Paleozoic magmatism and crustal recycling along the ancient Pacific margin of North America, northern Cordillera\textsuperscript{1,2}

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Abstract

Devonian to Permian igneous rocks in the Yukon-Tanana terrane (YTT) record six cycles of arc, arc-rift, continental rift and back-arc basin magmatism, each set apart from the others by changes in the locus and/or character of igneous activity, as well as deformational episodes and unconformities. The first four cycles, from mid-Devonian to Late Mississippian, record largely bimodal arc magmatism above a west-facing (east-dipping) subduction zone, with or without accompanying back-arc basin magmatism and continental margin rifting. The fifth, Pennsylvanian-Early Permian cycle, involved more primitive, mafic to intermediate volcanism in a west-facing arc with a corresponding marginal back-arc basin to the east. The sixth, Late Permian cycle reflects subduction reversal, and continental-arc

\textsuperscript{1}Data Repository items Piercey\_DR1.xls (Table DR1), Piercey\_DR2.xls (Table DR2) are available on the CD-ROM in pocket.

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An important aspect of many mafic rocks from YTT intra-arc rifts and back-arc basins is their high Nb/Th (signatures de type A ou hyperalcalines). It is remarkable that there have been no systematic temporal variations in the composition of most mafic and felsic rocks. Igneous source regions and igneous processes were essentially unchanged throughout the tectonic history of the YTT.

Les roches ignées dévoniennes à permiennes du terrane de Yukon-Tanana (YTT) témoignent de la présence de six cycles de formation d’arc, d’ouverture d’un fossé tectonique d’arc, d’ouverture d’un fossé tectonique continental et d’un magmatisme de bassin d’arrière-arc, chacun se distinguant des autres par des changements dans l’emplACEMENT ou le caractère de l’activité ignée, ainsi que par des épisodes de déformation et des discordances. Les quatre premiers cycles, du Dévonien moyen au Mississipien supérieur, témoignent en gros d’un magmatisme d’arc bimodal au-dessus d’une zone de subduction orientée à pendage vers l’est, avec ou sans magmatisme d’arrière-arc et ouverture d’un fossé tectonique continental. Le cinquième cycle, du Pennsylvanien au Permien inférieur, affiche un volcanisme de composition mafique à intermédiaire, plus primitif, dans un arc orienté vers l’ouest avec un arrière-arc correspondant vers l’est. Le sixième cycle, au Permien supérieur, reflète un renversement de la subduction, avec un magmatisme d’arc continental associé à une zone de subduction à pendage vers l’ouest. Les roches mafiques de tous les cycles, qu’elles soient de caractère d’arc ou non-arc provenaient de sources diversément enrichies, avec des apports provenant d’un prisme de matériau mantélique appauvri ou de l’asthénosphère d’arrière-arc, et d’un manteau lithosphérique enrichi, avec ou sans composante de plaque subductée. Les roches felsiques de contexte géodynamique d’arc, d’ouverture de fossé tectonique d’arc et d’arrière-arc proviennent principalement de la fusion et du recyclage de matériaux de la partie supérieure de la croûte continentale (UCC; La/Sm$_{CN}$ ≈ 1). Les roches felsiques de contexte d’arc ont des signatures calco-alcalines et tholéiitiques, alors que les roches qui ne sont pas de contexte d’arc sont enrichies en éléments à forts effets de champ et d’éléments de la famille des terres rares (signatures de type A ou hyperalcalines). Il est remarquable qu’il n’y ait pas eu de variations systématiques temporelles dans la composition de la plupart des roches mafiques et felsiques, durant l’ensemble de l’histoire magmatique du YTT au Paléozoïque tardif, s’étendant sur plus de 150 millions d’années. Pour l’essentiel, les régions sources et les processus ignées sont demeurés inchangés pendant toute la durée de l’histoire tectonique du YTT.
INTRODUCTION

The Yukon-Tanana terrane (YTT) is a vast pericratonic province exposed in the central Cordillera of northern British Columbia, Yukon and eastern Alaska. The terrane lies between rocks of the cratonic margin of North America, to the east, and oceanic and arc terranes that were accreted to the YTT and cratonic margin in the Mesozoic, to the west (Fig. 1). It contains metamorphosed and polydeformed mid- to late Paleozoic volcanic, plutonic and sedimentary rocks whose stratigraphic record differs significantly from coeval strata of the North American miogeocline (Colpron et al., this volume-a). However, the rocks of the YTT have some geochemical, isotopic, metallogenic and geochronological characteristics that suggest that its older parts may have originated as part of the North American craton (Mortensen, 1992a; Creaser et al., 1997; Paradis et al., 1998; Nelson et al., 2002; Villeneuve et al., 2003; Mortensen et al., this volume). Both allochthonous and parautochthonous elements, with respect to the North American craton, make up the YTT (Fig. 1). Yukon-Tanana terrane locally hosts important volcanicogenic massive sulphide deposits.

Regional studies undertaken under the auspices of the Ancient Pacific Margin NATMAP (National Mapping Program) project have generated a wealth of new stratigraphic, geochronological and biostratigraphical data from Yukon-Tanana and affiliated terranes in northern British Columbia, Yukon and eastern Alaska (e.g., Colpron et al., this volume-a, and references therein). These data provide a stratigraphic framework that is critical to elucidating the tectonic evolution of the terrane and onto which detailed metallogenic, geochemical and isotopic studies can be based. In particular, geochemical and isotopic data allow for determining the relationship between the timing of magmatism, magma composition, and, ultimately, the tectonic setting in which ancient terranes have formed (e.g., Stern et al., 1995; Swinden et al., 1997). These data give insights into the relative roles that the mantle, continental crust and subducted slab play in the genesis of magmatic rocks. They provide important information on the processes governing crustal growth and assembly of continents (e.g., Pearce, 1983; Pearce and Peate, 1995; Pearce and Parkinson, 1993), and therefore on the Paleozoic evolution of YTT and the ancient Pacific margin of North America.

In this paper we present an overview and synthesis of the petrologic, geochemical and (in some cases) isotopic attributes of over 500 volcanic and intrusive rocks from YTT and related terranes (Fig. 1). Our presentation follows a series of time slices that correspond to documented major tectonic and magmatic events within the terrane. In spite of significant dynamothermal metamorphism throughout much of the terrane, primary igneous textures are preserved in many areas, and immobile element signatures have not been changed appreciably (e.g., Creaser et al., 1999; Dusel-Bacon and Cooper, 1999; Piercey et al., 2001a, b, 2002a, b, 2003). Accordingly, where initial compositions can be determined, the protolith names are used to emphasize the pre-metamorphic igneous evolution of these rocks. Many of the data sets analyzed in this paper are published either in previous papers (e.g., Creaser et al., 1997, 1999; Piercey et al., 2001a, b, 2002a, b, 2003, 2004; Colpron, 2001; Simard et al., 2003; Dusel-Bacon et al., 2004; Nelson and Friedman, 2004) or elsewhere in this volume (e.g., Dusel-Bacon et al., this volume); the reader is referred to these original studies for more detailed treatment of the data.

Given the breadth of new stratigraphic information, geochronologic and biostratigraphic age constraints, and associated geochemical and isotopic data, YTT may well become one of the best understood ancient continental margin geodynamic systems in the world.

METHODOLOGY AND APPROACH

This paper synthesizes over 500 high precision major, trace and rare earth element analyses, and, where available, Nd-isotopic data, from various locations throughout YTT; where applicable, data from the Stikine, Quesnel and Slide Mountain terranes are also discussed. Representative analyses are presented in Tables DR1 and DR2 (see footnote 1) and were compiled from published and unpublished sources. Most samples were analyzed by a combination of X-ray fluorescence (XRF), inductively-coupled plasma emission spectroscopy (ICP-ES), and inductively-coupled plasma mass spectroscopy (ICP-MS; see Piercey et al., 2001b for an example). Descriptions of instrumentation, analytical techniques, precision and accuracy are found in the original studies cited in Tables DR1 and DR2.

The following sub-sections explain the trace element diagrams used in this paper and the rationale behind their usage. We will discuss, where available, Nd isotopic signatures, because they highlight the relative importance of continental crust and mantle sources in the genesis of YTT magmatic rocks.

Petrology and Geochemistry

Trace element and radiogenic isotope geochemistry are useful in establishing the tectonic setting and petrogenesis of rocks in both modern (e.g., Pearce, 1983; Pearce and Peate, 1995; Pearce and Parkinson, 1993) and ancient (e.g., Stern et al., 1995; Swinden et al., 1997) geodynamic environments. Trace elements are more sensitive to petrological processes than major elements, and provide a process-oriented fingerprint or signature suggestive of a given tectonic environment (e.g., Pearce and Cann, 1973; Pearce and Norry, 1979; Shervais, 1982; Wood, 1980). Furthermore, most major elements (except Al, Si, Ti, O) are highly mobile during hydrothermal alteration and metamorphism (e.g., Gibson et al., 1983; MacLean, 1990). In contrast, the high field strength elements (HFESE: Zr, Hf, Nb, Ta, Y), rare earth elements (REE: La-Lu, except Eu), transition elements (TE: Cr, Ni, Sc, V), and the low field strength element (LFSE) Th are considered immobile during metamorphism (up to mid-amphibolite facies) and hydrothermal alteration at low water-to-rock ratios (e.g., Campbell et al., 1984; Whitford et al., 1988; You et al., 1996; Jenner, 1996; Swinden et al., 1997; Johnson and Plank, 1999). Previous geochemical studies of YTT also have confirmed that these elements are immobile during regional metamorphism and alteration (e.g., Creaser et al., 1997; Dusel-Bacon and Cooper, 1999; Piercey, 2001; Piercey et al., 2001a, b, 2002a, b, 2003, 2004; Dusel-Bacon et al., 2004).

In addition to fingerprinting tectonic setting, trace element data can provide insights into the nature and type of mantle (e.g., enriched,
Figure 1. Paleozoic lithotectonic terranes and assemblages of the northern Cordillera: AA – Arctic Alaska (includes Endicott Mountains, North Slope and Skagit allochthon); AG – Angayucham; CA – Cassiar; CO – Coldfoot (schist belt of southern Brooks Range); DL – Dillinger; IN – Innoko; MN – Minchumina; MY – Mystic; NA – North American miogeocline; NX – Nixon Fork; PC – Porcupine; RB – Ruby; SD – Seward; SM – Slide Mountain - Seventymile (includes Chatanika); ST – Stikine (Asitka); TZ – Tozitna; WM – Windy-McKinley; WS – Wickersham (includes Chena River, Fairbanks schist); YT – Yukon-Tanana. Areas discussed in this paper: (1) Alaska Range; (2) Yukon-Tanana upland; (3) Fortymile River; (4) Stewart River (including Dawson); (5) Finlayson Lake; (6) Glenlyon; (7) Wolf Lake – Jennings River (including Big Salmon Complex); (8) Sylvester allochthon; (9) Tulsequah; (10) Lay Range. Other abbreviations: Ak – Alaska; B.C. – British Columbia; D – Dawson; E – Eagle; Fb – Fairbanks; NWT – Northwest Territories; Wh – Whitehorse; WL – Watson Lake; T – Tok; YT – Yukon Territory. Blueschists and eclogite occurrences are from Dusel-Bacon (1994) and Erdmer et al. (1998).
depleted), crust (e.g., continental, oceanic) and subducted slab components involved in mafic and felsic rock genesis. In this paper, trace element data have been plotted on diagrams that reflect the tectonic setting of these rocks, and diagrams that provide insight into the nature of mantle, crustal and subducted slab components in YTT igneous rocks.

Rocks of mafic to intermediate and felsic to intermediate compositions are treated separately. For mafic to intermediate rocks, data are presented on primitive mantle (PM)-normalized multi-element diagrams that depict elemental abundances for a suite of elements from different chemical groups, including REE, HFSE and transition elements (Fig. 2). In contrast with mafic rocks, the significance of geochemical signatures for felsic volcanic and intrusive rocks can be problematic in continental margin environments, as they may merely reflect the continental crust itself (e.g., Piercey et al., 2001b). Nevertheless, there are subtle differences between granitoids from different tectonic environments (e.g., Piercey et al., 2001b), and these differences, used in conjunction with the compositional characteristics of coeval mafic magmatism, and the nature of volcanic and sedimentary facies (e.g., Piercey and Murphy, 2000), can provide significant insight into the origin of the felsic rocks. Upper continental crust (UCC)-normalized trace element plots (Fig. 3) and the Nb-Y diagram of Pearce et al. (1984; Fig. 4) are used to portray the chemical characteristics of felsic rocks. The Nb-Y diagram is particularly useful in deciphering the relative HFSE enrichment (non-arc) or HFSE depletion (arc) in felsic rocks, which can be used to differentiate arc from non-arc rocks (e.g., Piercey et al., 2001b, 2003; Fig. 4). This diagram, however, cannot separate rocks that have a true arc signature from those that have inherited an arc signature from interaction or melting of pre-existing arc crust; thus, when we describe “arc” signatures in felsic rocks we have relied on a combination of field relationships and the lithogeochemical signatures of both mafic and felsic magmatic rocks.

Arc and Non-Arc Geochemical Signatures

Mafic Geochemical Signatures

Mafic rocks from non-arc settings (e.g., mid-ocean ridges, ocean islands, etc.) are characterized by very smooth PM-normalized trace element patterns and have flat to positive anomalies of Nb relative to Th and La (Fig. 2A). In the non-arc mafic group, there are three broad subdivisions, including normal mid-ocean ridge basalts (N-MORB), enriched mid-ocean ridge basalts (E-MORB) and ocean island (or oceanic intraplate) basalts (OIB).

Normal-MORB (N-MORB) magmas are typified by flat PM-normalized signatures with depleted light rare earth elements (LREE), Nb and Th contents relative to PM (Fig. 2A). They are derived from incompatible element depleted mantle source regions (McKenzie and Bickle, 1988; McKenzie and O’Nions, 1991; Sun and McDonough, 1989) and are commonly found at spreading ridges, but can also be found in back-arc basin settings (e.g., Gribble et al., 1996; Sun and McDonough, 1989).

So-called ocean island basalt (OIB or within-plate) signatures, are characterized by steep PM-normalized patterns with LREE-enrichment, high HFSE contents, and, in particular, positive Nb anomalies relative to Th and La (Fig. 2A). Rocks with OIB signatures can be derived from a variety of sources, but it is generally assumed that they represent melts from incompatible element-enriched mantle

![Figure 2. Primitive mantle-normalized plots for representative examples of arc and non-arc mafic rocks. (A) normal-mid-ocean ridge basalt (N-MORB), enriched-mid-ocean ridge basalt (E-MORB) and ocean island basalt (OIB; Sun and McDonough, 1989); (B) boninite (BON; Jenner, 1981), island arc tholeiite (IAT; Piercey, 2001; Piercey et al., 2004), LREE-enriched IAT (L-IAT; Shinjo et al., 2000) and calc-alkaline basalt (CAB; Stoltz et al., 1990); (C) back-arc basin basalt (BABB; Ewart et al., 1994) and Th-enriched OIB (T-NEB; Shinjo et al., 2000). Primitive mantle values in this diagram, and all other primitive mantle normalized plots in this paper, are from Sun and McDonough (1989).](image-url)
sources, such as mantle plumes, in within-plate environments such as ocean island (e.g., Hawaii), oceanic plateau (e.g., Ontong-Java) and continental flood basalt environments (e.g., Lassiter and DePaolo, 1997; Sun and McDonough, 1989). These signatures can also be found in basalts derived from enriched lithospheric mantle source melts formed in plate-margin environments during the rifting of continental arcs or continent margins (e.g., van Staal et al., 1991; Piercey et al., 2002a, b; Dusel-Bacon and Cooper, 1999; Shinjo et al., 1999; Shinjo and Kato 2000). These types of basalts also occur in modern arcs associated with slab windows (e.g., Kamchatka) and Archean greenstone belts, and have been termed Nb-enriched basalts (Kepezhinskas et al., 1997; Wyman et al., 2000). In this paper, rocks with OIB signatures may be described as “within-plate” in places. This reflects their positions on discrimination diagrams and does

Figure 3. Upper continental crust (UCC)-normalized plots for felsic rocks from YTT and recent analogues (Cenozoic and younger). Typical UCC-normalized signatures for non-arc felsic rocks: (A) A-type and peralkaline rhyolites from YTT (Piercey et al., 2001b; Dusel-Bacon et al., 2004); (B) within-plate (A-type) felsic rocks from the Yellowstone Plateau (Hildreth et al., 1991) and Quaternary peralkaline rhyolites from Ethiopia (Peccerillo et al., 2003). Typical UCC-normalized signatures for arc felsic rocks: (C) tholeiitic rhyolite (ThR) and calc-alkaline rhyolite from YTT (CAR; Piercey et al., 2001b); (D) calc-alkaline rhyolite built on mafic crust (similar to tholeiitic rhyolites in YTT?) from Crater Lake, Oregon (Bacon and Druitt, 1988) and calc-alkalic rhyolites from the central Andes, Chile (Lindsay et al., 2001). Upper continental crust values for this diagram, and all other continental crust normalized plots in this paper, are from McLennan (2001).

Figure 4. The Nb-Y discrimination diagram for felsic rocks from Pearce et al. (1984). Symbols and data sources are as in Figure 3.
not imply that they formed in a within-plate setting. Most, if not all, of these rocks formed along plate margins due to continental or arc rifting.

Enriched-MORB (E-MORB) signatures are hybrid signatures between N-MORB and OIB (Fig. 2A) due to mixtures of magmas from depleted and enriched mantle sources (Sun and McDonough, 1989). Enriched-MORB magmas can be found at so-called plume-centered spreading ridges (e.g., Iceland; Sun and McDonough, 1989), spreading centres with heterogeneous plum pudding-type mantle (e.g., East Pacific Rise; Niu et al., 1999), where there is mixing of enriched (OIB) “plums” within a depleted (N-MORB-type) mantle matrix, and can also be found in rocks derived from the mixing of enriched lithospheric/asthenospheric sources during the evolution of continental rifts and back-arc basins (e.g., Piercey, 2001; Piercey et al., 2002a).

Mafic to intermediate arc rocks, unlike non-arc rocks, do not have smooth PM-normalized signatures. They generally (although not universally; e.g., Piercey et al., 2002a) show a distinctive negative Nb anomaly relative to Th and La, the so-called “arc signature” (Fig. 2; Swinden et al., 1997). Arc volcanic rocks, like non-arc rocks, are derived from similar variably enriched mantle sources; however, unlike non-arc rocks, they have an additional component, the slab component, superimposed upon the mantle wedge (Pearce and Peate, 1995, and references therein). During the subduction of oceanic lithosphere, dehydration of hydrous silicate minerals within the slab (and from sedimentary rocks atop it) results in the transfer of fluid-mobile elements to metasomatize the sub-arc mantle wedge (e.g., Pearce, 1983; Pearce and Parkinson, 1993; Pearce and Peate, 1995; You et al., 1996; Johnson and Plank, 1999). Due to the high mobility of the LFSE in fluids, arc rocks are typically enriched in these elements; Th, the relatively immobile LFSE, also involved in slab metasomatism, is used here as the indicator of the slab-derived fluid flux. Thorium, however, can also be enriched in mafic rocks contaminated by continental crust. In these instances, Th often shows systematic relationships with other indicators of crustal contamination (i.e., tNd, SiO2, Zr), and these geochemical relationships were used to screen our data in order to identify samples that may have been influenced by crustal contamination versus samples that have a subducted slab signature (see Piercey et al., 2002a, 2004, for examples). In addition to Th-enrichment, arc rocks are typically characterized by HFSE depletions, in particular Nb, that result from retention of HFSE in accessory minerals within the subducted slab (e.g., rutile; Foley et al., 2000), which can be intensified by HFSE-depletions due to previous melting events in the sub-arc mantle wedge (e.g., Pearce et al., 1992; Woodhead et al., 1992). The elevated LFSE and depleted HFSE signatures lead to high LFSE/HFSE (Th/Nb) ratios in arc rocks (Fig. 2B). Within the arc group of YTT mafic rocks, there are broadly four common signatures: island arc tholeiites (IAT), LREE-enriched island arc tholeiites (L-IAT), calcalkaline basalts (CAB) and rare boninites (BON; Fig. 2B).

Island arc tholeiitic rocks represent melts of a source similar to N-MORB but include a subduction zone metasomatic component. The IAT suites have flat PM-normalized patterns and are characterized by a negative Nb anomaly relative to Th and La, HFSE depletion (Fig. 2B), and follow a tholeiitic (Fe-enrichment) differentiation trend (Swinden et al., 1997); in some cases they have very low Ti contents (Brown and Jenner, 1989). These suites of rocks are commonly associated with the early construction of island arcs, usually before the arc edifice accumulated up to a significant thickness.

Calc-alkaline basalts and andesites typically reflect the mature stages of arc development coincident with arc edifice growth to a significant thickness. They are also the most common mafic-intermediate rocks in arcs built on or near continental crust (Swinden et al., 1997). The CAB suite is characterized by PM-normalized patterns with LREE-enrichment, very strongly developed negative Nb and Ti anomalies (Fig. 2B), and calc-alkaline fractionation trends (e.g., early Fe-depletion). Calc-alcaline suites can also be formed when tholeiitic magmas, and possibly N-MORB magmas, become contaminated by continental crust early in their petrogenetic history (e.g., Swinden et al., 1997). Calc-alkaline magmatic suites can also be differentiated from tholeiitic suites by high Zr/Y, La/Yb and Th/Yb values (see Barrett and MacLean, 1999, and references therein), all of which can be observed on a standard primitive mantle normalized plot (Fig. 2).

The L-IAT suite of magmas is believed to be a hybrid and part of a continuum between the IAT and CAB suites, and likely represents either derivation from an E-MORB source with a subduction component (Shinjo and Kato, 2000; Shinjo et al., 1999), or a weakly crustally contaminated IAT (Piercey, 2001; Piercey et al., 2004).

Boninitic rocks are unique magmatic rocks that reflect derivation from high temperature melting of ultra-depleted mantle sources (i.e., more depleted than N-MORB). They commonly form during the initiation of island arcs or back-arc basins (Pearce et al., 1992; Crawford et al., 1989; and references therein), where they are commonly spatially associated with IAT. Although generally occurring in fore-arc environments within intra-oceanic arcs, in YTT boninites are associated with the initiation of intracontinental (ensialic) backarc basin magmatism (Piercey et al., 2001a). Boninitic rocks are characterized by U-shaped PM-normalized trace element patterns with negative Nb anomalies, middle-REE depletions, very low TiO2, HFSE and REE contents, high compatible element contents (Cr, Ni, Co, Sc, V), and often, but not always, they have positive Zr and Hf anomalies relative to Sm (Fig. 2B).

In addition, YTT also contains mafic to intermediate magmatic rocks that have geochemical signatures transitional between arc and non-arc. Typically they have non-arc affinities but have a weak subduction signature, these include: back-arc basin basalts (BABB) and Th-enriched, Nb-enriched basalts (T-NEB). The BABB suites are essentially similar to N-MORB but have a weak negative Nb anomaly (Fig. 2C). Similarly, the T-NEB suite has a signature very similar to OIB but with a flat to weakly negative Nb anomaly (Fig. 2C). The BABB suite commonly form during back-arc basin development and represent MORB-type magmatism with minor subduction zone fluid influence (Hawkins, 1995; Gribble et al., 1996; Piercey et al., 2004). The T-NEB suite is interpreted to represent arc rift rocks with a minor subduction signature (Kepezhinskas et al., 1997; Piercey et al., 2004).
Felsic Geochemical Signatures

Delineating arc and non-arc signatures in felsic rocks is very difficult. Commonly felsic rocks have non-distinctive geochemical signatures that may be derived in whole or in part from melting pre-existing continental crust (Piercey et al., 2001b). In order to understand the origin and setting of felsic rocks in YTT, an integrated approach is required. This approach includes a detailed documentation of volcanic and sedimentary facies, and using the geochemical signatures of associated mafic rocks. This approach has led to establishing empirical relationships for arc and non-arc felsic assemblages within YTT (e.g., Piercey et al., 2001b). In general, non-arc felsic rocks from YTT have PM- and UCC-normalized REE patterns that are somewhat different than that of arc rocks (Fig. 3). Non-arc rocks have higher total REE and HFSE contents and lower compatible element contents (Sc, V, Ti, Ni, Cr) and are similar to crustally-derived A-type and peralkaline felsic rocks (Figs. 3, 4; Piercey et al., 2001b, 2003; Dusel-Bacon et al., 2004). Shown for comparison on Figure 3B are within-plate (A-type) felsic rocks from the Yellowstone caldera (Hildreth et al., 1991) and Quaternary peralkaline rhyolitic rocks from Ethiopia (Peccerillo et al., 2003). Although some of the felsic rocks within this paper are described as having “within-plate” signatures, this reflects their position on a discrimination diagram, and does not imply that they formed in a within-plate setting. Most, if not all, of these rocks occur along plate margins due to continental or arc rifting.

Arc felsic rocks are present in many parts of the terrane throughout its evolution (Mortensen, 1992a; Colpron et al., this volume-a, and references therein). They typically are of crustally-derived, calc-alkaline affinity (e.g., Mortensen, 1992a; Piercey et al., 2003; Grant, 1997). Their UCC-normalized patterns are relatively flat, often with depletions in Ti, Sc, V, and in some cases Th ± Nb (Fig. 3), features consistent with derivation from upper crustal sources accompanied by oxide and/or accessory mineral fractionation (Piercey et al., 2001b, 2003). Compared to the non-arc suites in YTT, the calc-alkaline arc suites have lower HFSE and REE contents with volcanic-arc (I-type) signatures (Figs. 3, 4). Tholeiitic arc felsic rocks are rare within YTT (Grant, 1997; Piercey et al., 2001b, 2003). They have LREE- and HFSE-depleted UCC-normalized patterns when compared to calc-alkaline arc and A-type felsic rocks (Figs. 3, 4; Piercey et al., 2001b). The tholeiitic arc felsic rocks are interpreted to have been derived from melting of primarily mafic to andesitic crust more juvenile than the source for the A-type and calc-alkaline suites (Piercey et al., 2001b, 2003). Illustrated for comparison are arc felsic rocks built upon a thick crust of accreted oceanic crustal material from Mount Mazama, Crater Lake, Oregon (Bacon and Druitt, 1988), and arc felsic rocks erupted in a continental arc setting with thick crust (~calc-alkaline) from the Central Andes, Chile (Fig. 3; Lindsay et al., 2001). Notably, there are no transitional felsic suites in YTT.

Sm-Nd Isotopic Signatures: Crust vs. Mantle Components in Magmas

Samarium-neodymium isotope geochemistry is commonly used in ancient belts to decipher the relative contributions of crust and mantle in igneous rocks, because the Sm-Nd system is very resistant to alteration and metamorphism, and, in most cases, is insensitive to fractionation during magma crystallization or partial melting (e.g., DePaolo, 1988). This technique is based on the radiogenic decay of $^{147}$Sm to $^{143}$Nd; both of these isotopes are commonly presented as a ratio to the nonradiogenic $^{144}$Nd (e.g., $^{147}$Sm/$^{144}$Nd, $^{143}$Nd/$^{144}$Nd). The model isotopic evolution of the Sm-Nd system involves the separation of continental crustal (CC) and depleted mantle (DM) reservoirs from a chondritic uniform reservoir (CHUR). In the case of the DM reservoir, the separation from a CHUR reservoir resulted in a higher Sm/Nd ratio in DM than in CHUR reservoir, which leads to a time-integrated increase in $^{143}$Sm/$^{144}$Nd relative to CHUR (Fig. 5). In contrast, continental crust evolved with a lower Sm/Nd ratio, slower increase in $^{147}$Sm and, by association, resulted in a time-integrated $^{147}$Sm/$^{144}$Nd ratio lower than CHUR (Fig. 5A). Neodymium-isotopic data are commonly presented in the shorthand form, epsilon notation, which represents the $^{143}$Sm/$^{144}$Nd variation of a rock relative to CHUR at a given point in the past (i.e., $\varepsilon$Nd; Fig. 5B). Rocks derived from crustal reservoirs, or with recycled crustal components, have $\varepsilon$Nd <0, whereas rocks derived from depleted mantle-type reservoirs have $\varepsilon$Nd >0; values from chondritic reservoirs have $\varepsilon$Nd ~ 0 (Fig. 5B).

![Figure 5. Isotope evolution diagrams for the Sm-Nd isotope system. Modified from Swinden et al. (1997) based on concepts outlined in DePaolo (1988; and references therein). See text for discussion. CHUR = chondrite uniform reservoir.](image)
MAGMATIC AND PETROLOGIC EVOLUTION

The magmatic and tectonic evolution of Yukon-Tanana and related terranes in the northern Cordillera is recorded in six Paleozoic magmatic cycles (Tables 1 and 2; see also Colpron et al., this volume-a; Nelson et al., this volume). The magmatic cycles include two pulses of felsic (plus mafic and intermediate) magmatism, separated by Pennsylvanian – Early Permian mainly intermediate to mafic activity (Fig. 6). The first pulse of felsic magmatism in YTT corresponds primarily to voluminous Late Devonian to Mississippian igneous rocks of the Finlayson and lower Klinkit assemblages (Colpron et al., this volume-a). It contains four distinct cycles (I-IV; Fig. 6) that are punctuated by at least two episodes of deformation and erosion. The Pennsylvanian to Early Permian lull in felsic magmatism corresponds to a cycle dominated by intermediate to mafic magmatism and basinal sedimentation in the Slide Mountain and upper Klinkit assemblages (Cycle V; Fig. 6). The second pulse of felsic magmatism, and the last Paleozoic magmatic cycle, is represented by more localized Middle to Late Permian magmatism of the Klondike assemblage (Cycle VI; Figs. 1, 6).

Ecstall Cycle (Cycle I - Middle to Late Devonian - 390-365 Ma)

The first magmatic cycle is most widespread in the Alaska Range and Yukon-Tanana Upland of eastern Alaska (Fig. 1). Coeval felsic volcanism of probable within-plate affinity is locally present in mio-geoclinal strata of Selwyn basin in central Yukon (Hunt, 2002), and magmatism of probable arc affinity is documented in the Coast Mountains of British Columbia and southeastern Alaska, including parts of the Tracy Arm and Endicott Arm assemblages (Gehrels et al., 1992) and the Ecstall belt (Gareau and Woodsworth, 2000; Alldrick and Gallagher, 2000; Alldrick, 2001; Figs. 1, 7). Massive sulphide occurrences and deposits of the Ecstall belt, Selwyn basin and the Bonnifield district and Delta mineral belt formed during this cycle.

Limited trace element data presented by Dashevsky et al. (2003) for the Delta mineral belt of the Alaska Range suggest that these rocks formed in a continental arc setting. In the Coast Mountains, the predominance of tonalitic plutons, and limited isotopic data from the Ecstall belt, also suggest a continental arc setting for these rocks (Table 1; Gareau and Woodsworth, 2000). However, because the dataset from the Delta mineral belt is limited, and because high-precision geochemical data are not available for mafic rocks of the Ecstall belt, Tracy Arm and Endicott Arm assemblages, these rocks will not be discussed further here.

Cycle I magmatism in the Alaska Range and Yukon-Tanana Upland likely occurred in a continental rift setting (Dusel-Bacon et al., this volume-a; Figs. 1, 7). In the Bonnifield mining district of the Alaska Range, the Late Devonian to earliest Mississippian stratigraphic sequence consists of a varying successions (Healy schist, Keevy Peak Formation, Totatlanika Schist and Wood River assemblage) of felsic and mafic metavolcanic and shallow intrusive rocks associated with variably carbonateous metasedimentary rocks (Wahrhaftig, 1968; Dusel-Bacon et al., 2003, this volume). Mafic metavolcanic rocks within the Totatlanika Schist (Moose Creek and Chute Creek members) have alkalic, OIB-like PM-normalized signatures, with the one Moose Creek sample exhibiting a minor negative Nb and Ti anomaly (Fig. 8A). These features are consistent with derivation from a moderately enriched mantle source region; the Moose Creek sample exhibits minor crustal contamination. Felsic rocks in the Bonnifield mining district occur throughout the succession from the Healy Schist through the Totatlanika Schist (Dusel-Bacon et al., 2004, this volume). Most felsic rocks, with the exception of the Mystic Creek Member of the Totatlanika Schist and some Wood River assemblage samples, have very similar calc-alkaline patterns on UCC-normalized plots but with differing HFSE and REE abundances (Fig. 9A). In contrast, the Mystic Creek member contains a very distinctive suite of peralkaline rhyolites with very high HFSE and REE contents and positive Nb anomalies (Fig. 9A).

In the Yukon-Tanana Upland, Cycle I comprises bimodal mafic (amphibolites) and felsic (augen gneiss, felsic schist) meta-igneous rocks associated with variable amounts of metasedimentary rocks (quartzite, pelite, marble, phyllite of the Lake George assemblage; Weber et al., 1978; Smith et al., 1992; Dusel-Bacon et al., 2003). Mafic rocks in the Yukon-Tanana Upland show mostly OIB-like signatures (Fig. 8B), with some samples (unit Dag) exhibiting weak negative Nb anomalies that are likely due to crustal contamination of an OIB-like magma (Dusel-Bacon and Cooper, 1999; Dusel-Bacon et al., 2004, this volume). Felsic rocks in the Yukon-Tanana Upland exhibit greater variability, but generally have flat UCC-normalized
multi-element patterns with depletions in Ti, Al, Sc, V (±Zr, Hf and Eu) that are consistent with derivation from melting of continental crust accompanied by feldspar and oxide fractionation (Fig. 9). Data for these rocks cluster close to the boundary between within-plate and volcanic arc fields, with the exception of two felsic samples from the Nasina assemblage, which have peralkaline affinities and within-plate signatures (Fig. 9). On UCC-normalized plots, some felsic rocks from the Yukon-Tanana Upland (units Pzsq, MDmg and Butte assemblage of Dusel-Bacon et al., 2004) all have similar relatively flat calc-alkalic patterns, with depletions in Ti, Al, Sc, V (±Zr, Hf and Eu) consistent with derivation from melting of continental crust accompanied by feldspar and oxide fractionation (Fig. 9). In contrast, two samples of felsic rocks from the Nasina assemblage are much more depleted in LREE, but are highly enriched in Nb, Zr, Hf and HREE contents (Figs. 9D, E).

The distinctive occurrence of peralkaline rocks in the Bonnifield mining district, the dominance of OIB-like mafic rocks, the lack of mafic rocks with arc signatures, and the dominance of metasedimen-
Paleozoic magmatism and crustal recycling

Table 1. Summary of geochemical attributes of arc assemblages for Paleozoic magmatic cycles in YTT.

<table>
<thead>
<tr>
<th>Cycle</th>
<th>Age Range (Ma)</th>
<th>Location*</th>
<th>Geochemical Signatures</th>
<th>Isotopic Signatures</th>
<th>Sources of Data</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Delta mineral belt, Alaska Range</td>
<td>Felsic: No geochemical data; mafic to intermediate (minor felsic) metavolcanic and intrusive rocks; meta-tonalite plutons.</td>
<td>Felsic: εNd = -4.8; Mafic: εNd = -5.1 to -7.0.</td>
<td>N/A.</td>
</tr>
<tr>
<td>II</td>
<td>365-357 Ma</td>
<td>Finlayson Lake (5) (Fig. 10)</td>
<td>Felsic: Calc-alkaline intrusions.</td>
<td>Felsic: εNd = -7.4 to -12.9.</td>
<td>S. Piercey and J. Ryan (unpublished data).</td>
</tr>
<tr>
<td></td>
<td>357-342 Ma</td>
<td>Finlayson Lake (5) (Simpson Range suite)</td>
<td>Felsic: Calc-alkaline metaluminous intrusions; Mafic: L-IAT, L-IAT, MORB, NEB and BON.</td>
<td>Felsic: εNd = -2.5 to -6.2; Mafic: εNd = +4.1 to +6.4.</td>
<td>Piercey et al. (2001a, b, 2002b, 2003; Grant (1997).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Teslin (7)</td>
<td>Felsic: Calc-alkaline andesites, rhyolites, and intrusive rocks; Mafic: OIB-like mafic volcanic rocks.</td>
<td>Felsic: εNd = 0.1 to -4.8; Mafic: εNd = -2.5 to -6.2;</td>
<td>Stevens et al. (1995); Creaser et al. (1997).</td>
</tr>
<tr>
<td>IV</td>
<td>314-269 Ma</td>
<td>Glenlyon (6) (Little Salmon fm. and Tutmain suite)</td>
<td>Felsic: Calc-alkaline metaluminous intrusions; Mafic: rift-like mafic rocks?</td>
<td>Felsic: εNd = 0.1 to -4.8; Mafic: εNd = -2.5 to -6.2;</td>
<td>Piercey et al. (2001a, b, 2002b, 2003; Grant (1997).</td>
</tr>
<tr>
<td>VI</td>
<td>253-237 Ma</td>
<td>Wolf-Jennings (8) (Klondike suite; Sulphur Creek orthogneiss)</td>
<td>Felsic: Calc-alkaline intrusion.</td>
<td>Felsic: εNd = +6.7 to +7.4.</td>
<td>S. Piercey and J. Mortensen (unpublished data); Mihalynuk (unpublished data).</td>
</tr>
</tbody>
</table>

Note: * Bold characters indicate areas for which data are summarized in this paper; numbers refer to locations shown in Figure 1.

Finlayson Cycle (Cycle II - Late Devonian to Early Mississippian - 365-357 Ma)

The second magmatic cycle marks the onset of felsic magmatism in many parts of YTT and the formation of massive sulphide deposits in the Finlayson Lake district (Figs. 6, 10). Magmatic activity, which began during Cycle I in the Alaska Range and Yukon-Tanana Upland persisted during Cycle II (Fig. 10). Localized felsic magmatism of probable within-plate affinity also occurs in miogeoclinal strata of the Selwyn basin and Cassiar terrane (Figs. 1, 10). In YTT, this cycle...
ended with a deformational event recorded in rocks of the Finlayson Lake district (Murphy et al., this volume). This deformational event is recorded by fabric development, uplift and erosion, and is interpreted to represent contractional deformation in an intra-arc setting (Murphy et al., this volume).

The Finlayson cycle is characterized by arc, arc-rift and back-arc basin magmatism in most of YTT, with the exception of the Bonnifield district and Yukon-Tanana Upland, where magmatic activity is interpreted to have occurred in an extended continental margin setting (Dusel-Bacon et al., 2004; Fig. 10). Coeval back-arc basin magmatism occurred predominantly in the eastern Finlayson Lake district and in the Slide Mountain terrane in the Sylvester allochthon (Fig. 10).

The Finlayson Lake district is divided into two contrasting Devonian-Mississippian magmatic regions, which are juxtaposed along the Permian Money Creek thrust and associated faults (see Murphy et al., this volume). Rocks of arc affinity occur primarily to
the southwest, in the hanging wall of the thrusts; coeval back-arc facies occur exclusively to the east, in the footwall. Cycle II arc magmatism is represented by the ~365-360 Ma Waters Creek and Cleaver Lake formations, which comprise mafic volcanic, volcanioclastic and intrusive rocks, with lesser felsic volcanic and sedimentary rocks (Mortensen, 1992a, b; Murphy, 1998, 2001; Murphy and Piercey, 1999; Murphy et al., 2002, this volume). Mafic arc rocks in the Waters Creek and Cleaver Lake formations are geochemically diverse, with signatures that include BON, IAT, L-IAT and CAB (Fig. 11A). Their source magmas were derived from a variably enriched mantle wedge, with components from subducted slab and/or crustal contamination (Fig. 11; Piercey, 2001; Piercey et al., 2001a, 2004). They have εNd values that range from −5.0 to +7.1; however, the bulk of the samples (6 of 8) have εNd > 0, reflecting derivation from chondritic to juvenile sources (Piercey, 2001; Piercey et al., 2004). Volumetrically minor felsic rocks of the Cleaver Lake formation have tholeiitic and calc-alkaline affinities, consistent with an arc parentage (Figs. 12A, B; Piercey et al., 2001b, 2003). Neodymium isotopic data for the felsic rocks is limited. A tholeiitic rhyolite sample has εNd = +0.1, and a calc-alkaline rhyolite sample has εNd = −4.8 (Piercey et al., 2003). These variations in Nd isotopic attributes, coupled with various trace element ratios, suggest that the YTT in this region was built upon a heterogeneous basement of both oceanic and continental basement (Grant, 1997; Piercey et al., 2001a, 2003; see also Murphy et al., this volume).

In the Finlayson Lake district (Figs. 1, 10), bimodal back-arc magmatic activity is represented by non-arc and transitional volcanic and intrusive rocks of the Grass Lakes group (Fire Lake, Kudz Ze Kayah and Wind Lake formations; Murphy et al., this volume). The stratigraphically lower Fire Lake formation is dominated by mafic volcanic and plutonic rocks with non-arc to transitional signatures including E-MORB, back-arc basin basalts, OIB and Th-rich OIB which are likely crustally contaminated or slab-influenced OIB (Fig. 11B; Piercey, 2001; Piercey et al., 2004). Neodymium isotopic data for these rocks suggest derivation from mantle sources with near-chondritic to depleted affinities (εNd = −1.6 to +8.5), with the lower εNd values associated with the most enriched rocks, suggesting the influence of lithospheric mantle within these suites (Piercey, 2001; Piercey et al., 2004).

Back-arc magmatic rocks in the slightly younger Kudz Ze Kayah formation include felsic volcanic, volcanioclastic and volcanosedimentary rocks (Murphy et al., 2002, this volume; Piercey et al., 2001b, 2002a, 2003). Felsic rocks of the Kudz Ze Kayah formation and coeval intrusions of the Grass Lakes plutonic suite are characterized by within-plate (A-type) geochemical signatures, with very high HFSE and REE contents and low compatible element contents (Figs. 12C; D; Piercey et al., 2001b, 2003). The flat UCC-normalized patterns, coupled with negative Ti, Eu, Al, Sc and V anomalies (Fig. 12C) are consistent with their derivation from partial melting of continental crustal sources coupled with the fractionation of feldspar and oxide minerals. An origin by partial melting of continental crust is also supported by the Nb/Ta, Ti/Sc and La/Yb ratios in these rocks (Piercey et al., 2001b, 2003), neodymium isotopic data (εNd = −7.8 to −9.5; Piercey et al., 2003), ubiquitous Proterozoic-

Table 2. Summary of geochemical attributes of non-arc assemblages for Paleozoic magmatic cycles in YTT.

<table>
<thead>
<tr>
<th>Cycle</th>
<th>Age Range</th>
<th>Locations</th>
<th>Back-Arc or Non-Arc Setting (Rift?)</th>
<th>Geochemical Signatures</th>
<th>Isotopic Signatures</th>
<th>Sources of Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>II</td>
<td>365-357 Ma</td>
<td>Finlayson Lake (5)</td>
<td>Felsic: within-plate (A-type); Mafic: OIB-like, NEB, T-NEB, BABB, and E-MORB in Fire Lake formation.</td>
<td>Felsic: εNd = −7.8 to −9.5 (Kudz Ze Kayah); Mafic: εNd = −2.8 to +1.1 (Wind Lake); εNd = −1.6 to +8.5 (Fire Lake).</td>
<td>Piercey et al. (2001a, b, 2002a, 2003, 2004); Grant (1997).</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sylvester allochthon (9)</td>
<td>Felsic: within-plate (A-type) volcanic rocks; Mafic: N-MORB to BABB volanic rocks.</td>
<td>Felsic: εNd = −7.1 to −8.2; Mafic: εNd = +6.9.</td>
<td>Piercey et al. (2001a, b, 2002b, 2003); Grant (1997).</td>
<td></td>
</tr>
<tr>
<td>IV</td>
<td>342-314 Ma</td>
<td>No back-arc yet discovered.</td>
<td>N/A</td>
<td>N/A</td>
<td>Nelson (1993).</td>
<td></td>
</tr>
<tr>
<td>V</td>
<td>314-269 Ma</td>
<td>Finlayson Lake (5)</td>
<td>Mafic: N-MORB, E-MORB, OIB, and BABB; MORB-signatures in eclogites derived from Campbell Range.</td>
<td>Campbell Range equivalent eclogites; εNd = +5.4 to +9.3. Juvenile Pb-isotopes in VMS associated with Campbell Range.</td>
<td>Creaser et al. (1999); Mann and Mortensen (2000); S. Piercey and D. Murphy (unpublished data); Mortensen et al. (this volume).</td>
<td></td>
</tr>
<tr>
<td>VI</td>
<td>269-253 Ma</td>
<td>No back-arc yet identified.</td>
<td>N/A</td>
<td>N/A</td>
<td>Ferri (1997).</td>
<td></td>
</tr>
</tbody>
</table>

Note: *Bold characters indicate areas for which data are summarized in this paper; numbers refer to locations shown in Figure 1.
Archean zircon inheritance in geochronological samples (Mortensen, 1992a; Piercey, 2001), and very radiogenic (high-μ ~12) Pb isotopic systematics in sulphides associated with felsic volcanic-hosted massive sulphide deposits (Mortensen et al., this volume).

The culmination of back-arc magmatism during Cycle II is represented by the metamorphosed mafic volcanic rocks of the stratigraphically highest Wind Lake formation and accompanying high level intrusions (Murphy et al., this volume; Piercey et al., 2002a). This formation is characterized by OIB signatures, with a subsuite of rocks that have been contaminated by continental crust (CC-OIB; Fig. 11C; Piercey et al., 2002a). Both suites have high TiO₂, P₂O₅, HFSE and REE contents and LREE-enriched signatures. The uncontaminated suite shows positive Nb-anomalies (and εNd = +1.1), and the contaminated suite has slightly negative Nb anomalies and higher Th/Nb ratios (εNd = -2.8; Fig. 11C; Piercey et al., 2002a). The low volume of mafic magmatism in the upper parts of the Grass Lakes group, a ~357-360 Ma intra-arc deformation event, and an unconformity overlying the succession at ~357 Ma (Murphy et al., this volume), suggest that the Grass Lakes back-arc became an aborted rift and did not evolve to full seafloor spreading.

In the Stewart River area, Cycle II mafic magmatism is represented by amphibolitic rocks that are interlayered with a lower metasedimentary package (Ryan and Gordey, 2001, 2002; Ryan et al., 2003). Preliminary geochemical data (S.I. Piercey and J.I. Ryan, unpublished data) for amphibolitic rocks are characterized by IAT signatures, with one sample having a N-MORB signature (Fig. 11D), suggesting that these rocks were derived from depleted mantle source

**Figure 11.** (facing page, top) Primitive mantle normalized plots for Cycle II mafic rocks. (A) Average values for arc rocks of the Waters Creek and Cleaver Lake formations, Finlayson Lake district (BON = boninite; IAT = island arc tholeiite; L-IAT = LREE-enriched island arc tholeiite; CAB = calc-alkaline basalt; data from Piercey, 2001; Piercey et al., 2004); (B) Average values for non-arc rocks of the Fire Lake and Cleaver Lake formations, Finlayson Lake district (OIB-1 = Nb-enriched basalt with low La/Yb; T-OIB = Th-rich, Nb-enriched basalt; BABB = back-arc basin basalt); (C) Average values for non-arc basaltic rocks from the Wind Lake formation, Finlayson Lake district (data from Piercey et al., 2002a); (D) Arc and non-arc rocks from amphibolitic rocks from the Stewart River area (S. Piercey and J. Ryan, unpublished data).

**Figure 12.** (facing page, bottom) Upper continental crust (UCC) normalized and Nb-Y discrimination plots for Cycle II felsic rocks. (A, B) Cleaver Lake formation (CAR = calc-alkaline rhyolite, ThR = tholeiitic rhyolite; data from Piercey et al., 2001b); (C, D) Grass Lakes group (KZK-Rh = Kudz Ze Kayah formation rhyolite; KZK-FPI = Kudz Ze Kayah formation felsic porphyry intrusive; KZK-FT = Kudz Ze Kayah formation felsic tuff; GLS-Gr = Grass Lakes suite granitoid; data from Piercey et al., 2001b).
Figure 11. Caption on facing page.

Figure 12. Caption on facing page.
regions. Furthermore, the spatial, temporal and geochemical similarities to rocks of the Fire Lake, Cleaver Lake and Waters Creek formations in the Finlayson Lake district suggest that the Stewart River area represents a continuation of the arc front recorded in the Cleaver Lake and Waters Creek formations, and initial back-arc basin development recorded in the Finlayson Lake district during Cycle II (Fig. 10).

Wolverine Cycle (Cycle III - Early Mississippian - 357-342 Ma)

The third magmatic cycle corresponds to the climax of felsic magmatism in YTT (Fig. 6). It begins at a marked angular unconformity at the base of the Wolverine Lake group in the Finlayson Lake district (Murphy et al., this volume; Colpron et al., this volume-a), and is therefore named after this succession. The base of the Wolverine Lake group comprises quartz- and feldspar-pebble conglomerate, grit and sandstone derived from the underlying Late Devonian – Early Mississippian volcanic and plutonic rocks (Murphy et al., this volume). This unconformity represents a hiatus of approximately 3-4 m.y. The Wolverine massive sulphide deposit formed during this cycle. This cycle ends with another deformational event, best defined in the Glenlyon area of central Yukon (Colpron et al., this volume-b), but also inferred in many parts of YTT (Colpron et al., this volume-a). This deformational event is recorded by fabric development in ~347-343 Ma volcanic and plutonic rocks that are intruded by an undeformed ca. 340 Ma pluton, and by foliated quartzite clasts in basal conglomerate of the ca. 340 Ma and younger Little Salmon formation (Colpron et al., this volume-b).

Cycle III magmatism was of arc, arc-rift and back-arc types. Arc magmatism occurred in the western Finlayson Lake district, Stewart River, Fortymile River, Glenlyon, Teslin and Wolf Lake-Jennings River areas (Fig. 13). Magmatism of probable arc affinity is also documented in the Stikine assemblage of northwestern British Columbia (Logan et al., 2000; Gunning et al., this volume) and the Endicott Arm assemblage of southeastern Alaska (Fig. 13; McClelland et al., 1991; Gehrels, 2001). Back-arc magmatic activity was restricted to the eastern Finlayson Lake area, in the foothall of the Money Creek thrust (Fig. 13; Piercey et al., 2001b, 2002b, 2003).

Arc magmatism in the Finlayson Lake district is manifested primarily by hornblende- and biotite-bearing granitoids of the Simpson Range plutonic suite. They range in age from ~357-345 Ma, but are mostly in the range of ~350-345 Ma (Mortensen, 1992a; J.K. Mortensen and D.C. Murphy, unpublished data). The UCC-normalized plots for Simpson Range granitoids show slight LREE-enrichment, weakly negative Nb anomalies (Fig. 14A) and low Y, consistent with an arc affinity (Fig. 15A; Piercey et al., 2003; Grant, 1997). These patterns are also consistent with derivation from upper crustal sources. This is further supported by their Nd isotopic signatures (εNd = -7.4 to -12.9; Grant, 1997; Piercey et al., 2003), upper crust-like Pb-isotope systematics (Grant, 1997) and inherited Proterozoic (=Archean) zircon (Mortensen, 1992a, b; Grant, 1997).

Cycle III back-arc magmatic rocks occur in the Wolverine Lake group in the eastern Finlayson Lake district (Fig. 13). Wolverine
Paleozoic magmatism and crustal recycling

Figure 14. Upper continental crust (UCC) normalized plots for Cycle III felsic rocks. (A) Plutonic rocks from the Simpson Range plutonic suite (SRPS), Finlayson Lake district (SRPS-P from Piercey et al., 2003; SRPS-G from Grant, 1997); (B) Non-arc rocks from the Finlayson Lake district (WV-5f - Wolverine Lake group felsic volcanic rocks; WV-6f-FW - Wolverine Lake group - felsic volcanic rocks in footwall of Wolverine deposit; WV-6f-HW - Wolverine Lake group - aphyric rhyolites in hanging wall of Wolverine deposit; data from Piercey et al., 2001b); (C) Tonalitic-dioritic orthogneiss from Stewart River area (S. Piercey and J. Ryan, unpublished data); (D) Orthogneiss from the Fortymile River area (Dusel-Bacon et al., this volume); (E) Felsic plutonic and volcanic rocks from the Glenlyon region (LK-And = Little Kalzas andesite; LK-Rhy - Little Kalzas rhyolite; LKS-Gr - Little Kalzas suite granite; data from Colpron, 2001); (F) Felsic plutonic and volcanic rocks from the Wolf Lake-Jennings River area (data from Nelson and Friedman, 2004); and (G) Simpson Range plutonic suite equivalents from the Teslin area (data from Stevens et al., 1995).
Figure 15. Nb-Y discrimination plots for Cycle III felsic rocks. Data sources and symbology as in Figure 14.
Lake group strata below the Wolverine VHMS deposit contain felsic tuffs and flows. Above the deposit, silicified, aphyric rhyolites are overlain by massive mafic lava flows (Murphy and Piercey, 1999; Piercey et al., 2001b, c, 2002b). The felsic rocks in the footwall of the Wolverine deposit are characterized by HFSE- and REE-enriched signatures, which are virtually identical to the rocks of the Kudz Ze Kayah formation, and are consistent with formation within an ensialic back-arc basin environment (Figs. 14B, 15B; Piercey et al., 2001b, 2002b). Above the Wolverine deposit, the rhyolites become less HFSE- and REE-enriched exhibiting more arc-like patterns and Nb-Y contents (Figs. 14B, 15B; Piercey et al., 2001b, 2002b). The position of these thin, silicified, aphyric rhyolite flows, above non-arc felsic rocks and below MORB-type basalts (see below), would require an unacceptably transient arc location. The relatively depleted, arc-like signatures of the aphyric rhyolites could result from silicification, which could have diluted their trace element abundances (Fig. 15); however, this would not appreciably change the trace element profiles of these rocks. Alternatively, Piercey et al. (2001b) suggested that trace element depletions in the aphyric rhyolites may be due either to mixing of HFSE- and REE-depleted mafic magmas with continental crust, or to lower temperature melting of continental crust. In either case, a continental crustal influence is evident in both the hanging wall and footwall felsic rocks of the Wolverine deposit, as both have distinctly negative εNd values (εNd = −7.1 to −8.2; Piercey et al., 2003).

Capping the entire felsic back-arc sequence in the Wolverine Lake group are massive basalts (Murphy et al., this volume). They have MORB-like signatures that have been derived from a depleted to weakly enriched mantle source with a weak subduction signature (Fig. 16A; Piercey et al., 2002b); a single Wolverine basalt sample yielded εNd = +6.9, consistent with derivation from depleted mantle sources (Piercey, 2001). The presence of N-MORB to E-MORB like signatures in the Wolverine basalts suggests that they were derived by melting in response to crustal thinning during back-arc basin formation and seafloor spreading (Piercey et al., 2002b).

In the Stewart River area, Cycle III magmatism is dominated by an arc-like assemblage of tonalitic-dioritic-granodioritic orthogneiss, which intrudes a lowermost pre-Late Devonian metasedimentary package and Late Devonian amphibolites (Ryan and Gordey, 2001, 2002; Ryan et al., 2003). The orthogneissic rocks have preliminary U-Pb ages (Villeneuve et al., 2003) and petrological attributes that are similar to the Simpson Range plutonic suite in the Finlayson Lake district. Preliminary geochemical data for the tonalitic orthogneiss show relatively flat UCC-normalized patterns with flat REE but weak positive anomalies in Nb, Eu, Ti and Al, and arc-like Nb-Y systematics (Figs. 14C, 15C; S. Piercey and J. Ryan, unpublished data). There are no Sm-Nd data for felsic rocks of the Stewart River area, but their higher Ti, Nb and Al relative to UCC suggest they may have been derived from a source more depleted than UCC, or as a result of mantle-UCC mixing (Fig. 14C). Nevertheless, samples of the orthogneiss in the Stewart River area contain zircons with Proterozoic inheritance (Mortensen, 1990; Villeneuve et al., 2003), and have flat REE relative to UCC (Fig. 14C), suggesting that a component of continental material was involved in their genesis.

In the eastern Yukon-Tanana Upland, just northwest of the Stewart River area, Cycle III magmatism is represented by ~357-341 Ma (J. Mortensen and C. Dusel-Bacon, unpublished data; Day et al., 2002) mafic amphibolites (gabbro/diorite?), intermediate-composition quartz diorite/tonalite orthogneiss, and subordinate coeval granodiorite and felsic metatuff of the Fortymile River assemblage. These igneous rocks are interlayered with metasedimentary rocks (Dusel-Bacon et al., this volume, and references therein). Amphibolitic rocks have trace element signatures that have LREE-enriched island arc and, to a lesser extent, MORB affinities (Fig. 16B; Dusel-Bacon and Cooper, 1999; Day et al., 2002). Plutonic orthogoness samples are subdivided into those with intermediate compositions (~64% SiO2) and more felsic affinities (~70% SiO2). On UCC-normalized diagrams, the lower SiO2 group are depleted in LREE and Th relative to UCC and enriched in Al, Sc and V (Fig. 14D) suggesting derivation from sources more primitive than UCC. In addition, these samples likely have an affinity transitional between tholeiitic and calc-alkaline (see Barrett and MacLean, 1999) and were most likely intruded within an arc setting (Fig. 15D). The higher SiO2 group have calc-alkaline patterns with flat UCC normalized patterns and minor depletions in Nb, Sc and V, consistent with derivation from UCC-like sources within an arc setting (Figs. 14D, 15D). Derivation from UCC-like sources is supported by Precambrian inheritance in some zircons from this region (J. Mortensen and C. Dusel-Bacon, unpublished data), as is the case with nearby, and likely related, magmatic rocks of the Stewart River area. The occurrence of several amphibolites with MORB characteristics may be evidence for local intra-arc or back-arc magmatism.

Cycle III magmatism in the Glenlyon area is represented by volcanic and sedimentary rocks of the 347-344 Ma Little Kalzas formation and coeval granitoids of the Little Kalzas plutonic suite (Colpron et al., this volume-b). Volcanic rocks in the Little Kalzas formation consist of andesites with lesser rhyolites and basalts, interbedded with epiclastic and basinal sedimentary strata (Colpron, 2001; Colpron et al., this volume-b). Granitoids of the Little Kalzas suite are metalamious, in part K-feldspar megacrystic, diorite and tonalite, and resemble the Simpson Range plutonic suite (Colpron, 2001).

Little Kalzas formation andesites have relatively flat REE relative to the UCC but have lower Th, and higher Eu, Ti, Al and V relative to UCC (Fig. 14E), yet have Zr/Y ratios (~4-9) that are transitional to calc-alkaline in affinity (Barrett and MacLean, 1999). These features suggest possible derivation via mixing between mafic (i.e., Eu, Ti, V-enriched) material and continental crust, within a continental arc environment. Little Kalzas rhyolites have flat UCC-normalized patterns consistent with derivation from a UCC-like source (Fig. 14E). Depletions in Ti, V and Al to a lesser extent Nb (Figs. 14E, 15E) are consistent with oxide fractionation at high levels in the crust (cf. Lentz, 1998; Piercey et al., 2001b). Plutonic rocks of the Little Kalzas suite have a similar, but less erratic UCC-normalized pattern, likely reflecting derivation from continental crust-mantle mixing within a continental arc setting (Fig. 14E). Both intrusive and extrusive rocks have relative low Nb and Y, and plot within the volcanic arc field in Nb-Y space (Fig. 15E). Tracer isotopic data are not available for the Little Kalzas formation and Little Kalzas suite
rocks; however, most U-Pb geochronological samples from this region have inherited Proterozoic-Archean zircon (Colpron et al., this volume-b), suggesting that there is a significant older crustal component in these rocks. Mafic rocks associated with the Little Kalzas formation, and the underlying Pelmac formation, have smooth PM-normalized patterns with OIB-like signatures consistent with derivation from enriched mantle sources (Fig. 16C); no isotopic data are presently available for these rocks. These mafic rocks likely reflect rifting of the Little Kalzas arc (Colpron, 2001), which may have been related to coeval arc-rifting and ensialic back-arc magmatic activity in the Finlayson Lake district (Piercey et al., 2001b, 2002b).

In the Wolf Lake – Jennings River area, Cycle III is represented by intrusive rocks structurally associated with, and possibly basement to the Ram Creek Complex, metavolcanic rocks in the upper

**Figure 16.** Primitive mantle normalized plots for Cycle III mafic rocks. (A) Mafic rocks from the uppermost part of the Wolverine Lake group (data from Piercey et al., 2002b); (B) Amphibolites from the Fortymile River assemblage (Dusel-Bacon and Cooper, 1999); (C) Mafic rocks from the Glenlyon region (data from Colpron, 2001); (D) Arc and (E) non-arc mafic rocks from Wolf Lake – Jennings river area (data from Nelson and Friedman, 2004); (F) mafic rocks from the Teslin region (data from Creaser et al., 1997).
Dorsey Complex, and minor mafic-intermediate rocks in the Swift River Group (Roots et al., this volume; Nelson, 1999, 2000; Roots and Heaman, 2001). The Dorsey Complex, which structurally overlies the Ram Creek complex, consists of a lower unit of quartzofeldspathic metasedimentary rocks and metabasalts (basalt/gabbro flows/sills), and an upper unit of siliciclastic rocks, marble and minor felsic tuffs (Roots et al., this volume; Nelson and Friedman, 2004). Early Mississippian ages have been obtained for felsic metavolcanic rocks from the upper Dorsey Complex (Roots and Heaman, 2001). The Swift River Group lies in structurally modified depositional setting (Fig. 14G) and suggests derivation from a more primitive source than the Nelson, 2000; Roots et al., 2000). The Swift River Group consists predominantly of basinal sedimentary rocks, with minor volcanoclastic debris and tuffaceous rocks (Nelson, 2000; Roots et al., 2000).

Early Mississippian tonalitic intrusions associated with the Ram Creek Complex have relatively flat UCC-normalized patterns but lower Th, and higher Ti, Al and V relative to UCC (Fig. 14F). These features are consistent with potential derivation from a more mafic source, or from crust-mantle mixing within an arc environment (Figs. 14F, 15F; Nelson and Friedman, 2004). Early Mississippian UCC-normalized patterns for coeval rhyolitic metavolcanic rocks from the upper Dorsey Complex have UCC-normalized patterns, with relatively flat REE but with elevated Th, Zr and Hf relative to UCC, and Nb-Y systematics that are transitional between arc and non-arc (Figs. 14F, 15F). One sample of mafic rock from the Upper Dorsey complex has an L-IAT to CAB affinity (Fig. 16D). Mafic tuffs from the Swift River Group are dominated by L-IAT to CAB signatures consistent with derivation from variably enriched mantle sources coupled with a subduction zone component with or without minor crustal contamination. In addition, two samples have non-arc affinities with N-MORB and OIB signatures (Fig. 16E; Nelson and Friedman, 2004). Collectively, these geochemical data are consistent with arc magmatism that was interrupted by intra-arc rifting.

In the Teslin area, Cycle III magmatism is represented by Simpson Range plutonic suite equivalents (Stevens et al., 1995) and possibly by mafic metavolcanic rocks (unit PMgr of Stevens, 1994; “Anvil assemblage” of Creaser et al., 1997). The mafic rocks have uncertain age and stratigraphic relationships and are considered to be part of Cycle III due to their spatial association with intrusions of that age; however, they could be part of any cycle in YTT. The Simpson Range plutonic suite equivalents in the Teslin region consist of variably deformed hornblende-bearing tonalite to quartz-diorite (Figs. 14G, 15G; Stevens et al., 1995). These granitoids have very erratic signatures that are broadly flat with positive Eu, Ti and Al anomalies, low La and Th relative to Nb (Fig. 14G) and low Nb-Y (Fig. 15G) that are consistent with formation within an arc environment. Their relative LREE-depletions and high Nb, Eu, Ti and Al (Fig. 14G) suggest derivation from a more primitive source than the UCC; however, εNd values for these Simpson Range plutonic suite equivalents (εNd = -2.5 to -6.2; Stevens et al., 1995) clearly point to an ancient crustal component. Mafic rocks from the Teslin region have been interpreted to reflect calc-alkaline basaltic protoliths (Creaser et al., 1997); however, a re-evaluation of the geochemical data suggest that these rocks more closely resemble E-MORB with a weak subducted slab component (Fig. 16F). Their signatures are consistent with derivation from moderately enriched sources during arc development or arc-rifting, as they are transitional between arc and back-arc magmatism. Furthermore, the Nd-isotopic data for the greenstones of the Teslin area are consistent with derivation from a moderately enriched mantle source, as they exhibit εNd = +4.1 to +6.4 (Creaser et al., 1997), which is less than the depleted mantle at 350 Ma (εNd = +9.5; Goldstein et al., 1984).

**Little Salmon Cycle (Cycle IV - Late Mississippian - 342-314 Ma)**

The fourth magmatic cycle is mostly represented in the southern part of YTT (Fig. 17). It begins at an unconformity beneath the Little Salmon formation of central Yukon and is thus named after this succession. The sub-Little Salmon unconformity is locally marked by a basal conglomerate containing foliated quartzite clasts and Early Mississippian detrital zircons (Colpron et al., this volume-b). Cycle IV magmatism was dominated by arc, and lesser intra-arc rift magmatism; no corresponding back-arc magmatism has yet been identified for this cycle. It is represented by the Little Salmon formation and associated plutons in the Glenlyon area (Colpron et al., this volume-b), the Ram Creek and Big Salmon complexes in the Wolf Lake-Jennings River area (Roots et al., this volume; Nelson and Friedman, 2004), and bimodal magmatism in Stikine terrane (Sebert and Barrett, 1996; Logan et al., 2000; Gunning et al., 2001). The Tulsequah Chief VMS deposit in northwestern British Columbia formed during this cycle (Fig. 17).

Coeval successions of YTT in the Finlayson Lake district and parts of Teslin and Wolf Lake areas are characterized by deposition of carbonates in the arc marginal regions. Basinal sedimentary rocks are predominant in back-arc areas rather than igneous rocks (Fig. 17; Slide Mountain/Seventymile assemblages; Nelson, 1993; Murphy et al., this volume; Dusel-Bacon et al., this volume).

In the Glenlyon area, volcanic rocks of the Little Salmon formation consist of basalt, andesite, and lesser dacite, tuff and quartz-feldspar porphyry, interbedded with volcano-sedimentary rocks (Colpron, 2001). Plutons of the Tatlmain suite are coeval and cogenetic with them. Little Salmon andesites and volcaniclastic rocks have LREE-depleted UCC-normalized patterns with higher Eu, Ti, V and HREE relative to UCC (Fig. 18A). These features are consistent with derivation from a source more depleted than UCC, or with mixing between mantle-derived magma and crustal material. In contrast, the dacites and quartz-feldspar porphyries from the Little Salmon formation have broadly flat UCC-normalized patterns with relatively weak depletions in Nb and V, and higher LREE and total REE (Fig. 18A), all of which are consistent with derivation from melting of continental crust accompanied by high-level oxide fractionation. The Tatlmain suite plutons have fairly flat, slightly LREE-depleted signatures (Fig. 18A), reflecting derivation from mixed mantle-crust sources. All of the felsic to intermediate rocks of the Little Salmon formation and the Tatlmain suite intrusions have low Nb-Y, typical of arc settings (Fig. 19A). Mafic rocks from the Little...
Late Mississippian Little Salmon cycle (Cycle IV: 342-314 Ma). Events assemblage legend. Teeth indicate dip of subducting slab. See Figure 1 for tectonic as red line marks approximate limit between geodynamic environments. Prior to ~430 km of dextral displacement along Tintina fault. Dashed are located on a base map of Yukon-Tanana and related terranes.

Figure 17: Distribution of tectonic and magmatic events during the Late Mississippian Little Salmon cycle (Cycle IV: 342-314 Ma). Events are located on a base map of Yukon-Tanana and related terranes prior to ~430 km of dextral displacement along Tintina fault. Dashed red line marks approximate limit between geodynamic environments. Dashed blue line represents relative position of subduction zone; teeth indicate dip of subducting slab. See Figure 1 for tectonic assemblage legend.

Salmon formation are dominated by OIB-like alkalic signatures, which indicate derivation from enriched mantle source regions (Fig. 20A; Colpron, 2001). This geochemical signature likely reflects intra-arc rifting episodes, which are also indicated by the presence of spatially associated Mn-rich exhalative horizons (Colpron, 2001).

Only limited geochemical data are available for mafic volcanic rocks of the Big Salmon Complex (M. Mihalyuk, unpublished data; Nelson and Friedman, 2004). In the Wolf Lake-Jennings River area, the Big Salmon Complex contains a stratigraphic succession of greenstone overlain by marble, Mn-rich silicious exhalate and siliciclastic sedimentary rocks with minor felsic tuff (Mihalyuk et al., 1998, 2000, this volume). Mafic volcanic rocks from the Big Salmon Complex are dominated by LREE-enriched signatures with negative Nb anomalies, typical of L-IAT suite magmas (Fig. 20B); andesite with a calc-alkaline signature is locally present (Fig. 20B). The Big Salmon Complex also contains a few samples that exhibit E-MORB to OIB affinities with positive Nb anomalies (Fig. 20B). These non-arc magmas probably reflect intra-arc rift events. As in the Little Salmon formation, the presence of Mn-rich exhalative horizons, interpreted to have formed during volcanic hiatuses, within the Big Salmon Complex also supports this interpretation (Mihalyuk and Peter, 2001).

The Late Mississippian Ram Creek Complex in the Wolf Lake-Jennings River area comprises intermediate to felsic tuff, interbedded with basinal and siliciclastic sedimentary rocks and rare marble (Roots et al., this volume; Nelson and Friedman, 2004). Quartz-sericite schist from the Ram Creek Complex exhibits a calc-alkaline arc affinity (Figs. 18B, 19B), consistent with formation within an arc environment.

The Tulsequah Chief VMS deposit in northwestern British Columbia provides the most comprehensive geochemical dataset for the Stikine portion of Cycle IV magmatism. The Tulsequah Chief deposit is hosted by a Mississippian (327 ± 1 Ma and 330 ±0/1 Ma, U-Pb zircon on rhyolites; Childe, 1997) bimodal assemblage of basaltic and rhyolitic volcanic and volcaniclastic rocks (Sebert and Barrett, 1996). The footwall succession to the Tulsequah Chief deposit is bimodal: calc-alkalic arc rhyolitic rocks are interbedded with basalts that display E-MORB signatures, and in some cases include a subordinate subduction zone component (Figs. 18C, 19C); hanging wall basaltic rocks exhibit similar signatures (Fig. 20C). Collectively, these rocks probably represent derivation from a weakly enriched mantle source, coupled with crustal melting to form a bimodal volcanic assemblage in response to arc rifting (e.g., Sebert and Barrett, 1996).

Klinkit Cycle (Cycle V - Pennsylvanian to Early Permian - 314-269 Ma)

The fifth cycle is primarily represented by arc magmatism of the Klinkit Group and Fourmile succession in southern Yukon and northern British Columbia (Roots et al., 2002, this volume; Simard et al., 2003; Nelson and Friedman, 2004, Fig. 21) and the Lay Range assemblage of central British Columbia (Fig. 1; Ferri, 1997). Arc magmatism is also inferred in the Semenof Hills of south-central Yukon (Fig. 1; Tempelman-Kluit, 1984; Simard and Devine, 2003;
Paleozoic magmatism and crustal recycling

Figure 18. Upper continental crust (UCC) normalized plots for Cycle IV felsic rocks. (A) Tatmain suite arc felsic rocks from the Glenlyon region (data from Colpron, 2001); (B) Arc felsic rocks from the Wolf Lake – Jennings River area (data from Nelson and Friedman, 2004); and (C) Arc felsic rocks from the footwall of the Tulsequah Chief VMS deposit, Stikine terrane (data from Sebert and Barrett, 1996).

Figure 19. Nb-Y discrimination plots for Cycle IV felsic rocks. Data sources and symbology as in Figure 18.
This cycle corresponds to a lull in felsic magmatism in YTT (Fig. 6) and important deposition of carbonate and basinal sedimentary rocks in the Finlayson Lake district and Stikine terrane (Colpron et al., this volume-a).

Coeval, back-arc magmatism is present in the Campbell Range formation of the Finlayson Lake district (Murphy et al., this volume; Plint and Gordon 1997; Piercey et al., 1999) and Slide Mountain terrane in the Sylvester allochthon and Nina Creek Group (Nelson, 1993; Ferri, 1997; Figs. 1, 21). The beginning of this cycle is arbitrarily set at the Mississippian-Pennsylvanian boundary, a time marking the end of felsic magmatism, and the maximum preservation of fossil faunas in sedimentary strata of YTT (Fig. 6; Colpron et al., this volume-a).

The Klinkit Group outcrops discontinuously from northern British Columbia into southern Yukon, a strike length of over 250 km (Figs. 1, 21). It is characterized by the predominance of Pennsylvanian to Permian volcaniclastic rocks overlying carbonates of Viséan to Bashkirian age (Roots et al., this volume). The Klinkit Group unconformably overlies the rocks of the Swift River Group (Nelson, 2002; Roots et al., this volume). Within the Klinkit Group, the volcaniclastic members of the Butsish and Mount McCleary formations are composed of crystal and lithic tuff. They are mainly basaltic in composition, with calc-alkaline arc PM-normalized signatures (Fig. 22A), but TiO₂ and Ti/V ratios akin to modern island-arc tholeiitic basalts (Gamble et al., 1995), suggesting that the rocks have transitional signatures between tholeiitic and calc-alkaline (Simard et al., 2003). Their εNd values (+6.7 to +7.4) and Th/La ratios (0.13 to 0.17) indicate that these rocks were derived from depleted mantle sources with minimal crustal contamination (Simard et al., 2003).

Based on geochemical and geological evidence, the volcaniclastic members of the Klinkit Group are interpreted as primitive arc lavas erupted either through relatively young crust that consists of slightly older arc basement, or rapidly emplaced through coated conduits and/or relatively thin continental crust without significant crustal contamination (Simard et al., 2003). The alkali-basalt member of the Mount McCleary Formation comprises scarce discontinuous lenses within the volcaniclastic member (Roots et al., this volume). These mafic rocks have OIB-like alkalic signatures with high TiO₂, HFSE and LREE (Fig. 22A). They are interpreted to have been derived from an enriched mantle source associated with episodic intra-arc rifting of the Klinkit arc (Simard et al., 2003).

In the Sylvester allochthon, Cycle V magmatic rocks occur within the Fourmile succession. It occurs as thrust panels, in which undeformed volcanic and epiclastic units, minor limestone, overlie polydeformed black phyllite, siltstone and argillite (Nelson, 2002; Nelson and Friedman, 2004). It forms part of Division III, the structural division of the allochthon assigned to the Harper Ranch (late...
Paleozoic arc) subterrane of Quesnellia (Nelson, 1993). Basaltic to andesitic rocks of the Fourmile succession have predominantly L-IAT affinities, consistent with derivation from weakly-enriched mantle sources within a subduction zone environment (Fig. 22B). Andesitic to dacitic volcanic rocks of the Fourmile succession have calc-alkaline arc signatures, slightly depleted relative to the UCC (Fig. 23).

Pennsylvanian-Early Permian arc magmatism within the Lay Range assemblage of the Quesnell terrane (Ferri, 1997) is coeval with, and inferred to be correlative to, the Klinkit Group (Simard et al., 2003). In the Lay Range, Ferri (1997) has described both L-IAT (Fig. 22C) and MORB signatures in basaltic rocks from the Upper Mafic Tuff division (Fig. 22D). These signatures are very similar to magmatism in the Klinkit Group (Simard et al., 2003) and Fourmile succession (Nelson, 2002; Fig. 20). Simard et al. (2003) suggested that by the late Paleozoic, YTT was the basement to the Quesnell arc. Furthermore, they argued that the Klinkit Group represented distal turbiditic sedimentation in response to arc magmatic activity within a large YTT arc system. The presence of MORB-type magmatism, although minor, in the Lay Range assemblage suggests that the Lay Range, Klinkit Group and Fourmile succession reflect an arc system that was subsequently rifted during renewed back-arc extension, as recorded in the Campbell Range formation of the Finlayson Lake district (Fig. 21).

In the Campbell Range, Pennsylvanian-Permian massive and pillowled lava flows, chert and carbonates unconformably overlie older rocks of the Money Creek formation, the Wolverine Lake and Grass Lakes groups, and the Fortin Creek group east of the Jules Creek fault (Murphy et al., 2002, this volume). Diabase, gabbro and ultramafic rocks intrude both the basalt succession and its basement. Basalts of the Campbell Range formation have an array of non-arc signatures, including N-MORB, E-MORB and OIB, representing derivation from a variably enriched mantle wedge (Fig. 22E; S. Piercey and D. Murphy, unpublished data). Additional data from basalts of the Campbell Range formation (Plint and Gordon, 1997; Piercey et al., 1999), and from eclogitic rocks which are interpreted to be recycled Campbell Range basalts (Creaser et al., 1999), provide additional evidence in support of derivation from variably enriched mantle in a back-arc basin environment. At present, limited isotopic data exist to elucidate the nature of the crust-mantle interaction in the Campbell Range formation. Nd isotopic data for the MORB-like Permian eclogites show εNd = +5.4 to +9.3, suggesting derivation from depleted mantle to weakly enriched sources (Creaser et al., 1999). Furthermore, Pb-isotopic signatures on sulphides from VMS occurrences hosted by the Campbell Range basalts in the Finlayson Lake district are non-radiogenic and suggestive of mantle sources (Mann and Mortensen, 2000). These primitive isotopic signatures, coupled with the prevalence of N-MORB and E-MORB lavas, suggest that full seafloor spreading occurred in the Campbell Range back-arc basin.

In north-central British Columbia, magmatism coeval with the Campbell Range formation is recorded within the Slide Mountain terrane (Nina Creek Group and Sylvester allochthon; Ferri, 1997; Nelson, 1993, 2002). Overall, Slide Mountain basalts range from
Figure 22.
Caption on facing page.
latest Devonian through mid-Permain in age, corresponding to Cycles II-V in YTT. Pennsylvanian-Permian units include the Pillow Ridge succession in the Nina Creek Group, and Pennsylvanian-Permian units in Division II of the Sylvester allochthon. Basalts from the Pillow Ridge succession all show N-MORB signatures (Fig. 22F), as do rocks of the older Mount Howell and Mafic-Ultramafic units. N-MORB signatures are also ubiquitous in the Slide Mountain terrane in the Sylvester allochthon (Fig. 22G; Nelson 1993). These data are similar to the Campbell Range N-MORB; all are consistent with derivation from depleted mantle sources within an oceanic or back-arc basin environment (Nelson, 1993, 2002; Ferri, 1997). Back-arc magmatism in the Campbell Range, however, also included more enriched-mantle products (OIB and E-MORB).

The predominant N-MORB signatures and the juvenile nature of the isotopic systematics of Pennsylvanian – Early Permian mafic rocks in the Slide Mountain terrane suggest that magmatism occurred in a back-arc basin setting with no interaction with continental crust. However, local interbedding of terrigenous clastic rocks (e.g., Nelson, 1993), and evolved isotopic systematics from sedimentary rocks of Slide Mountain terrane (Patchett and Gehrels, 1998), indicate a continental source for this detritus. This implies that the Slide Mountain ocean was a marginal sea that developed in proximity to a cratonic region; this situation is similar to the Japan Sea relative to the island of Japan and the Sino-Korean craton (Pouclet et al., 1995; Creaser et al., 1999).

**Klondike Cycle (Cycle VI - Middle to Late Permian - 269-253 Ma)**

The last Paleozoic magmatic cycle is primarily represented in the Klondike region of western Yukon, and is therefore named after this area (Fig. 22). It corresponds to the second pulse of felsic magmatism in the pericratonic terranes (Fig. 6), and its beginning is marked by the formation of eclogites in Yukon (Erdmer et al., 1998). The Klondike cycle ends with the cessation of magmatism in YTT, and a period of depositional and igneous hiatus in the Early Triassic (Fig. 6). Subsequent deposition of Middle to Upper Triassic clastic sedimentary rocks overlap Yukon-Tanana and Slide Mountain terranes, and also the North American miogeocline (Colpron et al., this volume-a; Nelson et al., this volume).

Cycle VI magmatism is represented by the mid- to Late Permian Klondike Schist (~263-253 Ma), a sequence of felsic volcanic and volcanioclastic rocks with lesser interlayered mafic rocks, and coeval and probably co-genetic monzonite to quartz-monzonitic granitoids of the Sulphur Creek orthogneiss (Mortensen, 1990). Late Permian felsic schist layers occur also within carbonaceous rocks assigned to the Nasina assemblage in the Forty Mile River area of eastern Alaska (Figs. 1, 24; J. Mortensen and C. Dusel-Bacon, unpublished data; Dusel-Bacon et al., this volume). Biotite-bearing granitic orthogneiss of Late Permian age also intrudes Devonian-Mississippian metasedimentary rocks of the Nasina assemblage (Ryan et al., 2003; J. Mortensen, unpublished data). Although limited, geochemical data for felsic rocks of the Klondike Schist consistently exhibit flat calc-alkaline signatures, with low Nb, Eu, Ti, Sc and V relative to UCC (Fig. 25), indicative of an arc setting (e.g., Mortensen, 1990; S. Piercey and J. Mortensen, unpublished data). The UCC-profiles for the Sulphur Creek orthogneiss exhibit less systematic behaviour than the Klondike Schist (Fig. 25A), perhaps due to mobility of some

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**Figure 23.** (A) Upper continental crust normalized and (B) Nb-Y discrimination plots for felsic rocks in the Sylvester allochthon (data from Nelson and Friedman, 2004).

**Figure 22.** (facing page) Primitive mantle normalized plots for Cycle V mafic rocks. (A) Arc and non-arc mafic lavas from the Klinkit Group (data from Simard et al., 2003); (B) Arc rocks from the Sylvester allochthon (data from Nelson and Friedman, 2004); (C) Arc and (D) non-arc rocks from the Lay Range area (data from Ferri, 1997); (E) Non-arc rocks from the Campbell Range formation (S. Piercey and D. Murphy, unpublished data); (F) Non-arc rocks from the Nina Creek Group (data from Ferri, 1997); and (G) Non-arc rocks from the Sylvester allochthon (data from Nelson, 1993; Nelson and Friedman, 2004); (H) Cycle VI greenstones from the Slide Mountain terrane in Glenlyon area (data from Colpron et al., 2005).
Middle Permian – Early Triassic Klondike cycle (Cycle VI: 269–253 Ma). Events are located on a base map of Yukon-Tanana and related terranes prior to ~430 km of dextral displacement along Tintina fault. Dashed red line marks approximate limit between geodynamic environments. Dashed blue line represents relative position of subduction zone; teeth indicate dip of subducting slab. See Figure 1 for tectonic assemblage legend.

SYMBOLS

- Volcanism
- Plutonism
- Fossil
- Compressional deformation
- Blueschist / eclogite occurrence
- Permo-Triassic syn-orogenic clastics

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Figure 24. Distribution of tectonic and magmatic events during the Middle Permian – Early Triassic Klondike cycle (Cycle VI: 269-253 Ma). Events are located on a base map of Yukon-Tanana and related terranes prior to ~430 km of dextral displacement along Tintina fault. Dashed red line marks approximate limit between geodynamic environments. Dashed blue line represents relative position of subduction zone; teeth indicate dip of subducting slab. See Figure 1 for tectonic assemblage legend.

In the Fortymile River area, felsic schist layers within carbonaceous rocks of the “Nasina assemblage” have UCC-normalized profiles that can be broken into two groups: a group with relatively flat patterns with depletions in Eu, Zr, Hf and compatible elements, and a second group with similar characteristics but with additional LREE-depletion (Fig. 25B). The first group is consistent with an arc environment; however metarhyolite with LREE-depletion and slightly higher Nb (and Ta) contents suggest a possible within-plate origin for those samples (Fig. 25D). Alternatively, these may represent residual enrichment of Nb and Ta at the expense of LREE, due to loss of LREE during hydrothermal alteration leading to closure-related gains in Nb and Ta (see Stanley and Madeisky, 1994).

Permian magmatic activity is recorded by a suite of plutons that intrude YTT and related allochthons in the Wolf Lake-Jennings River area and the Sylvester allochthon (Fig. 24). A suite of mid-Permian (270-262 Ma) intrusions, including the Ram stock, and Nizi and Meek plutons, intrudes the Dorsey Complex and panels of Cycle V arc rocks in far northern British Columbia (Nelson and Friedman, 2004). One body in this suite stiches a post-Early Permian thrust fault in the Sylvester allochthon. Compositinally, the suite ranges from gabbro to granite; tonalite is the dominant phase. A sample of granite from the Meek pluton exhibits a calc-alkalic arc signature, like the rocks of the Klondike region (Fig. 25C); this scenario is consistent with a southerly continuation of arc magmatism during the Klondike cycle throughout eastern YTT. Carbonate deposition dominates the mid- to late Permian record of the Stikine terrane (Gunning et al., this volume).

Mafic volcanism of back-arc basin affinity (N-MORB) persisted into mid-Permian time in Slide Mountain terrane in the Sylvester allochthon and Seventymile terrane (Figs. 1, 24; Dusel-Bacon et al., this volume; Nelson, 1993). In the Glenlyon area, the Slide Mountain terrane is represented by a narrow belt of chert, argillite, greenstone and serpentinite of Middle Permian age (Colpron et al., 2005). Greenstones from Slide Mountain terrane in Glenlyon area have mixed N-MORB and calc-alkaline signatures indicative of crustal contamination of an N-MORB parent magma (Fig. 22H; Colpron et al., 2005).

DISCUSSION

Petrochemical Constraints on the Regional Tectonic Evolution of Yukon-Tanana and Related Terranes

Models for the tectonic evolution of Yukon-Tanana and Slide Mountain terranes have three common themes: (1) rifting of part of YTT from the western margin of Laurentia in mid-Paleozoic time (~390-365 Ma); (2) mid- to late Paleozoic arc activity, back-arc extension and opening of the Slide Mountain ocean (~365-269 Ma); and (3) subduction of Slide Mountain crust and lithosphere (~269-253 Ma), and the subsequent accretion of YTT back onto the western margin of North America (Tempelman-Kluit, 1979; Mortensen, 1992a; Hansen and Dusel-Bacon, 1998; Nelson, 1993; Nelson et al., this volume). The geochemical and isotopic data reviewed in this paper...
provide further constraints that help refine our understanding of the mid- to late Paleozoic evolution of Yukon-Tanana and related terranes.

It is important to note that the present configuration of YTT (Fig. 1) and the distribution of magmatic belts discussed in this paper are the result of the protracted deformational history of Yukon-Tanana, Slide Mountain and the North American miogeocline between late Paleozoic and early Cenozoic times (e.g., Murphy et al., 2002; Roddick, 1967). On Figures 7, 10, 13, 17, 21 and 24 we show the distribution of the main Paleozoic magmatic and tectonic events on a series of maps of Yukon-Tanana and related terranes prior to ~430 km of early Cenozoic dextral displacement along Tintina fault (Murphy and Mortensen, 2003; Gabrielse et al., in press). These maps essentially show the Late Cretaceous paleogeography of Yukon-Tanana and related terranes. The effects and magnitude of offset associated with late Paleozoic to Mesozoic faulting within YTT are for the most part poorly constrained, and have not been removed from the maps. It should also be noted that displacement along the Denali and Border Ranges faults have not been restored. In the following discussion we assume that the Late Cretaceous distribution of magmatic and tectonic events in YTT approximates their relative original distribution in the Paleozoic.

Alaska Range cycle (1 – 390-365 Ma) magmatic activity involved crustal extension and attenuation of the North American continental margin in the Alaska Range and Yukon-Tanana Upland (Dusel-Bacon et al., this volume, 2004), and continental arc magmatism in the Coast Mountains of southeastern Alaska and British Columbia (Fig. 7). Cycle I magmatic rocks in the Bonnifield mineral belt and correlative assemblages (Lake George assemblage and siliceous, carbonaceous and volcanic assemblage) have diagnostic crustally-derived peralkaline rhyolites and associated OIB-like mafic rocks (Fig. 8). These geochemical characteristics, and the predominantly older ages of these Alaskan rocks, resemble coeval volcanic rocks of the North American miogeocline (Selwyn basin and Cassiar terrane, Fig. 1; Mortensen, 1982; Mortensen and Godwin, 1982; Goodfellow et al., 1995). Mid-Paleozoic alkalic volcanism and extension of the continental margin is likely associated with coeval arc magmatism in YTT in the Coast Mountains, and initiation of the Slide Mountain rift (Paradis et al., 1998; Nelson et al., 2002).
Finlayson cycle (II – 365-357 Ma) magmatism was characterized by arc and back-arc environments in the Finlayson Lake district and Stewart River area. Rift-related magmatism in the Alaska Range and Yukon-Tanana Upland persisted during this cycle. Arc magmatism occurred predominantly in a western belt, now preserved primarily in the Stewart River region and in the southwestern Finlayson Lake district (Table 1; Fig. 10). Arc sequences are dominated by felsic magmatic rocks with calc-alkalic signatures and mafic rocks with arc signatures (Table 1; Figs. 11-12). This magmatic arc belt is coeval with back-arc magmatism in the eastern Finlayson Lake district and MORB-dominated marginal basin facies of the Slide Mountain assemblage in the Sylvester allochthon (Fig. 10; Nelson, 1993), implying a west-facing, east-dipping subduction zone (Mortensen, 1992a; Nelson, 1993; Nelson et al., this volume).

Wolverine cycle (III – 357-342 Ma) magmatism is most widespread in the central and southern parts of the YTT (Fig. 13). Its configuration is similar to that of Cycle II, with arc sequences located in the western portions of the terrane, in front of the more easterly back-arc region in eastern Finlayson Lake, and even more distal back-arc of the Slide Mountain assemblage in the Sylvester allochthon (Fig. 13). There is little Cycle III arc magmatic activity recorded in the Yukon-Tanana Upland of eastern Alaska. The intermediate to mafic volcanic rocks of the Fortymile River assemblage span the Alaska-Yukon border and have arc and MORB signatures (Fig. 16B). They may represent an arc and back-arc sequence, although the tectonic implications of these rocks have yet to be resolved. It is important to note the paucity of magmatism in the Alaska Range from Cycle III onward (Fig. 13). On the other hand, arc magmatic activity was voluminous and is well represented in the Glengyle and Teslin regions (Fig. 13; Table 1). These latter arc sequences also contain alkalic mafic rocks (Fig. 16; Table 1), suggesting that the arc or arcs underwent a period of extension during Cycle III. Furthermore, the presence of arc-dominated successions in the central and western portions of YTT, and the occurrence of seafloor spreading recorded by MORB-type and BABB-type basaltic rocks in the easternmost part of the terrane in the Finlayson Lake district (e.g., Wolverine basalt; Piercey et al., 2002b) and in the Sylvester allochthon (e.g., Slide Mountain basalts; Nelson, 1993), imply the likelihood of westerly rollback of the slab and the induration of back-arc seafloor spreading.

The Little Salmon cycle (IV – 342-314 Ma) marks a fundamental shift in the configuration of the arc system. Both arc and back-arc magmatism essentially ceased in the Finlayson Lake and Stewart River areas, and the locus of arc activity shifted southwards to the Glengyle and Wolf Lake-Jennings River areas (Fig. 17). Coeval arc magmatism also occurs in the Stikine assemblage of northwestern British Columbia (Fig. 17). Cycle IV sequences are dominated by crustally-derived, calc-alkalic magmatic rocks with OIB-like effusions of mafic material (Figs. 18-20), implying continental arc magmatism with intra-arc rift episodes. The lack of documented Late Mississippian magmatic activity in the Slide Mountain assemblage of the Sylvester allochthon has led Nelson and Bradford (1993) to suggest that there was a lull in volcanic activity in the back-arc basin. Alternatively, this may be a function of poor preservation of Late Mississippian Slide Mountain crust. Nevertheless, Cycle III magmatism most likely occurred above an east-dipping subduction zone, as with previous magmatic cycles (Fig. 17).

Klinkit cycle (V – 314-269 Ma) magmatism represents another shift in the style and nature of magmatism in YTT. Arc magmatism was widespread in the Wolf Lake-Jennings River areas, as represented by the Klinkit Group (Simard et al., 2003), and within the Fourmile succession in the Sylvester allochthon (Nelson and Friedman, 2004; Fig. 21). Arc volcanic rocks within the Lay Range assemblage of Quesnellia in north-central British Columbia are also correlated with this magmatic cycle (Fig. 1; Ferri, 1997; Simard et al., 2003). Coeval back-arc magmatism is present in the Campbell Range formation in the Finlayson Lake district and in the Slide Mountain assemblage and the Nina Creek Group, in the Sylvester allochthon and the Lay Range area respectively (Figs. 1, 21; Nelson, 1993; Ferri, 1997; Plint and Gordon, 1997; Piercey et al., 1999; Creaser et al., 1999). The location of these back-arc rocks to the northeast and north of the dominantly arc assemblages suggests that subduction polarity was east- to northeast-dipping (Fig. 21). This is likely a continued evolution of the east-dipping subduction zone that characterized all previous magmatic cycles.

An interesting feature of this phase of magmatism is the dominance of mafic to intermediate material with very little felsic magmatism compared to earlier YTT episodes (e.g., Nelson, 1993; Ferri, 1997; Plint and Gordon, 1997; Murphy and Piercey, 1999; Piercey et al., 1999; Simard et al., 2003; Fig. 6). Furthermore, Cycle V mafic magmatism is characterized by very juvenile isotopic characteristics. In the Klinkit Group, arc-related samples have juvenile εNd values of +6.7 to +7.4 (Simard et al., 2003). Permian eclogites near Faro and Ross River, north of the Finlayson Lake district (Figs. 1, 24), which have back-arc-related protoliths and are inferred to be Campbell Range equivalents, have εNd = +5.4 to +9.3 (Creaser et al., 1999). Back-arc-related VMS occurrences in the Campbell Range belt have juvenile Pb-isotopic systematics (Mann and Mortensen, 2000). Collectively, these features suggest that both arc and back-arc-related rocks had minimal interaction with continental crustal materials. Two potential mechanisms can be invoked to explain this: (1) both arc and back-arc regions were built (at least in part) upon juvenile substrates; and/or (2) that the rate of extension was rapid, and coupled with rapid effusion rates and/or conduit armouring that prevented any substantial interaction with continental crustal material. Nd isotopic and trace element geochemical evidence suggests that both mechanisms were operative in YTT (Piercey et al., 2001a; Simard et al., 2003); and it is likely that both were important in explaining the dominance of juvenile material during Cycle V magmatism.

Klondike cycle (VI – 269-253 Ma) tectonic/magmatic patterns mark a significant departure from that of all previous cycles in YTT. Cycle VI is characterized by a change in subduction polarity such that the arc now faced east above a west-dipping subduction zone (Fig. 24; Mortensen, 1990). This east-facing subduction geometry is indicated by pairing of a belt of mid-Permian eclogite and blueschists along the eastern edge of the terrane (Dusel-Bacon, 1994; Erdmer et al., 1998) with mid- to Late Permian arc rocks that occur primarily in the Klondike district to the west and southwest.
Mantle Sources and Mixing During YTT Evolution

Mafic magmatic rocks along convergent margins often have complex petrological histories involving the interaction of subducted slab, mantle wedge, lithospheric mantle and continental crust (e.g., Gill, 1981; Pearce, 1983; Rogers and Hawkesworth, 1989; Pearce and Peate, 1995; Shinjo and Kato, 2000; Shinjo et al., 1999). In order to evaluate the mantle sources of mafic rocks, discriminants that specifically characterize mantle, as opposed to slab-metasomatic and crustal components must be used. In Figure 26 the element ratio Zr/Yb is plotted against Nb/Yb for mafic rocks from each magmatic cycle in YTT. This diagram is particularly useful in discriminating the relative incompatible element enrichment of the mantle source region for basaltic rocks, because Zr, Nb and Yb are all moderate to highly incompatible elements and ratios of these elements to one another are largely insensitive to slab metasomatism, crustal contamination and fractionation.

The broad range of Nb/Yb and Zr/Yb in Figure 26 shows that, regardless of age, mafic rocks in YTT were derived from heterogeneous mantle sources that ranged from depleted to enriched. In YTT, a depleted end-member mantle source played a role in the genesis of arc, and to a lesser extent, non-arc rocks. On Figure 26, most arc rocks cluster near or slightly below the N-MORB end-member and extend toward E-MORB compositions, whereas non-arc rocks generally plot between E-MORB and OIB end-members. In Cycles III and V (Figs. 26B, D), non-arc rocks show more variability and have values that extend toward the N-MORB end-member, implying that some of these rocks were derived from depleted mantle sources (e.g., BABB). The role of the depleted mantle in many modern arc and back-arc magmatic environments is well documented (McCulloch and Gamble, 1991; Woodhead et al., 1992; Hawkins, 1995; Pearce and Peate, 1995; Gribble et al., 1996; Shinjo et al., 1999). The presence of depleted mantle wedge sources for YTT arc and back-arc magmatism suggests that the arc mantle wedge and the back-arc mantle beneath YTT were similar to modern magmatic arc and back-arc environments.

In addition to depleted mantle, a component of enriched mantle is evident in most YTT arc and non-arc rocks (Fig. 26). The occurrence of mafic rocks with more enriched compositions is a feature that is commonly present in continental arc and back-arc, and in continental rift geodynamic environments (e.g., Pearce, 1983; Pearce and Peate, 1995; Shinjo et al., 1999). However, the nature and origin of this enriched component is a matter of debate. Shinjo et al. (1999) proposed an asthenospheric source (i.e., new, upwelling asthenospheric mantle) for the enriched component, whereas Pearce (1983), Hawkesworth et al. (1990) and McDonough (1990) favour a lithospheric origin (i.e., ancient, subcontinental mantle). Pearce (1983) further argued that the higher HFSE contents of continental arc magmas relative to intra-oceanic arc magmas are a result of a subcontinental lithospheric contribution to continental arc mafic magmas, in addition to depleted mantle wedge and slab components (cf. Pearce and Parkinson, 1993). Rogers and Hawkesworth (1989) illustrated a correlation between increasing incompatible element enrichment (e.g., Nb/La, Nb/Th >1) and decreasing εNd and increasing εSr in Andean basalts, indicating the influence of old subcontinental lithosphere in their genesis. Data for enriched rocks in YTT favours a lithospheric origin. Most Nb-enriched rocks in YTT (e.g., OIB-rift, NEB-suite rocks), with positive Nb anomalies relative to Th and La (Nb/Th, Nb/La ＞1), have low εNd values ranging from +1.7 to -1.6. Furthermore, a broad relationship between increasing Nb/Th and Nb/La and decreasing εNd in some YTT mafic rocks are features that cannot be explained by crustal contamination and thus require an enriched, older, subcontinental lithospheric component (Piercey, 2001; Piercey et al., 2002a, 2004).

Any model that explains the enriched component in the YTT subarc mantle wedge requires that it operated for most of the mid- to late Paleozoic evolution of the terrane (Fig. 26). We favour a simple two-component mixing model between magmas derived from a depleted mantle wedge, or depleted back-arc asthenosphere, and an enriched lithospheric component, to explain the heterogeneity of YTT mafic rocks. The linear array between depleted and enriched components on the Zr/Yb-Nb/Yb diagram supports this hypothesis. Nevertheless, there are clearly end-member magmas that do not require a mixed component. For example, the boninitic, island arc tholeiitic and N-MORB-type magmas likely reflect derivation from solely depleted to ultradepleted mantle sources (Piercey, 2001; Piercey et al., 2001a, 2004), which would probably be involved in normal arc and back-arc magmatic activity. In contrast, OIB-rift type rocks (within-plate/extensional) probably represent derivation from solely enriched lithospheric sources in response to arc rifting and the initiation of back-arc magmatic activity in the bulk of the YTT, as well as continental margin extension in the Yukon-Tanana Upland and Alaska Range (Piercey et al., 2002b; Colpron, 2001; Simard et al., 2002).
Figure 26. Zr/Yb-Nb/Yb plots for mafic rocks from the various magmatic cycles in YTT evolution. Cycles I through VI represented in (A) through (F) respectively. Diagram modified after Pearce and Peate (1995). Average values of mantle reservoirs are from Sun and McDonough (1989): N-MORB = normal mid-ocean ridge basalt (depleted); E-MORB = enriched MORB (moderately enriched); and OIB = ocean-island basalt (enriched).
Mixing between these depleted and enriched end-members is an expected response to concurrent tectonic activity. The presence of an enriched component in arc rocks likely reflects mixing of material from a depleted mantle wedge with subcontinental lithospheric mantle during migration of the melts to their ultimate destination within and upon the crust. Similarly, the occurrence of E-MORB-type signatures within YTT back-arc basins would signal mixing between upwelling depleted N-MORB type asthenosphere and enriched lithospheric material, en-route to emplacement within the back-arc regions (e.g., Campbell Range basalts and Sylvester allochthon).

It is notable that there are no temporal variations in the incompatible element behaviour of YTT mafic rocks from the mid- to late Paleozoic. This suggests that mantle sources for mafic rocks of YTT did not change appreciably throughout the terrane’s evolution, and that the mantle heterogeneity exhibited by YTT is a fundamental feature of the terrane.

Importance of Recycled Continental Crust during YTT Evolution
The importance of continental crustal recycling in modern and ancient continental margin arc and back-arc, and in rifted continental margin geodynamic environments is well established (Rogers and Hakesworth, 1989; Shinjo et al., 1999; Whalen et al., 1998; Piercey et al., 2003). In YTT, the influence of evolved continental crust is shown by the Pb, Nd and Sr isotope systematics of granitic rocks (Mortensen, 1992a). Uranium-lead zircon data from geochronological studies of granitoid rocks from the terrane commonly indicate Proterozoic and Archean inheritance (Mortensen, 1990; 1992a; Dusel-Bacon et al., 2004). Similarly, Nd isotopic studies of YTT sedimentary and felsic igneous rocks exhibit strong evidence for recycled ancient continental crust in their genesis (eNd <0 and Proterozoic T<sub>DM</sub> ages; e.g., Stevens et al., 1995; Creaser et al., 1997; Piercey et al., 2003).

Figure 27 shows plots of La and Sm (both normalized to UCC values) for felsic rocks from all magmatic cycles in YTT. Rocks derived from crustal sources lie on or near the line with a slope of 1, equivalent to La/Sm<sub>UCN</sub> (UCN = Upper Crust Normalized) = 1 (Fig. 27). Samples that plot below this line, with La/Sm<sub>UCN</sub> < 1, likely come from sources that are depleted relative to UCC, for example, mafic crust; or they may have fractionated LREE-bearing accessory phases. The case of La/Sm<sub>UCN</sub> >>1 is very rare, as it would require crustal sources with much higher LREE than the UCC; however, high La to Sm ratios could be achieved through preferential mobilization of LREE in hydrothermal fluids, or through kinetically controlled melting of LREE-enriched accessory phases at high temperatures (e.g., Bea, 1996a, b).

Most felsic rocks from YTT cluster close to the La/Sm<sub>UCN</sub> ~1 control line, implying that recycling of continental crust has been an important process throughout the history of the terrane. Furthermore, although there are variations in the absolute La and Sm concentrations between arc and non-arc rocks, there is no significant separation in terms of La/Sm<sub>UCN</sub> ratios between the two groups, suggesting that both arc and non-arc felsic rocks were derived from predominantly recycled upper continental crustal material.

This geochemical data for felsic rocks, unfortunately, can only point to an UCC-like source and cannot specifically delineate what the source of that crust was. In particular, it cannot delineate whether this crust was once part of the North American craton, or an exotic crustal fragment with UCC geochemical attributes. The similarity in stratigraphic and geochemical attributes between rocks of the Yukon-Tanana Upland and parts of the Alaska Range and those of Selwyn basin in Yukon does suggest that the basement to YTT may have once been part of the North American miogeocline. Other aspects of YTT also share similarities with the North American miogeocline including: (1) Nd isotopic attributes for YTT felsic and sedimentary rocks (Mortensen, 1992a; Stevens et al., 1995; Boghossian et al., 1996; Grant, 1997; Creaser et al., 1997; Garzione et al., 1997; Patchett et al., 1999; Piercey et al., 2003); (2) detrital zircon geochronology (Gehrels et al., 1995); (3) Pb isotopic composition of syngegetic sulphide deposits (Nelson et al., 2002; Mortensen et al., this volume); and (4) broadly similar Devonian-Mississippian geodynamic history (Paradis et al., 1998; Nelson et al., 2002, this volume). Collectively these features point to a likely link between the North American miogeocline and YTT crust, at least prior to Devonian-Mississippian back-arc initiation along the ancient Pacific continental margin (see also Nelson et al., this volume).

Importance of Recycled Oceanic Crust
It is well established that basaltic rocks can provide probes to their past mantle history, their mantle sources, and the relative importance of oceanic and continental crustal recycling in their genesis (Zindler and Hart, 1986). The abundance of magmatic rocks with incompatible element-enriched OIB signatures throughout the evolution of YTT points to the potential for a recycled oceanic crustal component in YTT magmas.

It has been shown that mafic rocks erupted in ocean islands and other large igneous province (LIP) environments commonly exhibit geochemical and radiogenic isotopic evidence for past oceanic and continental crustal recycling (White and Hofmann, 1982; Zindler and Hart, 1986; Hart et al., 1992; Hofmann, 1997, and references therein). Of particular interest to this paper is the presence of excess Nb and Ta relative to Th, La and other LREE/LFSE compared to primitive mantle values (i.e., positive Nb and Ta anomalies), features interpreted to indicate a recycled oceanic crustal component in these magmas (e.g., McDonough, 1991; Niu and Batiza, 1997; Niu et al., 1999).

The origin of positive Nb and Ta enrichments is attributed to the recycling of oceanic crust in subduction zones and its re-incorporation into the mantle wedge via convective stirring into the upper mantle (e.g., Allegre and Turcotte, 1986; Meibom and Anderson, 2003), and/or by incorporation into mantle plumes (e.g., Weaver, 1991; McDonough, 1991; Hart et al., 1992; Stein and Hofmann, 1994; Hofmann, 1997; Niu and Batiza, 1997; Niu et al., 1999; Condie, 1998, 2000). As the subducted slab descends into the subduction zone, it undergoes dehydration and transport of Th, U, LFSE and in some cases LREE, to the overlying mantle wedge, leading to the characteristic “arc” signature in subduction related basalts (e.g., Pearce, 1983; You et al., 1996; Jenner, 1996; Swiden et al., 1997; Johnson...
Figure 27. La$_{UCN}$-Sm$_{UCN}$ plots for YTT felsic rocks. Cycles I through VI represented in (A) through (F). Details of the diagram provided in the text. Solid (filled) symbols in each plot represent non-arc rocks and open (unfilled) symbols represent arc rocks. Normalization values from McLennan (2001).
Figure 28. Nb/Th$_{pm}$-Nb/La$_{pm}$ diagram for YTT mafic rocks. Symbols as in Figure 22. Cycles I through VI represented in (A) through (F). Details of the diagram provided in the text. Diagram constructed from the concept of Niu et al. (1999) (e.g., Ta/U$_{pm}$-Nb/Th$_{pm}$ diagram).
and Plank, 1999). In contrast, the stability of phases such as rutil during eclogitization results in the retention of HFSE in the subducted slab (in particular Nb, Ta and Ti; Saunders et al., 1988; McDonough, 1991; Rudnick et al., 2000; Foley et al., 2000). The preferential retention of Nb, Ti and Ta in the slab results in arc basalts with characteristic Nb depletions relative to Th and La (and other LFSE and LREE), and primitive mantle normalized Nb/La and Nb/Th ratios of less than one (Nb/Th$_{pm}$, Nb/La$_{pm}$ $< 1$; Pearce and Peate, 1995). The subducted slab, in contrast, has antithetical Nb/La and Nb/Th ratios relative to arc basalts, with primitive mantle normalized values greater than 1 (Nb/Th$_{pm}$, Nb/La$_{pm}$ $> 1$). It has been argued that the Nb/Th$_{pm}$ and Nb/La$_{pm}$ $> 1$ in OIB and E-MORB imply a recycled oceanic crustal component in the source of these basaltic rocks (e.g., Saunders et al., 1988; McDonough, 1991; Niu and Batiza, 1997; Niu et al., 1999; Rudnick et al., 2000).

On Figure 28, the data for YTT mafic rocks are shown on a Nb/Th$_{pm}$ vs. Nb/La$_{pm}$ diagram. This diagram divides mafic rocks derived from magmas of arc parentage, or that have been contaminated by continental crust (Nb/Th$_{pm}$ and Nb/La$_{pm}$ $< 1$), from rocks that come from enriched sources with an inherited oceanic crustal component (Nb/Th$_{pm}$ and Nb/La$_{pm}$ $> 1$; Fig. 28). Arc and crustally contaminated rocks invariably lie within the arc/contamination field (Nb/Th$_{pm}$ and Nb/La$_{pm}$ $< 1$; Fig. 28). Non-arc rocks are more variable. Those with a weak subduction signature (e.g., BABB) or that are crustally contaminated typically plot in the arc/contamination field (Nb/Th$_{pm}$ and Nb/La$_{pm}$ $< 1$; Fig. 28). Mafic rocks with N-MORB $+$ E-MORB geochemical signatures lie near the junction between Nb/Th$_{pm}$ $< 1$ and Nb/La$_{pm}$ $> 1$, and those with OIB $+$ E-MORB signatures fall within the enriched field (Fig. 28).

These types of enriched rocks with oceanic crustal components are commonly associated with large igneous provinces (LIP). The occurrence of enriched rocks in all cycles of YTT magmatism suggests the influence of a recycled oceanic crustal component, but may also highlight the possibility that these rocks represent formation within a LIP. If this is the case, then this LIP must have existed during the entire mid- to late Paleozoic magmatic evolution of YTT. This is highly unlikely, as it would require that the LIP persisted for more than 150 m.y. and influenced an area in excess of 250,000 km$^2$ in the northern Cordillera. Furthermore, the geological events that dominated YTT history such as arcs, arc rifts, intra-arc compressional events, and extensional back-arc basin activity (e.g., Nelson et al., this volume), coupled with the lack of evidence for voluminous, high-temperature basaltic magmatism, are inconsistent with a LIP geodynamic environment such as an ocean island or oceanic plateau (e.g., Hawaii, Ontong-Java).

Although formation within a LIP can be ruled out for the mid-to late Paleozoic evolution of YTT, the presence of mafic rocks with signatures similar to rocks from such environments highlight the potential role that a LIP may have had in its ancestry. In particular, the recycled oceanic lithospheric signatures in enriched YTT basalts could be explained by inherited LIP melts residing as veins within a subcontinental lithospheric mantle domain in the terrane. For example, Stein and Hofmann (1992, 1994) argued that in the Arabian-Nubian shield, past LIP-related material (frozen plume heads) froze against and/or was incorporated into the lithospheric mantle. With subsequent rifting and melting of the lithospheric mantle, these ancient LIP components were recycled into younger basalts with signatures akin to the preexisting LIP (Stein and Hofmann, 1992, 1994). Pre-Paleozoic fossilized LIP material in the YTT subcontinental lithospheric mantle could also explain the recycled oceanic crustal component in YTT enriched mafic magmas. An interesting consequence of this model is the question of the provenance of the LIP magmatism (plume?) that fertilized the YTT lithospheric mantle.

**Figure 29.** The occurrence of OIB-like signatures in YTT rocks over 150 m.y. points to a potential plume component in their genesis. Enriched rocks associated with the Gunbarrel and Franklin large igneous provinces (LIP) resulted in widespread magmatism and igneous activity along the northwestern edge of Laurentia in Neoproterozoic time (~780-720 Ma; Heaman et al., 1992; Ernst and Buchan, 2001; Harlan et al., 2003). During this event, partial melts from the Franklin plume fertilized the subcontinental lithospheric mantle of northwestern Laurentia resulting in a veined lithospheric mantle, which retained the geochemical signature of plume-derived rocks (i.e., OIB to E-MORB). These veins were later liberated during partial melting related to YTT arc and back-arc magmatism, which developed on top of western Laurentian continental margin fragments during the middle to late Paleozoic. SCLM = Subcontinental lithospheric mantle.
The ancient Pacific margin of North America was the locus of a LIP in the Late Proterozoic. In particular, two Neoproterozoic events, the Gunbarrel (~780 Ma; Harlan et al., 2003) and Franklin igneous events (~720 Ma; Heaman et al., 1992; Park et al., 1995; Dupuy et al., 1995; Dudas and Lustwerk, 1997), resulted in more than 1,250,000 km² of mafic magmatism extending from Wyoming, along the cratonic margin of western North America, through Arctic Canada and into Greenland (Ernst and Buchan, 2001). These events resulted in widespread mafic dike swarms, mafic sills and flood basalt extrusions over this extended region and collectively have been interpreted to be a product of the Neoproterozoic breakup of Rodinia along the western margin of Laurentia (Fig. 29; Heaman et al., 1992; Park et al., 1995; Dudas and Lustwerk, 1997; Harlan et al., 2003). It is possible that the widespread mafic activity associated with these LIP events was responsible for the fertilization of the subcontinental mantle of the northern Cordillera, resulting in a lithospheric mantle with veins of Nb-enriched material within it, which upon later reactivation during continental rifting or arc rifting resulted in the OIB-like signatures observed in both the miogeoclone and YTT (e.g., Goodfellow et al., 1995; Dusel-Bacon and Cooper, 1999; Piercey et al., 2002a; Nelson and Friedman, 2004). Although equivocal, the presence of Neoproterozoic depleted mantle model ages in uncontaminated alkalic basalts supports the hypothesis that these rocks had their initial origins as part of the LIP associated with the Neoproterozoic breakup of Rodinia (e.g., Piercey et al., 2002a, 2004; S.J. Piercey and R.A. Creaser, unpublished data).

SUMMARY AND CONCLUSIONS

Yukon-Tanana terrane has had a varied petrological history with complex interactions between crust, mantle and subducted slab; ultimately these petrological variations can be tied to the geodynamic evolution of the terrane. The main conclusions of this paper can be outlined as follows:

1. YTT comprises six cycles of magmatic activity prior to its Mesozoic accretion to the North American craton. The first five cycles occurred above an east-dipping subduction zone that developed along the distal edge of the North American craton, whereas the sixth (Klondike) magmatic cycle took place above a west-dipping subduction zone.

2. Felsic magmatic rocks associated with the extending continental margin in the Alaska Range and Yukon-Tanana Upland (Cycle I) are dominantly crustally-derived peralkaline rocks with lesser crustally-derived subalkaline rocks; they are accompanied by mafic rocks of intraplate affinity.

3. During cycles II through V, magmatism was largely bimodal in nature, with mafic and felsic end-members. Mafic rocks with arc signatures are predominantly calc-alkaline and island-arc tholeiitic, with lesser LREE-enriched island arc tholeites and boninites. In corresponding back-arc environments, mafic rocks are dominated by normal mid-ocean ridge basalts, enriched mid-ocean ridge basalts and ocean island basalt signatures. Ocean island basalt-like rocks are also present in many arc-dominated successions, and are interpreted to record intra-arc rift events within YTT. Felsic rocks in arc environments are dominated by calc-alkaline affinities with lesser tholeiitic rocks; back-arc rocks are characterized by HFSE- and REE-enriched (A-type) affinities.

4. Mafic rocks from both arc and non-arc environments in YTT originate from variably enriched mantle domains, ranging from ultra-depleted (boninites) to enriched (OIB). There is a continuum of compositions between these end-members, but with arc rocks tending toward more depleted compositions, and non-arc rocks tending toward more enriched compositions. Rocks with intermediate signatures can be explained by mixing between the enriched and depleted end-members in the mantle source region. There are no inter-cycle variations in the composition of the YTT mafic rocks, implying that both the underlying mantle and the processes of magma generation remained similar throughout its mid- to late Paleozoic evolution.

5. Felsic rocks from the YTT, although varying in absolute HFSE and REE contents, are predominantly derived from recycled upper continental crust (UCC). Most felsic rocks have LREE-enrichment and relatively flat REE patterns relative to UCC (La/Sm$_{UCC}$ ≈1). This, coupled with published Nd, Sr and Pb isotopic data, and the common occurrence of inherited zircon, all imply derivation from crustal protoliths. This UCC protolith was similar to the North American craton; however, more data is required to establish definitively whether the basement of YTT was the North American craton.

6. Many mafic rocks from YTT intra-arc rifts and back-arc basins have Nb/Th$_{SM}$ and Nb/La$_{SM}$ > 1, implying excess Nb relative to Th and La relative to the primitive mantle. Excess Nb relative to Th and La in the primitive mantle in these rocks implies a recycled oceanic crustal component in their genesis. Recycled oceanic crust is a feature common in rocks associated with many large igneous provinces (LIP); however, since this component exists in all cycles of YTT magmatism, it is highly unlikely that these rocks represent a LIP, because YTT magmatism spans more than 150 m.y. and affects an area in excess of 250,000 km² over this time frame. This signature in YTT alkaline basalts likely reflects the reactivation of recycled LIP components resident in the lithospheric mantle that were originally derived via lithospheric fertilization by LIP magmatism during the Neoproterozoic breakup of Rodinia.

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