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The Bear River dykes (1265–1269 Ma): westward continuation of the Mackenzie dyke swarm into Yukon, Canada

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Abstract

The 1.27 Ga Mackenzie dyke swarm, the largest on Earth, radiates from a point in the western Canadian Arctic and extends across much of the Canadian Shield. Possible western extensions of the swarm are largely obscured by younger sedimentary cover. Paleoproterozoic inliers in northern Yukon host the Bear River dykes (BRD), herein dated by U–Pb baddeleyite and zircon methods at 1268.5 ± 1.5 Ma and 1264.6 ± 1.2 Ma. The BRD share similar geochemical and Nd-isotopic characteristics with the Mackenzie dykes and the coeval Coppermine River basalts, and are regarded as products of the same plume. Hydrothermal activity at ∼1270 Ma in breccia of the Nor mineral occurrence, northwest of the BRD, was probably generated by BRD at depth. The BRD generally strike northwest, approximately 90° to the orientation predicted by the radiating dyke model. This difference may be explained by an anomalous local stress field at the time of BRD emplacement, or reorientation of BRD during subsequent events of Cordilleran deformation. Using the Nor mineral occurrence as the westernmost locale of Mackenzie dyking, and assuming a model of uniform dyke radiation, the arc of the Mackenzie swarm is at least 50° greater than previously recognized, bringing the total arc of dyke radiation to >150°.

Keywords: Dyke; Cordillera; Yukon; Mantle plume; Proterozoic; Mackenzie

1. Introduction

Radiating mafic dyke swarms represent brief but voluminous injections of mantle-derived magma into dilatant crust. Many are linked to continental flood basalts, and are attributed to the arrival of a mantle plume head (Campbell and Griffiths, 1990; Ernst et al., 1995a) or a magmatic outburst above a steady-state hotspot (Johnston and Thorkelson, 2000). Anomalously high mantle temperatures and huge volumes of uprising melt, leading to updoming of the crust and outflow of magma from the hot-spot, are the most likely causes of the radiating dyke patterns.

The Mackenzie dyke swarm is the largest swarm on Earth, radiating with an arc of ∼100° across most of the Canadian Shield (Fahrig and West, 1986; Ernst

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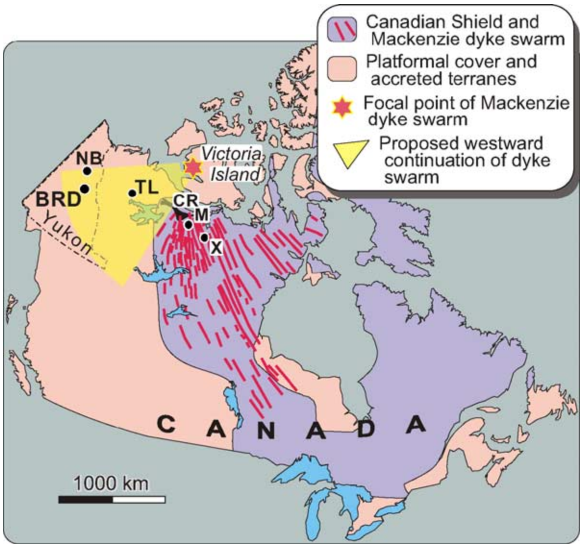

Fig. 1. Map of Canada showing main features of the Mackenzie igneous event (after Fahrig and West, 1986; Hoffman, 1989; LeCheminant and Heaman, 1989; Ernst and Baragar, 1992). Yellow fan-shaped area represents proposed, mainly subsurface continuation of the Mackenzie dyke swarm into western Canada. Canadian Shield is mainly Middle Proterozoic to Archean; Platformal cover and accreted terranes is mainly Late Proterozoic and Phanerozoic. BRD: Bear River dykes; CR: Coppermine River Group basalts (black area); M: Muskox intrusion; NB: Nor breccia; TL: Tweed Lake basalts (subsurface); X: xenolith locality of Davis (1997).

et al., 1995a). It fans from a focal point on Victoria Island, near the Arctic coast, and extends eastward for ∼1400 km to the eastern Canadian Arctic, and southward for ∼2400 km into northwestern Ontario (Fig. 1). The dykes were emplaced in the same event that led to eruption of the Coppermine River flood basalts, the Tweed Lake basalts (preserved in the subsurface ∼250 km to the west; Sevigny et al., 1991), and emplacement of the Muskox layered intrusion (Baragar, 1969; Dostal et al., 1983; LeCheminant and Heaman, 1989; Francis, 1994). All of the igneous activity is considered to have occurred within a few million years of 1267 Ma (LeCheminant and Heaman, 1989). The large volume of magma, the radial flow pattern, and the inferred short time span of emplacement point to a mantle plume origin (LeCheminant and Heaman, 1989; Ernst and Baragar, 1992; Ernst et al., 1995a).

In this paper we examine the field relations, age, and geochemistry of the mafic Bear River dykes (BRD) which are exposed in an Early Proterozoic inlier of northern Yukon (Fig. 2). The inlier is separated from the Mackenzie dyke swarm by a 700-km wide swath of Middle Proterozoic to Phanerozoic sedimentary cover (Fig. 1), nearly all of which was deposited after ca. 1.27 Ga Mackenzie magmatism. We characterize the BRD using petrography, major and trace element geochemistry, Nd isotopes, and U–Pb geochronology, and propose a correlation with the Mackenzie dykes. Relations with the 1.27 Ga Nor hydrothermal breccia of northeastern Yukon are explored, and an increase in the estimated size of the Mackenzie dyke swarm is corroborated.

2. Bear River dykes

The BRD occur in a Proterozoic inlier in northern Yukon (Fig. 2), where they crosscut Early Proterozoic sedimentary strata of the Wernecke Supergroup (Thorkelson, 2000). One of the dykes was mapped by Blusson (1974), who assigned it a provisional Cretaceous age. Subsequent work including comprehensive studies by Abbott (1997) and Thorkelson (2000) have not revealed any Cretaceous-aged dykes in the Proterozoic inliers, and the assignment by Blusson (1974) is considered to be incorrect. Thorkelson (2000) and Schwab and Thorkelson (2001) identified about 15 dykes in at least seven localities (simplified in Fig. 2). Some of the dykes are present as individual intrusions, whereas others occur in swarms of up to 8 dykes. The dykes are 5–15 m thick and up to 5 km long (Figs. 2 and 3). They are subvertical, and strike mainly to the northwest. They consist of medium to fine grained diorite and gabbro, variably altered to lower greenschist metamorphic grades. Mineralogy is dominated by subequal amounts of clinopyroxene and plagioclase, with accessory magnetite, apatite, interstitial granophyre, and local late-stage biotite. Plagioclase is variably altered to clay, mica, calcite, chlorite and epidote; clinopyroxene is commonly pseudomorphed by chlorite. One of the dykes (dyke 1 in Fig. 2) is crosscut by undated hematitic veins and is metasomatically enriched in Cu and U (Schwab and Thorkelson, 2001). Another dyke (dyke 3) contains a weak penetrative foliation of unknown age and origin. Despite these minor variations in fabric and alteration, the BRD have consistent compositions, orientations and dimensions, and are considered part of a single magmatic event.

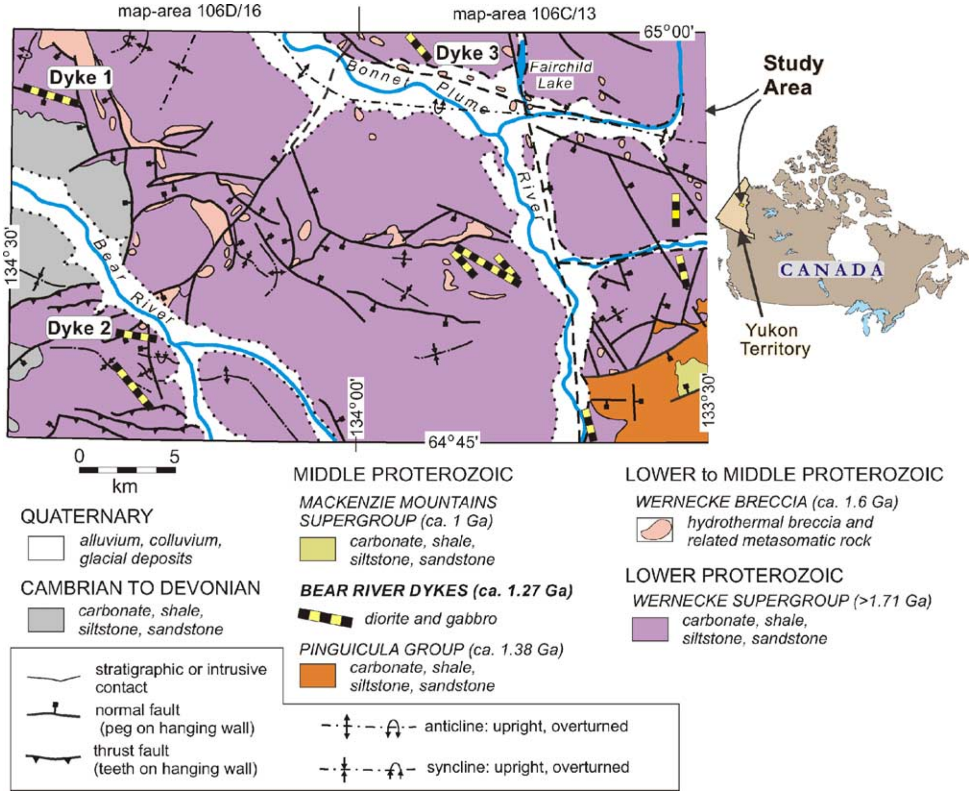

Fig. 2. Simplified geological map of the study area (map areas 106 D/16, 106 C/13) showing locations of the Bear River dykes (simplified from Thorkelson, 2000). Dykes 1, 2, and 3 are dated by U–Pb geochronology (this paper).

Fig. 3. Bear River dyke crosscutting steeply dipping, previously folded dolostone of the Wernecke Supergroup. Dyke is located in southwestern corner of Fig. 2, south of dyke 2. Dolostone in contact with dyke has been metamorphosed to white-weathering calcite marble. Dyke is approximately 15 m wide and dips away from viewer at 75°.

Details on age and geochemistry, provided below, support this suggestion.

2.1. Regional geology

The BRD intruded a complex Proterozoic orogen that developed over a preceding interval of at least 450 million years As shown in Figs. 2 and 4, the oldest exposed rocks are the Wernecke Supergroup, a 13-km thick succession of fine-grained basinal siliciclastic rocks and platformal carbonates which were deposited on a basement of unknown age and affinity. The Wernecke Supergroup was deformed and metamorphosed during three phases of deformation, collectively termed the Racklan orogeny (Brideau et al., 2002), intruded by the 1.71 Ga rift-related Bonnet Plume River Intrusions (Thorkelson et al., 2001a), and crosscut by the 1.60 Ga Wernecke Breccias, of hydrothermal origin (Thorkelson et al., 2001b). The

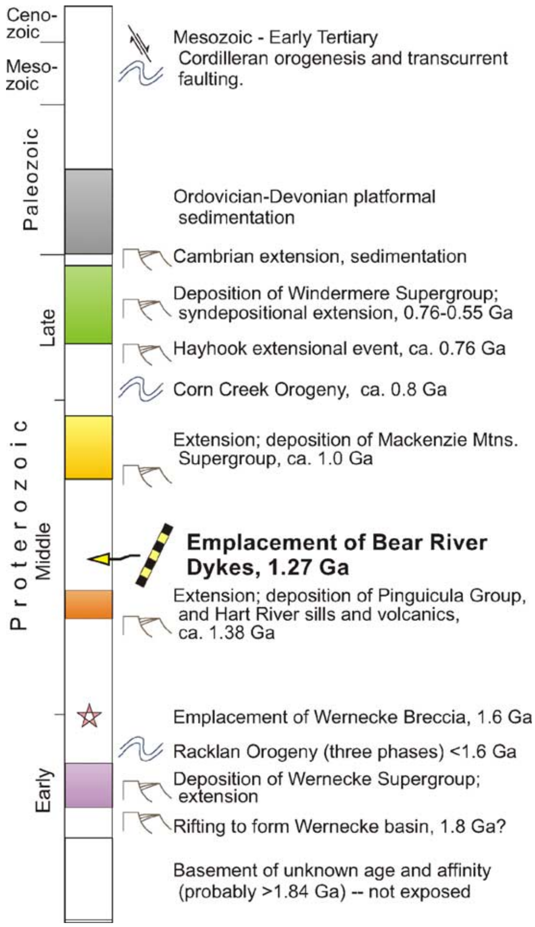

Fig. 4. Time-stratigraphic column for the study area (Fig. 2) and adjacent areas, showing Bear River dykes in context of geological evolution of northern Yukon (modified from Thorkelson, 2000). Note at least five events of extension, two events of contraction, and one event of strike-slip deformation following emplacement of Bear River dykes.

region was subsequently extended, intruded by the Hart River mafic sills, and overlain by the mafic Hart River volcanics and the clastic-carbonate Pinguicula Group. As indicated in Fig. 4, the BRD intruded the area following Pinguicula Group sedimentation, although all of the BRD identified so far are hosted by the Wernecke Supergroup, and none have yet been observed to crosscut the Pinguicula Group.

After emplacement of the BRD, the region underwent at least four events of extension leading to deposition of the Mackenzie Mountains Supergroup (Aitken et al., 1982; Norris and Dyke, 1997), the Windermere Supergroup (Eisbacher, 1981), and Paleozoic platformal strata (Cecile, 1982). West-verging folding and thrust-faulting occurred in the Middle Proterozoic between deposition of the Pinguicula and Mackenzie Mountains successions (Eisbacher, 1981; Thorkelson, 2000). North-verging folding and thrusting, and dextral strike-slip faulting (Cecile, 1984; Abbott, 1997), occurred during Late Cretaceous to Paleogene orogenesis, which affected most of the North American Cordillera (Norris, 1997). Thus, the crust in the study area (Fig. 2) remained mobile for nearly 1.3 billion years after intrusion by the BRD.

Selected folds and faults are illustrated in Fig. 2. Because of the reconnaissance nature of the geological mapping and investigations in the area thus far, many of these structures are poorly constrained in age, and their origins are uncertain. Some older structures have been reactivated or superposed by subsequent events (Thorkelson, 2000), complicating the structural evolution. Nevertheless, the history depicted in Fig. 4 is supported by observations in key localities and provides a rationale for the possible structural reorientation of the BRD, as considered below.

2.2. Geochemistry

Geochemical analyses were obtained by X-ray fluorescence and inductively coupled mass spectrometry for seven samples of the BRD from five of the six dyke localities. Nd isotopic data were obtained for four of the least-altered samples at the University of Alberta. All of the data are reported in Table 1. Mobility of the large-ion lithophile elements (LILE), such as K, Rb, Cs, U and Th, is evident from their large, incoherent variations in abundance. In contrast, the rare earth elements (REE) and high field strength elements (HFSE) show less scatter and yield consistent patterns on multi-element diagrams (Fig. 5).

The BRD are evolved quartz tholeiites with magnesium numbers ranging from 42 to 59 (Table 1). The range of Cr concentrations (109–409 ppm), and a trend toward iron enrichment (Schwab, 2001), also indicate evolved rather than primary character. Subalkaline composition is inferred by the ubiquity of interstitial quartz. Mantle-normalized trace element profiles (Fig. 5) and chondrite-normalized REE patterns (Schwab, 2001) resemble those of continental flood basalts (e.g., Saunders et al., 1992). BRD patterns

Table 1
Major and trace element geochemistry, and Nd isotope composition of the BRD Major oxide, trace element and Nd isotope data for the Bear River dykes

Sample ID		TOA-96-6-7-2B	DT-96-1-1-1B	DT-92-11-1B	CW-93-19-1B	DT-93-72-1B	DS-00-2-2-8B	DS-00-3-2-5B
Location	NTS	106D/16	106C/13	106D/16	106C/13	106C/13	106D/16	106D/16
	UTM E	527600	563700	532000	556900	569600	528500	527500
	UTM N	7203300	7181700	7186800	7192900	7196900	7203000	7203500
Major oxides (wt.%)	SiO_2	54.34	51.80	53.53	48.81	51.92	55.59	55.70
	TiO_2	1.05	2.29	1.09	1.30	2.28	1.18	1.15
	Al_2O_3	14.23	11.97	13.44	12.71	13.08	14.00	14.52
	FeO^*	10.7	14.9	11.4	17.0	14.0	12.12	10.54
	MnO	0.20	0.31	0.15	0.16	0.23	0.21	0.20
	MgO	7.20	6.93	8.05	6.19	5.89	6.67	5.83
	CaO	7.83	6.73	6.74	9.48	8.34	6.50	7.85
	Na_2O	2.17	3.86	2.21	1.61	2.91	2.23	2.55
	K_2O	2.16	0.99	3.24	2.65	1.11	1.36	1.52
	P_2O_5	0.11	0.21	0.12	0.14	0.23	0.13	0.12
Trace elements (ppm)	Ba	386.75	238.33	267.85	430.71	331.84	239.87	322.52
	Rb	76.02	24.24	164.65	35.31	25.19	32.73	41.41
	Cs	1.89	0.86	5.72	0.90	2.57	N.D.	N.D.
	Th	4.08	1.75	2.81	4.36	1.81	4.26	6.26
	Ta	0.47	0.54	0.52	1.09	0.59	N.D.	N.D.
	U	0.87	0.49	0.68	1.93	1.04	2.81	3.07
	K	17952	8258	26921	21989	9222	11271	12656
	Nb	7.70	9.53	6.44	15.01	10.30	8.00	9.13
	La	15.8	11.8	13.0	21.9	12.5	N.D.	N.D.
	Ce	32.7	29.8	27.3	49.3	31.6	53.4	24.2
	Pb	5.00	−5.00	21.69	15.73	22.94	7.36	−3.59
	Sr	145.5	122.6	102.9	294.5	256.0	93.7	157.5
	Pr	3.34	3.40	3.25	5.95	4.42	N.D.	N.D.
	Nd	15.34	17.37	13.79	22.26	20.19	N.D.	N.D.
	P	494	913	503	592	996	N.D.	N.D.
	Zr	100	122	73	113	137	105	118
	Sm	3.47	4.85	3.15	4.77	5.28	N.D.	N.D.
	Hf	2.94	3.77	2.22	3.55	3.82	N.D.	N.D.
	Eu	1.09	1.86	0.96	1.43	1.84	N.D.	N.D.
	Gd	3.80	5.10	3.50	4.78	6.18	N.D.	N.D.
	Ti	6303	13761	6550	7802	13706	5934	5537
	Tb	0.66	0.80	0.50	0.71	0.86	N.D.	N.D.
	Dy	3.69	4.84	3.36	4.26	5.23	N.D.	N.D.
	Ho	0.72	0.89	0.62	0.84	0.97	N.D.	N.D.
	Er	2.20	2.53	1.86	2.20	2.58	N.D.	N.D.
	Tm	0.29	0.35	0.25	0.31	0.37	N.D.	N.D.
	Y	21.0	25.0	15.5	19.2	23.4	17.7	21.0
	Yb	2.15	2.26	1.63	2.04	2.05	N.D.	N.D.
	Lu	0.32	0.32	0.24	0.29	0.29	N.D.	N.D.
	V	263	369	284	238	419	292	285
	Sc	N.D.	N.D.	31.3	25.8	28.2	29.0	25.6
	Cr	357	129	479	208	109	331	289
	Co	69.8	69.9	41.5	33.8	45.6	N.D.	N.D.
	Ni	162	678	151	82	78	84	87
	Ga	N.D.	N.D.	N.D.	N.D.	N.D.	16.22	18.76
	Cu	122	354	142	70	371	167	101
	Zn	50.0	267.0	102.0	120.6	156.4	30.4	15.7
	Be	N.D.	N.D.	0.54	0.54	1.50	N.D.	N.D.
	Li	N.D.	N.D.	32.06	27.33	17.36	N.D.	N.D.

Table 1 (*Continued*)

Sample ID		TOA-96-6-7-2B	DT-96-1-1-1B	DT-92-11-1B	CW-93-19-1B	DT-93-72-1B	DS-00-2-2-8B	DS-00-3-2-5B
Location	NTS	106D/16	106C/13	106D/16	106C/13	106C/13	106D/16	106D/16
	UTM E	527600	563700	532000	556900	569600	528500	527500
	UTM N	7203300	7181700	7186800	7192900	7196900	7203000	7203500
	Tl	0.35	1.05	0.62	0.30	0.19	N.D.	N.D.
	Bi	0.81	0.17	0.26	0.05	0.05	N.D.	N.D.
	Ag	−0.50	−0.50	0.66	1.44	0.95	N.D.	N.D.
	Sn	0.80	1.10	35.58	1.95	1.66	N.D.	N.D.
	Sb	2.09	2.58	7.39	2.40	1.19	N.D.	N.D.
	Mn	1517	2435	1139	1203	1770	N.D.	N.D.
T_{DM}^{Nd} (Ga)		2.72	N.D.	2.69	2.09	2.13	N.D.	N.D.
$\varepsilon_{Nd(T)}$ ($T = 1.27$ Ga)		−7.2	N.D.	−5.9	−1.9	1.4	N.D.	N.D.
Mg#		58	49	59	42	46	53	53

Notes. Major oxides reported on a volatile-free basis; N.D.: analysis not done; FeO*: total Fe as FeO; Mg#: $[100Mg^{2+}/(Mg^{2+} + Fe^{2+})]_{atomic}$.

show enrichments of LILE and the light REE, relative to the heavy REE, with chondrite-normalized Ce/Yb ranging from 3.4 to 6.1. Relative depletions of Ta, Nb, Sr and P, and a slight depletion in Ti, are also persistent features. La/Nb ratios range from 1 to 2, similar to those of modern mid-ocean ridge basalts (Schwab, 2001).

Modeling of closed-system crystal fractionation in the BRD was unsuccessful using mass-balance of major oxides and partitioning of trace elements. Fractionation of augite is suggested by the covariation of Cr and Sc, which are relatively insensitive to crustal assimilation, but modeling of major oxides could not reproduce this finding (Schwab, 2001). Fractionation of plagioclase, the other main mineral phase, is not indicated by either trace or major element systematics and is contra-indicated by a negative correlation between Sr and magnesium number (Schwab, 2001). The overall chemical variability of the BRD requires open-system processes such as assimilation, mixing of magma, and/or variable mantle sources.

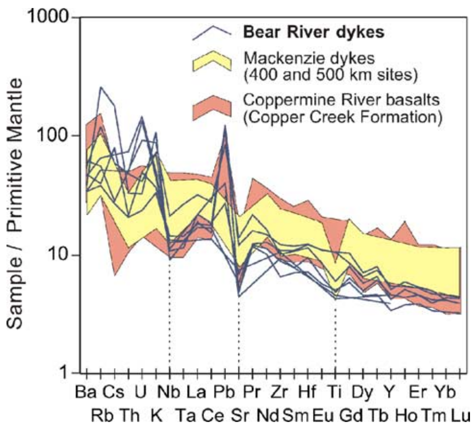

Fig. 5. Trace element profiles of Bear River dykes, showing similarities to Mackenzie dykes at distances of 400–500 km from focal point (Fig. 1) and the lower lavas of the Coppermine River basalts (Copper Creek Formation). Data sources: Bear River dykes: Thorkelson (2000), Schwab and Thorkelson (2001) (Table 1); Coppermine River basalts: Dupuy and Dostal (1992), Baragar et al. (1996), Griselin et al. (1997); Mackenzie dykes: Gibson et al. (1987), Baragar et al. (1996). Normalizing values from Sun and McDonough (1989).

The possibility of crustal contamination and/or variable source characteristics is highlighted by values of $\varepsilon_{Nd(T)}$ ($T = 1.27$ Ga) which range from +1.4 to −7.2 in four of the BRD (Table 1). If crustal contamination were evoked, this range in values could represent variable degrees of assimilation of old, evolved continental crust by more juvenile, mantle-derived magmas. However, assimilation of evolved crust cannot have been the main process involved, because the $\varepsilon_{Nd(T)}$ values correlate negatively with Mg-number, Cr, and Ni, and positively with incompatible elements such as the REE (example using Gd and Cr shown in Fig. 6). If crustal contamination were prevalent, these trends would imply that the geochemically more primitive dykes (with higher Mg-numbers) had undergone more crustal contamination than the more evolved dykes, the reverse of that expected. To account for these trends, derivation from two mantle sources with contrasting compositions is suggested. One of the mantle sources may have been the asthenosphere, with positive ε_{Nd} and a

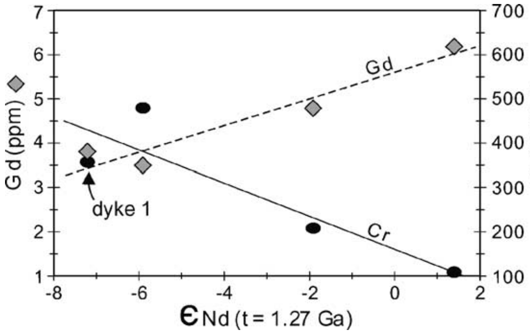

Fig. 6. Variations in Cr and Gd with increasing $\varepsilon_{Nd(T)}$, showing regression lines (data from Bear River dykes in study area, Fig. 2, Table 1).

depleted geochemical composition, and the other may have been enriched peridotite with negative ε_{Nd}, from either the plume or lithospheric mantle reservoirs.

2.3. U–Pb geochronology

U–Pb geochronology was carried out on three samples of fine- to medium-grained BRD. Two of the dykes yielded baddeleyite and one yielded zircon. The samples were dated using conventional U–Pb methods at the Pacific Centre for Isotopic and Geochemical Research at the University of British Columbia. The methodology for sample selection, dissolution, geochemical preparation and mass spectrometry is described in Mortensen et al. (1995). Procedural blanks were 2–5 pg for Pb and 1 pg for U. U–Pb data is presented in Table 2 and are shown on conventional U–Pb concordia plots in Fig. 7. Errors are given at the 2σ level.

Baddeleyite recovered from dykes 1 and 2 consisted of very fine-grained (<62 μm maximum dimension), pale to medium brown needles and tabular cleavage fragments. The grains were too fine to permit abrasion prior to dissolution; however they appeared to be inclusion-free and unaltered, with no evidence of secondary alteration to zircon. An initial 20-kg sample collected in 1996 from dyke 1 (Fig. 2) yielded enough baddeleyite for four fractions, and resampling of the same outcrop in 2000 provided enough for an additional two analyses. Despite the apparently unaltered nature of the baddeleyite, all fractions analyzed show evidence of Pb-loss. The six baddeleyite fractions from this dyke outcrop (fractions A1–A6) range from 1.4 to 39.3% discordant (Fig. 7a), and a Davis-type regression yields calculated upper and lower intercepts of 1268.5 ± 1.5 Ma and −10 Ma, with a probability of fit of 59.6%. The upper intercept is interpreted to give the crystallization age of the sample. The near-zero lower intercept is typical for discordant baddeleyite data arrays, as observed by Heaman (1997) and is interpreted as recent Pb-loss. Two fractions of baddeleyite (B1–B2) from dyke 2 (Fig. 2) are both discordant (Fig. 7a), however a regression through the two points yields a similar, although imprecise, upper intercept age of 1265 + 20/−11 Ma and a lower intercept of ~221 Ma, indicating that this dyke is similar in age to that of dyke 1.

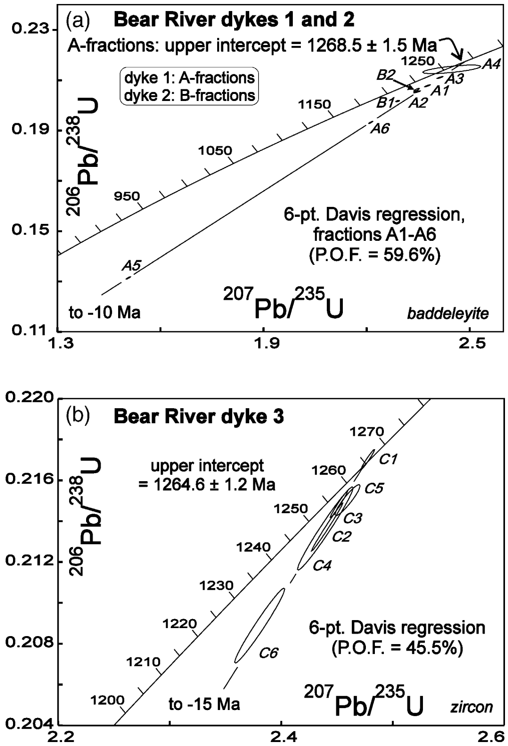

Fig. 7. U–Pb concordia diagram for samples from the Bear River dykes. (a) Dykes 1 and 2, where fractions A1 to A6 are baddeleyite from dyke 1, and fractions B1 and B2 are baddeleyite from dyke 2. (b) Dyke 3, zircon fractions C1 to C6. Errors are at the 2σ level. See Fig. 2 for locations.

Zircon rather than baddeleyite was recovered from dyke 3 (Fig. 2). The zircon consists of clear, pale brown, stubby to elongate prismatic grains with vague internal zoning and no visible cores. A total of six

Table 2
U–Pb isotope data on zircon and baddeleyite from dykes 1 to 3 of the BRD U–Pb analytical data

Sample description[a]	Weight (mg)	U content (ppm)	Pb[b] content (ppm)	$^{206}Pb/^{204}Pb$ (meas.)[c]	Total common Pb (pg)	%^{208}Pb[c]	$^{206}Pb/^{238}U$[d] (±1%)	$^{207}Pb/^{235}U$[d] (±1%)	$^{207}Pb/^{206}Pb$[d] (±1%)	$^{206}Pb/^{238}U$ age[e] (Ma; ±2%)	$^{207}Pb/^{206}Pb$ age[e] (Ma; ±2%)
Dyke 1(TOA-96-6-7-1 and DS-00-3-2-5C) [baddeleyite]											
A1	0.02	1397	275	3762	97	1.7	0.20779 (0.09)	2.3774 (0.17)	0.08298 (0.09)	1217.0 (2.0)	1268.8 (3.4)
A2	0.02	1389	271	1520	218	1.9	0.20540 (0.10)	2.3471 (0.20)	0.08288 (0.12)	1204.2 (2.2)	1266.3 (4.8)
A3	0.018	1177	234	2917	96	1.5	0.21024 (0.10)	2.4042 (0.17)	0.08294 (0.09)	1230.1 (2.1)	1267.8 (3.6)
A4	0.005	116	24	97	97	3.5	0.21431 (0.39)	2.4480 (1.72)	0.08284 (1.48)	1251.8 (9.0)	1265.5 (57.8)
A5	0.008	1768	219	2470	44	1.3	0.13140 (0.18)	1.5063 (0.20)	0.08314 (0.07)	795.8 (2.7)	1272.6 (2.8)
A6	0.005	4022	733	4451	50	1.3	0.19322 (0.11)	2.2121 (0.13)	0.08303 (0.05)	1138.8 (2.3)	1269.9 (1.8)
Dyke 2 (DT-95-1) [baddeleyite]											
B1	0.005	685	132	1263	35	2.9	0.20170 (0.07)	2.2905 (0.12)	0.08236 (0.09)	1184.4 (1.5)	1254.1 (3.4)
B2	0.005	1500	297	1537	63	1.7	0.20618 (1.5)	2.3455 (0.17)	0.08251 (0.08)	1208.4 (3.2)	1257.6 (3.3)
Dyke 3 (JL-02-01) [zircon]											
C1	0.02	1849	615	18790	26	39	0.21704 (0.11)	2.4778 (0.12)	0.08280 (0.03)	1266.2 (2.6)	1264.5 (1.1)
C2	0.025	2309	775	6872	110	40.5	0.21372 (0.28)	2.4412 (0.29)	0.08284 (0.04)	1248.6 (6.3)	1265.5 (1.4)
C3	0.019	2562	814	20420	31	36.9	0.21482 (0.15)	2.4516 (0.16)	0.08277 (0.03)	1254.4 (3.2)	1263.8 (1.1)
C4	0.029	3134	1025	8490	142	39	0.21364 (0.48)	2.4395 (0.51)	0.08282 (0.11)	1248.2 (10.9)	1264.9 (4.3)
C5	0.027	2170	657	12040	65	33.7	0.21503 (0.18)	2.4594 (0.23)	0.08295 (0.09)	1255.6 (4.0)	1268.1 (3.7)
C6	0.017	3707	1087	9588	85	33.5	0.20889 (0.44)	2.3807 (0.47)	0.08266 (0.10)	1222.9 (9.8)	1261.2 (4.1)

[a] N2, M2 = non-magnetic, magnetic at indicated number of degrees side slope on Frantz isodynamic magnetic separator; grain size given in microns.
[b] Radiogenic Pb; corrected for blank, initial common Pb (from Stacey and Kramer, 1975), and spike.
[c] Corrected for spike and fractionation.
[d] Corrected for blank Pb and U, and common Pb.
[e] Decay constants from Steiger and Jäger (1977); errors assigned using the numerical error propagation method of Roddick (1987).

fractions were analyzed (Table 2). The analyses form a linear array (Fig. 7b), and a Davis-type regression of all six analyses yields calculated upper and lower intercept ages of 1264.6 ± 1.2 Ma and 15 Ma, respectively, with a probability of fit of 45.5%. The upper intercept gives the crystallization age of the diorite body and the lower intercept is consistent with mainly recent Pb-loss from the zircons.

3. Correlations with the Mackenzie igneous event

Identical ages (within analytical uncertainty) of the BRD and the Mackenzie dyke swarm make inclusion of the BRD in the Mackenzie igneous event highly attractive. U–Pb ages of 1267 ± 2 Ma for 4 Mackenzie dykes, and 1270 ± 4 Ma for two Muskox intrusion samples were reported by LeCheminant and Heaman (1989), and supplant previous, younger K–Ar and Rb–Sr ages for the Mackenzie event. As suggested by the tight age constraints, and supported by the radiating pattern and flow paths of the dyke swarm (Ernst and Baragar, 1992), Mackenzie-age magmatism is considered to have taken place in a very short interval, perhaps no more than 2 or 3 million years (LeCheminant and Heaman, 1989). The enormity of the Mackenzie event clearly points to the BRD as disparate members of the Mackenzie swarm.

Petrography and geochemistry support the inclusion of the BRD in the Mackenzie igneous event. The Coppermine River basalts generally consist of microphenocrysts of plagioclase and augite with accessory Fe-oxides in an interstitial groundmass of potassium feldspar and quartz (Baragar et al., 1996). The Mackenzie dykes have a similar mineralogy consisting essentially of plagioclase, pyroxenes and Fe–Ti oxides, with quartz, potassium feldspar ± biotite and amphibole in the interstices (Baragar et al., 1996). These assemblages are similar to the plagioclase-augite-magnetite-granophyre ± biotite mineralogy of the BRD.

Geochemically, the Coppermine River basalts and Mackenzie dykes are mostly quartz tholeiites (Dostal et al., 1983; Griselin et al., 1997), and share similar trace element profiles with the BRD (Fig. 5). The trace element patterns of the BRD are most similar to those of the Mackenzie dykes sampled between 400 and 500 km from the proposed focal point of the swarm (the BRD are located ∼1100 km from the focal point), and basalts in the lower part of the Coppermine River Group (Copper Creek Formation; Fig. 1). Striking Pb-anomalies (positive) and Sr-anomalies (negative) are shared by the BRD and the Copper Creek Formation. Initial ε_{Nd} values of the Mackenzie dykes and Coppermine lavas range from +4.9 to –9.8 (Dupuy and Dostal, 1992; Dudás and Peterson, 1992; Griselin et al., 1997), encompassing those of the BRD (+1.2 to −7.2). The range in $\varepsilon_{Nd(T)}$ of the BRD is similar to the +1.7 to −5.5 range of the lower Coppermine River basalts (Griselin et al., 1997).

Crustal assimilation as a significant influence on the composition of the Coppermine lavas and Mackenzie dykes has been proposed on the basis of major and trace element abundances and Nd isotope geochemistry (Dudás and Peterson, 1992; Dupuy and Dostal, 1992; Dostal et al., 1983; Francis, 1994; Baragar et al., 1996; Griselin et al., 1997). Dykes and flows with strongly negative values of $\varepsilon_{Nd(T)}$ are considered to have undergone significant crustal contamination, a process evident in an individual Mackenzie dyke with $\varepsilon_{Nd(T)}$ values ranging from −0.6 in the interior of the dyke to −9.8 at the margin (Dudás and Peterson, 1992). However, details on the assimilation process remain uncertain because the more primitive igneous units commonly appear to have undergone a greater degree of crustal contamination than associated, less primitive ones (Dupuy and Dostal, 1992; Francis, 1994; Griselin et al., 1997). This notable trend of a decreasing "crustal" signature in more fractionated rocks is also evident in the BRD (Fig. 6).

4. Regional significance

The BRD, as correlative intrusions of the Mackenzie dyke swarm, provide key evidence that the Mackenzie igneous event was much more widespread than previously recognized. The BRD are the only dykes west of the Canadian Shield that have been correlated with the Mackenzie swarm, although numerous additional dykes may exist in the subsurface of western Canada, lying undetected beneath the extensive post-1.27 Ga sedimentary cover. In northern Yukon, the Proterozoic inliers which host the BRD provide a rare opportunity to evaluate the possible westward continuation of Mackenzie-age igneous rocks.

If the BRD belong to the Mackenzie dyke swarm they should strike northeast, according to the model of radiating dyke propagation from a focal point on Victoria Island (Ernst and Baragar, 1992; Fig. 1). None of the BRD strike in the expected direction, and instead strike north to northwest. However, this lack of concordance does not preclude correlation with the Mackenzie swarm. Post-Mackenzie tectonic events in the Cordilleran orogen, ranging in age from Mesoproterozoic to Tertiary, comprise contractional, extensional and strike-slip deformation of different ages (Fig. 4). Folds and faults representing all of these deformational styles are recorded in the vicinity of the BRD (Thorkelson, 2000). These events may have reoriented the BRD from positions that were originally concordant with the expected trend of the Mackenzie dyke swarm.

Alternatively, the BRD may not have been reoriented, but instead may have been influenced by a local stress regime whose minimum compressive stress was oriented northeast rather than the northwest direction predicted by the radiating dyke model. In this hypothesis, magma would have flowed outward from the vicinity of the focal point, following a radially controlled pattern of tension. Upon reaching the area of northern Yukon, the magma would have encountered a different stress regime and begun flowing northwestward. Two explanations for such a local stress field in the Yukon area are tendered. In one, the stress field may have belonged to a concentric pattern of tension (relative to the focal point) as suggested by some dyke-like features on Mars (Montesi, 2001). The absence of a concentric pattern of Mackenzie dykes in other parts of Canada, however, casts doubt on this idea. In the other explanation, the local stress field may have been related to recurring episodes of extensional tectonism and basin formation in the Yukon region (Fig. 4). In general support of the local stress-field option, we note that at least one Mackenzie dyke, located ∼1000 km away from the focal point of the swarm, has a trend perpendicular to the main trend of the swarm (Pehrsson et al., 1993). These authors speculated that the dyke orientation may have been controlled by the similar-trending Great Slave Lake shear zone in the Slave craton or that the trend of the dyke may have been influenced by the local stress field at the time of emplacement. Similarly, Ernst et al. (1995b) suggested that the transition from a radiating pattern to one of parallel dykes at the southeastern extremity of swarm (Fig. 1) occurred because the dykes in that area extended beyond the region dominated by plume-generated updoming.

Evidence for additional Mackenzie-age magmatism in the subsurface of northern Yukon is implied by a 1.27 ± 0.04 Ga U–Pb monazite date from a hematitic breccia zone at the Nor mineral occurrence (Parrish and Bell, 1987). This breccia zone, which is exposed in a small inlier 150 km north-northwest of the BRD (Fig. 1), has been correlated with the 1.60 Ga Wernecke Breccias (Figs. 2 and 4). The ca. 1.27 Ga monazite age was originally regarded as the age of breccia emplacement (Parrish and Bell, 1987) but is now regarded as a product of secondary hydrothermal fluid activity (Thorkelson et al., 2001b). This Mackenzie-age hydrothermal pulse may have been driven by crustal heating in response to local BRD magmatism. How far the BRD extend beyond their area of exposure, hidden beneath sedimentary cover, is unknown, but their presence in the vicinity of the Nor breccia is plausible. An earlier hydrothermal fluid event in other Wernecke Breccias, recorded by ca. 1.38 Ga U–Pb ages on rutile, has been linked to the nearby 1.38 Ga Hart River sills (Abbott, 1997; Thorkelson et al., 2001b). The hypothesis of hydrothermal activity related to Mackenzie-age magmatism is similar to the suggestion of Davis (1997) who explained 1.27 Ga thermal metamorphism in xenoliths from the subsurface of the western Canadian Shield (Fig. 1) by the presence of voluminous Mackenzie-age intrusions at depth.

The Tweed Lake basalts in the western Northwest Territories (Fig. 1) were previously regarded as the westernmost expression of Mackenzie magmatism. However, if the BRD are indeed Mackenzie dykes, as we suggest, then the extent of the Mackenzie swarm, in terms of both the angle of dyke radiation and the areal extent of magmatism, is significantly larger than previously recognized. Furthermore, if hydrothermal mineralization in the Nor breccia is accepted as a proxy for nearby Mackenzie-age intrusions, then the size of the Mackenzie igneous province is greater still. By using the Nor breccia as the western limit of Mackenzie dyking, the arc of the known swarm increases from 100° to a remarkable 150°.

The identification of disparate members of the Mackenzie dyke swarm in Yukon is important for two main reasons. First, it broadly supports the prediction by Ernst and Baragar (1992) and Ernst et al. (1995a) that the Mackenzie swarm extended outward from the Canadian Arctic across much of ancestral North America (Laurentia). Second, the extension of the dyke swarm into western Canada means that other continents which may have lain alongside western Laurentia in Middle Proterozoic or later times, and were subsequently rifted away, may contain distal members of the Mackenzie swarm. Candidates for formerly attached continents include Australia (Bell and Jefferson, 1987; Moores, 1991; Ross et al., 1992), Siberia (Sears and Price, 2000), and South China (Li et al., 1995). If Mackenzie-aged dykes were to be identified in one or more of these regions, the orientation, composition, and exact age of the dykes would serve as valuable tools in resolving the controversy over paleocontinental reconstructions and understanding magmatic processes of giant dyke swarms.

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