Eocene volcanism above a depleted mantle slab window in southern Alaska

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ABSTRACT

The Caribou Creek volcanic field lies along the continent-side edge of forearc basin rocks in south-central Alaska and consists of over 1000 m of shallow-dipping basalt and andesite lavas with minor mafic pyroclastic deposits. Dacite and rhyolite lavas along with shallow intrusions form dome complexes with associated pyroclastic deposits that overlie and crosscut the basalt and intermediate lavas. The basalts are tholeiitic and strongly depleted in the light rare earth elements (La/Yb = 0.18–1.5), with concentrations of high field strength elements (e.g., Zr, Hf, Ti, Y) similar to mid-ocean-ridge basalt and with variable enrichment in fluid-mobile elements (e.g., Cs, Ba, and Pb). Intermediate and felsic rocks show enrichment in the rare earth elements and fluid-mobile elements plus Rb and K, but retain low La/Yb ratios (0.48–3.6). A few andesite and dacite samples are strongly depleted in the heavy rare earth elements and are geochemically similar to adakites (e.g., Sr/Y up to 52). Ten $^{40}Ar/^{39}Ar$ ages for the Caribou Creek volcanic rocks range from 49.4 ± 2.2 to 35.6 ± 0.2 Ma. An adakite-like tuff beneath the other volcanic rocks yields an age of 59.0 ± 0.4 Ma.

Caribou Creek basalts were derived from mid-ocean-ridge–like depleted mantle that was emplaced beneath the southern margin of Alaska through a slab window following spreading ridge subduction. Caribou Creek volcanism was coeval with oblique subduction, oroclinal bending, and right-lateral strike-slip faulting in south-central Alaska, all of which could have induced crustal extension to allow adiabatic melting of the depleted mantle reservoir to form basaltic magmas. The basalts then evolved by fractional crystallization with moderate to high degrees of assimilation of Jurassic arc basement rocks to form the intermediate and felsic magmas. Enrichment of the basaltic parent magmas in fluid-mobile elements occurred by contamination from the Jurassic arc rocks and/or by contamination with metasomatic mantle remnant from preceding subduction. High heat flow through the slab window induced partial melting of garnet-bearing mafic parts of the Jurassic arc basement to form the adakite-like rocks. The Caribou Creek volcanic rocks demonstrate that slab windows can directly influence magmatism inboard of accretionary prism and forearc basin settings given a suitable deformation regime (e.g., crustal extension) and that the influence of a slab window on continental margin magmatism can be long-lived (>20 m.y.).

Keywords: Eocene, volcanism, Alaska, Talkeetna Mountains, slab window, adakite.

INTRODUCTION

Slab windows are gaps in subducted plates that allow the upwelling of asthenospheric mantle (e.g., Thorkelson, 1996) and have been documented in cases of spreading ridge–trench interactions (Dickinson and Snyder, 1979; Thorkelson and Taylor, 1989; Cole and Basu, 1995; Gorring and Kay, 2001; Breitsprecher et al., 2003; Gorring et al., 2003), detachment of subducted plates (Pe-Piper and Piper, 1994; Davies and von Blanckenburg, 1995), and along subducted plate edges (Yogodzinski et al., 2001). In all cases, slab windows can impart dramatic changes along convergent margins, including near-trench high heat flow (e.g., DeLong et al., 1979) accompanied by magmatism and/or high-temperature–low-pressure metamorphism, geochemical changes in, or cessation of arc magmatism, and changes in strain partitioning between the subducted and overriding plates. There is growing evidence that these types of changes occurred along the early Tertiary margin of southern Alaska. For example, a series of early Tertiary near-trench plutons of the Sanak-Baranof belt (Fig. 1) were intruded into accretionary prism rocks (Hudson et al., 1979) and are interpreted to record the eastward migration of a ridge-trench-trench triple junction (Bradley et al., 1993, 2000). High-temperature–low-pressure metamorphism in accretionary prism rocks (Sisson et al., 1989) and the demise of arc magmatism during late Paleocene time (Wallace and Engebretson, 1984; Moll-Stalcup, 1994) were coeval with Sanak-Baranof belt magmatism. Closely following these events, volcanic rocks in the southern Talkeetna Mountains (referred to herein as the Caribou Creek volcanic field) (Fig. 1) began to erupt just near the continent side of the forearc region. In this paper, we present new field, radiometric age, and geochemical data that demonstrate that these volcanic rocks represent a unique episode of continental margin magmatism, which occurred in response to slab window development. The Caribou Creek volcanic rocks are

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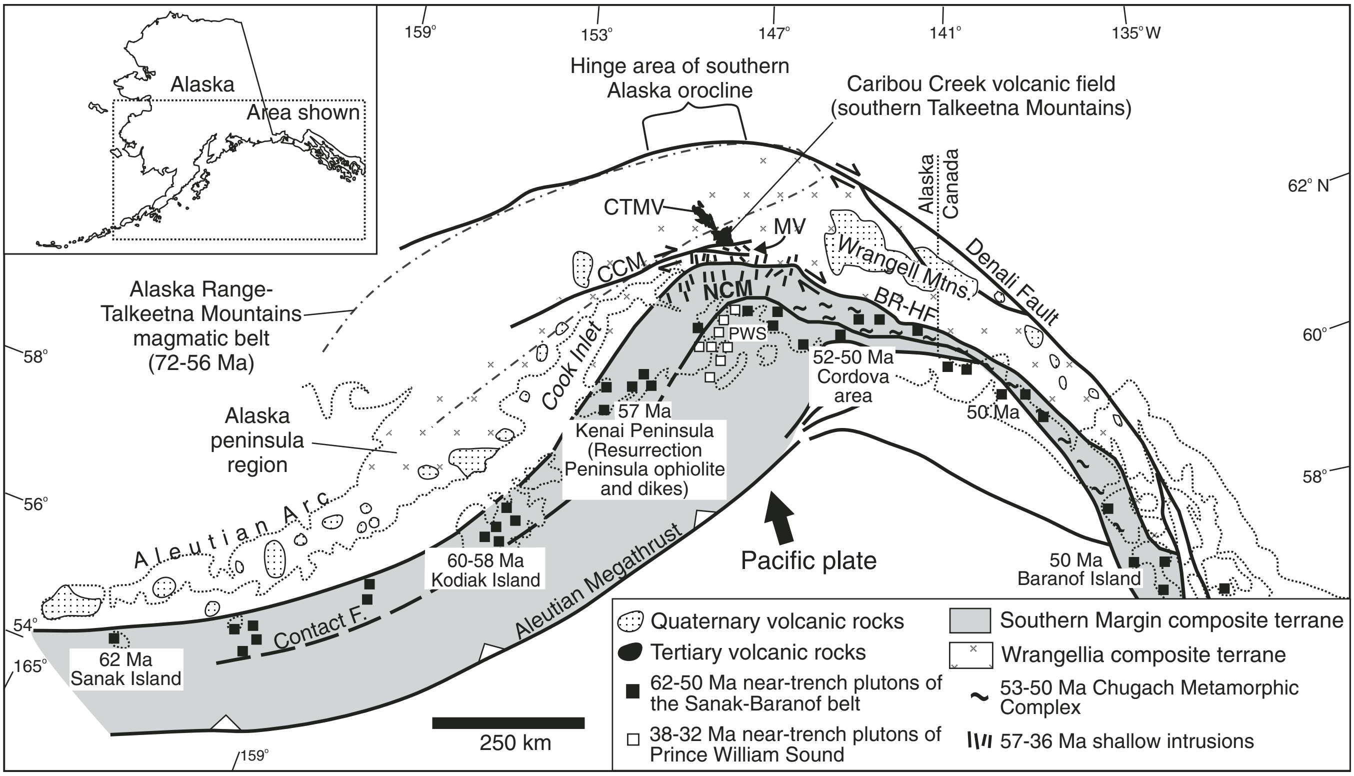

Figure 1. Map of south-central Alaska showing the Caribou Creek volcanic field, other volcanic and shallow intrusive rocks in the southern Alaska orocline, major fault systems, the Wrangellia and Southern Margin composite terranes, and regional magmatic and metamorphic belts. CTMV—central Talkeetna Mountains volcanic field, MV—Matanuska Valley, NCM—northern Chugach Mountains, PWS—Prince William Sound, BR-HF—Border Ranges–Hanagita fault system, and CCM—Caribou–Castle Mountain fault system. Geologic base is from Winkler (1992) and Wilson et al. (1998). Composite terrane boundaries are from Nokleberg et al. (1994) and Plafker et al. (1994). The age and area of the Chugach metamorphic complex is from Sisson et al. (1989) and Pavlis and Sisson (1995). Magmatic belts are from Moll-Stalcup et al. (1994), and ages for near-trench plutons are from Bradley et al. (1993, 2000).

unique because they include the most depleted continental basalts yet documented along the southern Alaska margin and are an example of slab window volcanism that occurred inboard of accretionary prism and forearc settings. Also, they provide an important link between early Tertiary magmatism and the regional tectonic events that shaped southern Alaska.

GEOLOGIC SETTING

The Caribou Creek volcanic field is located along the northern margin of the Matanuska Valley, which is a remnant forearc basin (Little and Naeser, 1989; Trop et al., 2003) situated between the 72–56 Ma Alaska Range–Talkeetna Mountains calc-alkaline magmatic belt to the north and accretionary prism rocks of the Southern Margin composite terrane to the south (Fig. 1). The Southern Margin composite terrane includes the Chugach and Prince William terranes and the Ghost Rocks Formation (of Kodiak Island) and consists of graywacke, argillite, mélange, chert, basalt, blueschists, glaucophanic greenschists, and ophiolitic rocks, which represent a late Mesozoic to early Tertiary subduction complex (Plafker et al., 1994). Late Paleocene to Eocene shallow intrusions and thin basalt flows are present within the Matanuska Valley (Silberman and Grantz, 1984) (Figs. 1 and 2) and represent an episode of magmatism within the forearc region. Silberman and Grantz (1984) inferred a depleted magma source for these rocks on the basis of strontium isotope data. More regionally, the Caribou Creek volcanic field lies within a N-NW–trending zone of Tertiary volcanic and shallow intrusive rocks in southern Alaska, which includes near-trench plutons of Prince William Sound, shallow intrusions in the northern Chugach Mountains, the shallow intrusions in the Matanuska Valley, and volcanic rocks in the central Talkeetna Mountains (Fig. 1). This zone is located along the axis of curvature of regional structures in southern Alaska; we refer to this zone as the hinge area of the southern Alaska orocline (Fig. 1).

The Caribou Creek volcanic rocks overlie Mesozoic rocks of the Peninsular terrane (Little, 1988; Winkler, 1992) and Tertiary nonmarine clastic rocks of the Wishbone Formation (Trop et al., 2003) (previously mapped in the study area as the Arkose Ridge Formation by Winkler, 1992) (Figs. 1 and 2). The Peninsular terrane is part of the Wrangellia composite terrane, which was accreted to southern Alaska between Late Jurassic and Late Cretaceous time (as summarized by Nokleberg et al., 1994; Ridgway et al., 2002), with final northward suturing by early Paleocene time (Cole et al., 1999; Ridgway et al., 2002). In the study area, the Peninsular terrane includes Middle Jurassic through Late Cretaceous marine clastic rocks that overlie Late Triassic(?) and Early to Middle Jurassic rocks of the Talkeetna arc (Nokleberg et al., 1994). The Jurassic Talkeetna arc forms the predominant crustal rocks beneath the Caribou Creek volcanic field and includes ultramafic and garnet-bearing cumulate rocks, gabbroic rocks, dioritic and tonalitic rocks, and volcanic and

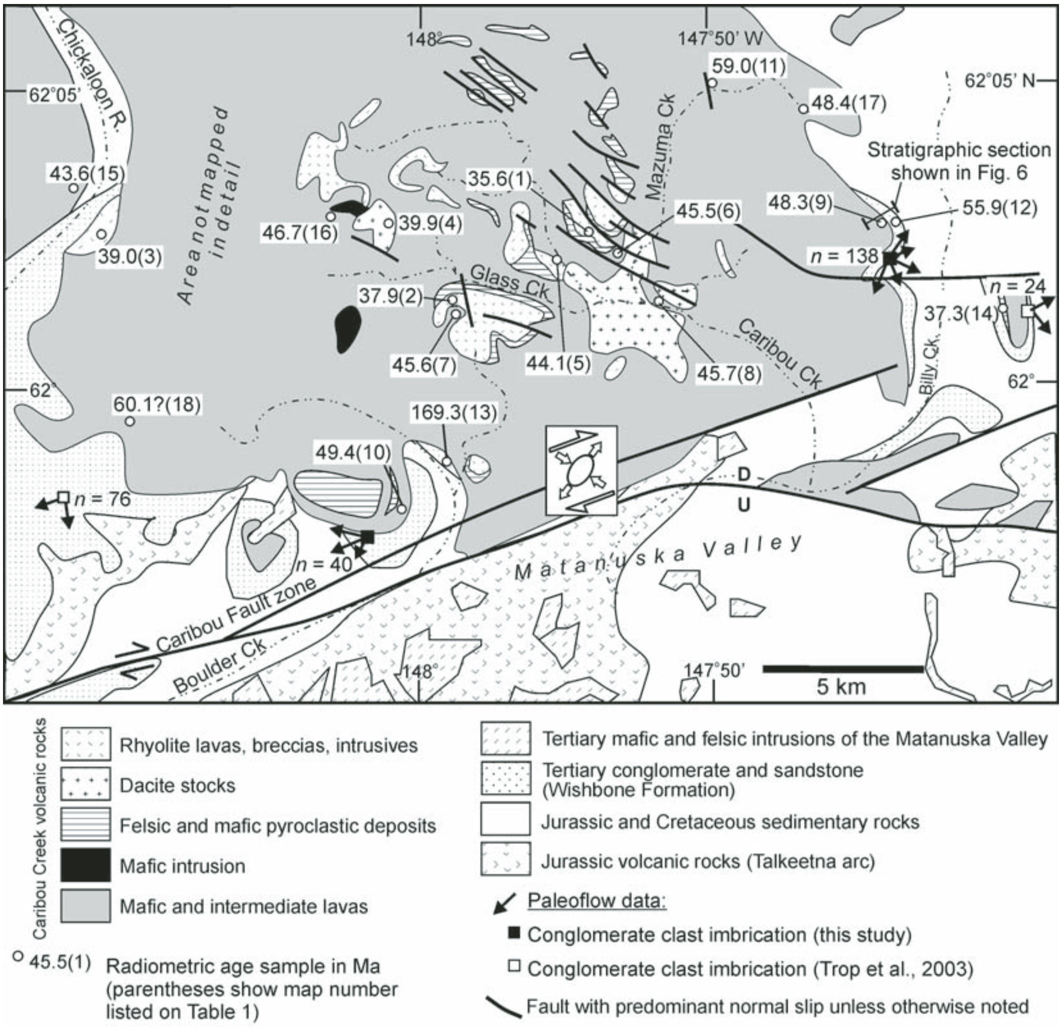

Figure 2. Simplified geologic map of the Caribou Creek volcanic field. Mapping of Caribou Creek volcanic rocks is from this study and Oswald et al. (2002); surrounding geology is from Winkler (1992) and Wilson et al. (1998). Strain ellipse shows predicted directions of shortening and extension that would occur in response to right-lateral simple shear along the Caribou strand of the Caribou–Castle Mountain fault system. D—downthrown side of fault, U—upthrown side of fault.

sedimentary rocks of the Talkeetna Formation (basaltic to andesitic lavas, silicic pyroclastic rocks, and lithic graywackes) (Burns, 1985; Plafker et al., 1989; DeBari and Sleep, 1991; Clift et al., 2005).

RADIOMETRIC AGE DATA

We report new $^{40}Ar/^{39}Ar$ ages for ten lava and shallow-intrusive samples, one tuff sample, and two conglomerate clast samples from the Caribou Creek volcanic field (Table 1; Figs. 2, 3, and 4). The $^{40}Ar/^{39}Ar$ age determinations were performed in the geochronology laboratory at the Geophysical Institute, University of Alaska, Fairbanks. We report plateau ages where plateaus were identified on the basis of a sample having three or more contiguous fractions constituting greater than 50% of gas release, with a mean square weighted deviation of plateau steps less than 2.5. Samples with fractions that are about the same age, but fail the plateau criteria are termed "pseudo-plateaus" as noted on Table 1 and in Figure 4. Measurements for the age determinations are given in Table DR1[1], and analytical procedures are outlined in Layer (2000).

The Caribou Creek lava and shallow intrusive samples have plateau ages that range from 49.4 ± 2.2 to 35.6 ± 0.2 Ma. An andesite tuff sample (99ANS-17A) found beneath a conglomerate at the base of the Caribou Creek volcanic rocks and above the Cretaceous Matanuska Formation yields a late Paleocene age of 59.0 ± 0.4. The two dated conglomerate clasts, a granitic clast and a porphyritic dacite clast, are from the Wishbone Formation at the base of the Caribou Creek volcanic rocks, and yield ages of 169.3 ± 0.9 and 55.9 ± 0.3 Ma, respectively. Other age data for the Caribou Creek volcanic field and volcanic and shallow intrusive rocks in adjacent areas are shown in Figure 3 and listed on Table 1. On the basis of the overlap in error ranges between our age dates and existing K-Ar ages, we interpret that the main phase of volcanism in the Caribou Creek volcanic field began during early Eocene time (ca. 52 Ma). The 59 Ma tuff is geochemically distinctive and, as described later, represents a volcanic event that preceded the main phase of Caribou Creek volcanism.

VOLCANIC UNITS AND STRATIGRAPHY

The Caribou Creek volcanic rocks consist of over 1000 m of shallow-dipping basalt and intermediate lavas, subordinate mafic and felsic pyroclastic deposits and felsic lavas, and mafic and felsic shallow intrusions (Figs. 2 and 5A). Mineralogically, the basalt lavas include modal olivine, clinopyroxene, and plagioclase

[1]GSA Data Repository item 2006014, Tables DR1–DR4, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

TABLE 1. RADIOMETRIC AGE DATA FOR TERTIARY VOLCANIC AND INTRUSIVE ROCKS OF THE CARIBOU CREEK VOLCANIC FIELD

Map no. on Figure 2[†]	Sample	Latitude (°N)	Longitude (°W)	Material dated[‡]	Age (Ma)
Caribou Creek volcanic field ($^{40}Ar/^{39}Ar$ dates of this study)[§]					
1	7-TMA3–6 Rhyolite lava	62.0417	147.9017	WR	35.6 ± 0.2
				KFS	33.7 ± 0.4[#]
2	GLA99–18C Rhyolite lava	62.0219	147.9942	WR	37.9 ± 0.2[#]
3	CHK-00–1 Fine granite	62.0544	148.1911	KFS	39.0 + 0.4
4	99ANS-9A Dacite dome	62.0472	148.0200	WR	39.9 ± 0.7
5	2-TMA3–3 Rhyolite dome	62.0389	147.9139	WR	44.1 ± 0.2[#]
6	8-TMA3–1 Andesite lava	62.0325	147.8861	WR	45.5 ± 0.3[#]
7	GLA99–17 Basalt lava	62.0219	147.9850	WR	45.6 ± 5.1
8	99ANS-18A Andesite lava	62.0242	147.8611	WR	45.7 ± 1.4
9	STM1–85 Basaltic andesite lava	62.0456	147.7239	WR	48.3 ± 1.6
10	BC98TY2A Basaltic andesite lava	61.8739	148.0272	WR	49.4 ± 2.2
11	99ANS-17A Andesite tuff	62.0825	147.8256	FS	59.0 ± 0.4
12	BIL99–14-C2 Dacite clast	62.0403	147.6103	BI	55.9 + 0.3[#]
13	BC98–7-C5 Granite clast	61.9792	147.7375	BI	169.3 ± 0.9
		Reference			
Caribou Creek volcanic field (other studies)					
14	Felsic tuff between conglomerate and mafic lavas, $^{40}Ar/^{39}Ar$ method	Trop et al. (2003)		FS	37.3 ± 2.1
na	Basalt lava near base of volcanic section, K-Ar method	Panuska et al. (1990)		WR	44.1 ± 2.6
na	Basalt lava, lower part of volcanic section, K-Ar method	Panuska et al. (1990)		WR	53.6 ± 3.2
na	Basalt lava, upper part of volcanic section, K-Ar method	Panuska et al. (1990)		WR	40.8 ± 2.4
15	Pyroclast in tuff breccia; sample 79AGz-116, K-Ar method	Silberman and Grantz (1984)		WR	43.6 ± 2.2
16	Mafic(?) plug; sample 79AGz-102, K-Ar method	Silberman and Grantz (1984)		WR	46.7 ± 2.3
17	Andesite sample 79AGz-107, K-Ar method	Silberman and Grantz (1984)		WR	48.4 ± 2.4
na	Basalt sample 79AGz-106, K-Ar method	Silberman and Grantz (1984)		WR	55.5 ± 3.5
18	Basalt sample 79AGz-112, K-Ar method	Silberman and Grantz (1984)		WR	60.1 ± 4.6[††]

[†]na—locality beyond area shown in Figure 2 or exact locality not available.
[‡] WR—whole rock; FS—alkali feldspar; KFS—K-feldspar; BI—biotite.
[§]Samples run against standard MMhb-1 with an age of 513.9 Ma. Except where noted, ages are plateau ages reported to 1σ. See text and Table DR1 (text footnote one) for details.
[#]Pseudo-plateau age. See text for plateau criteria.
[††]Uncertain, low %K_2O.

(Table 2). The intermediate lavas include basaltic andesite and andesite that consist mostly of plagioclase and generally contain less than 10% pyroxene (Table 2). A small volume of basaltic pyroclastic units, including base surge, spatter, and diatreme deposits, are associated with the mafic and intermediate lavas (Table DR2 [see footnote 1]; Figs. 5B and 5C).

Dacite and rhyolite lavas and shallow intrusions are found in the central part of the volcanic field where they form dome complexes (Figs. 2 and 5A). There are mutually crosscutting relationships between the dacite and rhyolite in different parts of the volcanic field, indicating that their magmas were formed and emplaced coevally. Most of the felsic lavas contain quartz, plagioclase, and alkali feldspar, but a subset with adakite-like geochemistry, including samples from several different lavas and small intrusions, is composed of 85%–95% plagioclase with only minor quartz and a few percent of modal pyroxene (Table 2). The pyroxene in these rocks contains inclusions of plagioclase, exhibits resorbed grain boundaries, and is not strongly altered. Felsic pyroclastic fallout and flow deposits (Table DR2 [see footnote 1]; Fig. 5D) are generally thickest in the central part of the volcanic field, where they are stratigraphically associated with felsic lava domes (Fig. 2).

Fluvial conglomerate and sandstone deposits of the Eocene Wishbone Formation (Trop et al., 2003) are interbedded with mafic and intermediate lavas and pyroclastic deposits along the margins of the volcanic field (Fig. 6). Trop et al. (2003) interpreted the Wishbone Formation as stream-dominated alluvial fan deposits consisting of two main lithofacies; organized imbricated conglomerate with cross-stratified sandstone and unorganized boulder-cobble conglomerate. We also recognize these lithofacies (Figs. 7A and 7B; Table DR2 [see footnote 1]) and found that the organized facies includes a wide mixture of metamorphic, sedimentary, and volcanic lithic detritus, while the unorganized facies contains mostly unaltered volcanic detritus (Table DR2 [see footnote 1]; Fig. 6) (Hooks et al., 2001). The depositional processes (Table DR2 [see footnote 1]), the high percentage of volcanic detritus, and the stratigraphic association with lavas and pyroclastic deposits indicate that the unorganized facies was deposited during active phases of Caribou Creek volcanism when there was an abundance of volcanic detritus supplied to surrounding drainage systems (e.g., Cole and Ridgway, 1993). The organized facies likely records the initial onset of Caribou Creek volcanism when there may have been thermal uplift and erosion of local Jurassic rocks as indicated by the presence of Jurassic-aged clasts (Table 1; Trop et al., 2003) and by metamorphic and sedimentary lithic grains that are characteristic of Jurassic Talkeetna arc source rocks (Hooks et al., 2001). Collectively, the association of these two facies with Caribou Creek volcanism is supported by early Eocene ages obtained on volcanic clasts (Table 1; Trop et al., 2003) and by conglomerate clast imbrication data, which show that paleoflow during Wishbone deposition was directed radially away from the central

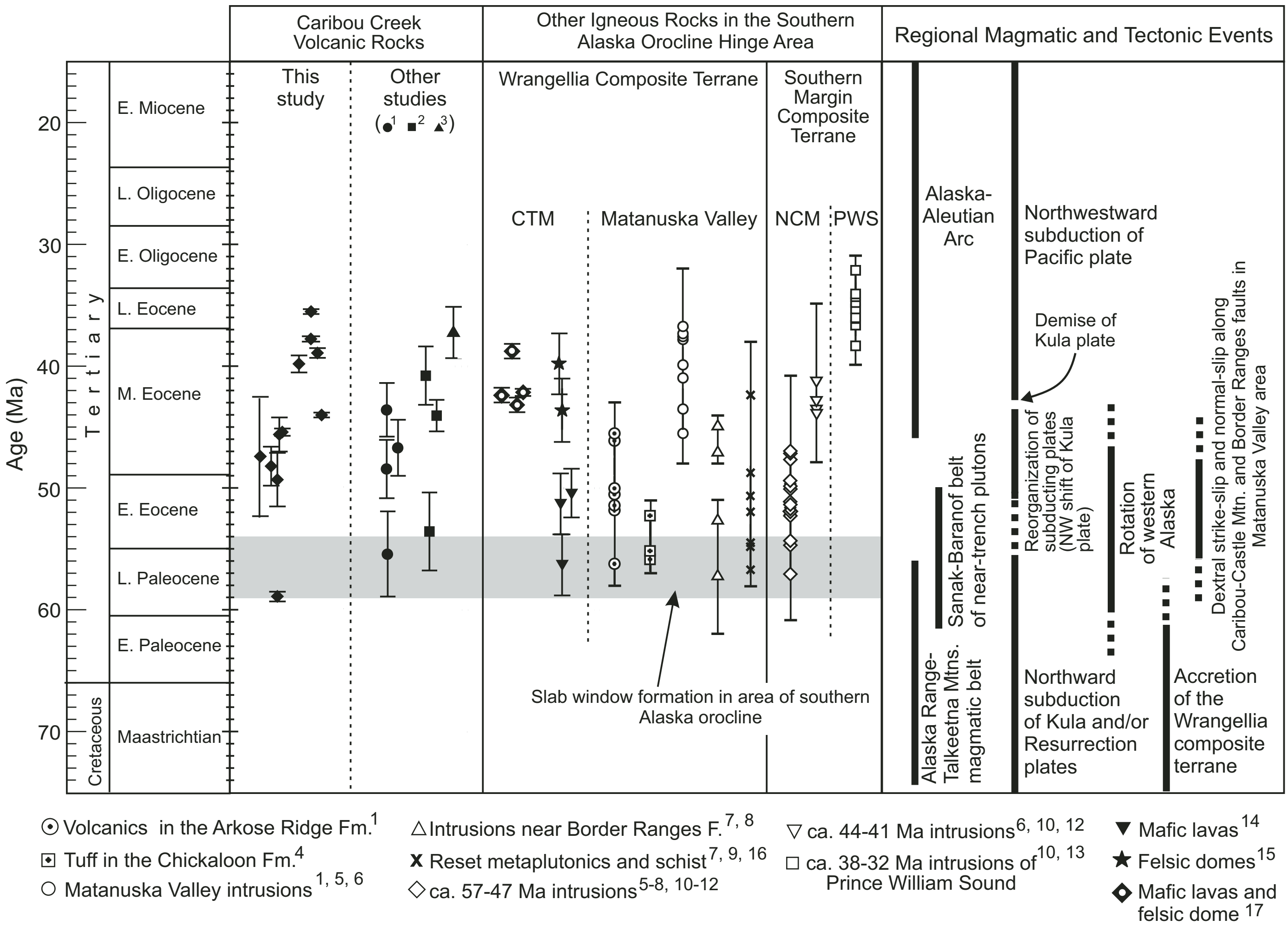

Figure 3. Age-event diagram showing radiometric ages of volcanic rocks in the Caribou Creek volcanic field and other volcanic rocks and shallow intrusions in the Wrangellia composite terrane and Southern Margin composite terrane in the hinge area of the southern Alaska orocline. CTM—central Talkeetna Mountains; NCM—northern Chugach Mountains; PWS—Prince William Sound. Sources for age data are as follows: 1—Silberman and Grantz (1984); 2—Panuska et al. (1990); 3—Trop et al. (2003); 4—Triplehorn et al. (1984); 5—Little (1988); 6—Little and Naeser (1989); 7—Plafker et al. (1989); 8—Winkler et al. (1981); 9—Madden-McGuire et al. (1990); 10—Nelson et al. (1985); 11—Updike and Ulery (1988); 12—Plafker et al. (1992); 13—Clark et al. (1976); 14—Csejtey et al. (1978); 15—Adams et al. (1985) and Little (1988); 16—Harlan et al. (2003); and 17—R.B. Cole and P.W. Layer (2002, personal commun.). The regional magmatic and tectonic events are discussed in the text. Time scale is from Cande and Kent (1992).

part of the volcanic field (Fig. 2). The two fluvial facies are not found together everywhere, but where they are, the deposits of the organized facies unconformably overlie Jurassic and Cretaceous sedimentary rocks and, in turn, are unconformably overlain by deposits of the unorganized facies, which is conformable and interbedded with Caribou Creek volcanic rocks (Figs. 6 and 7C).

GEOCHEMISTRY OF THE VOLCANIC ROCKS

Major element data for 66 samples and trace element data for 39 samples of volcanic and shallow-intrusive rocks from the Caribou Creek volcanic field are reported on Tables DR3 and DR4 (see footnote 1). Samples were powdered using an alumina ceramic mixer mill at Allegheny College. Major elements were determined by X-ray fluorescence, and trace elements were determined by inductively coupled plasma–mass spectrometry at ALS-Chemex Labs, Inc. About a third of the samples analyzed for major elements had relatively high loss on ignition (>3%), which most likely reflects posteruptive alteration. The major element data were therefore normalized to 100%, volatile free, to eliminate the alteration effects on major oxide characterization (e.g., as described by Johnson et al., 1999).

The volcanic and shallow-intrusive rocks are subalkalic and range between tholeiite and calc-alkaline series compositions (Fig. 8). This suite shows systematic decreases in MgO, Fe_2O_3, CaO, Al_2O_3, and increased K_2O with increased SiO_2 (Fig. 8). Most samples have low and narrow ranges of large ion lithophile element (LILE) (e.g., Ba, Pb, Th, and Sr), high field strength element (HFSE) (e.g., Zr), and light rare earth element (LREE) (e.g., La) to Yb ratios compared to rocks of the 72–56 Ma Alaska Range–Talkeetna Mountains magmatic belt and rocks of the Quaternary Aleutian arc (Fig. 8). The basalt samples have moderate to high MgO (3.6%–10.6%) and low

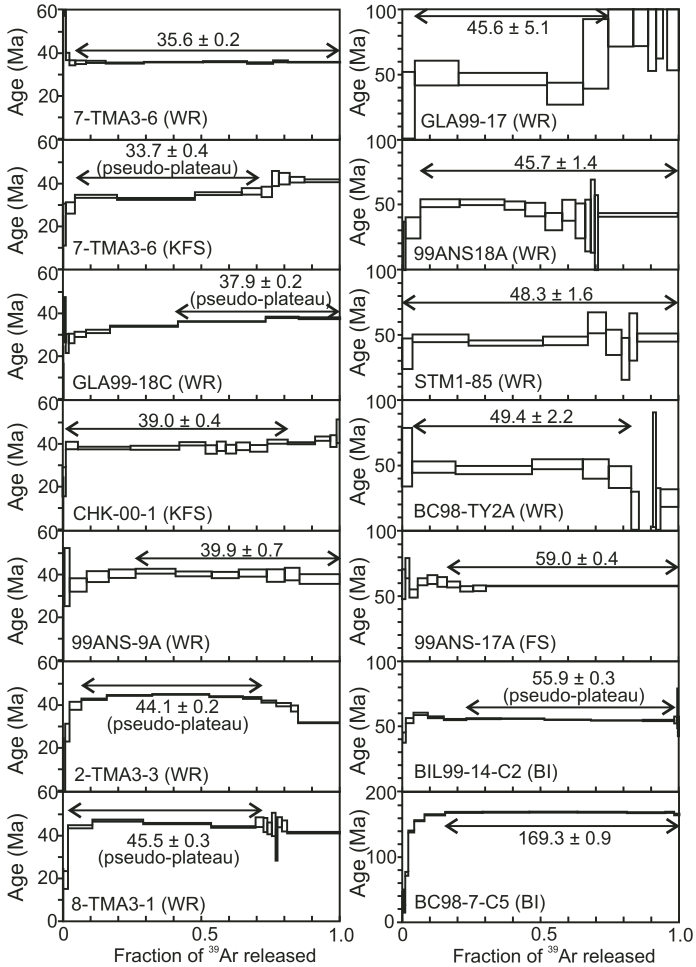

Figure 4. Plateau diagrams for the $^{40}Ar/^{39}Ar$ age samples of the Caribou Creek volcanic field. Plateau ages are shown except where noted. Criteria for plateau and pseudo-plateau ages are described in the text and on Table DR1 (see text footnote 1). WR—whole rock; KFS—potassium feldspar; FS—alkali feldspar; BI—biotite.

to moderate Mg# of 0.24–0.49 (Table DR3 [see footnote 1]).

The Caribou Creek volcanic rocks are especially distinctive in their rare earth element (REE) concentrations, with low La/Yb ratios for the basalt-andesite-rhyolite samples (Figs. 8 and 9). The basalts exhibit LREE depletion patterns and other trace element patterns that are similar to normal mid-ocean-ridge basalt (N-MORB), with the exception of some enrichment in Cs, Rb, Ba, Th, U, and Pb (Figs. 9 and 10). The basaltic andesites show slightly higher LREE concentrations, in the range of N-MORB and enriched mid-ocean-ridge basalt (E-MORB) (Fig. 9), and also show enrichment in Cs, Rb, Ba, Th, U, and Pb (Fig. 10).

The andesite samples exhibit flat and parallel REE patterns, with the exception of sample 99ANS-17A, which is more strongly depleted in the heavy rare earth elements (HREE) than the remaining andesite samples (Fig. 9). This sample is from a 59 Ma tuff found at the base of the volcanic sequence. The remaining andesite samples show patterns of REE and incompatible element enrichment that are similar to the basalts but with a greater degree of enrichment in Cs, Ba, Th, and U. The andesites also display Nb-Ta depletion anomalies relative to U and K (Fig. 10).

The rhyolite and dacite samples fall into two groups, an enriched group and a depleted group, on the basis of trace element concentrations. The enriched group, which includes most of the dacite and rhyolite samples, is more enriched in REEs compared to the mafic and intermediate samples, but has low La/Yb ratios (Fig. 9). This group also shows negative Eu anomalies, greater enrichment of the more incompatible trace elements (e.g., Cs, Rb, Ba, and K) compared to the basalts and andesites, as well as marked depletions of Sr, Eu, P, and Ti (Figs. 9 and 10).

The depleted group of felsic rocks, which are samples from stocks, a pyroclastic clast, and a lava, are depleted in the HREEs, as well as in a range of compatible and incompatible trace elements, compared to the enriched felsic rocks (Figs. 9 and 10). The depleted felsic rocks are geochemically similar to andesite sample 99ANS-17A, and as a group these rocks exhibit characteristics of adakites. Adakites are characterized as having $SiO_2 \geq 56\%$, $Al_2O_3 \geq 15\%$, MgO < 3%, Yb ≤ 1.9 ppm, Y < 18 ppm, Sr > 400 ppm, Sr/Y > 20–40, an absence of a negative Eu anomaly, and a high percentage of modal plagioclase (Kay, 1978; Defant and Drummond, 1990). The depleted intermediate and felsic samples meet most of these criteria with the exception of some samples that have lower Sr concentrations than other adakites (Table 2; Tables DR3 and DR4 [see footnote 1]; Fig. 11).

MAGMA EVOLUTION

The basalts of the Caribou Creek volcanic field are similar to mid-ocean-ridge basalt in their concentrations of REEs and in those trace elements that are typically attributed to mantle sources, such as the HFSE (e.g., Ta, Nb, Zr, Hf, and Ti) (e.g., Pearce and Parkinson, 1993; Davidson, 1996) (Figs. 9 and 10). The basalts also have depleted strontium isotope compositions in the range of mid-ocean-ridge basalt, with $^{87}Sr/^{86}Sr_{(i)}$ ratios between 0.7028 and 0.7032 (Cole and Stewart, 2005). From these data, we interpret that the Caribou Creek basaltic magmas were derived from a depleted mantle source. To test this interpretation, we compared representative Caribou Creek basalts with partial-melting models of a depleted spinel peridotite (representative of suboceanic upper mantle) calculated by Cousens et al. (1995) (Fig. 12). There is good agreement between the Caribou Creek basalts and the range of 5%–10% partial melts of spinel peridotite, with the exception of a higher range of Rb and La/Ce ratios in some Caribou Creek basalts (e.g., sample GLA99-17 in Fig. 12). This comparison supports our interpretation of a depleted mantle source for Caribou Creek basalt magmas but also indicates that the Caribou Creek basalts became enriched in LILEs and LREEs compared to primary partial melts of spinel peridotite. The basalts also show enrichment relative to N-MORB in fluid-mobile elements (e.g., Cs, Ba, U, and Pb) (Fig. 10). Enrichment in fluid-mobile elements and also in Rb, Th, K, Sr, and the LREEs is typical of arc volcanic rocks, where these elements are derived from fluids and sediments of the subducted slab (Pearce and Parkinson, 1993; Plank and Langmuir, 1993; Pearce et al., 1995; Davidson, 1996; Hochstaedter et al.,

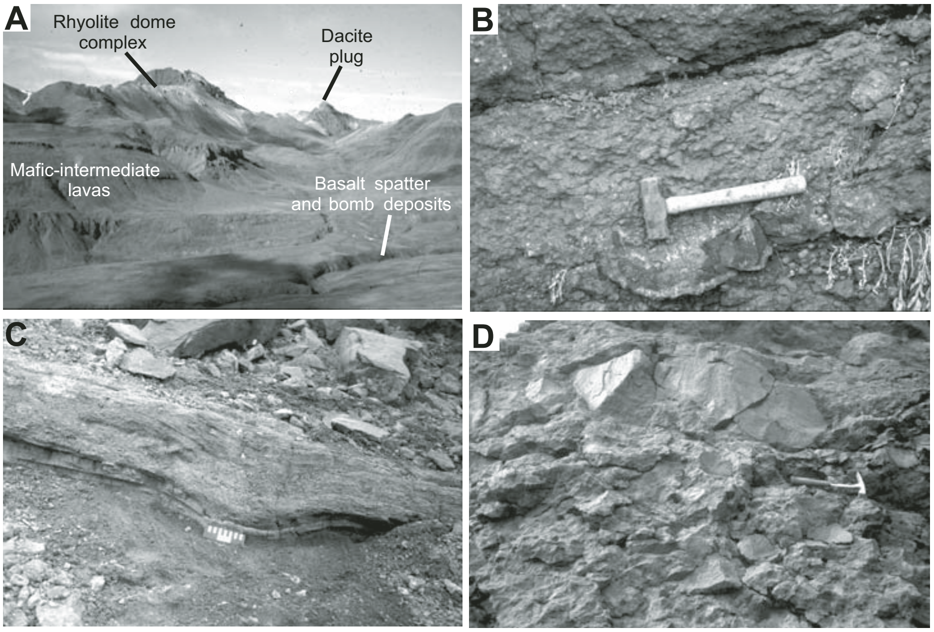

Figure 5. Photographs of the Caribou Creek volcanic rocks. (A) Mafic and intermediate lavas and shallow felsic intrusions in the upper part of Glass Creek (view is toward the northwest). Thickness of rocks shown is ~500 m. (B) Basalt lapillistone-breccia spatter deposit; hammer is 34 cm long. Note chilled margin on block beneath the hammer. (C) Trough in mafic pyroclastic surge deposit; scale is 15 cm. (D) Inversely graded felsic pyroclastic flow deposit; hammer is 28 cm long.

TABLE 2. MODAL MINERALOGY (IN PERCENT) OF CARIBOU CREEK VOLCANIC ROCKS, BASED ON VISUAL ESTIMATES

Unit	Ol	Cpx	Opx	Hbl	Bi	FeTi	Pl	Alk	Qtz
Basalt	5–10	10–15	<2	-	-	5–10	60–70	-	-
Basaltic andesite	<2	5–15	5–10	-	-	5–10	65–85	-	-
Andesite	-	<5	<5	5–10	5–10	5–10	80–90	-	-
Dacite and rhyolite	-	-	-	<5	5–10	<5	40–55	15–25	30–45
Adakite	-	1–2	-	-	<5	<5	85–95	<10	<10

Note: Mineral abbreviations are: Ol—olivine, Cpx—clinopyroxene, Opx—orthopyroxene, Hbl—hornblende, Bi—biotite, FeTi—iron-titanium oxides, Pl—plagioclase, Alk—alkali feldspar, Qtz—quartz.

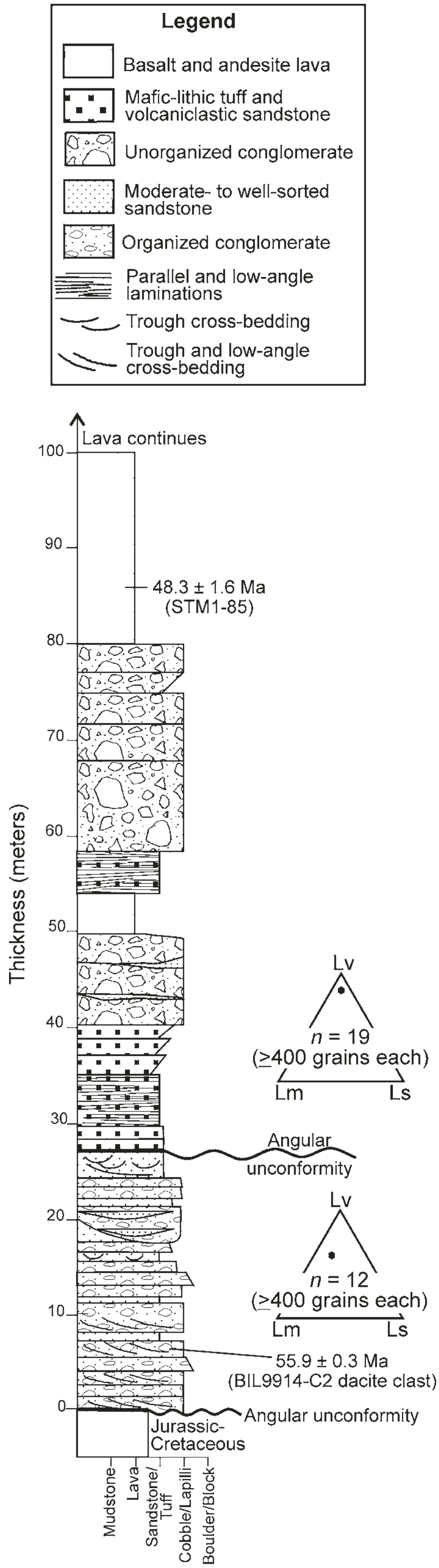

Figure 6. Representative stratigraphic section of fluvial deposits of the Wishbone Formation along the eastern margin of the Caribou Creek volcanic field (measured bed-by-bed). Section location is shown in Figure 2. Radiometric age samples are listed in Table 1. Average sandstone petrographic data shown on LvLmLs ternary diagrams are from Hooks et al. (2001); Lv—volcanic lithic grains, Lm—metamorphic lithic grains, Ls—sedimentary lithic grains.

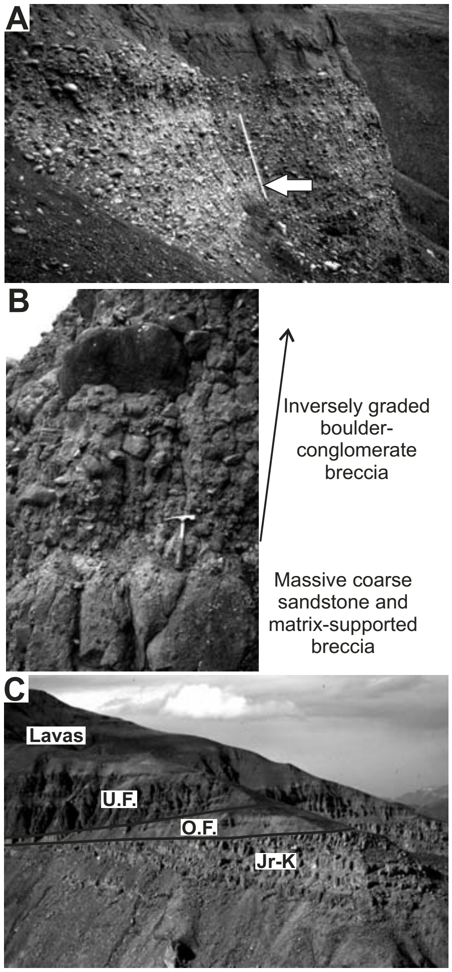

Figure 7. Photographs of fluvial deposits of the Wishbone Formation along the eastern margin of the Caribou Creek volcanic field near Billy Creek. (A) Conglomerate of the well-organized fluvial facies. Arrow points to Jacob staff that is 1.5 m long. (B) Conglomerate of the poorly organized fluvial facies. Note boulder-sized clast near top of photograph. Hammer is 28 cm long. (C) Unconformable contacts (black lines) between Jurassic-Cretaceous sedimentary rocks (Jr-K), rocks of the organized fluvial facies (braided fluvial deposits) (O.F.), and rocks of the unorganized fluvial facies (debris-flow and braided fluvial deposits) (U.F.). The unorganized facies is conformable (interbedded) with the overlying mafic and intermediate lavas. Thickness of rocks shown is ~350 m.

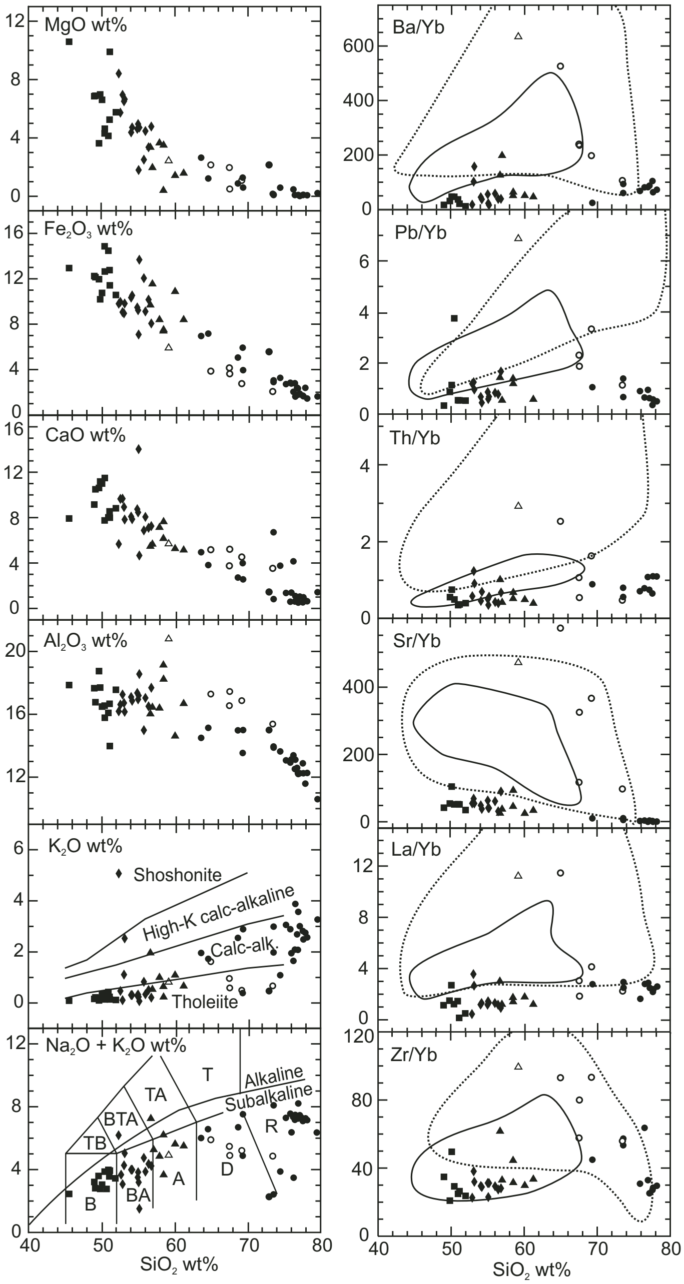

Figure 8. Major and trace element variation diagrams for rocks of the Caribou Creek volcanic field. Symbols are as follows: square—basalt, diamond—basaltic andesite, triangle—andesite, circle—dacite and rhyolite. Open symbols are samples that have geochemical characteristics similar to adakites. The rock classification scheme on the SiO_2 vs. $Na_2O + K_2O$ plot is from LeBas et al. (1986), and the rock series on the SiO_2 vs. K_2O plot are from Rickwood (1989). Rock name abbreviations are: B—basalt, BA—basaltic andesite, A—andesite, D—dacite, R—rhyolite, TB—trachybasalt, BTA—basaltic trachyandesite, TA—trachyandesite, T—trachyte. Dashed outlines show composition of volcanic rocks from the Quaternary Aleutian Arc on the Alaska Peninsula; data compiled from Hildreth (1983), Nye and Turner (1990), Nye et al. (1994), Till et al. (1994), and Johnson et al. (1996). Solid outlines show composition of volcanic and plutonic rocks from the 72–56 Ma Alaska Range–Talkeetna Mountains belt; data compiled from Reed and Lanphere (1973), Lanphere and Reed (1985), Reiners et al. (1996), Cole and Layer (2002), Cole (2003), Cole and Stewart (2005).

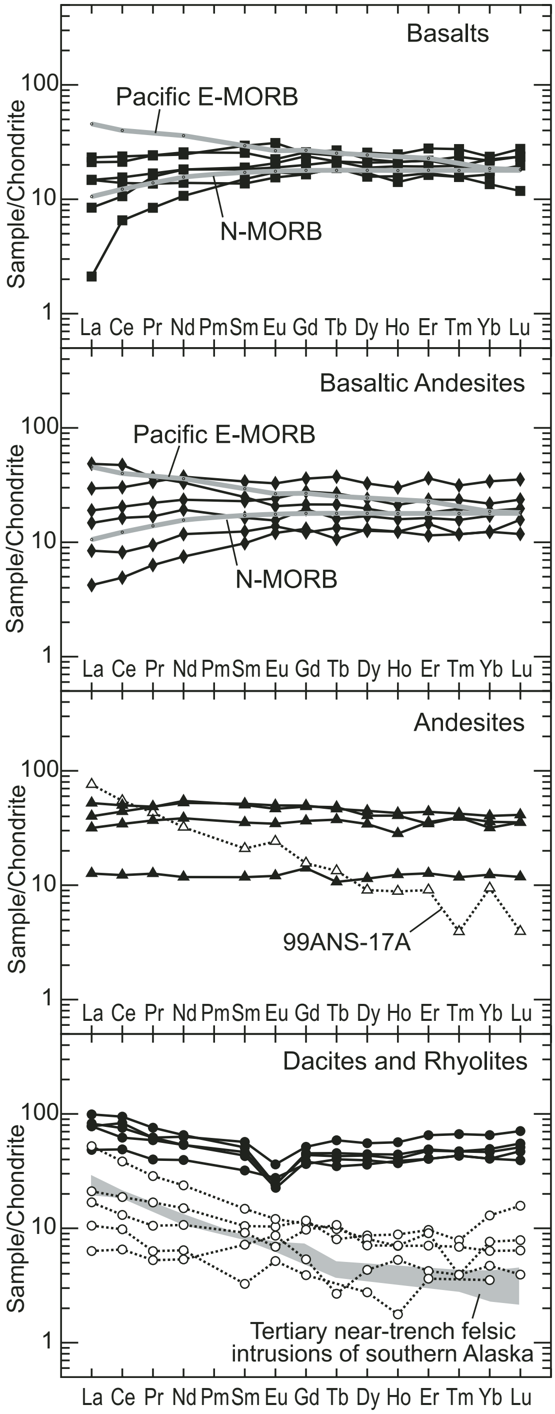

Figure 9. Chondrite-normalized rare earth element (REE) plots for rocks of the Caribou Creek volcanic field. Symbols are as shown in Figure 8. The shaded area showing Tertiary near-trench felsic intrusions of southern Alaska includes samples 84ANK-122C and 84APR219 from the northern Chugach Mountains (Plafker et al., 1989) and sample APR140 from the Cordova area (Barker et al., 1992). Normalizing values are from Sun and McDonough (1989). Values for normal mid-ocean-ridge basalt (N-MORB) are from Sun and McDonough (1989) and Pearce and Parkinson (1993), and values for Pacific enriched mid-ocean-ridge basalt (E-MORB) are from GERM (2005).

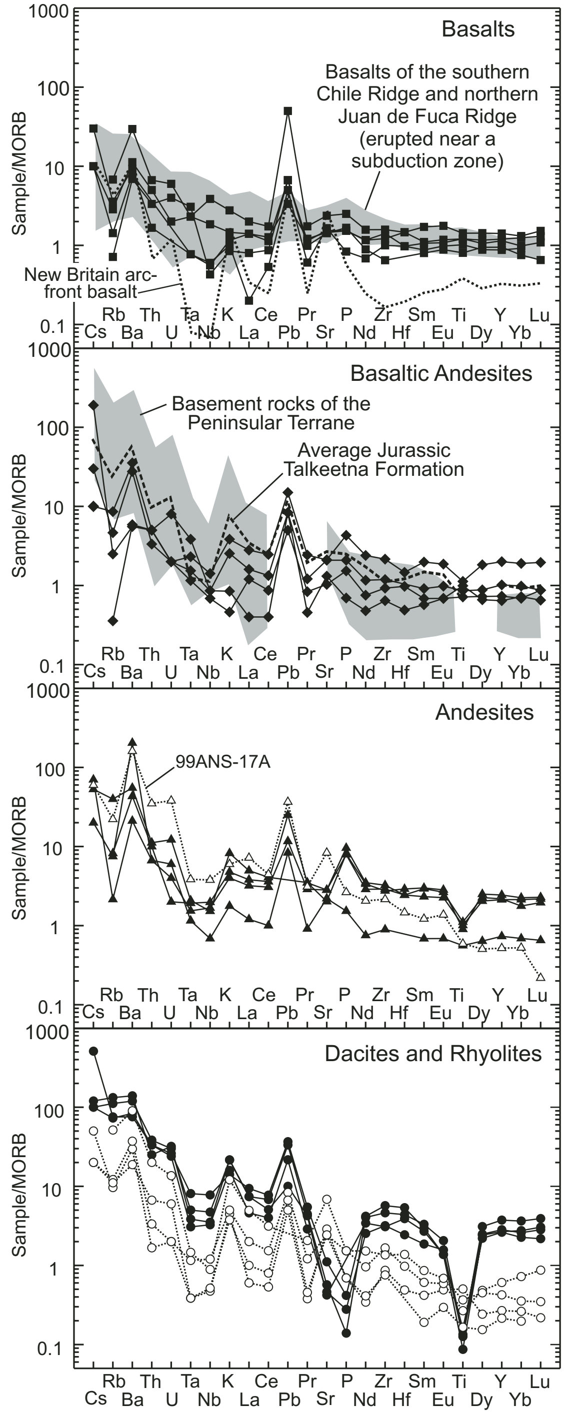

Figure 10. Mid-ocean-ridge basalt (MORB)–normalized trace element plots for rocks of the Caribou Creek volcanic field. Normalizing values and arrangement of elements are from Pearce and Parkinson (1993). Relative compatibility of elements increases from left to right. The shaded area showing the composition of basalts from the southern Chile and northern Juan de Fuca spreading ridges is from data of Klein and Karsten (1995) and Cousens et al. (1995). The New Britain arc front basalt is sample 2651A from Woodhead et al. (1998). The shaded area showing the composition of Peninsular terrane basement rocks is from data of Plafker et al. (1989), DeBari and Coleman (1989), and DeBari and Sleep (1991). The average composition of the Jurassic Talkeetna Formation is from Clift et al. (2005).

2001). While the Caribou Creek volcanic rocks do exhibit some of these arc-like geochemical characteristics, several observations indicate that the Caribou Creek magmas did not evolve as part of a continental margin arc system. First, the Caribou Creek volcanic field is located closer to the paleo-forearc (farther south) than most of the Late Cretaceous and Tertiary calc-alkaline arc-related rocks of southern Alaska (e.g., the Alaska Range–Talkeetna Mountains belt) (Fig. 1). Second, the Caribou Creek volcanic rocks began to erupt during a hiatus in regional arc magmatism across southern Alaska (Fig. 3). In fact, the Caribou Creek volcanic field was one of the only active volcanic areas in south-central Alaska between ca. 52 and 45 Ma (Moll-Stalcup et al., 1994). These observations, together with geochemical trends described later, suggest to us that the Caribou Creek volcanic field was not formed as part of a regional margin-parallel arc system.

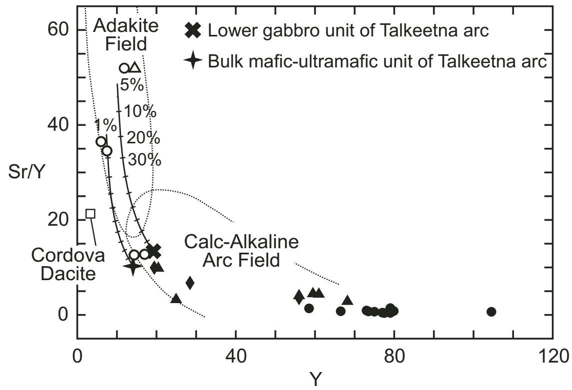

Figure 11. Graph of Sr/Y versus Y for rocks of the Caribou Creek volcanic field with SiO2 ≥ 56 wt%. Symbols for the Caribou Creek volcanic samples are as shown in Figure 8. Adakite and calc-alkaline arc fields are from Defant and Drummond (1990, 1993). The Cordova dacite is sample APR140 from Barker et al. (1992). Shown on the graph are equilibrium melting curves for different end-member compositions of the lower mafic-ultramafic portion of the Jurassic Talkeetna arc (as labeled on the graph). The tick marks show percentages of melting in 10% increments, except where labeled at ends of curves. Bulk partition coefficients of 0.52 and 1.9 for Sr and Y, respectively, are based on an average mineralogy of 38% olivine, 5% garnet, 15% clinopyroxene, 5% orthopyroxene, 10% amphibole, and 25% plagioclase for the mafic to ultramafic rocks of the lower Talkeetna arc (mineralogy is from Plafker et al., 1989; DeBari and Coleman, 1989; and DeBari and Sleep, 1991). Mineral–melt partition coefficients are from Rollinson (1993; their Table 4.1) and Lytwyn et al. (2000).

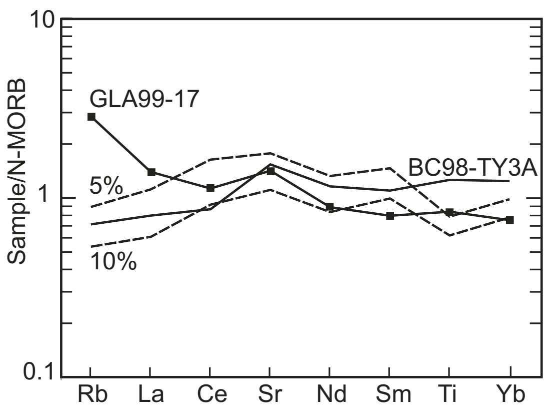

Figure 12. Comparison of Caribou Creek basalts (solid lines as labeled) with partial melting models of Cousens et al. (1995) for a spinel peridotite mantle reservoir. The 5% and 10% partial melt model curves (dashed lines) are labeled. The initial mineral assemblage used for the spinel peridotite is 65% olivine, 15% orthopyroxene, 15% clinopyroxene, and 5% spinel. An equilibrium nonmodal model is applied. Model parameters and results are given on Table 3 of Cousens et al. (1995). N-MORB—normal mid-ocean-ridge basalt.

Geochemically, the Caribou Creek basalts do not show the Ta-Nb depletion trend that is typically found in arc volcanic rocks, and they are more depleted in LREEs and other incompatible elements than would generally be expected for arc basalts (e.g., Davidson, 1996). There are, however, published examples of arc basalts that are strongly depleted. For example, basalts erupted along the front of the New Britain island arc of Papua New Guinea are enriched in fluid-mobile elements but are strongly depleted in the HFSEs and LREEs compared with N-MORB (Woodhead et al., 1998) (Fig. 10). Woodhead et al. (1998) interpreted these strongly depleted arc-front basalts to have formed after previous melt extraction from a depleted mantle source that was also contaminated by slab-derived fluid. The Caribou Creek basalts have much less-depleted HFSE concentrations (closer to N-MORB) than the New Britain arc-front basalts (Fig. 10) and so probably did not form from a mantle source that experienced a similar degree of previous melt extraction. Woodhead et al. (1998) further documented that there is a decrease in the flux of slab-derived fluids into the sub-arc mantle from the arc front toward the backarc. Ryan et al. (1996) documented a similar forearc to backarc decrease in fluid flux from subducted slabs across several arc systems. By analogy, if the Caribou Creek volcanic rocks were erupted in the frontal part of a volcanic arc (e.g., as part of the Alaska Range–Talkeetna Mountains magmatic belt), then we would expect the Caribou Creek rocks to exhibit greater fluid-mobile and LILE enrichment than rocks farther inboard (northward). Overall, the Caribou Creek rocks have a *lower* range of fluid-mobile and large ion lithophile elements (e.g., lower Ba/Yb, Pb/Yb, Th/Yb, Sr/Yb, and La/Yb ratios) than rocks of the Alaska Range–Talkeetna Mountains magmatic belt, which lies farther north (Fig. 8). This is further evidence that the Caribou Creek volcanic field was not simply part of a regional arc system. The Caribou Creek rocks also have lower LILE/Yb ratios than the Alaska Peninsula segment of the Quaternary Aleutian arc (Fig. 8), which formed on Peninsular terrane basement rocks, similar to those beneath the Caribou Creek volcanic field. The same pattern is found when the LILEs are compared to HFSEs, such as Zr. The lack of decoupling of fluid-mobile elements and LILEs (e.g., Ba, Pb) and those elements typically attributed to mantle sources (e.g., Yb, Zr) is relevant because arc volcanic rocks are generally more strongly enriched in LILE relative to HFSE and are characterized by higher LILE/HFSE ratios than those exhibited by the Caribou Creek rocks (Patino et al., 2000; Hochstaedter et al., 2001).

Although the Caribou Creek basalts retain characteristics of a depleted source, none of our samples can be considered as representative of a primary basalt that formed in equilibrium with depleted upper mantle mineralogies; the basalts have relatively low Mg# (0.28–0.49), low Cr (92–280 ppm), and low Ni (20.8–132 ppm) compared to values for magmas in equilibrium with typical upper mantle mineralogies (>0.7, >400 ppm, and >1000 ppm, respectively) (Wilson, 1989, p. 22). The low Mg, Cr, and Ni concentrations of the Caribou Creek basalts indicate that they formed after olivine and other mafic mineral phases had fractionated from the primary magma. Nevertheless, the Caribou Creek volcanic rocks exhibit uniform trends of major and trace elements from the basalts through rhyolites (e.g., Figs. 8–10), indicating that they are a cogenetic suite.

To test whether the basalts could have been parental to the range of more acidic Caribou Creek rocks, we applied assimilation-fractional crystallization (AFC) modeling of DePaolo (1981) using the Igpet for Windows software of M.A. Carr (version of June 12, 2002) distributed by TerraSofta, Inc. (Fig. 13). In our models, we selected basalt samples BC98-TY3A and GLA99-17 as representative parent magma compositions and the Jurassic Talkeetna Formation as the assimilant. The Talkeetna Formation consists mostly of basaltic to andesitic lavas, silicic pyroclastic rocks, and lithic graywackes with a minimum thickness range of 3–7 km (Burns, 1985; Plafker et al., 1989; Clift et al., 2005). These rocks form the upper part of the Jurassic Talkeetna arc, which once was an oceanic island arc (Burns, 1985; Clift et al., 2005) that now is part of the Peninsular terrane and represents the upper-crustal rocks beneath the Caribou Creek volcanic field (Plafker et al., 1989).

The modeling results are consistent for trace elements that exhibit a range of fluid mobility. For example, the graphs of Figure 13 are arranged in order of decreasing element mobility from Ba through Zr, or in order of their typical decreasing contribution to subduction zone magmas from a subducted slab (e.g., Pearce et al., 1995). In each case, the AFC modeling reveals that the Caribou Creek intermediate and felsic samples could have evolved from the basalt parent magmas by fractional crystallization with a moderate to high degree of crustal assimilation. With sample BC98-TY3A as the parent composition, a moderate ratio of assimilation to fractional crystallization (*r* value) of 0.4–0.5 satisfies most of the range of intermediate to felsic samples. A subset of data better fits a model in which sample GLA99-17 is the parent composition, and there is a higher degree of assimilation (*r* values of 0.8–0.95).

Overall, our geochemical models indicate that the Caribou Creek magmas could have inherited their arc-like geochemical characteristics (e.g., elevation of fluid-mobile elements) by contamination from Jurassic arc basement rocks, which are enriched in these types of subduction-related elements (Fig. 10). In this case, fluid flux from a subducted slab is not necessary to explain the subduction signature of the Caribou Creek rocks. Another example of continental margin volcanic rocks with a subduction signature (enriched LILE) inherited from crustal sources is the Colville Igneous Complex in Washington State (Morris et al., 2000).

The adakite-like rocks have a different petrogenetic history from the rest of the Caribou Creek volcanic rocks. The adakite-like rocks are strongly depleted in the HREEs (Fig. 9) and differ markedly in their chemistry from the enriched felsic rocks, including higher Ba/Yb, Th/Yb, La/Yb, and Zr/Yb ratios than can be explained with AFC models (Fig. 13). The depleted intermediate and felsic rocks are therefore not cogenetic with the other Caribou Creek volcanic rocks and are not associated with the same parental basaltic magmas. The strong depletion in the HREEs (less than 10× chondrite) is consistent with petrogenetic models for adakites that form by partial melting of metamorphosed basalt (amphibolite and eclogite) (Drummond and Defant, 1990). Adakites are commonly associated with the partial melting of young subducted oceanic lithosphere (Defant and Drummond, 1990, 1993), but adakite formation has also been attributed to melting of oceanic lithosphere along the edges of a slab window (Yogodzinski et al., 2001), melting of lower crustal rocks in continental collision zones (Chung et al., 2003), and melting of oceanic terranes that may form the upper plate of a subduction zone (Arculus et al., 1999).

The basement rocks beneath the Caribou Creek volcanic field include at least 3–4 km of garnet-bearing gabbroic and ultramafic rocks of the lower part of the Jurassic Talkeetna arc (Burns, 1985; DeBari and Sleep, 1991). Accordingly, the Caribou Creek adakite samples could have formed by partial melting of these crustal rocks (e.g., Arculus et al., 1999; Chung et al., 2003). Applying a simple equilibrium melting model shows that the highest Ba/Yb, Th/Yb, La/Yb, and Zr/Yb concentrations of the Caribou Creek adakites could be attained with less than 1% to ~5% partial melting of the gabbroic and ultramafic basement rocks (Fig. 13). A similar result is found in an equilibrium melting model using Sr and Y, which are diagnostic elements in characterizing adakites (Fig. 11). Kay et al. (1993) interpreted a similar range of partial melting (2%–5%) of oceanic rocks in the formation of backarc adakites in the southern Andean arc. The Caribou Creek adakite-like rocks, with lower Ba/Yb, Th/Yb, La/Yb, and Zr/Yb ratios, would require a higher percentage of partial melting of the gabbroic and ultramafic basement rocks (~15%–60%). Drummond and Defant (1990) and Defant and Drummond (1993) modeled a similar range of partial melting (<10% to ~50%–60%) of ultramafic oceanic rocks in the formation of several examples of Cenozoic adakites. Alternatively, these Caribou Creek adakitic magmas could have formed by mixing of low-percentage partial melts of the gabbroic and ultramafic basement rocks with nonadakitic Caribou Creek magmas. In this case, the adakite signature would have been diluted (e.g., Yogodzinski and Keleman, 1998) and higher percentages of partial melting would not be required. The low concentration of Sr (<400 ppm) in some of the Caribou Creek depleted dacites compared with typical adakites (>400 ppm) can be attributed to a small percent (<18%) of plagioclase fractionation (e.g., Prouteau et al., 1999).

TECTONIC IMPLICATIONS AND REGIONAL MAGMATISM

Tectonic and magmatic models to explain the origin of the Caribou Creek volcanic field need to account for: (1) a source of depleted basaltic magma, (2) eruption of depleted basalts with minimal crustal assimilation in order to retain low La/Yb ratios, (3) eruption of adakite-like rocks, (4) eruption closer to the paleo-forearc than regional magmatic arcs of south-central Alaska, and (5) eruption during early to middle Eocene time when there was a lull in regional margin-parallel (arc) magmatism across south-central Alaska. We propose that a mantle slab window along southern Alaska following migration of a ridge-trench-trench triple junction, coupled with regional crustal extension, can account for these characteristics.

Spreading Ridge Subduction Along Southern Alaska and Caribou Creek Volcanism

There is mounting geologic evidence that spreading ridge subduction was an integral process in the early Tertiary tectonic history of southern Alaska (e.g., Bradley et al., 2003). Plutons of the Sanak-Baranof belt, along with high-temperature–low-pressure metamorphic rocks of the Chugach Metamorphic terrane (Fig. 1), have long been recognized as representing an episode of near-trench high heat flow (Marshak

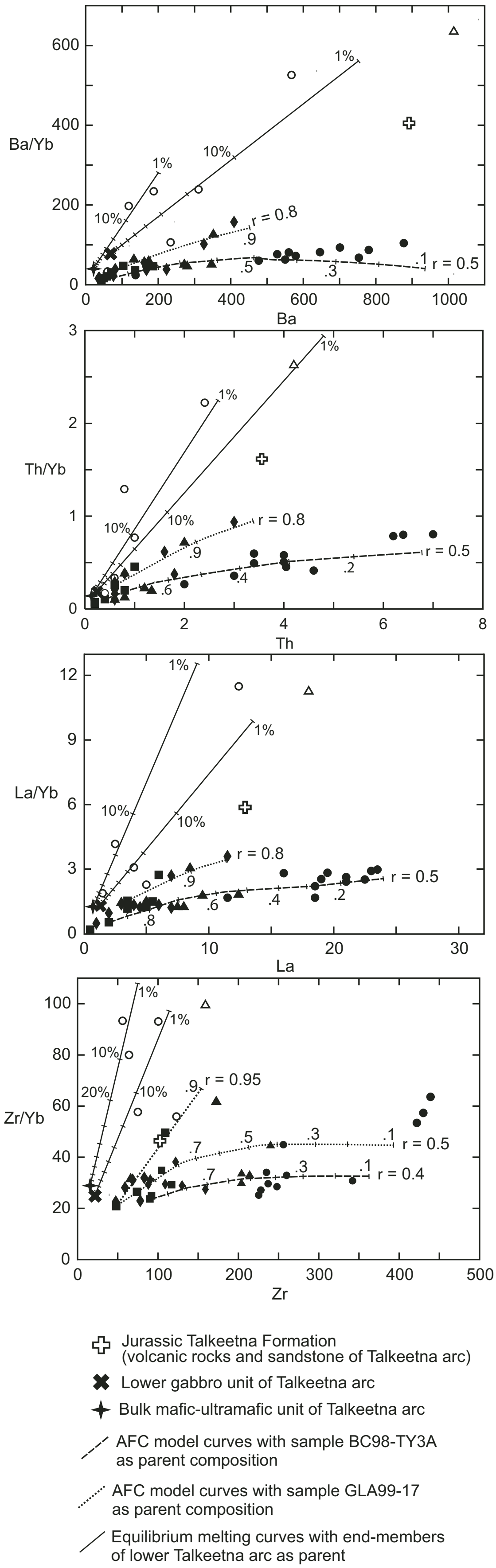

Figure 13. Graphs of Ba/Yb versus Ba, Th/Yb versus Th, La/Yb versus La, and Zr/Yb versus Zr for rocks of the Caribou Creek volcanic field and representative basement rocks of the Jurassic Talkeetna arc. Symbols for the Caribou Creek volcanic samples are as shown in Figure 8. Talkeetna arc compositions are compiled from Plafker et al. (1989), DeBari and Coleman (1989), and DeBari and Sleep (1991). Each graph displays the results of assimilation-fractional crystallization (AFC) and equilibrium melting models. The AFC model trends are for two representative parental basalt compositions (sample BC98-TY3A, dashed lines; and sample GLA99-17, dotted lines) undergoing assimilation of volcanic and sedimentary rocks of the Jurassic Talkeetna Formation. The *r* values (ratio of rate of assimilation/rate of crystallization) are given for each model curve. The tick marks show F values (percent of liquid remaining in 10% increments). The AFC models use the following changing composition of the fractionating liquid and corresponding bulk partition coefficients (D, written as X_D, where X = element): F—1.0–0.85 (8% olivine, 15% clinopyroxene, 1% orthopyroxene, 68% plagioclase, and 8% magnetite with $Ba_{0.16}$, Th_0, $La_{0.30}$, $Zr_{0.06}$, and $Yb_{0.24}$); F = 0.85–0.65 (1% olivine, 8% clinopyroxene, 8% orthopyroxene, 75% plagioclase, and 8% magnetite with $Ba_{0.18}$, Th_0, $La_{0.31}$, $Zr_{0.07}$, and $Yb_{0.23}$); F = 0.65–0.4 (2% clinopyroxene, 3% orthopyroxene, 75% plagioclase, 8% hornblende, 6% biotite, and 6% magnetite with $Ba_{0.51}$, $Th_{0.08}$, $La_{0.27}$, $Zr_{0.21}$, and $Yb_{0.20}$); F = 0.4–0.1 (40% quartz, 40% plagioclase, 8% K-feldspar, 8% biotite, 2% hornblende, 2% magnetite with $Ba_{1.4}$, $Th_{0.10}$, $La_{0.17}$, $Zr_{0.23}$, and $Yb_{0.23}$). The equilibrium melting curves are for different end-member compositions of the lower mafic-ultramafic portion of the Jurassic Talkeetna arc (as labeled on the graphs). Bulk partition coefficients of $Ba_{0.11}$, $Th_{0.05}$, $La_{0.10}$, $Zr_{0.12}$, and $Yb_{0.87}$ were based on an average mineralogy of 38% olivine, 5% garnet, 15% clinopyroxene, 5% orthopyroxene, 10% amphibole, and 25% plagioclase for the mafic and ultramafic rocks. The tick marks show percentages of melting in 10% increments except where labeled at the ends of curves. Mineral–melt partition coefficients in all models were compiled from Rollinson (1993; their Table 4.1) and Lytwyn et al. (2000).

and Karig, 1977; Hudson et al., 1979; Hudson and Plafker, 1982; Sisson et al., 1989; Hudson, 1994). A MORB-like magma source has been documented for several Sanak-Baranof intrusions (Hill et al., 1981; Moore et al., 1983; Harris et al., 1996; Lytwyn et al., 1997, 2000; Bradley et al., 2003; Sisson et al., 2003), which links the episode of high heat flow to a depleted mantle source of magmas. Based on a compilation of age data, Bradley et al. (1993, 2000) proposed that the Sanak-Baranof plutons are an eastward-younging belt that tracks the past migration of a ridge-trench-trench triple junction. In an alternative model, Hudson and Magoon (2002) view the same age data differently and interpret the western portion of the Sanak-Baranof belt as a coeval 61 ± 3 Ma thermal event and the eastern Sanak-Baranof belt as 53 ± 2 Ma thermal event. They attribute these thermal events to a slab window that formed due to a steepening and de-coupling of the subducting plate as opposed to the eastward migration of a spreading ridge. Haeussler et al. (2003b), however, outlined plate kinematic criteria against de-coupling of a subducted plate and in support of ridge subduction. Other evidence in support of ridge subduction includes deformation (Haeussler et al., 1995, 2003a; Pavlis and Sisson, 1995; Kusky et al., 1997, 2003; Pavlis et al., 2003), mineralization (Haeussler et al., 1995; Weinberger and Sisson, 2003), metamorphism (Sisson et al., 1989; Pavlis and Sisson, 1995; Bowman et al., 2003), sedimentation (Trop et al., 2003), and ophiolite emplacement (Bol et al., 1992; Crowe et al., 1992; Nelson and Nelson, 1993; and Kusky and Young, 1999) in accretionary prism and forearc rocks along southern Alaska. We favor the model of spreading ridge subduction as the mechanism to form a slab window beneath southern Alaska, but suggest that the spreading ridge may have been broken into multiple segments that intersected the subduction zone of southern Alaska at several locations. Such a model may help to resolve the alternative interpretations of the Sanak-Baranof age trends. For example, sets of similar-aged near-trench plutons could have formed at separate ridge-trench intersections, analogous to near-trench magmatism that occurred during the early Tertiary demise of the Farallon-Pacific spreading ridge along western California (Cole and Basu, 1995). In this case, sets of plutons, such as those in the western Sanak-Baranof belt, could have formed closely in time and their ages may not necessarily track the uniform migration of a single spreading ridge.

Past models for early Tertiary ridge-trench interactions along southern Alaska have invoked subduction of the Kula-Farallon spreading ridge. This is problematic though, because the Kula–Farallon–North America triple junction has been cited as the cause for early Tertiary geologic events both in southern Alaska and the Pacific Northwest (e.g., Marshak and Karig, 1977; Moore et al., 1983; Wells et al., 1984; Thorkelson and Taylor, 1989; Bradley et al., 1993; 2000; Breitsprecher et al., 2003). These two areas are ~2000 km apart, and so it is not possible for the same triple junction to exist in both places during the same time interval. Because the Kula plate has been almost entirely consumed by subduction (Engebretson et al., 1985; Lonsdale, 1988), there is no seafloor paleomagnetic record to resolve past positions of the Kula-Farallon spreading ridge. Several workers have proposed a viable solution for this disparity (Miller et al., 2002; Bradley et al., 2003; Haeussler et al., 2003b). In the absence of seafloor paleomagnetic data, they have compiled and outlined convincing geologic evidence that a separate oceanic plate, the Resurrection plate, existed in the northeast Pacific basin, which would permit separate ridge-trench-trench triple junctions to coexist during early Tertiary time both along southern Alaska and in the Pacific Northwest.

The ridge subduction model for southern Alaska provides a framework for Caribou Creek volcanism. The Caribou Creek basalts require partial melting of a depleted mantle source, similar to the source of mid-ocean-ridge basalts. Formation of a slab window in response to spreading ridge subduction would have allowed for upwelling of suboceanic upper mantle beneath the continental margin (e.g., Cole and Basu, 1995; Thorkelson, 1996). This depleted mantle reservoir could have undergone adiabatic melting to form the Caribou Creek basalts (facilitated by crustal extension, as described later). Assimilation of Jurassic arc crustal rocks would have imparted a subduction signature to the Caribou Creek volcanic rocks by the addition of LILE and fluid-mobile elements, which are in high abundance in the Jurassic rocks (Fig. 10). Lytwyn et al. (1997) described a similar model in which near-trench basalts of the southern Kenai Peninsula (Fig. 1) were derived from a depleted mantle source associated with ridge subduction, and which attained arc-like geochemical characteristics by assimilation of crustal rocks from the accretionary prism. In a different example, Gorring and Kay (2001) interpreted Neogene lavas in southern Patagonia to have formed in response to asthenospheric slab window processes associated with the collision of the Chile ridge with the Chile Trench. In this example, the asthenospheric source was enriched oceanic-island-basalt–like mantle instead of depleted MORB-like mantle, but similar to the Caribou Creek basalts, the Patagonia basalts attained arc-like geochemical characteristics by contamination from arc-modified continental lithosphere and/or from the supra-slab mantle wedge. Interestingly, in places where spreading ridges are adjacent to trenches today, such as the northern Juan de Fuca ridge and southern Chile ridge, there are ridge basalts that show arc-like geochemical characteristics (elevated fluid-mobile and large ion lithophile elements) (Klein and Karsten, 1995; Cousens et al., 1995) (Fig. 10). In these cases, the suboceanic mantle beneath the near-trench spreading ridges became contaminated by subduction zone elements. This could be facilitated with a slab window, which would provide a pathway between suboceanic mantle and metasomatized subarc mantle reservoirs (Klein and Karsten, 1995). At the time of Caribou Creek volcanism, a similar situation may have existed whereby depleted suboceanic mantle rising beneath a nascent slab window could have mixed with metasomatized mantle that had formed beneath the southern margin of Alaska during preceding subduction. The Caribou Creek basalts show a similar composition to the Chile ridge and Juan de Fuca ridge basalts (Fig. 10) and could have attained their fluid-mobile element enrichment in this manner (in addition to or in lieu of contamination from Jurassic basement rocks). Our AFC models indicate though that the Caribou Creek intermediate and felsic rocks did attain their enrichment in fluid-mobile and other incompatible elements by contamination from Jurassic basement rocks.

The adakite-like rocks of the Caribou Creek volcanic field are consistent with a slab window model. For example, the adakite magmas could have formed when high heat flow through the slab window (due to mantle upwelling) induced partial melting of garnet-bearing mafic-ultramafic basement rocks beneath the southern Talkeetna Mountains (e.g., Arculus et al., 1999; Chung et al., 2003) and/or melted the oceanic lithosphere along the edge of the slab window (e.g., Yogodzinski et al., 2001). In addition to the Caribou Creek adakitic rocks, other 50–52 Ma adakite-like rocks are found within the southern Alaska orocline. These rocks are depleted in HREEs, similar to the Caribou Creek depleted felsic rocks (Fig. 9) and are compatible with partial melting of a basaltic protolith (e.g., melting of oceanic crustal fragments within accretionary prism rocks of the Southern Margin composite terrane) (Barker et al., 1992). These other adakites provide further evidence for near-trench high heat flow, which is consistent with a model of slab window magmatism in southern Alaska. The formation of a slab window also provides a mechanism for renewed magmatism in southern Alaska, even while regional Late Cretaceous–early Tertiary arc magmatism was waning (Fig. 3).

Links Between Spreading Ridge Subduction, Accreted Terranes, and Late Paleocene–Eocene Magmatism in Southern Alaska

We can evaluate the slab window model for Caribou Creek volcanism by comparing the age and paleoposition of Caribou Creek volcanism and other igneous events with estimates for past positions of spreading ridges along southern Alaska. Bol et al. (1992) argued that the ca. 57 Ma Resurrection Peninsula ophiolite is a remnant of a spreading ridge that at 57 Ma, on the basis of paleomagnetic data, was located 13 ± 9° (equal to ~1745 ± 1205 km according to calculations of Bradley et al., 2003) south of its present position. In this scenario, the spreading ridge (and therefore the central part of the Southern Margin composite terrane onto which the ophiolite was emplaced) would have been more than 1000 km south of the Caribou Creek volcanic field at 57 Ma. Alternatively, the 95% confidence minimum estimate of Bol et al. (1992) places the Resurrection Peninsula ophiolite some 540 km south of its present position at 57 Ma. This is consistent with results of Kusky and Young (1999), who, on the basis of geologic data and tectonic reconstructions, interpreted that the Resurrection Peninsula ophiolite was transported 461 ± 79 km after its emplacement at 57 Ma. In this scenario, the spreading ridge (and therefore the central part of the Southern Margin composite terrane) would have been located just south of the position of the Caribou Creek volcanic field at 57 Ma (Fig. 14A). Accordingly, based on the ages of the Sanak-Baranof belt plutons that intrude the Southern Margin composite terrane, the spreading ridge would have migrated past the position of the Caribou Creek volcanic field between ca. 60 and 58 Ma, leaving a slab window in its wake (Fig. 14A).

This latter restoration involves northward margin-parallel displacement of the Southern Margin composite terrane by ~500 km or less since 57 Ma, which is consistent with displacement estimates for the Border Ranges–Hanagita fault system and for the adjacent Wrangellia composite terrane. The Border Ranges–Hanagita fault system, which separates the Southern Margin composite terrane from the Wrangellia composite terrane, originated as a Mesozoic subduction zone megathrust (Pavlis, 1982), which has been overprinted by Late Cretaceous and early Tertiary dextral slip (Little and Naeser, 1989; Roeske et al., 2003). On the basis of geologic mapping and geochronology, Roeske et al. (2003) interpreted a minimum of 600 km (possibly >1000 km) of dextral slip on the Border Ranges–Hanagita fault system in southeastern Alaska that began at ca. 85 Ma and was continuous between ca. 70 and 50 Ma. Subsequent dextral slip occurred on faults within the Southern Margin composite terrane (southward of the present-day Border Ranges–Hanagita fault), but the amounts of offset on these faults is not known (Pavlis and Sisson, 2003; Pavlis et al., 2003). Allowing the possibility for at least 1000 km of offset during the period of continuous slip on the Border Ranges–Hanagita fault (ca. 70–50 Ma) yields an average slip rate of 5 cm/yr. Because slip on the Border Ranges–Hanagita fault system accommodated most of the displacement of the Southern Margin composite terrane during this time (Roeske et al., 2003), we can use this average slip rate to estimate that the Southern Margin composite terrane moved northward by ~350 km since 57 Ma. Additional data to quantify the amount(s) of offset along faults within the Southern Margin composite terrane will serve to refine this estimate. This estimate is, however, consistent with the estimate for displacement of the Resurrection Peninsula ophiolite (e.g., Kusky and Young, 1999) and highlights that the Southern Margin composite terrane was most likely a few hundred kilometers south of its present position during late Paleocene to early Eocene time. This is also consistent with the interpretation of Little and Naeser (1989) and Little (1990) that the Border Ranges fault in the northern Chugach Mountains and Matanuska Valley areas (south of the Caribou Creek volcanic field) experienced less than 100 km of dextral slip during early Eocene time. A similar displacement history is found for the Wrangellia composite terrane. Paleomagnetic studies reveal that the Wrangellia composite terrane experienced 1650 ± 850 km of northward displacement between 80 and 55 Ma (Stamatakos et al., 2001) and no significant northward displacement since ca. 55 Ma (Hillhouse et al., 1985; Panuska et al., 1990). This equates to an average displacement rate of ~7 cm/yr and less than 200 km of northward motion of the Wrangellia composite terrane since 57 Ma. From these collective data, we surmise that there could have been large northward displacements (>1000 km) of the Southern Margin and Wrangellia composite terranes with respect to North America since Late Cretaceous time, and several hundred kilometers (<500 km) of northward motion since late Paleocene time.

We suggest then, in accordance with the Resurrection plate model (Haeussler et al., 2003b), that the northward position of the Resurrection Peninsula ophiolite during late Paleocene time indicates that it was formed and emplaced at the Kula-Resurrection ridge, instead of the Kula-Farallon ridge (Fig. 14A). While we have adopted the Resurrection plate model of Haeussler et al. (2003b), our interpretation is not dependent on this model. Our interpretation calls upon a depleted mantle source of magma explaced through a slab window, regardless of the mechanism for slab window formation. Nevertheless, with the restorations described herein and depicted in Figure 14A, there is good spatial and temporal correlation between the area of slab window formation due to ridge subduction and the onset of late Paleocene near-trench igneous activity and metamorphism across the Southern Margin and Wrangellia composite terranes in the southern Alaska orocline (Fig. 3). This includes the 59 Ma andesite adakite tuff sample in the Caribou Creek volcanic field (sample 99ANS-17A), which indicates that adakite magma formation was a nearly immediate response to slab window development. Younger Caribou Creek adakitic rocks (e.g., samples 00ANS4C and 99ANS9A) indicate that adakites continued to form over the duration of Caribou Creek volcanism (for ~20 m.y.). In another example, Kinoshita (2002) also documented a relatively long period of magmatism (~20 m.y.) that occurred above a slab window that formed in response to spreading ridge subduction beneath southwest Japan.

Crustal Extension and Caribou Creek Volcanism

Caribou Creek volcanism was coupled with an episode of crustal extension. The Caribou Creek volcanic field is crosscut by NW-SE–trending normal faults (Fig. 2), which are similar in trend to diabase dikes of the Matanuska Valley described by Fuchs (1980) and Little (1988). Synvolcanic fluvial deposits (Figs. 6 and 7; Table DR2 [see footnote 1]) at the base of the volcanic rocks thicken across some of these faults, and mafic dikes were intruded along some of these fault planes (Oswald et al., 2002). These observations indicate that the normal faults were active during Caribou Creek volcanism. While there is no direct link to specific faults, the rotated unconformities within synvolcanic fluvial deposits (Fig. 7C) lend support that an episode of deformation occurred at the onset of Caribou Creek volcanism. In addition, the nature of Caribou Creek volcanism itself is consistent with eruption in a zone of crustal extension. The main phase of Caribou Creek volcanism began with the outpouring of mafic and intermediate lavas, which was punctuated by small-volume intermediate to mafic pyroclastic eruptions (spatter and base-surge deposits) from maar or tuff cone style vents (Table DR2, see footnote 1). Overall, the predominance of lavas compared with pyroclastic deposits indicates a basaltic volcanic field in

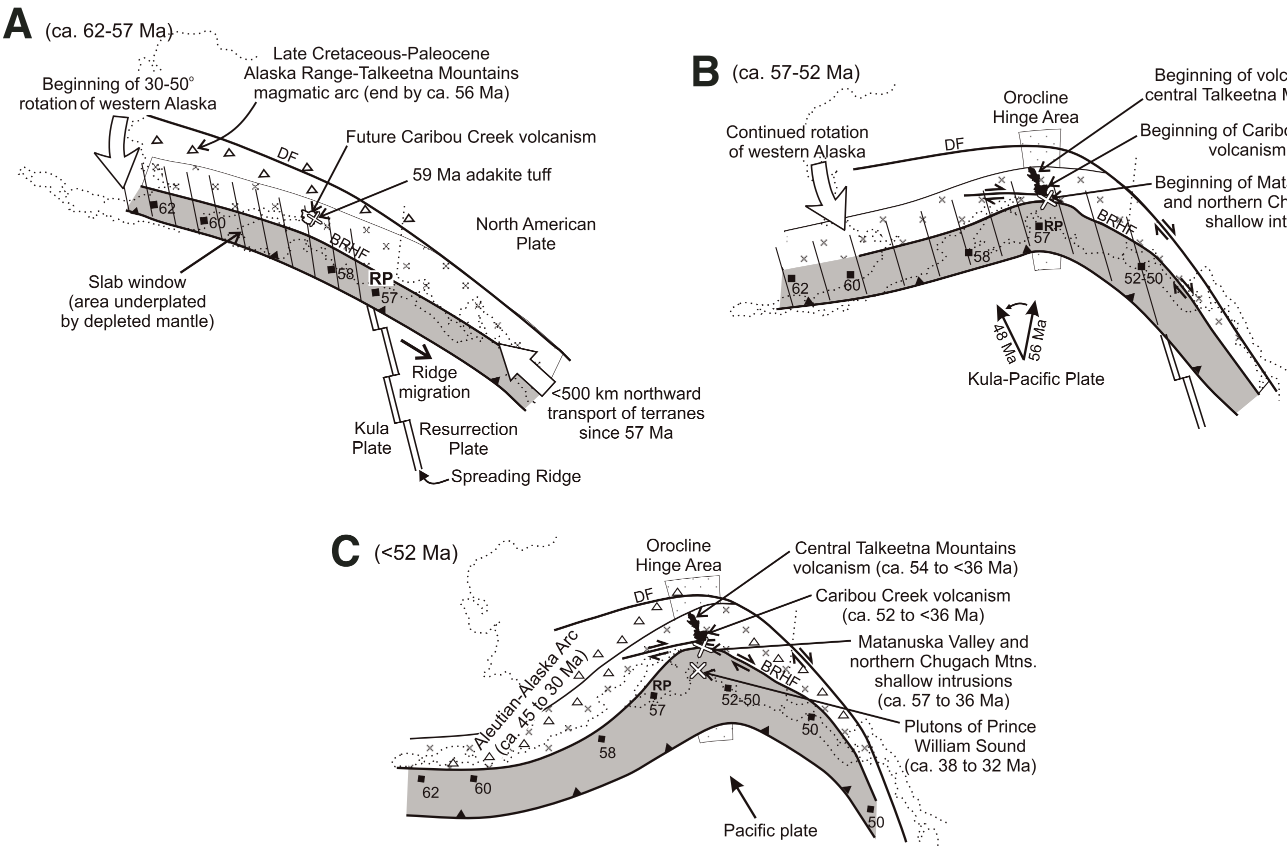

Figure 14. Tectonic reconstructions for southern Alaska, based on reconstructions of Wallace and Engebretson (1984), Engebretson et al. (1985), Lonsdale (1988), Bol et al. (1992), and Haeussler et al. (2003b). The restored position of the Southern Margin composite terrane (shaded region) is based upon the 95% minimum confidence limit of paleomagnetic data of Bol et al. (1992) and data of Kusky and Young (1999) for the Resurrection Peninsula ophiolite. The restored position of the Wrangellia composite terrane (x pattern) is based upon the combined paleomagnetic data of Hillhouse et al. (1985), Panuska et al. (1990), and Stamatakos et al. (2001). Representative near-trench plutons of the Sanak-Baranof belt are shown as black squares with their ages in Ma (from Bradley et al., 1993, 2000). Triangles represent subduction-related magmatic arcs. Abbreviations are as follows: RP—Resurrection Peninsula ophiolite; DF—Denali fault; BRHF—Border Ranges–Hanagita fault system. Other symbols are as shown in Figure 1. (A) 62–57 Ma reconstruction showing area along southern Alaska that was underplated by depleted mantle (diagonal lines) through a slab window, which formed during subduction and migration of the Kula-Resurrection spreading ridge. (B) 57–52 Ma reconstruction showing the coincidence between the onset of Caribou Creek volcanism, volcanism in the central Talkeetna Mountains, shallow near-trench intrusions of the Matanuska Valley and northern Chugach Mountains, and tectonic events that included oblique subduction beneath southern Alaska, counterclockwise rotation of western Alaska, and regional strike-slip faulting. (C) 52 Ma and younger reconstruction showing relationship between Caribou Creek volcanism, volcanic rocks in the central Talkeetna Mountains, 57 Ma and younger igneous rocks in the Matanuska Valley, Chugach Mountains, and Prince William Sound areas, and the hinge area of the southern Alaska orocline. See text for discussion.

which monogenetic vents were scattered along fissures (e.g., Swanson et al., 1975; Condit and Connor, 1996; Walker, 2000). Such types of basaltic volcanism are typically associated with regions of high differential stress and rifting (Takada, 1994). Magma emplacement through extended crust would have allowed the basalts to erupt with minimal crustal assimilation (e.g., Singer et al., 1992), which is consistent with their strong LREE depletion and their relatively low degrees of enrichment in LILE compared to the Caribou Creek felsic rocks (Figs. 8–10).

This extensional deformation was coeval with several regional tectonic events. First, early Eocene dextral and normal slip along the Caribou–Castle Mountain and Border Ranges fault systems in the Matanuska Valley area (Fuchs, 1980; Little and Naeser, 1989; Bunds, 2001) could have influenced extension in the Caribou Creek volcanic field. The NW-trending synvolcanic normal faults in the Caribou Creek volcanic field are consistent with extension in a zone of right-lateral simple shear along the adjacent Caribou strand of the Caribou–Castle Mountain fault system (Fig. 2). A similar interpretation was proposed by Fuchs (1980) to explain NW-trending dikes in the Matanuska Valley area. In addition, Little (1990) described a similar set of NW-SE–trending normal faults in the Matanuska Valley area. He proposed that these formed along a releasing bend of the Border Ranges fault system, which is another kinematic mechanism that may have caused extension in the area of the Caribou Creek volcanic field.

Second, crustal extension in the Caribou Creek volcanic field may have been, in part, a response to spreading ridge subduction. Sets of brittle faults in rocks of the Southern Margin composite terrane, including normal faults and conjugate strike-slip faults that trend perpendicular to regional strike, were coeval with Sanak-Baranof magmatism and indicate that the region above a subducting ridge will have a predominantly extensional and/or strike-slip structural history (Kusky et al., 1997; Bradley et al., 2003; Haeussler et al., 2003a). Accretionary prism and forearc rocks in places of modern ridge subduction (e.g., southern Chile, the Woodlark Basin), have a similar structural history, dominated by normal and strike-slip faulting (Forsythe and Nelson, 1985; Crook and Taylor, 1994). In the case of southern Alaska, it is not clear how far inboard this deformation extended, but Little and Naeser (1989) and Trop et al. (2003) have documented uplift and extension in the Matanuska Valley forearc region that is coincident with the timing of ridge subduction. If deformation related to ridge subduction extended inboard of the forearc region, then this could account for extension in the Caribou Creek region just prior to the start of Caribou Creek volcanism. Faults that formed at this time could then have been reactivated throughout Caribou Creek volcanism.

Third, and most speculatively, the locus of Caribou Creek volcanism and associated extension may have been influenced by counterclockwise rotation (oroclinal bending) of western Alaska. Paleomagnetic data from Late Cretaceous and early Tertiary igneous rocks reveal that western Alaska has rotated by ~30°–50° counterclockwise between ca. 68 and 44 Ma (Hillhouse and Coe, 1994), forming the southern Alaska orocline (Fig. 14). The oroclinal bend may have formed by some combination of rigid rotation of western Alaska (e.g., Hillhouse and Coe, 1994) and/or indentor tectonics (Scholl and Stevenson, 1991), and, together with oblique subduction following plate reorganization at ca. 52 Ma (Fig. 14B), could have been a driving force for strike-slip faulting in southern and western Alaska (as summarized by Miller et al., 2002). Note the coincident timing between the start of the main phase of Caribou Creek volcanism and plate reorganization, both occurring at ca. 52 Ma (Fig. 3). We have no field evidence to directly link crustal extension in the Caribou Creek volcanic field to plate reorganization or oroclinal bending, but Haeussler and Nelson (1993) described brittle faulting and block rotations in the orocline hinge area (Prince William Sound) that they relate to the kinematics of oroclinal bending. If oroclinal bending did cause brittle deformation, then perhaps this deformation acted to enhance crustal extension due to either strike-slip faulting and/or spreading ridge subduction in the hinge area of the southern Alaska orocline. Enhanced crustal thinning in the orocline hinge area could have allowed adiabatic melting of the depleted mantle reservoir of southern Alaska in a narrow, roughly N-S–trending zone, forming the Caribou Creek basalt magmas. Regionally, there is a close spatial and temporal relationship between the orocline hinge area and the magmatic events of the Caribou Creek volcanic field, the central Talkeetna Mountains, the Matanuska Valley, the northern Chugach Mountains, and Prince William Sound (Figs. 1, 3, and 14). In addition to the Caribou Creek volcanic rocks, some of these other igneous rocks show evidence for a depleted mantle source of magma, including the central Talkeetna Mountains volcanic rocks (Amos and Cole, 2003; Cole and Stewart, 2005) and the Matanuska Valley intrusions (Silberman and Grantz, 1984). The orocline hinge area may therefore have provided a unique pathway for magmatism along a zone that trended orthogonally to the continental margin and extended inboard of the forearc region.

CONCLUSIONS

1. The Caribou Creek volcanic rocks range in age from ca. 52 to younger than 36 Ma. A 59 Ma tuff with adakite geochemical characteristics marks eruptive activity that preceded the main phase of Caribou Creek volcanism. The 52–36 Ma phase of Caribou Creek volcanism began with mafic and intermediate lava flows with subordinate mafic and felsic pyroclastic eruptions in an extensional setting. This was followed by rhyolite and dacite lava flows, pyroclastic eruptions, and shallow intrusions. Some of the felsic lavas and intrusions have adakite-like geochemical characteristics.

2. A depleted mantle reservoir was the source of Caribou Creek basalt magmas. The basalt magmas evolved by fractional crystallization with moderate to high amounts of crustal assimilation from the Jurassic Talkeetna arc basement to form the andesite, dacite, and rhyolite magmas. The adakitic magmas did not evolve from the depleted basalt parent magmas, but were formed by partial melting of ultramafic-mafic basement rocks.

3. The depleted mantle reservoir was emplaced beneath the southern margin of Alaska through a slab window that formed during subduction and migration of a spreading ridge, possibly the Kula-Resurrection ridge. This slab window was the source of heat and mafic magmas for the Caribou Creek volcanic field, as well as other late Paleocene–Eocene igneous rocks in southern Alaska. The Caribou Creek magmas were erupted in a zone of crustal extension that formed in response to strike-slip faulting (driven by oblique plate subduction), ridge subduction, and/or rotation of western Alaska.

4. The Caribou Creek volcanic rocks provide new constraints on tectonic models for southern Alaska. First, the basalts of the Caribou Creek volcanic field are the most depleted Tertiary volcanic rocks yet documented along the southern margin of Alaska, inboard of accretionary prism rocks. The Caribou Creek basalts support the hypothesis that development of a depleted mantle slab window was important along the southern margin of Alaska during early Tertiary time. Second, the spatial and temporal relationship of these volcanic rocks with near-trench igneous rocks in both the Southern Margin and Wrangellia composite terranes favors estimates for northward displacement of these terranes of a few hundred kilometers as opposed to more than 1000 km since late Paleocene time. Third, the location of these volcanic rocks, as well as the location of late Paleocene to early Oligocene igneous rocks in the hinge area of the southern Alaska orocline (e.g., central Talkeetna Mountains, Matanuska Valley, northern Chugach Mountains, and Prince William Sound areas), supports the hypothesis that during rotation of western Alaska, the rotational hinge area exerted influence on magmatic processes. This model provides a solution for why late Paleocene through early Oligocene igneous activity of south-central Alaska occurred in a relatively narrow zone that extended inboard from a near-trench setting and is orthogonal to regional subduction-related arcs (Fig. 1).

5. Slab windows that form in response to spreading ridge subduction can cause magmatism in accretionary prism and forearc settings (e.g., Cole and Basu, 1995; Breitsprecher et al., 2003), but, as demonstrated with the Caribou Creek volcanic field, can also influence magmatism further inboard, given a suitable deformation regime (e.g., crustal extension). In addition, the Caribou Creek volcanic field shows that the magmatic effects of a slab window can be long-lived. Caribou Creek volcanism ranged between 59 and ca. 36 Ma, indicating that the influence of slab window volcanism in the southern Alaska orocline lasted for at least 20 m.y.

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