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Latest Triassic onset of the Central Atlantic Magmatic Province (CAMP) volcanism in the Fundy Basin (Nova Scotia): New stratigraphic constraints

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ABSTRACT

In this paper we investigate the stratigraphic relationship between the emplacement of the CAMP basalts and the Triassic–Jurassic (Tr–J) boundary in the Fundy Basin (Nova Scotia, Canada). This is one of the best exposed of the 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 Blomidon Formation) and just above the North Mountain Basalt (Scots Bay Member of the McCoy Brook Formation) 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 Blomidon Formation is characterised by the dominance of the Circumpolles group (e.g. *Gliscopollis meyeriana, Corollina murphyae, Classopollis torosus*) which crosses the previously established Tr–J boundary. The Triassic species *Patinasporites densus* disappears several centimetres below 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 previous definition of the Tr–J boundary in the ENA basins, which was 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.

1. Introduction

The Triassic–Jurassic (Tr–J) boundary, most recently dated by U/Pb at 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; Olsen et al., 2002; McElwain et al., 2007; Whiteside et al., 2007). A marked

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negative carbon isotope (δ^{13}C) anomaly in both carbonate and sedimentary organic matter (OM) observed in marine Tr–J boundary strata, e.g. from Hungary, England, U.S.A., Canada, Spain, Italy, Austria (e.g Pálfy et al., 2001; Hesselbo et al., 2002; Guex et al., 2004; Ward et al., 2004; Gómez et al., 2007; Galli 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 hydrate release (Pálfy et al., 2001; Tanner et al., 2004; Ward et al., 2004; Lucas and Tanner, 2008).

In summary, there is a general consensus that a biotic and environmental perturbation occurred during the latest Triassic and spanned the Tr–J boundary (Tanner et al., 2004; Lucas and Tanner, 2008). Some of the forcing mechanisms that have been invoked to explain this disruption of the ecosystem include: (1) rapid sea-level change and/or anoxia (Hallam and Wingall, 1999); (2) bolide impact (Olsen et al., 2002; Ward et al., 2004); and (3) atmospheric loading by CO_2 and SO_2 related to eruptions of the Central Atlantic Magmatic Province (CAMP), which potentially caused atmospheric warming of up to 3–4 °C or acidic atmospheric pollution (Marzoli et al., 1999; McElwain et al., 1999; Hesselbo et al., 2002; Marzoli et al., 2004; Guex et al., 2004; Knight et al., 2004; Cohen and Coe, 2007; Tanner et al., 2007; Schaltegger et al., 2008; Van de Schootbrugge et al., 2008).

Emplacement of the tholeiitic dikes, sills, and flood basalts of CAMP occurred in North and South America, Africa and Europe over a total surface area potentially in excess of 10 million km^2 (Marzoli et al., 1999, 2004; Verati et al., 2007) (Fig. 1). The available ^{40}Ar/^{39}Ar plateau ages for the basalts suggest brief durations (<1 Ma) of peak eruptive activity at specific locations (Deckart et al., 1997; Marzoli et al., 1999, 2004; Knight et al., 2004; Nomade et al., 2007; Verati et al., 2007), although the main interval of emplacement across the entire CAMP may have had a duration of 2 ± 1 Ma centered at ca. 199 Ma (Jourdan et al., 2009). These ^{40}Ar/^{39}Ar ages conform to the U–Pb age of the Tr–J boundary (201.6 ± 0.3 Ma, single zircon; Schaltegger et al., 2008) considering the well-established ca. 1% bias between the two isotopic dating methods (Min et al., 2000; Kuiper et al., 2008). However, taking into account analytical and intercalibration uncertainties, as well as the short duration of the eruptive events, the published radio-isotopic ages indicate proximity, but cannot ultimately define the relative timing of CAMP volcanism and the Tr–J boundary crisis. Therefore, correlation of the biotic turnover from the marine realm, where the boundary is defined by ammonites (Guex et al., 2004) with the continental CAMP volcanism is required. Various analytical methods have been applied, including palynology, magnetostratigraphy and geochemistry. However, multidisciplinary studies applied to the North American and African Tr–J sediments and interlayered CAMP basaltic flows have produced contrasting interpretations.

Since the 1970s the palynological Tr–J boundary in the ENA continental rift-basins has been placed in the strata immediately beneath the lowest CAMP basalt flow (e.g. the Orange Mt. Basalt in the Newark basin). Thus, all the strata overlying and interlayered with the CAMP lava flows were assigned to the Jurassic. This boundary is defined, according to both older and more recent studies (e.g. Cornet and Traverse, 1975; Cornet and Olsen, 1985; Fowell and Olsen, 1993; Fowell et al., 1994; Fowell and Traverse, 1995; Whiteside et al., 2007), by a marked palynological turnover event recognized on the basis of the following: (1) some last appearances of certain species (*Ovalipollis ovalis*, *Vallasporites ignacii*, and *Patinasporites densus*); (2) an increase in *Corollina* spp. (= *Classopollis* and *Gliscopollis*) percentage; (3) and a bloom of trilete spores (fern spike), considered as the expression of a renewed palynoflora after the Tr–J mass extinction. These palynostratigraphic criteria for placing the Tr–J boundary have been criticised (Gradstein et al., 1994; Van Veen, 1995; Lucas and Tanner, 2007) because locally (e.g. the European domain) the disappearance of the vesicate forms (e.g. *V. ignacii*, and *P. densus)* took place during the late Norian and occasionally early Rhaetian (e.g. Morbey, 1978; Schuurman, 1979; Krystyn et al., 2007; Kuerschner et al., 2007). Notably, Kozur and Weems (2005), relying on conchostracan assemblages from the Culpeper and Newark basins, concluded that the Tr–J boundary in the Newark Supergroup must occur above the lowest CAMP basalt.

In the Newark and Fundy basins, the putative Tr–J boundary is also characterized by a moderate Ir anomaly, the origin of which is disputed (Olsen et al., 2002; Tanner and Kyte, 2005; Tanner et al., 2008); more importantly, the Ir anomaly is immediately preceded by a magnetic

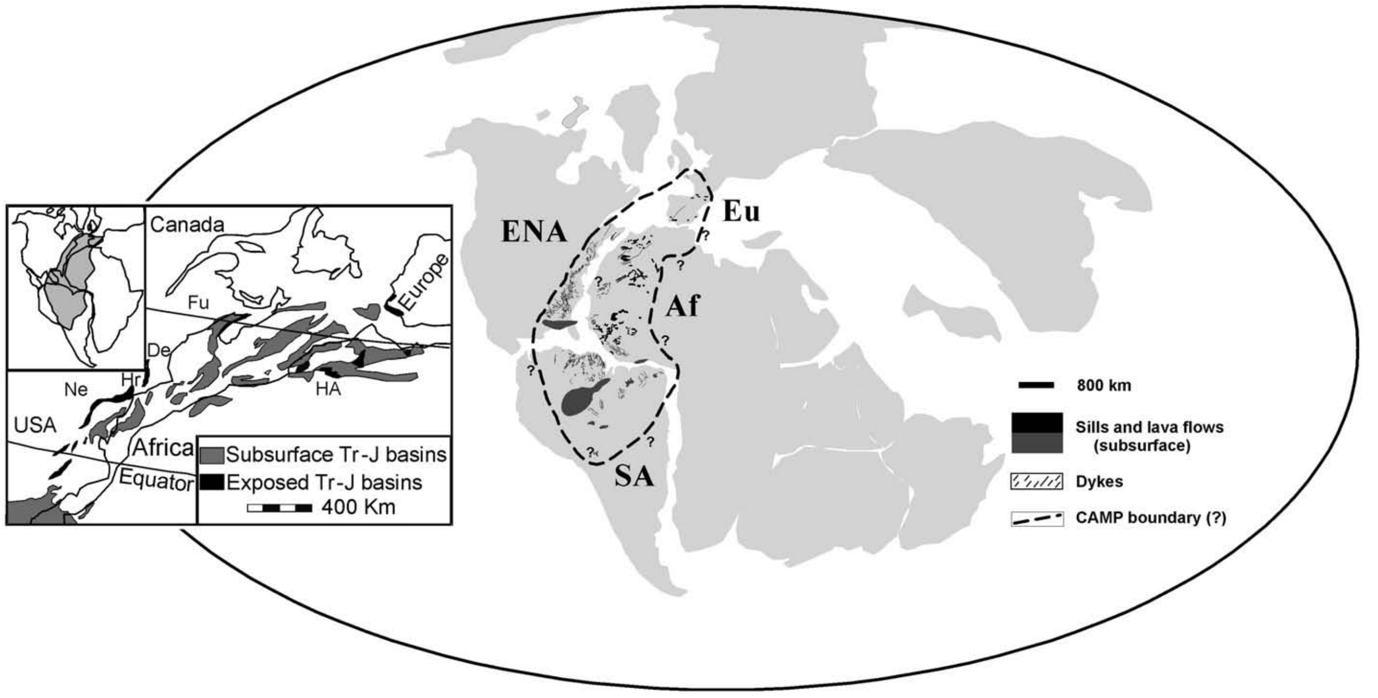

Fig. 1. Distribution of the CAMP at the Triassic–Jurassic transition. Eu: Europe (Portugal, Spain, and France); Af: Africa (from Morocco to Ivory Coast); SA: Southern America (Brazil, Bolivia, Suriname and Guyana); ENA: Eastern North America (USA and Canada); (left) distribution of the main rift basins in eastern North America and Morocco. Ne: Newark; Hr: Hartford; De: Deerfield; Fu: Fundy; HA: High Atlas.

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reversal (chron E23r; Kent and Olsen, 1999). Indeed, the most complete Late Triassic to earliest Jurassic magnetic polarity time scale has been calibrated on the non-marine section of the Newark basin (e.g. Kent and Olsen, 1999), which is anchored to the palynologically defined Tr–J boundary (see Lucas and Tanner, 2007 for a historical review), and thus represents a potentially powerful tool for global correlations. The E23r reversal has been correlated by Whiteside et al. (2007) to one of the brief reversals (chrons SA5n.3r and SA5r) of the marine St. Audrie's Bay section (England), which corresponds to the Tr–J boundary in the upper part of the Penarth Group (Hounslow et al., 2004). Recently, Gallet et al. (2007) noticed discrepancies between the Newark magnetostratigraphy and that of the Tethyan realm (e.g. the Oyuklu section of south-western Turkey), which suggests either a higher position of the Norian–Rhaetian boundary and/or a lack of the lowermost part of the Rhaetian in the Newark sequence.

The cyclostratigraphy of the Newark basin suggests that the onset of CAMP volcanism postdated the Tr–J boundary by 20–40 ky and that the entire basalt pile was erupted in the Early Jurassic during normal polarity chron E24n (Kent and Olsen, 1999; Olsen et al., 2003; Kent and Olsen, 2008). The next reversal, chron E24r, occurred about 1.6 Ma after the supposed Tr–J boundary, and above the youngest CAMP lava flow (Hampden Basalt), as recorded in the Hartford basin (Kent and Olsen, 2008). This reversal tentatively has been correlated with reversals detected in Jurassic strata from the Paris basin, France (Yang et al., 1996).

In contrast to the interpretation that all CAMP volcanism postdated the Tr–J boundary, multidisciplinary data suggest a latest Triassic onset of the volcanism in Morocco (Knight et al., 2004; Marzoli et al., 2004). This conclusion is based on: (1) evidence that the first lava flows erupted were synchronous with sediments bearing a Triassic palynological assemblage dominated by *P. densus, Corollina murphyae, Gliscopollis meyeriana* and *Classopollis torosus;* (2) the presence of a palaeomagnetic reversal within the lava pile (e.g. approximately halfway between the base and top) that has been correlated with the Tr–J boundary reversal from the marine St. Audrie's Bay section, and possibly with chron E23r from the Newark basin; (3) geochemical correlations of the basalts between basins, suggesting that CAMP volcanism started slightly earlier in Morocco than in the Newark basin. However, the data for the Moroccan CAMP have been questioned (Whiteside et al., 2007, 2008; *contra* Marzoli et al., 2008), which means that the relationship between the age of CAMP volcanism and the Tr–J boundary remains an unresolved question. This problem mainly results from the absence of a generally accepted chronostratigraphic definition of the Tr–J boundary,

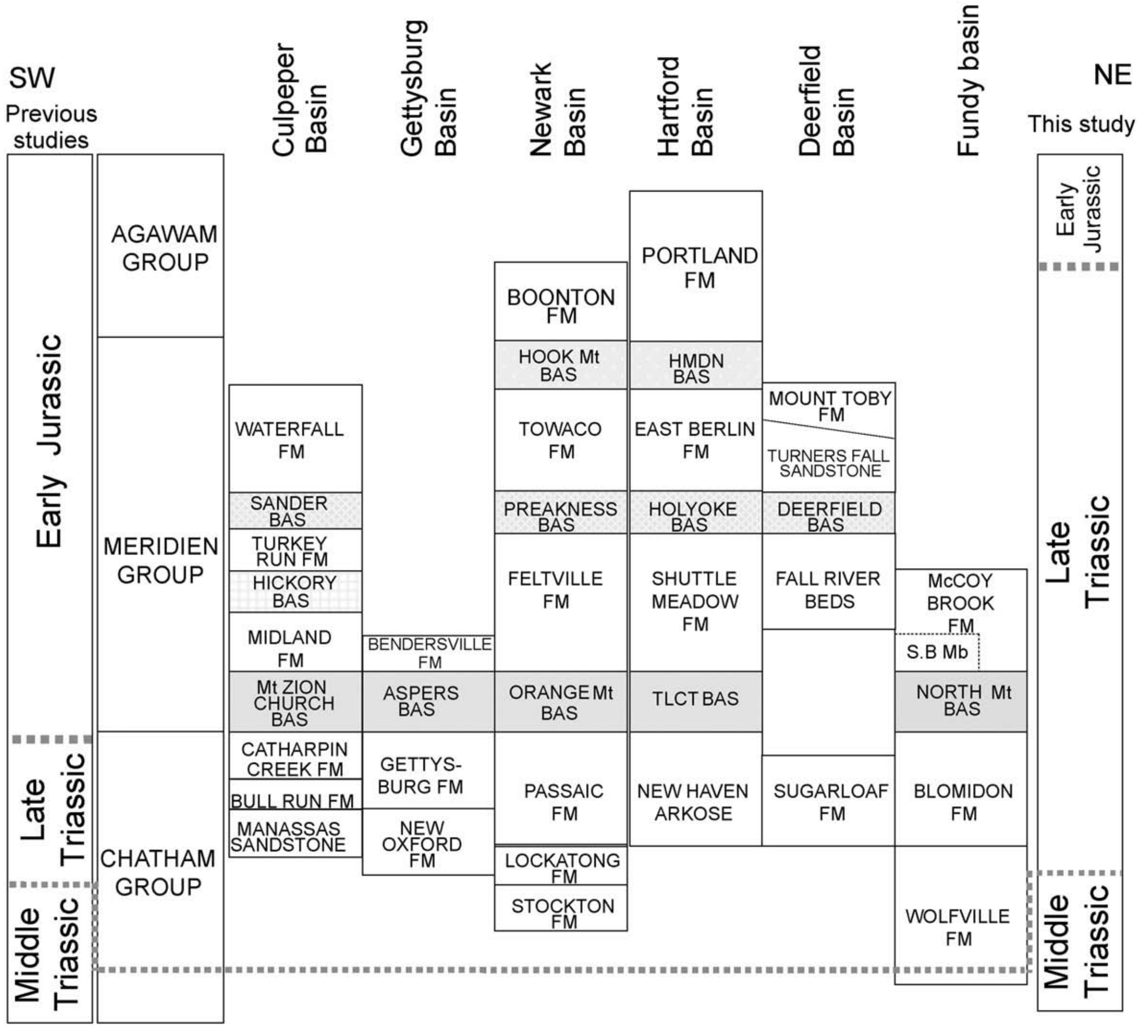

Fig. 2. Stratigraphic correlations, compiled from literature, among the Newark Super Group basins in the eastern North America (USA and Canada) yielding CAMP basalt flows. These elongated asymmetric basins, formed as a consequence of the rifting that led to the formation of the Atlantic Ocean, filled with very thick successions of continental sediments interlayered with tholeiitic basalts and intruded by sills and dykes. HMDN BAS: Hampden basalt; TLCT BAS: Talcott basalt; S.B. Mb: Scots Bay Member. Formational nomenclature and the group-level stratigraphy after Weems and Olsen (1997). The results of the present study are referred to the Triassic–Jurassic transition. Not to scale.

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the disputed biostratigraphic value of the palynological assemblages, and the lack of a well constrained age for the sediments interlayered with or covering the CAMP lava flows.

2. Area studied

In this paper we investigate the relationship between the emplacement of the CAMP basalts and the Tr–J boundary in the Fundy Basin (Nova Scotia, Canada), one of the best exposed ENA synrift basins. These basins, collectively referred to as the Newark Supergroup, are more or less continuously exposed on the east coast of North America, from Nova Scotia to South Carolina (Olsen and Schlische, 1990; Tollo and Gottfried, 1992; Olsen et al., 2003; Whiteside et al., 2007). Formational nomenclature and the group-level stratigraphy used in this paper are after Weems and Olsen (1997) (Fig. 2).

The Fundy Basin, which was located at 24° N palaeolatitude at the close of the Triassic (Kent and Tauxe, 2005), is the northernmost of the Mesozoic rift basins of ENA (Fig. 3D), and its sedimentary and volcanic sequence is well exposed on the shores of the Bay of Fundy in western Nova Scotia and eastern New Brunswick. The Fundy Basin, including the Minas, Fundy, and Chignecto structural sub-basins (Olsen and Schlische, 1990), consists of a thick siliciclastic succession, of which about 1 km of fluvial, lacustrine and eolian sediments crops out in Nova Scotia and New Brunswick, and 4 km are submerged beneath the Bay of Fundy (Olsen et al., 2005). The Blomidon Formation of the Fundy Group, which is assigned to the middle-late Norian and possibly Rhaetian on the basis of magneto-cyclostratigraphy (Kent and Olsen, 2000; Olsen et al., 2005), comprises 200–300 m of cyclically interbedded sandstone and mudstone of mostly lacustrine, playa, eolian and fluvial origin deposited during an interval of semi-arid to arid climate (Tanner, 2000; Tanner and Kyte, 2005). The uppermost Blomidon Formation (Partridge Island Member) contains the turnover event cited by Fowell and Traverse (1995) as the Tr–J palynological boundary.

The Blomidon Formation is overlain by the tholeiitic North Mountain Basalt, which is overlain in turn by the McCoy Brook Formation. The North Mountain Basalt (about 400 m thick) crops out on the north and south shores of the Minas Basin, along the western coast of Nova Scotia and on Grand Manan Island (New Brunswick). It is subdivided into three lava flow units, which are distinguished in the field by volcanological features (Papezik et al., 1988; Kontak, 2008). The lowermost unit, the East Ferry Member, is composed of a single massive basalt flow up to 180 m thick, which in its upper part locally contains sheets ($\leq$1–2 m) of coarse-grained mafic pegmatite and granophyre. The middle unit, the Margaretsville Member, is formed by up to 16 individual inflated, pahoehoe-type flows, up to 10 m thick, that are heavily vesiculated and altered. The uppermost unit, the Brier Island Member, consists of one to two massive basalt lava flows that are similar to the lowermost flow unit. Significant sedimentary intercalations are absent between the lava units or flows, although flow tops commonly are oxidized, suggesting only brief hiatuses between volcanic events. U/Pb and ^{40}Ar/^{39}Ar ages are available for the North Mountain Basalt. A single-zircon U/Pb age of 201.3 ± 0.3 Ma was obtained on a granophyric matrix in a mafic pegmatite sheet in the upper part of the East Ferry Member (Schoene et al., 2006). ^{40}Ar/^{39}Ar plateau ages on plagioclase from lower and upper unit lava flows range

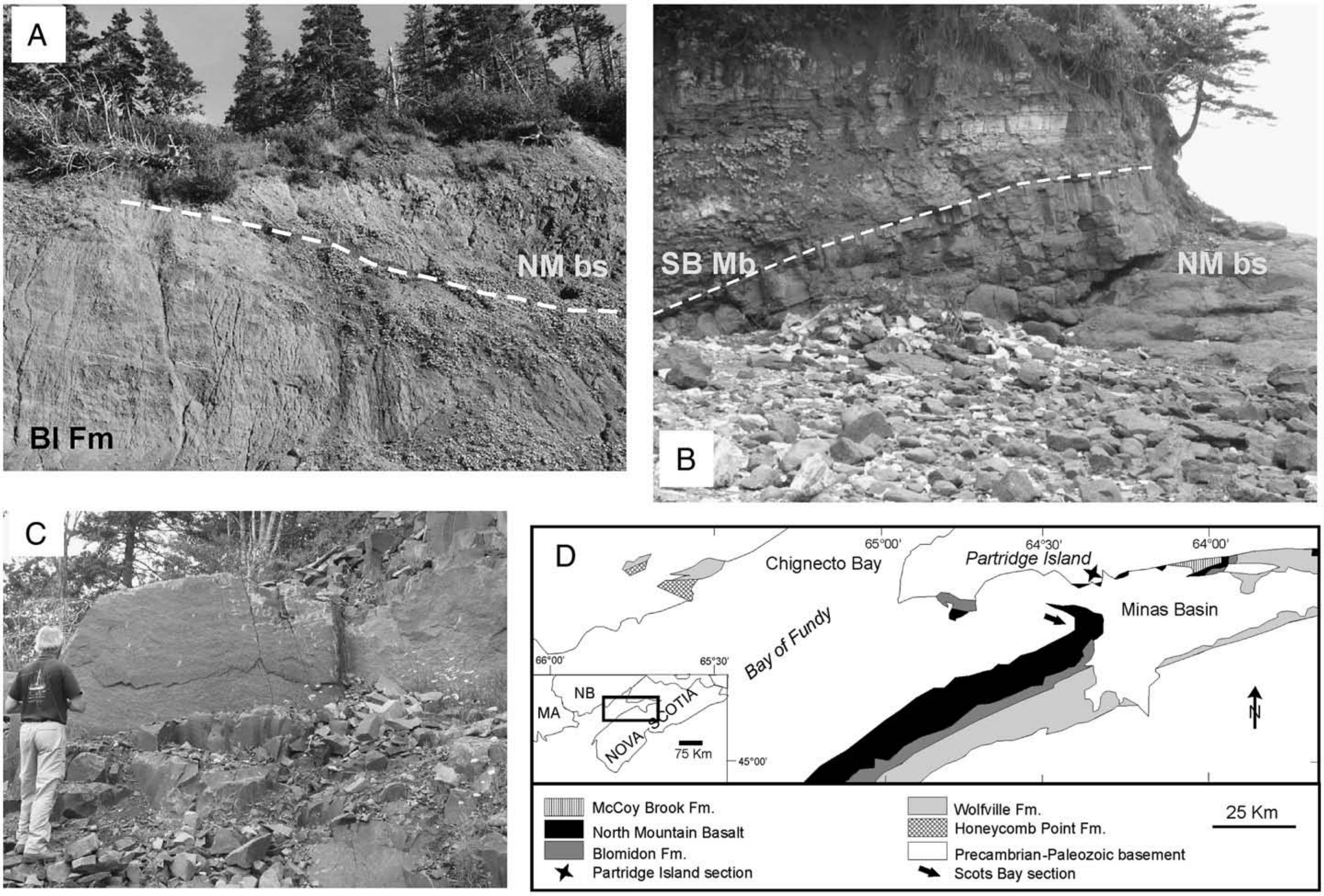

Fig. 3. Study area in the Fundy Basin (Nova Scotia); A) the studied section at Partridge Island: the Blomidon Formation (Bl Fm) is overlain by the North Mountain basalt (NM bs) with a stratigraphic contact; B) The Scots Bay section: the stratigraphic boundary between the North Mountain basalt (NM bs) and the lacustrine deposits of the Scots Bay Mb (SB Mb); C) the North Mountain basalt near Digby (note the coarse-grained massive mafic pegmatite, top of outcrop); D) Geologic sketch of the Fundy Basin area and location of the studied sections: Partridge Island section (45 22 09 N; 64 20 11 W) (star); Scots Bay 45 14 56 N; 64 27 11 W (arrow).

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from 198.6 ± 1.1 to 201.0 ± 1.4 Ma (Jourdan et al., 2009), supporting a duration of the peak eruptive event on the order of, or shorter than ~1 Ma. These ages overlap with the U/Pb age of Schoene et al. (2006), and with the U/Pb age of the Tr–J boundary (Schaltegger et al., 2008), if analytical and intercalibration uncertainties are considered. Furthermore, the ages are indistinguishable from high-quality isotopic ages for basalts from other areas of the northern CAMP (i.e, Morocco and the U.S.A.; Knight et al., 2004; Marzoli et al., 2004; Nomade et al., 2007; Verati et al., 2007). The North Mountain Basalt flows show broad geochemical similarities with other CAMP basalts, but in detail the geochemical correlation with those flows is not straightforward (Marzoli et al., 2008).

The McCoy Brook Formation is the youngest unit in the Fundy rift basin and is mainly composed of red beds and abundant sand-rich fluvial deposits, gypsiferous sand patch playa/lacustrine cycles and local eolian sandstones (De Wet and Hubert, 1989; Tanner and Hubert, 1992; Tanner, 2000; Olsen et al., 2005). The basal member, the Scots Bay Member, generally lies directly on the North Mountain basalt.

3. Previous work

The entire sedimentary succession of the Fundy Basin has been dated mainly on the basis of palynology (Bujak, 1977; Fowell and Traverse, 1995; Whiteside et al., 2007), vertebrate fossils and cyclostratigraphic correlations with the Newark Supergroup (Olsen et al., 2005 for references). Bujak (1977) located the Tr–J boundary within the McCoy Brook Formation based on palynological data from the Chinampas N-37 well (Bay of Fundy). Subsequently, new palynological data from the Minas subbasin (at Partridge Island) resulted in placement of the Tr–J boundary in the uppermost meter of the Blomidon Formation (Fowell and Traverse, 1995; Whiteside et al., 2007), where the palynological assemblage exhibits a transition from dominantly *Corollina torosa* and *P. densus* to mainly *Corollina* spp. (*C. meyeriana, C. torosa, C. simplex, C. murphyae*). Whiteside et al. (2007), on the basis of new palynological data, placed the Tr–J boundary at less than 20 cm below the contact with the North Mountain Basalt (i.e. near the top of the Blomidon Formation), between the bed that contains the highest occurrence of *P. densus,* and the overlying sample, which lacks *P. densus* or abundant bisaccates. The stratigraphic position of this presumed turnover is similar to that observed in the Newark and Hartford basins (Fowell et al., 1994; Olsen et al., 2002).

As in the Newark basin (Olsen et al., 2002), a moderate Ir (maximum measured value <400 pg/g) and PGE (Platinum Group elements) enrichment occurs in multiple gray-coloured mudstone layers within the uppermost meter of the Blomidon Formation in the Partridge Island section, which is roughly coincident with the presumed palynological transition (Tanner and Kyte 2005; Tanner et al., 2008). In contrast to the Newark basin, no magnetic reversal (equivalent to E23r) and no fern spike (Kent and Olsen, 2000; Olsen et al., 2002), have been found in this section. The spike in fern spores would be produced by *Clathropteris meniscoides,* which occurs in well-preserved specimens in some localities of the Hartford basin (Cornet and Traverse, 1975). In the Fundy Basin, the only recorded *Clathropteris* has been found in the Scots Bay Member, while the species *C. meniscoides* has not been reported (Olsen et al., 2005). The McCoy Brook Formation contains a large variety of fossils which are mostly considered to be of early Jurassic age (De Wet and Hubert, 1989; Shubin et al., 1994; Olsen et al., 2005).

4. Research methods

4.1. Lithostratigraphy of the sampled section

The section sampled at Partridge Island is a cliff face partially covered by colluvium (Fig. 3A). The uppermost 1.0 m of the Blomidon Formation was exposed in a trench and sampled at 5–10 cm intervals (Figs. 4A and 6). The lower part of the exposed section consists of interbedded red and gray mudstones. The red mudstone is thinly to medium bedded (2–15 cm thick) and composed mostly of hematite-stained clays and silt to sand-size immature arenite grains. The coarser components consist mostly of quartz and minor micas and feldspars. Gray mudstones are mostly thin-bedded (2–5 cm in thickness) and contain a higher proportion of sand-size grains, including quartz and rock fragments (igneous and metamorphic), intercalated with organic-rich dark clay beds. Finely laminated mudstone containing dark organic-rich laminae interlayered with gray to reddish laminae are also present in the lower part of the sampled section. The clay fraction of the mudstones comprises mostly illite, smectite and chlorite, with lesser amounts of kaolinite and mixed-layer clays and an abundance of chlorite in a greenish-gray mudstone 20 cm below the contact with the base of the North Mountain Basalt, the East Ferry Member (Tanner and Kyte, 2005).

At Scots Bay, the outcropping strata of the Scots Bay Member are in direct stratigraphic contact with the Margaretsville Member of the North Mountain Basalt (the Brier Island Member is not present here; Fig. 3B and C). The Scots Bay strata here consist of a basal carbonate-rich sequence of two white, green, purple, and red lacustrine cycles. The basal portion contains coarse immature sandstone with ostracod- and peloid-bearing carbonate matrix. Clastic grains mostly consist of quartz, mottled carbonate mudstone, siltstone and volcanic debris.

4.2. Sampling and analysis

Twelve samples were selected for organic matter (OM) analysis. Of these samples, nine are from the uppermost Blomidon Formation beneath the North Mountain Basalt (East Ferry Member) at the Partridge Island section (Fig. 3A and D). The Tr–J boundary identified at this location in earlier palynological studies (Fowell and Traverse, 1995; Whiteside et al., 2007) was sampled in the interval represented by sample PI-30 (i.e. 30–35 cm below the basalt, Fig. 4A). The three remaining samples are from the lower Scots Bay Member, which overlies the North Mountain Basalt, and are from the section at Scots Bay (Fig. 3). These three samples were also analyzed for palynological assemblages (Figs. 4 and 5), palynofacies (Fig. 6) and thermal maturity (see supporting online material, Figs. S1 and S2).

5. Results

5.1. Palynological assemblage

The palynological content of the sampled section at Partridge Island (Fig. 5) is characterised by a dominance of Circumpolles group species, represented in order of decreasing abundance by *Gliscopollis meyeriana* (Fig. 5D and F), *C. murphyae* (Fig. 5L and M), *C. torosus* (Fig. 5G, H, J and K), and *C. simplex* (in this paper, the generic name *Classopollis* is used instead of *Corollina* or *Circulina*, as recently formally proposed by Traverse (2004), for *C. torosus*). The total abundance of the Circumpolles group varies slightly through the section (Fig. 4A). Generally, a lower abundance occurs in the most oxidized beds (i.e. red mudstones), as demonstrated by the state of preservation of the palynofacies (Fig. 6), which is mostly dominated by inertinite and strongly oxidized vitrinite. At these low levels of preservation, degraded, oxidized, and badly preserved sporomorphs prevail (i.e. PI-70, PI-5, PI-0). The low total content of Circumpolles also correlates with a low total OM concentration, which is poorly preserved in oxidized beds.

As shown in Fig. 4A, Circumpolles cross the Tr–J boundary previously established by Fowell and Traverse (1995), Whiteside et al. (2007) without a marked decline in abundance. *P. densus* (Fig. 5A, B and C) is present from the base of the sampled section at Partridge Island up to sample PI-45, which is near the presumed Tr–J boundary. The percentage of this vesicate pollen is rather low when compared with the other sporomorphs, comprising at most 10–15% of the association. Among bisaccates, *Alisporites parvus* (Fig. 5P and R), *A. tenuicorpis, Ovalipollis septimus* and other undetermined bisaccates are generally quite common,

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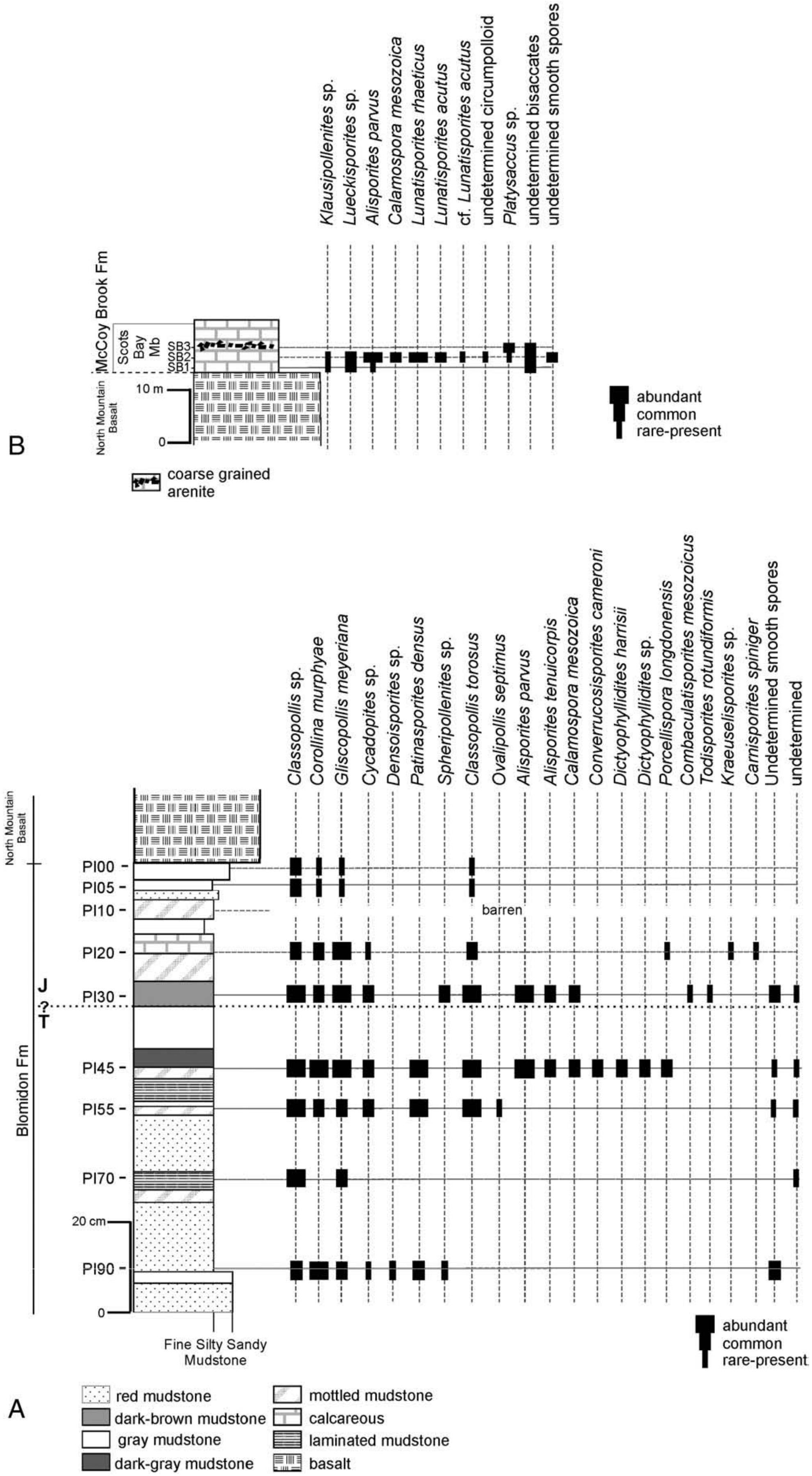

Fig. 4. Distribution range of the main sporomorphs. A) palynological assemblage from Blomidon Formation, Partridge Island section; B) palynological assemblage from the lowermost part of the Scots Bay Member; the dotted line indicates the Tr–J boundary as defined by Fowell and Traverse (1995) and Whiteside et al. (2007).

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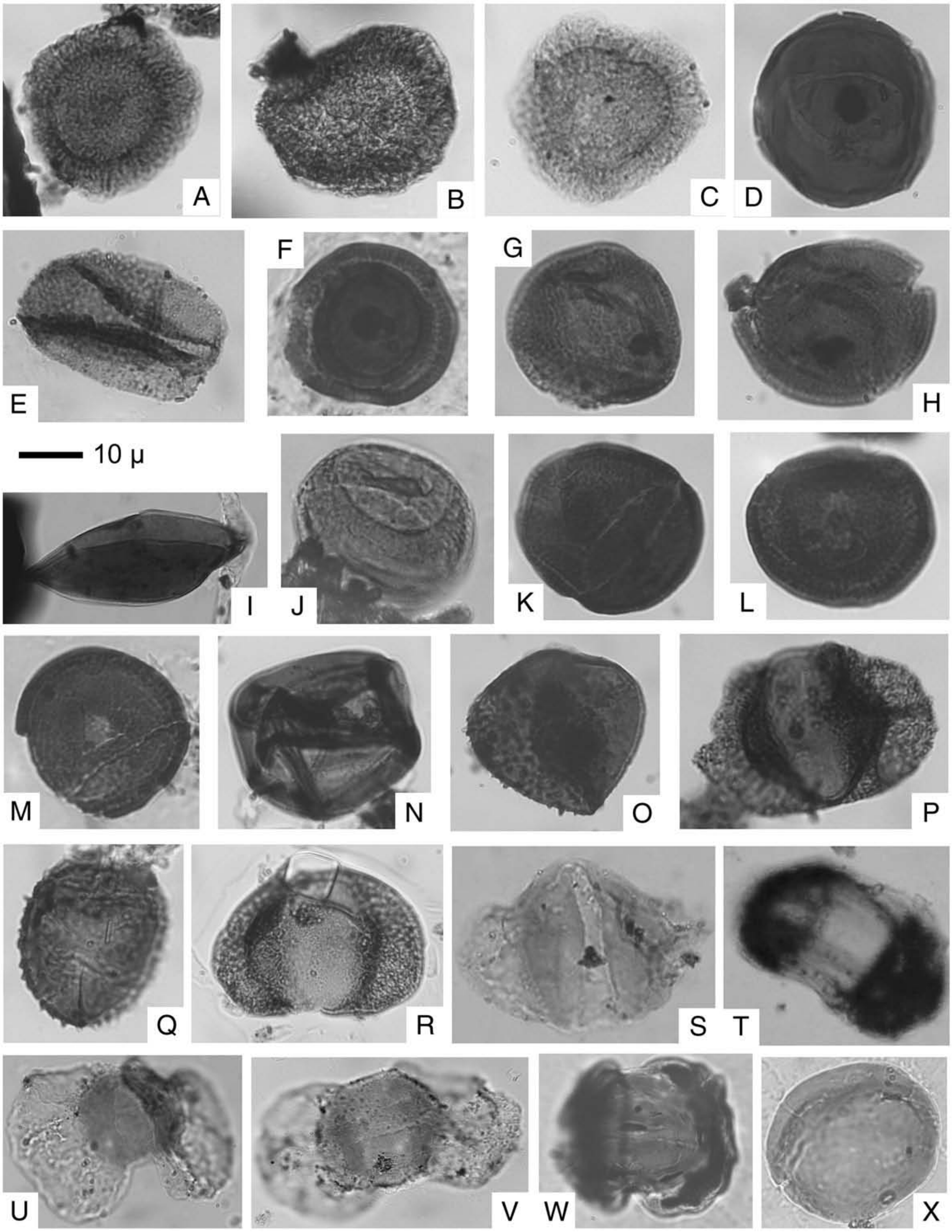

Fig. 5. Palynological assemblage from the Blomidon Formation (PI) and Scots Bay Member (SB); A) *Patinasporites densus* (Leschik) Scheuring 1970, sample PI45(4) England Finder coordinates (E.F.c.) S36(1), B) *P. densus*, PI55(4) E.F.c. W36(3), C) *P. densus*, PI55(4) E.F.c. K42, D) *Gliscopollis meyeriana* (Klaus) Venkatachala 1966, PI45(4) E.F.c. R33, E) *Conbaculatisporites mesozoicus* Klaus 1960, PI30(2) E.F.c. P34, F) *G. meyeriana*, PI45(4) E.F.c. O24(1), G) *Classopollis torosus* (Reissinger) Couper 1958, PI45(4) E.F.c. S46, H) *C. torosus*, PI45(4) E.F.c. R29(4), I) *Cycadopites* sp., PI55(3) E.F.c. N44(1), J) *C. torosus*, PI55(3) E.F.c. H28(4), K) *C. torosus*, PI45(4) E.F.c. L42(1), L) *Corollina murphyae* Cornet et Traverse 1975, PI45(4) E.F.c. L50, M) *C. murphyae*, PI45(4) E.F.c. X38(1), N) *Calamospora mesozoica* Couper 1958, PI45(4) E.F.c. P38(1), O) *Converrucosisporites cameroni* (de Jersey) Playford and Dettmann 1965, PI45(4) E.F.c. U40(3), P) *Alisporites parvus* de Jersey 1962, PI45(4) E.F.c. J31, Q) *Carnisporites spiniger* (Leschik) Morbey 1975, PI20(4) E.F.c. V33, R) *A. parvus*, PI45(3) E.F.c. V52(2), S) *A. parvus*, SB2(2) E.F.c. H40(3), T) *Lunatisporites rhaeticus* Leschik, 1955, SB2(1) E.F.c. G47(1), U) *Platysaccus* sp., SB2(2) E.F.c. F33(3/4); V) cf. *Lunatisporites acutus*, SB2(2) E.F.c. L34(1); W) *Lunatisporites acutus* (Leschik) Scheuring, 1970, SB2(2) E.F.c E44(3/4); X) undetermined circumpolloid, SB2(2) E.F.c L39(2).

particularly *Alisporites*. The bisaccates are absent in the 20-cm interval below the basalt, but are again present in the Scots Bay Member above the North Mountain Basalt. Less common are *Calamospora mesozoica* (Fig. 5N), *Conbaculatisporites mesozoicus* (Fig. 5E), *Porcellispora longdonensis, Todisporites rotundiformis, Kraeuselisporites* sp., *Carnisporites spiniger* (Fig. 5Q) and *Cycadopites* sp. (Fig. 5I), all of which are present in only a few samples. Noteworthy is the presence of a group of trilete spores, such as *Converrucosisporites cameroni* (Fig. 5O), *Dictyophyllidites harrisii* and *Dictyophyllidites* sp., and other trilete spores below the presumed Tr–J boundary. In the other sections of the ENA basins, these

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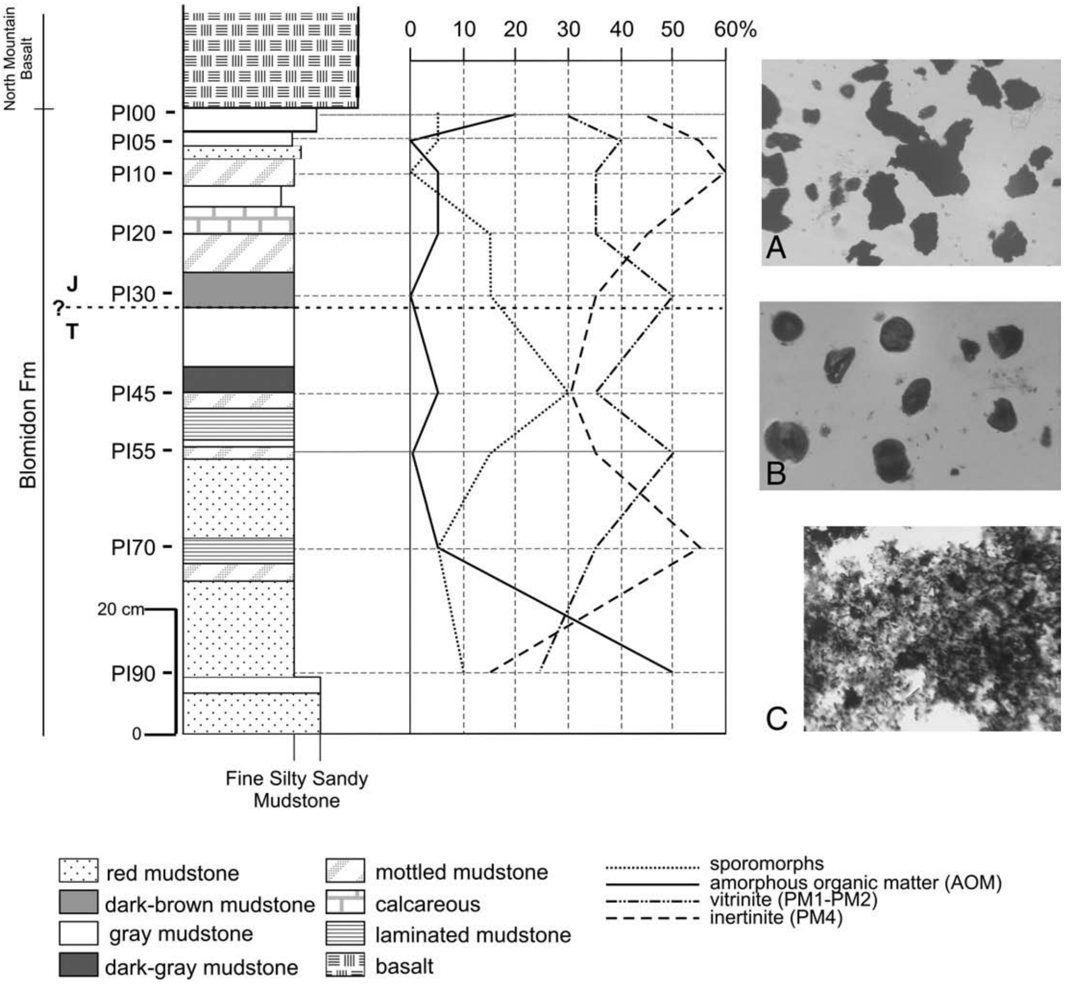

Fig. 6. Quantitative analysis of the palynofacies across the Blomidon Formation. Type and amount of organic debris show sensible variations through the section: palynofacies enriched in amorphous organic matter (C) are scarce and present only in the basal gray mudstone; sporomorphs (B) are present, although in different percentages, in the whole succession with a peak in correspondence of a dark gray mudstone, a few cm below the presumed Tr–J boundary of Whiteside et al., 2007; palynofacies dominated by intertinite (A) are common, showing a peak in abundance in the lower and upper part of the sampled section, where sporomorphs are absent and/or strongly degraded and oxidised.

fern spores characterize the fern spike just above the palynological boundary that has been interpreted as the floral turnover after mass-extinction.

Three samples of the Scots Bay Member (Fig. 4B) collected about 1 m above the top of the Margaretsville member basalt flow (the Brier Island Member is not present here) proved to be palynologically productive. Sample SB2 yielded the best preserved and most significant assemblage. The three samples yielded a palynological assemblage dominated by bisaccate pollens, e.g *Lunatisporites acutus* (Fig. 5W), *L. rhaeticus* (Fig. 5T), *Lueckisporites* sp., *A. parvus* (Fig. 5S), *Klausipollenites* sp., *Platysaccus* sp. (Fig. 5U) and other bisaccates in association with *C. mesozoica* and minor specimens of *G. meyeriana*, *C. torosus* and other circumpolloid (Fig. 5X). In general, the Circumpolles group is less abundant in the Scots Bay samples than in the Blomidon Formation samples.

5.2. Palynofacies

Palynofacies analysis addresses the potential for depositional and diagenetic control of OM preservation in the vicinity of the previously interpreted Tr–J boundary; i.e. rather than resulting from a catastrophic mass extinction event, is the observed palynological turnover an artifact of preservation? The type and amount of organic debris in the Blomidon Formation were measured quantitatively (Fig. 6). The lower total OM content of the Scots Bay Member prevented a detailed quantitative analysis, although a semi-quantitative estimation was possible. The palynofacies consist of terrestrial elements with high concentrations of palynomacerals, including debris of higher plants with minor amorphous organic matter (AOM). The type and amount of organic debris 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 gray-reddish laminated mudstone, gray 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 gray mudstone (PI-45) (Fig. 6B), which was likely deposited in low-oxygen conditions that favoured

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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 Blomidon Formation, and perennial lacustrine in the basal Scots Bay Member. The presence of abundant AOM within the darker intervals of the Blomidon Formation indicates that rarely transient conditions of very temporary nature made higher OM preservation possible.

5.3. Thermal alteration index (TAI)

Colour and degree of preservation of the OM were studies in order to determine the thermal history of the sampled section and to evaluate the possibility that the sporomorph content was recycled from older, thermally overmature strata (see online material). Given the nature of the depositional environment of the studied succession (i.e. fluvial–lacustrine), the potential for concentrating reworked palynomorphs from older strata was considered high. Fortunately, the transported or reworked fraction of OM can be distinguished by its opacity and darker colour (Traverse, 2008).

The Partridge Island section affords the opportunity to examine colour changes over a relatively short stratigraphic/time interval, thus minimizing evolutionary and/or environmentally driven changes in the sporomorph associations, and yielding an internally consistent dataset. The TAI does not reveal a significant trend in colour variation for most of the studied section (Fig. 6). The colour of the sporomorphs ranges from orange to brown down section, which corresponds to a TAI value of approximately 2+ to 3− (on a five point scale), suggesting a medium thermal maturity (about 60°–65° for this type of OM). Closer to the section top (the basalt flow), from sample PI-05 to sample PI-00, a modest increase occurs, with TAI values around 3 to 3+ (about 70°–80°). Thus, sampling indicates a mild increase in thermal alteration that is easily attributed to thermal overprint during the basalt emplacement.

The three samples of the 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 (e.g. 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 (e.g. *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 *P. 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 vesicate sporomorphs, is considered to occur at the upper Norian–lowermost Rhaetian boundary of the Late Triassic rather than at the Tr–J boundary, as documented in numerous independently dated key sections (see online material, Fig. S3). The last occurrence of this sporomorph in the Northern and Southern Hemispheres in the late Norian to Early Rhaetian (Sevatian–Rhaetian boundary) suggests a late Norian–early Rhaetian age for the palynological event in the Newark Basin rather than a Tr–J boundary age, as proposed by some (Gradstein et al., 1994; Van Veen, 1995, Lucas and Tanner, 2007). The possibility of a floral provincialism controlling a slightly diachronous distribution in North American and Tethyan domains of the parent plants (Fowell and Olsen, 1993) does not seem to be valid given the large distribution of these taxa in different palaeoclimate belts (Buratti and Cirilli, 2007). Furthermore, the updated palaeogeographic reconstructions based on palaeomagnetism (Kent and Tauxe, 2005) discount the possibility of explaining the discrepancy by a latitudinal floral gradient. *P. densus* is present in numerous Triassic sections located in similar palaeoclimatic and palaeogeographic belts (e.g. Carnian formations in Worcester, Tewkesbury and Reddict districts; Barclay et al., 1997). It is notable that in Northern Italy (Southern Alps) the last occurrence of *P. densus* is recorded (Jadoul et al., 1994) several hundred meters below the Tr–J boundary based on biostratigraphic, stratigraphic and geochemical constraints (Galli et al., 2007). Correlation of the palynological assemblage across the putative Tr–J boundary in the Fundy and Newark basins (Whiteside et al., 2007) with the section at St. Audrie's Bay (Hounslow et al., 2004) indicates that these sections are quite different, in particular at the latter where *P. densus* is absent in the Triassic portion of the section below the Tr–J boundary. Clearly, this lack of correlation cannot be explained by a provincialism of this taxon.

6.2. Does the compositional variation of Circumpolles group reflect an early Jurassic palynological assemblage?

The miospores from the extinct conifer family Cheirolepidiacea (*Classopollis, Gliscopollis, Corollina*) first appeared in the Late Triassic and became major elements in Jurassic and Cretaceous palynoflora. The widespread distribution of *G. meyeriana* and *C. torosus* commences in the Norian and continues into the Rhaetian and the Jurassic, as documented in several locations in the Northern and Southern Hemispheres (e.g. Cornet and Traverse, 1975; Schuurman, 1979; Batten and Koppelhus, 2002; Warrington, 2002; Hounslow et al., 2004; Barrón et al., 2006; Buratti and Cirilli, 2007). Recent geochemical and biostratigraphic constraints document that *C. torosus* and *G. meyeriana* were already abundant within the Rhaetian (*Rhaetipollis–Limbosporites* zone of Barrón et al., 2006; Gómez et al., 2007; Kuerschner et al., 2007). Thus, the presence or relative abundance of these Circumpolles taxa does not justify assigning a Jurassic age to the host strata, so long as they are not part of an assemblage with only Jurassic forms. Furthermore, a valid biozone should be defined by one or more first occurrences rather than solely by disappearances. In the Blomidon Formation, as well as in the Scots Bay Member, a distinctly Hettangian species has not yet been recorded. Also significant is the absence of *Cerebropollenites thiergartii,* which is considered a marker for the basal Jurassic in numerous palynological zonations that also are age-constrained by ammonoids, conodonts and geochemical data (Kuerschner et al., 2007; Hillebrandt et al., 2007; Gómez et al., 2007).

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6.3. Does the presumed palynological turnover and related fern spike mark the Tr–J boundary?

It is essential to emphasize that the sediments investigated in this study were deposited in continental environments (e.g. fluvial and lacustrine), which are unfavourable to OM preservation. Consequently, the OM likely was subjected to strong and episodic oxidizing conditions, as indicated by the palynofacies analysis. In the Partridge Island section, for example, sporomorph abundance varies as a function of OM preservation, which itself is associated with variable conditions of depositional/diagenetic oxidation. Correlation of the sporomorph distribution with the palynofacies and TAI variations demonstrates that the low percentage or absence of sporomorphs is coincident with higher thermal indices and/or with more degraded-oxidized palynofacies (Figs. 4 and 6). Therefore, no significant palynological decline or turnover occurs below the North Mountain Basalt that can be attributed to a mass extinction event. This observation is also reinforced by the fact that except for *P. densus* and a few other sporomorphs (Fig. 4) no other last occurrences of Triassic sporomorphs have been recorded at the presumed Tr–J boundary. A few last appearances (e.g. *C. cameronii, D. harrisii* and *Dictyophyllidites* sp.) are only local and most likely palaeoecologically controlled, since they are present above the putative Tr–J boundary as important elements of the so-called fern spike in other ENA basins (Fowell and Olsen, 1993; Fowell et al., 1994). Although a true fern spike is absent at the Partridge Island site, a level enriched in fern spores is recorded within a dark clay mudstone (Fig. 4) located 15 cm below the putative Tr–J boundary, but not above the boundary as in the Newark Basin. The co-occurrence of climate sensitive facies (dark clay) with abundant fern spores could be related to local, more humid climate conditions rather than to a re-colonization after mass-extinction. During the Mesozoic, ferns developed easily in warm and humid environments (e.g. coal swamps, river banks), preferring sheltered areas under the forest canopy, along creeks and streams and other sources of permanent moisture (Abbink et al., 2004). Alternatively, if the CAMP eruptions were slightly diachronous (Jourdan et al., 2009) and the environmental effects of the eruptions included severe releases of SO_2, with consequent terrestrial and marine acidification (Tanner et al., 2007), the “fern spikes” may record short-term ecological perturbations that were regional in scope.

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

One of our most significant findings is the presence of a palynological assemblage dominated by bisaccates, such as *L. acutus, L. rhaeticus, Lueckisporites* sp. with minor Circumpolles, within the Scots Bay Member. The presence of specimens considered as Triassic taxa (e.g. Schuurman, 1979; Jadoul et al., 1994; Batten and Koppelhus, 2002; Warrington, 2002; Hounslow et al., 2004) led to assigning these strata to the Triassic, although they overlie the East Ferry and Margaretsville Members of the ~201 Ma North Mountain Basalt. The colour of these sporomorphs, which are comparable to and often slightly lighter than sporomorphs from the underlying Blomidon Formation, excludes the possibility that they might be recycled from older strata. Therefore, while we cannot constrain the age of the Brier Island Member (one or two basalt flows), which is discontinuously present along the Bay of Fundy and has not been observed in contact with the Scots Bay sediments, palynostratigraphy convincingly constrains the age of the East Ferry and Margaretsville Members of the North Mountain Basalt to the Late Triassic.

6.5. Is chron E23r at the Tr–J boundary and how long is the Rhaetian?

The Newark magnetostratigraphy shows that the reverse chron E23r occurs just below the interval previously interpreted as the Tr–J boundary (Kent and Olsen, 2000; Olsen et al., 2002). The new data from Partridge Island do not support this interpretation, given the presence of Triassic sporomorphs in strata overlying the North Mountain Basalt, hence above the supposed Tr–J boundary. Unless the Newark magnetostratigraphy is missing several chrons (cf. Gallet et al., 2007) between the last occurrence of *P. densus* and the first CAMP lava flow, E23r must actually occur well before the end of the Rhaetian. Hounslow et al. (2004), in fact, noted the ambiguity in correlating palaeomagnetic records of the Newark basin and the St. Audrie's Bay marine section, and suggested that E23r might correlate with SA5.2r, below the base of the Lilstock Formation, rather than to SA5r, in the lower Blue Lias. In addition, it is significant that the terrestrial components of the palynological assemblage at St. Audrie's Bay (e.g. lack of *P. densus* and presence of typical uppermost Triassic taxa) are quite different from the microflora found at the Partridge Island and also at the presumed Tr–J boundary in the Newark basins.

This interpretation of the magnetostratigraphy is in good agreement with recent biogeochemical and magnetostratigraphic data that indicate the supposed Tr–J boundary of the Newark basin actually coincides with the Sevatian–Rhaetian boundary (Gallet et al., 2007). However, this correlation and the results of the present biostratigraphic study, as well as radio-isotopic ages, raise problems regarding the duration of the Rhaetian. Considering that the age of the North Mountain Basalt is constrained to ~201 Ma (Schoene et al., 2006; Jourdan et al., 2009), i.e. indistinguishable from the age of the marine Tr–J boundary (Schaltegger et al., 2008), the time which elapsed between the last occurrence of *P. densus*, the eruption of CAMP basalts and the Tr–J boundary would be extremely short. Although at face value the U/Pb age of the North Mountain Basalt (201.3 ± 0.3 Ma) predates that (201.6 ± 0.3 Ma) for the Tr–J boundary only at relatively low probability, we note that the former (obtained on a mafic pegmatite sheet in the upper part of the East Ferry Member) may not date the earliest North Mountain Basalt eruptions, and the latter (being determined from a silicic tephra) is more likely to be affected by pre-eruptive residence time of the zircons, which can produce apparent age bias of several hundred ka (Simon et al., 2008). Furthermore, if the last occurrence of *P. densus* can be considered as a Norian–Rhaetian or very early Rhaetian event, this would imply a very short Rhaetian (<1 Ma). A short Rhaetian event suggested here concurs with that proposed by Krystyn et al. (2007) and Gallet et al. (2007), but contrasts markedly with conclusions of Channell et al. (1993), and Muttoni et al. (2004) who consider that the Rhaetian had a duration of 6 Ma. Alternatively, it might be considered that the stratigraphic record of the Newark and Fundy basins are incomplete and that the Rhaetian is partly or wholly missing, as suggested by Kozur and Weems (2005).

7. Conclusions

New palynological data from the Fundy Basin of Nova Scotia, Canada, constrain the age of the sedimentary rocks below and just above the ~201 Ma North Mountain Basalt, part of the widespread CAMP event, to the Late Triassic. In particular, this is documented in the Fundy Basin by the presence of Triassic sporomorphs in the Scots Bay Member, which overlies the East Ferry and Margaretsville Members of the North Mountain Basalt. Significantly, examination of the state of preservation and thermal alteration of OM associated with the fossil assemblage precludes the possibility of recycling of the Triassic sporomorphs from the older Blomidon Formation strata, which underlies the basalt. From these new data, we draw the following conclusions:

1. Because the last occurrence of *P. densus* occurs 40 cm below the North Mountain Basalt, and strata above the basalt (e.g. the Scots Bay Member) are clearly Triassic, it follows that the palynological definition of the Tr–J boundary based on the last occurrence of *P. densus* is not valid. Consequently, it appears that the occurrence of a fern spike in the Newark Basin probably reflects a regional environmental perturbation, e.g. an abrupt but short-lived episode of climate change and atmospheric acidification, rather than a global event of recolonization after mass exctinction.
2. Given that the previously accepted definition of the Tr–J boundary in the Newark Basin cannot be supported palynologically, it follows that

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chron E23r, which is correlated with the last occurrence of *P. densus* in the Newark Basin, also cannot be considered as a marker of the Tr–J boundary. Instead, the chron E23r event merely marks a Late Triassic (probably Rhaetian) reversal. This is an important point to consider in correlating the Late Triassic magnetostratigraphic data from marine sections (Hounslow et al., 2004; Gallet et al., 2007) to the Newark continental section.

3. The occurrence of clearly Triassic strata in sharp contact with underlying North Mountain Basalt lava flows clearly indicates that CAMP volcanism, at least in Nova Scotia, is of Triassic (Rhaetian) age. Significantly, this temporal relationship must also apply to other areas of CAMP. In Morocco, for example, initiation of the basaltic eruptions was coeval with deposition of sedimentary strata containing *P. densus*. Cyclostratigraphy constrains the onset of CAMP volcanism elsewhere in the ENA (e.g. the Fundy and Newark basins) to about 20–40 ky after the last occurrence of *P. densus* if no unconformities are present.
4. Because the eruption of the basalt plus deposition of at least the first few meters of the Scots Bay Member occurred in a time span shorter than analytical resolution of the geochronological data, emplacement of the basalt took place in a very short span of time.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2009.07.021.

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