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ARTICLE

Sedimentology of the lower Serpukhovian (upper Mississippian) Mabou Group in the Cumberland Basin of eastern Canada: tectonic, halokinetic, and climatic implications

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Abstract: The Visean–Serpukhovian transition in Atlantic Canada was marked by a general humidification of the climate as the region drifted towards equatorial latitudes. It also corresponds to a time when ice volume was increasing on Gondwana, which marked the end of Mississippian marine incursions in the region. Glacioeustatic fluctuations of greater magnitude are thought to have increased the response of the regional climate to third-order cyclicity from orbital forcing. In the Cumberland Basin, fluvial grey beds of the lower Serpukhovian Shepody Formation were deposited in sub-humid conditions during highstands, whereas red playa deposits of the same unit were deposited under semi-arid conditions during lowstands. Basin reconstruction suggests that this unit was sourced from the fault-bounded Cobequid and Caledonia highlands and deposited in two separate salt-withdrawal minibasins. This fluvial system was seemingly discharging to the north into the broad lake that deposited the contemporaneous Hastings Formation. A disconformity separates the Shepody Formation from mid-Serpukhovian red beds of the Claremont Formation and is tentatively associated with another increase in ice volume on Gondwana followed by a recrudescence of fault activity and basin subsidence. A prolonged time of aridity, floral crisis, non-deposition, deep weathering and karstification in late Serpukhovian to early Bashkirian times is contemporaneous with abundant glacial deposits in higher latitudes, suggesting that globally low sea levels may have been at play in creating a situation of greater continentality in the study area.

Résumé : La transition du Viséen au Serpukhovien dans le Canada atlantique fût marquée par une humidification générale du climat alors que la région dérivait en direction des latitudes équatoriales. Cette époque correspond à un temps durant lequel le volume des glaces était croissant sur Gondwana, mettant ainsi fin aux incursions marines mississippiennes dans la région. L'amplitude croissante des fluctuations glacioeustatiques à l'époque aurait rendu le climat régional plus sensible aux cycles orbitaux de troisième ordre. Dans le bassin de Cumberland, les lits fluviaux gris de la formation de Shepody (Serpukhovien inférieur) se seraient déposés dans des conditions sub-humides durant des temps de hauts niveaux marins, tandis que les dépôts de playa de la même unité se seraient déposés dans des condition semi-arides durant des temps de bas niveaux marins. Cette unité se serait déposée à partir des hautes-terres de Cobéquid et de Calédonia à l'intérieur de deux mini-bassins distincts. Ce système fluviatile aurait connecté au nord à un grand lac dans lequel se serait déposée la formation contemporaine de Hastings. Une discordance d'érosion sépare la formation de Shepody des lits rouges de la formation de Claremont (Serpukhovien moyen) et corresponderait à une autre recrudescence des glaces sur Gondwana suivie par une augmentation des taux de subsidence et d'activité de failles. Des temps prolongés d'aridité, de crise florale, de non-déposition, d'altération profonde et de karstification durant le Serpukhovien supérieur et le Bashkirien inférieur sont contemporains à des dépôts glaciaires abondants en hautes latitudes, suggérant qu'un bas niveau marin global aurait pu aggraver les conditions de continentalité dans la région étudiée.

Introduction

In early Visean times, the composite Maritimes Basin of eastern Canada (Fig. 1; inset) accommodated an approximately 1 km thick succession of evaporites in many of its subbasins, including ~600 m of salt (Gibling et al. 2008). Because of its position in the most tectonically active zone of the composite basin, it is estimated that the Cumberland Basin of southern New Brunswick and northwest Nova Scotia (Fig. 1; inset) accumulated 2–3 km of salt and was characterized by the early onset of salt tectonics in late Visean times (Waldron et al. 2013; Jutras et al. 2015), whereas evidence of pre-Pennsylvanian salt tectonics is lacking in other subbasins of the Maritimes Basin.

Well-exposed upper Visean units in the western part of the Cumberland Basin allowed a detailed reconstruction to be made of its minibasins at that time interval (Jutras et al. 2015), setting the stage for our understanding of the more poorly exposed lower Serpukhovian units in that basin, which are here studied in detail for the first time. Significant differences between early (Visean) and late (Pennsylvanian) halokinetic settings were already identified (Jutras et al. 2015), and part of this study is to verify if minibasin dynamics were already changing in Serpukhovian times.

In northeast and central Nova Scotia, the early Serpukhovian interval is characterized by a succession of fine-grained continental clastic rocks that concordantly overlie the uppermost marine rocks of the Visean Windsor Group (Dawson 1873) and that were assigned to the Mabou Group by Belt (1964, 1965; Fig. 2). Hamblin (2001) studied in detail 17 outcrop sections distributed around northern Nova Scotia and produced a sedimentologic

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Fig. 1. Geology of the study area (modified from Ryan and Boehner 1994 and St. Peter and Johnson 2009), and studied localities. Inset: lateral extent of the Maritimes Basin, with close-up of the Cumberland Basin.

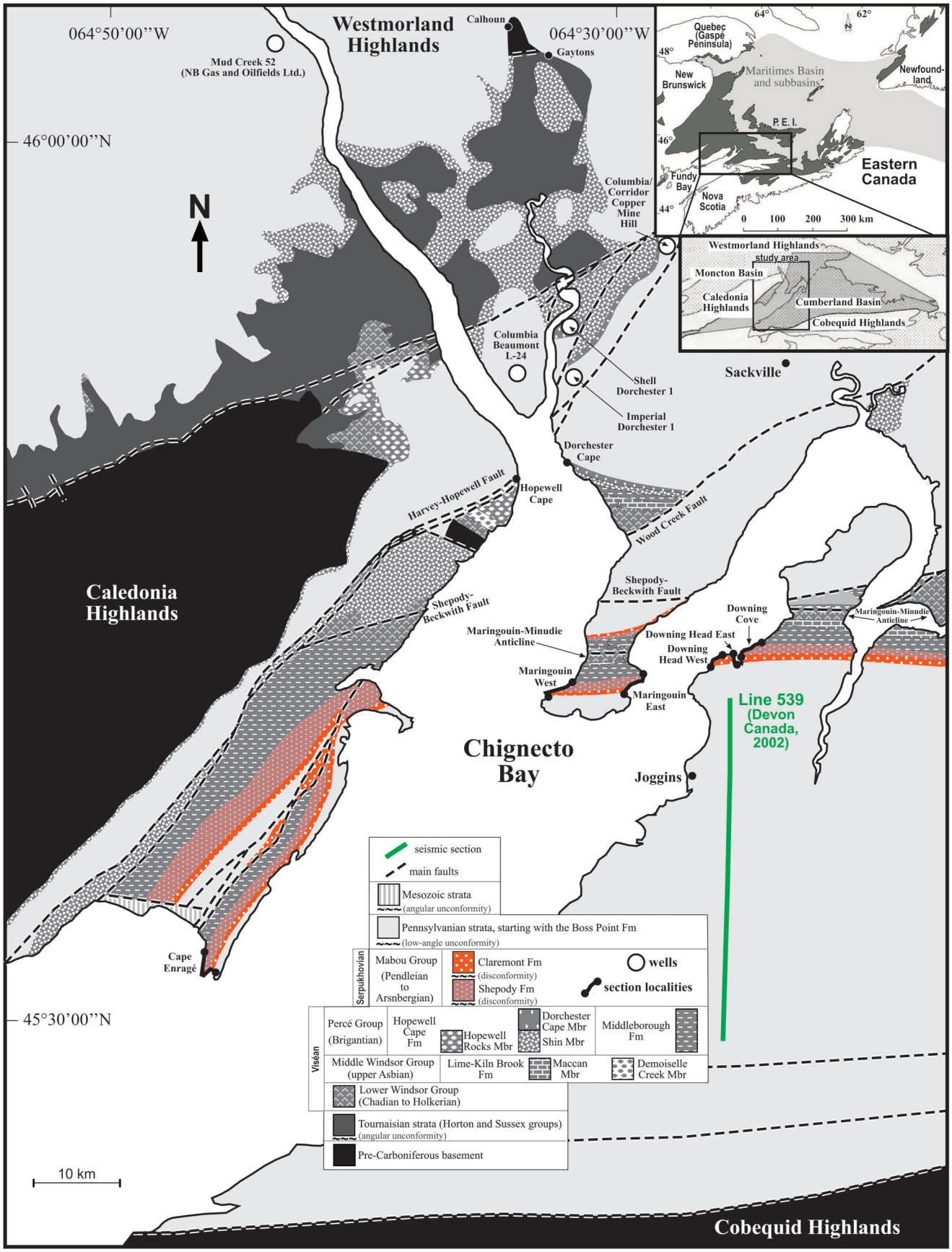
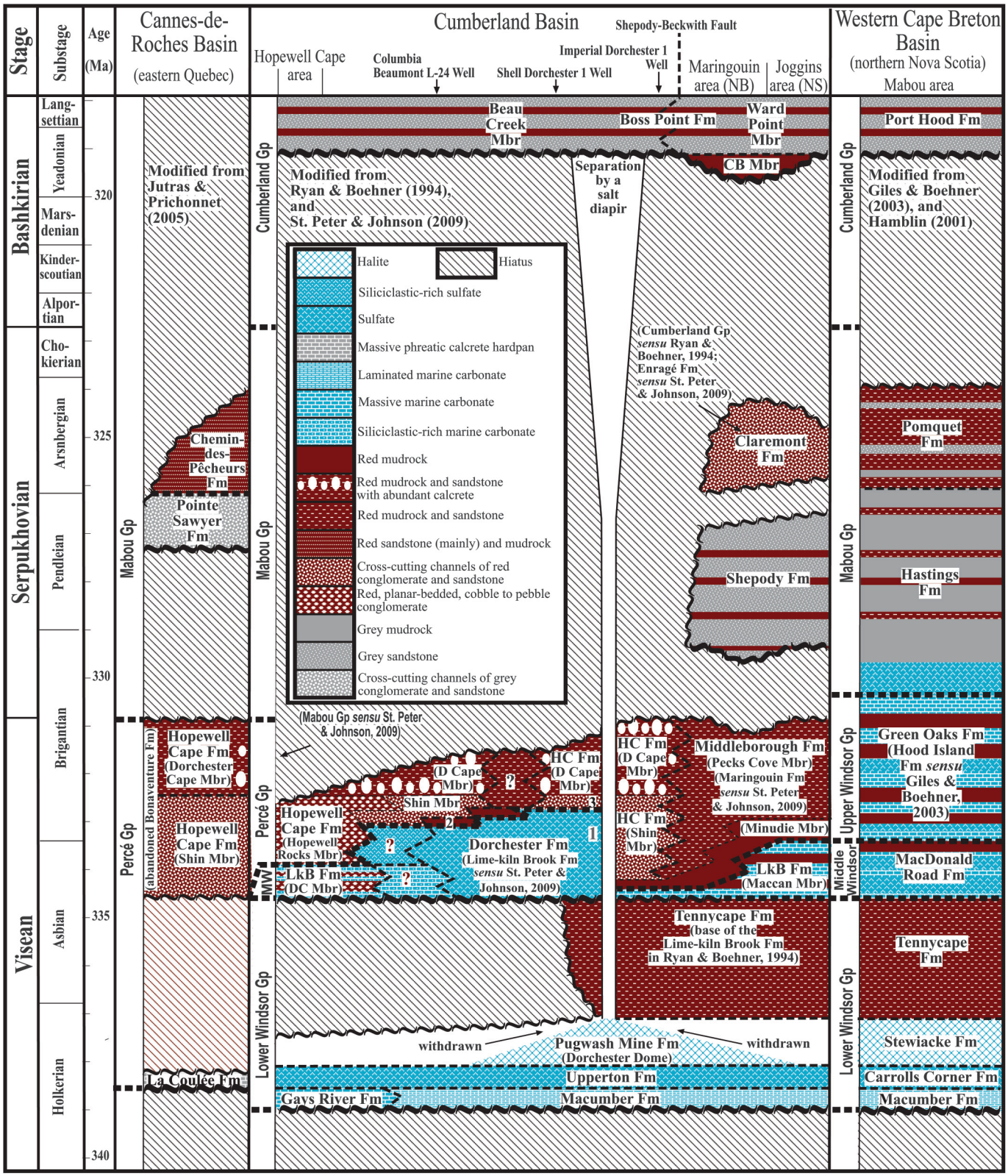


Fig. 2. Middle to upper Mississippian stratigraphy in the study area and adjacent areas of eastern Quebec, New Brunswick, and Nova Scotia (time scale after Richards 2013). CB, Chignecto Bay; D Cape, Dorchester Cape; HC, Hopewell Cape; LkB, Lime-kiln Brook; MW, Middle Windsor Group.



and paleogeographic model for units of this group in that general area. Of these sections, only one is from the Cumberland Basin, where the facies of Mabou Group units substantially differ from those found in the rest of northern Nova Scotia. This section is in the Joggins area (Fig. 1) and was also included by Allen et al. (2013) in their study of the ~7 km thick upper Palaeozoic section at Joggins.

Hamblin (2001) identified overlapping orders of sedimentary cyclicity related to variations in base level in the fine, dominantly lacustrine facies of lower Serpukhovian units in northeast Nova Scotia, whereas St. Peter and Johnson (2009) recognized only a gradual coarsening-upward pattern in the coarser, dominantly fluvial facies of equivalent units in southern New Brunswick. According to Giles (2009), the upper Visean succession in Nova Scotia was already affected by glacioeustatic variations related to the growth and decay of ice sheets on Gondwana, the amplitude of which increased at the Visean–Serpukhovian transition (Rygel et al. 2008). In this paper, lower Serpukhovian units are studied in the western part of the Cumberland Basin, where they are best exposed, to reconstruct the paleogeography of that basin and to evaluate the respective contribution of fault tectonics, halokinesis, and glacioeustatic variations at that interval. This study also includes insights on the late Serpukhovian to early Bashkirian paleogeography of the western part of the Cumberland Basin based on clues provided by eodiagenetic overprints in the upper part of the lower Serpukhovian succession.

Geological setting

The Cumberland Basin is part of the composite Maritimes Basin of eastern Canada, which developed during the Late Devonian in the aftermath of the Early to Middle Devonian Acadian Orogeny (Calder 1998; Gibling et al. 2008), when orogenic deformation migrated towards New England (Jutras et al. 2003). The Late Devonian to Mississippian interval is characterized by continental deposits in small strike-slip basins with local sources, except for the Visean portion, which includes epicontinental sea deposits (Calder 1998; Gibling et al. 2008). As in the rest of the composite Maritimes Basin, an unconformity is found between Mississippian and Pennsylvanian units in the Cumberland Basin. This unconformity marks the first pulse of Alleghanian deformation in the Maritimes Basin, which developed in association with the final closure of the Rheic Ocean to the south (Faure et al. 1996; Jutras et al. 2003, 2005). According to these authors, this resulted in the clockwise rotation of paleostresses in Atlantic Canada during early Pennsylvanian times (sensu Richards 2013), which generated transpressive deformation along the pre-existing fault system. In the Cumberland Basin, this translated as the tilting and partial erosion of Mississippian strata. Subsequent Pennsylvanian deposition was in part sourced from the developing Alleghanian Orogen in the eastern USA at the time (Gibling et al. 1992, 2008). Starting in the late Visean and extending into the rest of the Carboniferous, Maritimes Basin deposition was influenced by both halokinesis (Waldron et al. 2013) and glacioeustasy (Giles 2009). Of the two salt structures that are known in the study area, the Dorchester Salt Dome (reached by wells Imperial Dorchester 1, Shell Dorchester 1, and Copper Mine Hill) was active in late Visean times, whereas the south-southwest-striking Maringouin–Minudie Anticline was still inactive (Jutras et al. 2015; Fig. 1).

Stratigraphic framework

Visean units below the Mabou Group

The succession spanning from the first to the last Visean marine carbonate in eastern Canada was included within the Windsor Group by Bell (1929), whereas the overlying remainder of the Mississippian succession was originally assigned to the abandoned Canso Group of Bell (1944), but later included within the Mabou Group by Belt (1964, 1965). Hamblin (2001) later refined the definition of the Mabou Group by excluding from it the upper Visean red beds of the Middleborough Formation (Maringouin Formation in New Brunswick), which overlie the uppermost Windsor Group carbonate in northwest Nova Scotia, but which are not found in the Mabou Group type area of Cape Breton Island, northern Nova Scotia, where Windsor Group carbonates reach higher into the Visean (Fig. 2). In eastern Quebec, where the Visean interval is only characterized by continental clastics, similar and contemporaneous red beds were assigned to the Percé Group (Jutras and Prichonnet 2005), which is thought to be time equivalent to the entire Windsor Group.[1]

Based on precedence, we here use the term Middleborough Formation of Bell and Norman (1938) over the Maringouin Formation of Norman (1941) in reference to fine, planar-bedded uppermost Asbian to Brigantian red beds that underlie the Mabou Group (sensu Hamblin 2001 and Jutras et al. 2015) in central areas of the Cumberland Basin (Fig. 2). These beds grade laterally into coarse red beds of the Hopewell Cape Formation (Percé Group) towards source areas (Jutras et al. 2015; Fig. 2). The lower part of this formation is characterized by coarse, planar-bedded fan delta conglomerates in the northwest sector of the Cumberland Basin (Hopewell Rocks Member), and by cross-cutting channels of conglomerate and sandstone in other sectors of the basin (Shin Member; Fig. 2). The upper part of the Hopewell Cape Formation is characterized by finer red beds with abundant calcretes (the Dorchester Cape Member of St. Peter and Johnson 2009; Fig. 2).

Mabou Group units

In the most comprehensive study on the Mabou Group to date, Hamblin (2001) recognized only two units within that group, namely the grey and red mudrock with minor sandstone and carbonate of the upper Brigantian to Pendleian Hastings Formation (Rostoker 1960), and the red and minor grey mudrock and sandstone of the Arsnbergian Pomquet Formation (Belt 1965; Fig. 2; see "Palynology" for age constraints). Although the Serpukhovian interval is much coarser in the Cumberland Basin, Hamblin (2001) recognized a lateral consistency with contemporaneous successions of northern and central Nova Scotia, where a dominantly grey basal interval and a dominantly red upper interval are also found. Hence, this author still referred to these coarser stratigraphic units as, respectively, the Hastings and Pomquet formations. However, most other workers have been referring to these coarser equivalents as, respectively, the Shepody and Enragé formations of Norman (1941) in southern New Brunswick (e.g., St. Peter and Johnson 2009 and references therein), and as, respectively, the Shepody and Claremont (Bell and Norman 1938) formations in northwest Nova Scotia (e.g., Ryan and Boehner 1994; Waldron et al. 2013). Because the Enragé Formation is not exposed at the type locality of Cape Enragé (Fig. 1), and because the Claremont Formation has precedence, the latter appellation is herein used along with the term Shepody Formation (Fig. 2). We therefore refer to the grey-bed-dominated base of the Mabou Group as the Hastings Formation when mainly characterized by mudrock, but as the Shepody Formation when the grey bed intervals are

[1]Note: Despite these recent developments in stratigraphic nomenclature, many workers (e.g. St. Peter and Johnson 2009; Allen et al. 2013; Waldron et al. 2013) still lump units of the Percé and Mabou groups (sensu Hamblin 2001, and Jutras and Prichonnet 2005) within an expanded Mabou Group that includes Visean red beds of the Hopewell Cape and Middleborough formations.

mainly sandy; and refer to the red-bed-dominated upper part of the Mabou Group as the Pomquet Formation when mainly characterized by alternations of sandstone and mudrock, and as the Claremont Formation when mainly characterized by alternations of sandstone and conglomerate.

Basal Cumberland Group units

Throughout Atlantic Canada, the Mabou Group is unconformably overlain by the lower Pennsylvanian Cumberland Group (Bell 1944; Fig. 2). In the Cumberland Basin, the base of this group is occupied by the Boss Point Formation, which has been separated into several members by St. Peter and Johnson (2009). In most of the study area, the base of the Boss Point Formation is occupied by grey sandstone of the Ward Point Member or of the laterally equivalent and petrographically similar Beau Creek Member; but in the southwest part of the study area, these members are underlain by fine red beds of the Chignecto Bay Member (St. Peter and Johnson 2009; Fig. 2).

Studied sections

Six coastal sections in which rocks of the Serpukhovian interval are well represented were measured across the Cumberland Basin at Cape Enragé, Maringouin West, and Maringouin East, New Brunswick (Figs. 1, 3), and at Downing Head West, Downing Head East, and Downing Cove, Nova Scotia (Figs. 1, 4). The sections are positioned in line with stratigraphic markers high in the Shepody Formation (base of G1b in Fig. 3, and base of G1a in Fig. 4; see "Air-photo analysis"), and are measured from the base of this unit (Figs. 3, 4). Also included in this study to evaluate post-depositional erosion are key sections or localities in which thin remnants of the Mabou Group (the Dorchester Cape and Well Mud Creek 52 wells; Fig. 1) or older rocks (the Hopewell Cape, Calhoun, and Gaytons sections, and the Shell Dorchester 1, Imperial Dorchester 1, Columbia Beaumont L-24, and Columbia/Corridor Copper Mine Hill F-88-2329 wells; Fig. 1) are unconformably truncated by lower Pennsylvanian rocks.

Palynology

The Shepody Formation

Based on our new data, the lower part of the Shepody Formation straddles the *Schopfipollenites acadiensis – Knoxisporites triradiatus* (AT) and *Grandispora spinosa – Ibrahimispores magnificus* (SM) concurrent range zones (Figs. 3, 4), which are, respectively, Brigantian (late Visean to early Serpukhovian) and late Brigantian to Pendleian (early Serpukhovian; Utting and Giles 2004). According to recent estimations, the Brigantian substage itself straddles the Visean–Serpukhovian boundary (Richards 2013). Hence, the SM zone can be assumed to be entirely Serpukhovian, and the AT–SM transition possibly corresponds to the Visean–Serpukhovian transition.

At Cape Enragé, Maringouin West, and Downing Cove, basal beds of the Shepody Formation bear an SM assemblage (Figs. 3, 4) based on the diagnostic presence of *Ibrahimispores magnificus* within those beds, along with abundant *Crassispora trychera* and *Punctatsiporites glaber*, with *Auroraspora magna*, *Calamospora* sp., *Colatispora decorus*, *Discernisporites micromanifestus*, *Rugospora minuta*, *Schopfites claviger*, *Shopfipollenites acadiensis*, *Spelaeotriletes windsorii*, and *Spelaeotriletes bellii*. At Maringouin East (Fig. 3), basal beds of that unit bear a similar assemblage, but without *Ibrahimispores magnificus*, suggesting an AT affinity (Utting and Giles 2004). However, we acknowledge that a relative rarity of diagnostic taxa may result in some cases in applied zonal definitions that are tentative. Therefore, the absence of *Ibrahimispores magnificus* in basal Shepody beds at Maringouin East does not in itself preclude a late Brigantian or even Pendleian age for these beds. This is well demonstrated by the lack of this diagnostic taxon in a sample from near the top of the Shepody Formation at Cape Enragé, between intervals that do include that taxon (Fig. 3). Moreover, although *Ibrahimispores magnificus* was not found in several samples from basal grey beds of the Hastings Formation in the Mabou type area of Cape Breton Island (northern Nova Scotia), with the rest of the spore assemblage suggesting that they belong to the AT zone (Utting 2011), this taxon was later found in beds from the underlying uppermost Windsor Group, suggesting that the AT–SM boundary is significantly lower than previously thought in that section (Opdyke et al. 2014). At Maringouin East, the greater measured thickness of Shepody Formation beds below stratigraphic markers higher in the succession (Figs. 3, 4) supports the possibility that the most basal occurrences of this unit may be slightly older at that locality than at other localities, but assignment to the AT zone may nonetheless be erroneous.

In summary, a large amount of uncertainty remains regarding the age of beds bearing an AT assemblage, which could be either latest Visean or early Serpukhovian, but those bearing an SM assemblage can be more safely considered to be early Serpukhovian. In this study, because they are distinguished by significant environmental changes, rocks of the Percé and Mabou groups are presumed to be entirely Visean and Serpukhovian, respectively (Fig. 2), although this is neither demonstrated nor refuted by palynology at this stage.

The Hastings Formation

A similar assemblage, without *Ibrahimispores magnificus*, was identified by Dolby (1997) from grey mudrock cuttings of the Mabou Group in the Mud Creek 52 Well of New Brunswick Gas and Oilfields Ltd. (Fig. 1), ~4 m below basal Pennsylvanian strata (the Beau Creek Member of the Boss Point Formation, sensu St. Peter and Johnson 2009), and above a thin succession of Dorchester Cape Member red beds that onlaps basement rocks of the Westmorland Highlands (Fig. 1). Despite the absence of *Ibrahimispores magnificus*, Dolby (1997) considered the assemblage to be Serpukhovian. Considering the findings of Opdyke et al. (2014), we concur with this assessment. This contact is also observed at Dorchester Cape (Fig. 1), where small lenses (less than 30 cm thick and 1 m wide) of grey mudrock with a late Brigantian to Pendleian spore assemblage (SM zone) are found below the very low angle unconformity that separates the Dorchester Cape Member from the basal grey sandstones of the Pennsylvanian Beau Creek Member of the Boss Point Formation (Jutras et al. 2015). These thin mudrock occurrences above the Dorchester Cape Member of the Hopewell Cape Formation are interpreted to be small remnants of the Hastings Formation. As suggested by the high amount of reworked Brigantian to Pendleian spores in the basal beds of that unit at Dorchester Cape (Dolby 1996), the Hastings Formation was largely eroded from the northern sectors of the study area and incorporated within the lower Pennsylvanian Boss Point Formation.

The Claremont Formation

Utting et al. (2010) reported an SM assemblage for the base of the Claremont Formation at Downing Head. However, these spore-bearing beds are below the unconformity that marks the sharp boundary between the Shepody and Claremont formations (Figs. 3, 4) and are therefore no longer considered to belong to the latter formation.

In the type area of the Mabou Group, the top of the grey-bed-dominated Hastings Formation reaches into the *Reticulatisporites carnosus* Concurrent Range Zone (late Pendleian to Arsnbergian) based on the presence of *Dictyotriletes castaneaeformis* and *Savitrisporites nux*, and all productive samples within the overlying red-bed-dominated Pomquet Formation also bear this assemblage (Utting 2011; Opdyke et al. 2014). Hence, based on stratigraphic affinities with the Pomquet Formation of northern Nova Scotia (Hamblin 2001), the barren Claremont Formation is inferred to be upper Pendleian to Arsnbergian. To facilitate communication, we here refer to the Hastings and Shepody formations as upper Brigantian

Fig. 3. Serpukhovian interval at Cape Enragé, Maringouin West, and Maringouin East, New Brunswick, with palynological localities (Geological Survey of Canada samples "C-" numbers) and results indicated on the left side of the columns. Note: "Brigantian" spore dates refer to the *Schopfipollenites acadiensis – Knoxisporites triradiatus* (AT) Concurrent Range Zone assemblage, whereas "Pendleian" refers to the late Brigantian to Pendleian *Grandispora spinosa – Ibrahimispores magnificus* (SM) Concurrent Range Zone (Utting et al. 2010). Legend in Fig. 4. gran., granular; peb., pebble.

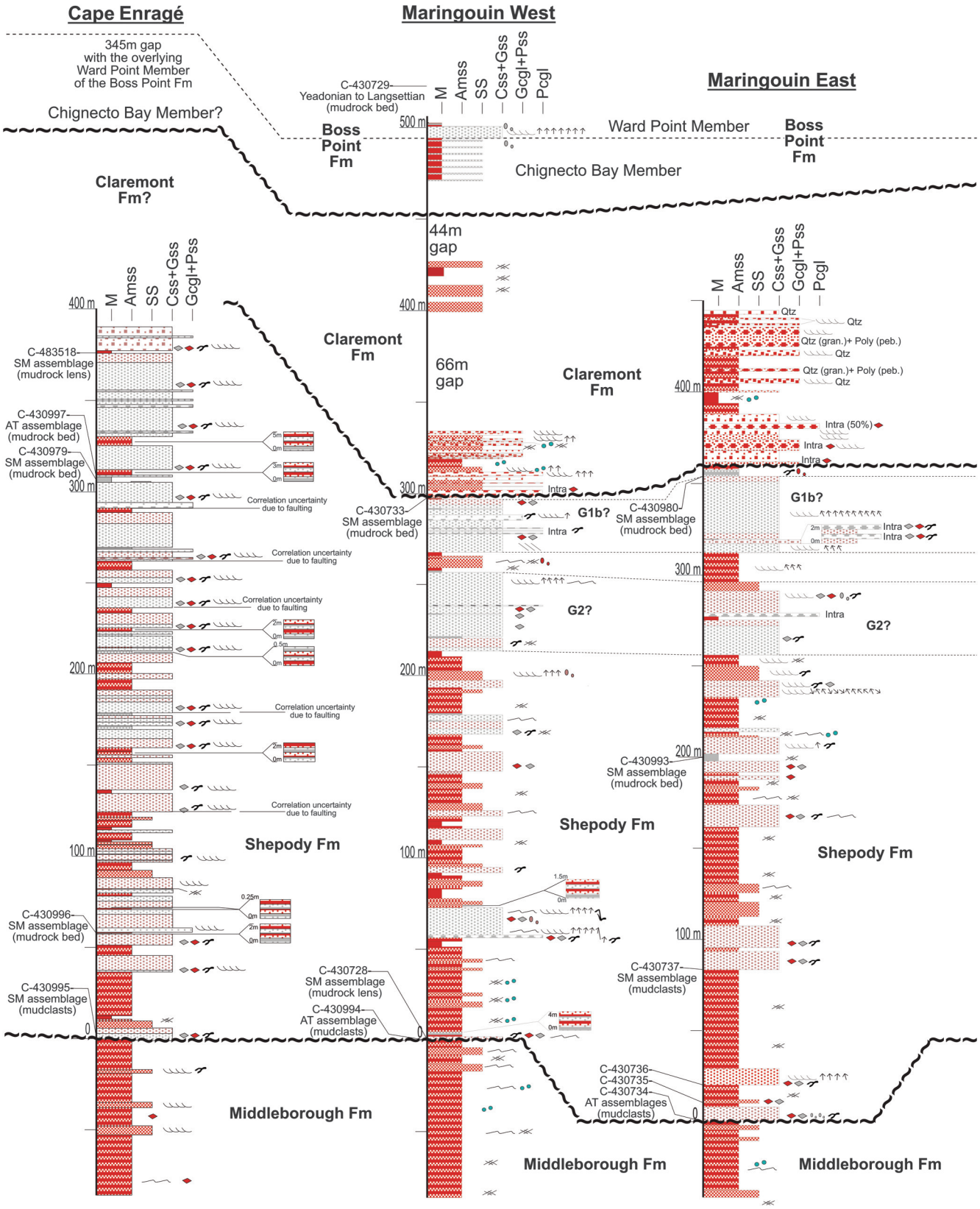


Fig. 4. Serpukhovian interval at Downing Head West, Downing Head East, and Downing Cove, Nova Scotia. See note regarding spore dates in the caption of Fig. 3.

to Pendleian, and the Pomquet and Claremont formations as Arsnbergian.

The Boss Point Formation (Cumberland Group)

In the four sections where productive samples near the base of the Cumberland Group (Boss Point Formation) could be retrieved (the Dorchester Cape, Maringouin West, and Downing Head East section), the presence of *F. pumicosus*, *P. elegans*, *Florinites*/*Potonieisporites* complex, and *Wilsonites* sp. suggest a Yeadonian to Langsettian age (lower Pennsylvanian; *Raistrickia saetosa* Concurrent Range Zone) for these beds (Figs. 3, 4). Hence, the hiatus between the Claremont and Boss Point formations spans the Chokierian, Alportian, Kinderscoutian, Marsdenian, and possibly the Yeadonian.

Sedimentology of Mabou Group units in the western Cumberland Basin

The Shepody Formation

The contact between the Middleborough and Shepody formations is sharp and disconformable throughout the west-central Cumberland Basin. The lowermost coarse grey sandstone bed of the Shepody Formation is observed to deeply downcut into the fine underlying red beds, especially at Cape Enragé (Fig. 5A) and at Maringouin West.

The Shepody Formation at all localities consists of a repetitive succession of fining-upward sequences that start with grey or red, coarse to granular sandstone with plant remains and grey or red mud clasts, followed by fine to tightly interbedded red mudrock and fine sandstone, or simply mudrock (Figs. 3, 4). Moving up section, the coarse sandstone beds that begin each fining-upward sequence become thicker, more abundant, and more often thoroughly grey (Figs. 3, 4). At all localities, these grey sandstone beds occasionally contain thin lenses of grey, intraformational pebble conglomerate (Figs. 3, 4). The sandstone beds are mostly cross-channelized, whereas the red intervals are planar-bedded and display moderate pedogenic overprints in the form of mottling, root traces, and minor calcitic nodules (Figs. 3, 4). The intraformational conglomeratic lenses and fossilized plant remains become increasingly abundant, moving west towards the Cape Enragé section (Figs. 3, 4).

The Shepody Formation at Maringouin West is ~60 m thinner than at Maringouin East, although only ~8 km separate the two sections (Figs. 1, 3). This is possibly explained by older occurrences of the Shepody Formation facies at Maringouin East, which bears an AT spore assemblage in basal beds at that locality, whereas the first occurrence of that facies bears an SM spore assemblage at Maringouin West (Fig. 3). As expressed in the palynology section, there are good reasons to doubt that the basal beds at Maringouin East truly belong to the AT zone, but they nonetheless appear to represent an earlier infill in a deeper part of the disconformable surface than that below the Maringouin West and Downing Cove sections. Overall, the thickness of the Shepody Formation increases from east to west, evolving from a full section of ~280 m at Downing Cove to an incomplete section of ~390 m at Cape Enragé (Figs. 1, 3, 4). It should also be noted that the Cape Enragé section is offset by numerous faults, with displacement along these faults on the order of potentially tens of metres. Because displacements exceed the depth of the cliff face, and because of the high degree of lateral variability, correlation is at best difficult. Error in thickness estimation is probable, but mainly a risk in terms of underestimation. Hence, measurements at Cape Enragé document a minimum thickness.

The Claremont Formation

Thanks to a shoreline clean-up by Hurricane Noel in 2007, the unconformable contact between the uppermost grey sandstone bed of the Shepody Formation and coarse red beds of the overlying Claremont Formation is now well exposed at both Downing Cove and Downing Head East and West (Figs. 4, 6), approximately 10 m above what Ryan and Boehner (1994) considered to be the contact between these two formations. The basal bed at each section is an intraclastic conglomerate composed of red or grey mudstone clasts and red or grey fine sandstone clasts. The clasts of these basal conglomeratic beds were scavenged from the underlying Shepody Formation by downcutting (Fig. 5B).

In the study area, the Claremont Formation consists mainly of red, cross-cutting channels of coarse to granular sandstone along with granular and pebble conglomerate, fine to medium sandstone, and mudstone. At all localities, the first few occurrences of conglomerate are intraclastic, whereas the other occurrences are quartzose granular conglomerate and polymictic pebble conglomerate that is also rich in quartz.

The Claremont Formation fines away from Downing Head West, both eastwardly and westwardly (Figs. 3, 4). At Downing Head West, despite a composition that includes less than 5% carbonate (McLeod 2010), large karstic shafts and galleries (up to several metres wide) occur throughout the upper half of the Claremont Formation in coarse to pebbly sandstone and are infilled by sandy mudstone (Fig. 5C). At Downing Head East, the same interval is characterized by deep weathering, but no large karst.

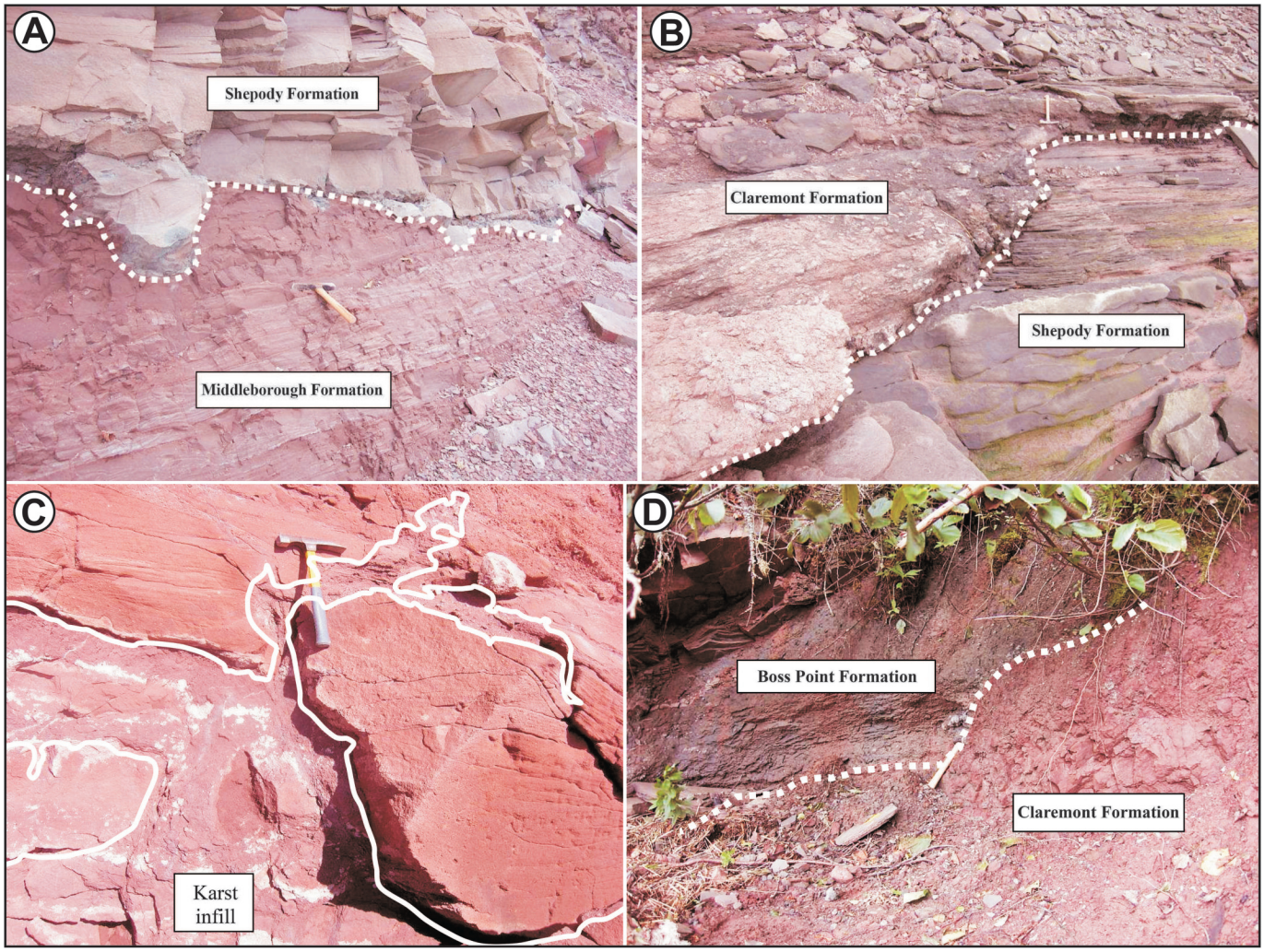
Provenance

Paleocurrent data were obtained from scour-and-fill trough orientations in sandstone intervals of the Shepody (Fig. 6A) and Claremont (Fig. 6B) formations at all sections where these units occur. The Shepody and Claremont formations show consistent north-northwest-verging paleocurrent vectors in the Maringouin, Downing Head, and Downing Cove sections, and the Shepody Formation shows southeast-verging paleocurrent vectors at Cape Enragé, which are altogether similar to those found in the underlying Middleborough Formation in those sections, and which suggest that the inferred Cobequid Highland and Caledonia Highland sources for the Visean unit (Fig. 1; Jutras et al. 2015) were still sourcing that sector of the Cumberland Basin during the Serpukhovian.

Note that our results from Downing Cove show a much more constrained range than that obtained by Hamblin (2001) in the same section. Our data also differ from those of Allen et al. (2013), who imply a dominance of southeast-verging vectors for the Shepody and Claremont formations in the Joggins area in their fig. 5, but mostly north–south vectors in their fig. 13, with no details on the nature of the paleocurrent data. We consider these southeast-verging vectors to be incompatible with data from seismic line 539 of Devon Canada (2002; in Waldron and Rygel 2005), which stretches north–south from the town of Joggins towards the Cobequid Highlands (Fig. 1), and in which an increase in the coherence of reflectors towards the north suggests that the source is to the south. This relationship is especially clear in the Langsettian interval, where the coarse clastic rocks of the Polly Brook Formation transition to the north into the finer clastic rocks of the Joggins Formation (Waldron and Rygel 2005), in which Allen et al. (2013) also note a dominance of highly improbable southeast-verging vectors.

Since our paleocurrent data differ so much from those of previous authors (Hamblin 2001; Allen et al. 2013), and since data from the latter authors also differ substantially from each other, an assessment of data reliability is here called for. In our experience of collecting broadly distributed paleocurrent measurements from braided fluvial systems (Jutras et al. 1999, 2001, 2005, 2007, 2015; Jutras and Prichonnet 2002, 2004, 2005), we now consider data collected from channel fills (clast imbrications, ripple marks, planar-cross bedding, and parting lineations) as highly unreliable, because deposition of the latter can be from weak secondary currents that were strongly influenced by the geometry of the channels, whereas scours are the products of stronger currents

Fig. 5. (A) Disconformable contact between the Middleborough and Shepody formations at Cape Enragé, New Brunswick. (B) Disconformable contact between a coarse grey sandstone bed at the top of the Shepody Formation and an intraclastic conglomerate bed at the base of the Claremont Formation at Downing Cove, Nova Scotia. (C) Karst structures within coarse sandstone in the upper part of the Claremont Formation at Downing Head West, Nova Scotia. (D) Disconformable contact between the Claremont and Boss Point formations at Downing Head West, Nova Scotia. Hammer for scale.



that reflect better the direction of the source area. Two-thirds of the paleocurrent measurements in Hamblin (2001) are from ripple mark orientations, which we consider as unreliable. The nature of the measurements made by Allen et al. (2013) is unknown, but as they have included 364 measurements from sections in which we could only find 51 reliable paleocurrent features, we conclude that their rose diagrams are dominated by data that we here consider as unreliable to determine the position of the source area in relation to the studied unit.

A Cobequid source is consistent with clast contents in the Shepody and Claremont formations, which include abundant rhyolite, phyllite, schist, and polycrystalline quartz (McLeod 2010). These lithologies are commonly found in the Cobequid Highlands shear zone (Pe-Piper and Piper 2002), although it should be noted that post-Mississippian dextral displacement along this fault is considered to be quite substantial and that the source area of the herein studied sections is most likely now located below the Bay of Fundy.

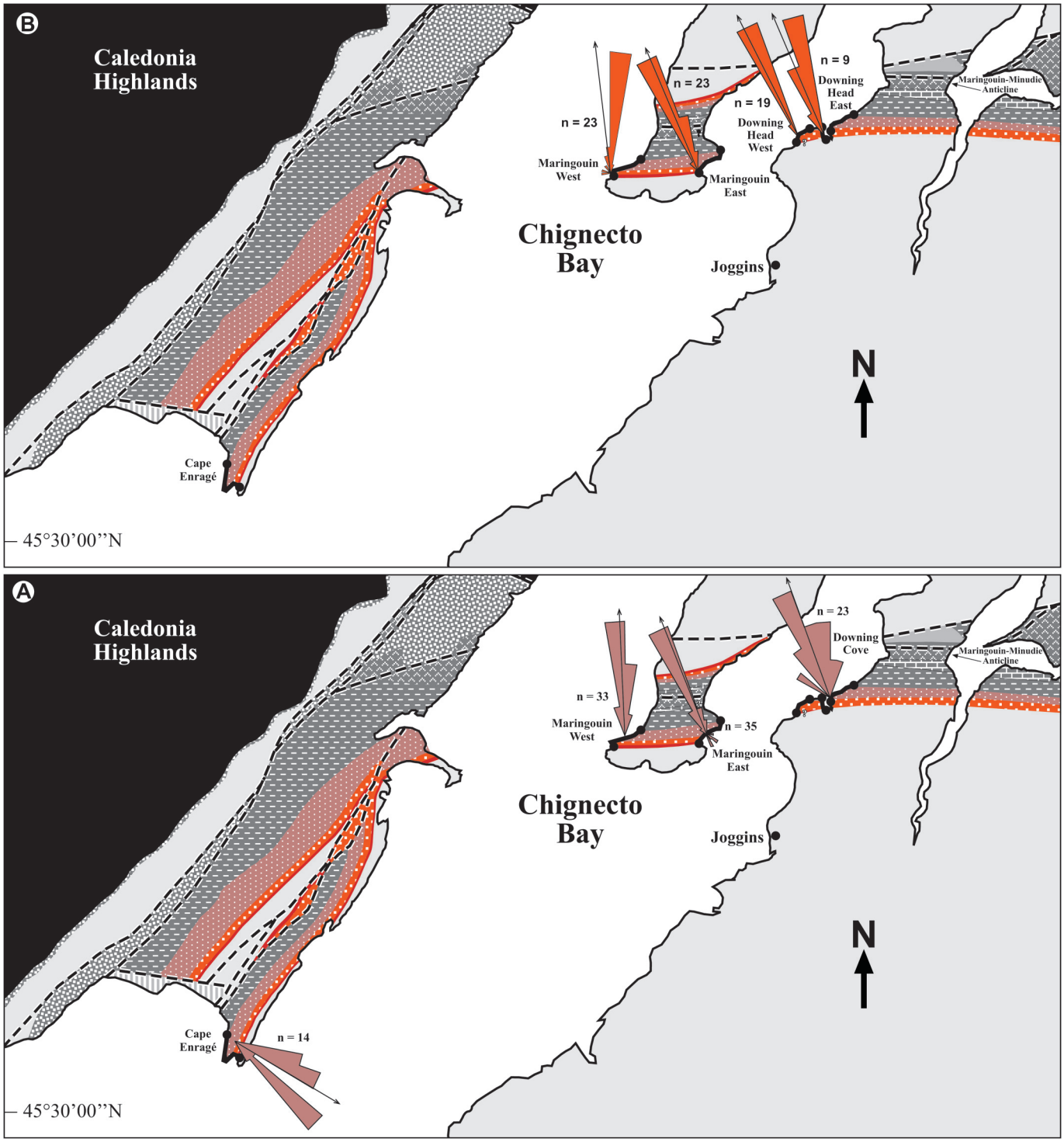
The Mississippian–Pennsylvanian transition

The uppermost Claremont Formation is exposed only at Downing Head East, where it terminates with a ~5 m thick, red, muddy paleosol (Fig. 4). The latter is disconformably downcut by an intraclastic grey conglomerate that evolves upward into grey sandstone and that corresponds to the base of the Ward Point Member of the Pennsylvanian Boss Point Formation (Fig. 5D). This is in contrast with the Maringouin West section, ~23 km to the west-southwest, where up to 65 m of fine red beds of the Boss Point Formation Chignecto Bay Member separate the Ward Point Member from the Claremont Formation (Figs. 1, 4).

At Cape Enragé, a ~345 m gap separates the uppermost exposure of the Shepody Formation from grey sandstone of the Ward Point Member (Fig. 3). Based on mapping extrapolations from available exposures inland from Cape Enragé (St. Peter and Johnson 2009), this gap may be occupied by the Claremont Formation and the Chignecto Bay Member of the Boss Point Formation (Figs. 1, 3).

Based on our data and those of St. Peter and Johnson (2009), an increasing amount of Mississippian strata is missing below the Cumberland Group towards the north (Fig. 1). Most of the Mabou Group is present below the Cumberland Group at Cape Enragé, Maringouin, Downing Head, and Downing Cove, to the south, but very little remains north of the Shepody–Beckwith Fault (Fig. 1).

Fig. 6. Paleocurrent vectors based on the measurement of scour-and-fill trough structures in (A) the upper Brigantian to Pendleian Shepody Formation and (B) the Arsnbergian Claremont Formation. *n*, number of measurements.



Pre-Cumberland erosion reaches deep down into the Visean Hopewell Cape Formation at Hopewell Cape, down into the base of the Upper Devonian to Tournaisian Horton Group at Gaytons, just south of the Gaytons Fault, and down into Westmorland basement rocks at Calhoun, just north of the latter fault (Fig. 1). As the Chignecto Bay Member is mainly present where the Mississippian succession is most complete (St. Peter and Johnson 2009), it is possibly a product of erosion of the latter.

No paleocurrent vectors could be retrieved from the fine red beds of the Chignecto Bay Member of the Boss Point Formation. In the overlying Ward Point Member at Maringouin West, paleocurrent vectors are still north-verging, suggesting a Cobequid Highland source for these beds, which is supported by the aforementioned trends in seismic line 539 (Waldron and Rygel 2005), but which contradict the dominantly east-verging paleocurrent vectors reported by Allen et al. (2013).

Air-photo analysis

Air photos obtained from the Map Library of Service Nova Scotia were used to verify the lateral extent of thick, coarse sandstone units in the Shepody Formation at the Downing Cove and Downing Head sections, and to see how the Shepody/Claremont and Claremont/Boss Point contacts evolve laterally. The uppermost three grey sandstone intervals can be traced from Downing Cove to Downing Head West, which are labelled from the top down as G1–G3 (Fig. 4), with G1 being subdivided as G1a and G1b on each side of a thin muddy interval (Fig. 7).

The sandstone units denoted as G3 and G2 display a relatively consistent thickness, with an equally thick succession of red mudrock separating the two units. These units appear to be laterally continuous over several kilometres, spanning the Downing Cove and Downing Head sections, and extending possibly as far as the Maringouin sections based on tentative correlations (Fig. 3).

On both air photos (Fig. 7B) and stratigraphic sections (Fig. 4), an important thinning of the G1 unit is observable between Downing Cove and Downing Head West due to the truncation of most of the G1a unit by the disconformably overlying Claremont Formation at the latter section. Globally positioning system coordinate data obtained in the field were used to plot the contact on both sides of Downing Cove in Google Earth. The coordinates matched the expected position of the contact based solely on the image. Hence, although only 10 m of G1a is missing at Downing Cove West, compared with Downing Cove East, this 10 m of downcutting is well observable on air photos due to the relatively shallow dip of the succession, which ranges between 35° and 40°, and which therefore exaggerates the disconformity. In contrast, the disconformity between the Claremont and Boss Point formations is much less significant at that scale based on observable lineaments on the air photo (Fig. 7), although the contact is not exposed on the west side of the headland, which makes the amount of downcutting difficult to estimate.

Discussion

Sedimentary environments of the Shepody and Hastings formations

The lateral persistency of the two alternating facies of the Shepody Formation suggests that they were controlled by climate cyclicity. The successions of coarse, grey, cross-channelized sandstone with plant remains, which are laterally uninterrupted on exposures that span up to 3 km in some cases, are interpreted as sandy braidplain deposits. We concur with Allen et al. (2013) that they must have been deposited by strongly seasonal fluvial systems. However, the preservation of coal suggests a non-arid setting, and the common occurrence of the trilete spores *Schopfites claviger* and *Colatisporites decorus* in those beds more specifically suggests a sub-humid setting (van der Zwan 1981). In contrast, the intervals of planar-bedded red mudstone and fine sandstone with pedogenic overprints are interpreted as playa deposits from more arid periods. The lack of evidence for overbank deposition in the fine red bed facies, such as splay deposits, also supports the conclusion that it is not the floodplain lateral equivalent of the coarse channel deposits.

Evidence of frequent channel avulsion, deep downcutting, and pervasive pedogenic overprinting in the upper Brigantian to Pendleian Shepody Formation suggest that it has evolved in a more purely continental environment than the planar-bedded, upper Asbian to lower Brigantian Middleborough Formation, which shows evidence for subtidal to supratidal environments (Jutras et al. 2015). This conclusion is also supported by the regional stratigraphy of Nova Scotia, which records in several areas a transition from marine to continental environments near the Visean–Serpukhovian boundary.

Paleoenvironmental implications

Giles (2009) demonstrated that Atlantic Canada was affected by significant glacioeustatic variations in late Visean times, starting in the upper Asbian. In the Cumberland Basin, this base-level cyclicity was partly masked by excessive subsidence in salt expulsion minibasins (Jutras et al. 2015). Glacioeustasy is probably responsible for alternations between carbonates and subtidal siliciclastic deposits in the upper Asbian Lime-kiln Brook Formation of the Cumberland Basin, but no clear cyclicity has been recognized in laterally equivalent sulfate and in overlying Brigantian red beds, which seemingly evolved in a consistently hyperarid setting (Jutras et al. 2015). Climate cyclicity is more apparent in the Shepody Formation, with grey sandstone intervals being interpreted as the record of warm cycles, which would have resulted in an increase of ocean water evaporation and atmospheric moisture, whereas the red bed intervals are interpreted to be the record of colder and globally more arid cycles. Hence, the Visean–Serpukhovian transition is not only marked by a final retreat of the Windsor Sea from Nova Scotia, but also by a significant humidification of the climate, which also became more sensitive to orbital cyclicity.

Based on the paleomagnetic studies of Scotese and McKerrow (1990) and Ziegler et al. (2002), Atlantic Canada was located in the paleosubtropics and migrating towards the paleoequator in late Mississippian times, which may partly explain the general humidification of the climate at the Visean–Serpukhovian transition (Allen et al. 2011). The Visean–Serpukhovian transition also corresponds to a time when ice volume was increasing on Gondwana (Rygel et al. 2008), which might explain why glacioeustatic Windsor Sea incursions in eastern Canada stopped occurring in early Serpukhovian times (Giles 2009; Opdyke et al. 2014). According to Rygel et al. (2008), an increase in the amplitude of glacioeustatic variations and therefore global climate fluctuations also occurred at the Visean–Serpukhovian transition, which may in part explain why such climate fluctuations are better recorded in continental clastic successions of the early Serpukhovian than in those of the Visean in Atlantic Canada.

The Shepody/Claremont disconformity

Postulating that the base of the Claremont Formation is younger than the uppermost beds of the Hastings Formation in Northern Nova Scotia, which bear an late Pendleian to Arsnbergian *Reticulatisporites carnosus* spore assemblage (Utting 2011), a substantial thickness of originally deposited Pendleian beds may be missing below the Shepody/Claremont unconformity, as a late Brigantian to Pendleian SM spore assemblage was retrieved less than 1 m below this unconformity at Maringouin East (Fig. 3). This erosion event may have been more substantial east of the study area, where the Shepody Formation is reportedly absent below the Claremont Formation (Ryan and Boehner 1994; Waldron et al. 2013).

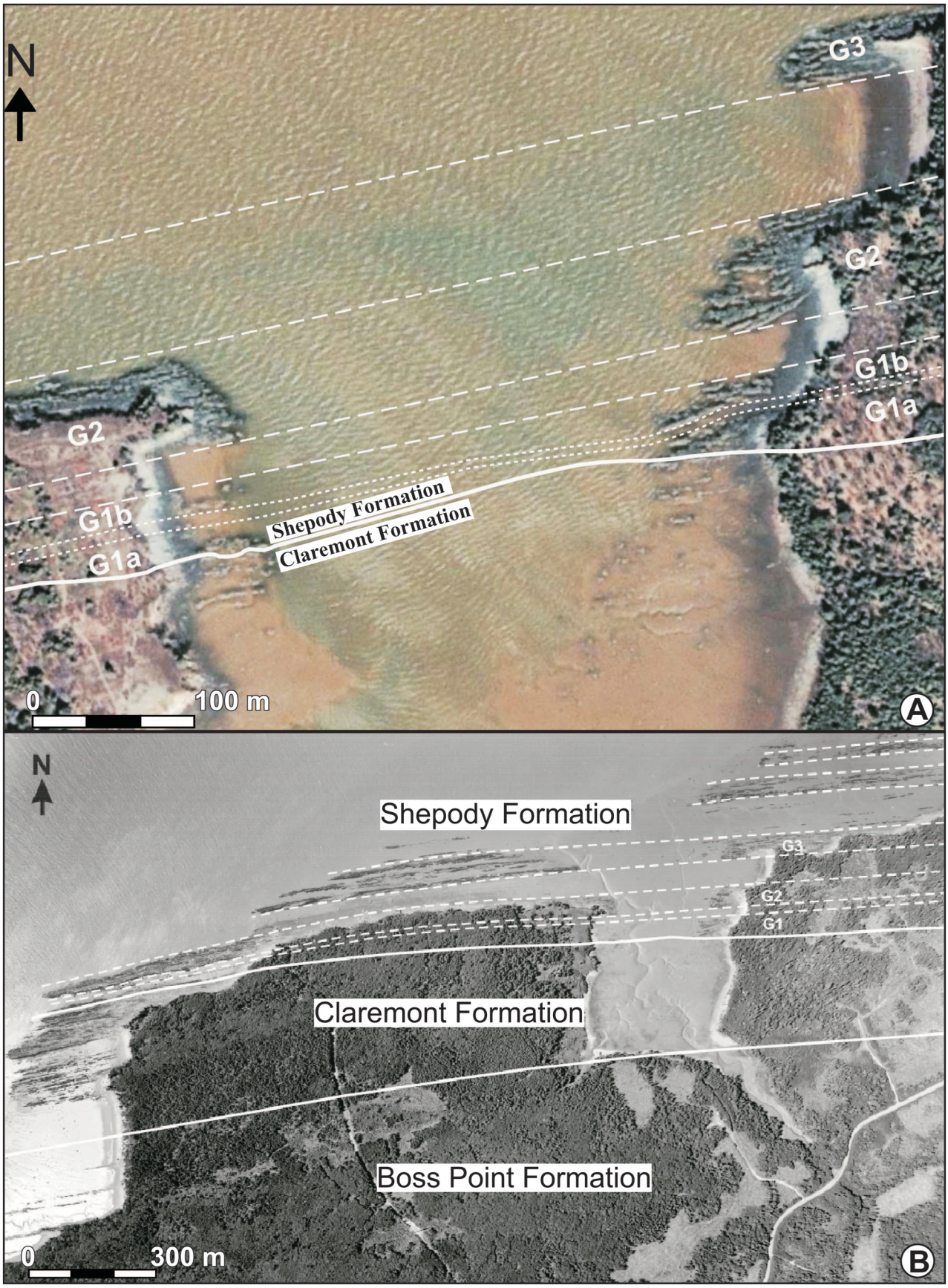
Paleoenvironmental implications

The erosion event that separates the Shepody and Claremont formations is possibly associated with a significant lowering of base level in relation to the ongoing size increase of Gondwanan glaciations in late Mississippian times (Isbell et al. 2003; Rygel et al. 2008). However, if erosion was indeed much more substantial east of the study area, a period of tectonic uplift is also possible. The glacioeustatic hypothesis is here favoured because of the significantly different climate signature that is found on each side of the disconformity.

Sedimentary environments of the basal Claremont Formation

Deposition of the coarse Claremont Formation is interpreted to mark an increase in fault activity and basin subsidence rates near the Pendleian–Arsnbergian boundary, which ended the preceding

Fig. 7. (A, B) Aerial views of the Downing Head and Downing Cove areas, showing the lateral continuity of thick grey sandstone sequences and intervening red beds in the Shepody Formation below the Shepody/Claremont unconformity.



period of basin erosion. The passage from coaly grey sandstones of the Shepody Formation to barren red conglomerates of the Claremont Formation is interpreted to be the record of a return to more arid conditions in the Arsnbergian, but not as arid as during the Visean, as reflected by mineralogically more mature conglomerates, intraclasts excluded, and as indicated by the presence of lycopsid spores in the laterally equivalent Pomquet Formation (Utting and Giles 2008). The vertically persistent succession of cross-cutting conglomerate channels suggests that the depositional environment of the Claremont Formation was that of a semi-arid gravelly braidplain, which is on par with the assessments of Hamblin (2001) and Allen et al. (2013).

Paleoenvironmental implications

Global climate degradation and a size increase of Gondwanan glaciations are inferred to explain the return to greater aridity near the Pendleian–Arsnbergian boundary. The Claremont Formation is interpreted to have been deposited farther from the sea than the underlying Shepody Formation. Moreover, no clear evidence of climate cyclicity could be found in this unit. This is also possibly explained by greater continentality, which may have made the regional climate less sensitive to global glacioeustatic fluctuations.

Sedimentary and eodiagenetic environments of the upper beds of the Claremont Formation

Very deep and pervasive weathering in the upper part of the Claremont Formation suggests that a substantial decrease in sedimentation rates occurred at that interval. Sedimentary quiescence seems to have been especially prolonged at the very top of the Claremont Formation, where deep weathering is observed just below the disconformity with the Pennsylvanian Boss Point Formation, which coincides with the Mississippian–Pennsylvanian transition. The development of karst below the Claremont/Boss Point unconformity at Downing Head West (Fig. 4) suggests that these units are separated by not only a significant period of non-deposition, weathering, and erosion, but also eodiagenetic cementation, as karst can only develop in consolidated rocks with generally low porosity (Ford and Williams 2007).

Paleoenvironmental implications

The long period of non-deposition, weathering, cementation, karstification, and erosion that followed deposition of the Claremont Formation and preceded deposition of the Boss Point Formation corresponds to a large hiatus that is found throughout Atlantic Canada between Arsnbergian (late Serpukhovian) and Yeadonian to Langsettian (mid-Bashkirian) beds, with the latest Mississippian and earliest Pennsylvanian intervals missing entirely. This hiatus has been associated with a major paleoenvironmental crisis that is also well recorded in Europe and that was followed by a substantial palynofloral change in the early Pennsylvanian (Utting 1987; Utting and Giles 2008). As suggested by well-recorded glacial deposits during that time period (Isbell et al. 2003), an increase in the ongoing deterioration of global climate and the associated aggravation of Gondwanan glaciations near the Mississippian–Pennsylvanian boundary may have promoted hyper-aridity and low base level in Atlantic Canada at the time, thus explain both the floral crisis and the hiatus.

Because of the siliciclastic composition of the Claremont Formation, very unusual groundwater conditions must have developed to account for karst development in these rocks. To explain a similar situation in the Hopewell Cape Formation at Hopewell Cape (Fig. 1), Jutras[2] proposed that high pH groundwater conditions may have developed in the arid Visean context, which are here also inferred for the late Serpukhovian to early Bashkirian occurrences. Above a groundwater pH of 9, siliciclastic material can go through nearly congruent solution due to a significant increase of both silica and aluminium solubility (Blatt et al. 1980).

As the karst occurrences in the Claremont Formation did not form as close to the source area as those in the coarse Hopewell Cape Formation, a less complex scenario is necessary to explain their formation. A semi-arid climate would best provide a sluggish but continuous groundwater flow that could become increasingly alkaline in central parts of the basin because of high evaporation rates, favouring early cementation by calcite, and eventually favouring the dissolution of silicate minerals in an alkaline groundwater flow concentrated between aquicludes. These well-developed karstic features are the record of a long period of non-deposition during which this part of the basin was lying above base level and was possibly dissected by fluvial cuts. In this context, karst would have best developed in areas of increased groundwater circulation towards fluvial cuts.

Paleogeographic model for the Serpukhovian interval in the Cumberland Basin

Late Visean prequel

The well-recorded late Visean setting of the Cumberland Basin (Fig. 8) provides a foundation for our reconstruction of the less well-recorded Serpukhovian setting, which was largely eroded prior to Pennsylvanian deposition. To explain lateral variations of facies and the lateral distribution of paleocurrent vectors, Jutras et al. (2015) proposed that at least two syndepositional salt diapirs (including the well-documented Dorchester Salt Dome of Martel 1987) and three salt expulsion minibasins were active during late Visean deposition and were controlled by differential clastic loading issued from the Cobequid Highlands to the south, the Caledonia Highlands to the west, and the Westmorland Highlands to the north (Fig. 8). In this setting, marine and evaporitic Visean deposits were concentrated in the three inferred areas of salt expulsion, northeast of Hopewell Cape, east of Downing Cove, and south of Cape Enragé (Fig. 8).

Grey intervals of the Shepody Formation

At all localities in which the Shepody Formation is exposed, it displays similar paleocurrent trends to those of the underlying Visean units (Figs. 6A, 8). This suggests similar tectonic and halokinetic environments in which sedimentation was still controlled by at least two of the fault-bounded source areas, namely the Cobequid and Caledonia Highlands (Fig. 9). The lack of convergence between river flows issued from the Caledonia Highlands with those issued from the Cobequid Highlands suggests that the latter were still funnelled in separate minibasins, as was inferred for the Visean successions (Jutras et al. 2015). However, the few remnants of time-equivalent grey units in the northwestern part of the Cumberland Basin, at Dorchester Cape and in Well Mud Creek 52 (Fig. 1), are fine grained and partly onlapping the Westmorland basement, which does not suggest that the latter was still exposed at the time. Moreover, there is sedimentological evidence in the time-equivalent Hastings Formation of northern Nova Scotia for wide and open lacustrine conditions towards the north of our study area (Giles and Utting 1999; Hamblin 2001), which suggests that the north-verging rivers that deposited the grey beds of the Shepody Formation at Maringouin, Downing Cove, and Downing Head were not far from their connection to an open lake. It is here interpreted that this fluviatile–lacustrine connection occurred in the area that separates these sections from the Dorchester Cape and Mud Creek 52 sections (Fig. 9), where

[2]P. Jutras. The role of salt tectonics, glacioeustatic variations and high pH evaporitic groundwater in the development of paleokarst within polymictic fanglomerate of the Carboniferous Hopewell Cape Formation, Atlantic Canada. [Manuscript in preparation.]

Fig. 8. Paleogeographic model for the western Cumberland Basin in latest Asbian times based on mapping data (Ryan and Boehner 1994; St. Peter and Johnson 2009), geophysical and sedimentological data (Jutras et al. 2015), and principal shortening direction based on Faure et al. (1996), Jutras and Prichonnet (2005), and Wilson and White (2006). Modified from Jutras et al. (2015).

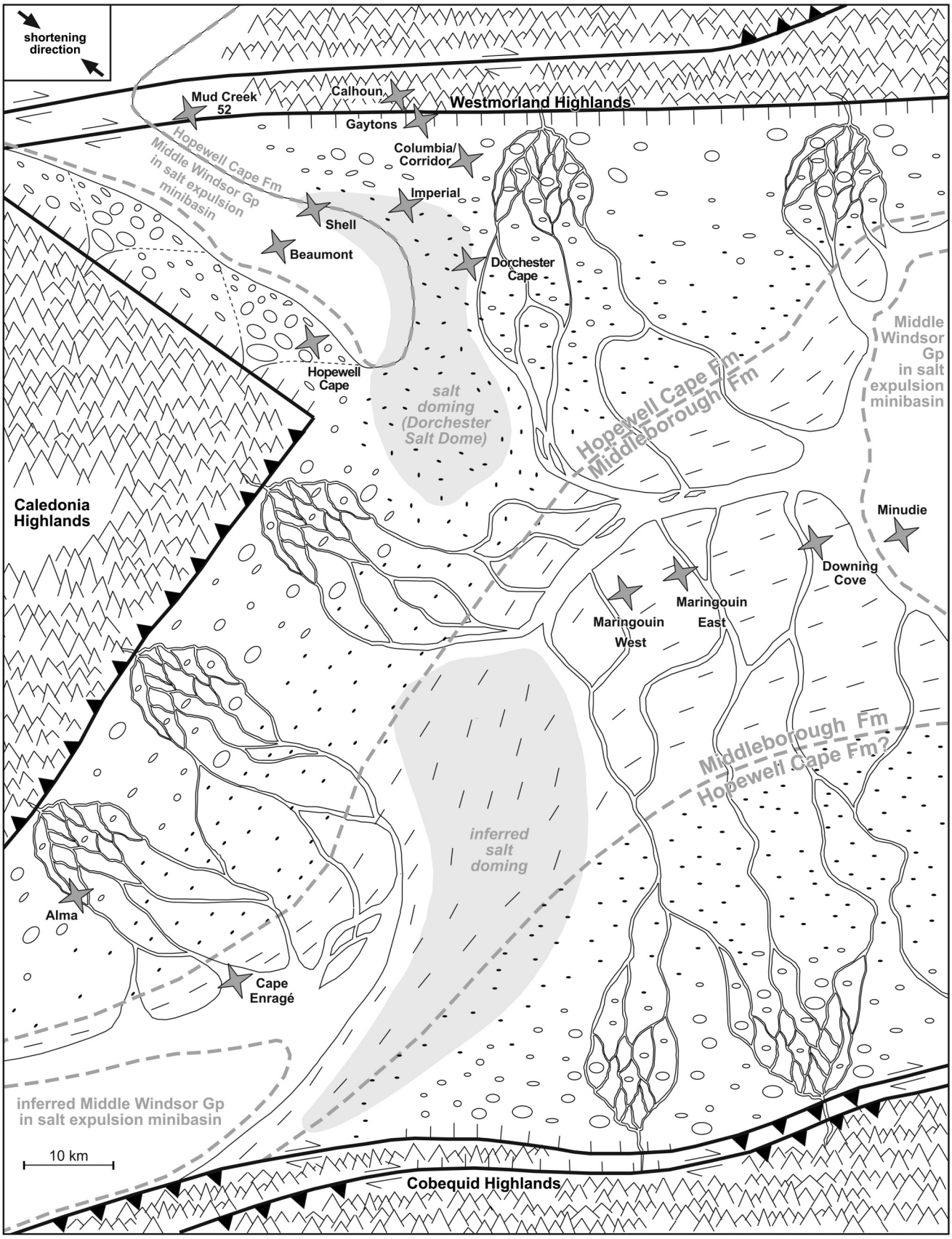


Fig. 9. Paleogeographic model for the grey intervals of the upper Brigantian to Pendleian Shepody Formation.

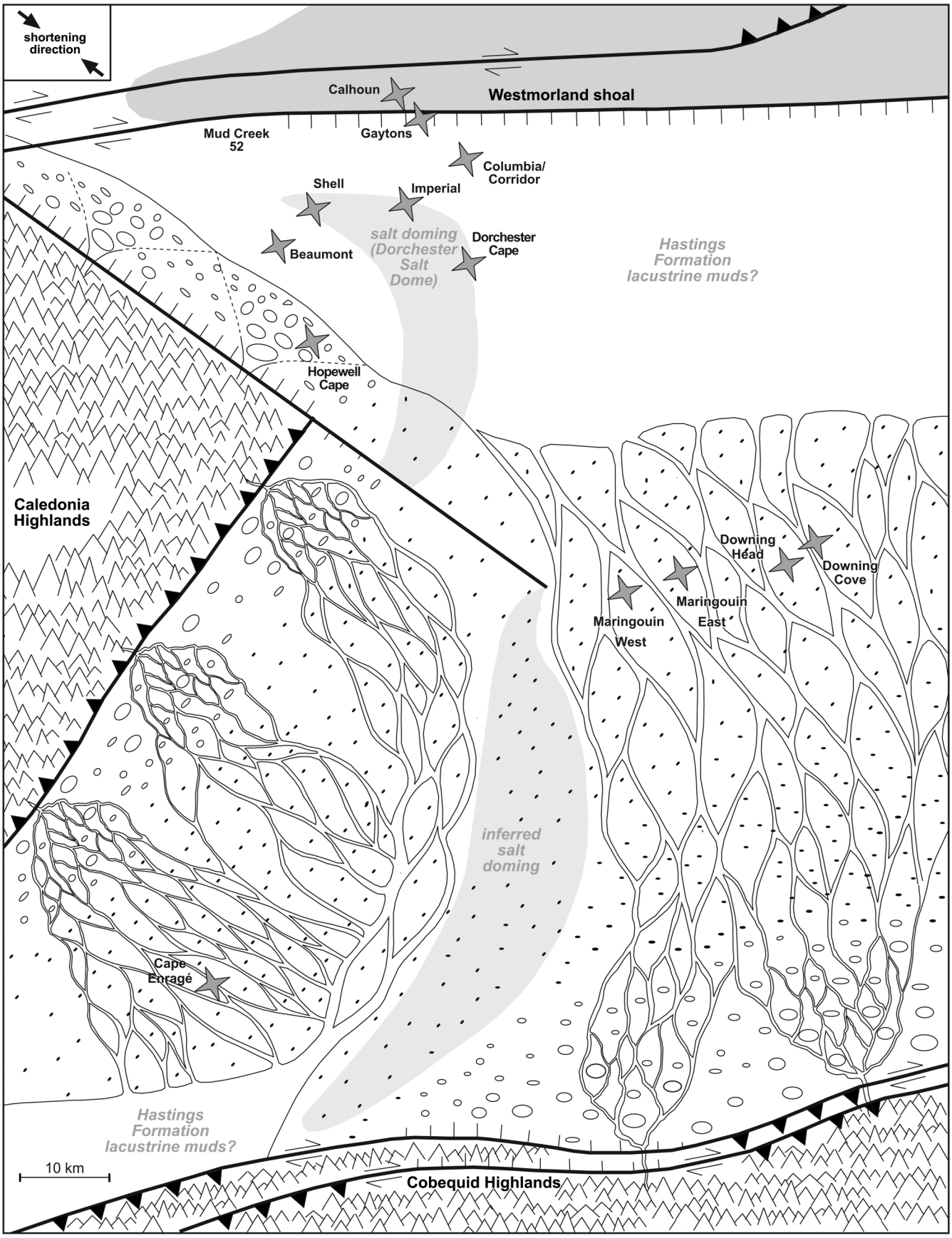


Fig. 10. Paleogeographic model for the red intervals of the upper Brigantian to Pendleian Shepody Formation.

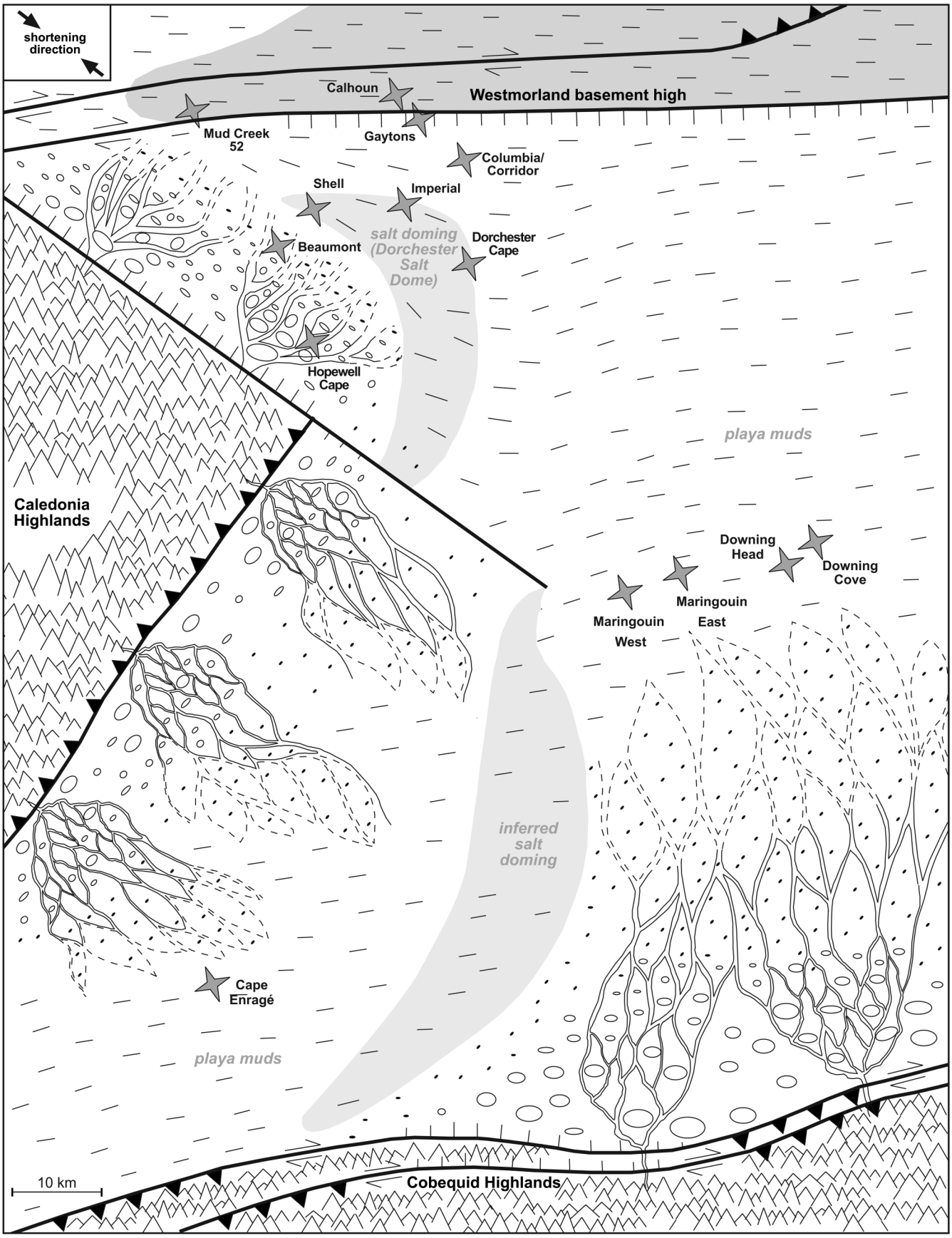
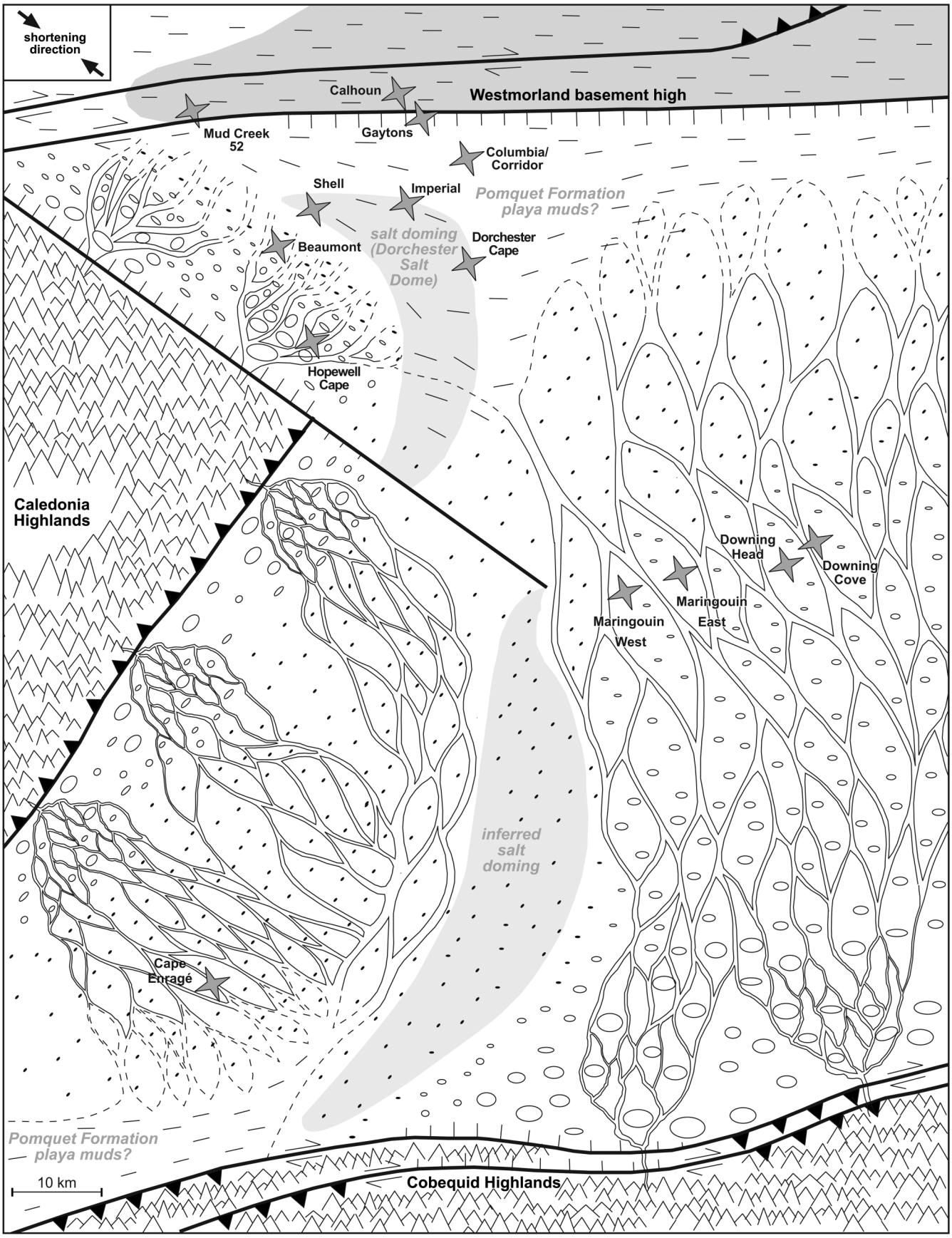


Fig. 11. Paleogeographic model for the red beds of the Arsnbergian Claremont Formation.



remnants of grey mudrock that correlate well with the Hastings Formation are found. In this context, the submerged Westmorland basement block may have formed a lacustrine shoal (Fig. 9).

Red intervals of the Shepody Formation

The lack of large fluvial channel fills in the red intervals of the Shepody Formation suggests that arid alluvial fans that were sourced from the Cobequid Highlands were passing laterally into playa deposits over a short distance (Fig. 10). This is interpreted to be the result of shorter-lived and more dispersed flow discharge, which is typical of arid continental environments.

The Claremont Formation interval

The similarity in paleocurrent directions between the Shepody and Claremont formations at Maringouin, Downing Head, and Downing Cove (Fig. 6) suggests that at least the Cobequid Highlands were still active as a source area when the latter unit was deposited (Fig. 11). Based on the conglomeratic nature of the few exposures of the Claremont Formation that are near the Caledonia Highlands (St. Peter and Johnson 2009), it can be inferred that the latter was also still active as a source area. However, the marked increase in coarseness above the Shepody–Claremont contact, especially if only red intervals are considered, suggests that these source areas were more quickly rejuvenated by more active faults during deposition of the latter unit. The gravelly braidplain deposits of the Claremont Formation are interpreted to have passed laterally into sandy braidplain and eventually playa deposits of the Pomquet Formation (Fig. 11), although these transitions are now fragmented and discontinuous due to early Pennsylvanian deformation and erosion.

Summary and conclusions

Serpukhovian events in the western Cumberland Basin of Atlantic Canada can be summarized in the following way:

- Significant humidification of the climate near the Visean–Serpukhovian boundary as the Maritimes Basin migrated closer to the paleoequator (Scotese and McKerrow 1990; Ziegler et al. 2002; Allen et al. 2011), which resulted in the preservation of organic matter in much of the fluvial sediments of the Shepody Formation.
- Increase in the magnitude of orbitally controlled glacioeustatic and climatic fluctuations at the Visean–Serpukhovian transition because of an increase in ice volume on Gondwana at the time (Rygel et al. 2008). In this new context, both highstands and lowstands would have been lower than equivalent cycles during the Visean, which explains why marine incursions no longer occurred during the Serpukhovian in Atlantic Canada. The organic-rich, non-oxidized channel-sand bodies of the Shepody Formation are here interpreted as having been deposited in sub-humid conditions during highstands, whereas red playa intervals of the same unit are interpreted as having been deposited in semi-arid conditions during lowstands.
- Although better documented in the more widely preserved upper Visean units of the study area (Jutras et al. 2015), the maintenance of peculiar paleocurrent patterns into the Serpukhovian, with a lack of convergence between paleoflows coming from different source areas, suggests that halokinesis in the Bay of Fundy was still ongoing in Serpukhovian times. However, as Serpukhovian paleocurrents are north-verging on the southern flank of the Maringouin–Minudie Anticline at Joggins and Maringouin (Figs. 1, 6), it is clear that deposition of Mabou Group units was not influenced by this south-southwest-striking salt wall, which therefore must have developed subsequently.
- The Cobequid, Caledonia, and Westmorland highlands were identified as three distinct source areas for upper Visean units (Jutras et al. 2015). Although facies distribution and paleocurrent data suggest that the Cobequid and Caledonia highlands were still active source areas in Serpukhovian times, the presence of fine, early Serpukhovian grey beds onlapping the Westmorland basement at Well Mud Creek 52 (Fig. 1) suggests that the latter source area was no longer active, but was instead forming a shoal in the broad lake that, according to Hamblin (2001), accommodated much of the upper Brigantian to Pendleian Hastings Formation of northern Nova Scotia (Fig. 9).
- A period of weathering and erosion occurred near the Pendleian–Arsnbergian boundary in the study area, possibly marking a significant step in the ongoing late Mississippian drop in global sea level in association with the aggravation of glacial conditions in low latitudes.
- Fault-controlled Serpukhovian sedimentation eventually resumed in Arsnbergian times in relation with an increase in basin subsidence, but under a different climate regime. The Arsnbergian Claremont Formation is substantially coarser than the underlying Shepody Formation, and it no longer includes non-oxidized deposits with plant remains nor convincing evidence of climate cyclicity. Greater distance from a globally receding sea may explain the reduced sensitivity of the sedimentary system to orbitally controlled climatic and glacioeustatic cyclicity.
- A long period of weathering and karstification took place between the Arsnbergian and the Yeadonian, which is a time that corresponds to an important floral crisis in eastern Canada and Europe (Utting 1987; Utting and Giles 2008). Again, an aggravation of Gondwanan glaciations is inferred, possibly paired with a slowdown of fault activity and basin subsidence rates.

Regarding the long-term evolution of the halokinetic setting in the western part of the Cumberland Basin, our preliminary data suggest that paleocurrents are still north-verging on the southern flank of the Minudie–Maringouin Anticline salt wall in the lower Pennsylvanian Boss Point Formation (Fig. 3, Maringouin West section). Hence, a sedimentological study of the rest of the Pennsylvanian succession would be needed to determine when the south-southwest-striking salt walls that are inferred to have partly controlled Pennsylvanian deposition in the Cumberland Basin (Waldron et al. 2013) first became active.

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