Revisiting the paleomagnetism of the 1.476 Ga St. Francois Mountains igneous province, Missouri

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[1] A paleomagnetic investigation of the St. Francois Mountains igneous province in southeastern Missouri provides a key 1476 ± 16 Ma paleomagnetic pole for Laurentia. The pole (13.2°S, 219.0°E; dp = 4.7°, dm = 8.0°) is considered primary on the basis of positive conglomerate, inverse baked contact, and fold tests. An analysis of 1470–1430 Ma poles from Laurentia highlights key differences between poles obtained from the Belt Supergroup, Electra Lake gabbro, and cratonic North America. Paleolatitudes based on the Lower Belt Supergroup poles are enigmatic, as two previous studies yielded a difference of 10°. Our new pole, combined with an analysis of previous results, favors the higher latitude interpretation for the Lower Belt Supergroup. Paleolatitudes from the younger Belt rocks indicate lower latitudes than coeval rocks from elsewhere in Laurentia for which there has been no adequate explanation. A comparison of the St. Francois Mountain pole with similar-age poles from Baltica, Siberia and Australia allow first-order tests of proposed continental configurations. Paleomagnetic data from Australia are compatible with proposed Rodinia reconstructions, whereas paleomagnetic data from Baltica are not. We are unable to rigorously test the alternative suggestion that places Siberia against the western margin of Laurentia due in part to large errors associated with Siberian paleomagnetic data. INDEX TERMS: 1525 Geomagnetism and Paleomagnetism: Paleomagnetism applied to tectonics (regional, global); 1527 Geomagnetism and Paleomagnetism: Paleomagnetism applied to geologic processes; 8110 Tectonophysics: Continental tectonics—general (0905); 8157 Tectonophysics: Plate motions—past (3040); KEYWORDS: Paleomagnetism, Mesoproterozoic, supercontinents, St. Francois Mountains, reconstructions

1. Introduction

[2] Mesoproterozoic continental configurations between Siberia, the elements of East Gondwana, and Laurentia are controversial. The controversy arises, at least in part, because of a paucity of high-quality paleomagnetic data from these continents as well as discordance of the extant results [see Harlan and Geissman, 1998]. On the one hand, Sears and Price [2000] argue for a northeastern Siberian conjugate margin with present-day western Laurentia, whereas others [Hoffman, 1991; Condie and Rosen, 1994; Frost et al., 1998, and references therein] link Siberia to the present-day northern margin of Laurentia with some variation in orientation. The placement of Siberia against the northern margin follows from suggestions that position Australia and Antarctica against the western margin of Laurentia in either the southwest United States-East Antarctica (SWEAT) or Australia-western United States (AUSWUS) configuration [Dalziel, 1997; Karlstrom et al., 2000; Barrett and Berry, 2000]. Interestingly, none of these configurations has strong paleomagnetic support [Meert, 1999; Torsvik et al., 2001; Meert and Powell, 2001], although ideally, paleomagnetism could distinguish among the various models if there are high-quality poles from the various cratonic elements in question [Ernst et al., 2000; Torsvik et al., 1996]. A recent attempt to test possible linkages between Siberia and Laurentia by Ernst et al. [2000] used a new 1503 ± 5 Ma paleomagnetic pole from the eastern Anabar shield region of the Siberian craton. The major limitation to their analysis was a lack of coeval paleomagnetic poles from Laurentia and the large error associated with their Siberian pole (see section 4.2). Furthermore, if one is to test alternative reconstructions for Siberia during this interval of Mesoproterozoic time, it would require coeval data from Australia and Antarctica since constituent cratons within these landmasses are also argued to occupy the region adjacent to present-day western Laurentia [e.g., Dalziel, 1997; Karlstrom et al., 2000].

[3] The St. Francois Mountains (SFM) region of Missouri is a 1476 ± 16 Ma igneous province in central Laurentia (Figure 1; mean age compiled from Van Schmus et al. [1993] excluding Munger granite porphyry). Previous paleomagnetic studies [Hsu et al., 1966; Hays and Scharon, 1966] were inconclusive in demonstrating a primary magnetization from these rocks, although directional comparisons between similar-aged units (Michikamau intrusion and Harp Lake Complex of the Canadian shield) suggested that a primary magnetization might be preserved. Nevertheless, the magnetization age of both the Michikamau and Harp Lake intrusive units may postdate their 1450–1460 Ma U-Pb age thereby negating a direct comparison with the SFM poles (see discussion in section 4.1). Harlan and Geissman [1998] compared paleomagnetic data from the Belt Supergroup (1400–1470 Ma), the Electra Lake gabbro (1433 Ma) and midcontinent poles from North America (including the early SFM studies) and argued that possible rotations of the Belt Supergroup, Electra Lake gabbro (1433 Ma), or both could explain the discrepancy in paleomagnetic poles from these units. In an effort to test some of these continental configurations and any
establish the tectonic relationships between paleomagnetic poles from Laurentia, we resampled the 1476 Ma SFM province in Missouri during the summer of 1999 (Figure 1). While we cannot, with a single well-dated pole, provide rigorous constraints on paleoreconstructions for this time interval, we can begin to build a database from which to test the various tectonic models and alternative reconstructions.

1.1. Geologic Setting and Age

[4] The St. Francois Mountains (SFM) of southeastern Missouri (Figures 1, 2a, and 2b) consist of nearly 40,000 km² of acidic volcanic and plutonic rocks [Berry, 1976; Kisvarsanyi, 1980] of which ~900 km² are exposed at the surface. The SFM form the uplifted core of the Ozark Dome and are overlain by flat lying or gently dipping lower Paleozoic rocks. The lowermost Paleozoic rocks overlying the SFM consist of either a boulder conglomerate of presumed Cambrian age or the Upper Cambrian Lamotte sandstone (~500 Ma). There are also rare occurrences of paleosol between the SFM rocks and the overlying Lamotte sandstone.

[5] The SFM represent the exposed northeastern terminus of a much larger, and mostly subsurface, Mesoproterozoic granite-rhyolite province in North America that flanks the southern and eastern margins of the 1.8–1.6 Ga Great Plains Orogen and the eastern margin of the Colorado Province. It extends in the subsurface from the Texas panhandle to southeastern lower Michigan [Van Schmus et al., 1987, Figure 1].

[6] The geologic history of the SFM begins with the main caldera-forming eruptions, caldera collapse, and intrusion of shallow magmas into their own ejecta at 1476 ± 16 Ma [Kisvarsanyi, 1980]. A second cycle of alkaline intrusion and magmatism occurred around 1.38 Ga [Kisvarsanyi and Kisvarsanyi, 1989; Lowell and Darnell, 1996], followed by a volumetrically smaller episode of primarily mafic magmatism (gabbroic intrusions, dike swarms, and minor flows) at ~1.33 Ga [Lowell and Young, 1999; Ramo et al., 1994; R. Tucker, personal communication, 2000]. The rocks show minor postemplacement metamorphism except in regions where (1) the volcanic rocks are intruded by their parent magmas following caldera collapse; (2) the rocks are intruded by a suite of younger granitic intrusions at 1.38 Ga or by mafic bodies at 1.33 Ga, and (3) there is mineralization and faulting associated with these younger magmatic episodes or subsequent reactivation [Clendenin et al., 1989; Lowell, 1991; Kisvarsanyi and Kisvarsanyi, 1989]. The dominant
structural features in the SFM are a series of caldera collapse structures that caused quasi-radial tilting of the volcanic rocks and, in some cases, intrusion of the collapsed volcanic rocks by their parental magmas [Kisvarsanyi, 1980; Sides et al., 1981; Lowell, 1991]. The main calderas in the sampling region are the Butler Hill caldera (Figure 2a) [Lowell, 1991] and the Taum Sauk caldera (Figure 2b) [Anderson et al., 1969]. There are also a number of Neoproterozoic faults related to the opening of the Reelfoot rift (Figure 1) that were reactivated during Paleozoic and younger times [Clendenin et al., 1989; Kisvarsanyi, 1980].

1.2. Previous Work

[?] Paleomagnetic studies in the SFM have a long history and appear to be one of the earliest paleomagnetic studies conducted in the United States (see Ph.D. thesis by Hays [1961] and M.A. thesis by Hsu [1962]). The results were eventually published by Hays and Scharon [1966], who calculated a paleopole at 5°N, 210°E (\(\alpha_{95} = 10^{\circ}\)). Although Hays and Scharon [1966] sampled almost exclusively in the volcanic units, they did not provide detailed site descriptions or apply any tilt correction to their data. Later that year, Hsu et al. [1966] published additional results from the SFM that agreed with the results of Hays and Scharon [1966] with a resultant paleopole at 0.9°S, 219°E (\(\alpha_{95} = 5.0^{\circ}\)). Hsu et al. [1966] did report a tilt-corrected direction for their samples, but they argued that the structures reflected primary flow features on the basis of increased scatter upon tilt correction. Their tilt-corrected paleomagnetic pole falls at 5°S, 214.4°E (\(\alpha_{95} = 6.2^{\circ}\)). Both studies applied blanket alternating field demagnetization to the samples between 50–80 mT, but as we show in section 2, these fields may not adequately resolve primary directions in all samples, whereas thermal demagnetization was more effective in separating components of magnetization in our study. As both previous studies were completed
prior to the widespread use of principal component analysis and orthogonal vector plots, details of demagnetization trajectories were not included. Subsequent geologic and structural studies showed that the tilting of the volcanic units occurred during collapse of the calderas, and therefore a tilt correction of the data might help determine the primary and secondary nature of the magnetization [Kisvarsanyi, 1980; Lowell, 1991, and references therein]. Given that the SFM province contains a number of mafic dikes, a boulder conglomerate, and tilting related to the collapse of the primary volcanic centers, we saw an opportunity to apply a number of stability tests to these rocks in order to ascertain the age of the magnetization in the rocks.

2. Methods

[8] A total of 154 samples from 22 sites within the St. Francois Mountains igneous province were collected with a water-cooled portable drill. Sites were distributed among both volcanic and intrusive rocks including samples from a younger (circa 1330 Ma) suite of mafic dikes and their host rocks. In addition, clasts of SFM volcanic and intrusive material were drilled from the Cambrian-age boulder conglomerate. All samples were oriented using both solar and magnetic compass in the field. Structural orientations were determined on the volcanic sequence of rocks for use in the fold test. The samples were then cut into individual cylindrical specimens, and the bulk susceptibility of each sample was measured using a Sapphire Instruments susceptibility bridge. A pilot selection of samples was chosen for stepwise thermal and alternating field demagnetization. In nearly every case, stepwise thermal demagnetization was able to more clearly define the individual vector components in the samples, and the remaining samples were treated using thermal methods. In an effort to determine the magnetic carriers within the samples, both isothermal remanence acquisition studies (IRM) and three-axis thermal demagnetization of IRM [Lowrie, 1990] tests were conducted. Samples were stored and measured in the shielded magnetic room at Indiana State University. Thermal and alternating field demagnetizations were carried out on an either an ASC-Scientific TD-48
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aMean site location, 37°30'N; 90°30'W. Abbreviations are as follows: n, number of samples used; N, number of samples run; Dec., declination; Inc., inclination; strike/dip using left-hand rule; k, kappa; precision parameter defined by Fisher [1953]; $\alpha_{95}$, cone of 95% confidence about the mean direction [Fisher, 1953]; VGP, virtual geomagnetic pole; dp, cone of 95% confidence about the paleomagnetic pole in the colatitude direction; dm, cone of 95% confidence about the paleomagnetic pole at a right angle to the colatitude direction; BH, Butler Hill Granite; KG, Knoblick Granite; SG, Slabtown Granite; GM, Grassy Mountain ignimbrite; CT, Basal conglomerate; SM, Silvermine Granite; FR, Lake Killarney Rhyolite; RG, Royal Gorge rhyolite; TS, Taum Sauk rhyolite; BM, Bell's Mountain rhyolite; LM, Lindsey Mountain rhyolite. Site 1S is not included in mean.

bDykes and baked contacts.
cCountry rock at dyke sites.
1Large dyke directions.
bBaked contact rocks.
cConglomeratic blocks.
thermal demagnetizer or Behlman alternating field demagnetizer. All samples were measured on a Minispin spinner magnetometer. IRM studies were conducted using the IM-10 impulse magnetizer (ASC-Scientific).

3. Results

[9] A total of 18 of 22 sites yielded consistent directional data. Samples from two of the granitic sites behaved erratically during thermal and alternating field demagnetization and yielded no consistent directions. Site 15 yielded directions that appeared to be approximately reverse (tilt corrected) to the characteristic direction as given in Table 1 (Figures 3a–3c). Because this was the only site that showed a reverse magnetization in the SFM and the relatively large \( \alpha_{95} \) (21°), we did not include this site in our mean calculation. Nevertheless, it does suggest that the magnetization of these rocks spanned an interval that included at least one field reversal.

[10] Samples from the remaining 18 sites showed stable magnetic behavior during both thermal and alternating field demagnetization; however, alternating field demagnetization was unable to fully demagnetize most samples, and the majority were treated using stepwise thermal treatment. Typical demagnetization behaviors are shown in Figures 4 and 5. In general, removal of either a present-day field or randomly directed overprint was followed by nearly linear decay of the characteristic component of magnetization. In situ directions were largely confined to the west-southwest quadrant and downwardly directed with considerable scatter (Table 1). The mean in situ direction is \( D = 232.8^\circ, I = +48.5^\circ \) (Table 1; \( \kappa = 7.8 \alpha_{95} = 13.2^\circ \)) and compares favorably to the directional data observed in the previous studies of the SFM by Hsu et al. [1966] and Hays and Scharon [1966]. Tilt correction of the data results in a considerable improvement in grouping and is discussed in section 3.3 along with other tests that help constrain the age of magnetization in the SFM rocks.

3.1. Conglomerate Test

[11] A boulder conglomerate overlies the St. Francois rocks at several locations and contains clasts of both granitic and volcanic material. The age of the boulder conglomerate is considered as Middle to Late Cambrian on the basis of stratigraphic relationships with the overlying Late Cambrian-aged Lamotte sandstone [Kisvarsanyi et al., 1981]. We sampled 13 boulders from this conglomerate. Individual clasts exhibited low-temperature unblocking directions consistent with a viscous overprint of recent origin and
Figure 4. Orthogonal vector plot and equal-angle stereoplot for a sample from the Bell’s Mountain rhyolite at site 17. (a) Sample BM-5 treated using alternating field demagnetization. (b) In situ directional data for thermally treated sample BM-2. (c) Same sample shown in tilt-corrected coordinates.
well-defined, but randomly distributed high-temperature unblocking components (Figure 6a–6e). We regard these random high-temperature components as a positive conglomerate test that constrains the minimum age of magnetization for the SFM rocks to older than Late Cambrian.

3.2. Baked Contact Test

[12] Both the granitic and volcanic rocks of the St. Francois Mountains are intruded by mafic dikes of variable width. These dikes are believed to represent the last igneous pulse in the St. Francois region and have been dated to ~1330 Ma [Lowell and Young, 1999; Ramo et al., 1994; R. Tucker, personal communication, 2000]. We sampled a detailed profile through one of the dikes and sampled several of the smaller dikes and the contact rocks at another site. The results of the baked contact test are somewhat ambiguous (Figure 7a–7f). Figure 7f shows a stereoplot of mean results from dikes and baked contacts at site 4 where the Grassy Mountain ignimbrite (GMI) is intruded by a 1.2 m wide dike (Figure 8a). Results from the GMI distant from the dike show directions consistent with the prefolding magnetization discussed in section 3.3 (D = 259°, I = +44°; Figures 7a and 7f). Samples from the dike at site 4 yield a mean direction that is different from the volcanics/granites (D = 325°, I = +59°; Figures 7b and 7f). In contrast, the baked country rock near the dike at site 4 exhibited stable behavior during demagnetization, and the directions are consistent out to one-half dike width distance (D = 76°, I = +45°; Figures 7c and 7f). At face value, these results suggest a negative baked contact test for site 4.

[13] Dike samples from a small dike swarm at site 3 (intruding the Slabtown granite), show similar directions to the baked country rocks at site 4 (Figures 7e and 8b). Samples taken from the host granites at site 3 show a weak overprint consistent with both the baked contact directions at site 4 and directions from the intruding dike swarm (Figures 7d and 7f). The mean direction from the smaller dikes and baked contacts at both sites 3 and 4 is D = 60.8°, I = +48.6° (2 sites, 17 samples). Our preferred interpretation is that these NE down directions are related to the timing of dike emplacement (circa 1330 Ma) and provide a positive baked contact test for the 1330 Ma rocks and a positive inverse baked contact test for the country rock. An inverse baked contact test [McElhinny and McFadden, 2000]

![Figure 5.](image-url)
provides evidence that the host rocks have retained their magnetization at least since the time of baking.

[14] We have no definitive explanation for the NW down directions observed in the larger dike; perhaps it is related to the presence of larger magnetic domain sizes that tend to be more unstable. We also note that the large dike is heavily fractured and mineralized and may have been the site of channeled fluid flow along its margin during a younger time. Many of the dike samples do not reach a stable endpoint (Figure 7b), but a few trend (at higher temperatures) toward the E-NE direction observed in the country rocks. In contrast, the dikes at site 3 have sharp and fresh contacts with their host rocks.

[15] A positive inverse baked contact test for the host rocks constrains the age of magnetization to older than 1330 Ma. There are few similar-age poles from North America at 1330 Ga. A virtual geomagnetic pole (VGP) obtained from seven samples in Kansas drill core [Kodama, 1984] yielded a similar inclination to our study; however, the declination was nearly 180° different. The declination in the Kodama [1984] study was based on the premise that the overprint was acquired in the present-Earth’s field, and the older component is primary. Thomas and Piper [1992] reported paleomagnetic data from the supposed circa 1300 Ma Ericksfjörd lavas of Greenland that are clearly different from the directions we observe in our dikes; however, Paslick et al. [1993] suggested the age of the Ericksfjörd lavas was closer to 1.2 Ga, and therefore a direct comparison to our VGP is not reasonable. In addition, Thomas and Piper [1992] noted reversal asymmetry of nearly 30° in these rocks and a number of intermediate directions attributed to nondipole components of the Earth’s field. The closest reliable pole to our dike VGP is derived from the 1267 ± 2 Ma MacKenzie dike swarm [Buchan and Halls, 1990]. This pole falls some 40° from our dike VGP, but the difference can be accounted for by normal plate motion during the time interval between the two poles. We have just completed a second field season to the area to sample additional sites in the younger magmatic suites to better constrain this magnetization.

3.3. Fold Test

[16] The tilts observed in the SFM volcanic rocks varied throughout the study region (Figure 8c). Maximum dips were nearly vertical in portions of the Lake Killarney region (Figure 8c,
site 9) and several of the units were nearly flat lying. We are unable to uniquely determine the amount of tilting in the granitic bodies but point out that the tilting of the volcanic units was likely synchronous with caldera collapse, and the intruding granites would have suffered only minor postemplacement tilting in regions of reactivated faults. Lowell [1991] argues on the basis of data provided by Clendenin et al. [1989] that the entire Butler Hill caldera (Figure 2a) was tilted a maximum of 10° to the southwest by Late Proterozoic or Phanerozoic faulting along the Simms Mountain Fault (Figure 2a). Styles [1980] also concluded on the basis of field and petrologic data that the Butler Hill batholith was tilted to the SW between 9°–11°, although the exact timing of the tilting was not discussed. Lower Paleozoic strata overlying the Butler Hill batholith are nearly flat lying at, and between, our sites 1–6, and dikes intruding the granites at site 3 are near vertical, as is the dike at site 4. We therefore have applied a regional tilt correction to sites 1–6 where we see no obvious bedding with a strike of 150° and a dip of 10° in accordance with the estimates of Styles [1980] and Bickford et al. [1981]. We applied the tilt test using data from 18 of 22 sites, and the results are shown in Figure 9a–9c. The in situ grouping improves significantly during stepwise unfolding of the rocks and reaches a maximum at 100% unfolding (see Figure 9c). The tilt-corrected direction in the SFM rocks is D = 233.4°, 1 = +36.9° (k = 27.0; α95 = 6.8°). Comparison of k values (unfolded/folded) in the classic McElhinny [1964] fold test yields a k unfolded/folded of 3.46 (critical value is 1.78), indicating that the magnetization was acquired prior to folding of the rocks. The fold test was also applied using the McFadden [1990] test. The critical value of ξ (n = 18 sites) is 6.919 (99%). The McFadden [1990] fold test assigns a probability to the null hypothesis that a particular magnetization was acquired prior to, synchronous with, or postfolding of the rocks. The SFM study yielded an SCOS value of 7.894 (in situ). Our result signifies that a postfolding magnetization can be rejected above the 99% confidence level. The combined result of these fold tests indicates that the magnetization in the SFM rocks was acquired prior to the event(s) that formed the folding in the rocks. Since the folding of these rocks was broadly synchronous with the formation of the caldera collapse features documented by Kisvarsanyi et al. [1980], the positive fold test in the SFM provides very powerful evidence that these rocks carry a primary magnetization dating to their emplacement age.

3.4. Rock Magnetism and Magnetic Mineralogy

[17] The opaque mineralogy of the SFM rocks has been discussed in detail by a number of authors [see, e.g., Hsu et al., 1966; Styles, 1976; Blades and Bickford, 1976; Kisvarsanyi et al., 1981; Lowell, 1991]. The primary opaque iron oxide minerals in the volcanic rocks are hematite and magnetite. The granitic rocks are dominated by magnetite. The magnetic minerals are thought to be primary igneous minerals, although some formed during late-stage eruption hydrothermal alteration [Hsu et al., 1966; Styles, 1976; Blades and Bickford, 1976; Lowell, 1991]. For example, the Grassly Mountain ignimbrite (sites 4, 5, and 10) contains an abundance of aligned hematite specks that have imparted a fabric to the rock as a result of rheoignimbritic flow [Styles, 1976]. Hematite has also been noted in the middle zone of the Lake Killarney unit (sites 8 and 9) and in a number of ash flow tuffs near our sites 16 and 18 [Styles, 1976; Blades and Bickford, 1976; Kisvarsanyi et al., 1981]. Lowell [1991] describes magnetite as an accessory mineral in the Silvermine granite (site 7) and the Butler Hill/Breadtray granite (site 1). Additional opaque mineralogy descriptions are given by Kisvarsanyi et al. [1981] and Hsu et al. [1966].

[18] Thermal demagnetization decay curves indicate that the stable remanence in the SFM rocks is carried by either magnetite or hematite (Figures 3–6 and Figure 10a). Isothermal remanence acquisition curves (IRM, Figure 10b) also show clear indications of both hematite and magnetite. Thermal demagnetization of three-axis IRM [Lowrie, 1990] confirms the presence of both intermediate and high-coercivity fractions dominated by hematite (Figures 10d, 10e, and 10g) or magnetite (Figure 10f and 10h). A few samples exhibit a broad range of coercivity fractions carried by hematite (Figure 10c). Collectively, the petrographic examinations cited earlier in section 3.4 and the rock magnetic behavior described in this study suggest that the primary remanence in our samples is carried by both hematite and magnetite.

4. Discussion

[19] New paleomagnetic data from the St. Francois Mountains igneous province in Missouri provide a key paleopole for Laurentia at 1476 ± 16 Ma. We consider the paleomagnetic pole primary on the basis of the un metamorphosed nature of the rocks, a positive conglomerate test (magnetization age > 500 Ma), positive inverse baked contact test (magnetization age > 1330 Ma), and a positive fold test (magnetization age > deformation ∼ caldera collapse). We also note a reversed direction at one site in the Taum Sauk rhyolite; however, in the absence of confirmation from other sites in the SFM, we did not use this site in our mean calculation (Table 1).

Figure 7. (opposite) (a) Orthogonal vector plot for the Grassly Mountain ignimbrite located ～50 m from the contact with a 1.2 m wide mafic dike at site 4. This sample shows the characteristic St. Francois Mountains (SFM) direction following the removal of a weak viscous overprint. (b) Orthogonal vector diagram of a dike sample taken from the middle of the 1.2 m dike at site 4. The direction trends toward the E-NE direction observed in the baked host at the site. (c) Orthogonal vector diagram of the Grassly Mountain ignimbrite taken 10 cm from the contact with the dike at site 4. The high-temperature component is significantly different from both the characteristic SFM direction and the direction observed in the dike at site 4 (see text for discussion). (d) Orthogonal vector diagram for the Slabtown granite host at site 3 located 20 cm from a swarm of 1–5 cm wide mafic dikes (see Figure 8). The low-temperature component is similar to the samples taken immediately adjacent to the dikelets and within the dike rocks. The high-temperature component is identical to the characteristic direction observed in unbaked SFM rocks. (e) Orthogonal vector diagram from a sample of dike material at site 3 showing a typical E-NE intermediate down component. (f) Equal-angle stereoplot of directional data relevant to the baked contact test (see text for details).
4.1. Comparison to Laurentian Paleomagnetic Poles

[20] Discrepancies in paleomagnetic data from Laurentia for the time period from 1400–1475 Ma were recently highlighted by Harlan and Geissman [1998] in their discussion of the Electra Lake gabbro (1433 Ma) pole. Harlan and Geissman [1998] noted that selected paleomagnetic data from igneous intrusions from eastern North America (including the previous SFM results; Figure 11a) yielded a mean direction of $0.5^\circ/C_{176}$, $214.7^\circ/C_{176}E$ ($\alpha_{95} = 6.2^\circ$) and was statistically different than the results from the Electra Lake gabbro ($21.1^\circ/S$, 221.1$^\circ/E$, $\alpha_{95} = 3.4^\circ$) and a mean pole from the Belt Supergroup (18.8$^\circ/S$, 207.2$^\circ/E$, $\alpha_{95} = 5.6^\circ$). New paleomagnetic data from the Belt Supergroup (R. Enkin, personal communication, 2001) do not statistically change the older results of Elston and Bressler [1980], although there is an indication of a slight eastward migration of the poles from the uppermost Belt rocks. The age of the Belt Supergroup is better constrained by new U-Pb ages on the interlayered volcanics and intrusions [Evans et al., 2000; Anderson and Davis, 1995]. Ages of the lower Belt sediments are constrained by U-Pb data from the Moyie sills. Anderson and Davis [1995] concluded that these sills were emplaced shortly after deposition of Aldridge/Pritchard Formations at 1468 ± 2 Ma. Age constraints for the Upper Belt rocks are provided by the Logan Pass bentonite (in the Helena Formation) with a U-Pb zircon age of 1454 ± 9 Ma, a rhyolite in the uppermost Purcell Lavas with a U-Pb age of 1443 ± 7 Ma and a U-Pb age of 1401 ± 6 Ma for a thin tuff between the Bonner quartzite and Libby Formation in the Upper Belt rocks [Evans et al., 2000].

[21] Paleomagnetic directions from the Lower Belt Supergroup rocks are enigmatic [Vitorello and Van der Voo, 1977; Elston and Bressler, 1980] (Table 2 and Figure 11a). Although both studies yield approximately the same declination, the inclinations of Vitorello and Van der Voo [1977] were some 20$^\circ$ steeper than the results of Elston and Bressler [1980]. Elston and Bressler [1980] commented that the difference in inclinations had no readily apparent explanation but noted that they appeared to be restricted to the northeastern part of the Belt Basin. Upon close inspection of their results, we note that Elston and Bressler [1980] calculated a mean direction on the basis of directional data at 550$^\circ$, whereas Vitorello and Van der Voo [1977] calculated their mean direction on the basis of high-temperature components (> 630$^\circ$) using orthogonal vector plots. The paleolatitude obtained by Vitorello and Van der Voo [1977] of $22.8^\circ\pm3.9^\circ$ is identical (within error) to that predicted by our new SFM results (Figure 11b).

[22] U-Pb crystallization ages of the Michikamau and Harp Lake intrusions are 1460 ± 5 Ma and 1450 ± 5 Ma, respectively. The predicted paleolatitude for the Belt Supergroup based on a combined Michikamau/Harp Lake (HL) pole is also systematically higher than the observed paleolatitude in the upper Belt rocks (~33$^\circ$ to 20$^\circ$). There is also a difference between the observed Electra Lake gabbro (ELG) paleolatitude and that predicted by the Michikamau/HL (24$^\circ$ versus 37$^\circ$); however, the Laramie anorthosite/Sherman granite pole [Harlan et al., 1994] is similar to the Michikamau/Harp Lake pole (Figure 11a).

[23] Collectively, the data from the SFM and Michikamau/HL rocks suggest that the paleolatitudes observed in the Belt rocks...
are lower than would be expected for their present-day location at the edge of the North American craton (see Figure 11c). There are a number of possible explanations for these observed differences. The first is that the SFM/HL/Michikamau poles do not accurately reflect the position of Laurentia because of possible tilting of the intrusions. Alternatively, we could argue for possible inclination shallowing in the Belt Supergroup rocks, but we have no way to accurately assess this possibility, and it remains an ad hoc explanation for the observed differences. R. Enkin (personal communication, 2001) has compiled new data from the Upper belt rocks that show a small eastward motion in the Belt apparent polar wander path (APWP) such that the cited differences in this paper may be somewhat less for the younger segment of the path (Figure 11c).

Harlan and Geissman [1998] propose a number of rotation and tilting possibilities to explain the differences in paleomagnetic poles from the central craton poles and those from the marginal ELG and Belt rocks. It is possible that undetected rotations of any, or all, of these poles can account for the observed differences, or that the timing of magnetization assigned to the intrusive bodies based on their U-Pb ages overestimates the true age of magnetization. More data from similar-age units may help distinguish among the possible explanations.

4.2. Comparison With Other Continents

[24] Accepting the SFM pole as representative for Laurentia at 1476 ± 16 Ma allows us to test proposed continental configurations for the Mesoproterozoic. Although the Rodinia hypothesis is considered valid from ~1100 Ma to ~750 Ma, the links between Australia-Antarctica, Siberia, and Baltica are considered valid for Mesoproterozoic and earlier time [Gower et al., 1990; Ross et al., 1992; Dalziel, 1997; Sears and Price, 2000]. The identities of the continents adjacent to the present-day western coast of Laurentia are controversial. For example, Ross et al. [1992] have argued for an Australian source for the Belt Supergroup on the basis of detrital zircon data, while Sears and Price [2000] argue that Siberia was the source continent for these zircons. Gower et al. [1990] have argued for a Mesoproterozoic linkage between Baltica and Laurentia (Nena). Ideally, paleo-
magnetic data can test these proposed configurations provided key poles are available for similar time periods from these continents. Paleomagnetic tests of continental coherence are more robust when closely spaced poles are available to define apparent polar wander paths (APWPs). Table 2 lists selected paleomagnetic poles available from Laurentia, Siberia, Australia, and Baltica for the interval from 1500 Ma to 1430 Ma. The data are limited in their power to distinguish among the various models as no APWPs can be constructed for this interval of time. Therefore the following analysis should be viewed with caution as we have assumed polarity choices for the models and attempted a “closest fit” approach to the reconstructions.

[25] Paleomagnetic data from Australia are derived from the mafic intrusions and host rocks of the Mt. Isa inlier [Tanaka and Idnurm, 1994] and the Gawler Range Volcanics [Chamalaun and Dempsey, 1978]. The Mt. Isa pole is assigned an age of 1525 ± 25 Ma [Tanaka and Idnurm, 1994]. There is some controversy about the age of magnetization recorded by the Gawler Range Volcanics (GRV). The emplacement age of the GRV is 1592 ± 2 Ma [Fanning et al., 1988]; however, recent work [Daly et al., 1998] suggests that the region underwent deformation and metamorphism during the 1540–1565 Ma interval. Tanaka and Idnurm [1994] and Idnurm [2000] noted the similarity between the GRV pole and the Mt. Isa pole and suggested that a purported postfolding magnetization in the GRV might be as young as 1525 Ma.

[26] New paleomagnetic data from the Anabar shield region of Siberia [Ernst et al., 2000] are derived from the 1503 ± 5 Ma Kuonamka dikes. Five of these dikes yield a paleomagnetic pole at 6°N, 234°E (dp = 14°, dm = 28°) that the authors considered primary on the basis of the extremely low metamorphic grade of the rocks and the fact that directional data from the host rocks are significantly different from the dikes.

[27] There are a number of paleomagnetic studies from the Baltic shield with ages between 1455–1530 Ma [Piper, 1980; Bylund, 1985]. These poles yield a grand mean at 27.9°S, 3.8°E (σ95 = 18.8°). Our best estimate for the age of this mean pole is 1520 ± 7 Ma.

[28] Figure 11d shows these paleomagnetic poles rotated to the Rodinia configuration of Dalziel [1997]. Both the Gawler Range Volcanics (GRV) and the Kuonamka dike pole (KD) fall close to coeval poles from Laurentia; however, the Baltica mean pole (BMP, including the individual poles used to derive BMP) and Mt. Isa pole (MI) fall well away from the Laurentian poles. Testing possible cratonic coherence or specific reconstructions with individual poles can be misleading since slight rotations or large errors can lead to misinterpretations about the validity of a particular reconstruction. A better, though nonunique, test is to rotate the continents to their correct paleolatitudes by assuming a polarity that will minimize the distance between cratons. Once the continents are placed at this paleolatitude and orientation, they can be moved longitudinally to a closest approach. Figure 11e shows the continents rotated to one possible closest approach fit using the SFM pole for Laurentia, for the KD pole, the Baltica mean pole, and the Gawler Range volcanic poles (using a South Pole option for the poles listed in Table 2). Figure 11f uses the same polarity choice for the Laurentian and Baltica poles, but uses the Mt. Isa pole from Australia (South Pole) and a North Pole option for the Siberian pole. The fit in Figure 11e approximately the Rodinia configuration of Dalziel [1997] for Siberia and Australia, whereas Figure 11f is close to the AUSWUS configuration of Karlstrom et al. [2000]. In fact, one cannot reject either the SWEAT or AUSWUS configurations on the basis of the extant paleomagnetic database for this time period. Nevertheless, it should be reemphasized that the age range of these poles may span over 50 Ma. Sears and Price [2000] have argued that Siberia was adjacent to the western margin of Laurentia rather than Australia-Antarctica on the basis of geological correlations between the two continents. At first glance, Figure 11e would appear to negate this possibility, but the paleomagnetic pole for Siberia has a large error (28°). It is therefore possible to position Siberia farther south in Figure 11e against the western margin of Laurentia. A choice of opposite polarity for the Siberia pole also brings Siberia closer to the western margin of Laurentia; however, the orientation precludes matching of geologic features critical to the arguments of Sears and Price [2000].

5. Conclusions

[29] The SFM igneous province in southeastern Missouri has provided a key 1476 ± 16 Ma paleomagnetic pole for Laurentia. The pole is considered primary on the basis of a positive conglomerate test (pole age > 500 Ma), a positive inverse baked contact test (pole age >1330 Ma) and a positive fold test (pole age > deformation ≈ caldera collapse ≈ 1476 Ma). An analysis of similar-aged poles from Laurentia highlights key differences
between the Belt Supergroup poles, the Electra Lake gabbro pole, and poles from cratonic North America. We note that the SFM pole is consistent with the paleolatitude suggested by the Lower belt rocks obtained by Vitorello and Van der Voo [1977]. An, as of yet, unexplained paleolatitudinal offset exists between data obtained from the Upper belt rocks and igneous intrusions from cratonic Laurentia.

[30] The debate regarding the Mesoproterozoic position of Siberia cannot be settled using the existing paleomagnetic data. Sears and Price [2000] have argued for a western Laurentia connection and described possible source rocks for the Belt Supergroup in Siberia. A strict interpretation of the new paleomagnetic data from Siberia cannot be settled using the existing paleomagnetic data. It is perhaps an understatement to suggest that additional high-quality paleomagnetic data are needed to provide a robust test of Mesoproterozoic paleogeographies.

[31] Acknowledgments. The authors wish to thank Karl Evans for a preprint of his paper on the age of the Belt Supergroup, R. Enkin for a discussion on new paleomagnetic data from the Belt Supergroup and Trond Torsvik, and Rob Van der Voo and Elizabeth Eide for comments on an early draft of the manuscript. Steve Harlan and Dave Evans are thanked for valuable suggestions that improved the manuscript. J.G.M. was supported by NSF grant EAR98-05306 and a Fulbright grant from the United States-Norway Fulbright Commission. Fieldwork support for W.S. and J.G.M. was provided by a grant from the Indiana State University Research Committee.

Figure 11. (opposite) (a) Paleomagnetic poles from Laurentia with well-constrained ages. Pole symbols are given in Table 2 along with their ages. (b) Paleolatitudinal construction for Laurentia at 1476 ± 16 Ma based on our SFM pole along with site locations of other paleomagnetic studies keyed to Table 2. Dashed shaded line shows the observed paleolatitude for the Lower Belt Supergroup on the basis of the results of Elston and Bressler [1980], and the thick shaded line shows the paleolatitude for this same formation based on the study by Vitorello and Van der Voo [1977]. (c) Paleolatitudinal construction for Laurentia at ~1450 Ma based on the combined Michikamau/Harp Lake poles along with site locations of other paleomagnetic studies keyed to Table 2. Dashed shaded line shows the observed paleolatitude for the Upper Belt rocks on the basis of the results of Elston and Bressler [1980]. (d) Paleomagnetic poles from Table 2 rotated into Laurentian coordinates based on the euler parameters of Dalziel [1997]. Dark shading, Baltica poles (note that the Dundret pole is not plotted on Figure 9d); light shading, Laurentian poles; stippled shading, Australian poles; no shading, Siberian pole. (e) One possible paleoconstruction showing the closest approach of the continental landmasses with proposed Laurentian links based on the data listed in Table 2. (f) An alternative reconstruction using the Mt. Isa pole for Australia and inverting the Kuonamka dike (KD) pole for Siberia.

Table 2. Selected Paleomagnetic Poles

<table>
<thead>
<tr>
<th>Pole Abbreviation</th>
<th>Pole</th>
<th>$\alpha_{95}$</th>
<th>Pole Latitude</th>
<th>Pole Longitude</th>
<th>Age ± Error (Ma)</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>MI</td>
<td>Mt. Isa inlier</td>
<td>8.4°</td>
<td>79.0°S</td>
<td>110.6°E</td>
<td>~1500</td>
<td>Tanaka and Idnurm [1994]</td>
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<td>GRV</td>
<td>Gawler Range Volcanics</td>
<td>5.2°</td>
<td>60.4°S</td>
<td>080.0°E</td>
<td>&lt;1540</td>
<td>Chamalaun and Dempsey [1978]</td>
</tr>
<tr>
<td>DUN</td>
<td>Dundret basic rocks</td>
<td>3.0°</td>
<td>22.0°N</td>
<td>203.0°E</td>
<td>1530 ± 35</td>
<td>Piper [1980]</td>
</tr>
<tr>
<td>HAL</td>
<td>Halleforsna dyke</td>
<td>9.3°</td>
<td>27.0°N</td>
<td>167.0°E</td>
<td>1518 ± 38</td>
<td>Piper [1980]</td>
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<td>BUN</td>
<td>Bunkris dolerite</td>
<td>3.5°</td>
<td>30.2°N</td>
<td>175.4°E</td>
<td>1516 ± 62</td>
<td>Bulyand [1985]</td>
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<tr>
<td>GLY</td>
<td>Glyson dolerite</td>
<td>8.2°</td>
<td>35.4°N</td>
<td>171.4°E</td>
<td>1516 ± 62</td>
<td>Bulyand [1985]</td>
</tr>
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<td>BMP</td>
<td>Baltica mean pole</td>
<td>18.8°</td>
<td>27.9°S</td>
<td>3.8°E</td>
<td>~1520</td>
<td>this study</td>
</tr>
<tr>
<td>SFM</td>
<td>St. Francois Mountains</td>
<td>6.8°</td>
<td>13.2°S</td>
<td>219.0°E</td>
<td>1476 ± 16</td>
<td>this study</td>
</tr>
<tr>
<td>H</td>
<td>St. Francois Mountains</td>
<td>5.0°</td>
<td>0.9°S</td>
<td>219.0°E</td>
<td>1476 ± 16</td>
<td>Hsu et al. [1966]</td>
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<td>HS</td>
<td>St. Francois Mountains</td>
<td>10.0°</td>
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<td>210.0°E</td>
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<td>Hays and Scharon [1966]</td>
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<td>SPF</td>
<td>Spokane Formation-Belt Supergroup</td>
<td>5.1°</td>
<td>15.5°S</td>
<td>225.5°E</td>
<td>≈1460</td>
<td>Vitorello and Van der Voo [1977]</td>
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<td>MK</td>
<td>Michikamau intrusion</td>
<td>6.5°</td>
<td>1.5°S</td>
<td>218.0°E</td>
<td>1460 ± 5</td>
<td>Emelie et al. [1976]</td>
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<td>HL</td>
<td>Harp Lake intrusive</td>
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<td>206.3°E</td>
<td>1450 ± 5</td>
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<td>Belt Supergroup mean</td>
<td>5.6°</td>
<td>18.9°S</td>
<td>207.2°E</td>
<td>1400–1470</td>
<td>Elston and Bressler [1980]</td>
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<td>Electra Lake gabbro</td>
<td>3.4°</td>
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<td>221.3°E</td>
<td>1433 ± 2</td>
<td>Harlan and Grissman [1998]</td>
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<td>LA</td>
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<td>3.5°</td>
<td>6.7°S</td>
<td>215.0°E</td>
<td>1429 ± 9</td>
<td>Harlan et al. [1994]</td>
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<td>KD</td>
<td>Kuonamka dikes</td>
<td>28.0°</td>
<td>6.0°N</td>
<td>234.0°E</td>
<td>1503 ± 5 Ma</td>
<td>Ernst et al. [2000]</td>
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*Pole DUN is not shown in Figure 11d, but it is used to calculate the mean pole.

*Baltica mean pole is given as a south pole.
References


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