Paleomagnetism and U–Pb geochronology of Franklin dykes in High Arctic Canada and Greenland: a revised age and paleomagnetic pole constraining block rotations in the Nares Strait region[1]

Steven W. Denyszyn, Henry C. Halls, Don W. Davis, and David A.D. Evans

Abstract: U–Pb baddeleyite ages and paleomagnetic poles obtained for dykes on Devon Island and Ellesmere Island in the Canadian Arctic and the Thule region of Greenland show that they are associated with the Franklin magmatic event. This study is the only one devoted to Franklin igneous rocks where a primary paleomagnetic remanence and U–Pb age have been obtained from the same rocks. Ages from this study range from 721 to 712 Ma, but paleomagnetic directional data show no clear age progression. The paleomagnetic poles from each of the two regional subsets are significantly different at the 95% confidence level from paleomagnetic results previously published for the Franklin event in the Canadian Shield. The difference in the pole locations can be accounted for, to first approximation, by a simple model of early Cenozoic block rotations among the North American plate, Greenland, and a hypothesized ancient microplate comprising Ellesmere, Devon, Cornwallis, and perhaps Somerset islands. A new grand-mean paleopole for the Franklin event, including restoration of Greenland and the proposed "Ellesmere microplate" to North America, is located at (8.4°N, 163.8°E, A_{95} = 2.8°, N = 78 sites) and is a key pole for Neoproterozoic supercontinent reconstructions.

Résumé : Des âges U–Pb, déterminés sur de la baddeleyite, et des pôles paléomagnétiques obtenus de dykes des îles de Devon et d'Ellesmere, dans l'Arctique canadien, et de la région de Thulé au Groenland, montrent que ces dykes sont associés à l'événement magmatique de Franklin. La présente étude est la seule consacrée aux roches ignées de Franklin où une rémanence paléomagnétique primaire et des âges U–Pb ont été obtenus des mêmes roches. Les âges dans cette étude varient de 721 à 712 Ma, mais les données sur les directions paléomagnétiques ne montrent aucune progression claire de l'âge. Les pôles paléomagnétiques de chacun des deux sous-ensembles régionaux diffèrent de manière significative, au niveau de confiance 95 %, des résultats paléomagnétiques publiés antérieurement pour l'événement de Franklin dans le Bouclier canadien. La différence d'emplacement des pôles peut être expliquée, en une première approximation, par un modèle simple de rotations de blocs, au Cénozoïque précoce, entre la plaque Nord-américaine, le Groenland et une ancienne plaque hypothétique qui comprenait les îles d'Ellesmere, de Devon, de Cornwallis et peut-être de Somerset. Un nouveau paléopôle de grande moyenne pour l'événement Franklin, incluant la restoration du Groenland et de la « microplaque d'Ellesmere » proposée à l'Amérique du Nord, est situé à (8,4°N, 163,8°E, A_{95} = 2,8°, N = 78 sites) et constitue le pôle clé pour les reconstructions du super continent au Néoprotérozoïque.

[Traduit par la Rédaction]

Introduction

Precisely dated rocks with well-defined paleomagnetic poles are scarce in the Proterozoic record, yet are critical for reconstructing the past positions of continents and correlating what are today the dispersed fragments of supercontinents (e.g., Buchan and Halls 1990). The Neoproterozoic of North America is currently represented by only a few widely

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S.W. Denyszyn.[2] University of Toronto, Department of Geology, 22 Russell St., Toronto, ON M5S.3B1, Canada; Present address: Berkeley Geochronology Center, 2455 Ridge Road, Berkeley, CA 94709, U.S.A.
H.C. Halls. University of Toronto, Department of Geology, 22 Russell St., Toronto, ON M5S.3B1, Canada.
D.W. Davis. Jack Satterly Geochronology Lab, University of Toronto, Department of Geology, 22 Russell St., Toronto, ON M5S 3B1, Canada.
D.A.D. Evans. Department of Geology and Geophysics, Yale University, New Haven, CT 06520-8109, U.S.A.

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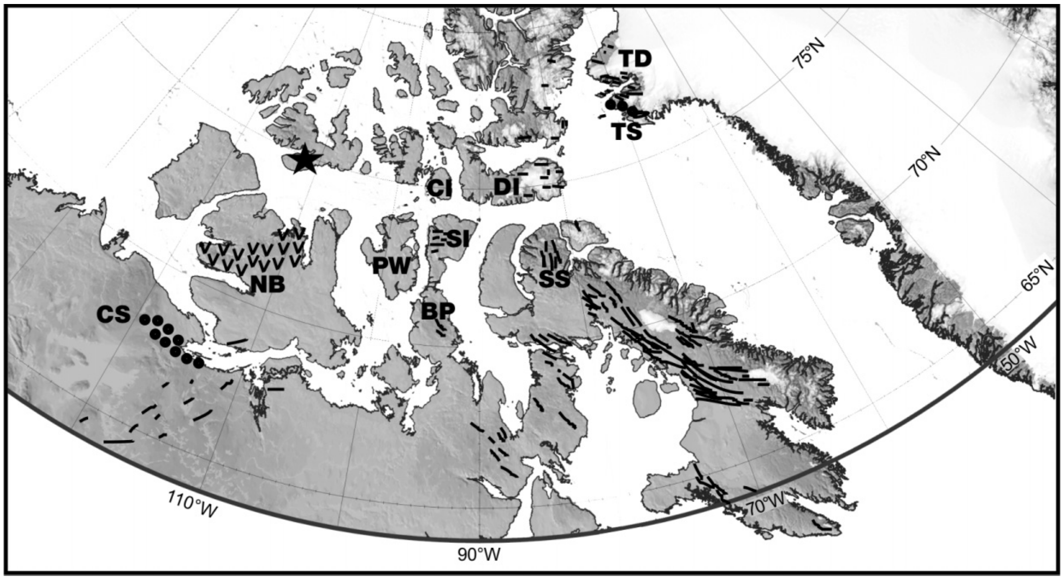
[2]Corresponding author (e-mail: sdenyszyn@bgc.org).

Fig. 1. Distribution of intrusions or lava flows associated with the Franklin magmatic event (modified after Fahrig (1987), with dyke locations also from Frisch (1984*a*, 1984*b*, 1984*c*) and Dawes (1991)). CS, Coronation sills; DI, Devon Island dykes; NB, Natkusiak basalts; SS, ,Strathcona Sound dykes; TD, Thule dykes; TS, Thule sills; placement of these labels indicates the locations of paleomagnetic studies referred to in text. Star denotes proposed plume source location based on radiating geometry of the dykes. Solid circles denote sills; v symbols denote volcanics.



spaced paleopoles with both high-precision U–Pb ages and well-characterized primary magnetizations (Buchan et al. 2000). The Franklin magmatic event of Arctic Canada (Fig. 1) is one such case. The currently accepted age of 723^{+4}_{-2} Ma (Heaman et al. 1992) and grand-mean paleomagnetic pole at (8.0°N, 163°E, A_{95} = 4.0°; Buchan et al. 2000) have been used extensively in Neoproterozoic plate reconstructions (Buchan et al. 2001; Li et al. 2008; Evans in press). The Franklin pole is particularly important as the Franklin magmatic event is related to a mantle plume north of Victoria Island (Fig. 1) that has been tied to the incipient breakup of the Neoproterozoic supercontinent Rodinia (Heaman et al. 1992; Shellnutt et al. 2004). Sturtian global glaciation is proposed to have occurred during this time period, and accurate knowledge of plate reconstruction from paleomagnetism may prove important for testing the "Snowball Earth" hypothesis (Hoffman et al. 1998). This study of the Franklin dyke swarm presents new paleomagnetic and U–Pb geochronological results, as well as a reinterpretation of published data from the Franklin magmatic event. This forms the basis for a new key paleomagnetic pole for Rodinia reconstructions, for a first-order tectonic reconstruction of the pre-Cenozoic Laurentian Arctic margin and for the demonstration that an extinct plate boundary exists between Ellesmere Island and Greenland.

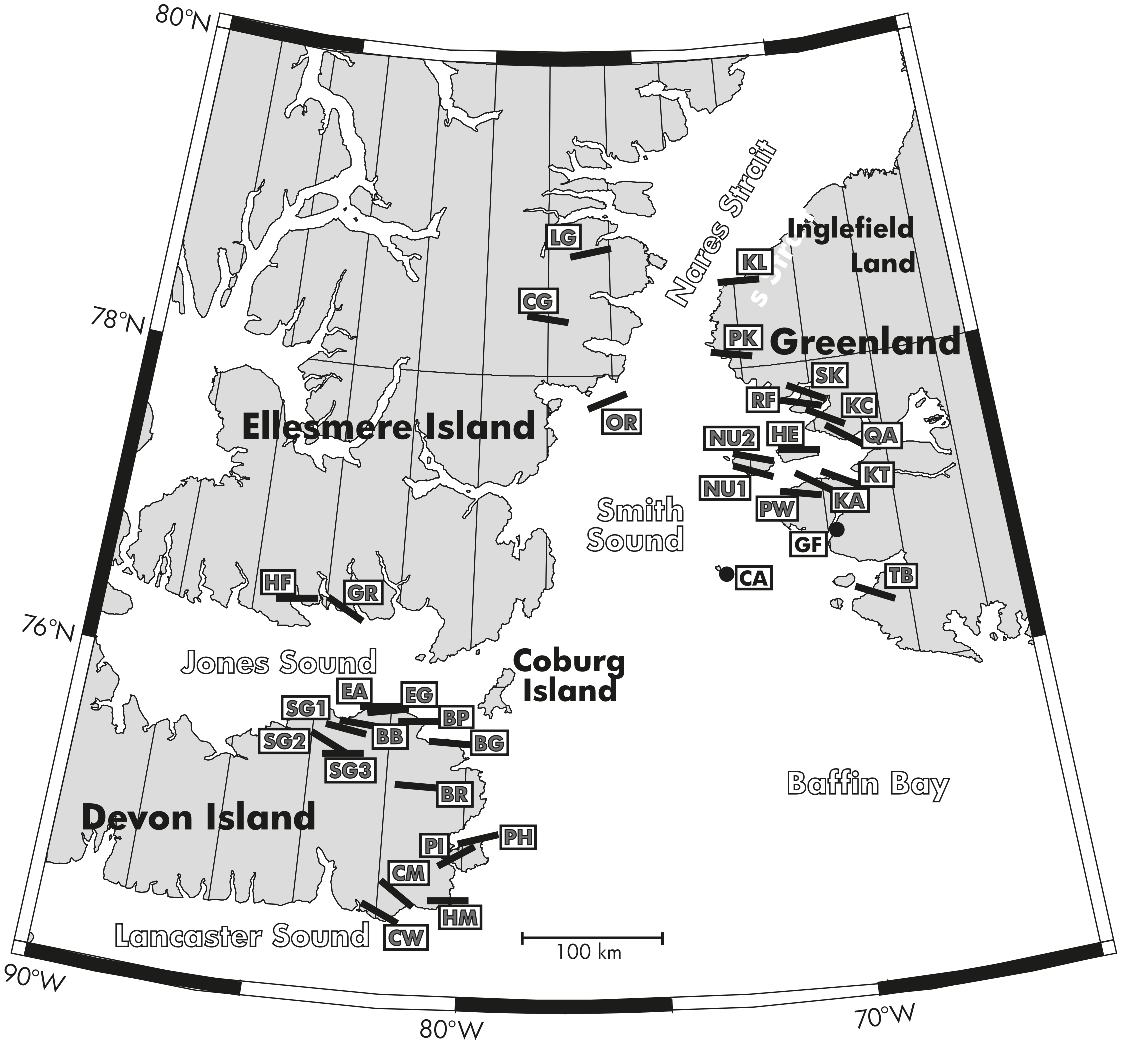
Geological setting

The Franklin igneous event emplaced sills and dykes across more than 2500 km of lateral distance in the Canadian Arctic and Greenland (Fig. 1), making it one of the largest dyke swarms on Earth. It includes large (≥20 m wide) diabase dykes in mainland Canada, Baffin Island, and Greenland. Also associated are the Victoria Island sills and Natkusiak volcanics of Victoria Island, the Coronation sills of the mainland Northwest Territories, and the Thule sills of Greenland (Fahrig et al. 1971; Frisch 1984*a*, 1984*b*, 1984*c*; Dawes 1991; Buchan and Ernst 2004; Shellnutt et al. 2004). In our study area (Fig. 2), the basement rocks that host the intrusions comprise high-grade granitic gneiss of Archean to Paleoproterozoic age, overlain in western and central Devon Island and western Ellesmere Island by flat-lying Paleozoic carbonate and mudstone that overlie the dykes, and in Greenland by flat-lying Mesoproterozoic sandstone of the Thule basin, which is cut by the dykes (Frisch 1988; Dawes 2004). The dykes are diabase with abundant clinopyroxene and lath-shaped plagioclase, minor biotite, hornblende, and oxide phases, as well as rare titanite and apatite. Unlike the Franklin dykes of similar trend on Baffin Island (Jefferson et al. 1994), olivine is absent. The dykes are, on average, 30 m wide with well-preserved chilled margins and appear unaltered, except for moderate sericitization of plagioclase and the occasional presence of chlorite. As they are all near vertical, are overlain by flat-lying strata in Canada, and cut flat-lying strata in Greenland, tectonic activity that might have remagnetized the rocks or locally rotated magnetization directions is likely minimal.

Methodology

Nineteen east–west-trending dykes in the High Arctic of Canada and 10 from northwest Greenland, as well as two sills, were sampled with 7–15 samples taken from each for paleomagnetic analysis. Blocks for geochronological analysis were taken from the interior of most dykes because coarser grained interiors are considered more favourable for the recovery of baddeleyite grains large enough for single-grain U–Pb analysis. The North America and Greenland data sets are treated separately because their remanence directions may have been offset by the Cretaceous movement of Greenland relative to Canada (Tessensohn and Piepjohn 1998),

Fig. 2. Map showing the locations of dykes (solid black bars) included in the study and their trends. Circles denote sills. Sample site names in boxes.



although they are certainly of the same swarm (Denyszyn et al. (2006) includes preliminary results for some of the dykes studied here).

Paleomagnetism

Samples were collected either as field-drilled oriented cores or as oriented block samples. Wherever possible, both magnetic and sun compasses were used for sample orientation, although cloud, fog, or high topographic relief made sun sighting impossible at some sites (CM, EA, HM, PI, SG2, KA). All sites were located by a hand-held Garmin Etrex global positioning system (GPS) unit or by onboard helicopter GPS. Block samples were subsequently drilled in the laboratory, and all drill cores were sliced into cylindrical specimens 2.45 cm in both length and diameter.

At least one specimen from each sample was subjected to detailed alternating field (AF) demagnetization, using a Schonstedt GSD-1 single-axis demagnetizer, to remove stray or present Earth field components. Field increments averaged 5 mT up to maximum fields of 95 mT. After each demagnetization step, the direction and intensity of remanent magnetization were measured using a modified Digico spinner magnetometer with repeatability of results for magnetization intensities above $\sim 10^{-3}$ Am^{-1}. Measurement procedures included an averaging procedure to reduce the dependence of results on the last sample axis to be demagnetized in the single-axis demagnetizer (description in Halls (1986)). Thermal demagnetization was also used for selected specimens using a Schonstedt TSD-1 thermal demagnetizer. Stepwise heating was conducted according to a measurement routine that involved taking measurements at temperature increments that decreased as the unblocking temperature of magnetite (580 °C) was approached. The paleomagnetic data were plotted on stereonets and vector diagrams (Zijderveld 1967) and then analyzed using principal component analysis (Kirschvink 1980). This included a search routine to find all linear segments on vector diagrams consisting of three or more points and passing a minimum acceptance given by a goodness-of-fit parameter; the maximum angle of deviation (Kirschvink 1980) was set at ≤10°.

Susceptibility versus temperature data were obtained from one or two specimens from every dyke using a Sapphire Instruments susceptibility meter to determine the magnetic mineralogy of the samples. During measurement, the specimens were heated from 20 to 700 °C and then cooled, with susceptibility measurements automatically recorded at 5 °C intervals during both heating and cooling stages.

U–Pb geochronology

At every stage of handling of the rock, heavy mineral concentrates, and baddeleyite grains, meticulous cleaning practices were followed to minimize any chance of sample contamination. Initially, baddeleyite grains were separated from the rock using "traditional" mineral separation methods (i.e., those used to separate zircon); these included density separation on a Wilfley table, heavy liquid separations, and dry magnetic separations in a Frantz isodynamic separator. The densest and least magnetic fraction was hand picked under a microscope to locate and isolate baddeleyite grains. This procedure, however, was found to be inefficient for extracting baddeleyite. The number of mineral separation steps was subsequently reduced to limit the opportunity for baddeleyite loss.

The powdered rocks were treated on the Wilfley water-shaking table using a technique modified after Söderlund and Johansson (2002) that involved adding only a spoonful of the powdered sample to the Wilfley table at a time as a slurry. Water and a small amount of liquid soap were mixed in to act as a surfactant to prevent the flat, tiny baddeleyite crystals from riding the surface tension of the water off the table. This was followed by concentrating the fine heavy minerals on the table, separating the magnetic material using a hand-held magnet, and handpicking baddeleyite from the residue. Only a few baddeleyite blades and fragments, most typically under 80 μm in length and weighing < 1 μg, were recovered from each sample, although every sample that underwent this procedure yielded a sufficient amount of baddeleyite for analysis. The recovered grains tended to be smaller than those from the "traditional" method, which were commonly over 100 μm in length. This leads to the conclusion that baddeleyite is in fact ubiquitous in these rocks despite its apparent absence using conventional mineral separation methods, indicating that the opportunity for loss while processing the sample using conventional methods is great, with only relatively large baddeleyite grains (if present) being recovered in the final concentrate.

All analyzed grains were photographed, but volume and therefore weight were difficult to determine because the grains were generally quite flat and could only be conveniently photographed perpendicular to the C axis with the optical camera. Model weights (w_{mod} in Table 2) were calculated, therefore, based on the assumption that the U concentrations were 300 ppm, an approximate average for many published baddeleyite analyses. Two exceptions were analyses 5 and 6 from CG-7, marked with an asterisk in Table 2. These were imaged along all three dimensions using a scanning electron microscope, which allowed for the calculation of U concentrations of 395 and 371 ppm, respectively.

Dissolution, isotope dilution, and sample loading methods were as described in Krogh (1973) using a ^{205}Pb–^{235}U spike and miniature bombs. No chemical separation procedures were required. The baddeleyite samples were then analyzed by a VG354 thermal ionization mass spectrometer using a Daly collector in pulse-counting mode. The mass discrimination correction for this detector was constant at 0.07%/amu (atomic mass units). Thermal mass discrimination corrections are 0.10%/amu. Dead time of the measuring system was about 20 ns and was monitored using the SRM982 standard.

Results

Paleomagnetism

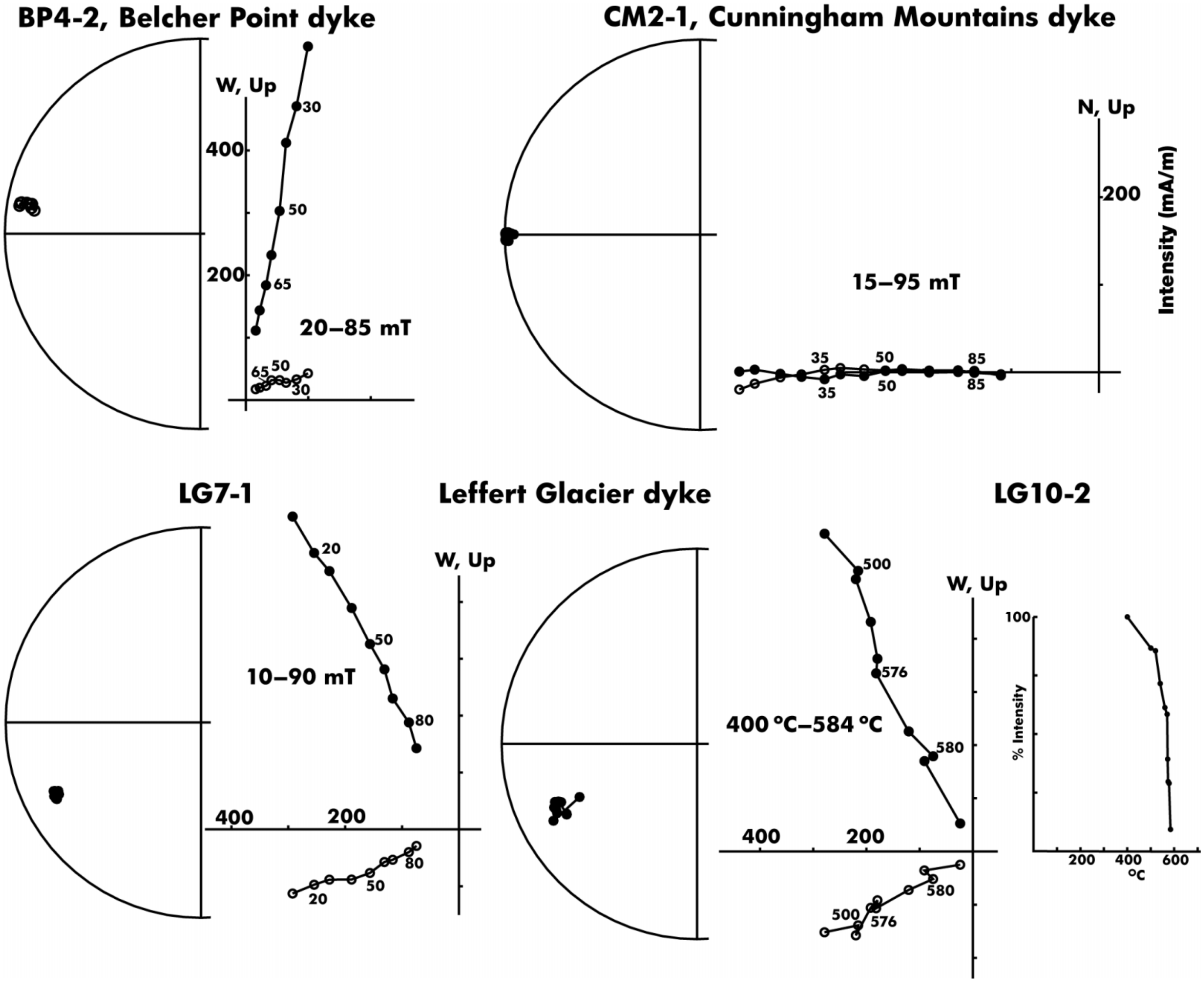
No useable results were obtained from seven of the 34 sampled sites, generally because magnetizations were too weak to obtain a meaningful remanence direction. At one site (OR), an inability to sight the sun combined with a magnetic storm made a reliable sample orientation impossible. From the remaining sites, paleomagnetic results (Fig. 3) indicate a stable high coercivity (H_c) and high unblocking temperature (T_{ub}) component of magnetization in the dykes that is revealed after removal of lower H_c and T_{ub} components (typically a low-coercivity present Earth field component) by AF and thermal demagnetization. The directions of this component for each site are given in Table 1 along with their corresponding virtual geomagnetic poles (VGPs) and show that most sites have a shallow inclination with westerly declination. One dyke from Canada (SG2) and two from Greenland (HE and NU1) have approximately antipodal (shallow east) declinations. The presence of opposing polarities indicates that the period of dyke injection spanned at least one reversal of the Earth's magnetic field. Westerly and easterly directed magnetizations are arbitrarily referred to as N and R polarities, respectively. Only samples with $N \geq 4$ and $\alpha_{95} \leq 15°$ are included in the calculations of the mean paleopoles.

A baked-contact test (Everitt and Clegg 1962) was performed on dyke and host rock samples from site GR (Denyszyn et al. 2006), where the host rock included anorthosite, part of a unit dated to 1972^{+41}_{-36} Ma (U–Pb, zircon) (Frisch 1988). The test was positive (Fig. 4) in terms of the change in paleomagnetic direction; the direction characteristic of the dyke was found in the baked host rock whereas that of the unbaked host rock was not seen in samples from the dyke. No hybrid magnetization in the thermal crossover regime was observed, but such "complete" baked-contact tests are rare in the literature (Buchan and Schwarz 1987). The VGP of the country rock anorthosite (P_{lat} = 01.6°N, P_{lon} = 272.1°E, dp = 7.1°, dm = 11.0°, N = 5) is similar to that expected from Laurentia at ca. 1800 Ma (Irving et al. 2004). The results of this test, combined with the presence of dual magnetic polarity, suggest that the characteristic magnetization components are primary.

Temperature versus susceptibility plots (Fig. 5) show a sharp drop at about 580 °C for all samples from east–west-trending dykes, indicating the presence of near-stoichiometric magnetite. Characteristic remanence unblocking spectra are also narrowly defined near 580 °C (Figs. 3, 4), demonstrating this magnetite as the principal carrier of the characteristic magnetization in all dykes.

The aggregate sample data set shows a moderate range of scatter in both declination and inclination (Fig. 6). Previous studies of Franklin-aged igneous rocks identified very large scatter, particularly in paleomagnetic inclination, most recently and cogently interpreted as an unresolved contamination from two-polarity Cenozoic overprints of variable strength (Pehrsson and Buchan 1999). In our data set, the range of inclination is not as pronounced, and evidence for

Fig. 3. Examples of stable end-points obtained from Franklin dykes on Devon Island. On equal-area stereoplots, solid and open symbols are downward- and upward-pointing magnetizations, respec tively. On vector diagrams, solid circles are projections of the tip of the magnetization vector on the horizontal plane, and open circles are projections on the vertical plane. Thermal data include an additional intensity decay plot.



contamination by a steep overprint is lacking; the results from specimens with either the steepest positive or negative inclinations (Fig. 7) show linear decay of components to the origin on both AF and thermal plots. This indicates that the remanence direction that they record is a single, well-isolated component, uncontaminated by an overprint. Much of our scatter in declination is due to three anomalously southwest-directed site means (Fig. 6). Local rotation at the sites is unlikely because those three are geographically interspersed among others that bear the modal direction. One of the anomalously directed sites (CG) yields exactly the same U–Pb age as that of one of the modal sites (QA; both 721 Ma as described later in the text), so motion of the Laurentian plate is also not a likely cause of the bimodality. The anomalously directed subset is best explained in terms of geomagnetic secular variation at the time of Franklin emplacement, and, therefore, those sites should be included in the mean directions from this study.

To address the issues of early Cenozoic block rotations and sinistral offset along the Nares Strait transform system, separate means have been calculated for the Canadian and Greenlandic subsets of paleomagnetic data. Among 12 sites passing quality filtering from Canada, the mean of VGPs is (05.8°N, 189.7°E, K = 17.4, A_{95} = 10.7°). Within this subgroup, nine sites from Devon Island alone yield a pole at (06.3°N, 184.0°E, K = 22.9, A_{95} = 11.0°), whereas three sites from Ellesmere Island alone yield a pole at (04.0°N, 206.8°E, K = 17.7, A_{95} = 30.2°). Those latter two means are substantially different, but statistically indistinct owing to the small number of site means from Ellesmere and the consequently large error circle. Two of the three useable Ellesmere sites bear the anomalous southwest-directed remanence vector noted earlier in the text (see Fig. 6), suggesting that the Ellesmere mean direction is biased by coincidentally concentrated sampling of the geomagnetic secular variation.

Table 1. Summary of paleomagnetic results.

	Latitiude (N)	Longitude (W)	Str (°)	W (m)	N_S	D (°)	I (°)	P_{lat} (°N)	P_{lon} (°E)	k	α_{95}
Canadian Franklin dykes											
BB	75°43.1	82°59.2	285	20	9	286.0	–18.0	–5.1	169.3	91	5.4
PH †	75°00.0	79°36.0	255	30	5	282.1	12.6	9.3	180.3	8	28.3
GR	76°25.9	83°01.2	315	40	8	272.9	23.3	12.5	187.1	190	4.0
BG	75°35.7	80°11.7	270	75	7	271.3	34.5	18.7	193.4	24	12.6
BP	75°46.1	81°19.4	265	34	9	279.2	–7.5	–1.4	178.8	57	6.9
BR	75°28.2	81°33.1	275	30	7	282.6	–7.5	–0.5	175.3	226	4.0
EG	75°49.4	82°04.2	255	21	7	280.1	24.8	15.1	181.4	134	5.2
LG	78°43.6	75°41.0	260	50	7	243.1	13.7	1.8	222.0	91	6.4
CG	78°17.1	77°06.6	280	25	7	251.9	1.8	–2.7	210.8	46	9.0
CM*	74°38.0	80°28.0	220	25	7	282.5	1.7	4.2	177.7	55	9.1
EA*	75°49.4	82°04.7	260	40	11	273.4	31.5	17.3	188.9	79	5.5
HM*	74°38.3	80°12.6	260	55	4	244.2	–4.7	–8.9	214.2	240	5.9
PI* †	75°00.4	79°46.2	225	50	5	293.3	35.8	25.1	172.5	22	16.7
CW* †	74°28.6	81°51.9	305	30	5	302.1	–0.2	8.2	157.0	6	35.4
SG2* (R)	75°42.0	84°00.0	305	20	6	101.2	–26.4	16.3	178.6	103	6.6
Greenlandic Franklin dykes											
KL	78°41.0	70°40.0	270	54	8	283.0	10.9	7.9	187.6	74	6.5
TB	76°27.5	69°14.4	295	30	9	296.5	26.4	19.6	178.0	37	8.6
NU2	77°22.7	71°29.0	295	40	8	289.3	16.3	12.3	181.4	123	5.0
PK	77°56.3	72°12.5	280	65	7	288.7	–18.5	–5.5	177.6	132	5.3
KC	77°33.9	70°15.4	300	16	4	290.6	15.7	12.2	181.2	327	5.1
PW	77°11.9	70°50.1	290	10	7	293.6	15.9	13.0	177.8	144	5.1
QA	77°29.1	68°51.9	305	33	6	298.8	26.1	19.5	175.7	106	6.5
KA* †	77°12.3	70°46.6	300	55	2	307.8	–1.6	7.0	162.0	435	12.0
CA (Sill)	76°44.3	73°13.4	n/a	20	5	295.4	9.3	10.2	173.0	234	5.0
GF (Sill)	76°51.4	69°55.5	n/a	10	6	295.6	17.3	14.4	177.0	68	8.2
NU1 (R)	77°24.5	71°33.7	290	40	8	108.8	37.7	–16.6	175.6	59	7.3
HE (R) †	77°25.3	70°08.5	280	21	5	118.3	11.9	0.0	171.0	14	21.5

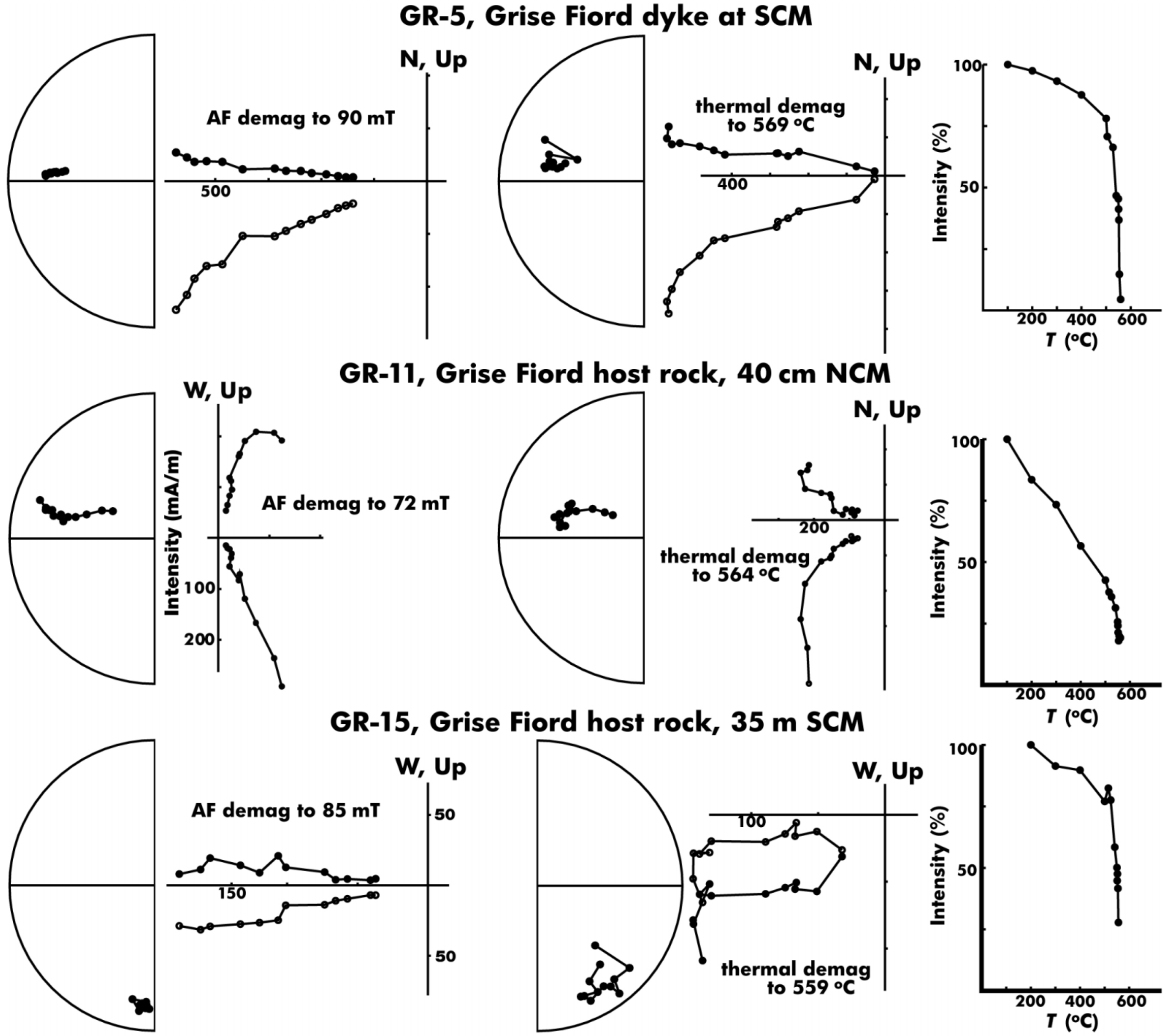
Note: Str, strike (°); W, width (m) of dyke; N_S, number of samples; D, mean declination; I, mean inclination of remanent magnetization; P_{lat} and $P_{lon,}$ north latitude and east longitude, respectively, of the VGP; k, Fisher precision parameter; α_{95}, radius of 95% confidence circle about the mean; (Sill), measurements for sills rather than dykes; (R), reversed-polarity,. corresponding antipoles are listed in subsequent columns; *, samples oriented using magnetic compass only; †, sites excluded from the means owing to quality filtering as discussed in the text.

Oakey and Damaske (2006) and Harrison (2006) proposed that Jones Sound, located between Devon and Ellesmere islands, contains the boundary between the North American and Greenland plates at the time of their early Cenozoic relative motions. Paleomagnetic discordance between Devon and Ellesmere data subsets in this study thus demand careful scrutiny. If early Cenozoic counterclockwise rotation of Ellesmere relative to Devon were the cause of the paleomagnetic discordance, then $\sim 20°$ of such rotation would be required to align the two subsets. The Jones Sound sedimentary basin has geophysically defined margins (Harrison et al. 2006; Oakey and Stephenson 2008) that do not appear capable of accommodating that amount of rotation. There is undoubtedly a small amount (some tens of kilometres) of extension localized in the Jones Sound region (Harrison 2006), but this amount of extension will not deflect the paleomagnetic directions or poles significantly: indeed, the lateral translations hypothesized for the entire Nares Strait region do not exceed a few hundred kilometres, equivalent to 2°–3° of pole arc distance. Only vertical axis components of motion, represented by Euler poles in close proximity to the region, can produce significant paleomagnetic pole discordance via a spherical trigonometric "levering" effect upon the low-inclination directional data.

The combined Devon + Ellesmere two-polarity paleomagnetic pole passes six of the seven of Van der Voo's (1993) quality criteria, allowing for the "structural control" criterion to be satisfied at the archipelago level rather than necessarily for the entire North American plate. The only criterion not satisfied is that of similarity to younger paleopoles, in this case Middle–Late Cambrian poles from Laurentia (Torsvik et al. 1996). Our positive baked-contact test, however, demonstrates that this similarity is coincidental.

The Greenland two-polarity paleomagnetic pole, based on 10 sites passing our quality filter, is (08.8°N, 178.5°E, K = 45.9, A_{95} = 7.2°). This result, by itself, passes five of Van der Voo's (1993) quality criteria, lacking a field stability test and showing similarity to the Middle–Late Cambrian poles from Laurentia. At the $p = 0.05$ confidence level, this pole is indistinct from the poles derived from Canadian Franklin data in this study. If all of our data are combined into a mean direction with no internal rotations, the resulting paleomagnetic pole from 22 sites is (07.2°N, 184.5°E, K = 22.5, A_{95} = 6.7°). As discussed later in the text, however, this combined result and all of the various subgroup poles in this study are distinct from previous Franklin paleomagnetic results from Baffin Island, Victoria Island, and the mainland autochthonous Canadian Shield.

Fig. 4. Results of a paleomagnetic baked-contact test at site GR (modified after Denyszyn et al. 2006). Note similar magnetization directions for the dyke and proximal host rock using either alternating field (AF) or thermal demagnetization (demag) techniques, and the different (p = 0.05) magnetization direction in the more distal unbaked host rock. SCM and NCM are the southern and northern chilled margins, respectively. Other symbols as in Fig. 3.



U–Pb geochronology

The uranium and lead isotopic results for each analysis were corrected for common Pb contamination using the isotopic composition of laboratory blank (see footnotes to Table 2). Because of the small size of the fractions (typically <0.5 μg), which was necessary because of the low yield and to analyze the best quality crystals, a significant proportion of the total ^{207}Pb is common. However, precise ages can still be determined using the ^{238}U–^{206}Pb system, which is only possible because of low (normally ca. 1 pg or less) laboratory blanks. The use of one system makes the age results susceptible to possible secondary Pb loss. This tends to be much less of a problem with baddeleyite than it is with zircon (Krogh et al. 1987) and reproducibility of $^{206}Pb/^{238}U$ ages on individual fractions provides a first-order test of accuracy. Averages of $^{206}Pb/^{238}U$ ages and fit parameters were calculated using the procedure of Davis (1982). U decay constants are from Jaffey et al. (1971) and all errors are reported as 2σ.

The Cadogan Glacier (CG) dyke (Fig. 8*a*) was analyzed using four single baddeleyite crystals and two fractions containing two crystals each. Grains were typically fresh, dark brown blade-like fragments, ranging from 50 to 150 μm in the largest dimension. However, some grains were relatively dark and may have been somewhat altered. Furthermore, this sample suffered from having unusually high blanks (up to 9.9 pg of common Pb) because of the difficulty in handling the tiny baddeleyite crystals, often no more than 20 μm in the largest dimension. Concordant $^{206}Pb/^{238}U$ ages were obtained, although some data scatter outside of the

Fig. 5. Plots of susceptibility (K) at temperature (T), normalized to room temperature susceptibility (K_0), versus T (in °C) for representative samples from the Devon Island swarm. Note the uniform drop in susceptibility at 580 °C, indicative of low Ti magnetite as the carrier of the magnetization in the east–west-trending dykes.

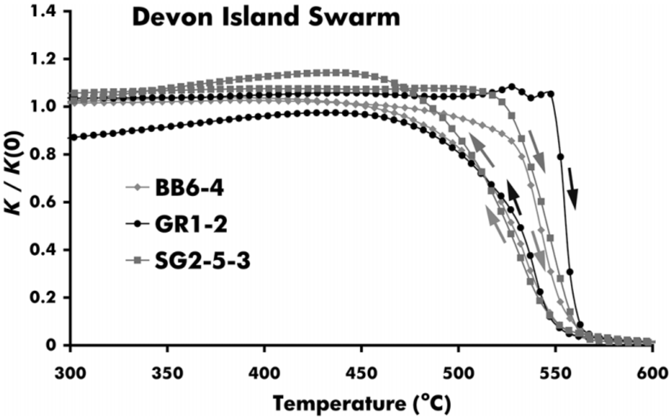


Fig. 6. Equal-area stereoplot of the quality-filtered, site-mean paleomagnetic characteristic remanence magnetization directions in the Franklin dykes and sills in this study. Circles,Greenland data; squares,Canada data; solid, lower hemisphere; open, upper hemisphere. Italics indicate south-seeking directions (reversed polarity). Asterisks indicate site orientations by magnetic compass only. U–Pb ages in Ma. See Fig. 2 for site locations.

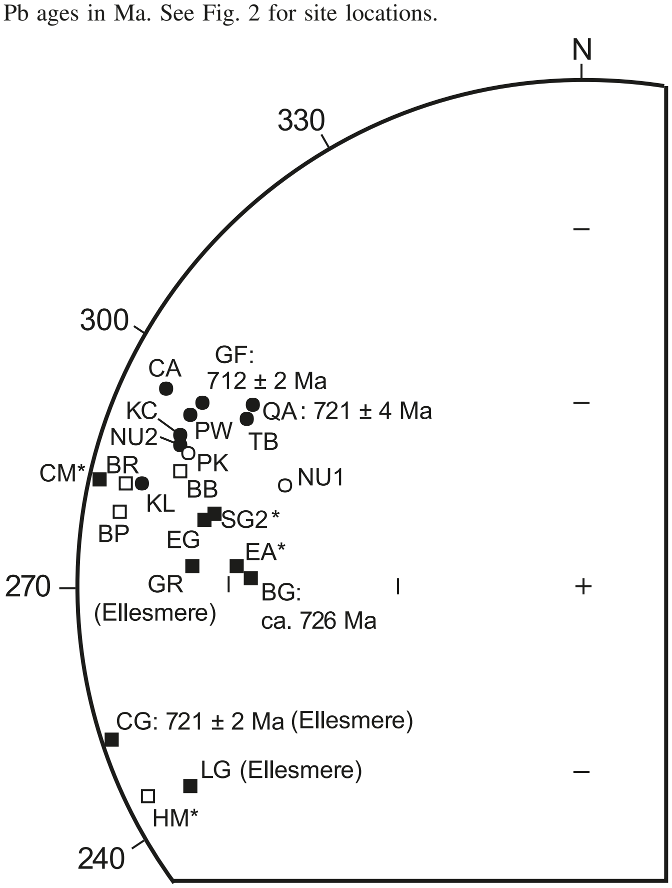
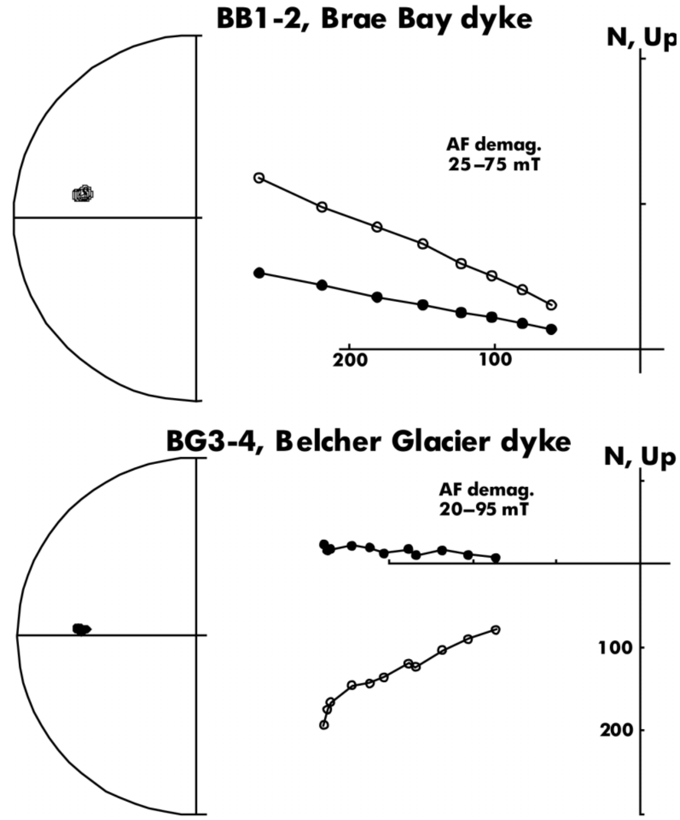


Fig. 7. Examples of stable paleomagnetic end-points obtained from the Franklin dykes with the steepest-up (BB) and steepest-down (BG) inclinations, showing linear decay to the origin. This indicates that a single component of magnetization is being removed, without a secondary steep overprint. AF demag., alternating field demagnetization. Symbols as in Fig. 3.



error. The oldest data (fractions 1, 5, and 6) give an average $^{206}Pb/^{238}U$ age of 721 ± 2 Ma (Table 2), whereas the other fractions define a resolvable range down to 714 ± 2 Ma. These different $^{206}Pb/^{238}U$ ages appear to be "stacked" below the fractions giving the oldest age, a consequence either of minor Pb loss that is noticeable in the more sensitive $^{206}Pb/^{238}U$ system or of loss of a mobile daughter product of ^{238}U, such as ^{222}Rn, possibly related to (or exacerbated by) the effects of alpha recoil (Denyszyn et al. 2009).

Baddeleyite grains from the Qaanaaq (QA) dyke in Greenland were rare, but fresh and relatively large (40–100 µm long). They tended to take the form of long, light brown, thin needles and fragments, the smallest of which were combined into fractions of two grains each. Four fractions of one or two grains each gave precise overlapping data with an average $^{206}Pb/^{238}U$ age of 721 ± 4 Ma (Fig. 8*b*).

Data from the Belcher Glacier (BG) dyke are discordant (Fig. 8*c*). The average $^{207}Pb/^{206}Pb$ age is 726 ± 24 Ma from four fractions of three or four small baddeleyite grains each. Most grains were small (<50 µm long), thin blades, pale brown in colour, and many had a "frosted" appearance, interpreted to be the result of a fine coating of zircon. Because the baddeleyite crystals were tiny and fragile, this coating could not be removed by any abrasion techniques. Some very small (<30 µm) skeletal crystals of what may have been zircon were observed in the final concentrate after mineral separation, but these were too small to abrade and analyze. The three most discordant fractions show calculated Th/U ratios well in excess of those that are typical of baddeleyite, which normally discriminates strongly against Th relative to U. It is likely that the grains of baddeleyite in this

Table 2. U–Pb isotopic data for baddeleyite from mafic dykes in Canada and Greenland.

Sample	Fraction	w_{mod} (mg)	Th/U	cPb (pg)	$\frac{^{206}Pb}{^{204}Pb}$ (meas.)	$\frac{^{206}Pb}{^{238}U}$	2σ	$\frac{^{207}Pb}{^{235}U}$	2σ	$\frac{^{206}Pb}{^{238}U}$ age (Ma)	2σ	$\frac{^{207}Pb}{^{206}Pb}$ age (Ma)	2σ	% disc.	Rho conc.
1. CG-7, Cadogan Glacier diabase, Ellesmere Island															
1	1 fresh frag	0.0001	0.001	4.9	314	0.118385	0.000	1.02526	0.034	721.3	2.1	701.8	66.9	–2.9	0.8372
2	1 fresh frag	0.0001	0.003	9.9	121	0.117032	0.001	1.01536	0.098	713.5	5.1	705.6	196.7	–1.2	0.9530
3	2 large dk blade frags	0.0001	0.005	5.3	86	0.116287	0.001	1.01982	0.146	709.2	7.5	728.5	296.3	2.8	0.9673
4	2 dk brn frags	0.0002	0.030	1.0	606	0.117100	0.000	1.02284	0.018	713.9	2.4	720.0	33.3	0.9	0.5417
5	1 dk brn frag	0.0002*	0.005	0.2	2179	0.118128	0.000	1.04055	0.006	719.8	2.1	737.9	9.9	2.6	0.6338
6	1 dk brn frag	0.0001*	0.007	0.2	1402	0.118934	0.001	1.04893	0.009	724.4	2.9	740.5	15.0	2.3	0.6058
2. QA-2, Qaanaaq diabase, Greenland															
1	1 large frag	0.0003	0.028	1.7	128	0.119016	0.001	1.12123	0.111	724.9	4.7	843.9	193.1	18.5	0.8049
2	1 needle, brn	0.0001	0.199	1.2	217	0.118201	0.001	1.02948	0.052	720.2	3.8	713.9	99.8	–0.9	0.7209
3	2 frags, brn	0.0001	0.006	3.2	45	0.116653	0.003	0.99875	0.374	711.3	19.4	677.4	854.8	–5.3	0.9490
4	2 frags, brn	0.0001	0.071	1.4	154	0.118206	0.001	1.02320	0.075	720.2	4.9	700.8	147.0	–2.9	0.7874
3. BG-6, Belcher Glacier diabase, Devon Island															
1	3 euh frags, med brn	0.0002	0.556	0.7	449	0.106825	0.000	0.93861	0.023	654.3	2.2	732.5	46.9	11.2	0.6179
2	3 frags, mod fresh	0.0002	1.153	2.9	90	0.084667	0.001	0.76765	0.101	523.9	5.4	798.7	267.1	35.8	0.9596
3	4 frags	0.0002	0.524	0.8	674	0.102523	0.000	0.89604	0.014	629.2	1.5	721.3	29.9	13.4	0.5941
4	4 small dk brn euh	0.0001	0.064	0.8	114	0.116064	0.002	1.04461	0.106	707.9	8.9	767.7	205.8	8.1	0.6675
4. GF-1, Granville Fjord diabase sill, Greenland															
1	3 fresh med brn frags	0.0002	0.129	1.3	275	0.116844	0.000	1.01598	0.040	712.4	2.8	710.4	77.3	–0.3	0.748
2	3 fresh frags, mod brn	0.0002	0.049	0.8	179	0.116834	0.001	1.04240	0.063	712.3	5.3	764.9	119.0	7.3	0.656
3	4 small frags	0.0001	0.015	3.2	50	0.117748	0.003	1.09664	0.335	717.6	18.4	854.7	644.7	16.9	0.876

Note: w_{mod}, estimated weights for the grains assuming 300 ppm U; Th/U calculated from radiogenic $^{208}Pb/^{206}Pb$ ratio and $^{207}Pb/^{206}Pb$ age assuming concordance; cPb, total measured common Pb assuming the isotopic composition of laboratory blank: 206/204 = 18.221; 207/204 = 15.612; 208/204 = 39.360 (errors of 2%); % disc., % discordance for the given age; Rho conc., error correlation coefficient for concordia data; brn, brown; dk, dark; euh, euhedral; frag, fragment; mod, moderately. Errors are given at 2σ. Sample locations are given in Table 1. Asterisks indicate sample weights calculated from actual measurement of grain dimensions.

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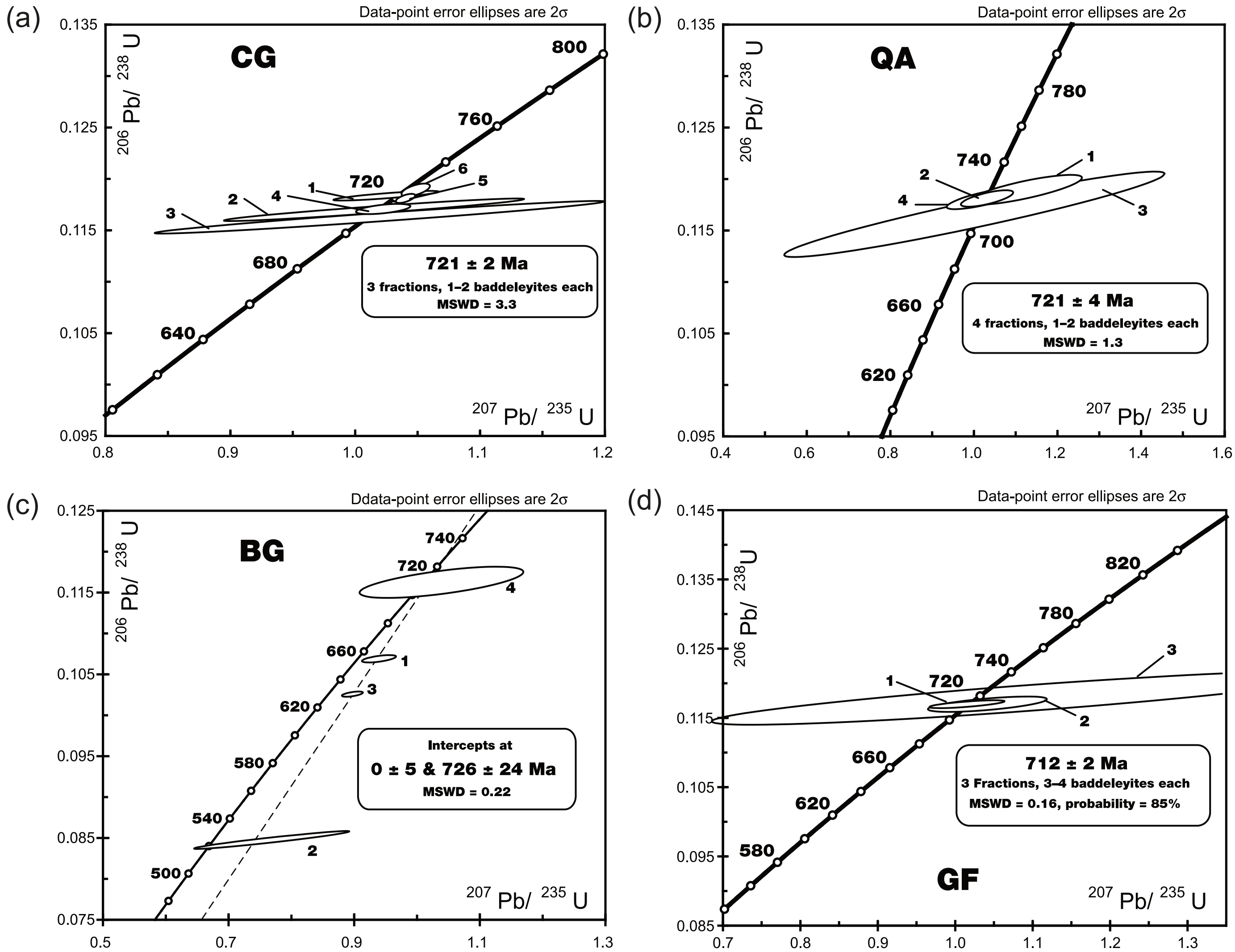
Fig. 8. U–Pb geochronological results for baddeleyite from east–west-trending Franklin dykes. (*a*) Cadogan Glacier (CG) dyke, Ellesmere Island. (*b*) Qaanaaq dyke (QA), Greenland. (*c*) Belcher Glacier dyke (BG), Devon Island. (*d*) Granville Fjord sill (GF), Greenland. MSWD, mean square of weighted deviates.



sample were coated with a thin layer of high-uranium, radiation-damaged zircon that may have suffered low-temperature Pb loss. This suggests that zircon crystallized during a late stage of crystallization of the magma, perhaps owing to increased silica activity.

A Thule sill (site GF) was analyzed using three fractions of three or four baddeleyite grains each. The sample yielded several very small (30–80 μm) thin, light brown blades with visible striations, and so multiple grains were analyzed in each fraction. Precise overlapping data have an average $^{206}Pb/^{238}U$ age of 712 ± 2 Ma (Fig. 8*d*). Mineralogically and paleomagnetically, the sill is indistinguishable from the east–west-trending dykes in the area, but its age is distinctly younger.

Discussion

Paleomagnetic data reported in this study are discordant relative to previously published poles from the Franklin igneous events in Canada. Table 3 lists the earlier results with their Van der Voo (1993) quality ratings. No fewer than nine independent studies have been published with new data between 1971 and 1983 from the Franklin igneous events in Canada. The groundbreaking work by Fahrig et al. (1971) included a broad sampling area from the Coronation Gulf to Baffin Island and produced a dual-polarity pole from 24 sites that were considered least affected by the steep inclination overprint. In the following year, Robertson and Baragar (1972) augmented the data set from Coronation mafic sills intruding the Shaler Supergroup of early Neoproterozoic age (Rainbird et al. 1996), obtaining only the east-directed polarity. A geochronology site contributing to the 723^{+4}_{-2} Ma (U–Pb, zircon) upper intercept age of Heaman et al. (1992) is on the same island in Coronation Gulf—and could therefore be from the same intrusion—as one of the paleomagnetic sites that was studied in detail. Fahrig and Schwarz (1973) returned to Baffin Island for more extensive sampling and localization of the steep overprint primarily to the northeast side of the island, strengthening the case that the overprint was related to early Cenozoic rifting and separation of Greenland from North America. Park (1974) added three more sites in dykes of presumed Franklin-aged mag-

Table 3. Paleomagnetic poles from the Franklin igneous event.

Pole	Description	N_N	N_R	P_{lat}(°)	P_{lon}(°E)	A_{95}(°)	1234567 Q	Reference
	Individual poles							
F+71	Coronation–Baffin intrusions	19	5	8.0	167.0	5.0	1100110 4	Fahrig et al. 1971†
RB72	Coronation sills	0	13	–1.0	163.0	9.0	1100100 3	Robertson and Baragar 1972†
FS73	Southern Baffin dykes	4	6	6.3	168.2	4.8	0100110 3	Fahrig and Schwarz 1973
P74	Miscellaneous dykes (N.W.T., Ont.)	1	2	3.0	161.3	9.2	0101110 4	Park 1974‡
PH75	Natkusiak lavas and sills (Victoria I.)	0	16	–7.0	163.0	4.0	1100100 3	Palmer and Hayatsu 1975†
P81	Brock Inlier sills	0*	5*	–2.0	165.0	12.0	0010110 3	Park 1981
CF83	Northwest Baffin dykes	4	9	9.2	153.3	3.8	0100110 3	Christie and Fahrig 1983
P+83	Natkusiak upper lavas (Victoria I.)	8	6	6.0	159.0	5.7	0110110 4	Palmer et al. 1983
	Grand-mean poles							
FS73 g	Coronation–Baffin intrusions	22	24	5.8	165.8	3.8	1100110 4	Fahrig and Schwarz 1973†
P94	Victoria I.–Baffin lavas and intrusions	?	?	5.0	163.0	5.0	1111110 6	Park 1994†
B+00	Victoria I.–Baffin lavas and intrusions	11	15	8.0	163.0	4.0	1110110 5	Buchan et al. 2000†
D+09can	Victoria I.–Baffin lavas and intrusions	28	28	6.7	162.1	3.0	1111110 6	*Site-filtered previous studies*
D+09cg	Canada and Greenland lavas and intrusions	48	30	8.4	163.8	2.8	1111110 6	*Site-filtered, variably rotated*

Note: N_N, samples with normal polarity; N_R, samples with reverse polarity; P_{lat}(°N), P_{lon}(°E), north latitude and east longitude, respectively, of the VGP; A_{95}, radius of cone of 95% confidence (in degrees). Q and preceding numbers refer to quality criteria of Van der Voo (1993): 1, present; 0, absent.

*Three sites contain mixed polarity; values shown here represent the dominant polarity at each site.

†Age criterion satisfied by subsequent data in Heaman et al. (1992).

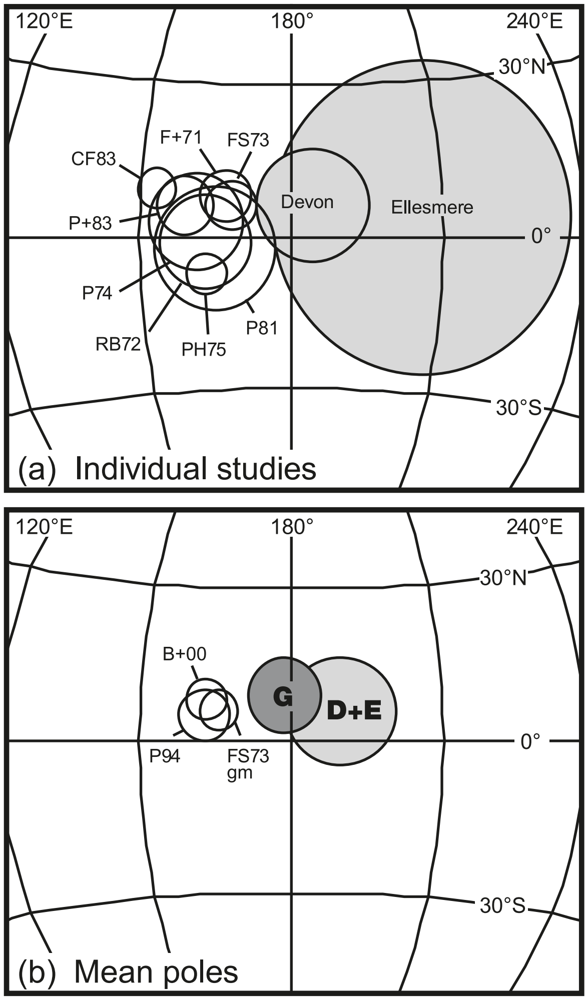
‡Positive baked-contact test implied by comparison with Park's (1973) study of Paleoproterozoic host rocks from the same area.

matism, but one of the dykes is located in southern Ontario and would represent an extreme geographic outlier of the Franklin large igneous province. Given the similarity between the Franklin poles to several low-inclination results from the Ediacaran–Cambrian apparent polar wander path of Laurentia (see McCausland et al. 2007), we suspect that the Ontario dyke, as yet undated, is not of Franklin age. Palmer and Hayatsu (1975) analyzed Victoria Island sills and lower lavas of the Natkusiak Formation, all bearing a single polarity of remanence matching that of the previously studied Coronation Sills. Jones and Fahrig (1978) added a few more sites of presumed Franklin-age dykes (but with a northeast trend) on Somerset Island, but the age assignment is purely based on comparison with the previous Franklin paleomagnetic poles.

Paleomagnetic research on the Franklin igneous event continued in the early 1980s, with Park's (1981) study of mafic sills in the Brock Inlier that found nearly identical results to those from the Coronation sills exposed just 200–300 km to the east. Christie and Fahrig (1983) presented new Franklin data in a study that focused on the so-called Borden dykes intruding the Bylot Supergroup on northwestern Baffin Island; that paper presented two apparently positive baked-contact tests on Franklin dykes intruding an older set of the so-called Borden swarm. Subsequently, Pehrsson and Buchan (1999) dated the Borden dykes as coeval to Franklin and showed that the baked-contact tests were not as conclusive as the original study averred. The more northerly trending dykes in this area are now referred to as the "Strathcona Sound" swarm (Buchan and Ernst 2004), continuing on Devon and Ellesmere islands as the Clarence Head swarm (Denyszyn et al. 2009). Somewhat inexplicably, they yield an anomalous paleomagnetic pole relative to those from all previous studies (Fig. 9), but the distinction is not large (5°–10°) and could arise from combinations of several factors ranging from anisotropic biases to local vertical-axis rotations in the Strathcona Sound area. A detailed sampling of the Natkusiak flood basalts and feeder sills on Victoria Island constituted the final published paleomagnetic work on the Franklin event prior to the present study (Palmer et al. 1983). Natkusiak volcanics yielded a very stable two-polarity remanence, whereas the underlying sills showed large amounts of directional scatter that may be attributable to Cenozoic overprints (Pehrsson and Buchan 1999).

Three studies reported grand-mean paleomagnetic poles for the Franklin event, representing the ever-increasing data sets (Fahrig and Schwarz 1973; Park 1994; Buchan et al. 2000). All three of these mean poles are within error of each other (Fig. 9*b*), a testament to the robustness of the paleomagnetic results. Several other paleomagnetic studies in northern Canada also bear on the reliability of the Franklin paleopole, namely on the host rocks to various Franklin intrusions. Park (1973) reported a distinct paleomagnetic direction from metamorphic rocks of the Melville Peninsula, in the same region as the two Franklin dykes presented the following year (Park 1974); although no hybrid magnetizations, necessary for a complete baked-contact test, were observed, the combined results of these two studies must constitute a positive field test by most standards. Fahrig et al. (1981) reported reliable data from the Bylot Supergroup that hosts both the Franklin and Strathcona Sound dykes, but with no baked-contact tests into those sedimentary country rocks. Park (1992) reported data from the Shaler Supergroup sediments in the Brock Inlier, intruded by the gabbro sills he previously studied (Park 1981). The later work claimed to distinguish between an "A" direction from sediments unaffected by Franklin magmatism, and a "B" direction from sediments either observed or inferred to lie in proximity to Franklin intrusions. This seemingly positive baked-contact test is invalidated by close examination of the combined data from the two Brock Inlier studies (Park 1981, 1992). Both of the so-called "A" and "B" directions are observed in the intrusions, and where "B"-bearing sediment sites are shown to be located near Franklin rocks, their directions are

Fig. 9. Locations of Franklin paleopoles with A_{95} (95% confidence) circles. (*a*) Individual results from regional subsets. (*b*) Mean poles from autochthonous Canada (unfilled) and new poles (shaded) from Greenland (G) and Devon + Ellesmere (D+E) as reported herein. Other abbreviations as in Table 3. Numbered ellipses refer to analyses in Table 2.



quite distinct from those of the adjacent intrusions. Park and Rainbird (1995) arrived at the same conclusion of widespread Franklin overprinting in the Brock Inlier, based on comparisons to data from correlative pre-Franklin sedimentary rocks on Victoria Island. Those rocks also showed widespread overprinting, with exception of one site in the Boot Inlet Formation, but this does not in itself constitute a rigorous field stability test. Finally, Symons et al. (2000) sampled Pb–Zn ore deposits at the Nanisivik mine, both distal and proximal to a likely Franklin-age mafic dyke. Although that paper claims a positive baked-contact test, the mean direction from the baked orebody is clearly distinct from that of the dyke, and the test is suggestive at best.

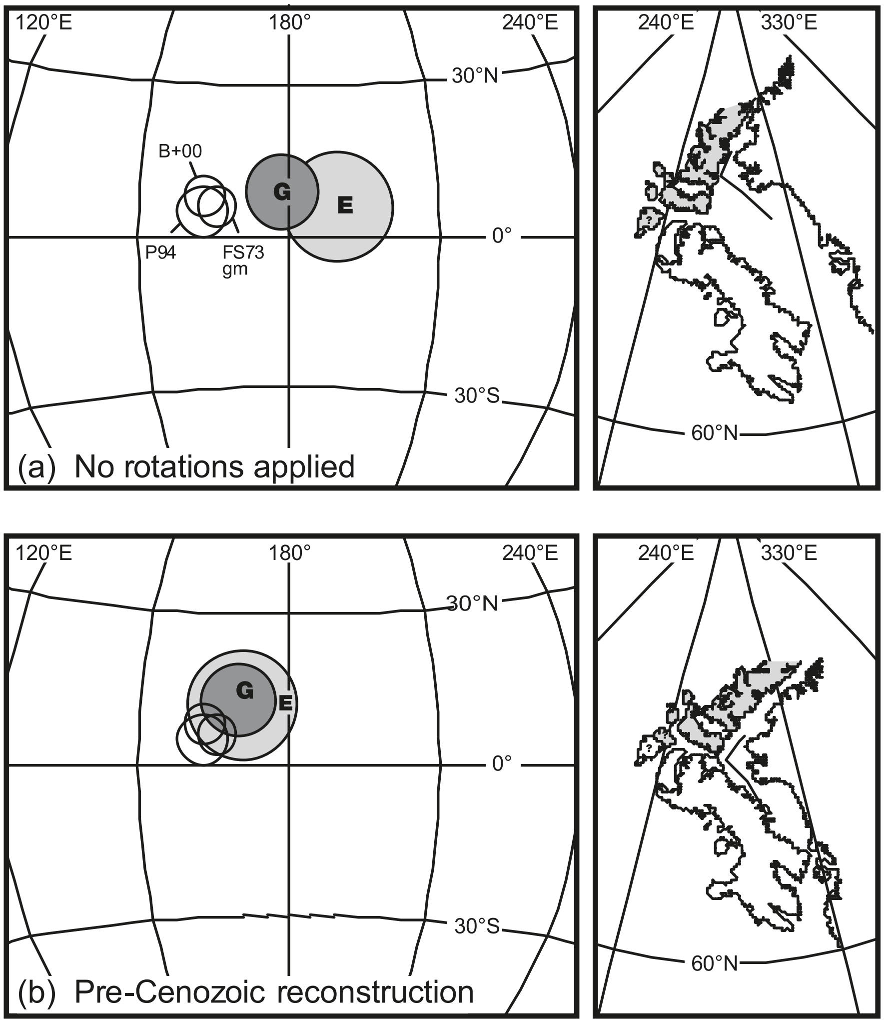
It is perhaps surprising that among all the previous work on Franklin paleomagnetism, only Christie and Fahrig (1983) attempted complete baked-contact tests, and these have been discussed and discounted by Pehrsson and Buchan (1999). Only the combined data from Park (1973) and Park (1974) in the Daly Bay area of the Churchill Province together constitute a positive, albeit weak, baked-contact test that supports primary remanence of the dykes, which are undated but presumably Franklin in age based on their remanence direction. With impressive reproducibility across an expansive geographic area, however, the Franklin grand mean is considered to be "key" in reliability by Buchan et al. (2000). Our baked-contact test at site GR includes stable directions from the target lithology, its baked zone, and distant unbaked host rock, adding further to the reliability of Franklin paleomagnetic directions. Our study is also the first to obtain primary paleomagnetic remanence and U–Pb ages from the same outcrops of Franklin rocks. Therefore, the discrepancy between our mean paleomagnetic poles and those from previous studies demands discussion.

There are many factors that could lead to directional discordance from one region to another. Directional biases from unresolved contamination by secondary overprints, as discussed earlier, have mainly involved inclinations owing to the steepness of the early Cenozoic remagnetizing episode (Pehrsson and Buchan 1999) and appear to be minimized between our results and those of earlier studies, which differ mainly in declination and east–west spread of poles (Fig. 9). A difference in ages of the poles, implying apparent polar wander from one pole to another, appears to be ruled out by the spread of U–Pb baddeleyite ages reported in this study (721 to 712 Ma), which spans those applied to other regions (Franklin sills and lavas dated at 723^{+4}_{-2} and 718 ± 2 Ma, Heaman et al. 1992; and Franklin dykes younger than 720 ± 8 and 716^{+4}_{-5} Ma, Pehrsson and Buchan 1999). Time-varying non-dipole geomagnetic field components could in principle contribute to directional scatter, but as noted earlier in the text, there is no prominent covariance of paleopole location with age. Also, the Proterozoic geomagnetic field appears to be dipolar according to the first-order constraints from large evaporite basins, including some of the best constraints from the mid-Neoproterozoic (Evans 2006). Hereafter we explore the possibility that most of the Franklin paleomagnetic pole discordance can be resolved by early Cenozoic block rotations around the Nares Strait region.

The "Nares Strait problem" has been a contentious issue plaguing Arctic geology for decades (Dawes and Kerr 1982; Okulitch et al. 1990; Tessensohn et al. 2006 and references therein). As noted earlier in the text, the maximum proposed offsets in the Nares Strait region do not exceed ca. 300 km by any estimates, and that amount of translation falls entirely within the uncertainty limits of all but the most precise paleomagnetic poles ever determined—certainly well within the uncertainties on data presented in this paper. However, our results are very sensitive to vertical-axis rotations because of their low inclination angles and consequent "levering" effect on the distant paleomagnetic poles.

Our data unequivocally support a local rotation on the order of 20° between autochthonous Canadian Shield and a microplate comprising the pre-Cenozoic basement areas of Ellesmere and Devon islands. Treating this motion rigorously in a plate tectonic framework requires application of Euler rotations, and a predominant sense of vertical-axis rotation is equivalent to an Euler pole that is proximal to the rotating block. An Euler pole location anywhere within a small circle cap of about 10°–20° radius (ca. 1000–2000 km from the sites) will adequately rotate the data in a vertical-axis sense. Other constraints must be used to pinpoint loca-

Fig. 10. Mean paleomagnetic poles (left) from the autochthonous Canadian Shield (unfilled 95% error circles, labelled in Fig. 9) and this study. E, Ellesmere and Devon islands; G, Greenland. Coastline reconstructions (right) according to variable rotation models for pre-Cenozoic Laurentia. (*a*) All poles and coastlines in present coordinates. (*b*) Greenland reconstructed according to Roest and Srivastava (1989), and the Ellesmere microplate (including Devon, Cornwallis, and Somerset islands) reconstructed by Euler parameters (72°N, 274°E, –20° counterclockwise) as proposed in this study. The piecewise smooth curve west of Greenland indicates the geophysical limit of that plate (Harrison et al. 2006; Oakey and Damaske 2006).



tion of the Euler pole within this cap region, and for this we rely on boundaries between rotating blocks and intervening pull-apart rift basins, as determined by regional crustal geophysics. Harrison et al. (2006), Neben et al. (2006), Oakey and Damaske (2006), and Oakey and Stephenson (2008) well illustrate these boundaries, including three features of particular relevance to this problem. First, Jones Sound (between Ellesmere and Devon islands) has experienced some extension but, as noted earlier, probably cannot account for substantial relative rotation. Second, and in contrast, Lancaster Sound (between Devon and Baffin islands) contains a substantial thickness of Cenozoic sediments in a westward-narrowing, wedge-shaped basin that does appear to be capable of accommodating such rotation. Third, an expansive area of thinned and submerged crust west of Inglefield Land in northwestern Greenland appears to be tied tectonically to the Greenland plate, most impressively by the aeromagnetic imaging of the Kap Leiper dyke (our site KL) almost completely across Smith Sound. The termination of its magnetic anomaly immediately east of the Ellesmere coast and absence of any correlative mafic dyke along its extrapolated linear trajectory on Ellesmere Island suggest placement of the so-called "Wegener fault" on that side of the waterway (Denyszyn et al. 2006). Figure 10 includes coastlines of Baffin, Devon, Ellesmere, and Greenland, as well as the approximate western margin of the Greenland continuous crust that honours the geophysical constraints.

In any pre-Cenozoic reconstruction of Arctic North America, Greenland must be brought closer to Canada, reducing the width of the Labrador Sea. Most restorations are broadly similar to each other. Seafloor spreading in the Labrador Sea was restored by Srivastava and Tapscott (1986), demonstrating shifting stage poles of rotation during the separation. The model was updated by Roest and Srivastava (1989), and both of these total reconstructions are still used by recent global kinematic models (e.g., Besse and Courtillot 2002; Torsvik et al. 2008). They both produce acceptable overlap of our Greenland Franklin pole with those of autochthonous Canada (Fig. 10). Alternative reconstructions include those of Torsvik et al. (2001), who provide a standard "model I" that unsurprisingly brings our Greenland data into alignment with those of the Canadian Shield, as well as a controversial "model II" that rotates our Greenland Franklin result far from the other Franklin poles. Finally, Müller et al. (2008) produce a model that only slightly varies from that of Roest and Srivastava (1989). We adopt the Roest and Srivastava (1989) reconstruction in this paper, which constrains the possible reconstructed positions of Devon and Ellesmere islands.

Our paleomagnetic data from Devon Island and Ellesmere Island subsets are consistent within error, suggesting that a joint Devon + Ellesmere plate may be considered in the reconstructions—the tens of kilometres of extension across Jones Sound being neglected for simplicity. Given that anticlockwise rotation of this block, relative to autochthonous Canada, is implied by our data, its western margin should be marked by compressional features. The Central Ellesmere Fold Belt, along with its foreland Sverdrup Basin, are often included in discussions on the Nares Strait problem (Higgins and Soper 1989; Tessensohn and Piepjohn 1998; Harrison 2006). The southern continuation of this orogen, through Cornwallis Island and extending southward to the Boothia Peninsula as defined by the Cornwallis foldbelt (Mayr et al. 2004), is less often described in this context. We consider all of these compressional structures to define the western margin of an early Cenozoic "Ellesmere microplate" that includes southern Ellesmere, Devon, and Cornwallis islands, as well as possibly Somerset Island.

The geographic extent of the Ellesmere microplate may also be assessed paleomagnetically. Previous work on Ellesmere Island has provided ample evidence for anticlockwise rotation during Cenozoic time, amounts varying between about 15° and 25° (Wynne et al. 1983, 1988; Jackson and Halls 1988; Tarduno et al. 1997, 2002). Much of the interpretation of those results was given in the context of local oroclinal bending in the Franklin foldbelt, but our results suggest that the rotations are characteristic of the entire Ellesmere microplate, including its basement exposures. The western boundary should correspond with the Cornwallis foldbelt as previously noted. On Prince of Wales Island, both the Devonian Peel Sound Formation (Dankers 1982) and the 1268 Ma Savage Point Sill (Jones and Fahrig 1978; Mayr et al. 2004) appear to be autochthonous to Laurentia; this suggests that the outermost thrust sheets of the Cornwallis foldbelt have not rotated with the Ellesmere microplate.

With the geographic extent of the Ellesmere microplate thus defined, we investigate possible locations of the early Cenozoic Euler pole that can accommodate its motion relative to Canada. The wedge-like closing of Lancaster Sound, plus avoidance of overlap between Ellesmere and Greenland, suggest a pole location at 72°N, 86°W. Such a restoration is similar to that provided by Rowley and Lottes (1988), but that earlier model only included 15.6° of Ellesmere rotation, which is not sufficient to align our Franklin data optimally relative to those of previous studies. An Euler pole in this location explains the decrease in Eurekan orogenic shortening from north to south, as indicated by the transition from a well-developed fold-thrust belt on central Ellesmere Island to a mere anticlinorium on northern Boothia Peninsula (Frisch 1988; Mayr et al. 2004). The Ellesmere microplate Euler pole is also similar to the "Stage 1" pole determined for Greenland by Geoffroy et al. (2001), based on marine magnetic anomalies in Baffin Bay and Labrador Sea. This suggests that Greenland and the Ellesmere microplate were in fact joined and rotating together relative to autochthonous Canada, from Late Cretaceous to Paleocene time, followed by a propagation of the plate boundary through Nares Strait region around Chron 25 near the end of the Paleocene (ca. 55 Ma).

This brings us at last to the Nares Strait problem. Dawes (2009) reiterated the apparent difficulty in driving a major (ca. 200 km offset) transform boundary through Smith Sound because of the geological similarities between Ellesmere Island and northwest Greenland. Following Harrison (2006), he treated Ellesmere + Greenland as a rigid block that must be considered a "linchpin" in pre-Cenozoic reconstructions. Our data strongly suggest otherwise. It is clear that Greenland cannot be restored by the same 20° rotation that must be applied to the Ellesmere microplate because that would greatly exceed the seafloor spreading extension in the Labrador Sea. It is also clear that the mere 14° of Euler rotation to restore Labrador Sea spreading (Roest and Srivastava 1989) will not be sufficient to align our Devon + Ellesmere Franklin pole to its autochthonous Canadian counterparts, and, more importantly, rotation of Devon + Ellesmere as rigidly attached to Greenland would result in substantial overlap atop northern Baffin Island, which is not tectonically permitted. Thus, an extinct plate boundary between Ellesmere and Greenland is required by our Franklin paleomagnetic data, in the context of any reasonable closure of the Labrador Sea. We note here that our solution brings the main dyke concentrations on Greenland and Devon Island into alignment, thus removing a major anomaly in the Franklin swarm in this area—a 200 km sinistral offset that would, without a Nares Strait fault displacement, be a unique feature in comparison with major swarms elsewhere in the world. Although detailed discussion of all the piercing points relevant to the Nares Strait problem is beyond the scope of this paper, further assessments of fault offsets in this region must include the 20° rotation of Ellesmere microplate as a starting point for subsequent analysis.

We combine paleomagnetic data from earlier studies of the Franklin igneous event with our new data from Devon and Ellesmere islands and Greenland, according to the three-plate, pre-Cenozoic total reconstruction as discussed earlier in the text. A grand-mean paleomagnetic pole is calculated from all studies, previously and current, according to the uniform quality filtering that we used to restrict our own data set. The new grand-mean pole is (8.4°N, 163.8°E, A_{95} = 2.8°), comprising results from 48 sites of "normal" polarity and 30 sites of "reversed" polarity. The only Van der Voo (1993) criterion not met by this pole is its similarity to younger (Cambrian) results from North America. The pole's primary age of remanence is demonstrated by baked-contact tests into Paleoproterozoic basement country rock from both our current study and earlier work (Park 1973, 1974). Our new Franklin grand-mean pole differs only slightly from earlier grand means (Buchan et al. 2000), but it is based on three times as many sites, all of which pass quality filtering. We consider the present compilation to represent best the full extent of Franklin magmatism, providing not only a key paleomagnetic pole for Rodinia reconstructions but also a first-order tectonic reconstruction of the pre-Cenozoic Laurentian Arctic margin.

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