West African proximity of the Avalon terrane in the latest Precambrian

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ABSTRACT

Considerable debate surrounds the Late Neoproterozoic paleogeographic position of the Avalon terrane, specifically whether it was adjacent to West Africa or Amazonia. New paleomagnetic results from upper Neoproterozoic rocks in the Avalon terrane challenge the latter position. Samples collected from the ca. 580–570 Ma Marystown Group in the southern part of the Burin peninsula of Newfoundland, Canada, yield high-temperature magnetic components, including dual-polarity directions, which are considered to be primary, on the basis of positive fold and agglomerate tests. The resultant tilt-corrected inclination is 53°, representing a paleolatitude of deposition of 34° +8°/–7° for the Marystown Group. Given the likelihood that Amazonia and Laurentia were still juxtaposed around 580–570 Ma, the Marystown Group results reveal that the paleolatitude of Avalon is significantly lower than would be expected if it was part of Amazonia. In fact, Avalon was separated from northern Amazonia by at least 1100 km at ca. 580 Ma. If West Africa was juxtaposed to Amazonia, opposite Laurentia, by this time, these results place Avalon at the same paleolatitude as the northern margin of the West African craton.

Keywords: Appalachians, paleogeography, paleomagnetism, Precambrian, terranes.

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INTRODUCTION

Elements of the early Paleozoic Avalonian continent are now exposed along the length of the Appalachian orogen (Fig. 1). The paleogeography and tectonic history of the continent are relatively well established for Ordovician and Silurian time (e.g., Torsvik et al., 1996; Cocks et al., 1997); however, its Neoproterozoic and earliest Paleozoic paleogeographic characteristics are less well determined. Avalon appears to have been consolidated in the latest Precambrian, and the consensus is that it originated as a peri-Gondwanan terrane (e.g., Schenk, 1971; Murphy and Nance, 1989; O'Brien et al., 1996). It subsequently rifted from the Gondwanan margin in the early Paleozoic and drifted northward across the Iapetus Ocean in Ordovician time. The original location of Avalon on the Gondwanan margin remains uncertain. Whereas it was earlier advocated that Avalon originated from a position adjacent to West Africa (e.g., Schenk, 1971; Rast and Skehan, 1983; Fig. 2A here), more recent paleogeographic reconstructions have placed elements of western Avalonia adjacent to northern South America (e.g., Dalziel et al., 1994; Nance and Murphy, 1994; Keppie et al., 1996; Fig. 2B). The earlier models were primarily based on lithotectonic correlation with the northern edge of the West African craton. The more recent reconstructions have been based on basement neodymium isotopic signatures and the ages and possible sources of detrital zircons in Neoproterozoic metasedimentary rocks of western Avalonia (e.g., Nance and Murphy, 1994, 1996).

This controversy has important implications for the drift history of Avalon in the early Paleozoic; Avalon being off Africa requires a different journey for it to arrive at its Silurian collision with Laurentia and Baltica than does a position off South America. Depending on the longitudinally relative positions of Laurentia and Gondwana, this journey may have been longer or shorter than the alternative. The rift-to-drift transition of Avalon from Gondwana is generally thought to have occurred in the Ordovician; considerable debate concerns the exact timing within this period (e.g., Prigmore et al., 1997). Some authors have alternatively proposed an even earlier, latest Precambrian separation (Landing, 1996). The existing paleomagnetic data for Avalon do not provide constraints on the original position of Avalon relative to the Gondwana elements because most of the available Ordovician data pertain to Avalon as an already rifted entity. Moreover, the Ordovician reference paleopoles for Gondwana are not precise enough ($\alpha_{95} > 30°$; Van der Voo, 1993) to enable a rigorous test of Avalon's position relative to Gondwana at this time.

To discriminate between the alternative paleogeographic positions of Avalon relative to the Gondwanan margin, we carried out a detailed paleomagnetic investigation of late Neoproterozoic Avalonian rocks: the ca. 580–570 Ma Marystown Group of southeastern Newfoundland, which pertain to prerift Avalonia. A reconnaissance paleomagnetic study on the Marystown Group by Irving and Strong (1985) yielded evidence for stable magnetizations at three localities; our results

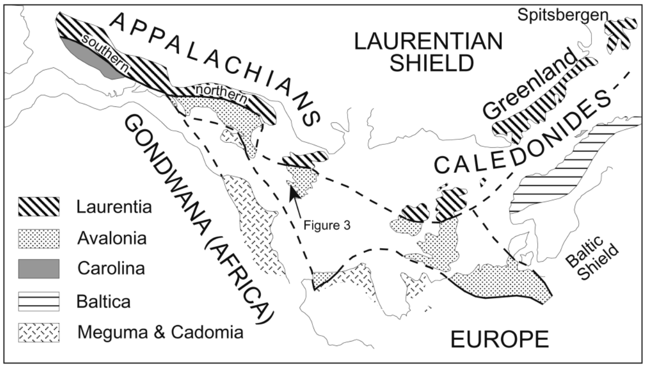


Figure 1. Illustration of the Appalachian-Caledonian belt, in a Pangea fit, depicting the locations of the various elements of Avalonia. The boundaries of Avalonia are taken from Cocks et al. (1997).

will be compared to theirs. Accurate positioning of Avalon along the Gondwanan margin at this time relies, of course, on knowledge of where the constituent elements of Gondwana were located. The precise timing of Gondwana assembly is poorly constrained because paleomagnetic poles for the Gondwanan continents are sparse for the time interval 815–550 Ma. Available poles indicate that 550 Ma is the latest date for the assembly of Gondwana (Meert and Van der Voo, 1997). Given that there is a lack of reliable poles for West Africa and Amazonia, poles taken from the remaining Gondwanan cratons may be used for a combined Gondwana path provided they are younger than 550 Ma. Laurentian poles may be used as a proxy for Amazonia for the time before the Laurentian-Amazonian break-up (Meert and Van der Voo, 1997), but this requires that the timing of the rift-to-drift transition and the predrift position of Amazonia relative to Laurentia are known. Igneous and stratigraphic evidence suggests that the rift-to-drift transition took place between 600 and 545 Ma (Williams and Hiscott, 1987; Torsvik et al., 1996). The uncertainty in these ages indicates that this approach must be used with caution.

GEOLOGY AND SAMPLING

The Marystown Group, as defined by O'Brien et al. (1977) and Strong et al. (1978), comprises the dominantly subaerial volcanic-sedimentary sequences that underlie much of the Burin peninsula of southeastern Newfoundland. It is dominated by felsic pyroclastic rocks with interbedded basaltic flows of tholeiitic composition, and clastic sediments, disconformably overlain by a sequence of Eocambrian to Cambrian sedimentary rocks that include the Precambrian-Cambrian stratotype at Fortune Head (Myrow and Hiscott, 1993). The volcanics of the Marystown Group are interpreted to have been extruded subaerially, as evidenced by the reddened nature of many of the flows, ubiquitous ignimbrites, and the facies of interbedded red sediments (e.g., Strong et al., 1977, 1978; O'Brien et al., 1977, 1983). Previous age determinations had bracketed the group as being 628–601 Ma (Krogh et al., 1988), but more recent work brackets the age as being c. 580–570 Ma (O'Brien et al., 1996; Sean O'Brien and Tom Krogh, 2000, written. commun.; see Fig. 3 here for locations of dated rocks and of sites). Furthermore, recent Pb-Pb ages from the lithologically and geochemically correlative St. Pierre Group of the St. Pierre and Miquelon islands, 20 km to the west of the Burin peninsula, have yielded similar age ranges (ca. 559–600 Ma) (Rabu et al., 1996; Fig. 3).

Standard 2.5 cm paleomagnetic core samples, or oriented hand samples, were collected (Fig. 3) from 21 sites in the Marystown Group as follows: samples of reddened porphyritic basalts were collected from 12 sites (sites 2–3, 6–7, 9–13, 16, 18, and 21), samples of red mudstones were collected at five sites (sites 4, 5, 8, 14, and 15), and samples of red siltstone were collected at two sites (sites 1 and 19). In addition, samples were collected from six clasts of porphyritic lava and siltstone in a conglomerate at site 17, and from seven clasts of porphyritic lava in an agglomerate horizon at site 20. At 17 sites, the bedding (Table 1) ranges from slightly to moderately dipping, al-

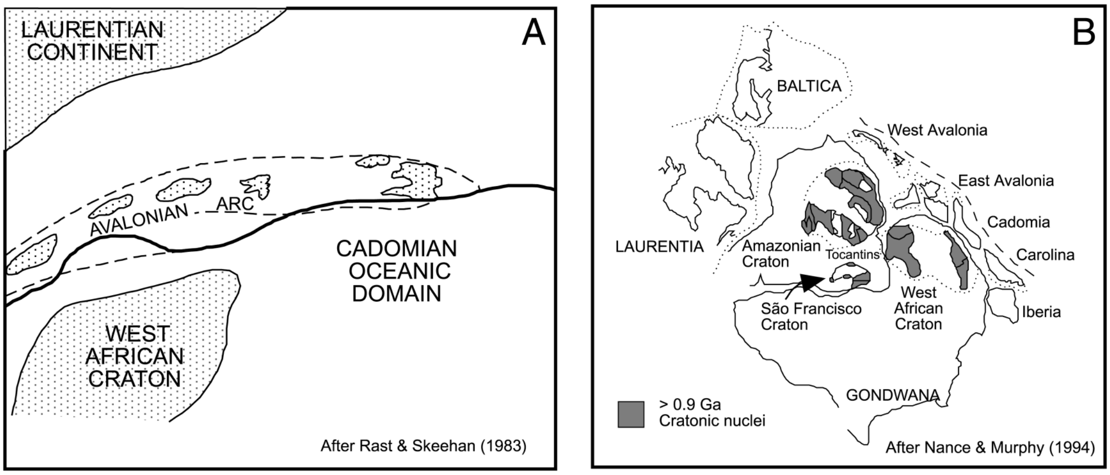


Figure 2. Two alternative paleogeographic positions for Avalonia in the late Precambrian. (A) Adjacent to the West African craton, on the basis of lithotectonic correlation (Rast and Skeehan, 1983). (B) To the north of the Amazonian craton, on the basis of matching the ages of detrital zircons and Nd isotopic signatures from Avalonia with a source in the Tocantins province of Brazil (Nance and Murphy, 1994).

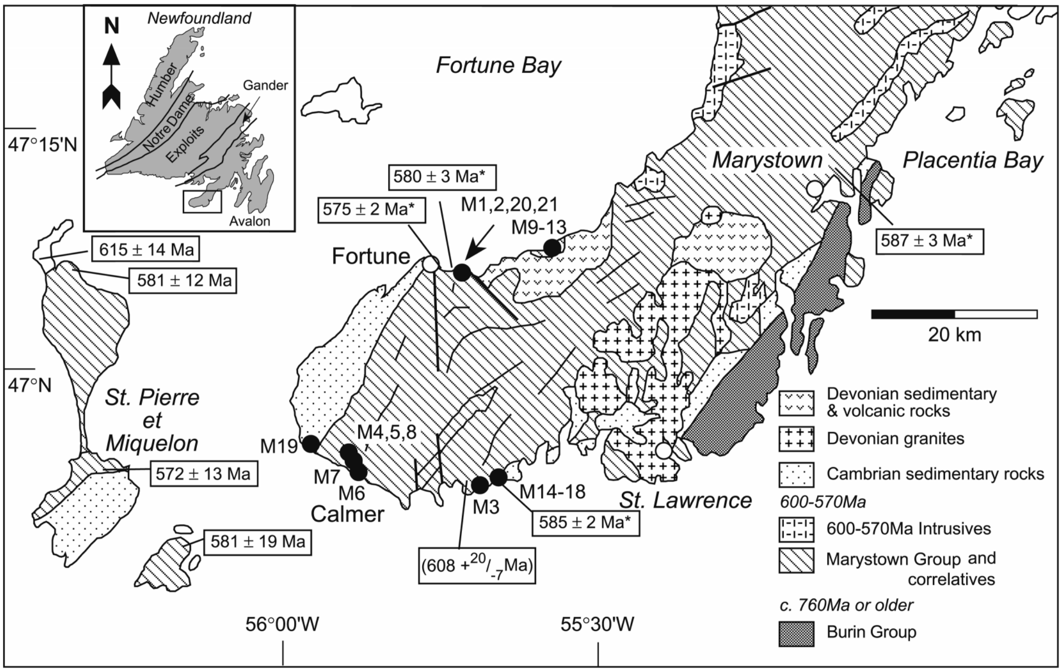


Figure 3. Generalized geological map of the Burin peninsula (modified after O'Brien et al., 1977, 1996; Strong et al., 1978) showing the sampling locations and structural trends. The age determinations, from Krogh et al. (1988) and Rabu et al. (1996), are based on U-Pb and Pb-Pb analyses; they also include unpublished data (asterisk) from Sean O'Brien and Tom Krogh (2000, written commun.).

TABLE 1. HIGH-TEMPERATURE MAGNETIC COMPONENTS FROM THE MARYSTOWN GROUP, SOUTHERN BURIN PENINSULA, NEWFOUNDLAND

Site	Latitude (N)	Longitude (W)	Strike/dip	N	Nsp	Nd	D/I (in situ)	D/I (t-c)	D/I (t-s-c)	κ	α_{95}	Paleolatitude
M1	47°06′	55°43′	182/58	8	8	8	292/−79	87/−42	85/−42	94	6	25
M2	47°06′	55°43′	185/60	6	6	8	157/79	263/35	258/35	26	14	19
M3	46°52′	55°41′	220/45	6	6	8	344/50			168	5	
M4	46°54′	55°54′	185/35	6	8	6	47/84	283/59	278/59	174	5	39
M5	46°54′	55°54′	190/30	7	8	7	186/88	275/60	265/60	40	10	41
M6	46°53′	55°53′	224/70	7	7	7	345/76			90	6	
M7	46°54′	55°53′	209/55	4	5	7	191/−6			53	13	
M8	46°54′	55°54′	189/20	6	6	9	283/76	281/56	272/56	45	8	37
M9	47°07′	55°36′	095/04	9	14	9	172/51	173/47	258/47	771	2	28
M10	47°07′	55°36′	Horizontal	10	10	10	173/51	173/51	263/51	559	2	32
M11	47°07′	55°35′	Horizontal	8	8	8	182/52	182/52	272/52	277	3	33
M12	47°07′	55°35′	090/05	8	9	8	186/52	186/47	276/47	93	6	28
M13	47°07′	55°35′	068/10	10	11	10	209/55	200/48	312/48	412	2	29
M14	46°53′	55°39′	014/23	4	4	9	112/−27	116/−49	102/−49	37	15	30
M15	46°53′	55°39′	015/25	0	9	9	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
M16	46°53′	55°39′	025/08	5	5	5	090/−59	082/−66	057/−66	41	12	49
M18	46°53′	55°39′	020/20	6	6	6	081/−45	065/−61	045/−61	106	7	42
M19	46°54′	55°57′	230/12	0	8	8	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
M21	46°57′	55°59′	183/60	7	8	8	070/20			19	14	
		means					217/72 κ = 8 α_{95} = 16	237/61 κ = 8 α_{95} = 15	268/53 κ = 30 α_{95} = 8			34

Note: Strike/dip is the strike and dip of bedding (down dip to the right of strike); N and Nsp are the number of samples and specimens used in the analysis; Nd is the total number of samples that were thermally demagnetized; D/I is declination and inclination (degrees), t-c is tilt-corrected and t-s-c is tilt and strike-corrected; κ is the Fisher precision parameter; α_{95} is the 95% cone of confidence around the mean direction. Paleolatitude is calculated after tilt and strike corrections. Results from sites M3, M6, and M7 are not used in the calculation of the means. N.D.—not determined.

lowing tilt and strike tests to be performed. Bedding at sites 10 and 11 is horizontal. Samples were oriented using a magnetic compass, because no discernible compass deflections were observed in the field and the magnetic intensities of the samples make it improbable that local magnetic effects have caused erroneous orientation measurements.

PALEOMAGNETIC ANALYSIS

The natural remanent magnetization (NRM) was measured with a 2G cryogenic magnetometer housed in a magnetically shielded room (rest field <200 nT) at the University of Michigan and with a CCL cryogenic magnetometer at the University of Oxford. The stability of NRM for a total of 150 samples (out of 158 collected) from the stratigraphic sections was tested by stepwise thermal demagnetization; between 12 and 24 demagnetization steps were used. Stepwise thermal demagnetization was also carried out on nine samples from the conglomerate at site 17, and 11 samples from the volcanic agglomerate at

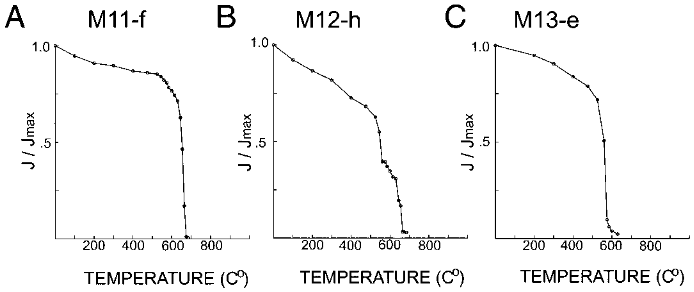


Figure 4. Magnetization versus temperature plots for three representative samples of the Marystown Group during progressive thermal demagnetization. J is the magnetization of the sample at a given temperature; J_{max} is the maximum magnetization (NRM). A: Results from sample M11-F show a sharp decrease in magnetization at ~ 680 °C, indicating hematite as the carrier of the magnetic remanence. B: Sample M12-H exhibits sharp decreases in magnetization at ~580 °C and ~680 °C, suggesting that both hematite and magnetite are carriers of the magnetic remanence. C: Sample M13-e has a sharp intensity decrease at ~580 °C, indicating magnetite as the carrier of the remanence.

site 20. Alternating field demagnetization up to 130 mT was performed on a small batch of samples but failed to adequately demagnetize them.

Demagnetization results were visually inspected in orthogonal and stereographic projections; characteristic remanence components were determined using least-squares algorithms (Kirschvink, 1980). Two or more components of magnetization were present in almost all samples, and occasionally these components overlapped in their stability to thermal treatments, yielding curved trajectories in orthogonal plots. No coherent components of magnetization were forthcoming from two sites (M15 and M19).

Some of the samples consist of multiple specimens; the magnetization directions from separate specimens are averaged, and the resulting direction is assigned to that sample. The directions so obtained from the samples within a site are averaged to produce the site mean of that site (Table 1).

PALEOMAGNETIC RESULTS

Most samples either lose their magnetization completely at 680 °C (Fig. 4A), indicating hematite as the carrier of the magnetic remanence, or exhibit sharp decreases of magnetization at 580 °C and at 680 °C (Fig. 4B), suggesting both magnetite and hematite as carriers of the remanence. On the basis of the subaerial nature of the basalt flows, it is probable that the hematite formed, due to oxidation, shortly after deposition (e.g., Geissman and Van der Voo, 1980). Samples from site M13 lose their magnetization at 580 °C, indicating magnetite as the main carrier of the remanence (Fig. 4C).

The anisotropy of magnetic susceptibility of samples from 15 sites was measured on a Kappabridge KLY-2 magnetic susceptibility meter to examine whether a preferred magnetic direction could be responsible for a deflection of the remanence direction from the true paleofield. Only a few sites display magnetic fabrics consistent with bedding. Most sites are characterized by an anisotropy pattern, unrelated to bedding, which is inconsistent between sites at the same locality. The grouping of principal directions, moreover, is generally poor, and the magnitude of anisotropy of magnetic susceptibility is weak ($K_{max}/K_{min} < 1.035$). Moreover, there is no evidence of penetrative deformation at any of our sites; therefore, we conclude that anisotropy does not influence the remanence direction.

The NRM intensity of most samples ranges from 3 to 210 mA/m. The NRM directions do not differ greatly from the high-temperature directions obtained after demagnetization, indicating that the high-temperature component is the dominant magnetization of the NRM. Demagnetization results for representative samples (Fig. 5) are described in tilt-corrected coordinates.

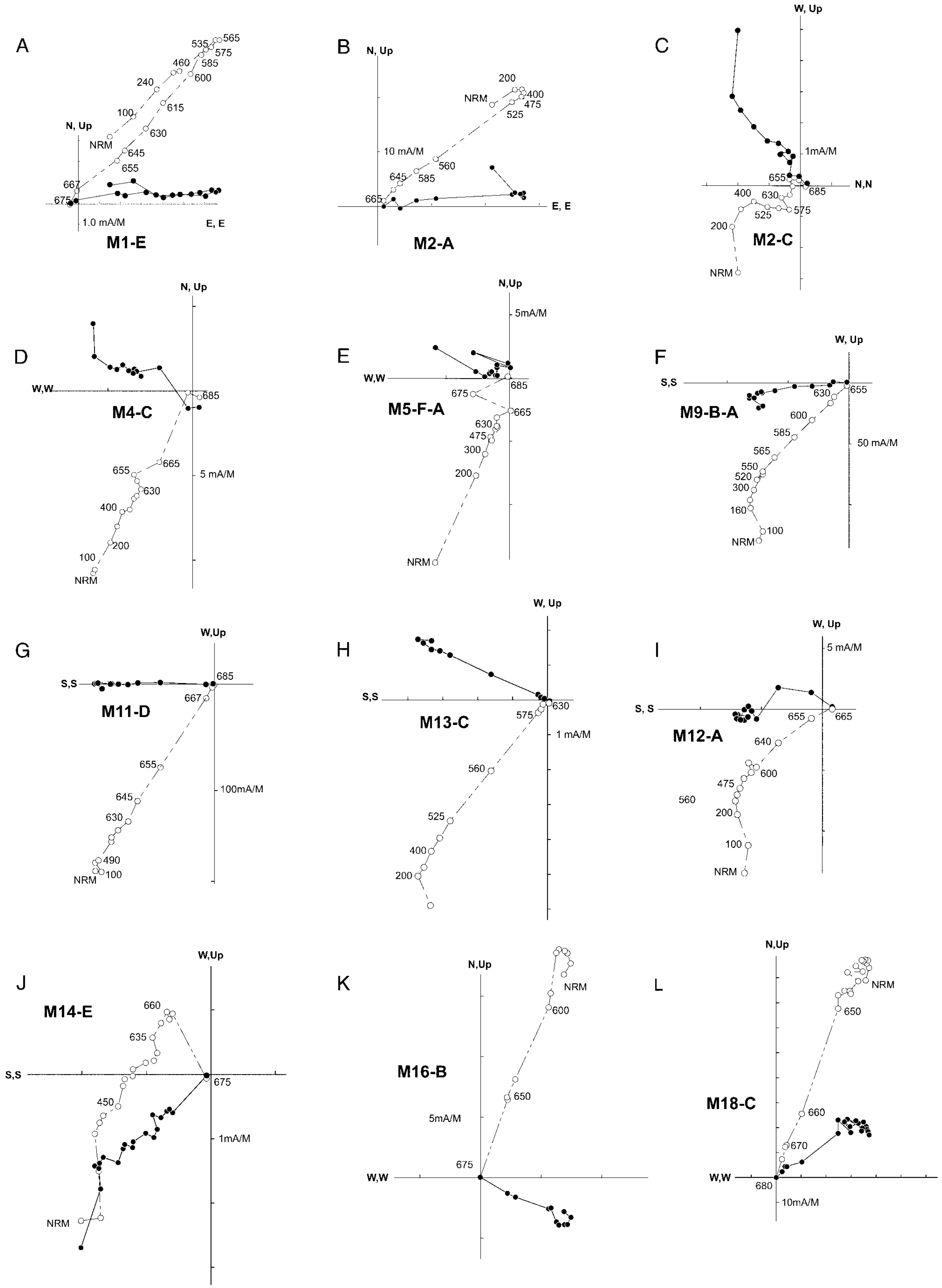
Samples from site M1 are derived from red siltstones of the Grand Bank sequence of O'Brien et al. (1977); all display two components that are opposite in direction. Sample M1-E is representative of this group (Fig. 5A). The first component is removed by 565 °C, and its direction is westward, with an intermediate downward inclination. The remaining high-temperature magnetization trends eastward, with an intermediate upward inclination.

Site M2 consists of samples taken from a sequence of red basaltic flows. High-temperature directions are upward for three samples and downward for three others. Sample M2-A (Fig. 5B), a representative sample, shows a soft downward component that is removed by 475 °C, whereas the resulting high-temperature direction is eastward and has an intermediate upward inclination. Sample M2-C (Fig. 5C) shows multiple components of magnetization that partially overlap, creating a curved trajectory from 200 to 575 °C. The remaining high-temperature component lacks a clear trajectory but has an overall southwestward and intermediate downward direction. Two samples appear to have multiple components that were demagnetized simultaneously, so a single high-temperature component could not be derived from them.

Samples from sites M4, M5, and M8 were collected from red mudstones at Calmer. Above 630 °C, the magnetizations became erratic, possibly owing to growth of new magnetic minerals; the magnetic susceptibility was remeasured after thermal demagnetization for a few specimens and was found to have increased by a factor of 4 to 8. The demagnetization behavior of samples M4-C and M5-F-A (Fig. 5, D and E) is typical of these sites. A soft component is removed by 200 °C for sample M4-C and by 300 °C for sample M5-F-A. Westward directions with intermediate downward inclinations are well defined below 630 °C.

Samples from sites M9, M10, M11, and M13 are derived from porphyritic basalts at Famine Back Cove; all reveal a strikingly similar and easily interpreted demagnetization behavior (Fig. 5, F–H). The demagnetization of M9-B-A unveils at least two components. The first is removed by 550 °C. The remaining direction is southward and has an intermediate downward inclination. Sample M11-D also reveals minor low-temperature components,

Figure 5. Representative orthogonal plots in tilt-corrected coordinates illustrating the thermal demagnetization behavior of samples from the Marystown Group. Open symbols represent projections in the vertical plane; solid symbols are projections in the horizontal plane. Numbers beside the data points represent treatments, in degrees centigrade.


A
M1-E
N, Up
E, E
1.0 mA/M
B
M2-A
N, Up
E, E
10 mA/M
C
M2-C
W, Up
N,N
1mA/M
D
M4-C
N, Up
W,W
5 mA/M
E
M5-F-A
N,Up
W,W
5mA/M
F
M9-B-A
W, Up
S,S
50 mA/M
G
M11-D
W,Up
S,S
100mA/M
H
M13-C
W, Up
S,S
1 mA/M
I
M12-A
W, Up
S, S
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J
M14-E
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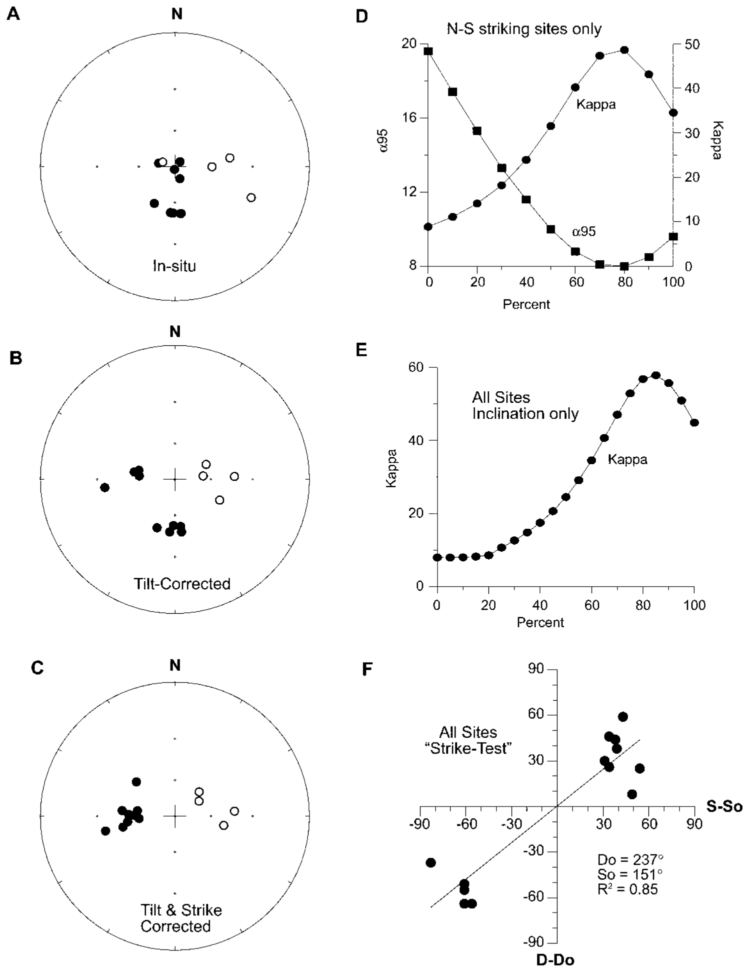

Figure 6. Site mean directions and structural tests for the Marystown Group, plotted in equal-angle projections. Open symbols represent upward directions; solid symbols are downward directions. (A) In situ site means. (B) Site means after tilt correction. (C) Tilt-corrected site means, rotated about a vertical axis relative to an arbitrary reference strike of 180°. (D) Incremental fold test of north-south–striking sites where κ is the Fisher precision parameter and α^{95} is the 95% cone of confidence. Zero percent unfolding represents in situ; 100% unfolding represents fully tilt corrected. (E) Inclination-only incremental fold test. (F) Strike test where D_o is average declination of all sites, D is declination of individual site, S_o is average strike of all sites, and S is strike of individual site. R^2 is the coefficient of determination where values range from 0 to 1, and 1 signifies a perfect correlation. For all of the field tests performed, reversed sites were first inverted through the origin to the antipodes.

which are removed by 490 °C. For other samples in this site, the temperature at which this component is removed varies from 400 to 600 °C. The remaining direction is also southward and has an intermediate downward inclination. In sample M13-C, the low-temperature magnetization is removed by 200 °C. A few samples of site 13 lack this component entirely. The remaining direction is south-southwestward, with an intermediate downward inclination.

Samples from site M12 are from porphyritic basalts at Famine Back Cove and reveal at least two components, which appear to be demagnetized simultaneously, judging from the curved trajectory (Fig. 5I). At temperatures above 600 °C, however, a single component becomes recognizable, but its direction is determined by only a few data points, and its mean trajectory bypasses the origin. Sample M12-A is representative of this site; its high-temperature component is south-southwestward and has an intermediate downward inclination.

Samples from sites M14 to 18, from red mudstone and porphyritic lava at Pump Cove, generally reveal two components of magnetization. After removal of a steeply dipping component, at temperatures of up to 600 °C, a stable high-temperature component is isolated, generally directed upward to the east in tilt-corrected coordinates (e.g., Fig. 5J–L).

This high-temperature component is not as well defined for site M14 as for the other sites, due to the extremely narrow unblocking temperature range (Fig. 5J).

Rejected Sites

Sites M3, M6, M7, and M21 could not be used in the analysis. All but one of the samples from site M3 exhibit high-temperature directions similar to the present-day field, so they are most likely viscous remanent magnetizations. Site M6 has steeply down and northward directions in situ (somewhat steeper than the present-day field), but after tilt-correction, M6 has very shallow northwestward directions. These directions do not resemble any tilt-corrected directions from any other site, and they have been excluded. Site M7 reveals three distinct high-temperature directions, which are discussed below. Four samples exhibit shallow, near-horizontal south-southwestward components; such directions elsewhere have been ascribed to late Paleozoic overprints (Irving and Strong, 1985; Lombard et al., 1991; Potts et al., 1993). Another high-temperature component is observed in two specimens from the same core. This component has a southeastward and intermediate downward direction, but there are not enough demagnetization steps to confidently assign a direction. Five samples have a high-temperature direction that is southward and steeply down. This direction, when tilt-corrected, would coincide with directions obtained from other sites, but the directions do not approach the origin and their trajectories are not well defined, so these results were not included. Site M21 yields a high-temperature component that is directed to the east, with a shallow downward inclination of in situ coordinates. This direction steepens upon tilt-correction and is unlike any direction observed at the other sites at this locality. Data from this site have, therefore, been excluded from the overall analysis.

Tilt and Strike Tests

We plotted site mean directions for the high-temperature components (Table 1) on stereonets in both in situ and tilt-corrected coordinates (Fig. 6, A and B). Because M2 consists of mixed upward- and downward-direct-

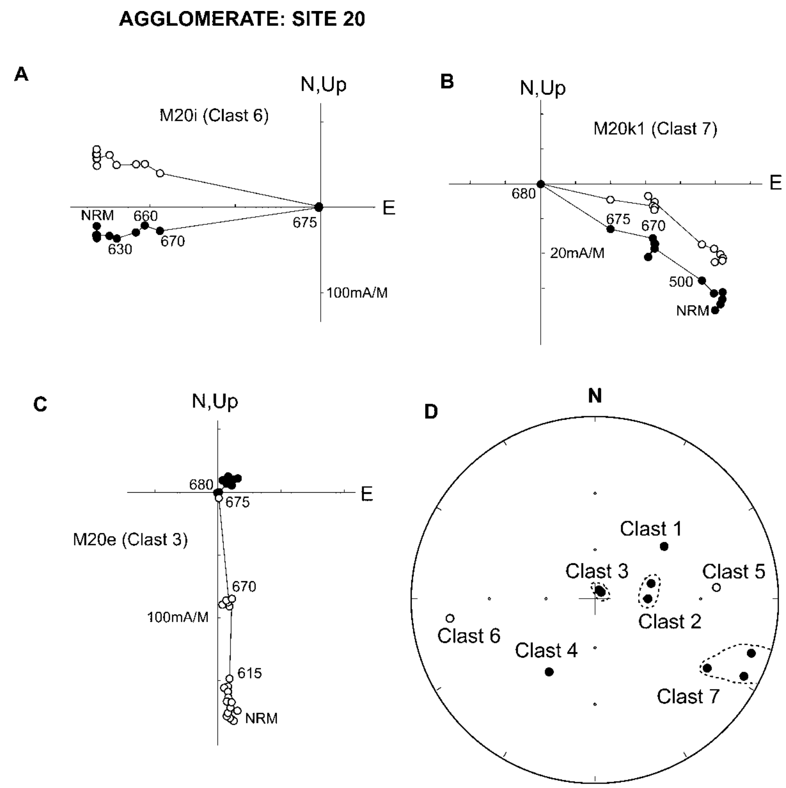

Figure 7. (A–C) Representative orthogonal plots of thermal demagnetization behavior of samples from the clasts in the agglomerate at site 20. Conventions and symbols are as for Figure 5. Marystown Group. (D) Stereographic projection illustrating the high unblocking temperature components from the clasts in the agglomerate. Conventions and symbols as for Figure 6. Directions from the same clast are circled (dashed) and yield a positive agglomerate test at the 95% confidence level (N = 7, R = 3.18, R_c = 4.18).

ed high-temperature magnetizations, all upward directions are inverted through the origin to downward antipodes when calculating the site mean. In each stereoplot (Fig. 6), separate sets of directions are observed. For the in situ site means, one set has a very steep downward or upward inclination and one set has an intermediate upward and eastward direction, whereas the final set has a southward direction and an intermediate downward inclination (Fig. 6A). For the tilt-corrected site means, two groups with similar intermediate inclination are observed, but the declinations of one grouping are east-west, of dual polarity, whereas those of the other are southward (Fig. 6B). The presence of these two groups with similar inclinations suggests possible rotations about a vertical axis. Such rotations interfere with a standard fold test, so we first analyzed only those sites with a north-south strike. These sites yield a positive fold test at the 95% confidence level (Fig. 6D), although we note that there is a peak in the precision parameter, and a corresponding minimum in the α_{95} (α is the 95% cone of confidence around the mean direction) at ~80% unfolding. This peak, however, is not statistically significant when compared with that obtained at 100% unfolding. We then considered all sites using the McFadden and Reid (1982) inclination-only fold test, because of the large differences in the strike (e.g., Parés et al., 1994). This test is again positive, maximum κ (the Fisher precision parameter) at 85% unfolding (Fig. 6E). The difference in κ values at 85% and 100% unfolding is small and statistically similar to the prefolding magnetization. We therefore conclude that the magnetization is prefolding.

Several scenarios may cause different declinations yet similar inclinations after tilt-correction: plunging folds, pretilting, and post-tilting vertical rotations. In the case of plunging folds, the strikes of the units have been rotated during plunging, complicating the method of simply untilting units to obtain the prefolding magnetization direction because present strikes do not represent the original strikes. For plunging folds and pretilting vertical rotations, if the strike of each site were brought into alignment with an arbitrary reference strike direction, the tilt-corrected directions would not be expected to cluster. In the case of post-tilting vertical rotations, however, the tilt-corrected directions would be expected to cluster when the strikes are brought into alignment. We note that there is a lack of field evidence for plunging folds and that analysis of bedding orientations indicates that plunges are $<2°$. The strike test that was performed is described below. It allows us to make a choice between these scenarios and was found to be positive (Fig. 6F), although we note that we are actually dealing with only two groups of strike-declination pairs. The presence of a positive inclination-only fold test and a positive, though limited, strike test indicates that the region underwent folding followed by later rotations about a vertical axis.

The strike of each site was rotated to an arbitrary reference strike of 180°. The tilt-corrected mean magnetic directions for each site were passively rotated along with the bedding orientation, except for sites M10 and M11, which have horizontal bedding. To overcome this problem for sites M10 and M11, strikes of 90° are assigned to them, on the basis of strike values of the nearby site M12. This is reasonable because all of the sites in that particular area have similar strikes (Table 1). In Figure 6C, the stereonet projection after strike correction, the individual azimuths are arbitrary, but their relative declinations allow statistical analysis. The resultant mean inclination is 53°, κ = 30 and α_{95} = 8°.

Agglomerate and Conglomerate Strike Tests

As noted previously, we analyzed nine samples from six clasts collected from a conglomerate at site 17, and 11 samples from seven clasts from a volcanic agglomerate at site 20. The samples from the conglomerate at site 17 yielded stable low-temperature directions, which were pointed steeply down in in situ coordinates. At temperatures above 500 °C, demagnetization behavior became erratic. Although the trends are not sufficiently well developed to confidently assign individual directions, they do appear to have random orientations. The samples from the agglomerate at site 20 yielded stable high-temperature components, isolated at temperatures above 650 °C (Fig. 7, A–C), which were consistent within clasts but randomly directed between clasts (Fig. 7D). A statistical analysis of the clast mean directions yielded a positive agglomerate test at the 95% confidence level (N [number of clasts] = 7, R [length of the resultant vector] = 3.18, R_c [critical value of R for N = 7] = 4.18). This indicates a primary origin for the high-temperature component in the clasts; thus, we can attribute a primary age of magnetization to our results for the Marystown Group as a whole.

DISCUSSION AND CONCLUSIONS

The high-temperature component in the Marystown Group is likely a primary magnetization. It predates middle Paleozoic (Silurian-Devonian; Dallmeyer et al., 1983; O'Brien et al, 1996) deformation, as evidenced by convergence of the directions after correction for the tilts and strikes of bedding and the positive agglomerate test. We note that Hodych (1991) also reported a preliminary positive conglomerate test for the Famine Back Cove basalts, although the precise age of the conglomerate was not reported. Given the positive field tests, the presence of reversals (sites M1, M2, M14, M16, and M18), and the high stability of the magnetizations, we conclude that the magnetization of the Marystown Group dates to the time of deposition; that is 580–570Ma. Our results agree with those reported by Irving and Strong (1985) from the Famine Back Cove basalt and Calmer mudstone, having a mean paleolatitude of ~35°, although their results were based on a much smaller collection of samples, and they did not perform or discuss the necessary strike test. Their results, therefore, appeared to lack coherence and were not considered as evidence for Avalon's paleogeographical position. Our detailed results, however, similarly indicate a paleolatitude of ~34° for western Avalon at ca. 580–570 Ma.

We first examine the situation in which Amazonia is rotated relative to Laurentia using the rotation parameters of Dalziel et al. (1994) (Fig. 8A). Second, we consider an alternative Laurentia-Amazonia fit in which western Amazonia is juxtaposed to Labrador and western New England, as suggested by Hoffman (1991) and paleomagnetically indicated by Weil et al. (1998) (Fig. 8B). These reconstructions assume, of course, that Laurentia and Amazonia were still juxtaposed at this time while recognizing the debate about the precise timing of the rift-to-drift transition of the opening of the Iapetus ocean. Nevertheless, available estimates indicate that the separation of Amazonia and Laurentia and the rift-to-drift transition on the Appalachian margin took place in the latest Precambrian or at the Precambrian-Cambrian boundary (e.g., Williams and Hiscott, 1987); thus, the use of Laurentian paleopoles as proxies for Amazonia at 580 Ma is appropriate. If West Africa was already completely or nearly juxtaposed to Amazonia by ca. 580 Ma, the two cratons would be approximately in the positions shown in the reconstructions. However, recall that the timing of the assembly of West Gondwana is not well determined because the existing paleomagnetic data older than 550 Ma are not abundant enough to test this issue (Meert and Van der Voo, 1997). Trompette (1997) proposed that the West African, Amazonian, and Rio de la Plata cratons were the constituents of a continental assembly that was a fragment of Rodinia. Fragmentation of this assembly, containing short-lived, continental basins, did not occur until 600 Ma or slightly later, and was not successful in breaking up the newly formed West Gondwana supercontinent. Thus, although West Africa may not have been in its later West Gondwana configuration relative to Amazonia at ca. 580 Ma, the two cratons were likely close to one another.

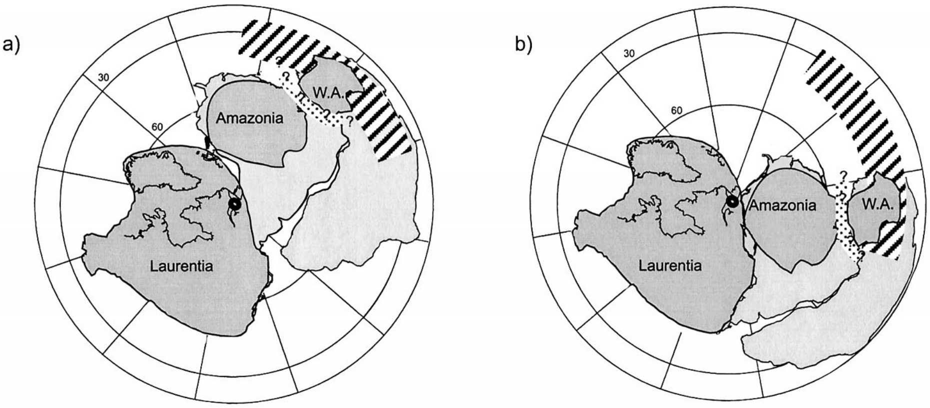

Figure 8. Reconstructions of Laurentia, Amazonia, and West Africa (W.A.) at ca. 580 Ma based on a Laurentian pole taken from Torsvik et al. (1996). Although the position of West Africa is uncertain at 580 Ma, we have positioned it relative to Amazonia using the rotation parameters of Rabinowitz and LaBrecque (1979) (Africa relative to South America: Euler pole; lat 45.5°, long 327.8°, rotation angle −57.5°) on the assumption that West Gondwana had already been assembled at this time (Trompette, 1997). A paleolatitudinal band (diagonal lines) illustrates the likely position of Avalonia at 580 Ma, based on the mean inclination (53° ± 8°). The dot pattern represents the uncertainty associated with the precise paleogeographical connection between Amazonia and West Africa. (A) Laurentia-Amazonia fit based upon the rotation parameters of Dalziel et al. (1994) (Amazonia relative to Laurentia: Euler pole; lat 2.3°, long 336.4°, rotation angle −99.34°). (B) Laurentia-Amazonia fit suggested by Hoffman (1991) and paleomagnetically supported by Weil et al. (1998) (Amazonia relative to Laurentia: Euler pole; lat 9.5°, long 315°, rotation angle −96.5°).

Given a date for the Marystown Group of ca. 580 Ma and accepting a more recent rift-to-drift transition between Laurentia and Amazonia, our determination of a 34° +8°/–7° paleolatitude precludes the possibility of an Amazonian origin for Avalon. In fact, the minimum latitudinal distance between Avalon and Amazonia would be more than 1100 km in the reconstruction of Dalziel et al. (1994) and more than 2200 km in the reconstructions of Hoffman (1991) and Weil et al. (1998). Instead, the paleolatitude herein derived for the Marystown Group coincides with that of the northern margin of West Africa.

Although paleolongitude indeterminacy provides a degree of freedom, geological evidence further delimits the position of the Avalon terrane. Avalonian rocks are of Gondwanan affinity, on the basis of lower Paleozoic cover sequences containing distinctive Acadian-Baltic faunas that rest above Proterozoic basement (e.g., O'Brien et al., 1996). The predominance of silicic, pyroclastic volcanics and comagmatic plutonic rocks within Avalon indicates the presence of continental basement (O'Brien et al., 1983), and geochemical data also indicate a continental origin for basalt within the Marystown Group (Strong et al., 1977). Because the Avalon terrane has Gondwanan continental affinity and an Amazonian origin is discounted paleomagnetically, the West African cratonic margin is the most suitable candidate for the position of Avalon. This is further supported by the observation that rocks from the Avalon terrane also have ages, lithologies, and tectonic histories similar to those of the Pan-African belts in West Africa (O'Brien et al., 1983; Nance et al., 1991).

Our location of Avalonia in Figure 8 contrasts with the reconstruction of Nance and Murphy (1994, 1996) (Fig. 2B), who have argued, on the basis of initial ϵN_d values, Nd model ages and the possible provenances of detrital zircons in Avalonian rocks, that Avalonia was close to the northern margin of Amazonia in the latest Precambrian. Initial ϵN_d values mark departures of the measured $^{143}Nd/^{144}Nd$ ratio from that of a chondritic uniform reservoir (CHUR; considered to be representative of the bulk earth) at the time of formation of the rock (expressed in parts per 10^4 deviation from the CHUR evolution line; DePaolo and Wasserberg, 1976). If the CHUR line defines the initial ratios through time, measurements of $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ should also yield a model age for the formation of that rock (or its precursor) from the chondritic uniform reservoir (T_{CHUR} [age at which the material was extracted from the chondritic reservoir]), assuming that the Sm/Nd systematics have not subsequently been disturbed. However, because the mantle preferentially retains Sm over Nd, initial ϵN_d values exhibit a departure from the CHUR line, when plotted against time, such that young mantle-derived rocks have high positive initial ϵN_d values. With a knowledge of the initial ϵN_d values of the mantle through time, model ages can also be calculated for when a given rock (or its precursor) was extracted from the mantle (T_{DM} [age at which the material was extracted from the depleted mantle]; DePaolo, 1981).

Nance and Murphy (1994, 1996) pointed out that Neoproterozoic volcanic and intrusive rocks from western Avalon and the Tocantins province of central Brazil (see Fig. 2B for location) have similar positive initial ϵN_d values (–0.4 to +5.0 and +0.2 to +6.9, respectively) and similar T_{DM} model ages (ca. 0.8–1.1Ga and 0.9 to 1.2 Ga, respectively). This comparison, however, only suggests a similar mantle extraction age of the protolith, positive initial ϵN_d values indicating a relatively juvenile basement. The depleted nature of these rocks may also be used as evidence for a similar tectonic setting, namely an arc devoid of basement older than ca. 1.2 Ga. This comparison, therefore, only indicates a similar tectonic regime; it does not establish a spatial or geographic connection between the two areas in the late Precambrian, especially given that arc magmatism in Avalonia is significantly younger than in the Tocantins province. Thus, although comparisons of the initial ϵN_d and T_{DM} model ages permit a link between Avalon and Amazonia, they are not unique and do not preclude other possibilities.

Nance and Murphy's (1994, 1996) linkage of western Avalonia with the Amazonian craton is, therefore, based primarily on the range of ages of detrital zircons obtained from metasedimentary units of western Avalonia. Several of these studies have yielded zircons that match the Liberian, Eburnian, and Pan-African ages of West Africa, although some have yielded ages between 1.7 and 1.0 Ga for which there is no apparent source in West Africa. However, African proximity of Avalonia in the late Precambrian does not preclude the possibility of Amazonian sources of (1.7–1.0 Ga) detrital zircons in late Neoproterozoic Avalonian sediments. Placing Avalonia in a position adjacent to the West African craton in the latest Precambrian, as suggested by the paleomagnetic data, does not result in a location that is farther removed from a source area in the Tocantins province than a position adjacent to the northern margin of Amazonia (Fig. 2B). Whereas detrital zircons may yield evidence of potential sediment sources, they do not yield information on how far they have been transported prior to deposition. We conclude, therefore, that west Avalonia is best located next to the West African cratonic margin at 580–570 Ma.

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