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1 Trilobite extinctions, facies changes and the ROECE carbon isotope excursion at the
2 Cambrian Series 2 - 3 boundary, Great Basin, western USA.

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13

14 Abstract

15 The mass extinction of the olenellid trilobites occurred around the Cambrian Series 2 - Series 3
16 boundary. Like many other crises, it coincided with a negative carbon isotope excursion but the
17 associated palaeoenvironmental changes remain unclear. To investigate the causal mechanism for
18 this event, we report facies changes, pyrite framboid petrography and carbon isotope values from
19 Cambrian Series 2 - Series 3 (traditionally Early - Middle Cambrian) boundary strata of the Carrara
20 Formation (Death Valley region, California) and Pioche Formation (Nevada). These data reveal
21 regionally changing water depths from high-energy, nearshore facies (oolitic grainstone) to more
22 offshore silty marl and finer-grained carbonate mudstone. In the Carrara Formation, the series
23 boundary occurs within a deepening succession, transitioning from high-energy, nearshore facies
24 (oolitic grainstone and oncolitic packstone) to offshore marl, the latter of which contains pyrite

25 frambooid populations indicative of low-oxygen (dysoxic) depositional conditions. Intermittent
26 dysoxia persisted below sub-wave base settings throughout the early and middle Cambrian, but did
27 not intensify at the time of extinction, arguing against anoxia as a primary cause in the olenellid
28 trilobite extinction. Within both field areas, the extinction interval coincided with a minimum in
29 $\delta^{13}\text{C}_{\text{carb}}$ values, which we interpret as the regional manifestation of the Redlichiid-Olenellid Extinction
30 Carbon isotope Excursion (ROECE). The Series 2 - Series 3 boundary is reported to closely coincide
31 with a large-amplitude sea-level fall that produced the Sauk I/II sequence boundary, but the
32 placement of the Series 2 - Series 3 boundary within a transgressive interval of the Carrara
33 Formation shows that this is not the case. The main sequence boundary in the succession occurs
34 much lower in the succession (at the top of the Zabriskie Quartzite) and therefore precedes the
35 extinction of the olenellids and ROECE.

36 Keywords: Olenellid extinction, Carrara Formation, Pyramid Shale Member, Pioche Formation, C-
37 Shale Member

38

39 1. Introduction

40 The first major biotic crisis of the Phanerozoic occurred during the Cambrian Series 2, an
41 interval that saw the collapse of archaeocyathan reefs (Newell, 1972; Boucot, 1990; Debrenne, 1991;
42 Zhuravlev and Wood, 1996). This was followed, at the Series 2 - Series 3 boundary, by severe
43 generic-level losses of olenellid and redlichiid trilobites (Palmer, 1998; Zhu et al., 2004; Zhu et al.,
44 2006; Fan et al., 2011; Wang et al., 2011; Zhang et al., 2013). This trilobite extinction has been used
45 to delineate the Series 2 - Series 3 boundary; however, the boundary remains unratiated as
46 international correlation is confounded by a lack of globally-distributed taxa at this time (Sundberg
47 et al., 2016). The trilobite extinction coincides with a major negative $\delta^{13}\text{C}$ excursion that has been
48 termed the Redlichiid-Olenellid Extinction Carbon Isotope Excursion or ROECE (Zhu et al. 2004,
49 2006).

50 In the western Great Basin of the United States, a Cambrian sedimentary succession
51 developed on a rapidly subsiding passive margin (Prave, 1999; Stewart, 1972; Fedo and Cooper,
52 2001; Hogan et al., 2011; Keller et al., 2012; Morgan, 2012). Sections in the southern Nopah Ranges
53 (Keller et al., 2012) expose strata from Cambrian Sauk I and Sauk II supersequences that are of
54 importance to this study (Prave, 1991). These are widespread, large-scale Laurentian sequences that
55 provide a regional stratigraphic framework. The transition from Sauk I to Sauk II records a major
56 lithological change that saw the siliciclastic deposition of the Zabriskie Quartzite replaced by
57 carbonate deposition of the Carrara Formation (Keller et al., 2012; Morgan, 2012). The contact
58 between these two units is considered to be the Sauk I/II sequence boundary (Keller et al., 2012;
59 Morgan, 2012).

60 Throughout the Phanerozoic the relationship between environmental perturbation and
61 extinction is a common focus of studies, including those in the Cambrian (Hallam and Wignall, 1997;
62 Wignall, 2015). In particular, sea-level change, marine anoxia, carbon isotope excursions and
63 eruptions of LIPs (large igneous provinces) often coincide with mass extinctions (Zhuravlev and
64 Wood, 1996; Wignall, 2001; Glass and Phillips, 2006; Jourdan et al., 2014). Thus, Zhuravlev and
65 Wood (1996) noted the temporal link between widespread deposition of black shales and the
66 disappearance of the archaeocyathans in the Cambrian of Siberia, and trilobite extinctions at
67 “biomere” boundaries are also ascribed to dysoxia (Palmer, 1984). However, the role of anoxia in
68 Cambrian extinctions has to be viewed in the context of persistently oxygen-restricted oceans at this
69 time (e.g. Montañez et al., 2000; Hurtgen et al., 2009; Pruss et al., 2010; Gill et al., 2011; Saltzman et
70 al., 2015; Tarhan et al., 2015).

71 Volcanism may also have played a role in ROECE (Glass and Phillips, 2006). The Kalkarindji
72 LIP is a Cambrian flood basalt province of northern and central Australia with an estimated original
73 surface area of $\sim 2.1 \times 10^6 \text{ km}^2$ (Glass and Phillips, 2006; Jourdan et al., 2014; Marshall et al., 2016).

74 Latest dating efforts yield a zircon age of 510.7 ± 0.6 Ma, which is close to that of the Cambrian
75 Series 2 - Series 3 boundary (Jourdan et al., 2014).

76 In order to improve our understanding of the events associated with ROECE, this study
77 examines sections spanning the Cambrian Series 2 - Series 3 boundary interval in the western Great
78 Basin. The olenellid extinction horizon has been located within the Pioche Formation in Nevada
79 (Palmer, 1998). We have examined this level and the correlative levels in the Carrara Formation in
80 California in order to examine changes of lithofacies, carbon isotope variability and pyrite
81 petrography.

82

83 **2. Geological background and biostratigraphy**

84 Following breakup of the Rodinia supercontinent in the late Neoproterozoic, the
85 northwestern margin of Laurentia subsided rapidly (Bond and Kominz, 1984; Levy and Christie-Blick,
86 1991; Prave, 1999; Howley et al., 2006). By the early Cambrian, the western Great Basin (USA) was
87 positioned along the western margin of Laurentia where a wide, clastic-dominated shelf developed
88 in an equatorial setting (Palmer and Halley, 1979; MacNiocaill and Smethurst, 1994; Fig. 1). Clastic
89 input decreased in late Series 2 and was replaced by carbonate deposition (Erdtmann and Miller,
90 1981; Howley et al., 2006; Landing, 2012).

91 We have assessed environmental conditions across the Series 2 - Series 3 boundary from
92 sections in the Great Basin including Emigrant Pass (Nopah Range, Death Valley, eastern California),
93 and Oak Springs Summit (Burnt Springs Range, eastern Nevada; Fig. 1). The regional lithostratigraphy
94 of the Pioche Formation (at Oak Springs Summit) was described in detail by Merriam and Palmer
95 (1964), and the Carrara Formation (at Emigrant Pass) by Palmer and Halley (1979). In addition, we
96 examined a section of the Pioche Formation at Ruin Wash in eastern Nevada.

97

98 2.1. *Emigrant Pass*

99 At Emigrant Pass the Zabriskie Quartzite and Carrara Formation are easily accessible and
100 well exposed. The strata are seen on the north side of Old Spanish Trail Highway as a continuous
101 section of quartz arenite and shale forming slopes and moderately steep hillsides with limestone
102 forming prominent ledges (Figs. 1 and 2). The Zabriskie Quartzite is dominated by burrowed and
103 hummocky cross-bedded quartz arenite beds (Prave, 1991; Keller et al., 2012). It lacks age-diagnostic
104 fossils, but rocks immediately above and below have yielded fauna from the *Bonnia-Olenellus*
105 trilobite zone of Series 2 (Diehl, 1974; Palmer and Halley, 1979; Prave, 1991; Peng et al., 2012).

106 The Carrara Formation comprises cycles of silty marl and limestone with a trilobite fauna
107 spanning the *Bonnia-Olenellus* to the *Glossopleura* zones, and thus the Series 2 - Series 3 boundary
108 (Adams, 1995; Palmer and Halley, 1979; Sundberg and McCollum, 2000; Babcock et al., 2012; Keller
109 et al., 2012; Fig. 3). The unit has been divided into nine members in the western Great Basin: Eagle
110 Mountain Shale, Thimble Limestone, Echo Shale, Gold Ace Limestone, Pyramid Shale, Red Pass
111 Limestone, Pahrump Hills Shale, Jangle Limestone and the Desert Range Limestone (Palmer and
112 Halley, 1979). However, there is significant lateral variation within the Carrara Formation and at
113 Emigrant Pass, the Thimble Limestone is not present (Palmer and Halley, 1979; Adams and
114 Grotzinger, 1996). Overall the carbonate content of the Carrara Formation decreases in more
115 basinward settings to the west (Hogan et al., 2011; Keller et al., 2012; Foster, 2014). Within the
116 Carrara Formation at Emigrant Pass, five members are important to this study.

117 The Eagle Mountain Shale Member consists of green to grey-brown silty shale with
118 interbedded lenses and beds of bioclastic limestone developed towards the top (Palmer and Halley,
119 1979). This is overlain by the Echo Shale Member which consists of green, platy shale and brown-
120 orange limestone. The Echo Shale is correlated with the basal Combined Metals Member of the
121 Pioche Formation in eastern Nevada (Palmer and Halley, 1979). The succeeding Gold Ace Limestone
122 is a prominent, cliff-forming limestone (Cornwall and Kleinhampl, 1961). The strata include thin to

123 medium-bedded lime mudstone, with dolomitic beds and oncolitic limestone (Palmer and Halley,
124 1979). Based on shared trilobite zones, the Gold Ace Limestone correlates with the Combined
125 Metals Member of the Pioche Formation at Oak Springs Summit (Merriam and Palmer, 1964).

126 The overlying Pyramid Shale Member is a green shale with interbeds of brown and maroon
127 silty marl and lenses of oncolitic and bioclastic limestone. Trilobite biostratigraphy indicates the
128 Pyramid Shale is equivalent to two members of the Pioche Formation in eastern Nevada: the C-Shale
129 Member and the Susan Duster Limestone Member (Palmer and Halley, 1979; Palmer, 1998). The Red
130 Pass Limestone Member is the youngest unit examined in Death Valley. It forms prominent cliffs of
131 oncolitic and bioclastic limestone, laminated lime mudstone and fenestral lime mudstone (Palmer
132 and Halley, 1979). There is no equivalent limestone unit in the Pioche Formation of Nevada (Palmer
133 and Halley, 1979).

134

135 2.2. *Oak Springs Summit*

136 At Oak Springs Summit, the Pioche Formation crops out to the west of a parking area in a dry
137 river bed. Limestone forms more prominent ledges and platforms whilst shale forms recessively
138 weathered outcrop (Figs. 1 and 2). The Combined Metals Member is composed of silty, oncolite-
139 bearing dark limestone with *Olenellus* (Palmer, 1998; Sundberg and McCollum, 2000; Hollingsworth
140 et al., 2011). It is overlain by the C- Shale Member (formerly the Comet Shale), a series of shale and
141 thin-bedded limestone beds with pinch-and-swell bed boundaries (Palmer, 1998; Sundberg and
142 McCollum, 2000). The Series 2 - Series 3 boundary is placed at the base of the C-Shale due to the
143 sudden disappearance of the Olenellidae, and their replacement by a fauna dominated by
144 *Eoptychoparia piochensis* at this level (Palmer, 1998; Sundberg and McCollum, 2000). The
145 succeeding Susan Duster Limestone Member is a well-bedded, grey marl with occasional argillaceous
146 and bioclastic limestone beds composed of trilobite fragments (Merriam and Palmer, 1964).

147

148 2.3. Biostratigraphy

149 Trilobite assemblages from the Carrara and Pioche formations belong to the *Olenellus*,
150 *Eokochaspis nodosa*, *Amecephalus arrosensis* and the *Plagiura-Poliella* zones that provide a
151 framework for regional correlation (Merriam and Palmer, 1964; Palmer and Halley, 1979; Fig. 3). The
152 *Olenellus* Zone ranges from the Zabriskie Quartzite to the basal portion of the Pyramid Shale
153 Member within Death Valley (Palmer and Halley, 1979; Fig. 2). In Nevada, this zone spans the
154 Delamar Member to the base of the C-Shale (Merriam and Palmer, 1964; Sundberg and McCollum,
155 2000). All olenellid trilobites disappear over a ~ 2 cm interval at the top of the zone, forming a
156 distinct extinction horizon (Palmer, 1998; Fig. 2). This is immediately followed by first appearance of
157 the ptychopariid trilobite *Eokochaspis nodosa*, which defines both the base of the *E. nodosa* Zone,
158 and the Series 2 - Series 3 boundary (Sundberg and McCollum, 2000; Fig. 2). *E. nodosa* Zone faunas
159 also occur in the Pyramid Shale Member in Death Valley (Sundberg and McCollum, 2000).

160 The succeeding *Amocephalus arrosensis* Zone contains *A. arrosensis*, *Mexicella robusta*
161 and *Kochina? walcotti*. The zone is best defined at Hidden Valley, Nevada, where its base is 30 m
162 above the base of the C-Shale Member (Merriam and Palmer, 1964), but it has not been recorded in
163 the Carrara Formation due to a paucity of fossils above the *E. nodosa* Zone in the Pyramid Shale
164 (Palmer and Halley, 1979). However, trilobites, from the *Plagiura- Poliella* Zone, reappear in the
165 uppermost Pyramid Shale and lower Red Pass Limestone (Palmer and Halley, 1979).

166

167 3. Methods

168 Sedimentary logging of the Carrara and Pioche formations was undertaken. At Emigrant
169 Pass, 170 m of the Carrara Formation was logged from the base of the formation (at the contact with
170 the Zabriskie Quartzite) up to the Red Pass Limestone Member. At Oak Springs Summit a 53 m-thick
171 section of the Pioche Formation was logged, ranging from the basal Combined Metals Member to
172 Susan Duster Limestone Member, an interval correlative with the Emigrant Pass section based on

173 trilobite biostratigraphy (Merriam and Palmer, 1964; Palmer and Halley, 1979; Fig. 2). From these
174 logs, four facies were defined (discussed below, Table 1). At the two study sections, 30 samples from
175 the Pioche Formation and 57 samples from the Carrara Formation were analysed for $\delta^{13}\text{C}_{\text{carb}}$ (Table
176 2). In Lincoln County, Nevada, we also sampled the Ruin Wash location (Palmer, 1998; Lieberman,
177 2003) for additional facies and framboid analysis. Ruin Wash provided a second section (after Oak
178 Springs Summit) where the extinction horizon of the olenellids is clearly seen (see Supplementary
179 Material, Fig. S1). Facies analysis was undertaken in the field and complemented by petrographic
180 examination of 49 thin sections. In order to evaluate redox conditions, pyrite framboid size
181 distribution was also assessed on 21 samples using a scanning electron microscope (FEI Quanta 650
182 FEG-ESEM) in backscatter mode (see Bond and Wignall (2010) for procedure).

183 The calcite carbon ($^{13}\text{C} / ^{12}\text{C}$) and oxygen ($^{18}\text{O} / ^{16}\text{O}$) isotope values of powdered bulk
184 sediment samples were measured on a total of 98 samples at the GeoZentrum Nordbayern, FAU
185 Erlangen-Nurnberg, Germany (27 samples) and the School of Environmental Sciences, University of
186 Liverpool, UK (71 samples). Carbon dioxide was prepared by reaction with phosphoric acid either at
187 70°C using a Gasbench II preparation system (FAU) or at 25 °C using the classical, ‘sealed vessel’
188 method (UoL). Mass ratios of the resultant purified gases were measured with a ThermoFisher Delta
189 V plus mass spectrometer operating in continuous flow mode (FAU) or a VG SIRA 10 dual-inlet mass
190 spectrometer (UoL). Raw gas data were corrected for ^{17}O effects and calibrated to the VPDB scale
191 using a combination of international reference materials ($\delta^{13}\text{C}$ values are assigned as +1.95 ‰ to
192 NBS 19 and -46.6 ‰ to LSVEC and $\delta^{18}\text{O}$ values of -2.20 ‰ to NBS19 and -23.2 ‰ to NBS18) and
193 laboratory quality control materials and reported as conventional delta (δ) values. Analytical
194 precision (1σ) is estimated to be better than 0.1 ‰ for both isotope ratios based on replicate
195 analysis of standards. Some notable differences in oxygen isotope values were reported where
196 specific samples were duplicated by both laboratories. The reason for these differences is uncertain.
197 Although some discrepancies were found to be significant, they do not impact on either the
198 palaeoenvironmental or chemostratigraphic interpretation of the carbon isotope data.

199 The topmost 8 samples of the Carrara Formation were analysed at the University of
200 Massachusetts, Amherst. Powdered, homogenized samples were analysed for $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$
201 values using a Finnigan Delta XL+ isotope ratio mass spectrometer with an automated carbonate
202 prep system (Kiel III). We report results as the per mille difference between sample and the VPDB
203 standard in delta notation where $\delta^{18}\text{O}$ or $\delta^{13}\text{C} = (R_{\text{sample}} / R_{\text{standard}} - 1) \times 1000$, and R is the ratio of the
204 minor to the major isotope. Results were calibrated using a house standard (crushed, washed and
205 sieved marble) with VPDB values of +1.28 for $\delta^{13}\text{C}$ ‰ and -8.48 ‰ for $\delta^{18}\text{O}$. Reproducibility of
206 standard materials is 0.1 ‰ for $\delta^{18}\text{O}$ and 0.05 ‰ for $\delta^{13}\text{C}$.

207 Total carbon (TC) and total organic carbon (TOC), following removal of calcite by acid
208 decomposition of bulk sediment samples, was measured using a LECO SC-144DR Dual Range carbon
209 and sulphur analyser at the University of Leeds. Total inorganic carbon (TIC) was subsequently
210 calculated by difference (TIC = TC - TOC). An estimate of the calcite content for each sample was
211 made by assuming that all TIC is hosted by calcite (wt % calcite = TIC x 8.333).

212

213 **4. Results**

214 *4.1. Facies Analysis*

215 Four facies and nine sub-facies were identified (Table 1): grainstone, packstone, silty marl
216 and marl and they have been grouped into an onshore-offshore trend spanning shallow subtidal to
217 outer shelf environments (Table 1). The shallowest strata consist of grainstone facies with common
218 shell hash that is often abraded. Beds are typically decimetres thick and can show a hummocky top
219 surface and sharp, erosive bases. Inclined stacks of flat-pebble intraclasts with herringbone-like cross
220 stratification are present, suggesting storm wave processes (Fig. 4D). Bioturbated packstone, with
221 sub-facies of oncolitic, bioclastic and silty packstone varieties (Fig. 5B), are interpreted to be a
222 deeper facies based on the presence of a micritic mud matrix. Deeper water silty marl include fissile,
223 homogenous and thoroughly bioturbated variants. Deepest-water, most offshore successions are

224 dominated by fine grained marl, including sub-facies of laminated, pyritic dolomicrite and
225 bioturbated marl with ichnofabric index (II) values of 2 - 3 (II2 and II3) in the scheme of Droser and
226 Bottjer (1986).

227 The facies distribution reveals consistent trends in the two principal study sections. The base of
228 the Carrara Formation, seen at Emigrant Pass, is dominated by deeper water facies (Facies 4) of the
229 Eagle Mountain Shale. Commonly, the marl has a grey-green colour produced by the abundance of
230 chlorite and clinocllore in the matrix (Figs. 4F, 5A and 6E). Laminated intervals are common,
231 although these occur interbedded with burrowed strata suggesting there were frequent fluctuations
232 of redox conditions.

233 The exception to the generally quiet, low-energy deposition of the Eagle Mountain Shale is
234 recorded by a sharp, erosive-based bed of Facies 1 developed just over 30 m above the base of the
235 section. This shelly, oncolitic packstone contains rip-up clasts of the underlying marl and sole marks
236 on its base (Fig. 7A). Internally, thin intraclasts and shells display a chevron-stacking pattern (Fig.
237 4D). A major storm event seems likely to have produced this horizon with the shell-stacking
238 produced by powerful bi-directional currents. The succeeding Echo Shale and Gold Ace members
239 record shallowing. Grainstone and packstone dominate this 20-m-thick interval which includes
240 erosive-based oncolitic packstone beds (Fig. 7B). Above this in the Pyramid Shale Member marl
241 facies dominate, taken to be indicative of a sustained deepening. Grainstone and packstone facies
242 developed in the lower ~ 15 m of the member are interpreted to have been transported during
243 storm events. It is within this transgressive phase that the Series 2 - Series 3 boundary is recorded,
244 along with the olenellid extinction (Foster, 2014). Deep-water sedimentation is abruptly terminated
245 by the development of shallow-water grainstone at the base of the Red Pass Limestone Member
246 (Fig. 7). Ooids and abraded fossil material (Table 1, Sub-Facies 1.1) suggest a nearshore setting.

247 The Pioche Formation at Oak Springs Summit records a more distal version of the succession
248 seen within the Carrara Formation with relatively deep-water Facies 3 and 4 dominating (Fig. 2),

249 though the same overall deepening-upwards trend is seen. Thus, the lower half of the Combined
250 Metals Member consists of alternating silty marl and packstone. Above this, the remainder of the
251 section is dominated by deeper-water facies (Fig. 2). The uppermost Combined Metals Member and
252 the majority of the C-Shale Member record a similar transgressive deepening seen within the
253 Pyramid Shale Member of the Carrara Formation. The olenellid extinction level occurs within this
254 transgressive succession between a marl and a silty marl in the base of the C-Shale Member (Fig. 4E).
255 This minor facies shift does not represent a significantly different environment and as such
256 extinction is not thought to be a function of facies change. Immediately above the extinction
257 horizon, chlorite in the form of both rounded grains and cement becomes common (Fig. 5A). The
258 remainder of the C-Shale Member is a thick package of marl that transitions to silty marl at the base
259 of the Susan Duster Limestone.

260

261 *4.2. Pyrite Framboid Analysis*

262 Framboid size analysis was performed on the Series 2 - Series 3 boundary strata (and thus
263 the extinction horizon) from the Pioche Formation at Oak Springs Summit, where 11 samples were
264 collected in a 7 m interval spanning 3.5 m either side of the extinction horizon. All samples contained
265 abundant scattered crystals of pyrite ranging in size from 1-10 μm , often found agglomerated in
266 clustered patches. Five samples yielded framboids preserved as iron oxyhydroxides due to
267 weathering, with only minor amounts of original pyrite preserved in their core. The framboids
268 showed a size distribution spanning an anoxic-dysoxic range (Fig. 8). The most dysoxic sample
269 (smallest mean framboid diameter size and size range, lowest standard deviation) occurred in a marl
270 approximately 1 m below the extinction horizon. Dysoxic framboid populations also occurred in the
271 1 m of strata overlying the extinction level. However, a sample from 20 cm below the extinction
272 level did not yield any pyrite framboids suggesting fully oxygenated conditions. This variable degree

273 of oxygen-restriction suggested by the framboid analysis is also seen in the variability of the
274 associated sedimentary fabrics, which varies from laminated to slightly burrowed (II2).

275 Seven samples were also analysed from the Pioche Formation at Ruin Wash where the
276 olenellid extinction horizon has been located within a succession of marls (Palmer, 1998; Lieberman,
277 2003; Fig. S1). Generally, framboidal pyrite was absent at this location with the exception of two
278 samples from 10 and 15 cm below the extinction horizon where they had size ranges that plot in the
279 anoxic field (Fig. 8). An additional four samples from around the extinction horizon at Emigrant Pass
280 were also analysed. In this case, all samples only yielded scattered pyrite crystals but not framboids,
281 suggesting better oxygenated conditions in this shallower-water section.

282

283 4.3. Chemostratigraphy

284 In the basal 20 m of the Carrara Formation $\delta^{13}\text{C}$ values are highly variable and do not show a
285 clear trend (Fig. 7), but they then begin to stabilise around -2‰ before a consistent positive trend
286 develops. In the Pyramid Shale Member the base of the negative excursion begins with at -0.1‰ ,
287 above this $\delta^{13}\text{C}_{\text{carb}}$ values begin a decline to a lowpoint at 105 m of -3.5‰ (a negative shift of 3.4
288 ‰). In the overlying 25 m, no data was obtained because carbonate values were too low for
289 analysis. Above this gap, $\delta^{13}\text{C}_{\text{carb}}$ values show a positive trend, returning to values around -0.1‰ ,
290 similar to those from the base of the section.

291 Barring one outlier at the base of the section, $\delta^{13}\text{C}_{\text{carb}}$ values from the first 13 m of the
292 Pioche Formation remain around -2.5‰ before there is a sharp, positive shift to -1.0‰ over the
293 next 15 m (Fig. 7). From this value of -1.0‰ a negative shift occurs, resulting in peak negative
294 values of -4.8‰ . The nadir at the base of the C-Shale Member marks an overall shift of -3.8‰ , a
295 similar size to that found at Emigrant Pass. At the top of the section values return to around 0‰ .

296

297 5. Discussion

298 5.1. Carbon isotopes and diagenesis

299 In order to evaluate the reliability of our isotope data we assess the preservation of a
300 primary carbon isotope signal in our samples. In both the Carrara and Pioche formations the isotope
301 analyses are derived from samples with a wide range of carbonate values (Table 2). The high TIC
302 samples are likely to record primary carbon isotope signatures since they are buffered from external
303 change by a large carbonate-carbon reservoir (Saltzman and Thomas, 2012). Lower TIC samples are
304 more susceptible to post depositional isotopic alteration or addition of carbonate with a non-
305 primary carbon isotope composition (Brand and Veizer, 1981; Banner and Hanson, 1990; Marshall,
306 1992). In both sections, the excursion to the lowest $\delta^{13}\text{C}_{\text{carb}}$ values occurs at the level of the trilobite
307 extinction (mid-Pyramid Shale, Carrara Formation and basal C-Shale, Pioche Formation) where TIC is
308 < 2 wt. %. Here we assess the preservation of a primary carbon isotope composition, particularly in
309 samples with low TIC.

310 Two diagenetic processes can alter the primary isotopic composition: recrystallization of
311 carbonate or precipitation of additional authigenic carbonate with a distinct isotope composition
312 (Marshall, 1992). Both marine pore fluids and meteoric waters can have dissolved inorganic carbon
313 (DIC) enriched in ^{12}C from the oxidation of organic matter and these mechanisms have differing
314 predictions of the $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ values preserved (Marshall, 1992). Both the Carrara and
315 Pioche formations display commonalities in their relationships between their $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$
316 ratios and their TIC and TOC concentrations. Firstly (point 1), neither formation shows a clear
317 relationship between $\delta^{13}\text{C}_{\text{carb}}$ and $\delta^{18}\text{O}_{\text{carb}}$ (Figs. S2 and S3). Secondly (point 2), samples with the most
318 negative $\delta^{13}\text{C}_{\text{carb}}$ and the most positive $\delta^{18}\text{O}_{\text{carb}}$ are mostly characterised by low TIC (defined as < 2
319 wt. %). Both the Carrara and Pioche formations exhibit generally low TOC (point 3). In the Carrara
320 Formation TOC concentrations range from 5.17 to 0.0 wt. % TOC with a mean concentration of 0.14
321 wt. % TOC. In the Pioche Formation concentrations range from 2.69 to 0.0 wt. % TOC, with a mean of

322 0.12 wt. % TOC. Finally (point 4), high TOC samples are characterised by more positive $\delta^{13}\text{C}_{\text{carb}}$. The
323 major difference between the sections for these parameters is a much clearer positive relationship
324 between TIC and $\delta^{13}\text{C}_{\text{carb}}$ within the Pioche Formation.

325 These observations rule out wholesale recrystallization in a meteoric fluid since neither
326 section displays a positive correlation between $\delta^{18}\text{O}_{\text{carb}}$ and $\delta^{13}\text{C}_{\text{carb}}$ (point 1, Figs. S2 and S3). The
327 generally low TOC concentrations and the relationship between TOC and $\delta^{13}\text{C}_{\text{carb}}$ (point 3 and 4) also
328 makes localised precipitation of organic-carbon derived DIC doubtful. From the relationships
329 between $\delta^{18}\text{O}_{\text{carb}}$ and TIC (point 2) it is likely that a proportion of the low TIC samples (< 2 wt. %)
330 have undergone variable resetting of their $\delta^{18}\text{O}_{\text{carb}}$ towards more positive values. This observation is
331 not consistent with precipitation of additional carbonate from unmodified meteoric or marine early
332 diagenetic pore fluids, where the expectation would be a change towards more negative $\delta^{18}\text{O}_{\text{carb}}$
333 values (Marshall, 1992; Knauth and Kennedy, 2009; Cochran et al., 2010; Saltzman and Thomas,
334 2012). The remaining possibility to explain the oxygen isotope relationships is variable exchange
335 with, or precipitation of carbonate from, a hypothetical high $\delta^{18}\text{O}$ fluid (Glumac and Walker, 1998).
336 Since the climate at both sites is currently arid, one possibility is that the fluid in question is derived
337 from evaporated modern meteoric water, but other possibilities exist (Saltzman and Thomas, 2012).

338 The relationships between $\delta^{13}\text{C}_{\text{carb}}$ and TIC differ somewhat from those between $\delta^{18}\text{O}_{\text{carb}}$ and
339 TIC: from the Carrara Formation, the range of $\delta^{13}\text{C}_{\text{carb}}$ in the < 2 wt. % TIC samples overlaps strongly
340 with the range found in near pure limestone samples suggesting that the influence of diagenetic
341 process on $\delta^{13}\text{C}_{\text{carb}}$ at this site is likely to be minimal (Fig. S4). In contrast, samples from the Pioche
342 Formation display a much clearer division between these two groups (TIC groups annotated in Fig.
343 S5). This suggests that the influence of post-depositional process on $\delta^{13}\text{C}_{\text{carb}}$ may have been more
344 pronounced at this site. However, the $\delta^{18}\text{O}_{\text{carb}}$ ranges of both high and low TIC samples of the Pioche
345 Formation overlap (Fig. S6), indicating that at least some of the carbon isotope values have
346 undergone minimal resetting.

347 In summary, there is clear evidence for a variable degree of oxygen isotope resetting
348 towards more positive values, which is particularly pronounced in samples with low TIC (< 2 wt. %).
349 There is also some evidence of concurrent variable resetting of carbon isotopes to more negative
350 values in low TIC samples, with this being somewhat more pronounced in the Pioche Formation.
351 Nonetheless, the presence of the negative $\delta^{13}\text{C}_{\text{carb}}$ values and the consistency of the magnitude of
352 the excursion at level of the Series 2 - Series 3 boundary (e.g., Zhu et al., 2006; Faggetter et al.,
353 2016), correlated independently by biostratigraphy between the two sections suggest that these
354 samples record a predominantly primary signal. As such, we conclude that the negative $\delta^{13}\text{C}_{\text{carb}}$
355 excursion within the Carrara and Pioche formations preserves a primary record, given its co-
356 occurrence with the olenellid extinction horizon, we interpret it to be ROECE.

357

358 *5.2. Extinction and palaeoenvironmental change*

359 Identification of the ROECE in the Pioche and Carrara formations (Fig. 7) confirms the close
360 temporal relationship between trilobite extinctions and carbon isotope excursions (Zhu et al., 2006;
361 Faggetter et al., 2016). It also allows examination of the associated facies and relative sea-level
362 changes at this time. Initially the Sauk I/II supersequence boundary was placed around the Series 2 -
363 Series 3 boundary (Sloss 1963). However, more recently this has been placed lower in the succession
364 at the top of the Zabriskie Quartzite, underlying the Carrara Formation (Prave, 1991). Thus the
365 Carrara Formation falls entirely within Sauk II (Keller et al., 2003, 2012; Morgan, 2012). Nonetheless,
366 there are alternative regression surfaces in the Carrara Formation. A candidate for a sequence
367 boundary occurs at the base of the Red Pass Limestone where there is a sharp transition from deep-
368 water to shallow-water. This level lies around 45 m above ROECE in the Carrara Formation.

369 Rather than regression, the olenellid extinction occurs within a deepening succession.
370 Transgression and shelf anoxia often go hand-in-hand, and oxygen stress has been implicated in
371 ROECE extinction (Montañez et al., 2000). However, at Oak Springs Summit, pyrite framboid analysis

372 suggests dysoxic but not euxinic conditions in the extinction interval, and the shallower study
373 locations show no evidence for oxygen restriction. The evidence for intensified oxygen-restricted
374 deposition at the trilobite extinction level is therefore weak. It also noteworthy that low-oxygen
375 conditions were common in Cambrian oceans (Hurtgen et al., 2009; Pruss et al., 2010; Gill et al.,
376 2011), and there is no suggestion that anoxia was intensified at the level of ROECE.

377 The Series 2 - Series 3 boundary interval saw the eruption of the Kalkarindji flood basalt
378 province (Glass and Phillips, 2006; Jourdan et al., 2014; Marshall et al., 2016). In younger intervals of
379 the Phanerozoic, the formation of large igneous provinces frequently coincides with mass
380 extinctions (Wignall, 2015; Bond and Grasby, 2016) and the eruption of large volumes of volcanic
381 volatiles provides a causal mechanism for driving biologic crises. The contemporaneous negative
382 $\delta^{13}\text{C}$ signal of ROECE is often seen at times of LIP eruptions and may record the influx of isotopically-
383 light volcanic CO_2 (e.g., Payne et al. 2004). Thus, in many regards the ROECE has the hallmarks of
384 later Phanerozoic LIP-related mass extinctions although evidence for the commonly associated
385 environmental changes such as the spread of anoxia (Wignall, 2015), is not clearly established for
386 this Cambrian example.

387

388 **6. Conclusions**

389 In the western Great Basin, USA, the extinction of the dominant olenellid trilobites occurs
390 within a deepening-upward shelf succession. A major -3.5% negative carbon isotope excursion
391 (ROECE) occurs at the same level. This extinction/isotope event occurs around the Cambrian Series 2
392 - Series 3 boundary interval. Pyrite framboid size distribution data and laminated facies suggest
393 periodic dysoxia occurred in the facies immediately surrounding the extinction horizon. However,
394 these conditions were neither widespread (shallower-water boundary sections in Death Valley do
395 not record oxygen starvation) nor especially unusual (laminated strata are sporadically developed
396 throughout the offshore units of the Carrara Formation) suggesting dysoxia did not play a major role

397 in the extinction. The environmental effects of the contemporaneous Kalkarindji flood basalt
398 province of Australia provide a better potential causal for the extinction at the Series 2 - Series 3
399 boundary, although detailed correlation with the sections in North America is required.

400

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408

409 **Figure captions**

410 Figure 1. Location map showing study sections at Emigrant Pass, Death Valley region, California (35°

411 53' 29.24"N, 116° 04' 39.08"W) and Oak Springs Summit, Burnt Spring Range, Lincoln County,

412 Nevada (37°37'04.32"N 114°43'17.20"W). Star indicates approximate location of field area during

413 the Cambrian Series 2 (after McKerrow et al., 1992).

414 Figure 2. Biostratigraphic correlation of the trilobite zones of the Carrara and Pioche formations

415 (Palmer and Halley, 1979; Sundberg and McCollum, 2000). Facies column is based on field and

416 petrographic observations, and numbers relate to facies detailed in Table 1. A generalised

417 stratigraphic column of Precambrian and Cambrian formations in Death Valley is given (from Corsetti

418 and Hagadorn, 2000).

419 Figure 3. Correlation of trilobite biozones within the Carrara and Pioche formations (Merriam and

420 Palmer, 1965; Palmer and Halley, 1979; Sundberg and McCollum, 2000). *O.* is *Olenellus*, *P.* is *Poliella*.

421 Figure 4. Field photographs.

422 A. Trilobite debris (spines and carapaces and hyoliths) in a bioclastic hash on bedding planes of
423 oolitic grainstone, Carrara Formation. B. Oncolitic packstone facies at Emigrant Pass. C. Bifurcating
424 burrows in well bioturbated silty marl at Emigrant Pass, notebook for scale. D. Oolitic grainstone
425 facies showing inclined chevron-style packing of thin intraclast and bioclasts (hyolith, ooid and other
426 detrital fragments). E. Olenellid extinction level at the base of the C-Shale Member at Oak Springs
427 Summit. Red line indicates extinction horizon from Palmer (1998). F. Fissile, laminated marl and silty
428 marl in the lower Eagle Mountain Shale at Emigrant Pass. G. *Thalassinoides* in fine-grained, silty marl
429 of the Carrara Formation. H. Vertical burrows (at the hammer tip) in silty marl beds of the Carrara
430 Formation.

431 Figure 5. Scans of thin sections and photomicrographs.

432 A. Photomicrograph of a range of chlorite in the silty marl facies immediately above the extinction
433 horizon at Oak Springs Summit. Chlorite occurs as elongate grains and also as cement. B:
434 Photomicrograph of a silty bioclastic packstone in the upper Eagle Mountain Shale. C: Scan of slide of
435 oncolitic packstone (Eagle Mountain Shale) showing oncoids with bioclastic nucleus of echinoderm
436 plates amongst a matrix of shell detritus and micrite. D: Scan of slide of bioclastic grainstone.
437 Elongate, trilobite fragments dominate this facies alongside hyolith remains and echinoderm plates.
438 Dark brown mineral growth shows iron oxide preferentially replacing shell material.

439 Figure 6. Scans of thin sections and photomicrographs.

440 A: Scan of silty marl showing quartz grains and detrital chlorite grains (green). B: Photomicrograph of
441 marl facies in the Combined Metals Member, Pioche Formation. Trilobite carapace exhibits brown
442 needle like iron oxide replacement of the calcite shell. C: Photomicrograph of peloidal grainstone
443 facies in the Combined Metals Member, Pioche Formation. Well rounded micrite pellets alongside
444 rounded quartz grains amongst a fine micrite matrix. D: Photomicrograph of oolitic, bioclastic
445 grainstone with iron oxides partially replacing ooids. E: Photomicrograph of silty chloritic limestone

446 showing rounded chlorite grains (white dashed lines). F: Photomicrograph of chloritic silty marl
447 facies showing sub-angular to angular quartz sand grains alongside hololith and trilobite debris.

448 Figure 7. Carbon isotope chemostratigraphy of the Carrara Formation at Emigrant Pass and Pioche
449 Formation at Oak Springs Summit. A. Inset log shows contact between silty micrite and an erosive-
450 based oncolitic packstone with rip up clasts of the underlying silty micrite. This horizon grades
451 laterally into an oolitic grainstone. B. Inset log of contact between silty bioclastic packstone and an
452 erosive-based oncolitic packstone. Both erosional surfaces mark the transport of shallow water
453 bioclastic material during storm events.

454 Figure 8. Size versus standard deviation for framboids from Series 2 - Series 3 boundary strata of
455 California and Nevada showing the presence of oxygen-restricted facies. The threshold separating
456 euxinic/anoxic and dysoxic/oxic size ranges in modern environments is from Wilkin et al. (1996).

457 Table 1: Facies of the Carrara and Pioche formations.

458 Table 2. Geochemical and framboid measurements for the Carrara and Pioche formations at
459 Emigrant Pass (EP) and Oak Springs Summit (OSS) and framboid data from the Pioche Formation at
460 Ruin Wash (RW).

461

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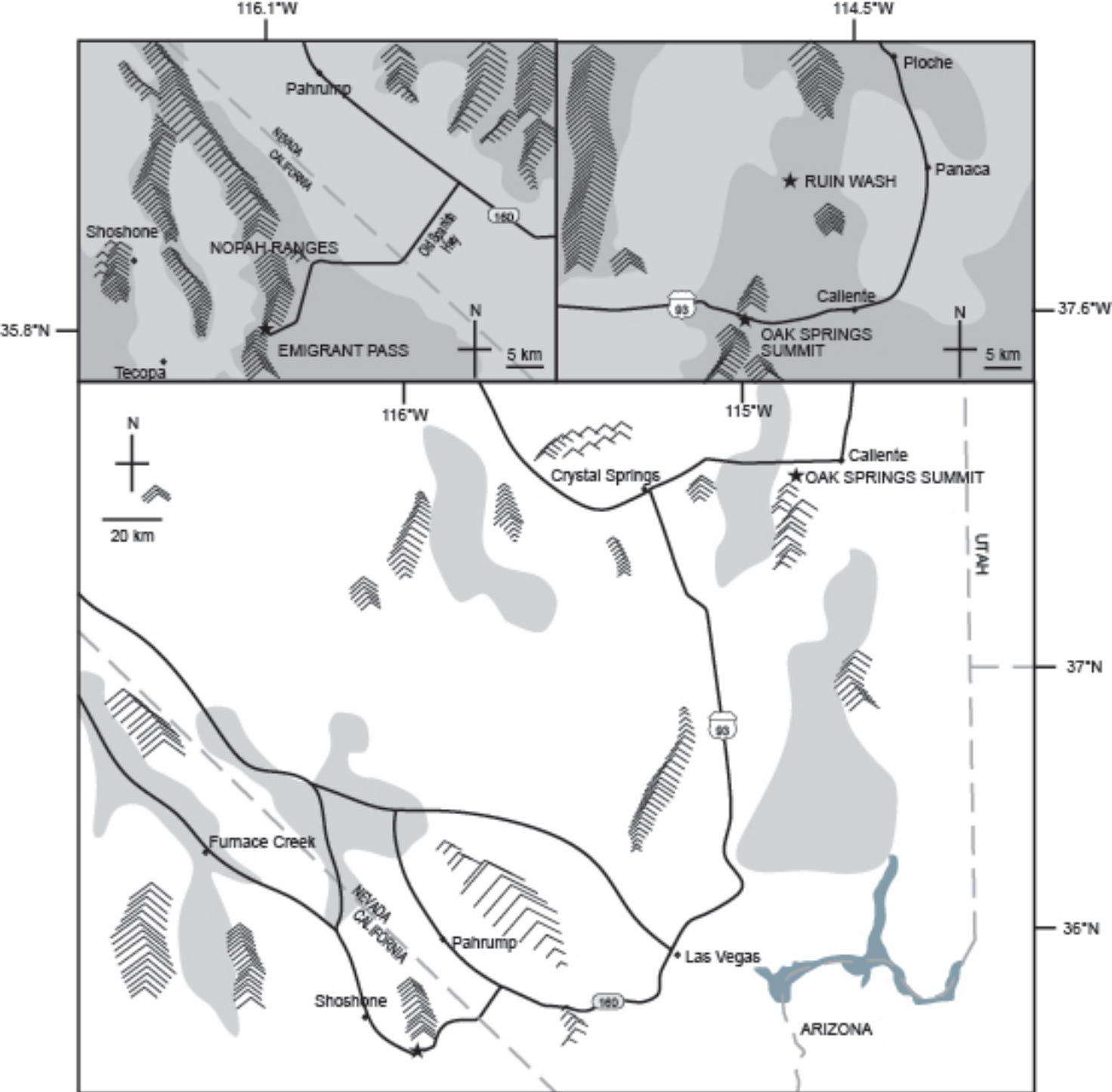
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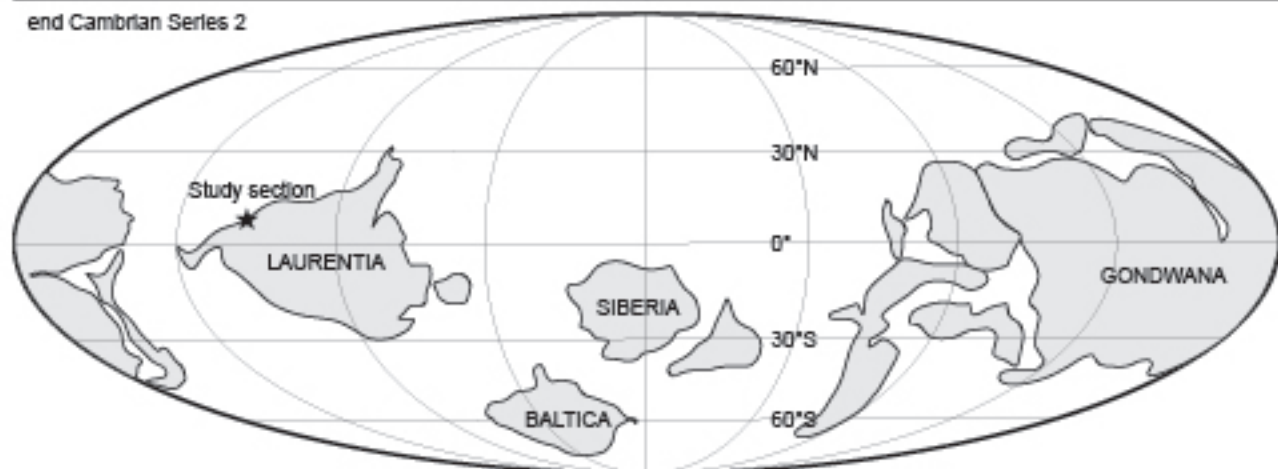
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end Cambrian Series 2



CARRARA FM,
EMIGRANT
PASS, CA

116°W

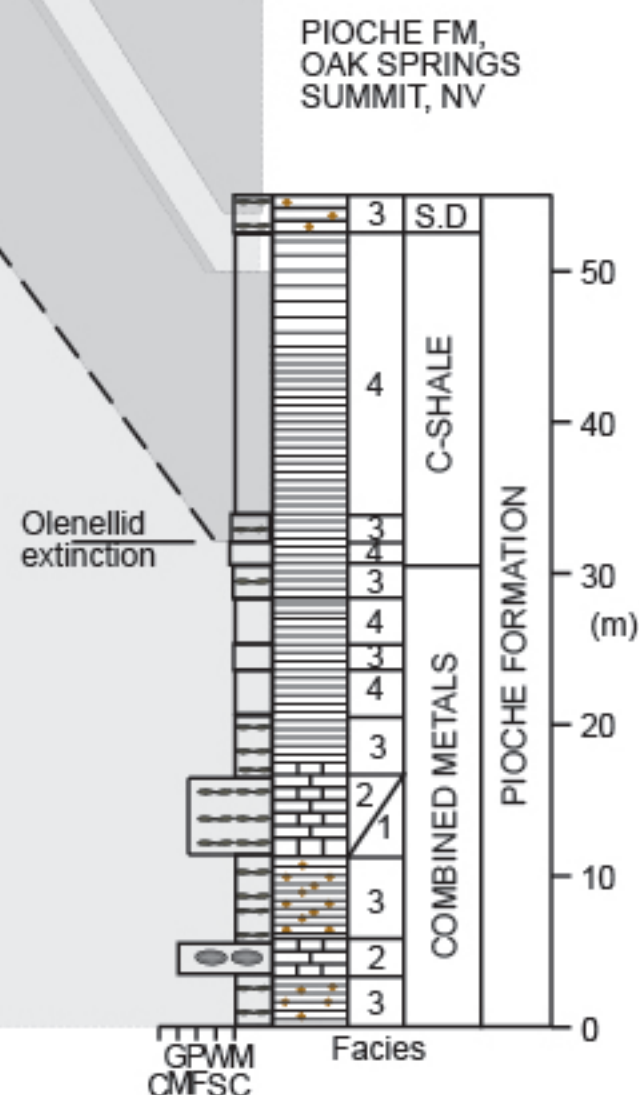
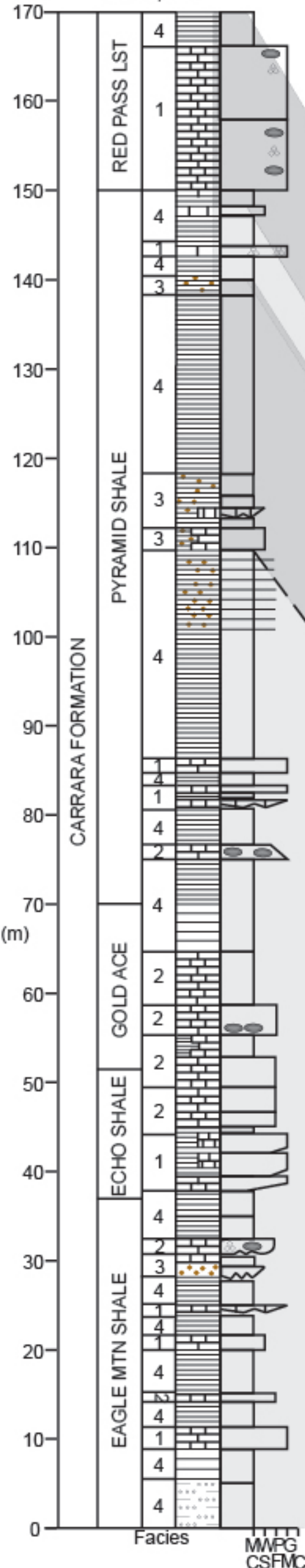
37°N



BIOZONES

1. *Olenellus* Zone
2. *Eokochaspis nodosa* Zone
3. *Amecephalus arjosensis* Zone
4. *Plagiura-Poliella* Zone

E Series 3	Bonanza King
	Carrara Fm
E Series 2	Zabriskie Quartzite
	Wood Canyon Fm
Pre E	Stirling Quartzite



(m)

50

40

30

20

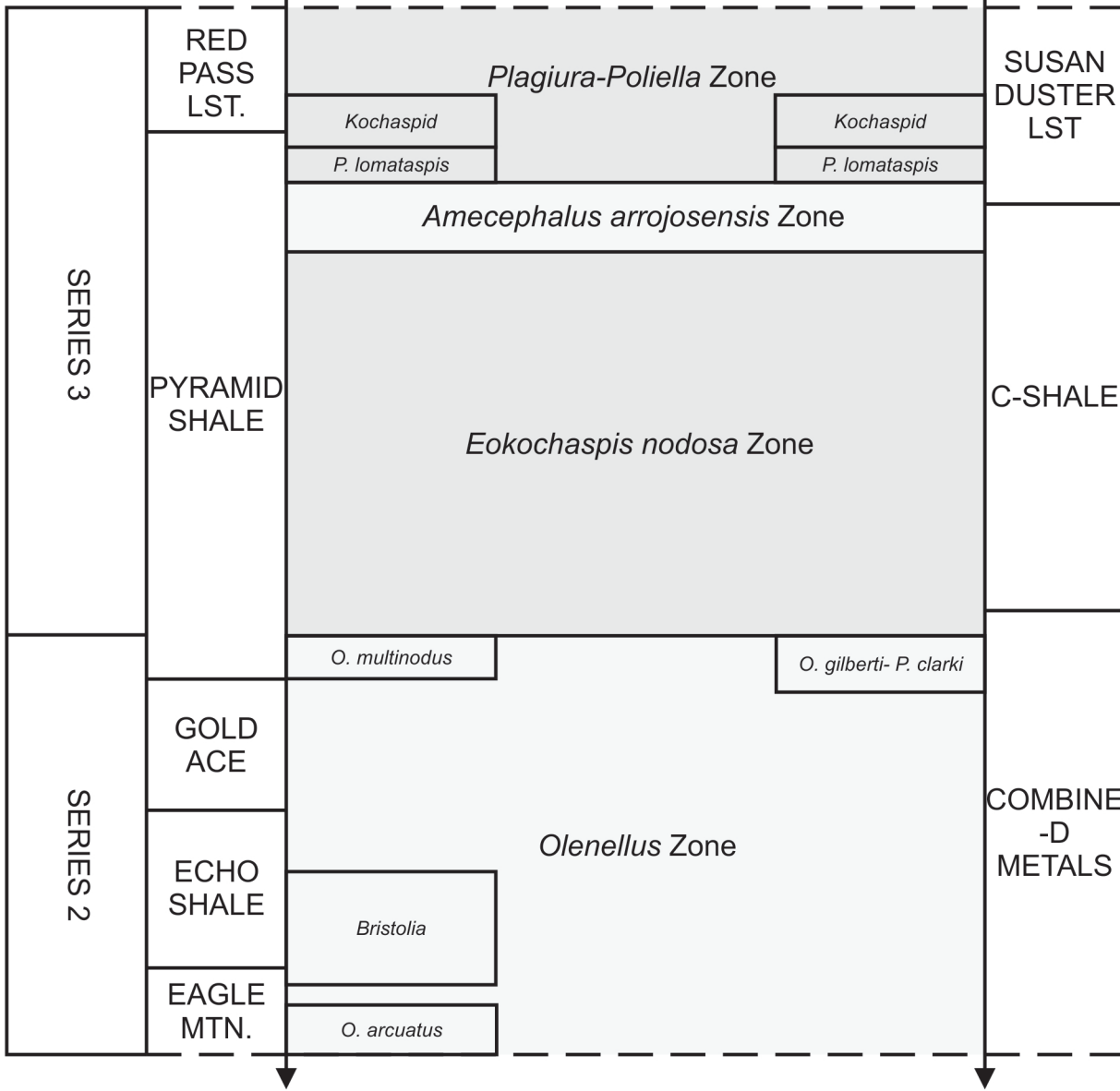
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(m)

CARRARA
FM

PIOCHE
FM



SERIES 3

SERIES 2

RED
PASS
LST.

PYRAMID
SHALES

GOLD
ACE

ECHO
SHALES

EAGLE
MTN.

SUSAN
DUSTER
LST

C-SHALES

COMBINE
-D
METALS

Plagiura-Poliella Zone

Kochaspis

Kochaspis

P. lomataspis

P. lomataspis

Amecephalus arrojosensis Zone

Eokochaspis nodosa Zone

O. multinodus

O. gilberti- P. clarki

Olenellus Zone

Bristolia

O. arcuatus

CAMBRIAN
SERIES

MEMBER

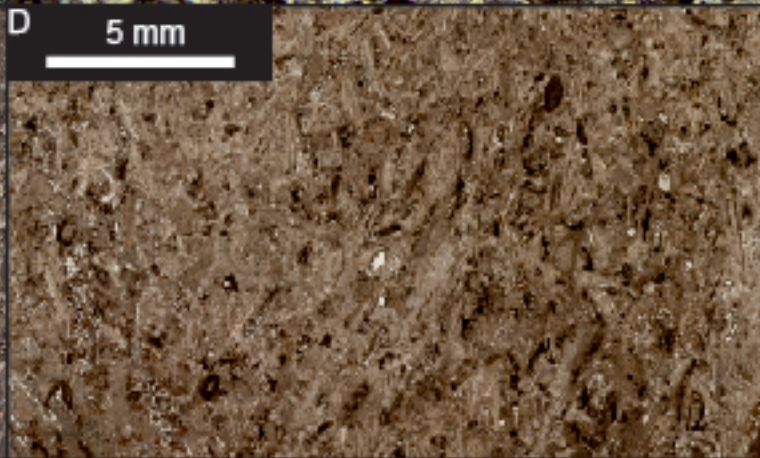
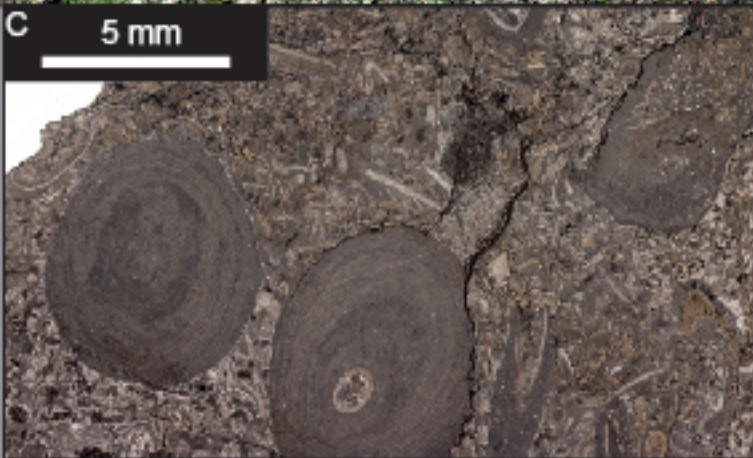
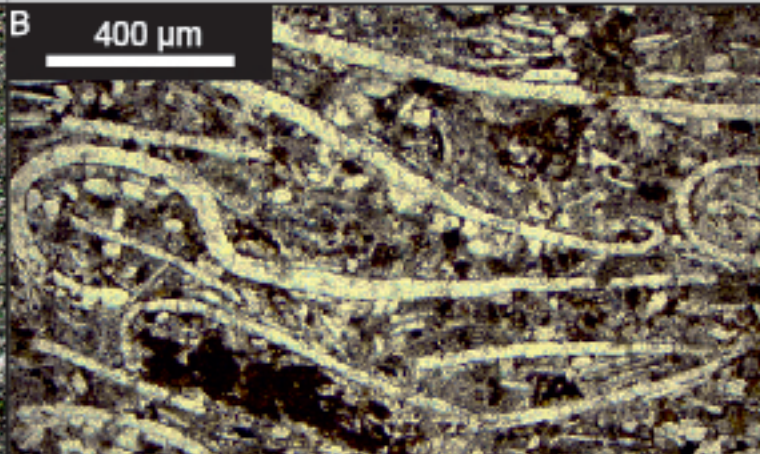
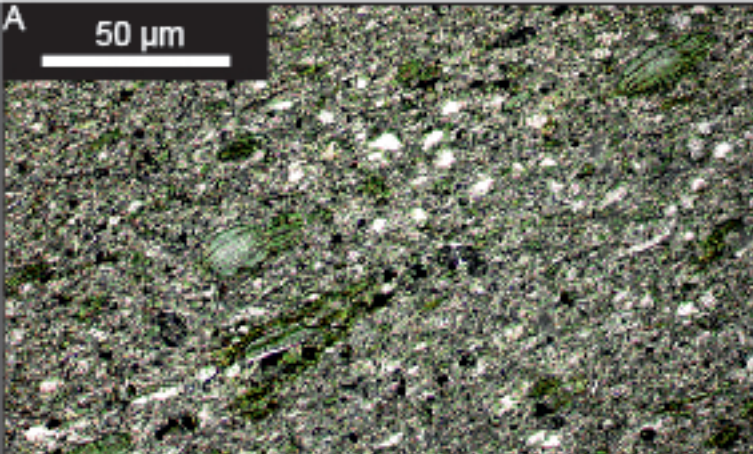
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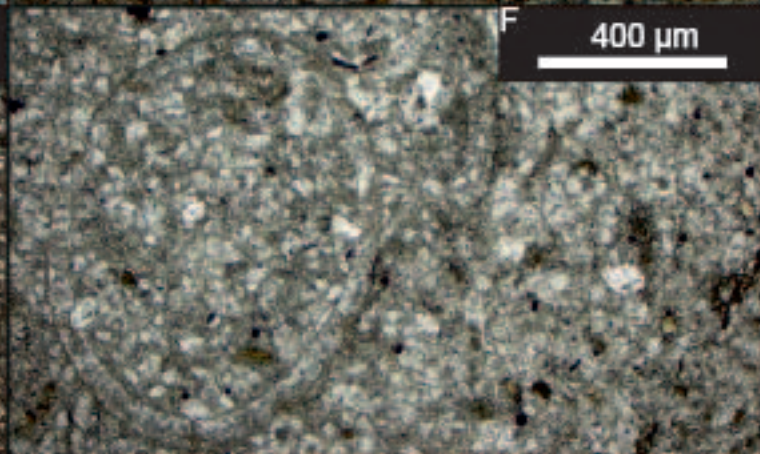
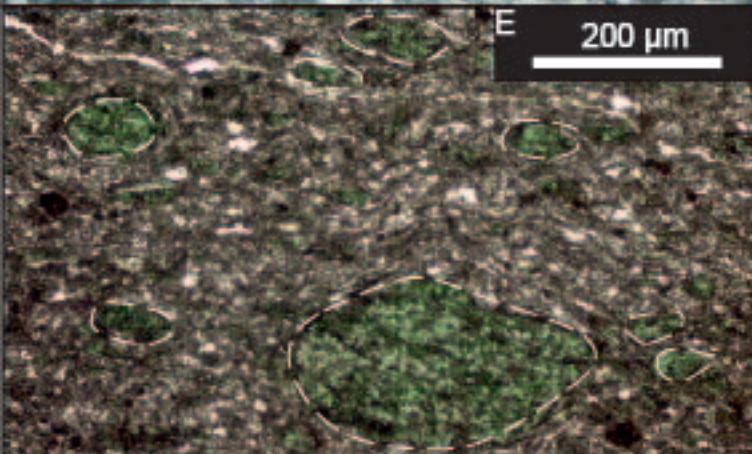
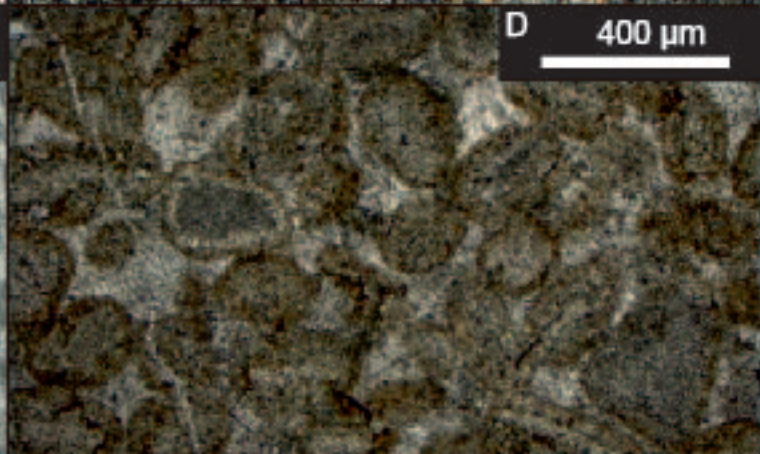
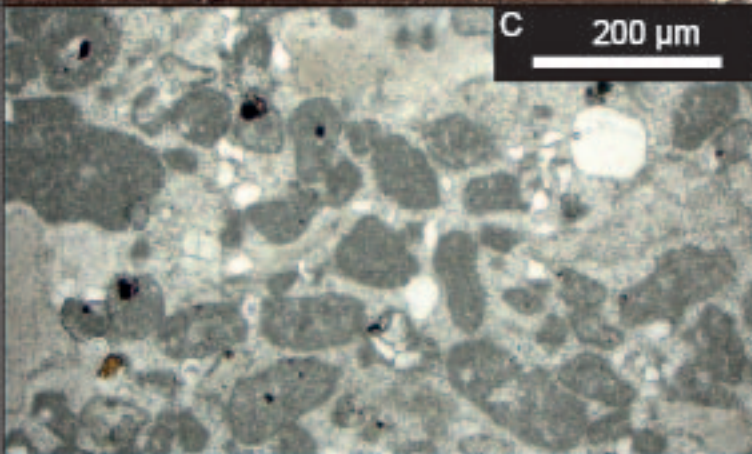
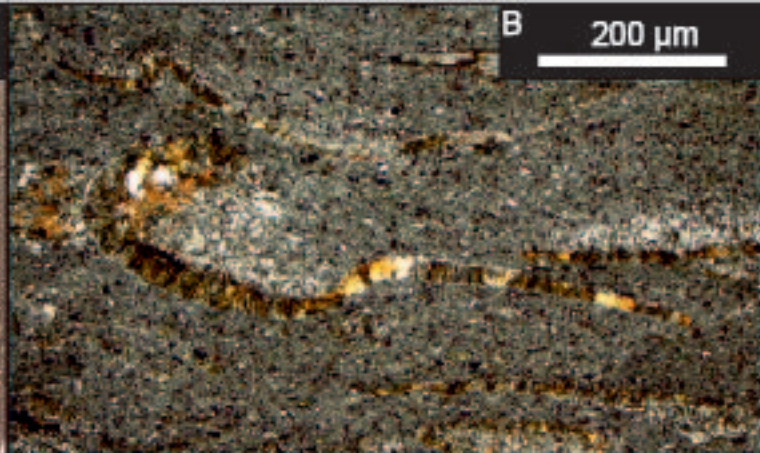
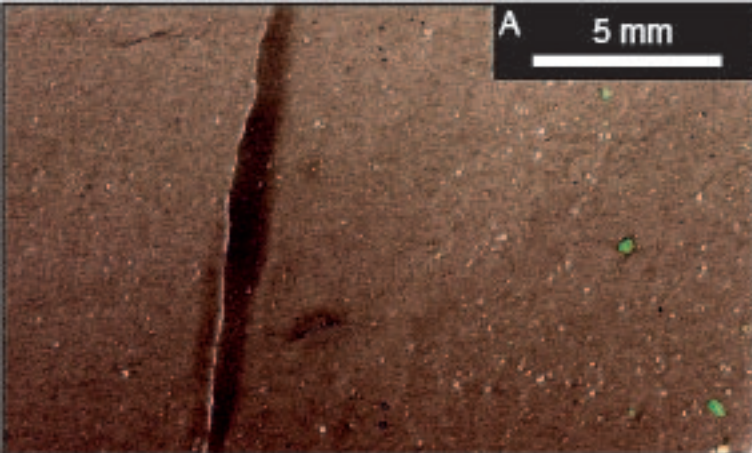
LAURENTIAN
TRILOBITE ZONE

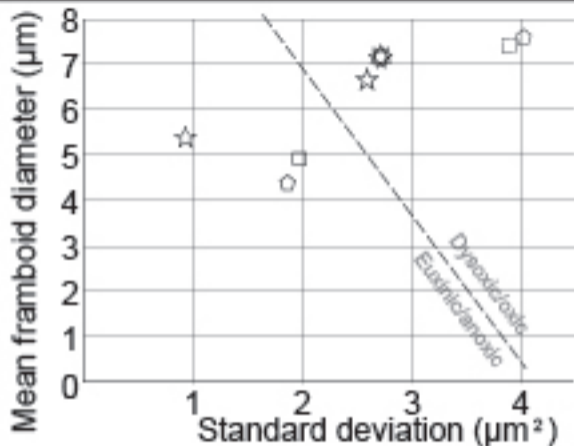
“ZONULE”

MEMBER



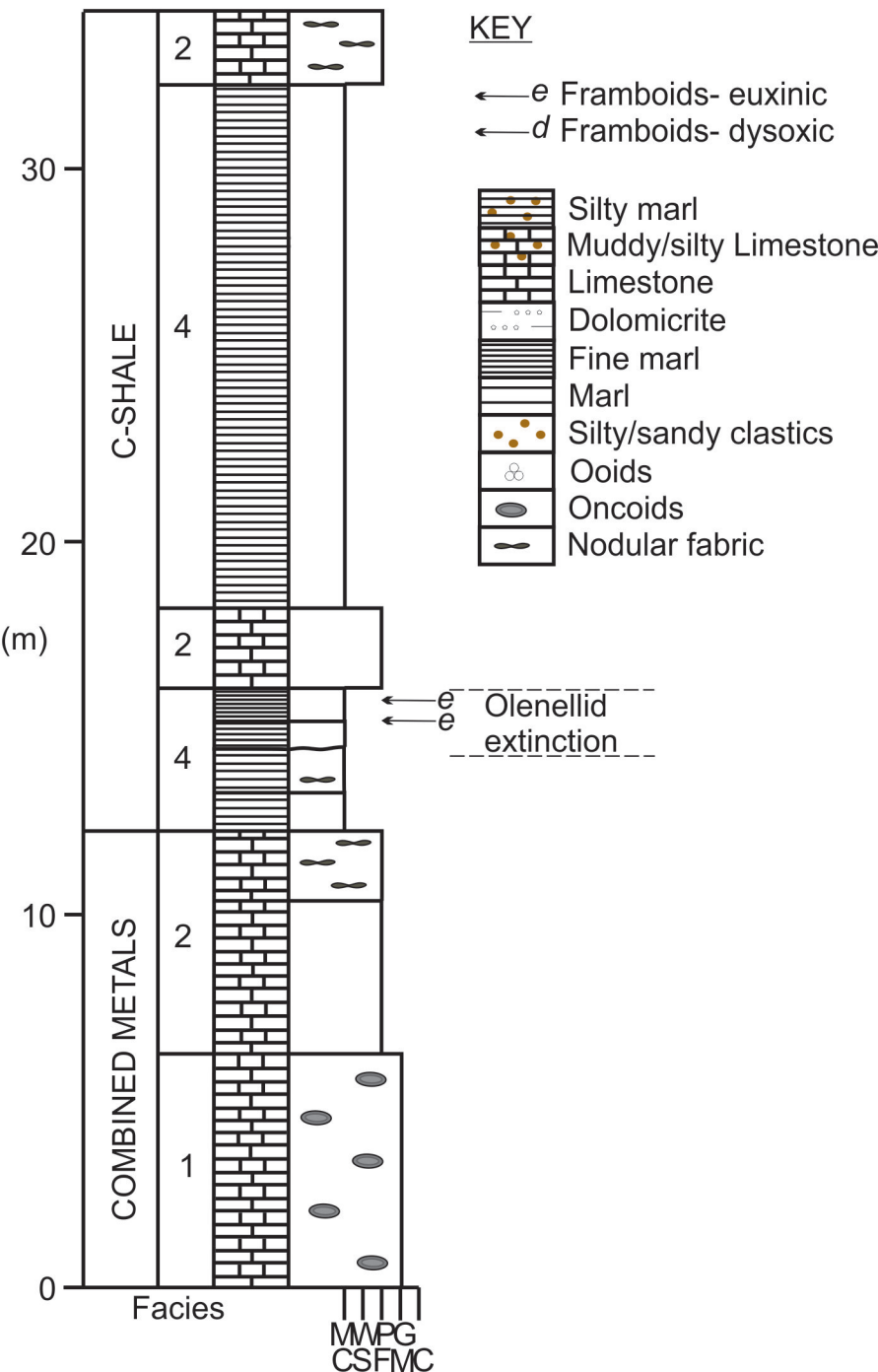




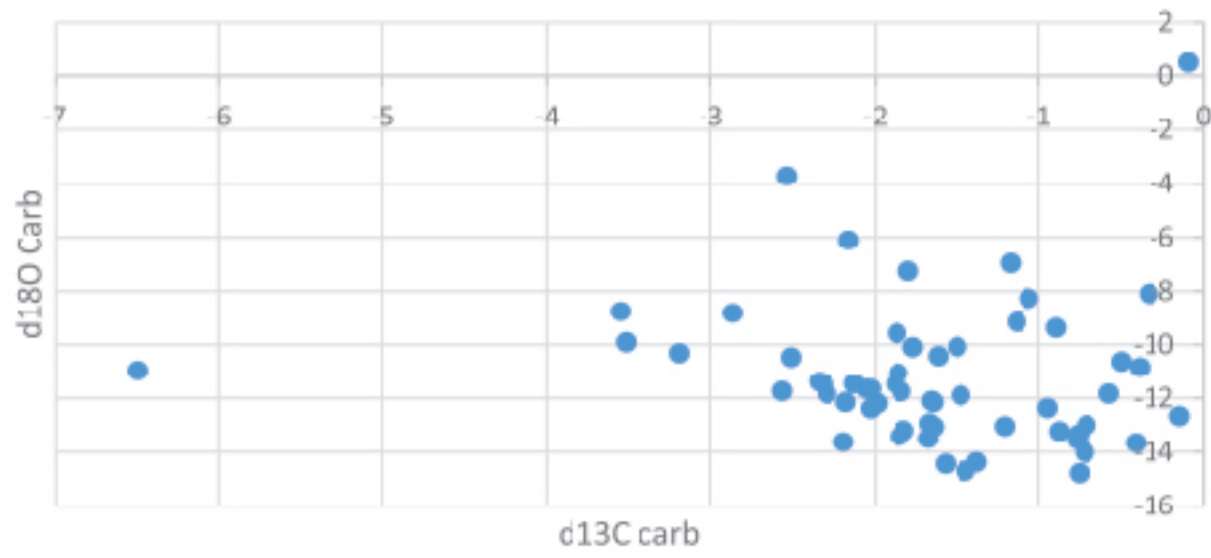


- ⚙ ☆ □ ○ Combined Metals Mbr, Oak Springs Summit
- ☆ C-Shale Mbr, Oak Springs Summit
- ○ C-Shale Mbr, Ruin Wash

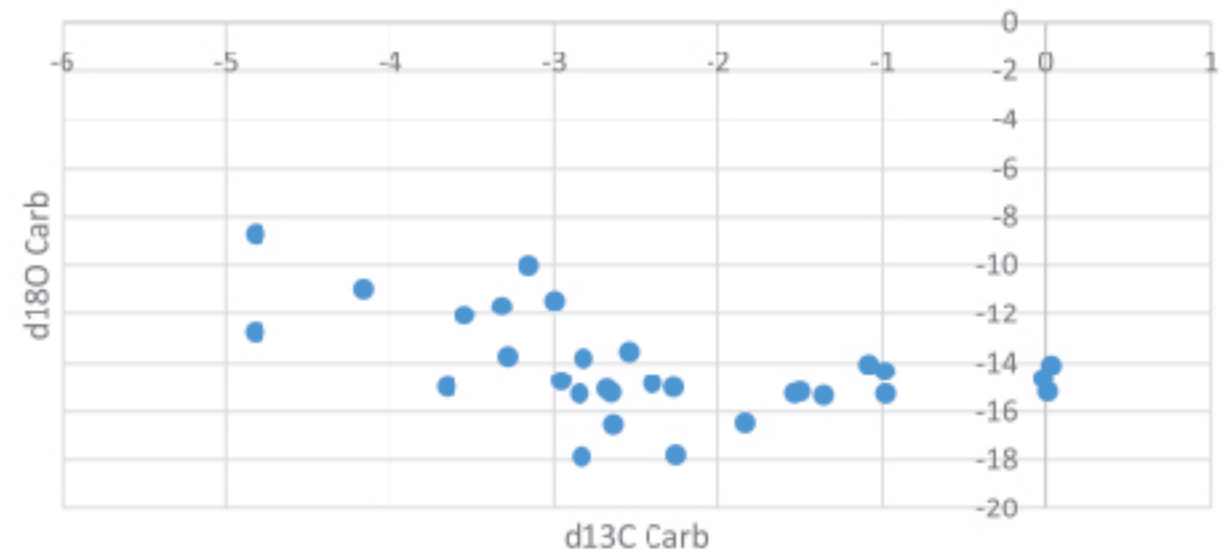
PIOCHE FM,
 RUIN WASH,
 NV



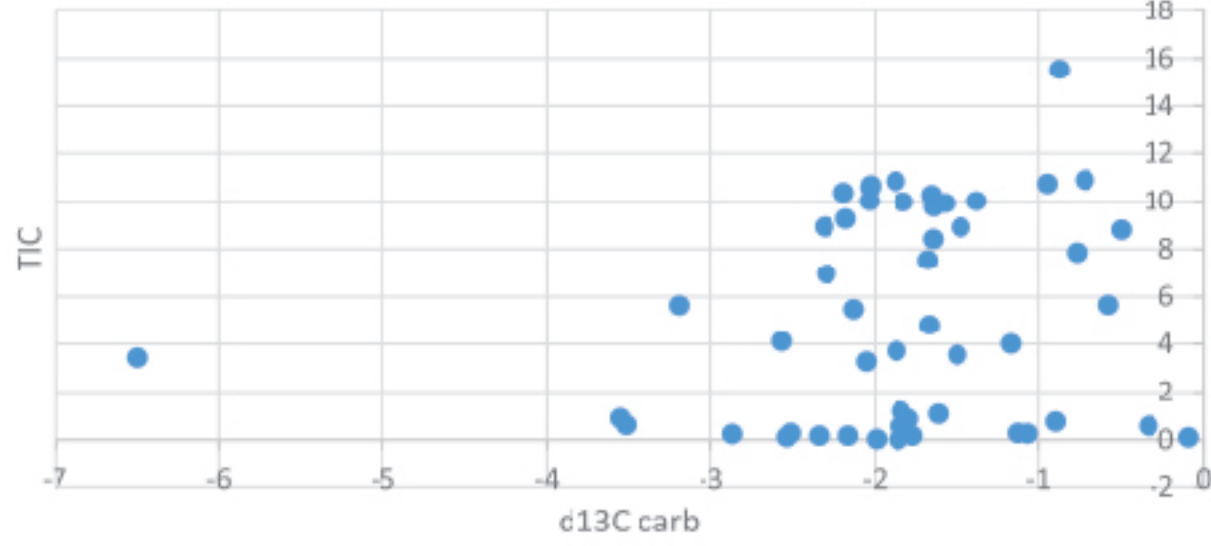
2 Carrara Formation



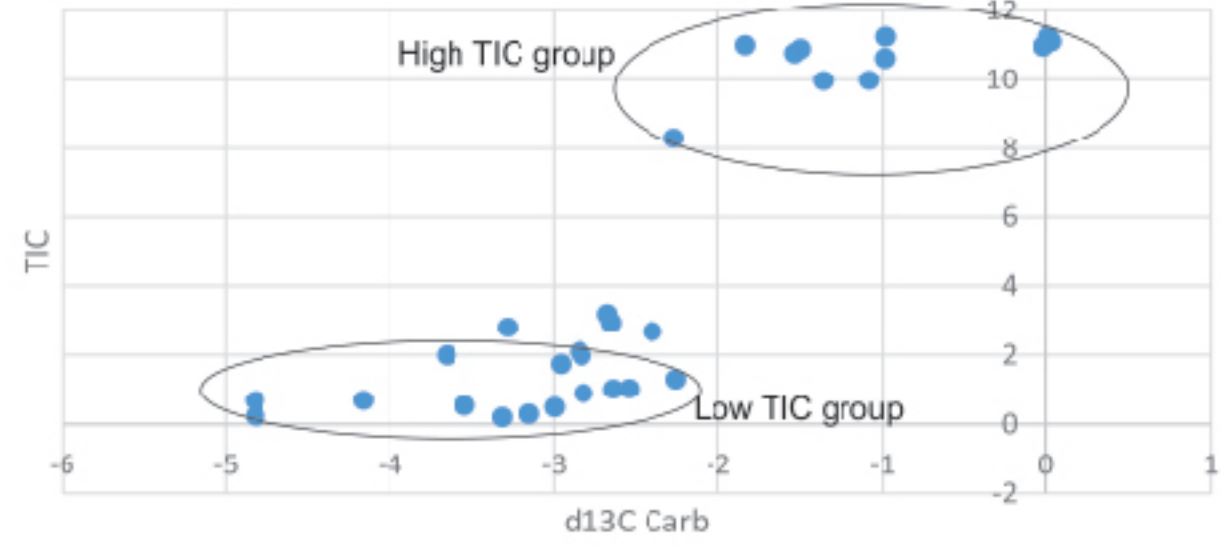
3 Pioche Formation



4 Carrara Formation



5 Pioche Formation



6 Pioche Formation

