapp Starostin and the basal Vardebukta formations (solid
273 circles in Fig. 5b) are consistently lower than average marine shale values, relative to Post-

274 Archean average Australian shale (PAAS) (Taylor and McLennan, 1985). However, Mo values
275 increase to > 20 ppm within a narrow interval of the laminated black shale at ~ 8 m above the
276 Kapp Starostin Formation (Figs. 2b,c; 5b). A similar trend is observed in U and V data (solid
277 circles in Figs. 5c,d). Through the upper Kapp Starostin Formation and basal Vardebukta
278 Formation, U and V concentrations are consistently below average marine values (except peaks
279 associated with the first ash layer). While there is an initial minor increase at the Kapp Starostin/
280 Vardebukta contact, values do not consistently exceed average marine shale until after the level
281 where burrowing organisms are lost (~5 m above the formation contact). The U and V
282 concentrations peak in the same laminated black shale ~8 m above the formation contact,
283 where Mo and TOC values are also highest.

284 Given the change from chert to shale at the formation boundary, redox sensitive
285 elements were also normalised to Al to account for potential lithologic affects (dashed lines in
286 Fig. 5). As with absolute values, Al normalised values show a decline (Mo, U) in the lower
287 Vardebukta Formation, or remain at low values (V) relative to the underlying Kapp Starostin
288 Formation chert. The only significant increase in the metal/Al ratio is associated with the
289 laminated black shale interval at 8 m above the formation contact.

290

291 **4.4 Trace Metals**

292 The variability of trace metals (Cu, Pb, As, Co, Ni, Hg) is illustrated in Figure 7. Absolute
293 concentrations are plotted (dots) along with values normalized to Al (lines) to account for
294 potential lithologic changes. Mercury deposition over geologic time is strongly controlled by
295 organic matter (Grasby et al., 2013b), and therefore anomalous Hg deposition is best observed
296 by normalizing relative to TOC (Sanei et al., 2012). Similarly, Ni is strongly scavenged by organic

297 matter (Tribovillard et al., 2006) and is also normalized relative to TOC. Trace metal values in the
298 lowest chert dominated portion of the section are relatively constant and below PAAS values
299 (vertical dotted lines in Fig. 7). At 12 m below the formation contact, where the section is
300 marked by the final disappearance of calcareous macrofauna, there is a small but distinct shift
301 to lower values of all metals as it transitions from shaly chert to pure chert. However, there is no
302 notable shift in normalized values at this level. The metals (concentrations and ratios) remain at
303 low levels through to the top of the Kapp Starostin Formation. In the basal 5 m of the
304 Vardebukta Formation there is a significant increase in all trace metals, to values well above
305 PAAS. This increase is also observed in the Al normalized values. The Ni/TOC and Hg/TOC ratios
306 also show a significant spike at in the basal Vardebukta Formation. After this trace metal spike,
307 there is a gradual decline to values near or below PAAS. One exception is a brief increase
308 associate with a zone defined by high Mo values ~ 8 m above the formation contact, where
309 there is also a peak in TOC values (Fig. 7). In general, trace metals in the basal Vardebukta
310 Formation show much greater variability than in the upper portion of the Kapp Starostin
311 Formation.

312

313 **4.5 Nitrogen Isotopes**

314 The $\delta^{15}\text{N}$ of total nitrogen has been used to assess changes in nutrient cycles across the LPE
315 (Knies et al., 2013; Schoepfer et al., 2013). Major changes in nutrient cycling, through shifts in
316 denitrification and/or atmospheric nitrogen fixation can strongly influence the $\delta^{15}\text{N}$ signal of the
317 marine nitrate pool. For the levels of thermal maturity found in the Festningen section, there
318 are negligible effects on the stable isotope values of N (Ader et al., 1998; Bebout and Fogel,
319 1992). The $\delta^{15}\text{N}$ values of total N are illustrated in (Fig. 4d). Results show that through the Kapp

320 Starostin that $\delta^{15}\text{N}$ values are consistently around 7‰. There is then a progressive decline in
321 $\delta^{15}\text{N}$ values through the basal 5 m of the Vardebukta Formation until the level at which
322 bioturbation was lost. Above the level where bioturbation is lost, values remain consistently
323 around 5‰.

324

325 **4.6 Chemical Index of Alteration**

326 The Chemical Index of Alteration (Sydeman et al.) (Nesbitt and Young, 1982) provides a proxy for
327 the degree of chemical weathering as recorded in siliciclastic sedimentary rocks whereby
328 increased chemical weathering mobilizes Na, K, and Ca during the transformation of feldspar
329 minerals to clays. However, the CIA index needs to be corrected for potential Ca from inorganic
330 carbon (Fedo et al., 1995). For Festningen, the CIA index shows almost no variation through the
331 section analyzed, with values consistently near 80 (Fig. 4e). Towards the top of the section the
332 CIA values drop within the zone of thermal influence.

333

334 **5.0 DISCUSSION**

335

336 **5.1 Age dating**

337 The bentonite layer 2.6 m above the top of the Kapp Starostin Formation had one dominant
338 population with a weighted mean age of 252 ± 3 Ma, which we interpreted as the crystallisation
339 age of the volcanic ash. Six older ages were obtained in zircons which are slightly rounded and
340 likely represent xenocrysts (Fig. 3b). The source of these older ages may be: 1) Silurian to Early
341 Devonian granites, located in the Nordaustlandet Terrane of northeast Svalbard whose ages
342 range from 410 to 440 Ma (Johansson et al., 2002); and 2) a Neoproterozoic quartz-monzonite,

343 located in the Ny-Friesland, northern Svalbard, which yielded an upper intercept of 2709 ± 28
344 Ma, considered the best estimate of the crystallisation age (Hellman et al., 2001).

345

346 **5.2 Carbon records and the LPE Boundary**

347 While not having sufficient precision to be definitive, the 252 Ma age of the ash layer suggests
348 that the top of the Kapp Starostin Formation represents the global LPE boundary. This is further
349 constrained by carbon isotope data. The organic carbon isotope record at Festningen shows a
350 distinct negative $\delta^{13}\text{C}$ excursion initiated at the basal most Vardebukta Formation (Fig. 4),
351 consistent with negative excursions associated with the LPE horizon observed in inorganic
352 carbon isotope records (Korte and Kozur, 2010). This negative carbon isotope shift is also
353 consistent with organic carbon isotope records from other boreal settings (e.g. Sverdrup Basin -
354 Grasby and Beauchamp (2008;2009)); NE British Columbia - (Wang et al., 1994; Wignall and
355 Newton, 2003), East Greenland (Twitchett et al., 2001); and Norway/Spitsbergen (Dustira et al.,
356 2013; Hermann et al., 2010). In their review, (Korte and Kozur, 2010) showed that the initial
357 negative decline in $\delta^{13}\text{C}$ values started at the LPE and reached a minimum $\delta^{13}\text{C}$ value after the
358 extinction event. (Shen et al., 2011) also show that the negative excursion in the carbon isotope
359 record occurs after the main extinction event in the Tethys. As well, at Meishan (Burgess et al.,
360 2014) show that after the initial negative peak, the broad decline in $\delta^{13}\text{C}$ values occurs after the
361 main extinction event. Therefore, we interpret the negative shift at Festningen as being
362 consistent with the global pattern for the negative carbon isotope excursion initiating at the LPE
363 event. Based on this interpretation, we follow previous workers who used the $\delta^{13}\text{C}$ minimum as
364 the approximate P/T boundary in Spitsbergen (Dustira et al., 2013; Wignall et al., 1998), and the
365 onset of the major $\delta^{13}\text{C}$ decline above the last chert beds to mark the LPE horizon. This makes

366 the LPE horizon coincident with the top of the Kapp Starostin Formation, that also marks the
367 loss of sponges as well as collapse of well-developed ichnofauna, including *Zoophycus* and
368 *Nereites* (Wignall et al., 1998).

369 The tops of the Kapp Starostin Formation also marks a shift in palynological assemblages,
370 from those dominated by gymnosperms to an assemblage dominated by lycopsids, as described
371 at Festningen and other sections on Spitsbergen (Mangerud and Konieczny, 1993). This lycopsid
372 “spore peak” in latest Permian strata is well documented elsewhere in the northern hemisphere
373 (Hochuli et al., 2010; Twitchett et al., 2001). These changes in palynoassemblages across the
374 boundary represent major vegetation community collapse of Late Permian gymnosperm-
375 dominated ecosystems followed by re-colonization by pioneering lycopsids and bryophytes and
376 components of typical Early Triassic shrubland communities (Hochuli et al., 2010; Twitchett et
377 al., 2001), representing a terrestrial response to environmental stress followed by rapid, but
378 short lived, recovery. (Twitchett et al., 2001) noted that the synchronous collapse of the marine
379 and terrestrial ecosystem preceded a sharp negative carbon isotope excursion at the LPE
380 boundary in East Greenland. As well, (Hermann et al., 2010) showed that in the Trøndelag and
381 Finnmark Platform, Norway, the marine extinction level was immediately followed by the
382 increase in spore abundance and a sudden drop of C-isotope values. Thus, the Latest Permian
383 terrestrial collapse observed across NW Pangea is coincident with the marine extinction marked
384 by the loss of chert forming siliceous sponges.

385 The loss of chert was a global feature at the LPE (Beauchamp and Baud, 2002; Beauchamp
386 and Grasby, 2012) that has been correlated with the extinction event in Meishan (Wignall and
387 Newton, 2003) and onset of the Early Triassic Chert Gap. Previous workers also place the LPE
388 boundary at the top of the last chert beds in correlative strata from the Sverdrup Basin (Embry

389 and Beauchamp, 2008; Grasby and Beauchamp, 2008; Proemse et al., 2013) and western
390 Canada (Schoepfer et al., 2013). However, this placement of the LPE boundary contrasts with
391 the claims of (Algeo et al., 2012) who speculated that that the loss of sponges and complex
392 ichnofabric represents an earlier extinction than the LPE event itself (their “arctic event”). The
393 level that they assign as the LPE horizon at the West Blind Fiord section of the Sverdrup Basin is
394 marked by minor geochemical changes in the overlying shales (Fig. 8). These are more
395 consistent with those observed in Festningen at the level of the first ash bed. In fact the samples
396 that (Algeo et al., 2012) analyzed from this level at WBF (collected by two of us, S.G. and B.B.)
397 were in fact ash layers and thus the chemistry is not representative of marine conditions as they
398 assumed.

399

400 **5.3 Redox Proxies**

401 Multiple proxies for anoxia examined as part of this work, including redox sensitive trace
402 elements, Fepy, and TC/TC, show similar trends. In the chert-dominated upper Kapp Starostin
403 Formation redox sensitive elements are consistently lower than average PAAS values, and
404 TS/TOC values plot along the oxic boundary for oxic/suboxic waters. Above the Kapp Starostin
405 there is a slight shift to higher concentrations of redox sensitive elements, however they remain
406 below PAAS values, suggesting a largely oxic system in the basal 2.6 m of the Vardebukta
407 Formation. Such an oxic environment is consistent with Fepy that remain low through the Kapp
408 Starostin and basal Vardebukta formations. These data suggest then that the LPE boundary,
409 marked by the loss of siliceous sponges, occurs under oxic to dysoxic conditions at Festningen.

410 Above the first ash layer at 2.6 m TS values increase and Fepy values plot in the sub-oxic
411 zone of Raiswell and Berner (1985). The peak values of redox sensitive elements, exceeding

412 PAAS, as well as peaks in Al normalized values occurs at ~ 8 m in association with the black
413 laminated shale above the zone where burrowers are lost. These increased concentrations of
414 redox sensitive elements, both absolute and normalized to Al are strong indicators of marine
415 anoxia (Tribovillard et al., 2006), suggesting that conditions at Festningen transitioned to a more
416 anoxic environment after the LPE extinction boundary. This is supported by the progressive shift
417 to lower $\delta^{15}\text{N}$ values that suggests increased fixation of atmospheric N_2 , possibly in response to
418 increasing anoxia (Schoepfer et al., 2013; Proemse et al., 2013; Knies et al., 2013). This
419 interpretation of anoxia is consistent with original work by Wignall et al. (1998) who suggested
420 onset of anoxia at this level, in addition to recent work by (Bond and Wignall, 2010) who showed
421 pyrite framboid data at Festningen consistent with transition to anoxic conditions at the same
422 level.

423

424 **5.4 Trace metals**

425 A key aspect of the Festningen section is the significant increase in metals at the LPE boundary
426 that occurs at a level where anoxia has not yet developed. In fact, metal concentrations right
427 above the LPE are greater than when anoxic conditions eventually develop higher in the section.
428 These high metal concentrations argue against these anomalous metal loads being associated
429 with increased drawdown into sediment. Previously it has been suggested that metal
430 enrichments at the LPE boundary could be related to high loading rates from the Siberian Trap
431 eruptions (Grasby et al., 2011, Sanei et al., 2012). Similarly, we interpret the anomalous metal
432 concentrations at the LPE boundary, both absolute and Al normalized, to be related to enhanced
433 metal flux from the Siberian Traps. While described here for Festningen, similar trace metal
434 spikes have been observed in the Sverdrup Basin (Grasby et al., 2011) as well as at Meishan,
435 where Ni concentrations show a significant increase just prior to the carbon isotope shift (Kaiho

436 et al., 2001; Rothman et al., 2014), implying that increased metal loading at the LPE is a global
437 phenomenon.

438

439 **6.0 Progressive environmental deterioration**

440 Results from our Festningen study demonstrate evidence for progressive environmental
441 deterioration leading up to and across the LPE event. This can be characterised by three main
442 events: 1) lysocline shoaling driving loss of carbonate producers, 2) volcanic metal loading
443 related to volcanics, and 3) onset of anoxia.

444

445 **6.1 Loss of Carbonate Producers**

446 The first notable event in the Festningen section is the loss of carbonate producers (i.e.
447 brachiopods, bivalves, corals, bryozoans, foraminifers) around 12 m below the top of the Kapp
448 Starostin Formation (Wignall et al., 1998); marking the last appearance of any carbonate
449 secreting organisms prior to the LPE event. Not only are carbonate fossils absent after this point,
450 but TIC values drop to near zero (Fig. 4), indicating a complete absence of carbonate sediment.
451 The loss of carbonate producers is also marked by a small negative shift in $\delta^{13}\text{C}$ and drop in TOC
452 (Fig. 4).

453 Early work had interpreted the loss of carbonate producers as being driven by a shift to
454 cooler water temperatures (Beauchamp and Baud, 2002; Reid et al., 2007; Stemmerik and
455 Worsley, 1995). However, reduced ocean temperatures are insufficient to account for loss of
456 carbonate production in clastic-starved, well-lit, aerobic environments (Beauchamp and Grasby,
457 2012). As well, temperatures in the Boreal Realm were already increasing during latest Permian

458 time (Beauchamp and Grasby, 2012) when silica producers became the dominant sediment
459 producer. Instead, the transition from carbonate to silica factories most likely relates to lysocline
460 shoaling driven by increasing atmospheric CO₂ (Beauchamp and Grasby, 2012). Carbon cycle
461 modelling suggests progressive increase in atmospheric CO₂ through the Late Permian (Berner,
462 2006) with values as high as 4000 ppm prior to the LPE (Cui and Kump, 2014). Given the inverse
463 solubility of CaCO₃ with temperature, high latitudes would be most susceptible to increasing
464 atmospheric CO₂ levels, becoming understaturated with respect to carbonates, while lower
465 latitudes maintained shallow water carbonate factories.

466

467 **6.2 Metal loading**

468 The eruption of the Siberian Traps, which roughly coincides with the LPE extinction (Burgess et
469 al., 2014), could have had both positive and negative impact on global ecosystems through
470 release of both nutrients and toxic metals (Frogner Kockum et al., 2006; Hoffmann et al., 2012;
471 Jones and Gislason, 2008). Metal loading from volcanic eruptions can serve as a significant input
472 of limiting nutrients (e.g. Fe, Ni: (Boyd et al., 2000; Konhauser et al., 2009; Langmann et al.,
473 2010), increasing primary productivity, that may relate to microbial blooms which occur at the
474 LPE (Lehrmann, 1999; Xie et al., 2010; Xie et al., 2005). At the same time, high rates of metal
475 loading could exert a toxic shock to both the marine and terrestrial system. While increased acid
476 rain related to the Siberian Trap eruptions has been argued to have significant impact on the
477 terrestrial environment (Black et al., 2014), metal loading would also be deleterious as it
478 dramatically decreases photosynthetic efficiency in vascular plants (Odasz-Albrigtsen et al.,
479 2000). Although there is pollen evidence for significant impact to the terrestrial system, CIA does
480 not change across the boundary, indicating no significant changes in chemical weathering rates
481 as suggested for lower latitudes (Sephton et al., 2005; Sheldon, 2006). This is consistent with

482 (Hochuli et al., 2010) who show a rapid recovery of plant ecosystems from records in the
483 southern Barents Sea, and suggests that in the Boreal realm terrestrial impact was relatively
484 short term at this level.

485 Volcanic eruptions are associated with release of metals to the atmosphere (Vie le Sage,
486 1983), that can form significant components of global element cycles (e.g. volcanoes account for
487 40% of the modern natural component of the global Hg budget (Pyle and Mather, 2003). Volatile
488 metals released from the magma (e.g Cu, Zn, Ni, Pb, Cd, Hg, As) can form stable compounds (e.g.
489 $CdCl_g$, CdS_g , (Symonds et al., 1987)) that condense onto ash particles, creating notable metal
490 enrichments in ash relative to the source magma (Bagnato et al., 2013). Leaching experiments of
491 ash fall show significant subsequent release of these metals into water (Olsson et al., 2013;
492 Ruggieri et al., 2011). Whether the resultant dissolved concentrations can be significant enough
493 to create toxicity, or in some cases nutrient influx (e.g. Fe), would be a function of the ash
494 loading rate (Olsson et al., 2013). In any case, ash loading would represent an anomalous metal
495 load to a system that can be used as a proxy for enhanced volcanic activity in the geologic
496 record (Grasby et al., 2011; Grasby et al., 2013b; Sanei et al., 2012; Sial et al., 2013; Silva et al.,
497 2013).

498 The Siberian Traps also intruded through the Tunguska sedimentary basin, and it has been
499 suggested that this induced combustion of coal and organic rich shales, causing release of over 3
500 trillion tons of carbon (Grasby et al., 2011; Korte et al., 2010; Ogden and Sleep, 2012; Reichow et
501 al., 2009; Retallack and Jahren, 2008; Saunders and Reichow, 2009; Svensen et al., 2009). As
502 with volcanoes, volatile metals released during combustion (e.g., Be, Zn, As, Cd, Tl, Pb, and U)
503 condense and concentrate onto the resulting fly ash that is composed dominantly of SiO_2 , Al_2O_3
504 and Fe_2O_3 particles (Gieré et al., 2003). Enrichment factor of metals, relative to the source

505 organics, can range from 30x up to 100x (Gieré et al., 2003; Klein et al., 1975; Papastefanou,
506 2010). Similar concentration of metals onto fly ash has been observed during combustion of oil
507 shales (Blinova et al., 2012). Metal enrichment is much greater in the smaller size fraction, as
508 they have the largest surface area for condensation of volatiles per unit mass (Davison et al.,
509 1974; Furuya et al., 1987; Kaakinen et al., 1975b; Martinez-Tarazona and Spears, 1996; Smith et
510 al., 1979). The smallest size fraction also has the longest atmospheric residence times, and
511 consequently the greatest spatial distribution during atmospheric transport (Kaakinen et al.,
512 1975a; Smith et al., 1979). Similar to volcanic ash, metals condensed onto the surface of fly ash
513 particles are also released when ash is submerged in water (Bednar et al., 2010). Evidence for
514 coal ash loading and metal release at the LPE was observed in the Sverdrup Basin by Grasby et
515 al. (2011), suggesting that coal ash dispersal was widespread in the northern hemisphere during
516 the latest Permian.

517 The largest volcanic eruption in Earth history, the Siberian Traps, combined with
518 combustion of organics in the Tunguska Basin, would have had an extremely high metal loading
519 rate that far exceeds normal background. As an example (Sanei et al., 2012) calculated a Hg
520 loading rates from the Siberian Traps that would be 4x above modern anthropogenic emissions,
521 assuming a 500 ky eruption period. Similar estimates for other metal fluxes can be made based
522 on the metal/S ratio for modern volcanic emissions (Nriagu, 1989), and estimates of total SO₂
523 release of 3.8×10^{13} Mg from the Siberia Trap eruptions (Beerling et al., 2007). Averaged over an
524 assumed maximum 500 ky eruption history gives a conservative minimum increase. Based on
525 this, Siberian Trap eruptions may have increased global metal flux to the atmosphere from 9%
526 (Se) to 78% (Co) above modern natural background flux (Mather et al., 2013; Nriagu, 1989)
527 (Table 3). However, Siberian Trap magmatism was more likely episodic over the total eruption
528 interval (Pavlov et al., 2011). Such episodic eruption would mean that rather than an overall

529 average background increase, the extinction interval would be better characterized by pulses of
530 extreme metal loading, significantly higher than those estimated here. Pavlov et al. (2011)
531 estimated that the total eruption intervals may represent as low as 8% of the total eruption
532 history (suggesting a net ~40 ky for metal release). Based on this, metal flux by the Siberian
533 Traps may have ranged from 107% (Se) to 977% (Co) above background.

534 While estimates of metal loading rates related to the Siberian Trap contain large
535 uncertainties, it is interesting to note that even conservative estimates are of the same order of
536 magnitude as modern anthropogenic metal release (Pacyna and Pacyna, 2001) that are subject
537 of global concern. Whereas, higher rates based on a more likely pulsed eruption history are one
538 to two orders of magnitude greater than modern anthropogenic emissions. Such extreme
539 loading rates may readily explain the metal anomalies at the LPE boundary, and likely
540 represented a toxic shock to both marine and terrestrial ecosystems.

541

542 **6.3 Anoxia**

543 Our study suggests that the main LPE horizon at Festningen occurs under oxic to dysoxic
544 conditions, but that anoxia developed soon after and is associated with a final extinction of
545 benthic life. There have been suggestions that the initial extinction event occurred under at
546 least local, and perhaps regional, oxic conditions in other NW Pangean (Algeo et al., 2010; Knies
547 et al., 2013; Proemse et al., 2013) and Neotethyan locations (Korte et al., 2004; Loope et al.,
548 2013; Richoz et al., 2010). However, such conditions are often only encountered in shallower
549 proximal settings. In the somewhat more distal setting of Tschermakfjellet, 60 km to the
550 northwest of Festningen, the redox record indicates the gradual onset of oxygen-restricted
551 deposition in the upper Kapp Starostin Formation (Dustira et al. 2013), whereas dysoxia is not

552 seen in the shallower Festningen section until the latest Permian in the lower Vardebukta
553 Formation. Similarly, in the Sverdrup Basin Proemse et al. (2013) show at the LPE a strongly
554 developed oxygen minimum zone with euxinic conditions in deep water settings and oxic
555 shallow water environments. This suggests a gradual expansion of dysoxic bottom waters into
556 shallow water environments (Grasby et al., 2009; Proemse et al, 2013). It is during this
557 expansion phase that the LPE occurs, even in locations like Festningen where oxic waters
558 remained. As the habitable seafloor area shrank, the additional stress caused by intense trace
559 metal poisoning may have driven the extinction of the low pH-tolerant benthos of NW Pangea.
560 This relative timing of anoxia is consistent with paleo-marine temperature records that show
561 rapid warming of global oceans (which would drive enhanced anoxia) occurred after the main
562 extinction event (Joachimski et al., 2012; Sun et al., 2012). This may imply then that the initial
563 eruption of the Siberian Traps had an initial short term toxic metal loading effect of global
564 ecosystems that was followed by a delayed rapid global warming related to emissions of
565 greenhouse gases (Dustira et al., 2013; Grasby and Beauchamp, 2009).

566

567 **7.0 Summary and Conclusions**

568 The Festningen section shows a remarkable record of progressive environmental deterioration
569 through latest Permian time. Three major steps are observed, which we interpret as reflecting
570 progressive ecological damage. First was the gradual lysocline shoaling along the NW margin of
571 Pangea leading to the final loss of carbonate producers at 12 m below the top of the Kapp
572 Starostin Formation. Such loss of carbonate producers has been recorded over much of NW
573 Pangea, where carbonate factories contracted into increasingly narrow mid to inner shelf areas
574 throughout the Middle Permian, and were nearly eradicated by Late Permian time except for

575 nearshore environments (Beauchamp and Grasby, 2012; Bugge et al., 1995; Ehrenberg et al.,
576 2001; Gates et al., 2004). While these carbonate factories were lost, silica productivity was
577 maintained with the result that the nearshore siliceous limestones were replaced by across-the-
578 shelf spiculites (Beauchamp and Baud, 2002; Beauchamp and Desrochers, 1997; Beauchamp and
579 Grasby, 2012). This Lysocline shoaling would reflect a gradual process related to long term
580 changes in atmospheric CO₂, that was most strongly manifest along the NW margin of Pangea in
581 Late Permian time. However such affects would not be expressed at low latitude shelves in the
582 Tethys that maintained productive carbonate factories.

583 If correct, evidence for lysocline shoaling suggests that the Late Permian oceans were
584 under progressive increasing stress of marine systems leading up to the LPE event. Although,
585 even if the loss of carbonate producers may reflect a progressive shift to a more stressed marine
586 environment, siliceous sponges were able to still thrive and diverse bioturbators continued to
587 produce a pervasively burrowed fabric.

588 The second major environmental impacts is recorded at the LPE event itself, when the
589 loss of sponges and major loss of burrowing organisms occurs during oxic conditions. We argue
590 that high metal loading rates at this time reflects onset of massive eruption of the Siberian traps
591 and associated volatile and toxic element release to the global atmosphere. Although burrowing
592 animals still survived, trace fossils became limited to *Planolites* and small burrows (Wignall et al.,
593 1998). Coincidental with the marine LPE, pollen records at this time indicate dramatic shifts to
594 highly stressed terrestrial environments, that implies simultaneous collapse of both marine and
595 terrestrial systems.

596 The third major impact oobserved at Festningen is a distinct shift to anoxia 2.6 m above
597 the LPE horizon, associated with a distinct loss of remaining burrowers. We suggest that

598 development of anoxia provided the third and final blow to the survivors. The continued spread
599 of anoxia could have several causes. Rapid increasing sea temperatures occurred just after the
600 main extinction that would have decreased oxygen solubility (Romano et al., 2013; Sun et al.,
601 2012), and could have also driven release of any remaining deep-marine gas hydrates, which
602 would also consume dissolved oxygen in marine waters (Majorowicz, et al., 2014; Ruppel, 2011).

603 Results from this study show a remarkable record of environmental deterioration
604 associated with the LPE event that struck progressively down ecologic systems, and
605 demonstrates the need for high resolution studies to characterize the nature of rapid change in
606 global biogeochemical cycles during this dramatic period of Earth history.

607

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609

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1046 **Table captions**

1047 Table 1: SHRIMP U-Pb zircon data for the ash layer. Ages are quoted as $^{206}\text{Pb}/^{238}\text{U}$ dates for
1048 Paleozoic zircons and as $^{207}\text{Pb}/^{206}\text{Pb}$ dates for Archean zircons.

1049 Table 2: Geochemical data from the Festningen section. TC= total carbon, TOC = total organic
1050 carbon, TIC = total inorganic carbon, CIA = chemical index of alteration, n.d. = not
1051 determined.

1052

1053 Table 3: Calculated increase in metal loading rates due to the Siberian Trap eruptions, based on
1054 metal/S ratio of Nriagu (1989) (mid point of the range given was used) and total S flux of
1055 Siberian Trap Volcanism (Beerling et al., 2007). Two loading rates are calculated for a
1056 constant eruption rated over 500 ky, or a sporatic eruption over a net 40 ky time period.
1057 Natural modern flux from Pacyna and Pacyna (2001).

1058 **Figure Captions**

1059 Figure 1 Location maps of field area, showing A) global Late Permian reconstruction base map
1060 after R. Scotese, B) the location of the Festningen section on Spitsbergen, and C) the
1061 paleo locations of important sedimentary records on the NW margin of Pangea at the
1062 time of the LPE event (Embry, 1992).

1063 Figure 2 Field photographs of the Festningen section, showing: A) the top resistant bedding
1064 plane of the Kapp Starostin Formation and overlying sediments of the Vardebukta
1065 Formation, B) basal shales of the Vardebukta Formation, location shown in Fig. 2a, and

1066 C) close up of finely laminated shales that mark the loss of burrowers in the section,
1067 location in Fig. 2b.

1068 Figure 3 Results of age dating, showing A) Cathodoluminescence images and ages of selected
1069 zircons of the ash layer, and B) concordia plot of SHRIMP data for zircon grains from
1070 the ash layer.

1071 Figure 4 Plots of geochemical data from Festningen, including: A) $\delta^{13}\text{C}$ of organic carbon, B)
1072 percent total organic carbon (TOC), C) percent total inorganic carbon (TIC), D) nitrogen
1073 isotope values, E) chemical index of alteration (Sydeman et al.).

1074 Figure 5 Plots of redox sensitive indicators for Festningen, including: A) percent Fe pyrite
1075 (Fepy), B) Molybdenum (Mo), C) uranium (U), and D) vanadium (V). Solid circles show
1076 absolute concentrations whereas dashed lined represent normalised values.

1077 Figure 6 Geochemical plot of: A) percent total sulfur (TS) versus total organic carbon (TOC)
1078 showing the significant shift to more anoxic state after the first ash bed (white circles),
1079 B) inverse relationship between carbon isotope values and redox proxies across the
1080 extinction horizon.

1081 Figure 7 Plot of trace metals along with A) carbon isotope values, and B) Mo for reference.
1082 Trends in metals across the extinction horizon include: C) Cu, D) Pb, E) As, F) Co, G) Ni,
1083 E) trends of Hg normalised to total organic carbon (TOC). Average shale values are
1084 shown as vertical dashed lines derived from Taylor and McLennan (1985) for Pb, Co,
1085 Ni, and Wedepohl (1991) for Cu, and As. Colour bars represent interpreted major
1086 impacts on life across the extinction event, including initial marine water acidity,
1087 development of toxicity, then finally anoxia.

1088 Figure 8 Comparative plot of key sections from NW Pangea, Festningen (this study) and West
1089 Blind Fiord, Sverdrup Basin (Proemse et al., 2013).