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1   Latest Triassic onset of the Central Atlantic Magmatic Province (CAMP) volcanism in

2   the Fundy Basin (Nova Scotia): new stratigraphic constraints

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24   **Abstract**

25   In this paper we investigate the stratigraphic relationship between the emplacement of the

26   CAMP basalts and the Triassic-Jurassic (Tr-J) boundary in the Fundy basin (Nova Scotia,

Canada), one of the best exposed synrift basins of eastern North America (ENA) formed as a consequence of the rifting that led to the formation of the Atlantic Ocean. The Triassic palynological assemblages found in the sedimentary rocks below (uppermost Blomindon Formation) and just above the North Mountain Basalt (Scots Bay Member) indicate that CAMP volcanism, at least in Nova Scotia, is entirely of Triassic age, occurred in a very short time span, and may have triggered the T-J boundary biotic and environmental crisis. The palynological assemblage from the Blomindon Formation is characterised by the dominance of the Circumpolles group (i.e. *Gliscopollis meyeriana, Corollina murphyae, Classopollis torosus*) which crosses the previously established Tr-J boundary. The Triassic species *Patinasporites densus* disappears several centimetres before the base of the North Mountain basalt, near the previously interpreted Tr-J boundary. The lower strata of the Scots Bay Member yielded a palynological assemblage dominated by Triassic bisaccate pollens (e.g *Lunatisporites acutus, L. rhaeticus Lueckisporites* sp., *Alisporites parvus*) with minor specimens of the Circumpolles group. Examination of the state of preservation and thermal alteration of organic matter associated with the microfloral assemblages precludes the possibility of recycling of the Triassic sporomorphs from the older strata. Our data argue against the previously definition of the Tr-J boundary in the ENA basins, based mainly on the last occurrence of *P. densus*. Consequently, it follows that the late Triassic magnetostratigraphic correlations should be revised considering that chron E23r, which is correlated with the last occurrence of *P. densus* in the Newark basin, does not occur at the Tr-J boundary but marks rather a late Triassic (probably Rhaetian) reversal.

52

1. Introduction

The Triassic-Jurassic (Tr-J) boundary, most recently dated by U/Pb 201.6 +/- 0.3 Ma (Schaltegger et al., 2008) is characterized by the well known end-Triassic extinction, (Sepkoski, 1996; Hallam, 2002; Tanner et al., 2004). According to several authors, the severity of the Tr-J event at the genus and family level is higher than for the end-Cretaceous extinction, and second only to the end-Permian extinction. Other authors, however, have documented that this extinction is characterized by a progressive, and possibly punctuated, reduction of diversity from the Late Triassic to the Early Jurassic (Hallam, 2002; Tanner et al., 2004; Lucas and Tanner, 2007). Notably, the record of turnover among continental flora remains unclear (Hallam, 2002). For example, in most of the European domain, the Tr-J boundary seems to be characterized by only a minor extinction/turnover of the macroflora and microflora, against a background of more gradual change (e.g. Schuurman, 1979; Hallam, 2002; Warrington, 2002; Tanner et al., 2004; Lucas and Tanner, 2007, 2008; Traverse, 2008). In contrast, an abrupt floral change has been documented for eastern North America (ENA) and Greenland (Fowell and Olsen, 1993; Fowell et al., 1994; Fowell and Traverse, 1995; McElwain et al., 1999; 2007; Olsen et al., 2002; Whiteside et al., 2007). A marked negative carbon isotope ($\delta^{13}C$) anomaly in both carbonate and sedimentary organic matter (OM) observed in marine and terrestrial Tr-J boundary strata, e.g. from Hungary, Canada, Spain, Italy, Austria, and England (Pàlfy et al., 2001; Hesselbo et al., 2002; Guex et al., 2004; Ward et al., 2004; Gomez et al., 2007; Kuerschner et al., 2007; Williford et al., 2007). Additionally, an apparently synchronous decrease of the stomatal index of fossil leaves (McElwain et al., 1999) has suggested a disruption of the global carbon cycle, potentially involving some combination of global warming, productivity decline and methane

77 hydrate release (Pàlfy et al., 2001; Tanner et al., 2004; Ward et al., 2004; Lucas and

78 Tanner, 2008).

79   In summary, there is a general consensus that a biotic and environmental

80 perturbation occurred during the latest Triassic and spanned the Tr-J boundary (Tanner et

81 al., 2004; Lucas and Tanner, 2008). Some of the forcing mechanisms that have been

82 invoked to explain this disruption of the ecosystem include: (1) rapid sea-level change

83 and/or anoxia (Hallam and Wignall, 1999); (2) bolide impact (Olsen et al., 2002; Ward et

84 al., 2004); and (3) atmospheric loading by CO_2 and SO_2 related to eruptions of the

85 Central Atlantic Magmatic Province (CAMP), which potentially caused atmospheric

86 warming of up to 3-4 °C or acidic atmospheric pollution (Marzoli et al., 1999, 2004;

87 McElwain et al., 1999; Hesselbo et al., 2002; Guex et al., 2004; Knight et al, 2004;

88 Cohen and Coe, 2007; Tanner et al., 2007; Schaltegger et al., 2008; Van de Schootbrugge

89 et al., 2008).

90   Emplacement of the voluminous tholeiitic dikes, sills, and flood basalts of CAMP

91 occurred in North and South America, Africa and Europe over a total surface area

92 potentially in excess of 10 million km^2 (Marzoli et al. 1999, 2004; Olsen et al., 2003;

93 Verati et al., 2007) (Fig.1). The available $^{40}Ar/^{39}Ar$ plateau ages for the basalts suggest

94 brief durations (< 1 Ma) of peak eruptive activity at specific locations (Deckart et al.,

95 1997; Marzoli et al., 1999; 2004; Knight et al., 2004; Nomade et al., 2007; Verati et al.,

96 2007), although the main interval of emplacement across the entire CAMP may have had

97 a duration of 2 ± 1 Ma centered at ca. 199 Ma (Jourdan et al., in press). These $^{40}Ar/^{39}Ar$

98 ages conform to the U-Pb age of the Tr-J boundary (201.6 ± 0.3 Ma, single zircon;

99 Schaltegger et al., 2008) considering the well established ca. 1% bias between the two

100 isotopic dating methods (Min et al., 2000; Kuiper et al., 2008). However, taking into

101 account analytical and intercalibration uncertainties, as well as the short duration of the

102 eruptive events, the published radio-isotopic ages indicate proximity, but cannot

103 ultimately define the relative timing of CAMP volcanism and the Tr-J boundary crisis.

104 Therefore, correlation of the biotic turnover from the marine realm, where the boundary

105 is defined by ammonites (Guex et al., 2004) with the continental CAMP volcanism is

106 required. Various analytical methods have been applied, including palynology,

107 magnetostratigraphy and geochemistry. However, multidisciplinary studies applied to the

108 North American and African Tr-J sediments and interlayered CAMP basaltic flows have

109 produced contrasting interpretations.

110   Since the 1970s the palynological Tr-J boundary in the ENA continental rift-

111 basins has been placed in the strata immediately beneath the lowest CAMP basalt flow

112 (e.g. the Orange Mt. Basalt in the Newark basin). Thus, all the strata overlying and

113 interlayered with the CAMP lava flows were assigned to the Jurassic. This boundary is

114 defined, according to both older and more recent studies (e.g. Cornet and Traverse, 1975;

115 Cornet and Olsen, 1985; Fowell and Olsen, 1993; Fowell et al., 1994; Fowell and

116 Traverse, 1995; Whiteside et al., 2007), by a marked palynological turnover event

117 recognized on the basis of the following: (1) some last appearances of certain species

118 (*Ovalipollis ovalis*, *Vallasporites ignacii*, and *Patinasporites densus*); (2) an increase in

119 *Corollina* spp. (= *Classopollis* and *Gliscopollis*) percentage; (3) and a bloom of trilete

120 spores (fern spike), considered as the expression of a renewed palynoflora after the Tr-J

121 mass extinction. These palynostratigraphic criteria for placing the Tr-J boundary have

122 been criticised (Gradstein et al., 1994; van Veen, 1995; Lucas and Tanner, 2007) because

123 locally (e.g., the European domain) the disappearance of the vesicate forms (i.e. *V.*

124 *ignacii*, and *P. densus)* took place during the late Norian and occasionally early Rhaetian

125 (e.g., Morbey, 1978; Schuurman, 1979; Krystyn et al., 2007; Kuerschner et al, 2007).

126 Notably, Kozur and Weems (2005), relying on conchostracan assemblages from the

127   Culpeper and Newark basins, concluded that the Tr-J boundary in the Newark

128   Supergroup must occur above the lowest CAMP basalt.

129   In the Newark and Fundy basins, the putative Tr-J boundary is also characterized

130   by a moderate Ir anomaly, the origin of which is disputed (Olsen et al., 2002; Tanner and

131   Kyte, 2005; Tanner et al., 2008); more importantly, the Ir anomaly is immediately

132   preceded by a magnetic reversal (chron E23r; Olsen et al., 2002). Indeed, the most

133   complete Late Triassic to earliest Jurassic magnetic polarity time scale has been

134   calibrated on the non-marine section of the Newark basin (e.g. Kent and Olsen, 1999),

135   which is anchored to the palynologically defined Tr-J boundary (see Lucas and Tanner,

136   2007 for an historical review), and thus represents a potentially powerful tool for global

137   correlations. The E23r reversal has been correlated by Whiteside et al. (2007) to one of

138   the brief reversals (chrons SA5n.3r and SA5r) of the marine St. Audrie's Bay section

139   (England), which corresponds to the Tr-J boundary in the upper part of the Penarth Group

140   (Hounslow et al., 2004). Recently, Gallet et al. (2007) noticed discrepancies between the

141   Newark magnetostratigraphy and that of the Tethyan realm (e.g., the Oyuklu section of

142   south-western Turkey), which suggests either a higher position of the Norian-Rhaetian

143   boundary and/or a lack of the lowermost part of the Rhaetian in the Newark sequence.

144   The cyclostratigraphy of the Newark basin suggests that the onset of CAMP

145   volcanism postdated the Tr-J boundary by 20-40 ky and that the entire basalt pile was

146   erupted in the Early Jurassic during normal polarity chron E24n (Kent and Olsen, 1999,

147   2008; Olsen et al., 2003). The next reversal, chron E24r, occurred about 1.6 Ma after the

148   supposed Tr-J boundary, and above the youngest CAMP lava flow (Hampden Basalt), as

149   recorded in the Hartford basin (Kent and Olsen, 2008). This reversal tentatively has been

150   correlated with reversals detected in Jurassic strata from the Paris basin, France (Yang et

151   al., 1996).

152   In contrast to the interpretation that all CAMP volcanism postdated the Tr-J

153   boundary, multidisciplinary data suggest a latest Triassic onset of the volcanism in

154   Morocco (Knight et al., 2004; Marzoli et al., 2004). This conclusion is based on: (1)

155   evidence that the first lava flows erupted were synchronous with sediments bearing a

156   Triassic palynological assemblage dominated by *P. densus, C. murphyae, G. meyeriana*

157   and *C. torosus;* (2) the presence of a paleomagnetic reversal within the lava pile (i.e.

158   approximately halfway between the base and top) that has been correlated with the Tr-J

159   boundary reversal from the marine St. Audrie's Bay section, and possibly with chron

160   E23r from the Newark basin; (3) geochemical correlations of the basalts between basins,

161   suggesting that CAMP volcanism started slightly earlier in Morocco than in the Newark

162   basin. However, the data for the Moroccan CAMP have been questioned (Whiteside et

163   al., 2007, 2008; *contra* Marzoli et al., 2008), which means that the relationship between

164   the age of CAMP volcanism and the Tr-J boundary remains an unresolved question. This

165   problem mainly results from the absence of a generally accepted chronostratigraphic

166   definition of the Tr-J boundary, the disputed biostratigraphic value of the palynological

167   assemblages, and the lack of a well constrained age for the sediments interlayered with or

168   covering the CAMP lava flows.

169

170   **2. Study Area**

171   In this paper we investigate the relationship between the emplacement of the CAMP

172   basalts and the Tr-J boundary in the Fundy basin (Nova Scotia, Canada), one of the best

173   exposed ENA synrift basins,,. These basins, collectively referred to as the Newark

174   Supergroup, are more or less continuously exposed on the east coast of North America,

175   from Nova Scotia to South Carolina (Tollo and Gottfried, 1992; Olsen and Schlische,

176   1990; Olsen et al., 2003; Whiteside et al., 2007). Formational nomenclature and the

177   group–level stratigraphy used in this paper are after Weems and Olsen (1997) (Fig.2).

178   The Fundy basin, which was located at 24° N palaeolatitude at the close of the

179   Triassic (Kent and Tauxe, 2005), is the northernmost of the Mesozoic rift basins of ENA

180   (Fig.3d), and its sedimentary and volcanic sequence is well exposed on the shores of the

181   Bay of Fundy in western Nova Scotia and eastern New Brunswick. The Fundy Basin,

182   including the Minas, Fundy, and Chignecto structural sub-basins (Olsen and Schlische,

183   1990), consists of a thick siliciclastic succession, of which about 1 km of fluvial,

184   lacustrine and eolian sediments crops out in Nova Scotia and New Brunswick, and 4 km

185   are submerged beneath the Bay of Fundy (Olsen et al., 2005). The Blomidon Formation

186   of the Fundy Group, which is assigned to the middle-late Norian and possibly Rhaetian

187   on the basis of magneto-cyclostratigraphy (Kent and Olsen, 2000; Olsen et al., 2005),

188   comprises 200-300 m of cyclically interbedded sandstone and mudstone of mostly

189   lacustrine, playa, eolian and fluvial origin deposited during an interval of semi-arid to

190   arid climate (Tanner, 2000; Tanner and Kyte, 2005). The uppermost Blomidon Formation

191   (Partridge Island Member) contains the turnover event cited by Fowell and Traverse

192   (1995) as the Tr-J palynological boundary.

193   The Blomidon Formation is overlain by the tholeiitic North Mountain Basalt,

194   which is overlain in turn by the McCoy Brook Formation. The North Mountain Basalt

195   (about 400 m thick) crops out on the north and south shores of the Minas Basin, along the

196   western coast of Nova Scotia and on Grand Manan Island (New Brunswick). It is

197   subdivided into three lava flow units, which are distinguished in the field by

198   volcanological features (Papezik et al., 1988; Kontak, 2008). The lowermost unit, the

199   East Ferry Member, is composed of a single massive basalt flow up to 180 m thick,

200   which in its upper part locally contains sheets (≤1-2 m) of coarse-grained mafic

201 pegmatite and granophyre. The middle unit, the Margaretsville Member, is formed by up

202 to 16 individual inflated, pahoehoe-type flows, up to 10 m thick, that are heavily

203 vesiculated and altered. The uppermost unit, the Brier Island Member, consists of one to

204 two massive basalt lava flows that are similar to the lowermost flow unit. Significant

205 sedimentary intercalations are absent between the lava units or flows, although flow tops

206 commonly are oxidized, suggesting only brief hiatuses between volcanic events. U/Pb

207 and $^{40}Ar/^{39}Ar$ ages are available for the North Mountain Basalt. A single-zircon U/Pb age

208 of 201.3 +/-0.3 Ma was obtained on a granophyric matrix in a mafic pegmatite sheet in

209 the upper part of the East Ferry Member (Schoene et al., 2006). $^{40}Ar/^{39}Ar$ plateau ages on

210 plagioclase from lower and upper unit lava flows range from 198.6 ± 1.1 to 201.0 ± 1.4

211 Ma (Jourdan et al., in press), supporting a duration of the peak eruptive event on the order

212 of, or shorter than ~1 Ma. These ages overlap with the U/Pb age of Schoene et al. (2006),

213 and with the U/Pb age of the Tr-J boundary (Schaltegger et al., 2008), if analytical and

214 intercalibration uncertainties are considered. Furthermore, the ages are indistinguishable

215 from high-quality isotopic ages for basalts from other areas of the northern CAMP (i.e,

216 Morocco and the U.S.A. Knight et al., 2004; Marzoli et al., 2004; Nomade et al., 2007;

217 Verati et al., 2007). The North Mountain Basalt flows show broad geochemical

218 similarities with other CAMP basalts, but in detail the geochemical correlation with those

219 flows is not straightforward (Marzoli et al., 2008).

220   The McCoy Brook Formation is the youngest unit in the Fundy rift basin and is

221 mainly composed of red beds and abundant sand-rich fluvial deposits, gypsiferous sand

222 patch playa/lacustrine cycles and local eolian sandstones (De Wet and Hubert, 1989;

223 Tanner and Hubert, 1992; Tanner, 2000; Olsen et al., 2005). The basal member, the Scots

224 Bay Member, generally lies directly on the North Mountain basalt.

225

226 **3. Previous Work**

227   The entire sedimentary succession of the Fundy Basin has been dated mainly on

228 the basis of palynology (Bujak, 1977; Fowell and Traverse, 1995; Whiteside et al., 2007),

229 vertebrate fossils and cyclostratigraphic correlations with the Newark Supergroup (Olsen

230 et al., 2005 for references). Bujak (1977) located the Tr-J boundary within the McCoy

231 Brook Formation based on palynological data from the Chinampas N-37 well (Bay of

232 Fundy). Subsequently, new palynological data from the Minas sub-basin (at Partridge

233 Island) resulted in placement of the Tr-J boundary in the uppermost meter of the

234 Blomidon Formation (Fowell and Traverse, 1995; Whiteside et al., 2007), where the

235 palynological assemblage exhibits a transition from dominantly *Corollina torosa* and *P.*

236 *densus* to mainly *Corollina* spp. (*C. meyeriana, C. torosa, C. simplex, C. murphyae)*.

237 Whiteside et al. (2007), on the basis of new palynological data, placed the Tr-J boundary

238 at less than 20 cm below the contact with the North Mountain Basalt (i.e. near the top of

239 the Blomidon Formation), between the bed that contains the highest occurrence of *P.*

240 *densus,* and the overlying sample, which lacks *P. densus* or abundant bisaccates. The

241 stratigraphic position of this presumed turnover is similar to that observed in the Newark

242 and Hartford basins (Fowell et al., 1994; Olsen et al., 2002).

243   As in the Newark basin (Kent and Olsen, 2000; Olsen et al., 2002), a moderate Ir

244 (maximum measured value > 400 pg/g) and PGE (Platinum Group elements) enrichment

245 occurs in multiple grey-coloured mudstone layers within the uppermost meter of the

246 Blomidon Formation in the Partridge Island section, which is roughly coincident with the

247 presumed palynological transition (Tanner and Kyte 2005; Tanner et al., 2008). In

248 contrast to the Newark basin, no magnetic reversal (equivalent to E23r) and no fern spike

249 (Kent and Olsen, 2000; Olsen et al., 2002), have been found in this section. The spike in

250 fern spores would be produced by *Clathropteris meniscoides,* which occurs in well-

251   preserved specimens in some localities of the Hartford basin (Cornet and Traverse, 1975).

252   In the Fundy basin, the only recorded *Clathropteris* has been found in the Scots Bay

253   Member, while the species *C. meniscoides* has not been reported (Olsen et al., 2005). The

254   McCoy Brook Formation contains a large variety of fossils which are mostly considered

255   to be of early Jurassic age (De Wet and Hubert, 1989; Shubin et al., 1994; Olsen et al.,

256   2005). Lucas and Tanner (2007) suggested that the Tr-J boundary in the Fundy Basin

257   occurs within the McCoy Brook Formation, based on the presence of only one CAMP

258   extrusive unit in the Fundy basin (compared to the three units present in the Newark and

259   Hartford basins).

260

261   **4. Research methods**

262   4.1 Lithostratigraphy of the sampled section

263   The section sampled at Partridge Island is a cliff face partially covered by

264   colluvium (Fig. 3a). The uppermost 1.0 m of the Blomidon Formation was exposed in a

265   trench and sampled at 5-10 cm intervals (Fig. 4a, 6). The lower part of the exposed

266   section consists of interbedded red and grey mudstones. The red mudstone is thinly to

267   medium bedded (2–15 cm thick) and composed mostly of hematite-stained clays and silt

268   to sand-size immature arenite grains. The coarser components consist mostly of quartz

269   and minor micas and feldspars. Grey mudstones are mostly thin-bedded (2–5 cm in

270   thickness) and contain a higher proportion of sand-size grains, including quartz and rock

271   fragments (igneous and metamorphic), intercalated with organic-rich dark clay beds.

272   Finely laminated mudstone containing dark organic-rich laminae interlayered with grey

273   to reddish laminae are also present in the lower part of the sampled section. The clay

274   fraction of the mudstones comprises mostly illite, smectite and chlorite, with lesser

275   amounts of kaolinite and mixed-layer clays and an abundance of chlorite in a greenish-

276 grey mudstone 20 cm below the contact with the base of the North Mountain Basalt, the

277 East Ferry Member (Tanner and Kyte, 2005).

278   At Scots Bay, the outcropping strata of the Scots Bay Member are in direct

279 stratigraphic contact with the Margaretsville Member of the North Mountain Basalt (the

280 Brier Island Member is not present here; Fig. 3b,c). The Scots Bay strata here consist of a

281 basal carbonate-rich sequence of two white, green, purple, and red lacustrine cycles. The

282 basal portion contains coarse immature sandstone with ostracod- and peloid-bearing

283 carbonate matrix. Clastic grains mostly consist of quartz, mottled carbonate mudstone,

284 siltstone and volcanic debris.

285

286 4.2 Sampling and analysis

287   Twelve samples were selected for organic matter (OM) analysis. Of these

288 samples, nine are from the uppermost Blomidon Formation beneath the North Mountain

289 Basalt (East Ferry Member) at the Partridge Island section (Fig.3a,d). The Tr-J boundary

290 identified at this location in earlier palynological studies (Fowell and Traverse, 1995;

291 Whiteside et al, 2007) was sampled in the interval represented by sample PI-30 (i.e., 30–

292 35 cm below the basalt Fig.4a). The three remaining samples are from the lower Scots

293 Bay Member, which overlies the North Mountain Basalt, and are from the section at

294 Scots Bay (Fig.3).These three samples were also analyzed for palynological assemblage

295 (Fig. 4, 5), palynofacies (Fig.6) and thermal maturity (see supporting online material, Fig.

296 S1, Fig. S2).

297

298 **5. Results**

299 5.1 Palynological assemblage

300   The palynological content of the sampled section at Partridge Island (Fig. 5) is

301 characterised by a dominance of Circumpolles group species, represented in order of

302 decreasing abundance by *G. meyeriana* (Fig. 5d,f)*, C. murphyae* (Fig. 5l,m)*, C. torosus*

303 (Fig. 5g, h, j, k), and *C. simplex* (in this paper, the generic name *Classopollis* is used

304 instead of *Corollina* or *Circulina*, as recently formally proposed by Traverse (2004), for

305 *Classopollis torosus*). The total abundance of the Circumpolles group varies slightly

306 through the section (Fig.4a). Generally, a lower abundance occurs in the most oxidized

307 beds (i.e. red mudstones), as demonstrated by the state of preservation of the palynofacies

308 (Fig. 6), which is mostly dominated by inertinite and strongly oxidized vitrinite. At these

309 low levels of preservation, degraded, oxidized, and badly preserved sporomorphs prevail

310 (i.e. PI-70, PI-5, PI-0). The low total content of Circumpolles also correlates with a low

311 total OM concentration, which is poorly preserved in oxidized beds.

312 As shown in the Fig.4a, Circumpolles cross the Tr-J boundary previously

313 established by previous workers (Fowell and Traverse, 1995; Whiteside et al., 2007)

314 without a marked decline in abundance. *P. densus* (Fig. 5a,b,c) is present from the base of

315 the sampled section at Partridge Island up to sample PI-45, which is near the presumed

316 Tr-J boundary. The percentage of this vesicate pollen is rather low when compared with

317 the other sporomorphs, comprising at most 10%-15% of the association. Among

318 bisaccates, *Alisporites parvus* (Fig. 5p, r)*, A. tenuicorpis, Ovalipollis septimus* and other

319 undetermined bisaccates are generally quite common, particularly *Alisporites*. The

320 bisaccates are absent in the 20-cm interval below the basalt, but are again present in the

321 Scots Bay Member above the North Mountain Basalt. Less common are *Calamospora*

322 *mesozoica (*Fig. 5n)*, Conbaculatisporites mesozoicus* (Fig. 5e), *Porcellispora*

323 *longdonensis, Todisporites rotundiformis, Kraeuselisporites sp., Carnisporites spiniger*

324 (Fig. 5q) and *Cycadopites* sp. (Fig. 5i), all of which are present in only a few samples.

325 Noteworthy is the presence of a group of trilete spores, such as *Converrucosisporites*

326   *cameroni* (Fig. 5o)*, Dictyophyllidites harrisii* and *Dictyophyllidites* sp., and other trilete

327   spores below the presumed Tr-J boundary. In the other sections of the ENA basins, these

328   fern spores characterize the fern spike just above the palynological boundary that has

329   been interpreted as the floral turnover after mass-extinction.

330   Three samples of the Scots Bay Member (Fig. 4b) collected about 1 meter above

331   the top of the Margaretsville member basalt flow (the Brier Island Member is not present

332   here) proved to be palynologically productive. Sample SB2 yielded the best preserved

333   and most significant assemblage. The three samples yielded a palynological assemblage

334   dominated by bisaccate pollens, e.g *Lunatisporites acutus* (Fig.5w), *L. rhaeticus* (Fig. 5t)*,

335   Lueckisporites sp., A. parvus* (Fig.5s)*, Klausipollenites sp., Platysaccus sp.* (Fig. 5u) and

336   other bisaccates in association with *C. mesozoica* and minor specimens of *G. meyeriana*,

337   *C. torosus* and other circumpolloid (Fig. 5x). In general, the Circumpolles group is less

338   abundant in the Scots Bay samples than in the Blomidon Formation samples.

339

340   5.2 Palynofacies

341   Palynofacies analysis addresses the potential for depositional and diagenetic

342   control of OM preservation in the vicinity of the previously interpreted Tr-J boundary;

343   i.e. rather than resulting from a catastrophic mass extinction event, is the observed

344   palynological turnover an artifact of preservation? The type and amount of organic debris

345   in the Blomidon Formation were measured quantitatively (Fig. 6). The lower total OM

346   content of the Scots Bay Member prevented a detailed quantitative analysis, although a

347   semi-quantitative estimation was possible. The palynofacies consist of terrestrial

348   elements with high concentrations of palynomacerals, including debris of higher plants

349   with minor amorphous organic matter (AOM). The type and amount of organic debris

350   show consistent variations through the Blomidon Formation (Fig. 6). Despite the

expected oxidizing conditions, the OM preservation and content are quite high, except in

several samples. The peak abundances of inertinite (Fig. 6a), occur in the lower and

upper parts of the sampled section (PI-10 and PI-70), coinciding with lower amounts of

AOM. Vitrinite, , varies slightly within the section, ranging from 25% to 50% of the OM

content.

Sporomorphs are present in varying abundances through the entire section, with

the exception of sample PI-10 (which was barren), a mottled limestone (Fig. 6b) just

above the previously interpreted Tr-J boundary. The sedimentary structures within this

latter sample may reflect bioturbation, a process that is typically detrimental to OM

preservation. Strongly degraded and oxidized sporomorphs occur in samples PI-70, PI-05

and PI-00; notably, the coarse and/or oxidized lithologic character of these samples was

not conducive for OM preservation (i.e., interlayered grey-reddish laminated mudstone,

grey coarser sandstone, and coarse siltstone at the contact with basalt, respectively). This

inference is supported by their high content of inertinite, which is the product of strong

degradation-oxidation processes. The highest abundance of sporomorphs occurs within a

dark grey mudstone (PI-45) (Fig. 6b), which was likely deposited in low-oxygen

conditions that favoured preservation of OM. The overall scarcity of AOM (Fig. 6c),

however, except in a few samples as noted, supports the interpretation of generally

strongly oxidizing conditions during deposition of the sedimentary section, as is typical

in many continental environments.

The three samples from the Scots Bay Member contain low concentrations of total

OM. The dominant palynomacerals are inertinite, and minor vitrinite. Several yellow to

dark orange cutinite debris are also present.

Variations in the palynofacies through the studied section are consistent with a

depositional environment that varied from playa/lacustrine to fluvial in the uppermost

376 Blomidon Formation, and perennial lacustrine in the basal Scots Bay Member. The

377 presence of abundant AOM within the darker intervals of the Blomindon Formation

378 indicates that rarely transient conditions of very temporary nature made higher OM

379 preservation possible.

380

381 5.3 Thermal alteration index (TAI)

382   Colour and degree of preservation of the OM were studies in order to determine

383 the thermal history of the sampled section and to evaluate the possibility that the

384 sporomorph content was recycled from older, thermally overmature strata (see online

385 material). Given the nature of the depositional environment of the studied succession

386 (i.e., fluvial-lacustrine), the potential for concentrating reworked palynomorphs from

387 older strata was considered high. Fortunately, the transported or reworked fraction of

388 OM can be distinguished by its opacity and darker colour (Traverse, 2008).

389   The Partridge Island section affords the opportunity to examine colour changes

390 over a relatively short stratigraphic/time interval, thus minimizing evolutionary and/or

391 environmentally driven changes in the sporomorph associations, and yielding an

392 internally consistent dataset. The TAI does not reveal a significant trend in colour

393 variation for most of the studied section (Fig. 6),. The colour of the sporomorphs ranges

394 from orange to brown down section, which corresponds to a TAI value of approximately

395 2+ to 3- (on a five point scale), suggesting a medium thermal maturity (about 60°-65° for

396 this type of OM). Closer to the section top (the basalt flow), from sample PI-05 to sample

397 PI-00, a modest increase occurs, with TAI values around 3 to 3+ (about 70°-80°). Thus,

398 sampling indicates a mild increase in thermal alteration that is easily attributed to thermal

399 overprint during the basalt emplacement.

The three samples of Scots Bay Member contain Triassic sporomorphs in which colour ranges from dark yellow to orange (TAI from 2 to 2+) (around 60°-55°). The fact that these sporomorphs are generally lighter in colour (i.e. less mature) than those in the Blomidon Formation strongly suggests that they were not recycled from the older Blomidon strata. (see online material, Fig.S2)

6. Discussion

The hypothesis that CAMP eruptions triggered the end-Triassic climatic and biotic perturbations requires that the basaltic eruptions commenced before and temporally overlapped the Tr-J boundary. This issue is debated in regards both to the ENA basins, as well as other CAMP localities (i.e. Morocco). Some authors (e.g. Fowell and Traverse, 1995; Whiteside et al., 2007) placing the Tr-J boundary below, hence before, the earliest basalt eruptions define the boundary location based on the following observations: (1) the disappearance of several Triassic sporomorphs (i.e., *P. densus*); (2) compositional variation of the Circumpolles group; and (3) a bloom of fern spores in the overlying strata, which is interpreted as a pronounced palynofloral turnover after the Tr-J mass extinction. However, the data presented in this study do not support a significant palynological turnover before emplacement of the oldest CAMP lava flows. In the following discussion we focus on five points that address specifically the end-Triassic age of CAMP volcanism in the Fundy basin and how it relates to the Tr-J boundary.

6.1 Is the last occurrence of *Patinasporites densus* a marker of the Tr-J boundary?

In the ENA basins, including the Fundy basin of the present study, the Tr-J boundary is currently defined, in part, based on the last occurrence of the Triassic sporomorph *P. densus*. However, the last appearance of *P. densus,* as well as other

425 vesicate sporomorphs, is considered to occur at the upper Norian - lowermost Rhaetian

426 boundary of the Late Triassic rather than at the Tr-J boundary, as documented in

427 numerous independently dated key sections (see online material, Fig. S3). The last

428 occurrence of this sporomorph in the Northern and Southern Hemispheres in the late

429 Norian to Early Rhaetian (Sevatian-Rhaetian boundary) suggests a late Norian-early

430 Rhaetian age for the palynological event in the Newark Basin rather than a Tr-J boundary

431 age, as proposed by some (Gradstein et al., 1994; Van Veen, 1995, Lucas and Tanner,

432 2007). The possibility of a floral provincialism controlling a diachronous distribution in

433 North American and Tethyan domains of the parent plants (Fowell and Olsen, 1993) does

434 not seem to be valid given the large distribution of these taxa in different palaeoclimate

435 belts (Buratti and Cirilli, 2007). Furthermore, the updated palaeogeographic

436 reconstructions based on palaeomagnetism (Kent and Tauxe, 2005) discount the

437 possibility of explaining the discrepancy by a latitudinal floral gradient. *P. densus* is

438 present in numerous Triassic sections located in similar palaeoclimatic and

439 palaeogeographic belts (e.g.: Carnian formations in Worcester, Tewkesbury and Reddict

440 districts, Barclay et al., 1997). It is notable that in Northern Italy (Southern Alps) the last

441 occurrence of *P. densus* is recorded (Jadoul et al., 1994) several hundred meters below

442 the Tr-J boundary based on biostratigraphic, stratigraphic and geochemical constraints

443 (Galli et al., 2007). Correlation of the palynological assemblage across the putative Tr-J

444 boundary in the Fundy and Newark basins (Whiteside et al., 2007) with the section at St.

445 Audrie's Bay (Hounslow et al., 2004) indicates that these sections are quite different, in

446 particular at the latter section where *P. densus* is absent in the Triassic portion of the

447 section below the Tr-J boundary. Clearly, this lack of correlation cannot be explained by

448 a provincialism of this taxon.

449

450 6.2 Does the compositional variation of Circumpolles group reflect an early Jurassic

451 palynological assemblage?

452   The miospores from the extinct conifer family Cheirolepidiacea (*Classopollis,*

453 *Gliscopollis, Corollina*) first appeared in the Late Triassic and became major elements in

454 Jurassic and Cretaceous palynoflora. The widespread distribution of *G. meyeriana* and *C.*

455 *torosus* commences in the Norian and continues into the Rhaetian and the Jurassic, as

456 documented in several locations in the Northern and Southern Hemispheres (e.g. Cornet

457 and Traverse, 1975; Schuurman 1979; ; Batten and Koppelhaus 2002; Warrington, 2002;

458 Hounslow et al. 2004; Barron et al., 2006; Buratti and Cirilli, 2007). Recent geochemical

459 and biostratigraphic constraints document that *C. torosus* and *G. meyeriana* were already

460 abundant within the Rhaetian (Rhaetipollis-Limbosporites zone of Kuerschner et al.,

461 2007; Barron et al. 2006; Gomez et al., 2007). Thus, the presence or relative abundance

462 of these Circumpolles taxa does not justify assigning a Jurassic age to the host strata, so

463 long as they are not part of an assemblage with only other Jurassic forms. Furthermore, a

464 valid biozone should be defined by one or more first occurrences rather than solely by

465 disappearances. In the Blomidon Formation, as well as in the Scots Bay Member, a

466 distinctly Hettangian species has not yet been recorded. Also significant is the absence of

467 *Cerebropollenites thiegarthi*, which is considered a marker for the basal Jurassic in

468 numerous palynological zonations that also are age-constrained by ammonoids,

469 conodonts and geochemical data (Kuerschner et al., 2007; von Hillebrandt et al., 2007;

470 Gomez et al., 2007).

471

472 6.3 Does the presumed palynological turnover and related fern spike mark the Tr-J

473 boundary?

474   It is essential to emphasize that the sediments investigated in this study were

475 deposited in continental environments (e.g. fluvial and lacustrine), which are

476 unfavourable to OM preservation. Consequently, the OM likely was subjected to strong

477 and episodic oxidizing conditions, as indicated by the palynofacies analysis. In the

478 Partridge Island section, for example, sporomorph abundance varies as a function of OM

479 preservation, which itself is associated with variable conditions of depositional/diagenetic

480 oxidation. Correlation of the sporomorph distribution with the palynofacies and TAI

481 variations demonstrates that the low percentage or absence of sporomorphs is coincident

482 with higher thermal indices and/or with more degraded-oxidized palynofacies (Fig.4,6).

483 Therefore, no significant palynological decline or turnover occurs below the North

484 Mountain Basalt that can be attributed to a mass extinction event. This observation is also

485 reinforced by the fact that except for *P. densus* and few other sporomorphs (Fig. 4) no

486 other last occurrences of Triassic sporomorphs have been recorded at the presumed Tr-J

487 boundary. A few last appearances (i.e. *C. cameronii, D. harrisii* and *Dictyophyllidites*

488 sp.) are only local most likely and palaeoecologically controlled, since they are present

489 above the putative Tr-J boundary as important elements of the so-called fern spike in

490 other ENA basins (Fowell and Olsen, 1993; Fowell et al., 1994). Although a true fern

491 spike is absent at the Partridge Island site, a level enriched in fern spores is recorded

492 within a dark clay mudstone (Fig. 4) located 15 cm below the putative Tr-J boundary, but

493 not above the boundary as in the Newark Basin. The co-occurrence of climate sensitive

494 facies (dark clay) with abundant fern spores could be related to local, more humid climate

495 conditions rather than to a re-colonization after mass-extinction. During the Mesozoic,

496 ferns developed easily in warm and humid environments (i.e. coal swamps, river banks),

497 preferring sheltered areas under the forest canopy, along creeks and streams and other

498 sources of permanent moisture (Abbink et al., 2004).

499

500   6.4 Are the Triassic sporomorphs in the Scots Bay Member reworked?

501   One of our most significant findings is the presence of a palynological assemblage

502 dominated by bisaccates, such as *L. acutus, L. rhaeticus, Lueckisporites* sp. with minor

503 Circumpolles, within the Scots Bay Member. The presence of specimens considered as

504 Triassic taxa (e.g. Schuurman, 1979; Batten and Koppelhous, 2002; Warrington, 2002;

505 Jadoul et al 1994, Hounslow et al., 2004) led to assigning these strata to the Triassic,

506 although they overlie the East Ferry and Margaretsville Members of the ~201 Ma North

507 Mountain Basalt. The colour of these sporomorphs, which are comparable to and often

508 slightly lighter than sporomorphs from the underlying Blomindon Formation, excludes

509 the possibility that they might be recycled from older strata. Therefore, while we can not

510 constrain the age of the Brier Island Member (one or two basalt flows) which is

511 discontinuously present along the Bay of Fundy and has not been observed in contact

512 with the Scot's Bay sediments, palynostratigraphy convincingly constrains the age of the

513 East Ferry and Margaretsville Members of the North Mountain Basalt to the Late

514 Triassic.

515

516   6.5 Is chron E23r at the Tr-J boundary and how long is the Rhaetian?

517   The Newark magnetostratigraphy shows that the reverse chron E23r occurs just

518 below the interval previously interpreted as the Tr-J boundary (Kent and Olsen, 2000;

519 Olsen et al., 2002). The new data from Partridge Island do not support this interpretation,

520 given the presence of Triassic sporomorphs in strata overlying the North Mountain

521 Basalt, hence above the supposed Tr-J boundary. Unless the Newark magnetostratigraphy

522 is missing several chrons (cf. Gallet et al., 2007), E23r must actually occur well before

523 the end of the Rhaetian. Hounslow et al. (2004), in fact, noted the ambiguity in

524 correlating palaeomagnetic records of the Newark basin and the St. Audrie's Bay

525 section, and suggested that E23r might correlate with SA5.2r, below the base of the

526 Lilstock Formation, rather than to SA5r, in the lower Blue Lias. In addition, it is

527 significant that the terrestrial components of the palynological assemblage at St. Audrie's

528 Bay (e.g. lack of *P. densus* and presence of typical uppermost Triassic taxa) are quite

529 different from the microflora found at the Partridge Island and also at the presumed Tr-J

530 boundary in the Newark basins.

531 This interpretation of the magnetostratigraphy is in good agreement with recent

532 biogeochemical and magnetostratigraphic data that indicate the supposed Tr-J boundary

533 of the Newark basin actually coincides with the Sevatian-Rhaetian boundary (Gallet et

534 al., 2007). However, this correlation and the results of the present biostratigraphic study,

535 as well as radio-isotopic ages, raise problems regarding the duration of the Rhaetian.

536 Considering that the age of the North Mountain Basalt is constrained to ~201 Ma

537 (Schoene et al., 2006; Jourdan et al., in press), i.e. indistinguishable from the age of the

538 marine Tr-J boundary (Schaltegger et al., 2008), the time which elapsed between the last

539 occurrence of *P. densus*, the eruption of CAMP basalts and the Tr-J boundary would be

540 extremely short. Although at face value the U/Pb age of the North Mountain Basalt

541 (201.3 ±0.3 Ma) predates that (201.6 ±0.3 Ma) for the Tr-J boundary only at relatively

542 low probability, we note that the former (obtained on a mafic pegmatite sheet in the upper

543 part of the East Ferry Member) may not date the earliest North Mountain Basalt

544 eruptions, and the latter (being determined from a silicic tephra) is more likely to be

545 affected by pre-eruptive residence time of the zircons, which can produce apparent age

546 bias of several hundred ka (Simon et al., 2008). Furthermore, if the last occurrence of *P.*

547 *densu*s can be considered as a Norian-Rhaetian or very early Rhaetian event, this would

548 imply a very short Rhaetian (<1 Ma). A short Rhaetian event suggested here concurs with

549 that proposed by Krystyn et al. (2007) and Gallet et al. (2007), but contrasts markedly

550 with conclusions of Channell et al. (1993), and Muttoni et al. (2004) who consider that

551 the Rhaetian had a duration of 6Ma. Alternatively, it might be considered that the

552 stratigraphic record of the Newark and Fundy basins are incomplete and that the Rhaetian

553 is partly or wholly missing, as suggested by Kozur and Weems (2005).

554

555 **7. Conclusions**

556 New palynological data from the Fundy Basin of Nova Scotia, Canada, constrain

557 the age of the sedimentary rocks below and just above the ~201 Ma North Mountain

558 Basalt, part of the widespread CAMP event, to the Late Triassic. In particular, this is

559 documented in the Fundy Basin by the presence of Triassic sporomorphs in the Scots Bay

560 Member, which overlies East Ferry and Margaretsville Members of the North Mountain

561 Basalt. Significantly, examination of the state of preservation and thermal alteration of

562 OM associated with the fossil assemblage precludes the possibility of recycling of the

563 Triassic sporomorphs from the older Blomidon Formation strata, which underlies the

564 basalt. From these new data, we draw the following conclusions:

565 1. Because the last occurrence of *P. densus* occurs 40 cm below the North Mountain

566 Basalt, and strata above the basalt (e.g. the Scots Bay Member) are clearly Triassic, it

567 follows that the palynological definition of the Tr-J boundary based on the last

568 occurrence of *P. densus* is not valid. Consequently, it appears that the occurrence of a

569 fern spike in the Newark Basin probably reflects a regional environmental perturbation ,

570 e.g. an abrupt but short-lived episode of climate change, rather than a global event.

571 2. Given that the previously accepted definition of the Tr-J boundary in the Newark Basin

572 cannot be supported palynologically, it follows that chron E23r, which is correlated with

573 the last occurrence of *P. densus* in the Newark Basin, also cannot be considered as a

574   marker of the Tr-J boundary. Instead, the chron E23r event merely marks a Late Triassic

575   (probably Rhaetian) reversal. This is an important point to consider correlating the Late

576   Triassic magnetostratigraphic data from marine sections (Hounslow et al., 2004; Gallet et

577   al., 2007) to the Newark continental section.

578   3. The occurrence of clearly Triassic strata in sharp contact with underlying North

579   Mountain Basalt lava flows clearly indicates that CAMP volcanism, at least in Nova

580   Scotia, is entirely of Triassic (Rhaetian) age. Significantly, this temporal relationship

581   must also apply to other areas of CAMP. In Morocco, for example, initiation of the

582   basaltic eruptions was coeval with deposition of sedimentary containing *P. densus*.

583   Cyclostratigraphy constrains the onset of CAMP volcanism elsewhere in the ENA (e.g.

584   the Fundy and Newark basins) to about 20-40 ky after the last occurrence of *P. densus*.

585   4. Because the eruption of the basalt plus deposition of at least the first few meters of the

586   Scots Bay Member occurred in a time span shorter than analytical resolution of the

587   geochronological data, emplacement of the basalt took place in a very short span of time.

588

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856

857

858   **Figure captions**

859   Fig. 1 Distribution of the CAMP in the lower Mesozoic. Eu: Europe (Southern Portugal

860   and Spain); Af: Africa (from Morocco to Ivory Coast); SA: Southern America (Brazil,

861   Bolivia and Guyana); ENA: Esatern North America (USA and Canada); (left) distribution

862   of the main rift basins in eastern North America and Morocco. Ne: Newark; Hr: Hartford;

863   De: Deerfield; Fu: Fundy; Ar: Argana; HA: High Atlas; Ou: Oujda

864

865   Fig. 2 Stratigraphic correlations, compiled from literature, among the main Newark

866   Super Group basins in the eastern North America (USA and Canada). These elongated

867   asymmetric basins, formed as a consequence of the rifting that led to the formation of the

868 Atlantic Ocean, filled with very thick successions of continental sediments interlayered

869 with tholeiitic basalts and intruded by sills and dykes. HMDN BSLT: Hampden basalt;

870 TLCT : Talcott S.B. Mb: Scots Bay Memberformational nomenclature and the group–

871 level stratigraphy are referred to Weems and Olsen (1997). Not to scale

872

873 Fig. 3 Study area in the Fundy basin (Nova Scotia); a) the studied section at Partridge

874 Island: the Blomidon Formation is overlain by the North Mountain basalt with a

875 stratigraphic contact; b) The Scots Bay section: the stratigraphic boundary between the

876 North Mountain basalt (bottom) and the lacustrine deposits of the Scots Bay Mb.; c) the

877 North Mountain basalt ; d) Geologic sketch of the Fundy basin area and location of the

878 studied sections: Partridge Island section (45.22.09 N; 64.20.11. W) (star); Scots Bay

879 (rhombus).

880

881 Fig. 4 Distribution range of the main sporomorphs. 4a)palynological assemblage from

882 Blomidon Formation at Partridge Island section; 4b) palynological assemblage from the

883 lowermost part of the Scots Bay Member;   the dotted line indicates the previous

884 established Tr-J boundary (Fowell and Traverse, 1995; Whiteside et al, 2007).

885

886 Fig. 5 Palynological assemblage from the Blomidon Formation (PI) and Scots Bay

887 Member (SB); A) *Patinasporites densus* (Leschik) Scheuring 1970, sample PI45(4)

888 England Finder coordinates (E.F.c.) S36(1), B) *P. densus*, PI55(4) E.F.c. W36(3), C) *P.*

889 *densus*, PI55(4) E.F.c. K42, D) *Glyscopollis meyeriana* (Klaus) Venkatachala 1966,

890 PI45(4) E.F.c. R33, E) *Conbaculatisporites mesozoicus* Klaus 1960, PI30(2) E.F.c. P34,

891 F) *G. meyeriana*, PI45(4) E.F.c. O24(1), G) *Classopollis torosus* (Reissinger) Couper

892 1958, PI45(4) E.F.c. S46, H) *C. torosus*, PI45(4) E.F.c. R29(4), I) *Cycadopites* sp.,

893   PI55(3) E.F.c. N44(1), J) *C. torosus*, PI55(3) E.F.c. H28(4), K) *C. torosus*, PI45(4)

894   E.F.c. L42(1), L) *Corollina murphyae* Cornet et Traverse 1975, PI45(4) E.F.c. L50, M)

895   *C. murphyae*, PI45(4) E.F.c. X38(1), N) *Calamospora mesozoica* Couper 1958, PI45(4)

896   E.F.c. P38(1), O) *Converrucosisporites cameroni* (de Jersey) Playford andDettmann

897   1965, PI45(4) E.F.c. U40(3), P) *Alisporites parvus* de Jersey 1962, PI45(4) E.F.c. J31,

898   Q) *Carnisporites spiniger* (Leschik) Morbey 1975, PI20(4) E.F.c. V33, R) *A. parvus*,

899   PI45(3) E.F.c. V52(2), S) *A. parvus*, SB2(2) E.F.c. H40(3), T) *Lunatisporites rhaeticus*

900   Leschik, 1955, SB2(1) E.F.c. G47(1), U) *Platysaccus* sp., SB2(2) E.F.c. F33(3/4); V) cf.

901   *Lunatisporites acutus*, SB2(2) E.F.c. L34(1); W) *Lunatisporites acutus*, Leschik, 1956

902   emend. Scheuring, 1970, SB2(2) E.F.c E44(3/4); X) undetermined circumpolloid, SB2(2)

903   E.F.c L39(2).

904

905   Fig. 6 Quantitative analysis of the palynofacies across the Blomidon Formation. Type and

906   amount of organic debris show sensible variations through the section: palynofacies

907   enriched in amorphous organic matter (c) are scarce and present only in the basal gray

908   mudstone; sporomorphs (b) are present, although in different percentage, in the whole

909   succession with a peak in correspondence of a dark grey mudstone, few centimetre below

910   the presumed Tr-J boundary of Whiteside et al., 2007; palynofacies dominated by

911   intertinite (a) are common, showing a peak in abundance in the lower and upper part of

912   the sampled section, where sporomorphs are absent and/or strongly degraded and

913   oxidised.

914
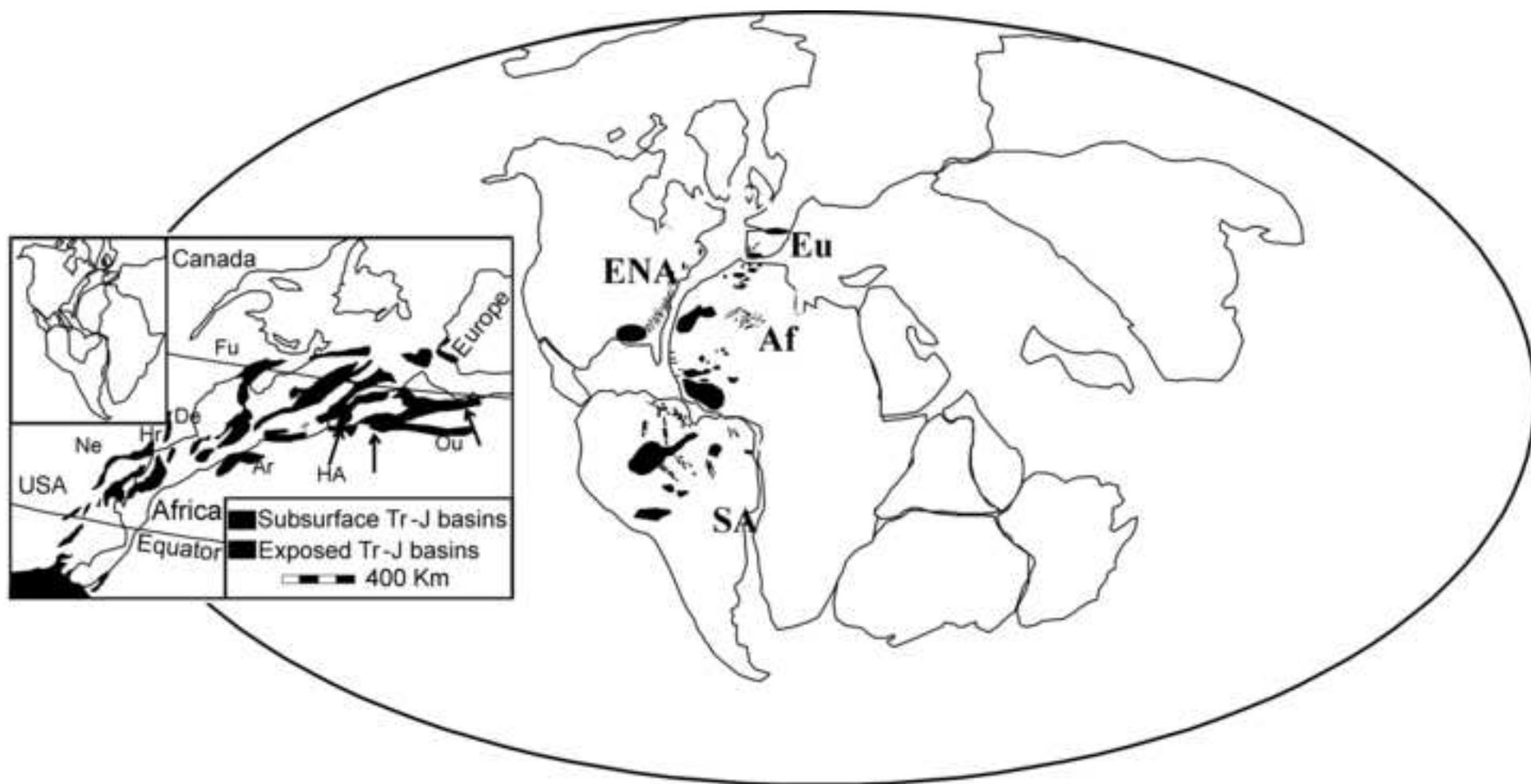
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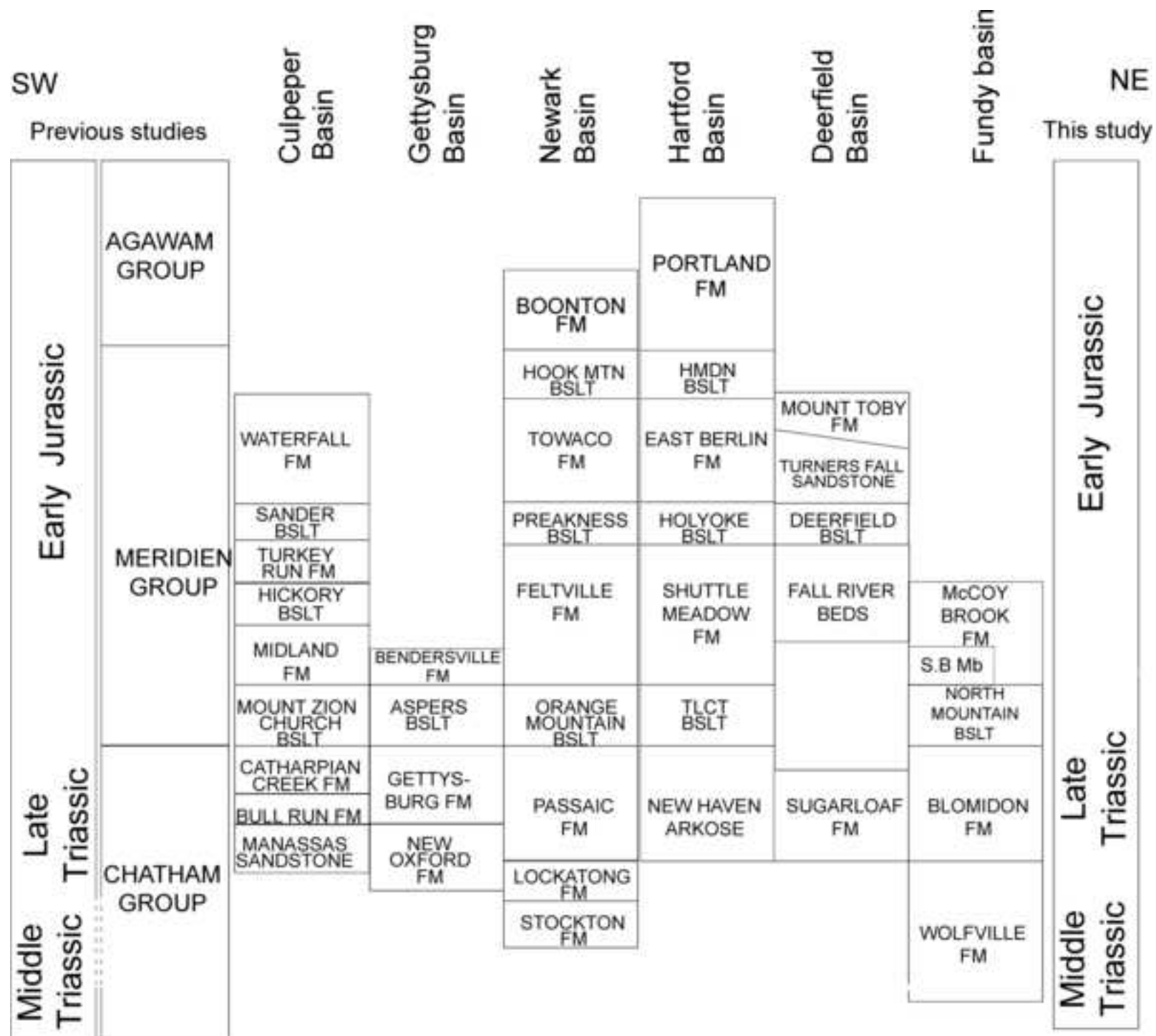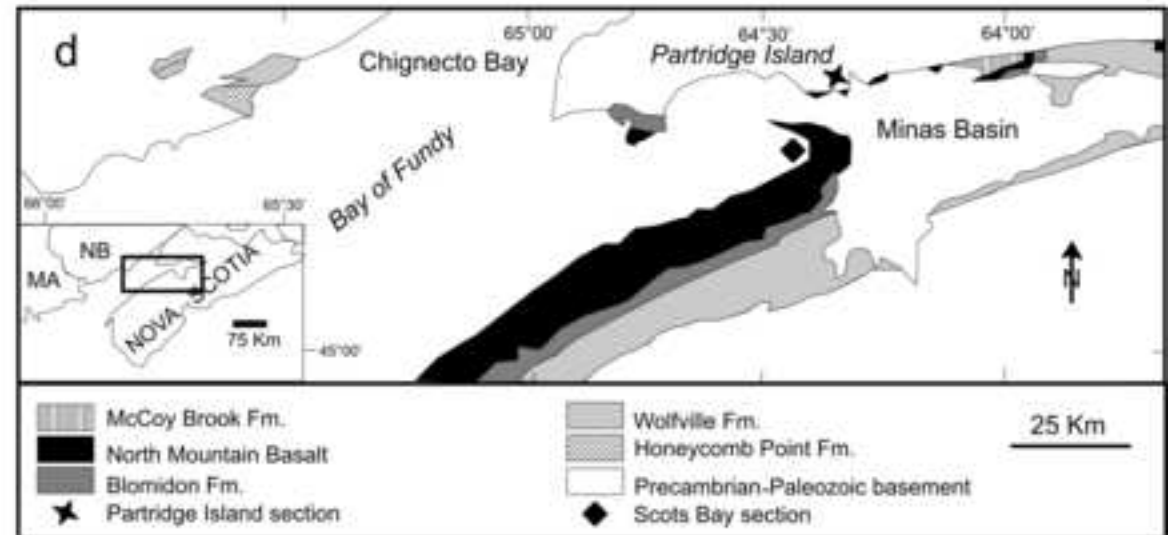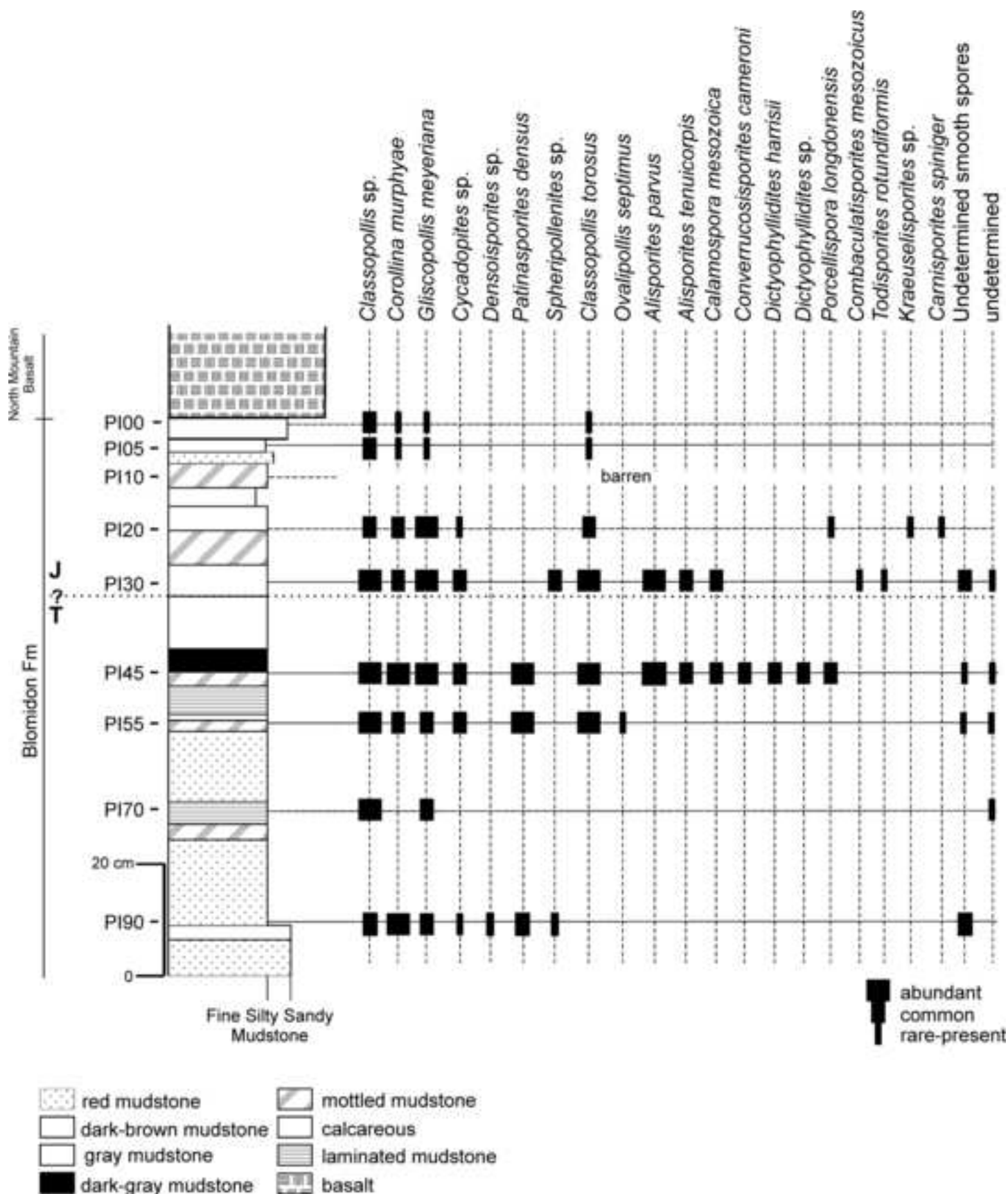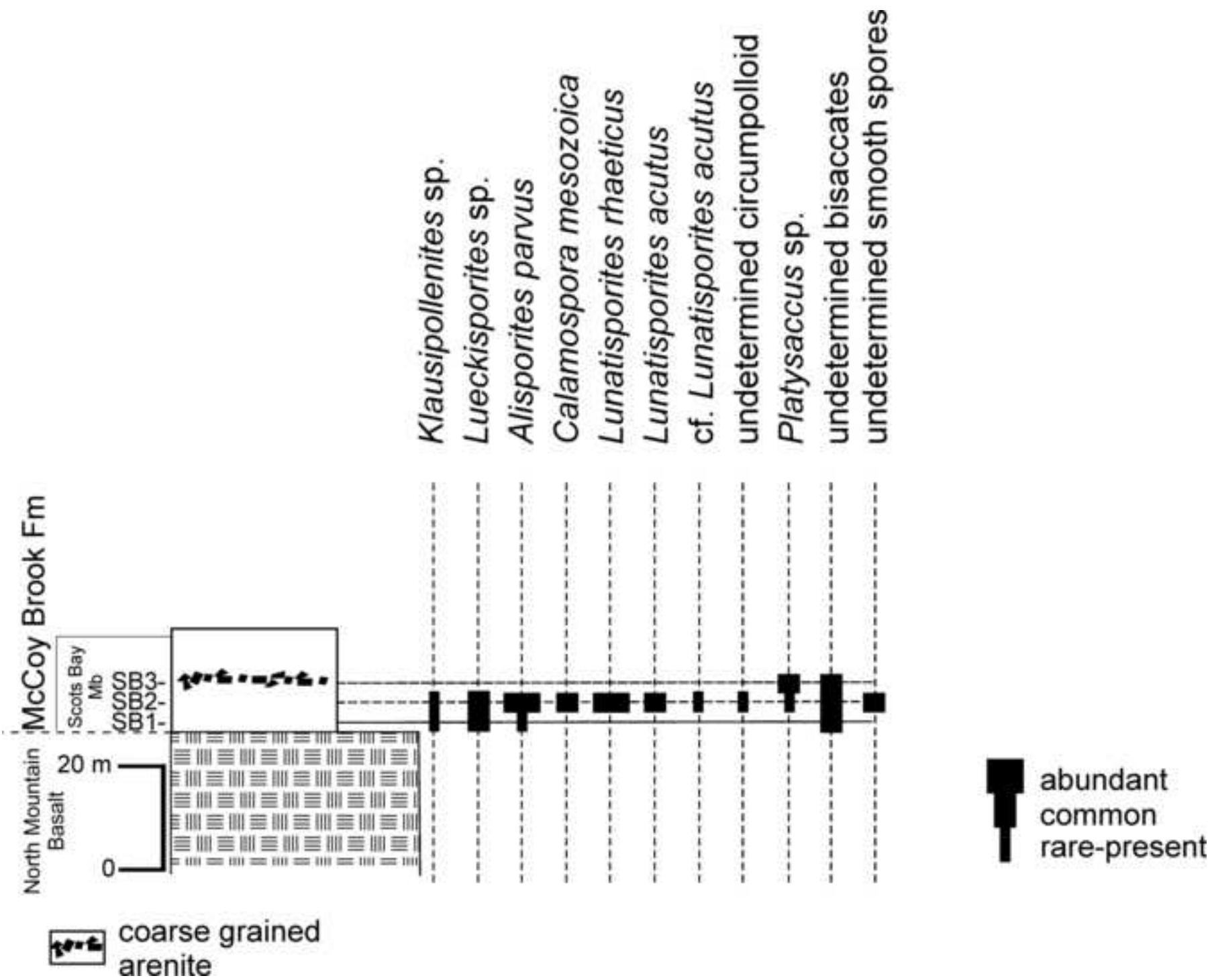
red mudstone   mottled mudstone   ·········· sporomorphs
dark-brown mudstone   calcareous   —— amorphous organic matter (AOM)
gray mudstone   laminated mudstone   ·—·—· vitrinite (PM1-PM2)
dark-gray mudstone   basalt   – – – inertinite (PM4)

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origin		group	constituent	approx. coal maceral equiv.
ALLOCHTONOUS CONTINENTAL FRACTION	higher plant debris	phytoclasts	PM4	Inertinite
			PM1-PM2	Vitrinite
			PM3	Cutinite
	pollen and spores	sporomorphs	saccates	Sporinite
			non-saccates	
	degraded plant debris	amorphous phytoclasts	non-fluorescent AOM	Vitrinite
		amorphous organic matter (AOM)		

Fine Silty Sandy Mudstone

red mudstone
dark-brown mudstone
gray mudstone
dark-gray mudstone
coarse grained arenite

mottled mudstone
calcareous
laminated mudstone
basalt

- Van der Zwan and Spaak, 1981 - NETHERLANDS - Carnian-early Rhaetian
- Boutet, 1981 - FRANCE - Carnian
- Scheuring, 1970 - SWITZERLAND - late Carnian
- Goczan and Oravec Scheffer, 1996 - HUNGARY - early Carnian
- Dunay and Fisher, 1978 – AUSTRIA – Carnian
- Hauschke and Heunisch, 1990 - GERMANY - late Ladinian-Carnian
- Van der Eem, 1983 - ITALY Dolomites – Carnian
- Petrunova-Olova, 1990 - BULGARIA - early Carnian
- Broglio Loriga et al., 1999 - ITALY Dolomites – early Carnian
- Blendinger, 1988 - ITALY Dolomites – Carnian
- Visscher and Krystyn, 1978 – ITALY Sicily – late Carnian
- Martini et al., 1991 - ITALY Sicily - late Carnian
- Cirilli and Montanari, 1994 - ITALY Sicily- late Carnian
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- Besems, 1983 - SPAIN - early Carnian
- Doubinger et al., 1990 - SPAIN - Carnian
- Besems and Simon, 1982 - SPAIN - late Carnian
- Besems, 1981 - SPAIN - Norian
- Brugman and Visscher, 1988 - LIBYA- late Ladinian to Carnian
- Bourmouche et al.,1996 - ALGERIA - late Carnian
- Peybernes et al., 1993 - TUNISIA - early Norian
- Hankel, 1993 - TANZANIA - Ladinian-Norian

- Fisher and Bujak, 1975 - ARCTIC CANADA - Carnian Fisher, 1979
- Leschik, 1955 – SWITZERLAND - Carnian
- Warrington, 1970 - ENGLAND - late Carnian - Warrington, 1971 - ENGLAND - middle to late Carnian
- Fisher, 1972 – ENGLAND - late Carnian
- Dunay and Fisher, 1974 - USA - Carnian
- Dunay, 1973 - USA - Norian
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- Litwin and Ash, 1993 - USA – late Carnian
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- Eshet, 1990 – ISRAEL - Carnian
- Eshet and Cousminer, 1986 – ISRAEL - late Carnian
- Manspeizer et al., 1978 - MOROCCO - middle Carnian
- Baudelot et al., 1993 - MOROCCO - early Carnian

- Martini et al., 2000 - TIMOR - early Carnian-early Norian
- Dolby and Balme, 1976 – AUSTRALIA - Carnian to Norian
- de Jersey and Raine, 1990 - NEW ZEALAND - late Norian-Rhaetian