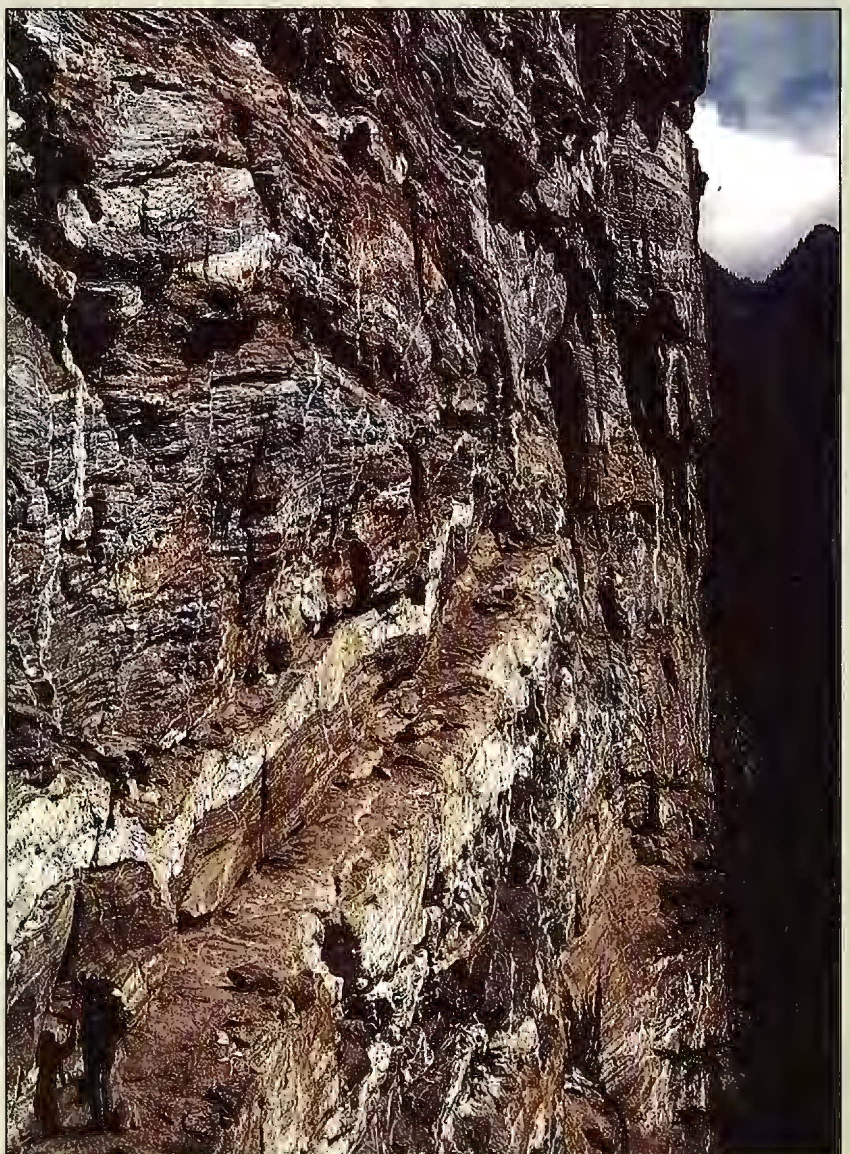
GEOLOGICAL SURVEY OF CANADA
PAPER 91-2

RADIOGENIC AGE AND ISOTOPIC STUDIES: REPORT 5



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Cover description

In the Valhalla Range of the southeastern Canadian Cordillera, the Gwillim Creek shear zone is exposed in this cliff face on the north side of the Gwillim Creek drainage. Highly deformed paragneiss with concordant pegmatitic intrusions about 80-90 Ma old are in the foregound of the photograph whereas grey 100 Ma Mulvey mylonitic orthogneiss is visible in the highest parts of the cliff face. The Gwillim Creek shear zone is a zone of major mid-crustal thrust displacement separating these two lithologies, and it was active between about 80 and 52 Ma ago. Photograph courtesy of R.R. Parrish.

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CONTENTS

RADIOGENIC AGE AND ISOTOPIC STUDIES: REPORT 5

INTRODUCTION

"Radiogenic Age and Isotopic Studies" is an annual collection of reports presenting data from the Geochronology Laboratory of the Continental Geoscience Division of the Geological Survey of Canada. The main purpose of this collection is to make geochronological and other radiogenic isotope data produced by the laboratory available promptly to the geological community. Reports make full presentation of the data, relate these to field settings, and make comparatively short interpretations. Readers are cautioned that some data reported here are part of work in progress, and more extensive publications may follow at a later date. Other geochronological and isotope data produced in the laboratory but published in other scientific journals or separate GSC publications are summarized at the end of this report.

Report 5 contains papers from most regions of Canada, but particularly from British Columbia. The Geochronology Laboratory has, over the years, provided substantial U-Pb dating for the Cordilleran Division of the Geological Survey of Canada in Vancouver, and the results of a number of these studies are presented this year. Many of these Cordilleran dating projects were begun more than four years ago, and we are pleased to be able to bring many of these to a conclusion with this volume.

The geochronology laboratory depends not only on the financial resources and scientific expertise of the Continental Geoscience Division of which it is part, but also on other groups within the Geological Survey. The Mineralogy Section of the Mineral Resources Division provides us with mineral separations and rock powders which are carefully and tediously prepared from generally large (10-30 kg) rock samples. For this we thank G. Gagnon, B. Machin, R. Christie, and R. Delabio. It also provides us with very high quality scanning electron photomicrographs of mineral grains for morphological studies. Some of these have been prominantly displayed on previous covers of this publication series. For mineralogical assistance we thank A. Roberts, D. Walker, and L. Radburn. Finally, the Analytical Chemistry Section of the Mineral Resources Division allows us access to an atomic absorption spectrometer for potassium analyses for K-Ar dating. We are thankful for all of this collective assistance.

INTRODUCTION

«Âges radiométriques et études isotopiques» est une collection de rapports annuels qui présentent des données provenant du Laboratoire de la géochronologie de la Division géoscientifique du continent. Ces rapports ont pour but de rendre les données géochronologiques et les autres données sur les isotopes radiogéniques produites par ce laboratoire facilement accessibles à la communauté géologique. On trouve dans ces rapports une présentation complète des données, le lien qui existe entre ces dernières et la situation sur le terrain ainsi que des interprétations comparativement courtes. Le lecteur doit toutefois savoir que certaines données reproduites dans ces rapports proviennent de travaux en cours et que des publications plus détaillées pourraient suivre. D'autres données géochronologiques et isotopiques produites au laboratoire, mais publiées dans d'autres revues scientifiques ou des publications distinctes de la CGC sont résumées à la fin du présent rapport.

Le rapport n° 5 contient des textes portant sur la plupart des régions du Canada, mais en particulier sur la Colombie-Britannique. Le Laboratoire de la géochronologie a, au cours des ans, produit de nombreuses datations U-Pb pour la Division de la Cordillère de la Commission géologique du Canada à Vancouver, et les résultats d'un certain nombre de ces études sont présentés cette année. Nombre des projets de datation dans la Cordillère ayant été entrepris il y a plus de quatre ans, il nous fait plaisir de tirer une conclusion à plusieurs d'entre eux dans le cadre du présent volume.

Le Laboratoire de la géochronologie dépend non seulement des ressources financières et des compétences scientifiques de la Division géoscientifique du continent, dont elle fait partie, mais aussi d'autres groupes de la Commission géologique. La Section de la minéralogie de la Division des ressources minérales lui fournit des séparation de minéraux et des poudres de roches qui ont été soigneusement et minutieusement préparées à partir d'échantillons généralement grands (de 10 à 30 kg). Nous tenons donc à en remercier MM. G. Gagnon, B. Machin, R. Christie et R. Delabio. Cette section a également fourni d'excellentes microphotographies à balayage électronique de grains de minéraux destinées à des études morphologiques. Certaines de ces photographies ont été avantageusement représentées en page couverture d'anciens numéros de la série. Nos remerciements s'adressent donc également à MM. A. Roberts, D. Walker et L. Radburn. Enfin, la Section de la chimie analytique de la Division des ressources minérales nous a donné accès au spectromètre d'absorption atomique pour y effectuer des analyses du potassium aux fins de datation par la méthode K-Ar. Cette aide collective nous a été, soulignons-le, d'une grande utilité.

Randall R. Parrish
Vicki J. McNicoll

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U-Pb geochronology of Yellowknife Supergroup felsic volcanic rocks in the Russell Lake and Clan Lake areas, southwestern Slave Province, Northwest Territories

J.K. Mortensen[1], J.B. Henderson[1] , V.A. Jackson[2], and W.A. Padgham[2]

Mortensen, J.K., Henderson, J.B., Jackson, V.A., and Padgham, W.A., 1992: U-Pb geochronology of Yellowknife Supergroup felsic volcanic rocks in the Russell Lake and Clan Lake areas, southwestern Slave Province, N.W.T.; in Radiogenic Age and Isotopic Studies: Report 5, Geological Survey of Canada, Paper 91-2, p. 1-7.

Abstract

U-Pb zircon ages of 2658.0 +1.2/-0.8 Ma and 2661.3 +1.2/-1.1 Ma were determined for felsic volcanic rocks in the Russell Lake and Clan Lake areas, respectively, in the southwestern Slave Province. These are some of the youngest supracrustal rocks dated thus far in the Slave Province. Lithological and age similarities between the volcanic and iron-formation-bearing sedimentary rocks in the Russell Lake area and those in the Contwoyto Lake area suggest that these sequences may be related.

Résumé

Des datations U-Pb sur zircon de 2658,0 +1,2/-0,8 Ma et de 2661,3 +1,2/-1,1 Ma ont été obtenues pour des roches volcaniques felsiques dans les régions des lacs Russell et Clan, respectivement, dans le sud-ouest de la province des Esclaves. Certaines roches supracrustales, parmi les plus récentes dans la province des Esclaves, remontent d'aussi loin. Les roches volcaniques et sédimentaires ferrifères des régions du lac Russell et du lac Contwoyto présentent une lithologie et un âge semblables, indiquant une corrélation entre ces séquences.

INTRODUCTION

The Yellowknife Supergroup in the Slave Province (Fig. 1) consists of variable proportions of volcanic rocks, both mafic and felsic, and turbiditic greywacke that typically overlies and dominates the sequence. Minor iron-formation and rare quartzite and conglomerate occur locally. U-Pb dating studies of the supracrustal rocks (e.g. Henderson et al., 1987; van Breemen et al., 1987, 1988; van Breeman, pers. comm., 1991; Mortensen et al., 1988, 1992; van Breemen and Henderson, 1988; Isachsen et al., 1991a,b) indicate that the bulk of the volcanic rocks in the Yellowknife Supergroup were deposited in the interval of 2715-2663 Ma. Felsic volcanic rocks older than 2715 Ma have been recognized by Isachsen et al. (1991a,b) in the Yellowknife greenstone belt. These older units mainly comprise the Dwyer Formation, which may pre-date the Yellowknife Supergroup (Isachsen et al., 1991c). Relatively old volcanic rocks may also be present in the Point Lake area (Mortensen et al., 1988).

In this study we have dated felsic volcanic units from two sequences in the southwestern Slave Province. One date is from the Clan Lake volcanic complex, which is a small volcanic accumulation interlayered within the metaturbiditic Burwash Formation in the Yellowknife area (Fig. 1). The second date is from the Russell Lake belt near the western edge of the Slave Province (Fig. 1). These two ages provide new constraints on the evolution of the younger part of the Yellowknife Supergroup.

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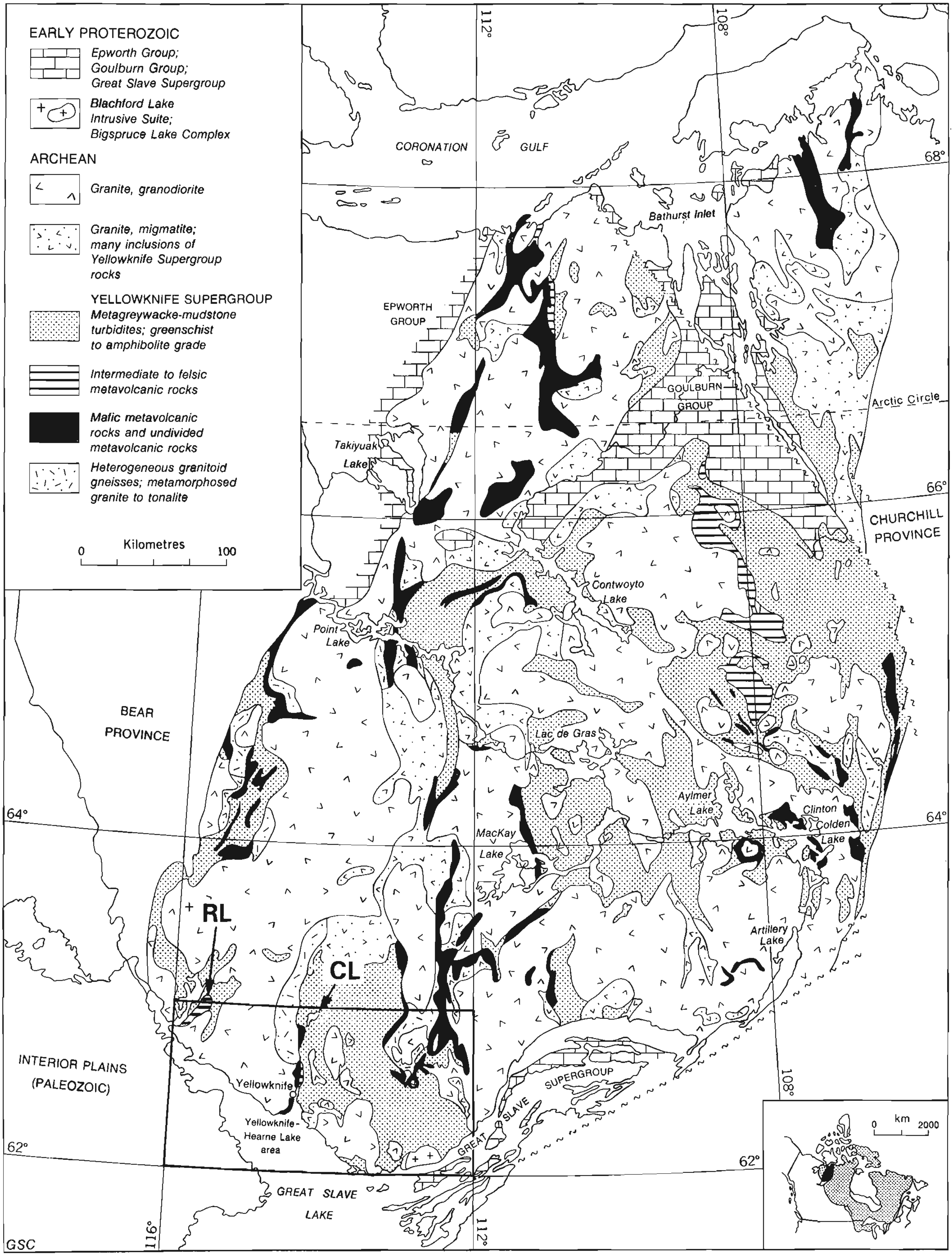

Figure 1. Simplified geological map of the Slave Province (modified from Henderson, 1985). RL = Russell Lake; CL = Clan Lake.

GEOLOGICAL SETTING AND SAMPLE DESCRIPTIONS

A continuous belt of supracrustal rocks extends for more than 300 km along the western edge of the Slave Province, stretching north from Russell Lake in the south to the west end of Point Lake (Fig. 1). This belt consists mainly of metaturbiditic rocks; however significant volcanic accumulations occur near the northern end of the belt and in the Indin Lake area (Fig. 1). A northeast-trending belt of mainly dacitic to rhyolitic flows and volcaniclastic rocks occurs to the south in the Russell Lake area (Fig. 2). Narrow belts of mafic volcanic rocks also occur in the area, but are separated from the main felsic sequence by metagreywacke. The sedimentary rocks resemble those of the turbiditic Burwash Formation northeast of Yellowknife (Henderson, 1985), but also include scattered occurrences of iron-formation. Most of the iron-formations in the Russell Lake belt are of the silicate facies type; however, carbonate facies iron-formation is found locally in the southern part of

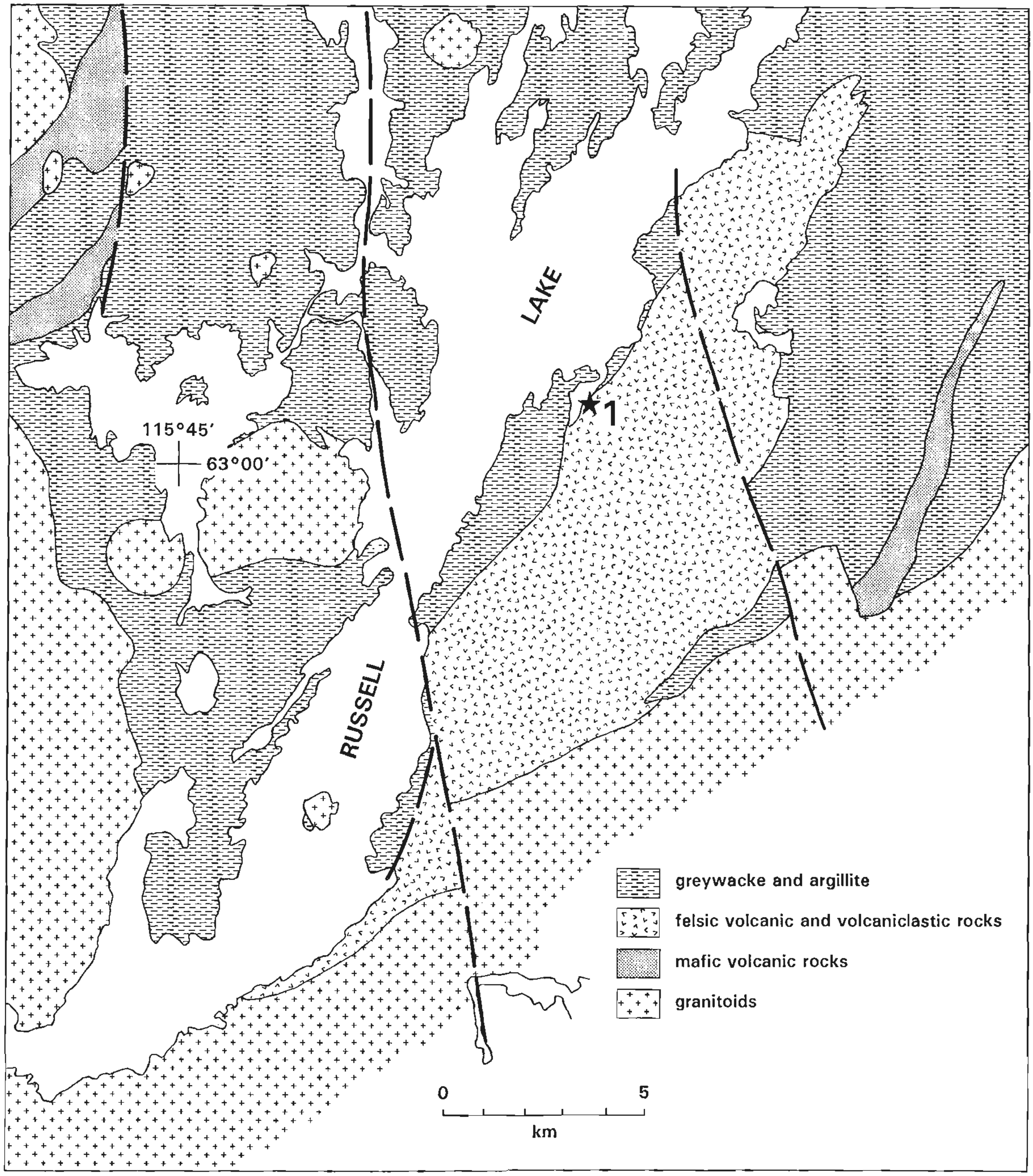

Figure 2. Simplified geological map of the Russell Lake belt (modified from Henderson (1985) and Jackson (1990). Numbered star shows location of Sample 1.

the belt. In rare instances thin sulphide-bearing layers that carry anomalous gold values have also been observed (Henderson, 1985; Jackson, 1990). In this sense the region has certain affinities with the Point Lake-Contwoyto Lake gold-bearing iron formation district (Padgham, 1990a).

Sample 1 was collected in the Russell Lake belt from near the stratigraphic top of a vertically dipping, thick sequence of dacitic to rhyolitic volcaniclastic rocks, stratigraphically below the contact with overlying greywacke (Fig. 2). The sample is from a vertically-dipping sequence of moderately foliated, fine grained felsic tuff and lapilli tuff which is interlayered with very rusty weathering carbonate. Similar rusty weathering carbonate is also commonly present as matrix to the tuff. The sampled unit is a layer 1 m thick of fine feldspar-phyric rhyolitic tuff with rare, fine, stretched quartz eyes.

In the Clan Lake area (Fig. 3), an irregularly shaped volcanic complex occurs within Burwash Formation metaturbidites (Hurdle, 1984; Henderson, 1985). The belt consists mainly of felsic (dacitic to rhyolitic) volcanic and volcaniclastic rocks, although basalt flows also occur along the eastern and southwestern margins of the complex, and andesite is found locally within it. The felsic rocks include massive flows and a small dome; however, the bulk of the complex comprises a variety of tuffs and volcanic breccias. The volcanic rocks are gradational over a narrow stratigraphic interval into the underlying and overlying metaturbidites.

Sample 2 was collected from a well foliated, thinly bedded sequence of felsic quartz-feldspar crystal tuffs and lapilli tuffs near the western edge of the Clan Lake complex (Fig. 3).

ANALYTICAL TECNHIQUES

Zircons were separated from 25 kg samples using conventional Wilfley table and heavy liquid methods. Analytical techniques employed in this study have been summarized by Parrish et al. (1987). Regressions of discordant U-Pb data use the model of Davis (1980). Compositions for initial common Pb in analyses are taken from the model of Cumming and Richards (1975). Errors in ages are quoted at the 2σ level.

ANALYTICAL RESULTS

Zircons from Sample 1 form clear, pale pink, unzoned, stubby prisms with no visible cores but abundant clear rod- and bubble-shaped inclusions. Most grains are euhedral, but some show slightly resorbed surfaces. Three fractions were analyzed (Table 1). They contain relatively low

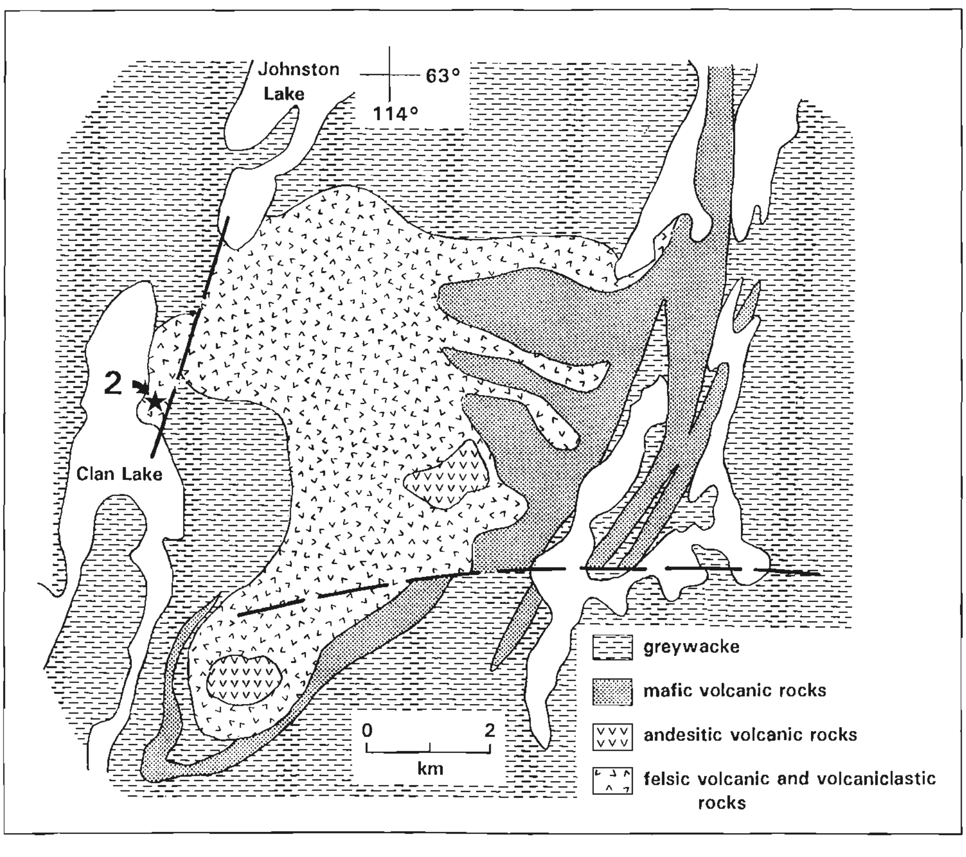

Figure 3. Simplified geological map of the Clan Lake volcanic complex (modified from Hurdle, 1984). Numbered star shows location of Sample 2.

Table 1. U-Pb analytical data

	Fraction, Size[1]	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$ [3]	^{208}Pb[2] (%)	$^{206}Pb/^{238}U$ [4]	$^{207}Pb/^{235}U$ [4]	$^{207}Pb/^{206}Pb$ [4]	$^{207}Pb/^{206}Pb$ age[5]
	Sample 1 (MLB-88-006; 63°00.7'N, 115°43.1'W)									
A	N2,+149,<M	0.030	50.1	29.2	9006	11.3	0.50890(.10)	12.668(.11)	0.18054(.03)	2657.9(1.0)
C	N2,+105,>M	0.026	64.7	37.4	6531	10.7	0.50861(.09)	12.656(.10)	0.18048(.03)	2657.3(0.9)
D	N2,+149,>M	0.008	108	60.3	4516	8.4	0.50304(.12)	12.509(.13)	0.18035(.03)	2656.1(1.1)
	Sample 2 (MLB-88-010; 62°56.9'N, 114°15.0'W)									
A	N1,+149,<M	0.044	214	120.3	32490	14.7	0.47192(.08)	11.760(.10)	0.18073(.03)	2659.6(0.9)
B	N1,+105,>M	0.021	207	121.1	20230	14.0	0.49405(.08)	12.318(.10)	0.18083(.03)	2660.5(0.9)
C	N1,+105,<M	0.021	161	93.7	15910	13.6	0.49412(.09)	12.319(.10)	0.18082(.03)	2660.4(0.9)
D	N1,+105,clr	0.014	247	145.8	12000	18.1	0.47529(.09)	11.838(.10)	0.18065(.03)	2658.8(0.9)

[1]sizes (+62-74 refer to size of zircons in microns; N1,2=non-magnetic cut with Frantz at 1 or 2 degrees side slope; <M,>M = magnetic with "pin", clr=clear
[2]radiogenic Pb
[3]measured ratio, corrected for spike and fractionation
[4]corrected for blank Pb and U and common Pb (errors quoted are 1σ in percent)
[5]corrected for blank and common Pb (errors are 2σ in Ma)
Initial common Pb compositions from Cumming and Richards (1975)

concentrations of U (50-108 ppm). The analyses range from 0.3-1.3% discordant, and define a short linear array (probability of fit = 49%) with upper and lower intercept ages of 2658.0 +1.2/-0.8 Ma and 322 Ma, respectively (Fig. 4a). The upper intercept gives the crystallization age of the rock, and the lower intercept indicates mainly recent Pb-loss.

Zircons recovered from Sample 2 form pale to medium pinkish brown, stubby to elongate prisms. Most grains are euhedral, although a small proportion of them have slightly resorbed surfaces. No cores were visible, but the zircons typically contain abundant clear rod- and bubble-shaped inclusions, and display vague of prominently developed zoning. Four fractions were analyzed (Table 1). Measured U contents are moderate (161-247 ppm). The analyses range from 3.3-7.6% discordant, and define a linear array (probability of fit = 37%) with upper and lower intercept ages of 2661.3 +1.2/-1.1 Ma and 69 Ma, respectively (Fig. 4b). The upper intercept gives the crystallization of the rock, and the lower intercept indicates mainly recent Pb-loss.

DISCUSSION AND CONCLUSIONS

The two ages reported here for volcanic rocks from the Russell Lake and Clan Lake areas are the youngest ages obtained thus far for Archean supracrustal rocks in the Slave Province. The young age for the Clan Lake sample is consistent with models in which the Burwash Group metaturbites mainly overlie volcanic rocks of the Yellowknife greenstone belt. The age obtained for the Russell Lake belt is very similar to that from the Clan Lake volcanic complex, indicating that these two supracrustal sequences formed at the same time, and may be related. Metaturbidites in the Russell Lake area resemble those of the Burwash Group, but also include iron formation. The volcanic/sedimentary sequence of the Russell Lake belt is also similar to the metaturbidites and felsic volcaniclastic rocks in the Fingers Lake area south of the Lupin Mine in the north-central Slave Province, for which a maximum depositional age of 2661 Ma has been inferred from detrital zircon dating studies (Mortensen et al., 1992).

The tectonic significance of extensive belts of supracrustal rocks in the western part of the Slave Province is uncertain. In recent arc-continent collision models for the Slave Province suggested by Kusky (1989, 1990a,b; see also discussions by King et al. (1989) and Padgham (1990b), these belts are considered to be either: 1) klippe thrust westwards over a continental block now represented by scattered remnants of pre-Yellowknife Supergroup crystalline rocks in the western part of the craton or 2) remnants of arc (or rift) volcanic and sedimentary sequences that are unrelated to those in the central and eastern Slave Province. The lithological and age similarities between the Russell Lake belt and supracrustal rocks in the central part of the Slave Province suggest that these sequences may be related. Supracrustal rocks may have originally formed a stratigraphically continuous sequence extending from the Yellowknife greenstone belt to the western margin of the craton. Pre-Yellowknife Supergroup basement appears to be scarce or absent in the region between the Yellowknife and Russell Lake belts, which is mainly underlain by granitoids (and their deformed equivalents) that post-date the Yellowknife Supergroup (e.g. Dudas et al., 1990; Atkinson and Fyfe, 1991; Isachsen et al., 1991c). Much more work is needed, however, to completely understand the nature and tectonic significance of supracrustal rocks in the western Slave Province.

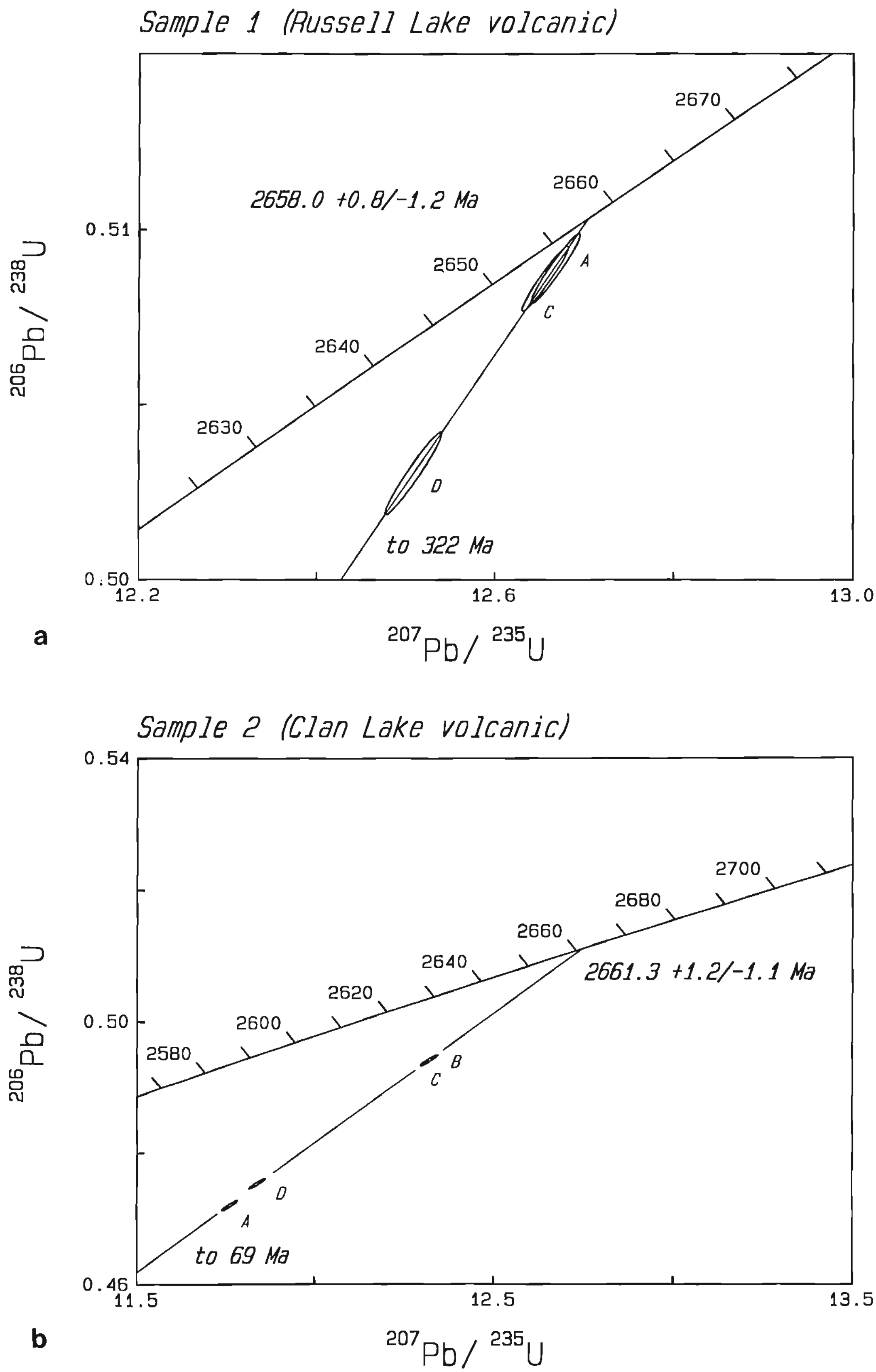

Figure 4. U-Pb concordia plots for Sample 1 (Russell Lake volcanic; Fig. 4a) and Sample 2 (Clan Lake volcanic; Fig. 4b).

ACKNOWLEDGMENTS

We thank the staff of the Geochronology Laboratory at the Geological Survey of Canada (Ottawa) for assistance in the analytical procedures. The manuscript was improved by critical reviews by J.E. King, O. van Breemen and R.I. Thorpe.

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U-Pb zircon ages from the Shallow Bay volcaniclastic belt, Contwoyto Lake area, Northwest Territories: age constraints for Lupin-type iron-formation

J.K. Mortensen[1], C. Relf[2], W.J. Davis[3], and J.E. King[1]

Mortensen, J.K., Relf, C., Davis, W.J., and King, J.E., 1992: U-Pb zircon ages from the Shallow Bay volcaniclastic belt, Contwoyto Lake area, Northwest Territories: age constraints for Lupin-type iron-formation; in Radiogenic Age and Isotopic Studies: Report 5, Geological Survey of Canada, Paper 91-2, p. 9-15.

Abstract

Single detrital zircon grains from a sample of volcaniclastic rock of intermediate to felsic composition from the Shallow Bay volcaniclastic belt south of the Lupin gold mine yield a range of U-Pb ages. One grain is concordant at 3005.9 Ma, and four are concordant to slightly discordant in the range of 2698.7-2724.2 Ma. A single broken grain fragment that may be either detrital or a microphenocryst crystallized from the Shallow Bay magmas themselves gives a slightly discordant age of 2661.0 Ma. This is a maximum depositional age for the volcaniclastic unit, and as such indicates that the stratigraphic sequence that hosts iron-formations and the gold deposits in the Lupin Mine area is younger than most of the greenstone belts in the Slave Province. This sequence may correlate with lithologically and temporally similar rocks in the Russell Lake belt in the southwestern Slave Province.

Résumé

La datation U-Pb de grains de zircon détritiques individuels provenant d'un échantillon de roche volcano-clastique de composition d'intermédiaire à felsique reposant dans la zone volcano-clastique de Shallow Bay au sud de la mine d'or Lupin donne un intervalle d'âges. Un grain est concordant à 3005,9 Ma et quatre sont concordants à légèrement discordant dans l'intervalle de 2698,7-2724,2 Ma. Un fragment de grain cassé qui est d'origine détritique ou un microphénocristal cristallisé provenant des magmas de Shallow Bay donne un âge légèrement discordant de 2661,0 Ma. L'unité volcano-clastique comporte un âge de sédimentation maximale, indiquant que la séquence stratigraphique qui renferme les formations ferrifères et les gisements aurifères dans la région de la mine Lupin est plus récente que la plupart des zones de roches vertes de la Province des Esclaves. Cette séquence pourrait être corrélée avec les roches de lithologie et d'âge semblables dans la zone du lac Russell dans le sud-ouest de la Province des Esclaves.

INTRODUCTION

Silicate (± oxide) facies iron-formation is widespread within variably metamorphosed turbiditic sediments of the Late Archean Contwoyto Formation in the Contwoyto Lake area, north-central Slave Province (e.g. Bostock, 1980; King et al., 1989; Lhotka and Nesbitt, 1989) (Fig. 1). Locally, sulphidic zones within these iron-formations host significant gold mineralization, such as at the Lupin Mine (Fig. 1) and in a great number of smaller occurrences (e.g. Kerswill, 1986; Lhotka and Nesbitt, 1989). The presence of iron-formation has been used as a first order exploration criterion for defining

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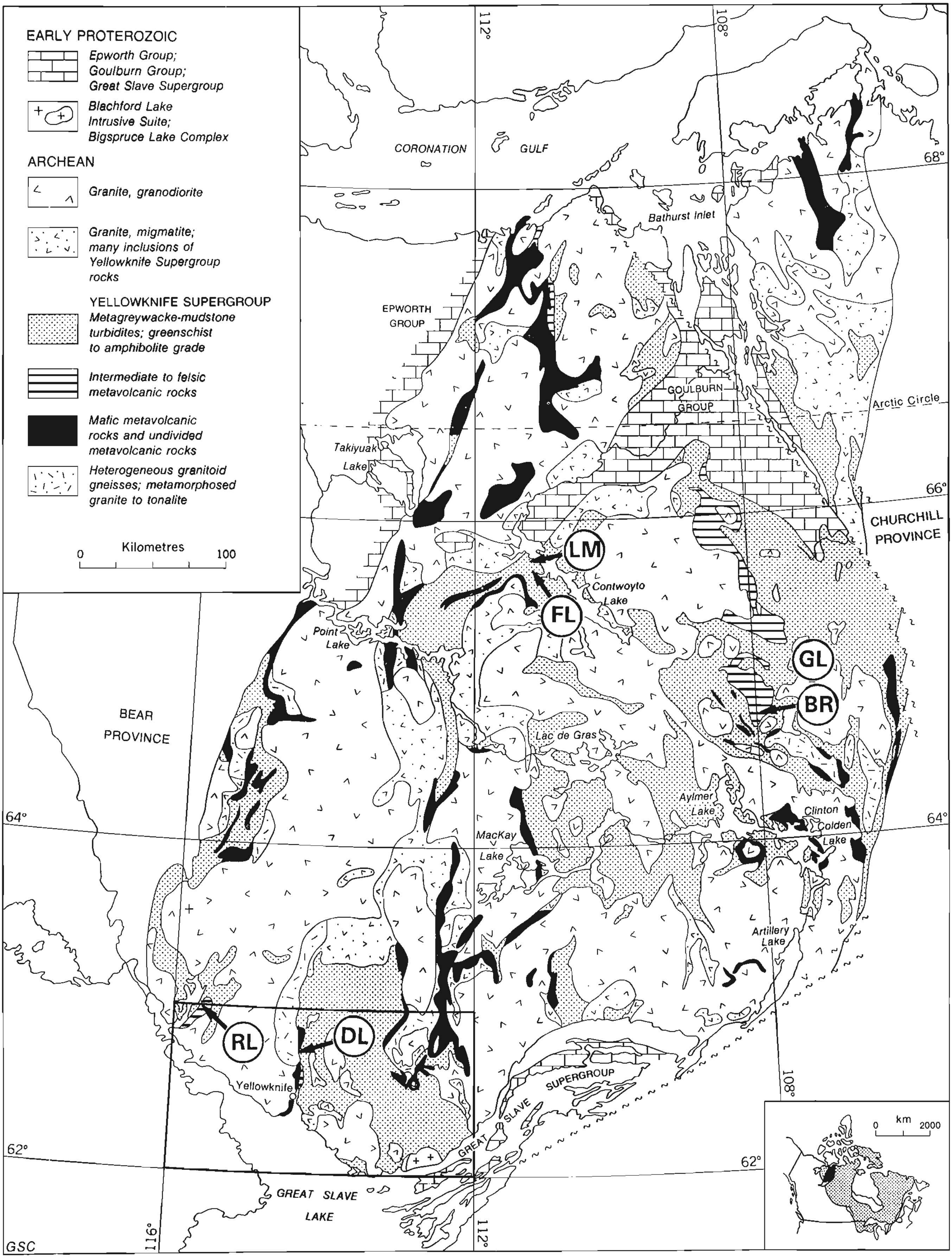

Figure 1. Locations of dated iron-formations in the Slave Province (modified from Henderson, 1985). FL = Fingers Lake; LM = Lupin gold mine; GL = George Lake; BR = Back River; DL = Dwyer Lake; RL = Russell Lake.

areas of high gold potential in the Contwoyto Lake area and elsewhere in the Slave Province. It is clear, however, that more than one age of iron-formation occurs in the Slave Province. In the Back River area (Fig. 1), Jefferson et al. (1989) identified a minimum of three separate stratigraphic units of chemical sedimentary rocks containing variable amounts of oxide, carbonate, silicate and sulphide facies iron-formation. These iron-formations, some of which contain anomalous levels of gold, occur both as intercalations within volcanic and sedimentary rocks of the Back River Complex, and in the overlying turbidites. One felsic unit of the Back River Complex has been dated by U-Pb (zircon) methods at 2692 ± 2 Ma (van Breemen et al., 1987). Gold-bearing iron-formation at George Lake (C.W. Jefferson, pers. comm.; Fig. 1) has been dated at 2683 ± 2 Ma by O. van Breemen (pers. comm.). Isachsen et al. (1991a,b,c) obtained an age of 2.8 Ga for a sequence of sub-Yellowknife Supergroup felsic metavolcanic rocks that contain thin bands of iron-formation in the Dwyer Formation near Dwyer Lake north of Yellowknife (Fig. 1). Mortensen et al. (1992)

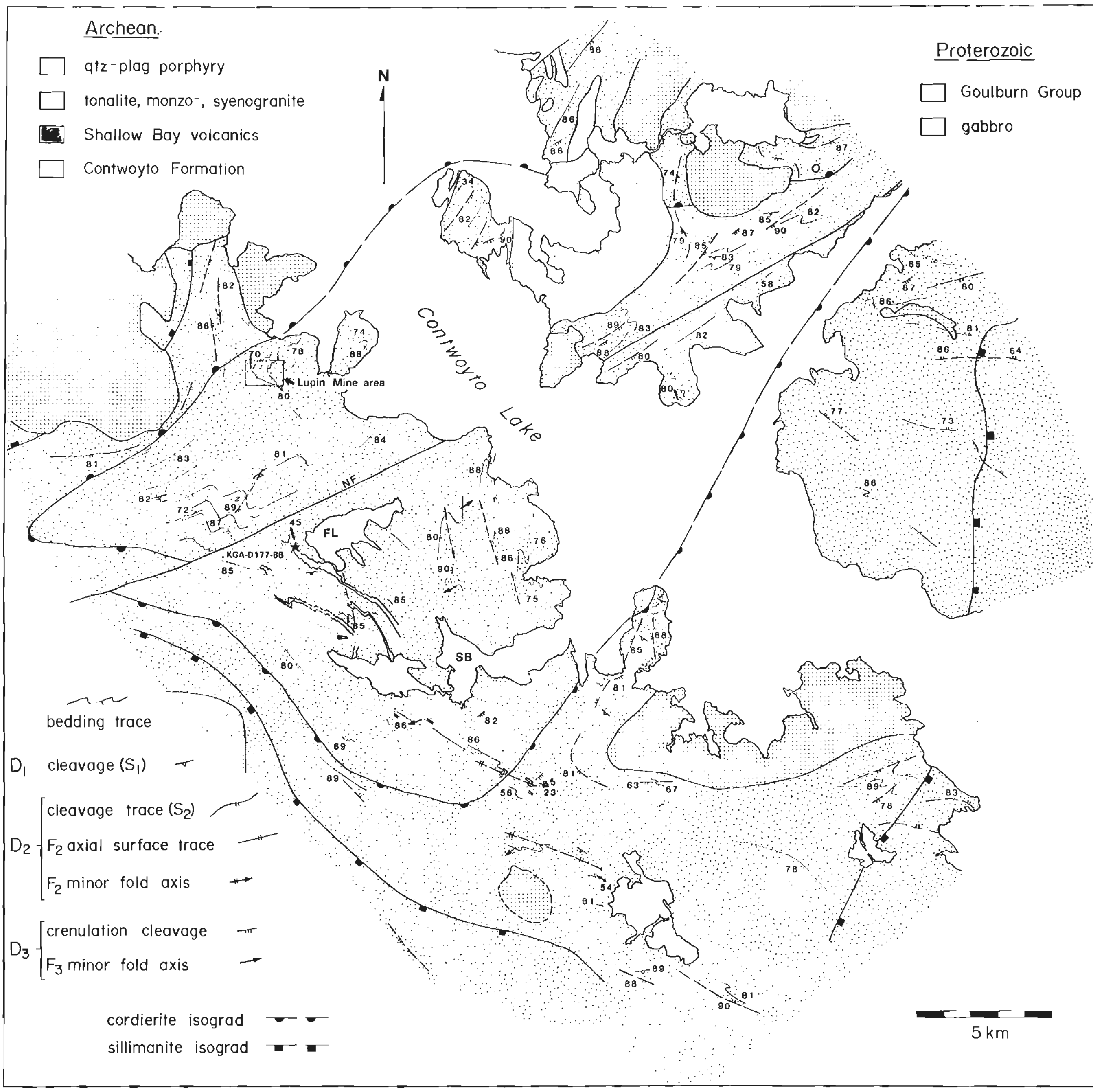

Figure 2. Simplified geological map of the Fingers Lake area. Sample locality is shown by the star. Geology by Relf (1989). FL = Fingers Lake; SB = Shallow Bay; NF = Norma Fault.

reported a U-Pb zircon age of 2658 ± 1 Ma for felsic metatuffs associated with metagreywackes containing silicate-sulphide and carbonate facies iron-formation in the Russell Lake area (Fig. 1).

The origin of gold mineralization within iron-formations in the Slave Province is controversial; both syngenetic and epigenetic models have been advanced (e.g. Kerswill, 1986, 1990; Lhotka and Nesbitt, 1989; Bullis, 1990). If the gold in the stratiform, Lupin-type deposits is in fact syngenetic, one might reasonably expect that iron-formations of different ages might have different potential for hosting significant gold mineralization. In this study we have attempted to constrain the age of the host rocks for the Lupin Mine, which is the largest iron-formation-hosted gold deposit in the Slave Province. We report U-Pb zircon data from a sequence of volcaniclastic rocks that are interbedded with the iron-formation-bearing part of the Contwoyto Formation in the Fingers Lake area just south of the Lupin Mine (Fig. 1, 2). Our results place constraints on the possible age of these iron-formations, and suggest possible relationships with volcanic belts elsewhere in the Slave Province.

GEOLOGY OF THE FINGERS LAKE AREA

Fingers Lake is located about 7 km south-southeast of the Lupin Mine (Fig. 2). In this area, thin (< cm) layers of silicate facies iron-formation occur within a sequence of mainly thin bedded turbidites. The coarser grained beds reach medium sand size, and consist of 30-40% quartz and abundant feldspar. The metamorphic grade in this area is relatively low (biotite grade), and primary sedimentary structures are very well preserved. The Shallow Bay volcaniclastic sediments (Lhotka, 1988; Relf, 1989) form several distinct layers, as much as 30 m thick, within the metaturbidite sequence. They consist mainly of epiclastic sediments of intermediate composition. Feldspar phenocrysts are commonly preserved. The thickness of individual beds ranges from 2 mm to 10 cm in some areas, to 5 m or more in the thicker bedded portions of the section. Volcaniclastic units with lapilli- and block-sized clasts are also present, and a mafic to intermediate unit with possible pillow structures was observed in one locality.

The area has been multiply folded, and most lithological contacts are steeply dipping to vertical. The contacts between the Shallow Bay volcaniclastic rocks and the adjacent sedimentary rocks are conformable and locally gradational, however, and Relf (1989) concluded that the volcanic and sedimentary rocks are in primary depositional contact.

ANALYTICAL TECHNIQUES

Zircon separation was done using conventional Wilfley table and heavy liquid methods. Analytical techniques employed in this study have been summarized by Parrish et al. (1987). All analyses reported are from strongly abraded single grains. Errors in ages are quoted at the 2σ level.

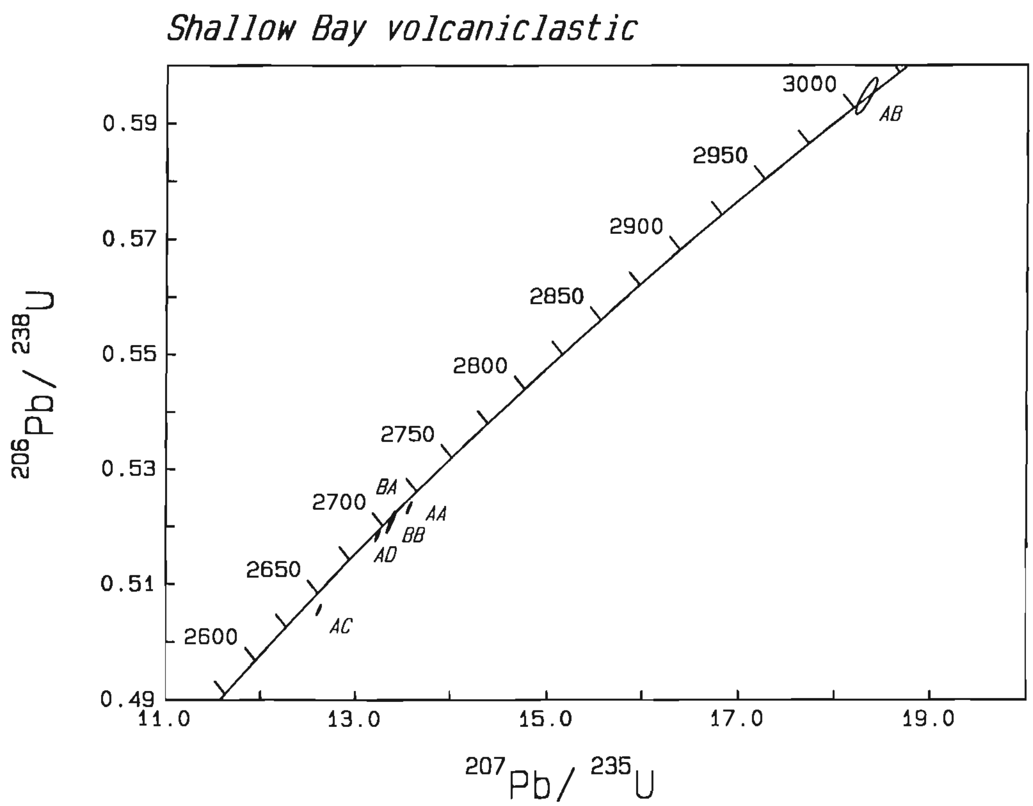

Figure 3. U-Pb concordia plot for single zircons from the Shallow Bay volcanics.

ANALYTICAL RESULTS

Zircon was separated from a 25 kg sample of massive volcaniclastic rock of intermediate composition collected west of Fingers Lake (see Fig. 2 for location). The sample contains <5% white feldspar phenocrysts from 2-10 mm in diameter. The zircon is heterogeneous in appearance, ranging from well rounded to irregular, anhedral grains. Most grains are very clear, with rare clear bubble- and rod-shaped inclusions. They are unzoned to very faintly zoned, and range from very pale pink to medium pinkish brown in color. Poorly preserved facets are present on a few grains. The surfaces of some of the rounded grains show a slight frosting. This observation, together with the absence of any euhedral grains, and the mainly intermediate composition and volcaniclastic to epiclastic nature of the rocks, suggest that the zircons are detrital in origin. The age(s) of the zircons therefore can only be used to provide a maximum age for the volcanism and associated turbidite deposition.

Fourteen single grains, representing the range of grain colour and morphology observed, were selected. The six best quality, coarsest grains remaining after strong abrasion were analyzed. Analytical results for these grains are given in Table 1 and shown in Figure 3. The analyses range from 0 to 1.1% discordant. Four of the analyses give ^{207}Pb-^{206}Pb ages in the range of 2698.7 to 2724.2 Ma. A nearly colorless, well rounded grain is concordant with a ^{207}Pb-^{206}Pb age of 3005.9 Ma. A large, medium pinkish brown grain fragment with concoidally fractured surfaces has a ^{207}Pb-^{206}Pb age of 2661.1 Ma. This analysis is 1.1% discordant; thus the actual crystallization age of this grain may be slightly older than the ^{207}Pb-^{206}Pb age of the analysis.

DISCUSSION

Our results are consistent with the inferred detrital origin for most of the zircon grains in this sample. The very small degree of discordance of each of the analyses, together with the lack of evidence for inherited cores in any of the grains analyzed, indicates that each age should correspond to a specific igneous event in the source region(s). Potential sources for 2724-2699 Ma zircons include felsic igneous rocks in the Anialik River area (e.g. Abraham et al., 1991), the Yellowknife greenstone belt (e.g. Isachsen et al., 1991a,b), and possibly the Point Lake area (e.g. Mortensen et al., 1988). Igneous zircon ages of 3.0 Ga have been obtained from many widely separated exposures of basement in the western part of the Slave Province (e.g. Williams et al., 1991).

The relatively young age of 2661 Ma for fraction C (Table 1) is the most significant result of this study. Although this single grain is interpreted to be detrital, its morphology does not preclude the possibility that the grain is a microphenocryst from the Shallow Bay magmas. Thus the age of the grain is a *maximum* age for deposition of the Shallow Bay volcaniclastic rocks, and may actually be the crystallization age for the unit. The data indicate that the Shallow Bay volcaniclastic rocks are slightly younger than volcanic rocks in the Gondor Lake area about 32 km southwest of Fingers Lake, which were dated by Mortensen et al. (1988) at 2667.6 +3.0/-1.6 Ma. The Shallow Bay volcaniclastic rocks may be synchronous with felsic volcanic rocks in the Clan Lake and Russell Lake areas (Fig. 1), dated at 2661 Ma and 2658 Ma, respectively (Mortensen et al., 1992).

Table 1. U-Pb analytical data

Fraction, Size[1]	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$[3]	^{208}Pb[2] (%)	$^{206}Pb/^{238}U$[4]	$^{207}Pb/^{235}U$[4]	$^{207}Pb/^{206}Pb$[4]	$^{207}Pb/^{206}Pb$ age[5]
Sample KGA-D175-88 (65°42.3'N, 111°13.0'W)									
AA N2,+149,clr	0.011	117	75.1	2117	16.6	0.52329(.10)	13.560(.11)	0.18794(.04)	2724.2(1.2)
AB N2,+149,clr	0.007	31.2	23.1	310	15.1	0.59461(.27)	18.322(.31)	0.22348(.14)	3005.9(4.6)
AC N2,+149,pink	0.018	79.9	45.6	4671	10.0	0.50554(.09)	12.609(.10)	0.18089(.03)	2661.1(1.0)
AD N2,+149,brn	0.007	286	182.8	3175	17.0	0.51851(.09)	13.230(.10)	0.18505(.03)	2698.7(1.1)
BA N2,+105-149	0.005	114	67.1	916	9.7	0.52120(.15)	13.378(.16)	0.18616(.06)	2708.5(2.0)
BB N2,+105-149	0.004	97.2	56.3	1246	8.3	0.52025(.13)	13.355(.14)	0.18617(.05)	2708.7(1.6)

[1]sizes (+62-74 refer to size of zircons in microns; N1,2=non-magnetic cut with Frantz at 1 or 2 degrees side slope; clr=clear, brn=brown

[2]radiogenic Pb

[3]measured ratio, corrected for spike and fractionation

[4]corrected for blank Pb and U and common Pb (errors quoted are 1σ in percent)

[5]corrected for blank and common Pb (errors are 2σ in Ma)

Initial common Pb compositions from Cumming and Richards (1975)

The Shallow Bay volcaniclastic rocks are interpreted to be in depositional contact with the adjacent metaturbidites, and there is no evidence for any significant unconformity between the two units. Our age data for the volcanics therefore indicate that the iron-formations within the metaturbidites in the Fingers Lake area, and likely those that host gold mineralization at the nearby Lupin Mine (8 km to the north) as well, are <2661 Ma in age, and therefore represent some of the youngest rocks of the Yellowknife Supergroup. Iron-formations in this area are therefore much younger than those at George Lake or in the Back River complex, and are temporally correlated with iron-formations exposed at Russell Lake (Jackson, 1990; Mortensen et al., 1992). This observation may have important implications for syngenetic gold exploration models, particularly because stratiform, Lupin-type gold mineralization occurs in the Russell Lake area (J.A. Kerswill, pers. comm.).

CONCLUSIONS

Single detrital zircon grains from the Shallow Bay volcaniclastic belt in the Fingers Lake area of the north-central Slave Province yield concordant to slightly discordant ages that range from 3006 to 2661 Ma. The older grains may be derived from volcanic and/or plutonic terranes in the western Slave Province. The youngest grain, at 2661 Ma, gives a maximum age for deposition of the Shallow Bay volcanics, and, by inference, for the iron-formations that occur within turbiditic sediments that are associated with the volcanic units. This indicates that the iron-formations which host the stratiform gold mineralization at Lupin are significantly younger than iron-formations in either the Dwyer Formation north of Yellowknife, or in the Back River and George Lake areas. The iron-formations in the Contwoyto Lake area are more similar in age to those associated with felsic metavolcanic rocks in the Russell Lake area in the southwestern Slave Province.

ACKNOWLEDGMENTS

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Geochronology of granites along the margin of the northern Taltson Magmatic Zone and western Rae Province, Northwest Territories

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Abstract

U-Pb zircon ages for quartz diorite gneiss and an unfoliated syenogranite within mixed gneisses of the western Rae Province yield ages of 2400 ±10 Ma and 2394 ± 2 Ma respectively. These ages are from the extensive paragneiss-free mixed gneiss region which is distinguished from a narrow sliver of paragneiss-bearing mixed gneiss along the western margin of the Rae Province. The new data, together with previously published geochronology, suggest that the paragneiss-free region is characterized by igneous rocks in the 2.44 - 2.39 Ga interval; apparently contrasting with the paragneiss-bearing mixed gneisses where plutonic rocks as young as 2.27 Ga have been documented.

U-Pb zircon and monazite data show that a spatially continuous body of granite, hitherto called Gagnon granite, which is separated from the Taltson Magmatic Zone (TMZ) to the west by a 5 km wide N-S zone of mylonite, is probably made up of two intrusions. These comprise a megacrystic monzogranite (1940 ± 8 Ma) similar to Konth batholith within TMZ, and a significantly younger equigranular biotite granite (1882 +7/-6 Ma) which is similar in age to granite veins within Archean rocks within Great Slave Shear Zone.

Résumé

La datation par la méthode U-Pb sur zircon d'un gneiss à diorite quartzique et d'un syénogranite contenus dans des gneiss mélangés reposant dans l'ouest de la province de Rae donne respectivement 2400 ± 10 Ma et 2394 ± 2 Ma. Ces âges s'appliquent à la région de gneiss mélangé ne contenant pas de paragneiss, qui se distingue d'un fragment étroit de gneiss mélangé renfermant du paragneiss le long de la bordure occidentale de la province de Rae. Les nouvelles données, combinées à celles publiées antérieurement, indiquent que la région sans paragneiss est caractérisée par des roches ignées dont l'intervalle d'âge est de 2,44 à 2,39 Ga, ce qui semble contraster avec l'âge des gneiss mélangés renfermant du paragneiss où des roches plutoniques récentes (2,27 Ga) ont été documentées.

Les données U-Pb sur zircon et monazite indiquent qu'un massif de granite spatialement continu, appelé ici granite Gagnon, qui est séparé de la ZMT à l'ouest par une zone nord-sud de mylonite de 5 km de large, est probablement composé de deux intrusions : un monzogranite mégacristallin (1940 ± 8 Ma), semblable au batholite de Konth dans la ZMT, et un granite à biotite isogranulaire beaucoup plus récent (1882 +7/-6 Ma) qui est d'âge équivalent aux filons de granite dans les roches archéennes de la zone de cisaillement de Great Slave.

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INTRODUCTION

The western part of the Rae Province (Fig. 1) is for the most part separated from Taltson Magmatic Zone (TMZ) by a complex, northerly trending, mostly ductile, and predominantly strike slip shear zone. Previous geochronology has been reported by Bostock et al. (1987,1991) and Bostock and Loveridge (1988). Although granite bodies of the Taltson Magmatic Zone have yielded ages in the range 1.99 Ga to 1.92 Ga, those within the ductile strike slip zone and mixed gneisses immediately to the east feature an extended range of previously reported igneous ages from 2.44 Ga to 1.81 Ga. Pre-TMZ igneous ages range from 2.44 Ga to 2.26 Ga.

Recent mapping has covered a corridor of some 30 km width east of the eastern fault boundary of TMZ (Bostock, 1982, 1986, 1987, 1988). Extensive early ductile shear within this corridor was later concentrated in discrete ductile to brittle zones which anastomose, thereby breaking the corridor into a number of major wedge-like fault blocks (Bostock, 1988). Rocks consist largely of mixed gneisses, but a large granite complex, the Gagnon granite, which bears some lithological similarity to the major plutons of TMZ, intrudes the mixed gneisses at the northern end of the corridor, and is separated from the main batholiths of TMZ by a N-S trending dextral mylonite belt, part of the major fault zone. The mixed gneiss is unconformably overlain by Nonacho Group (Aspler,1985) and both are intruded by diabase dykes of the Sparrow Dyke Swarm (McGlynn et al., 1974) for which a preliminary U-Pb age of 1.83 Ga has been obtained (unpublished data on baddeleyite).

Farther south, smaller plutons of Taltson (1.99 - 1.92 Ga) and younger age have been emplaced along the boundaries of some of the fault blocks. They include the Natael muscovite monzogranite pluton (1935 +/-3 Ma; Bostock et al., 1987) at Hill Island Lake (south of the area shown in Fig. 1); the Benna Thy monzogranite (1906 +3/-1 Ma; Bostock

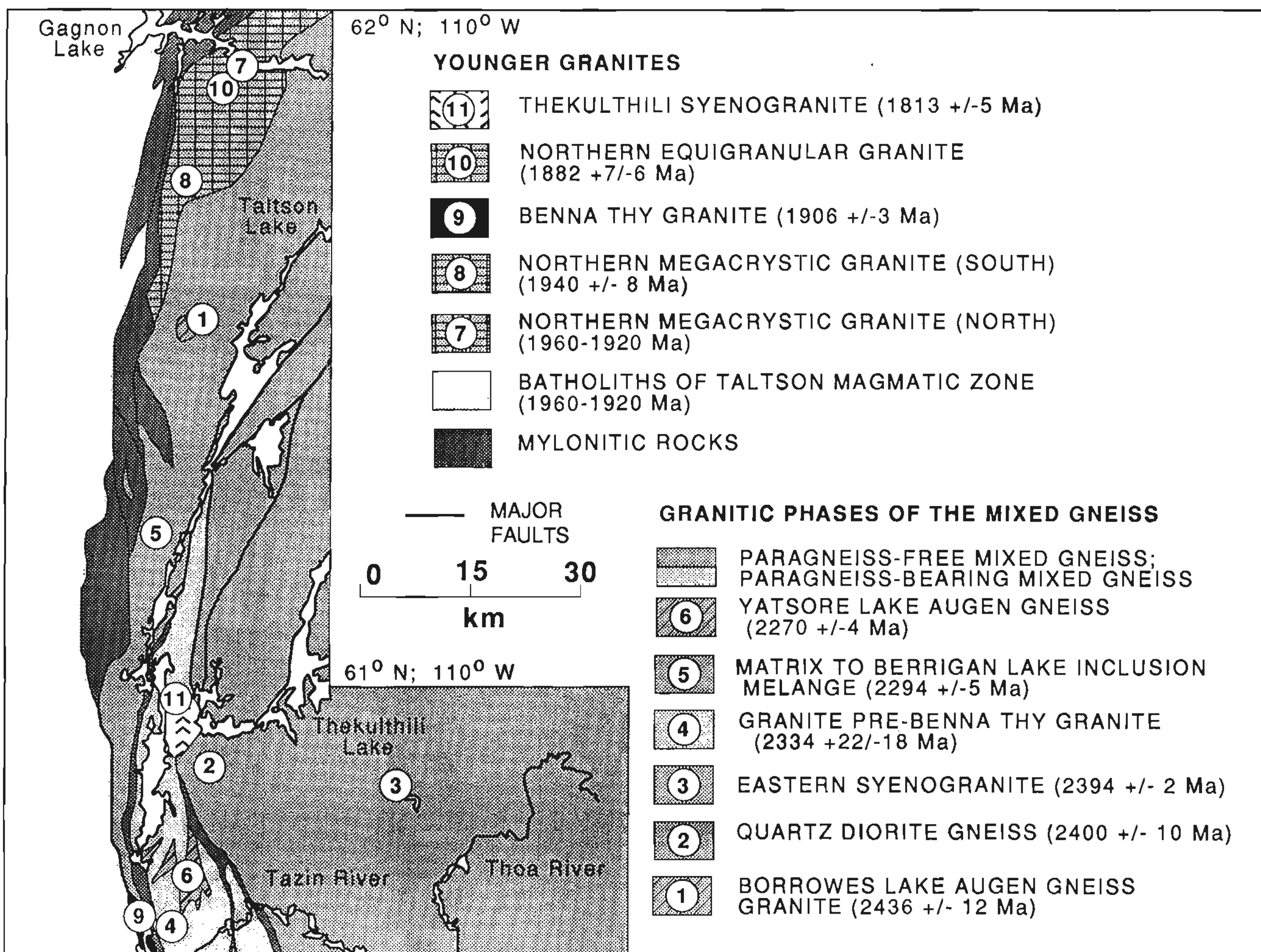

Figure 1. Plutonites of the mapping corridor along the western margin of the Rae Province. These plutonites comprise two zones of mixed granite and gneiss characterized by the presence and near abscence of paragneiss. Granitic phases dated are listed in the legend in order of decreasing U-Pb zircon, monazite or sphene age. Sources: localities 2,3,7,8 and 10 (this study); localities 1,4 and 9 (Bostock and Loveridge, 1988); localities 5 and 6 (Bostock et al., 1991). Supracrustal rocks (eg. Nonacho Group) are not shown.

and Loveridge, 1988; locality 9, Fig. 1) at Benna Thy Lake; and the unfoliated Thekulthili syenogranite (1813 ± 5 Ma, Bostock et al., 1991; locality 11, Fig. 1) at the southeast corner of Thekulthili Lake.

The Natael pluton like the Konth Batholith of TMZ farther west, locally displays sinistral shear bands and thus provides a maximum age for at least part of the sinistral shear that has affected the gneisses of the corridor. The Benna Thy pluton is less sheared and does not show any sinistral kinematic indicators despite the presence of sinistrally sheared gneisses and a major band of mylonite along its western side. Rather it has a dextral C and S fabric and is bordered on the west by mylonite showing dextral shear bands. Furthermore, dextral kink bands were found to deform the mylonites on the east margin of the pluton. The minimum age of the sinistral shear and the maximum age of the dextral shear are therefore thought to be given by the 1906 Ma age of the Benna Thy pluton.

The present study addresses two aspects of the regional geochronology. Firstly, two new samples have been investigated in an attempt to better understand the earlier data and define the age of granitic plutonism evident in the mixed gneisses east of TMZ. Secondly, two phases of the unit heretofore called Gagnon granite have been dated. In this work these two granites will be collectively referred to as the northern granites. Analytical techniques for U-Pb zircon and monazite analysis (Table 1) are summarized in Parrish et al. (1987). Treatment of analytical errors follows Roddick (1987). Uncertainties for ages are quoted at the 2σ level.

MIXED GNEISSES

The mixed gneisses consist predominantly of foliated, medium- to fine-grained granitic rocks of monzogranitic to dioritic composition with a varied abundance of undeformed to strongly deformed mafic inclusions incorporated into the gneissosity of their host. The dominant mafic mineral phase in the inclusions, schlieren or bands may be hornblende, biotite or chlorite-epidote. Scattered throughout these mixed gneisses are small, conformable (but locally crosscutting), medium- to fine-grained bodies of syenogranite, and pegmatite with gradational contacts. Conformable plutons of foliated, variably mafic, potash feldspar megacrystic granite to augen gneiss (here referred to as augen gneiss) are also present.

Recognizable paragneisses consisting predominantly of retrograded sillimanite-bearing metapelites and quartz-rich gneisses are a common component only within a major crustal fault wedge in the southern half of the corridor (Fig. 1).

Table 1. U-Pb data

Fraction,[1] Size	Weight (mg)	U (ppm)	Pb^2 (ppm)	$^{206}Pb/^{204}Pb$ [3]	Pb_c [4] (pg)	^{208}Pb [2] (%)	$^{206}Pb/^{238}U$ ±SEM%[5]	$^{207}Pb/^{235}U$ ±SEM%[5]	R	$^{207}Pb/^{206}Pb$ age, error (Ma)[6]
Quartz diorite gneiss (87BK-366X; Easting 540500, Northing 6751820, Zone 12)										
A, +105,N2.5	0.009	165	84	2039	19	11.7	0.4474 ± (.12)	10.108 ± (.13)	0.96	2495.7 (1.3)
B, -105,N2.5	0.011	1091	499	3604	82	4.2	0.4406 ± (.09)	9.494 ± (.10)	0.94	2415.9 (1.2)
C, -105,N2.5	0.019	1371	571	1681	340	3.8	0.4046 ± (.09)	8.422 ± (.12)	0.87	2356.8 (2.0)
H, 300*100,N2.5, S	0.002	2268	1074	5279	22	6.7	0.4448 ± (.09)	9.483 ± (.10)	0.96	2397.6 (1.1)
F, -105,N2.5	0.011	865	373	1388	155	4.6	0.4151 ± (.09)	8.808 ± (.12)	0.85	2389.7 (2.2)
G, -105,N2.5	0.005	903	384	2992	38	5.3	0.4059 ± (.09)	8.666 ± (.11)	0.94	2400.3 (1.3)
D, sphene	0.240	9	5	135	616	2.6	0.6089 ± (.27)	9.397 ± (1.1)	0.61	1831.0 (34)
E, sphene	0.159	13	4	108	426	2.9	0.3248 ± (.38)	4.992 ± (1.3)	0.66	1823.2 (38)
Syenogranite (88BK-172X; Easting 565870, Northing 6744900, Zone 12)										
A, ≈120*100,N0.5	0.029	191	93	8769	17	9.2	0.4467 ± (.09)	9.565 ± (.10)	0.96	2405.3 (1.0)
C, ≈200*90,N0.5	0.011	555	262	3752	41	7.5	0.4378 ± (.09)	9.553 ± (.10)	0.95	2437.0 (1.1)
E, ≈150*80,M0.5	0.002	171	87	846	12	12.0	0.4498 ± (.17)	9.563 ± (.18)	0.95	2393.1 (1.9)
G, ≈300*180,M0.5,S	0.015	281	137	2901	38	10.5	0.4409 ± (.08)	9.364 ± (.10)	0.94	2391.4 (1.2)
H, ≈150*100,M0.5	0.008	168	83	2464	14	9.5	0.4495 ± (.09)	9.569 ± (.11)	0.95	2395.1 (1.2)
I, ≈150*100,M0.5	0.015	106	53	261	160	11.4	0.4433 ± (.12)	9.394 ± (.39)	0.66	2387.5 (11)
Megacrystic granite (north) (87BK-276X; Easting 539430, Northing 6864800, Zone 12)										
A, -74,	0.014	1315	488	3638	102	10.1	0.3458 ± (.08)	5.792 ± (.10)	0.93	1977.8 (1.4)
B, -74,N2	0.020	839	307	677	482	10.5	0.3400 ± (.09)	5.561 ± (.20)	0.70	1935.7 (5.4)
C, sphene	0.134	117	71	153	2243	45.5	0.3429 ± (.22)	5.619 ± (.83)	0.70	1939.2 (25)
D, sphene	0.174	108	64	162	2509	44.0	0.3441 ± (.22)	5.639 ± (.79)	0.67	1939.0 (24)
Megacrystic granite (south) (89-BK-10X; Easting 528930, Northing 6845620, Zone 12)										
A, ≈140*70	0.008	1317	534	4099	56	12.2	0.3682 ± (.08)	6.242 ± (.10)	0.94	1999.6 (1.3)
B, ≈140*50	0.006	2397	979	4751	66	16.8	0.3524 ± (.08)	5.799 ± (.10)	0.94	1946.6 (1.2)
C, ≈100*40	0.008	1140	444	6175	29	11.9	0.3566 ± (.09)	5.842 ± (.10)	0.95	1938.7 (1.1)
D, ≈110*50	0.005	2808	1055	2776	91	13.5	0.3380 ± (.09)	5.471 ± (.11)	0.91	1917.1 (1.6)
F, monazite,S	0.002	2824	4270	6758	21	77.8	0.3471 ± (.08)	5.671 ± (.10)	0.95	1933.7 (1.1)
Z, monazite	0.007	5252	7058	27508	29	74.9	0.3487 ± (.09)	5.692 ± (.10)	0.96	1932.1 (1.0)
X, monazite	0.008	4699	6105	28355	29	73.9	0.3501 ± (.09)	5.752 ± (.10)	0.96	1943.7 (1.0)
Equigranular granite (89BK-12X; Easting 534710, Northing 6857010, Zone 12)										
A, ≈120*70,N4	0.009	5466	1874	6519	143	10.1	0.3216 ± (.09)	5.051 ± (.10)	0.95	1862.8 (1.2)
B, ≈140*70,N4	0.004	2141	743	6119	23	17.4	0.2974 ± (.09)	4.923 ± (.10)	0.95	1956.9 (1.1)
C, ≈200*90,N4	0.005	3953	1316	6422	51	10.2	0.3120 ± (.09)	4.881 ± (.10)	0.95	1855.6 (1.1)
E, ≈100,N4	0.008	4987	1808	5327	149	10.5	0.3381 ± (.09)	5.368 ± (.10)	0.95	1882.1 (1.2)

Notes: [1]sizes in microns for zircons before abrasion (either -74+62; through 74 micron sieve but not the 62 micron sieve or ≈100*50; average length and breadth aspects); M and N refer to magnetic and non-magnetic at side slope indicated in degrees; S indicates analysis of a single crystal [2]radiogenic Pb; [3]measured ratio, corrected for spike and fractionation; [4]total common Pb in analysis corrected for fractionation and spike; [5]corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; R correlation of errors in isotope ratios; [6]corrected for blank and common Pb, errors are two sigma in Ma.

Quartz diorite gneiss (87BK-366X; locality 2, Fig. 1)

This sample represents one of the more mafic-rich gneisses within the mixed gneiss about 4 km south of Thekulthili Lake and 1.5 km east of Thekulthili River. The rock is medium grained and weakly foliated with gneissosity defined by lenticular concentrations of mafic minerals. Major minerals, quartz, andesine, olive biotite, and blue-green hornblende are anhedral. Minor epidote with traces of apatite, small rounded zircon, and carbonate are also present. The rock is cut by small irregular granite pegmatites.

Zircons analyzed are prismatic with length to breadth (L/B) ratios of 2:1 and feature strong igneous zoning as well as numerous fine cracks. Cores are common. Uranium concentrations are variable (Table 1). On a concordia diagram (Fig. 2) the data points are strongly scattered. Fraction A with the lowest U content has the oldest ^{207}Pb-^{206}Pb model age of 2496 Ma which is interpreted as an Archean inherited component. The least discordant data point H corresponds to a single zircon grain with the highest U content of 2268 ppm. Inspection of the zircons under transmitted light suggests that grain H and fraction F are least likely to contain core material. A line through these points intersects the concordia at 2399 and 253 Ma. Lines drawn through data points H and G and H and C yield respective ages of 2397 Ma and 2404 Ma. The age of crystallization of igneous zircon is tentatively assigned at 2400 ± 10 Ma.

Syenogranite (88BK-172X; locality 3, Fig. 1)

This sample represents a larger example of one of the many small, to medium sized largely unfoliated syenogranite bodies which are conformable to discordant within the mixed gneiss. The rock is medium grained, massive and pink. It contains rare angular, fine grained, altered mafic inclusions. Major minerals are microcline perthite, albite, and quartz. Muscovite and chlorite are minor minerals. Small amounts of carbonate are disseminated in plagioclase and along grain boundaries. Zircon is common. Traces of opaque are partially rimmed by sphene.

Zircons are clear, stubby and subhedral to irregular in shape. Uranium concentrations range from 555 ppm to 106 ppm. Data points are scattered on a concordia diagram (Fig. 3) and it is likely that fractions C and A contain an inherited component. Fractions E and H yield concordant data points, both in agreement with an age of 2394 ± 2 Ma. This age is interpreted in terms of the time of igneous crystallization and emplacement of the syenogranite.

THE NORTHERN GRANITES

The northern granites considered here involve a single spatially continuous plutonic body heretofore called the Gagnon granite. Geochronology (this paper) now suggests there are two distinct granites involved: 1) equigranular

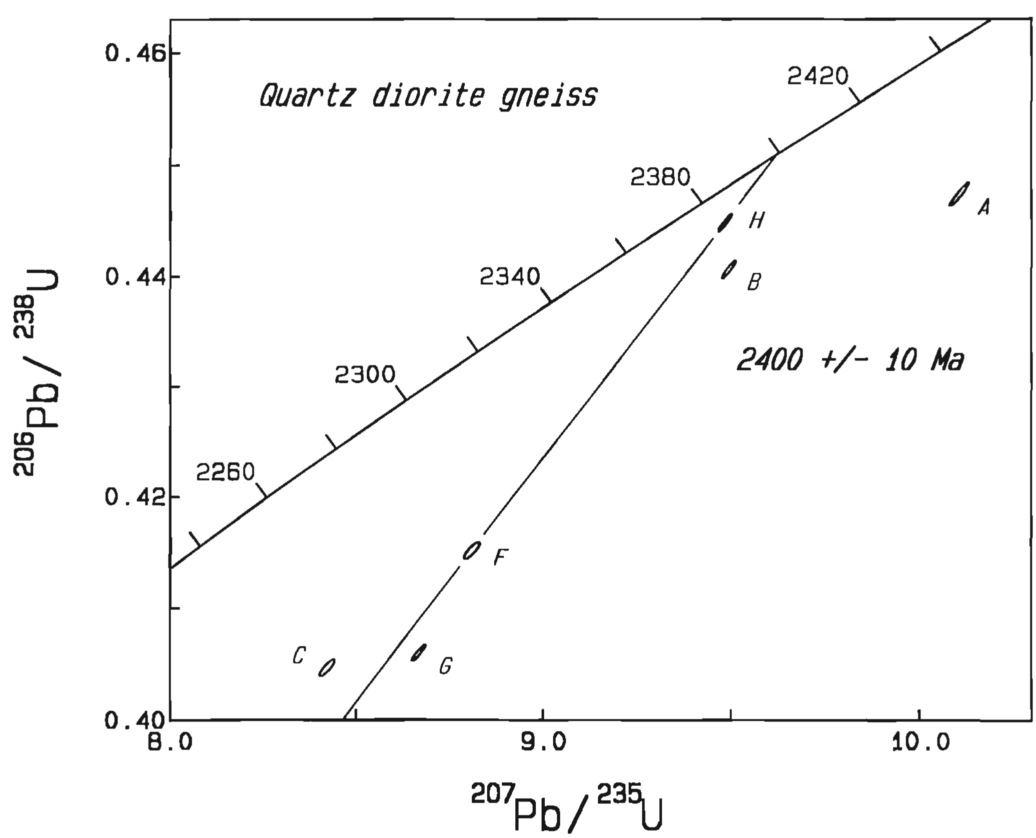

Figure 2. U/Pb isotope ratio plot showing zircon data points with error envelopes for Quartz diorite gneiss (87BK-366X). Letters with data points correspond to those of Table 1.

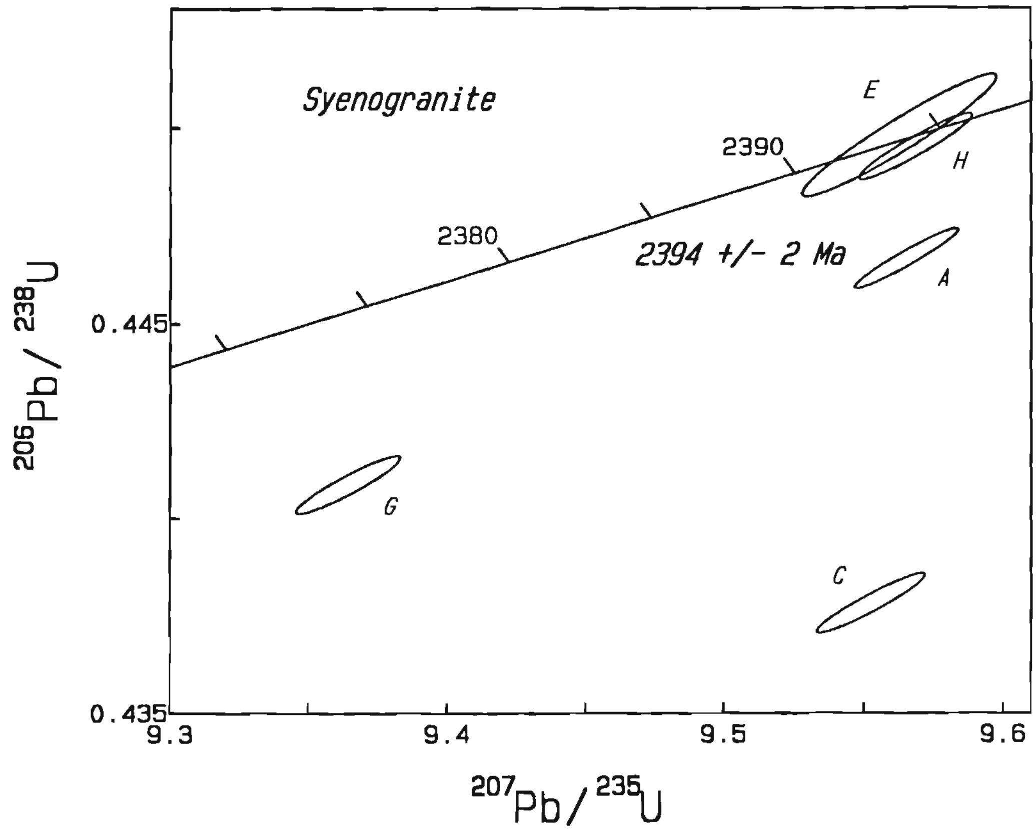

Figure 3. U/Pb isotope ratio plot showing zircon data points with error envelopes for Syenogranite (88BK-172X). Letters with data points correspond to those of Table 1.

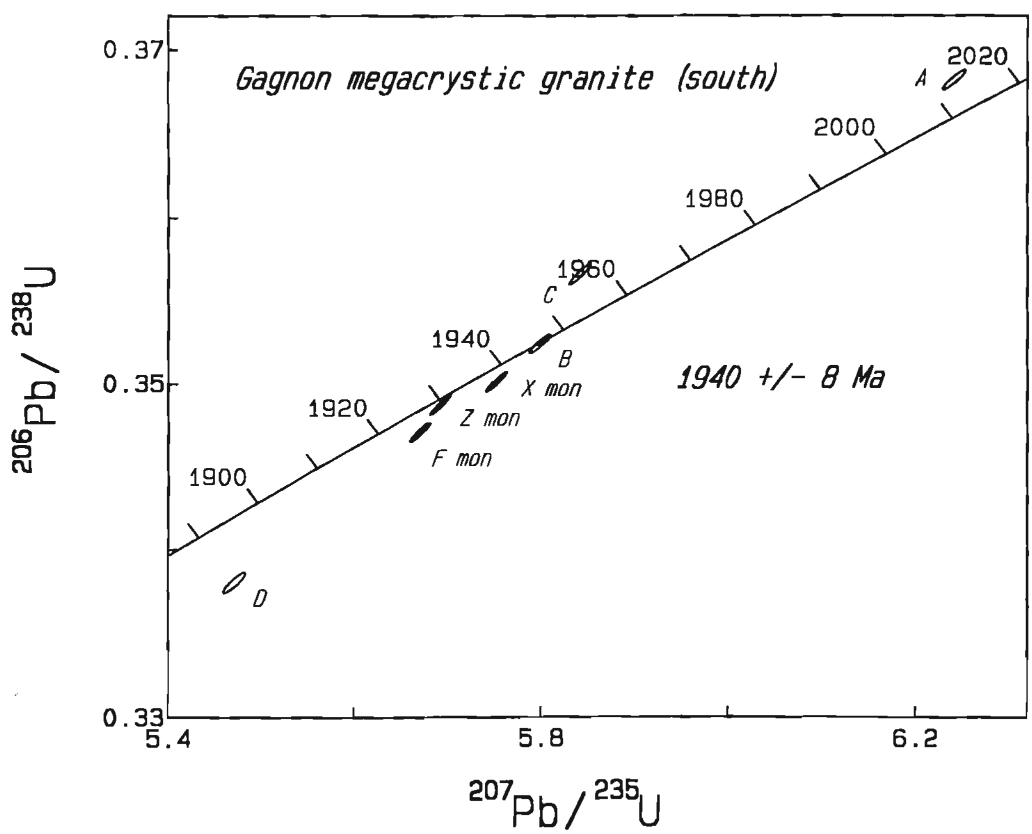

Figure 4. U/Pb isotope ratio plot showing zircon and monazite (Mon) data points with error envelopes for megacrystic northern granite (south, 89BK-10X). Letters with data points correspond to those of Table 1.

fine- to medium-grained biotite-chlorite granite; and 2) biotite ± hornblende megacrystic monzogranite. The megacrystic phase appears to be predominant in the southern part of the body, and the equigranular one in the northern part, but because the two occur in ill defined patches together, in places in the same outcrop, they have heretofore been regarded as consanguinous and essentially coeval. Megacrystic granite locally cuts the adjacent gneisses to the east, but the structural relations of the equigranular granite are uncertain. Large masses of hornblende chlorite diorite and chloritic gneiss with irregular gradational boundaries exist within the pluton particularly along its eastern margin. Small bodies of albitic tourmaline-bearing granitic rocks, irregular pegmatite and late felsic dykes are present locally near the contacts of the major pluton.

The northern granites are separated from the batholiths of the TMZ by a 5 km wide N-S zone of mylonite. A sample of megacrystic granite from the north shore of Gagnon Lake (Fig. 1) has yielded a maximum 1.94 Ga age for dextral mylonitization (unpublished U-Pb monazite data).

Megacrystic granite (North) (87BK-276X; locality 7, Fig. 1)

This sample is typical of the megacrystic monzogranite phase of Gagnon granite, coming from the south shore of Gagnon Lake about 5 km west of its eastern contact. It is medium grained, unfoliated, and white with common subhedral megacrysts of microcline perthite about 1 cm in length. Quartz is interstitial with commmon myrmekite. Oligoclase forms discrete anhedral grains and common subhedral inclusions in microcline megacrysts along with lesser biotite and quartz. Brown biotite and blue-green hornblende are minor minerals and traces of chlorite, epidote, rounded prismatic finely zoned zircon, apatite, opaque and sphene are present.

Zircons are prismatic with L:B of 3:1. Crystals are of poor quality reflecting high U contents and contain ubiquitous cores. In fraction B, all grains were broken in the middle in order to abrade away the cores and this procedure has resulted in a younger ^{207}Pb-^{206}Pb model age of 1935 Ma. Two slightly discordant sphene fractions (Table 1) yield Pb-Pb model ages of 1939 ± 25 Ma. The closure temperature of sphene has been estimated at $\approx 600°C$ (Tucker et al., 1987) and it is possible that the U-Pb sphene system has been reset. Given the general agreement of ages for zircon (fraction B) and sphene ages, the age for this phase of megacrystic granite is likely to be in the 1960 - 1920 Ma interval.

Megacrystic granite (south) (89BK-10X; locality 8, Fig. 1)

This sample represents the megacrystic monzogranite phase of Gagnon granite, and was taken from the central part of the map unit about 5 km northeast of the constriction at its southern end. It is medium grained, slightly foliated, biotite granite with subhedral-euhedral potash feldspar megacrysts up to 2 cm. Pegmatite dykes and patches are common with sharply defined borders, and local patches of equigranular fine grained granite are present. Major minerals are quartz, microcline perthite, and albite. Minor minerals are brown biotite, muscovite, chlorite and epidote, with traces of zircon, monazite, anatase and fluorite. The zircon is present as strongly zoned prisms with irregular cores and scattered larger metamict prisms.

Zircons are prismatic and euhedral. Uranium concentrations are high, ranging from 1140 ppm to 2810 ppm. Fraction A with a ^{207}Pb-^{206}Pb model age of 2000 Ma clearly contains an inherited component (Fig. 4). Both fractions A and C plot above the concordia. This effect may be due to high U content and migration of radiogenic Pb within the grains. It is possible, therefore, that following abrasion not all of the radiogenic Pb is supported by U. The ^{207}Pb-^{206}Pb model age of 1946.6 ± 1.2 Ma for concordant fraction B is considered a maximum age. Monazites, which are concordant to slightly discordant, yield ^{207}Pb-^{206}Pb model ages ranging from 1944 Ma to 1932 Ma. It is not possible to determine whether this scatter is the result of inheritance or later resetting. The closure temperature of monazite has been estimated at 725°C by Parrish (1990). On the basis of both the zircon and monazite data, the age of this phase of the megacrystic granite is assigned at 1940 ± 8 Ma.

Equigranular granite (89BK-12X; locality 10, Fig.1)

This sample represents the equigranular monzogranite phase of Gagnon granite in the central part of the pluton about 8 km south of Gagnon Lake. The rock is unfoliated, medium grained, equigranular, and homogeneous but is cut locally by irregular dykes and patches of pegmatite. Major minerals are quartz, perthite, and albite. Minor chlorite, epidote and a few flakes of altered brown biotite are present. Zircon, commonly cracked with hematitic haloes, is present along with scattered grains of anhedral opaque.

Zircons are euhedral simple prisms. Uranium concentrations are high, ranging from 2140 ppm to 5470 ppm. Three data points (A,C, and E) are linearly aligned (Fig. 5), although significant geological scatter is indicated by a mean square of weighted deviates of 19. A regression line intercepts the concordia at 1882 +7/-6 Ma and 552 ± 122 Ma (York, 1969). The upper intercept age is interpreted as the age of igneous crystallization. Zircon fraction B with a ^{207}Pb-^{206}Pb model age of 1957 Ma is likely to contain an inherited component.

DISCUSSION

Early plutonic rocks

Zircon geochronology from granitic rocks within the mixed gneiss corridor east of TMZ reported both here and previously (Bostock and Loveridge, 1988; Bostock et al., 1987) has shown a wide spread of interpreted ages from 2.44 Ga to 2.27 Ga. It might appear at first glance that they represent a period of orogeny lasting nearly 200 Ma. There is, however, an apparent spatial distribution as the three youngest granitic phases (2.34 - 2.26 Ga) are associated with the marginal zone of ductile shearing. Of these younger ages, two are based on single concordant data points from complex zircon populations with either large amounts of inheritance (locality 5, Fig. 1) or later Proterozoic disturbance (locality 4, Fig. 1). The igneous age for the Yatsore Lake augen gneiss (locality 6, Fig. 1) at 2270 ± 4 Ma is, however, reliable.

East of the marginal zone of ductile shearing, within the region of paragneiss free mixed gneiss (Fig. 1), three distinct plutonic rock types yield ages in the range 2.44-2.39 Ga. Of special significance is the 2394 ± 2 Ma age for unfoliated syenogranite (88BK-172X), which is likely to have been emplaced late within the tectonic development of the area. The available radiometric evidence, therefore, suggests that granitic phases of the more extensive paragneiss-free mixed gneiss of this area within the western Rae Province were emplaced in a more restricted time interval 2440-2390 Ma ago.

The northern granites

Zircon data from the megacrystic granite suffer from the same discordant dispersion as do the mixed gneisses. At two separate localities megacrystic granite has yielded identical ages of 1.94 Ga and emplacement appears to have been coeval with that of the Konth Granite within TMZ to the west (1940-1935 Ma; Bostock and Loveridge, 1988).

The Konth granite and the megacrystic phase of the northern granites differ in that the former contains abundant monazite and metasedimentary inclusions, and is typically foliated whereas the latter is characterized by the presence of sphene, local amphibole, common mafic inclusions, rare monazite and the lack of foliation. These contrasts may have arisen in part as a result of tectonic isolation of the northern granites in a fault block derived from the northeast, where the composition of assimilated material may have been different. This block was subsequently inserted at its present location along major faults, dextral to the west and sinistral along Taltson valley to the southeast. Thus the megacrystic northern granite may have been emplaced in a somewhat different environment in a part of TMZ far removed from the Konth Granite, now exposed immediately to the west of it.

The equigranular phase of the northern granites, hitherto considered to be a coeval consanguinous phase of megacrystic northern granite, gives a much younger zircon age at 1882 +7/-6 Ma and is clearly not related to the megacrystic phase. This age is very similar to that reported by S. Hanmer et al. (pers. comm.) for granite veins preserved within an included wedge of Archean mylonite within Great Slave Lake Shear Zone to the north. Unfortunately no exposures of the equigranular granite have yet been found along the west boundary of the map unit where they might provide critical data on the time of late shear both in TMZ and along Great Slave Lake Shear Zone.

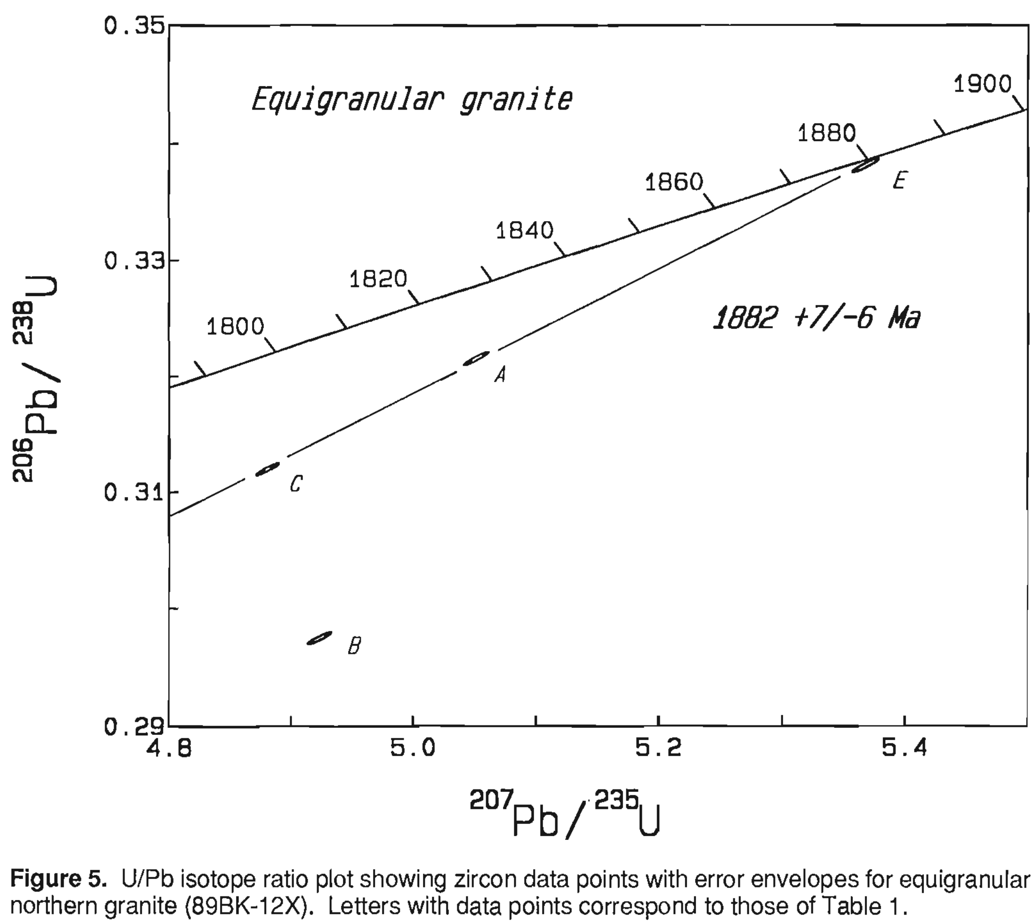

Figure 5. U/Pb isotope ratio plot showing zircon data points with error envelopes for equigranular northern granite (89BK-12X). Letters with data points correspond to those of Table 1.

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U-Pb zircon ages from an Archean orthogneiss and a Proterozoic metasedimentary gneiss of the Thelon Tectonic Zone, District of Mackenzie, Northwest Territories

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Abstract

A 2598 ± 6 Ma U-Pb zircon age for a granitoid gneiss unit helps to delimit the eastern extent of Archean rocks of the Slave craton within the Proterozoic Thelon Tectonic Zone in the Healey Lake and Artillery Lake map areas. This age confirms previous field interpretations that a straightened zone of Yellowknife Supergroup rocks is separated from Proterozoic granitoid gneiss of the Thelon Tectonic Zone by a 1 to 3 km wide zone of Archean granitoid gneiss. Single grain U-Pb zircon ages also have been determined on apparent core material as well as on metamorphic tips and multi-faceted metamorphic grains from Mary Frances migmatitic metasedimentary gneiss of the central and eastern Thelon Tectonic Zone. Metamorphic zircon growth is indicated at 2000 ± 10 Ma marking the beginning of thermal activity in the 2.0-1.9 Ga Thelon Tectonic Zone, although the possibility of a slight age shift resulting from minor inheritance cannot be ruled out. Two cores, considered to be detrital, yield model ^{207}Pb-^{206}Pb ages of about 2.2 Ga which provide a maximum age for Mary Frances sedimentation. Ages for some 'cores' overlap the metamorphic data and may have been misidentified as probable detrital cores.

Résumé

La datation à 2598 ± 6 Ma par la méthode U-Pb sur zircon d'une unité de gneiss granitoïde permet de délimiter l'extension orientale des roches archéennes du craton des Esclaves dans la zone tectonique de Thelon du Protérozoïque dans les zones cartographiées des lacs Healey et Artillery. Cette datation confirme des interprétations faites sur le terrain selon lesquelles une zone rectiligne de roches du supergroupe de Yellowknife est séparée du gneiss granitoïde protérozoïque de la zone tectonique de Thelon par une zone de gneiss granitoïde archéenne de 1 à 3 km de large. On a également daté par U-Pb sur grains individuels de zircon un matériau de noyau apparent provenant du gneiss métasédimentaire migmatitique de Mary Frances dans les parties centrale et orientale de la zone tectonique de Thelon. La croissance du zircon métamorphique est indiquée à 2000 ± 10 Ma marquant le début de l'activité thermique dans la zone tectonique de Thelon, même s'il est impossible d'établir un léger décalage d'âge causé par un faible héritage. Deux noyaux, considérés détritiques, ont donné des âges modèles $^{207}Pb/^{206}Pb$ d'environ 2,2 Ga permettant de déterminer un âge maximal pour la sédimentation de Mary Frances. Les âges de certains «noyaux» chevauchent les données métamorphiques et on pu être faussement identifiés comme des noyaux détritiques probables.

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INTRODUCTION

The purpose of this U-Pb zircon geochronological study is twofold. Firstly, it is intended to more accurately delimit the eastern limit of rocks of the Archean Slave craton which have been deformed within the north-northeasterly trending Proterozoic Thelon Tectonic Zone (Fig. 1). Due to extreme deformation, differentiating between Archean gneiss and foliated granitoid derived from the Slave, and granitoid gneiss of Proterozoic age to the east is difficult on the basis of field criteria alone. Geochronological data are presented from a granitoid gneiss, believed to be part of the Slave craton, which was collected 1 km west of the proposed eastern limit of deformed Slave rocks (James, 1985; Henderson, 1989; Henderson and Thompson, 1982).

The second part of the study concerns the highly metamorphosed and deformed supracrustal gneiss that occurs within the central and eastern Thelon Tectonic Zone, a terrane dominated by Proterozoic granitoid gneiss. Little is known of the age of the protolith of the supracrustal gneiss. From a sample of migmatitic metasedimentary rock containing complex zircons, U-Pb ages on a suite of cores and their overgrowths are presented in an attempt to gain some insight into the age of the source terrain from which the original sediments were derived, as well as the timing of their metamorphism.

ANALYTICAL METHODS

The concordance of all zircon U-Pb isotopic analyses has been maximized by the careful selection and strong abrasion of the individual mineral grains analyzed (Krogh, 1982). Analytical techniques for U-Pb zircon and monazite analysis are summarized in Parrish et al. (1987) which also described the modified form of the regression analysis technique of York (1969) that was used. Analytical data are presented in Table 1 with the individual data points also displayed in isotope ratio plots in Figures 2 and 3. In Table 1 the data for individual fractions are arranged in order of decreasing ^{207}Pb-^{206}Pb model ages for each of the two rocks analyzed. Treatment of analytical errors follows Roddick (1987). Uncertainties for ages are quoted at the 2σ level.

EASTERN LIMIT OF SLAVE CRATON ROCKS

Geological setting

In the Healey Lake and Artillery Lake area of the southeastern Slave Province the pattern of bulbous plutons and irregular areas of Yellowknife supracrustal rocks, characteristic of the Slave Province, gives way to a pronounced 020° trending, steeply dipping 'straight zone' as the Archean Slave rocks become involved in the Proterozoic (2.0-1.9 Ga) deformation of the Thelon Tectonic Zone (Fig. 1). This zone is dominated by mylonitized migmatitic metasedimentary rocks and amphibolitic rocks that to the north can be traced out of the straight zone and into an extensive terrane of Yellowknife Supergroup metasedimentary rocks, large parts of which are metamorphosed only to greenschist grade (Frith, 1982). U-Pb ages of zircon from Yellowknife metavolcanic rocks within the Healey Lake and Artillery Lake areas have been dated at two localities at 2692 ± 2 Ma (van Breemen et al., 1987a) and 2671+2/-4 Ma (van Breemen and Henderson, 1988). There is little doubt that the supracrustal rocks within the straight zone, although deformed in the Proterozoic Thelon Tectonic Zone, are Archean and were originally part of the Slave craton.

The main part of the Thelon Tectonic Zone lies to the east and consists of a similarly oriented assemblage of granitoid and supracrustal gneiss, largely at granulite grade and containing extensive mylonite zones. Several units in this terrane, dated with U-Pb methods, range in age between 1.9 and 2.0 Ga (James et al., 1988; van Breemen and Henderson, 1988; van Breemen et al., 1987b). The main syn- to pre-metamorphic igneous suites appear to have been emplaced in the interval 2.00 - 1.96 Ga (Frith and van Breemen, 1990). The oldest ages from within the Thelon Tectonic Zone have come from a tonalitic granulite gneiss at Moraine Lake (Fig. 1) which has been investigated by both single and multi-grain conventional methods and ion probe U-Pb analysis (van Breemen et al., 1987b; Roddick and van Breemen, 1989). This gneiss contains zircons with 2.3 Ga cores mantled by euhedral igneous rims at 2.15 Ga and minor metamorphic overgrowths at 1.95 Ga. The Sifton granitoid gneiss, a Proterozoic unit that contains 1.97 Ga zircons (van Breemen and Henderson, 1988) occurs less than 3 km from the easternmost Archean Yellowknife supracrustal rocks.

Between the easternmost Yellowknife and the Proterozoic gneiss is a zone of granitoid gneiss. In the Moraine Lake area, the lower grade gneiss was recognized by James (1985, 1989) as being distinct from the granulite grade gneiss to the east and was interpreted as Archean. To the south, similar gneiss in the same structural position was also interpreted as Archean in age (Henderson, 1989). In order to test this interpretation, a sample of this gneiss was collected at Smart Lake between the easternmost Yellowknife supracrustal rocks and the westernmost Sifton gneiss (location A Fig. 1).

Geochronology

The rock collected (MLB-88-18) was a strongly deformed, pink and grey, megascopically sparsely porphyroclastic, granitic gneiss with fine, straight layering. In thin section, the rock has a strongly developed mylonitic foliation defined by narrow quartz ribbons and wispy, discontinuous aggregates of fine greenish brown biotite. Anhedral porphyroclasts of both plagioclase and microcline that grade from 2 mm to the 0.1 mm grain size of the quartzofeldspathic matrix. Epidote associated with the biotite is abundant. Also present are apatite, pyrite, and elongate, finely-polycrystalline aggregates of sphene. Subhedral to anhedral zircon, up to 0.2 mm in size, can contain cores with well developed, generally fractured overgrowths.

The zircon separate consisted of about 70% pyrite which was floated off in heated 3N HNO_3. In general the zircons analyzed were brown, elongate, subhedral prisms between

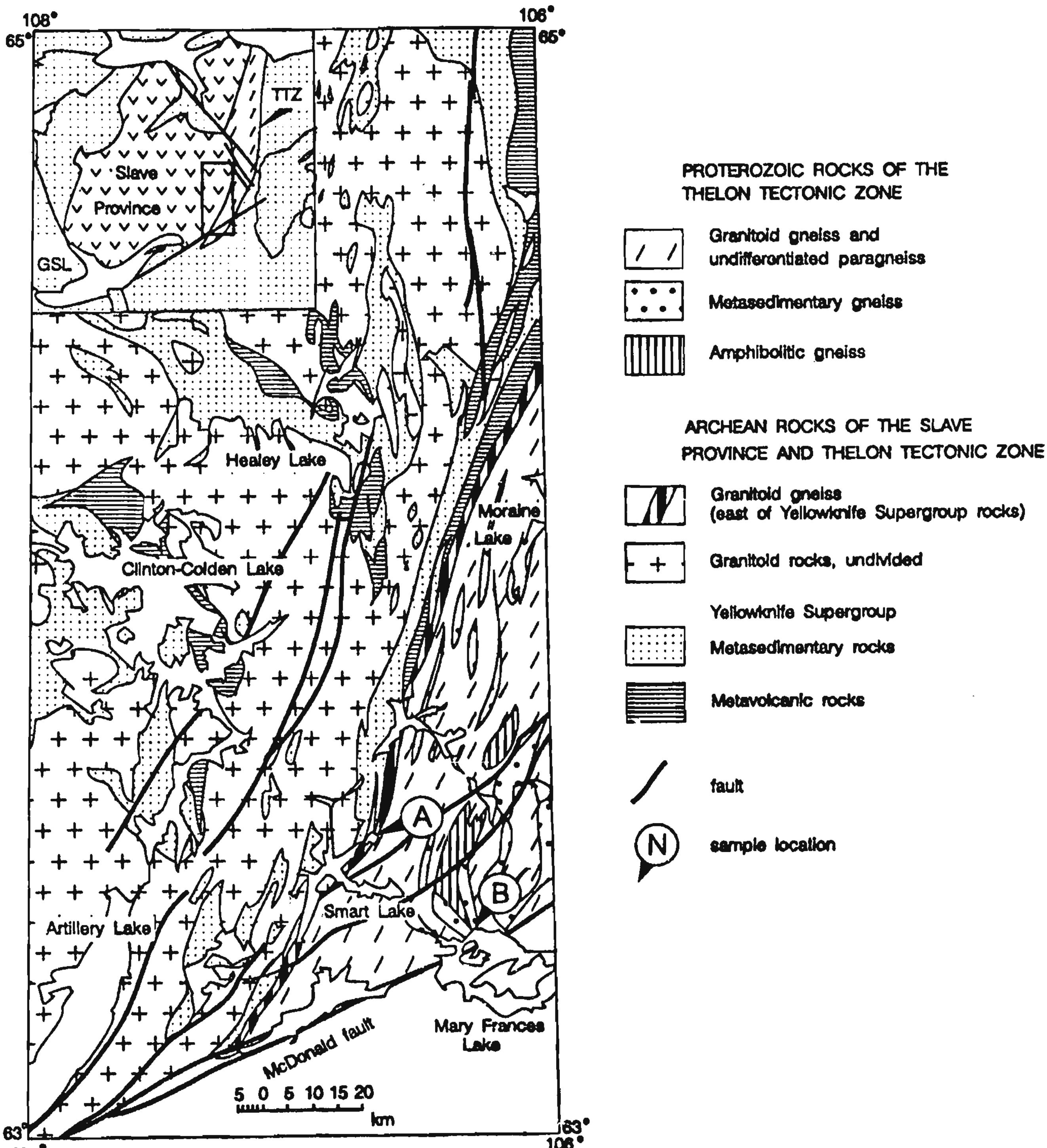

Figure 1. Geology of the Healey Lake - Artillery Lake area at the boundary between the Slave Province and the Thelon Tectonic Zone (inset). Location of the two samples analyzed are indicated. Sample A - granitoid gneiss east of last Archean Yellowknife Supergroup supracrustal rocks at Smart Lake; Sample B - Mary Fances metasedimentary gneiss north of Mary Frances Lake. TTZ - Thelon Tectonic Zone; GSL - Great Slave Lake.

200 and 500 μm long containing minor fractures, no apparent zoning and with the simple morphology suggestive of an igneous origin. Fraction D consisted of clear, colourless more elongate zircons less than 200 μm in length. The crystal faces are etched, possibly due to leaching during subsequent metamorphism. Cores were apparent in some zircons and these were avoided in selecting grains for analysis.

The analytical data are shown in Table 1 and portrayed in Figure 2. All of the 4 fractions analyzed are discordant but three of the fractions fit a regression line (MSWD = 0.1) with an upper intercept of 2598 ± 6 Ma and a lower intercept of about 1.35 Ga. Fraction D is displaced to the right of the regression line, possibly as a result of the presence of some older core material. The lower intercept age of 1.35 Ga has no known significance but since the rock was involved in Thelon Tectonic Zone deformation and metamorphism at 1.9 - 2.0 Ga (van Breemen et al., 1987b,c) and again during the indentation of the Slave Province and part of the western Thelon Tectonic Zone into the Churchill Province between about 1.78 Ga and 1.74 Ga (Henderson et al., 1990), the zircon has probably had a complex lead loss history.

Discussion

There is little doubt that the rock was originally crystallized during the Archean. The upper intercept age at 2598 ± 6 Ma is close to that of the nearby lithologically similar but less deformed Smart gneiss dated at 2602 ± 2 Ma (van Breemen and Henderson, 1988) located immediately west of the Yellowknife metasedimentary belt (van Breemen et al, 1987a). The age of this gneiss confirms the existence of Archean granitoid rocks in this structural position as was originally interpreted on the basis of the contrast in metamorphic grade with the Proterozoic granulites of the Thelon Tectonic Zone to the north (James, 1985). As the first appearance of orthopyroxene east of the Yellowknife supracrustal belt is commonly separated by a 2 to 3 km wide zone of granitoid gneiss (Henderson and Thompson, 1982; Henderson, 1989), this age strengthens the probability that the gneiss in this position was originally Archean.

MARY FRANCES SUPRACRUSTAL GNEISS

Geological setting

Supracrustal gneiss consisting of amphibolite of possible mafic volcanic origin and migmatitic metasedimentary gneiss are an important component of that part of the Thelon Tectonic Zone that is dominated by Proterozoic granitoid gneiss and other gneiss of indeterminate origin (Fig. 1). They have been named the Mary Frances gneiss in the vicinity of Moraine Lake (James, 1989) and occur throughout the region. They are less abundant in the north where they form long narrow units and are at granulite grade. To the south they are more abundant, less deformed and at lower metamorphic grade. They are everywhere in tectonic contact with adjacent gneissic units (Henderson et al., 1987). The migmatitic metasedimentary gneiss consists of biotite-rich paleosome and white quartz-plagioclase leucosome. An introduced pink granitic phase is commonly present. Rarely do the metasedimentary migmatites contain sillimanite but where present, the rocks resemble the Archean Yellowknife metasedimentary migmatites of the Slave Province to the west. More commonly, however, hornblende and clinopyroxene are present in the biotite rich rocks, suggesting a more calcic parent.

Little is known of the age and origin of the metasedimentary gneiss. Based on their similarity to high grade Yellowknife metasediments particularly in the north and their association, like the Yellowknife metasediments, with amphibolites of possible mafic volcanic origin, a suggested correlation has been made with the Yellowknife Supergroup of the Slave Province (Fraser, 1972; Henderson

Table 1. U-Pb zircon data

Fraction,[1] Size	Weight (mg)	U (ppm)	Pb^2 (ppm)	$\frac{^{206}Pb}{^{204}Pb}$ [3]	Pb_c[4] (pg)	$^{208}Pb^2$ (%)	$\frac{^{206}Pb}{^{238}U}$ ±SEM%[5]	$\frac{^{207}Pb}{^{235}U}$ ±SEM%[5]	R	$\frac{^{207}Pb}{^{206}Pb}$ age, error (Ma)[6]
Granitoid gneiss at Smart Lake (MLB-88-18; 63° 32.8′ 106°40.0′)										
B, 400*150,N2,S	0.025	223	115	3900	38	9.2	0.4674 ± (.08)	10.918 ± (.10)	0.95	2551.8 (1.1)
D, ≈200*75,N2	0.026	212	106	6765	21	10.9	0.4457 ± (.08)	10.242 ± (.09)	0.96	2524.3 (1.0)
C, ≈200*100,N2	0.012	443	214	3062	44	7.2	0.4468 ± (.08)	10.200 ± (.10)	0.93	2513.6 (1.2)
A, 550*200,N2,S	0.046	267	132	6814	46	9.9	0.4443 ± (.09)	10.117 ± (.10)	0.96	2509.1 (1.0)
Mary Francis metasedimentary gneiss (MLB-88-17; 63° 23.0′ 106° 17.5′)										
D, core,110,M1,S	0.006	1479	622	7657	25	6.2	0.4029 ± (.09)	7.720 ± (.10)	0.96	2214.2 (1.0)
I, core,250,M1,S	0.021	746	297	7197	49	4.5	0.3890 ± (.09)	7.299 ± (.10)	0.96	2178.0 (1.1)
G, core,300*150,M1,S	0.020	765	287	7332	44	6.3	0.3634 ± (.09)	6.276 ± (.10)	0.96	2032.4 (1.1)
H, core,200,M1,S	0.012	876	321	15720	14	4.5	0.3620 ± (.09)	6.185 ± (.10)	0.96	2013.5 (1.0)
C, tip,M1,S	0.004	403	153	2247	15	7.6	0.3629 ± (.09)	6.185 ± (.11)	0.93	2009.1 (1.4)
E, tip,250*150,M1,S	0.015	348	138	13400	9	11.4	0.3622 ± (.10)	6.160 ± (.11)	0.97	2005.3 (1.0)
K, core,200*150,M1,S	0.011	336	131	4455	18	10.1	0.3639 ± (.09)	6.187 ± (.10)	0.94	2004.8 (1.3)
F, tip,200*75,M1,S	0.006	629	234	3066	27	8.4	0.3535 ± (.10)	5.975 ± (.12)	0.86	1994.3 (2.1)
A, round,120,N0,S	0.004	615	231	4004	13	9.2	0.3523 ± (.09)	5.953 ± (.10)	0.95	1993.6 (1.2)
J, core,250*200,M1,S	0.024	376	142	3320	56	10.8	0.3488 ± (.09)	5.794 ± (.10)	0.93	1963.6 (1.4)
B, round,130,N0,S	0.005	467	154	4892	10	8.0	0.3149 ± (.09)	5.217 ± (.10)	0.95	1958.4 (1.1)

Notes: [1]fractions C and D are tip and core from the same zircon grain, sizes in microns for zircons before abrasion, i.e. ≈100*50; average length and breadth aspects); M and N refer to magnetic and non-magnetic at side slope indicated in degrees; S indicates analysis of a single crystal, [2]radiogenic Pb; [3]measured ratio, corrected for spike and fractionation; [4]total common Pb in analysis corrected for fractionation and spike; [5]corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; R correlation of errors in isotope ratios; [6]corrected for blank and common Pb, errors are two sigma in Ma.

and Macfie, 1985). However, given the generally more calcic composition of the Mary Frances metasedimentary gneiss, such a correlation is considered less likely (Henderson, 1989; James, 1989). Preliminary Sm-Nd analyses of samples from this unit indicate 'depleted mantle' model ages between 2.7 and 2.9 Ga, implying that these rocks may contain a large Archean component (Ernst Hegner, pers. comm., 1989).

Geochronology

A sample of the Mary Frances gneiss (MLB-88-17) was collected north of Mary Frances Lake (Fig. 1) from sub granulite grade terrane with the hope that an analysis of the core and rim material from a suite of zircons from the rock would provide a better understanding of the provenance and history of the metasedimentary rock. Rim material would potentially give an estimate of the age of migmatization while the cores, if detrital remnants, would provide age estimates of sediment source terrains or syn-sedimentary volcanism.

Many of the metasedimentary migmatites in the area of the sample contain a high proportion of leucosome. To improve the chances of recovering a detrital zircon component, the most quartzofeldspathic, least migmatitic material was sampled. The rock is a dark greenish grey, medium-fine grained moderately foliated gneiss with layering defined by variations in the proportion of mafic to quartzofeldspathic minerals on a millimetre to several centimetre scale as well as by fine layers to lenses of yellowish white leucosome that make up less than 10% of the rock. In thin section the rock consists of the mineral assemblage plagioclase-quartz-biotite-clinopyroxene-hornblende-garnet. Hornblende locally replaces the clinopyroxene. Other minerals present include apatite, zircon, an opaque phase and a yellow mineraloid. The leucosome is dominated by quartz with lesser amounts of plagioclase and a small amount of biotite. The rock has a granoblastic texture with no trace of any detrital textures. Zircon is moderately abundant as stubby anhedral to rounded grains less than 0.1 mm in size, in some cases with apparent cores.

The analytical data for 11 single zircons or parts thereof are summarized in Table 1 and Figure 3. Fractions A and B are clear, colourless, equidimensional, multifaceted zircons with no visible inclusions, cores or fractures and are presumed to be metamorphic in origin (van Breemen et al., 1987b). The original 150 μm diameter grains were reduced in diameter by about 1/7 by abrasion. Fractions C (tip) and D (core) were derived from a single zircon with a prominent core which separated cleanly. Fraction C is a clear, colourless

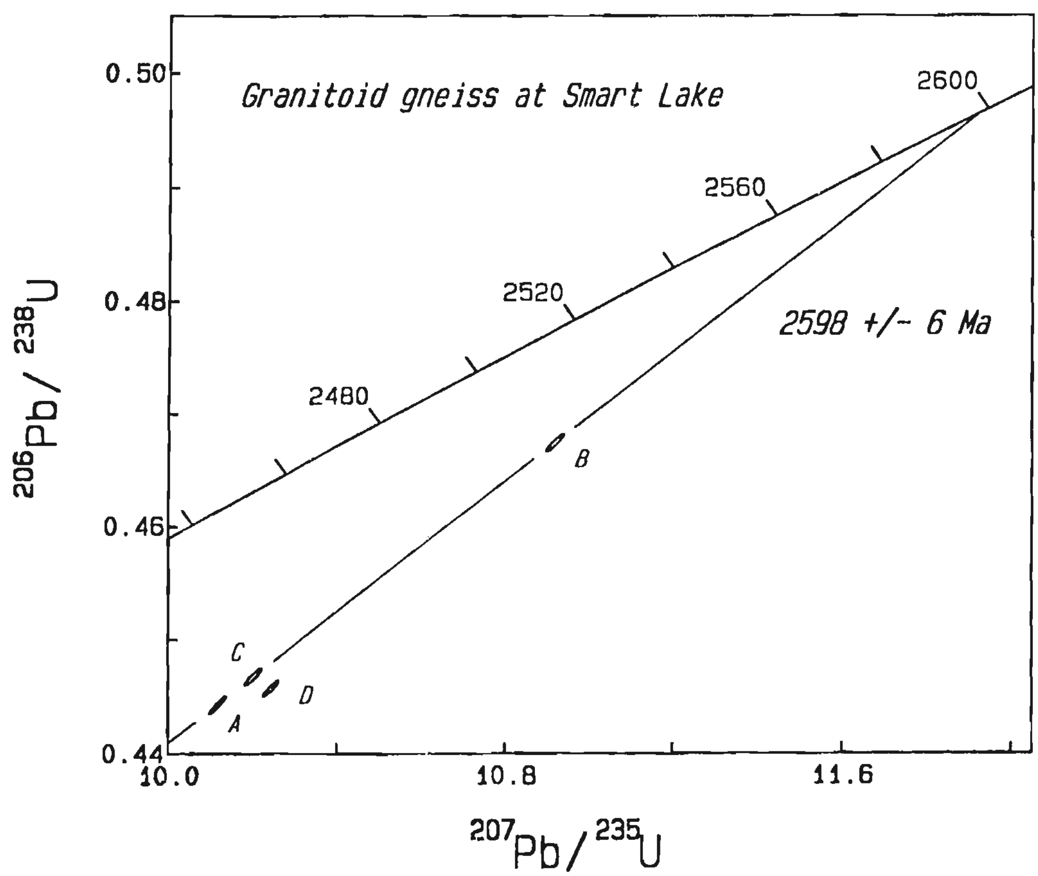

Figure 2. Concordia diagram for the granitoid gneiss at Smart Lake. Although the data are moderately discordant, the protolith of the deformed granitoid is Archean in age. Fraction D was excluded from the regression. Error ellipses are 2σ.

tip fragment with no apparent inclusions or zoning that was abraded for 2 hours. The equidimensional, anhedral with finely pitted surfaces, 140 m core material of Fraction D was reduced by abrasion by about 1/6 of its original diameter. Fractions E and F are clear, colourless broken tips that were abraded for 3 hours. Fractions G through K are 5 coarse cores in the range of 200 to 250 µm, that vary in colour from dark brown to clear and colourless, and in shape from equidimensional to slightly elongate with a simple crystal form. In the course of 9 hours of abrasion, their diameters were decreased by about one third.

Polished and etched zircon grain mounts were made of grains judged to be similar, but of lesser quality to those selected for analysis. All the grains showed the typical simple crystal morphology with few large crystal faces characteristic of zircons crystallized from a melt. Although most of the grains appeared to contain cores when viewed in transmitted light through alcohol, only one grain had a core in which the zoning was discordant to the mantle. The zircons ranged from those with all zones composed of equal uranium content to those with most zones consisting of higher uranium zircon, the relative uranium content based on the degree of etch. A few consisted of mainly higher uranium zones with a few lower uranium zones that in transmitted light might easily be mistaken for zircons with cores.

It is evident from Table 1 and Figures 3 and 4 that what was interpreted as zircon core and rim material and zircon of metamorphic origin do not separate into distinct age populations. Core fractions D and I are distinctly older with ^{207}Pb-^{206}Pb ages of 2214 Ma and 2178 Ma. The corresponding tip material (Fraction C, associated with core Fraction D) together with the other two tip fractions (Fractions E and F) group together with ^{207}Pb-^{206}Pb ages between 1994 and 2010 Ma. One of the clear, round, multifaceted zircons interpreted as metamorphic in origin (Fraction A) has a similar age of 1994 although the other, otherwise identical, metamorphic zircon (B) is much more discordant and younger with a ^{207}Pb-^{206}Pb age of only 1958 Ma. Two of the core fractions (Fractions K and H) plot within or slightly older than this grouping while core fraction G is somewhat older at 2032 Ma. The discordant core fraction J is significantly younger than the main grouping with a ^{207}Pb-^{206}Pb age of 1964 Ma, more similar in age to the discordant metamorphic fraction B.

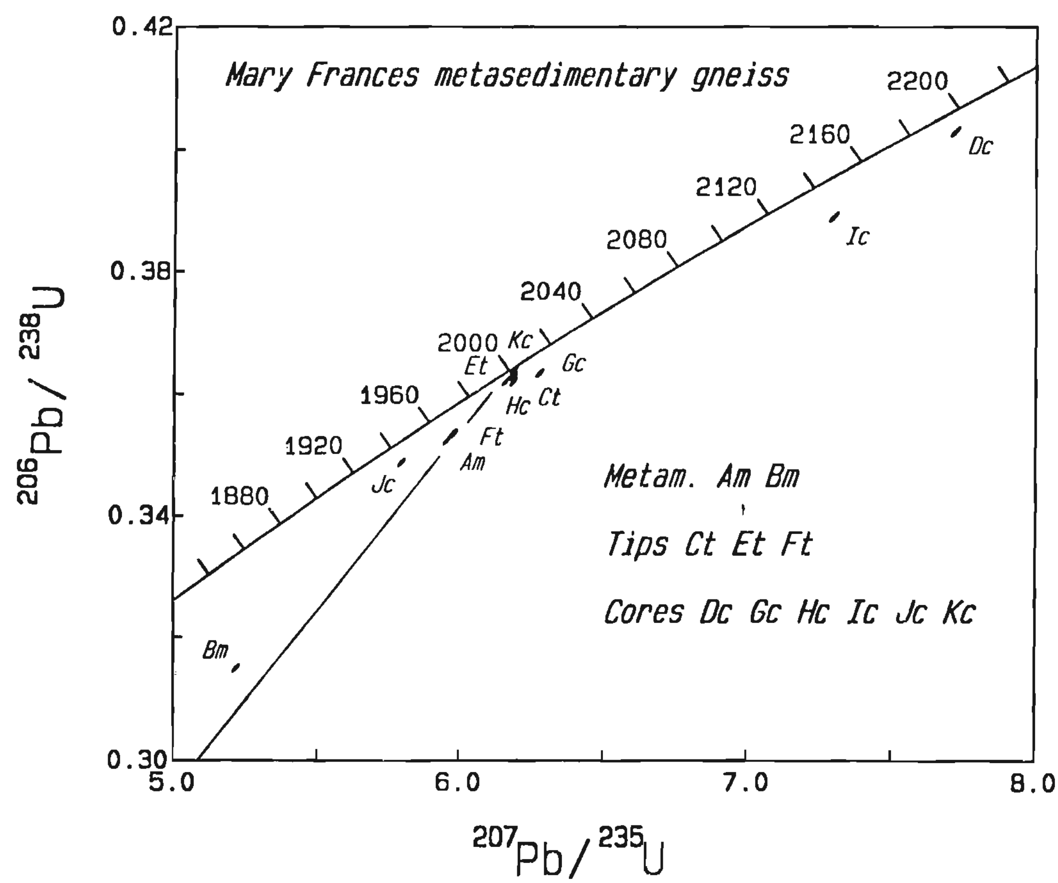

Figure 3. Concordia diagram for the Mary Frances metasedimentary gneiss. Although most of the fractions interpreted as core material are not distinguishable in age from those fractions interpreted as metamorphic in origin, two fractions are distinctly older. The data do not suggest any of the core material is Archean in age. Zircon fractions are classified as follows: Metam = metamorphic zircon; Tips = zircon tips removed from core bearing grains; Cores = zircon cores isolated by removal of rim and tip material by breakage or abrasion. Error ellipses are 2%.

The data for fractions C,E,F, and A suggest that a significant amount of metamorphic zircon grew at 2000 ± 10 Ma, although the model age for metamorphic grain B at about 1958 Ma may indicate some later growth of zircon. 'Core' fractions G, H, J and K are clearly younger than the oldest two cores and may be metamorphic zircon formed as part of the metamorphic process that was misidentified as detrital core material. The ^{207}Pb-^{206}Pb ages of the two oldest cores at 2214.2 ± 1.0 Ma and 2178.1 ± 1.1 Ma are significantly older than most ages of rocks previously dated from the Thelon Tectonic Zone which range between 2.0 and 1.9 Ga (Fig. 4). These ages are intermediate to the 2.3 Ga and 2.15 Ga core and mantle ages of the granulite grade tonalitic gneiss approximately 70 km to the north northeast at Moraine Lake (Roddick and van Breemen, 1989). It is possible that they have been partly reset during the 2.0 - 1.9 Ga tectonothermal events. As these cores are considered to be detrital, a line

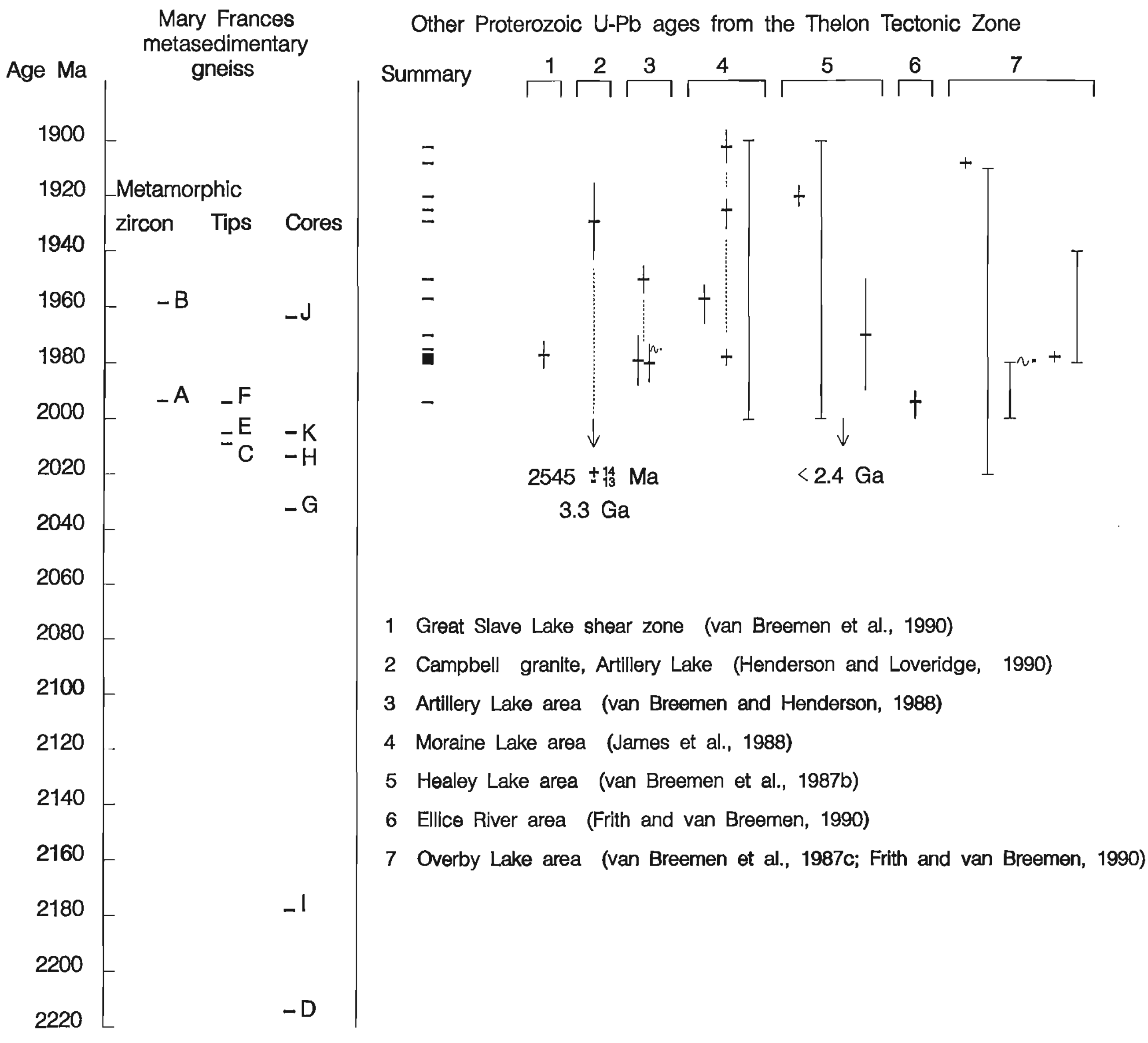

Figure 4. Summary of Mary Frances metasedimentary gneiss ^{207}Pb-^{206}Pb ages shown in context with all other Proterozoic U-Pb ages from the Thelon Tectonic Zone or equivalents. Ages to date for the Thelon Tectonic Zone as a whole range between 1.90 Ga and 2.02 Ga with a concentration at about 1.98 Ga which may represent a major metamorphic event. Mary Frances metasedimentary gneiss zircon fractions interpreted as metamorphic in origin (metamorphic zircon and tips) range between 1.96 and 2.01 Ga in age; with one exception, somewhat older than the concentration at 1.98 Ga for the Thelon Tectonic Zone as a whole. The ages of 'core' fractions range between 1.96 Ga and 2.21 Ga but only two are significantly older than the metamorphic age range within the same rock. For the other Thelon Tectonic Zone ages, each entry represents a sample analyzed. In some samples, several ages, each with a vertical error bar, were separated (ie, cores and rims). These are joined by a dotted line. For some others, only a range is indicated.

through 2.0 Ga and point D (Fig. 3) intercepts the concordia at 2.4 Ga which would represent the maximum age of sedimentation.

Discussion

The more concordant metamorphic zircon fractions include C, E and F (tips) and A (multi-faceted zircon) (Fig. 3) whose ages fall at and slightly beyond the older end of the age spectrum of metamorphic and plutonic ages for the Thelon Tectonic Zone (Fig. 4). The grouping of ages at about 2.00 Ga suggests metamorphism at that time, somewhat older than the pronounced grouping of ages at about 1980 Ma for the rest of the Thelon Tectonic Zone (Fig. 4). However, the presence of an unrecognized inherited component in this material cannot be ruled out. Younger model ages for the more discordant fractions (J and B) suggest that metamorphic zircon growth, possibly localized along fractures or fluid conduits, continued until about 1.96 Ga.

The presence of 2.2 Ga zircon cores in the metasedimentary rock implies a maximum age of sedimentation of 2.2 Ga if indeed the zircon cores are detrital. Since the sample was selected for its dominantly quartzofeldspathic composition in order to enhance the probability of detrital zircon being present in the original sediment and the recognized ability of zircon to survive melting conditions given the common occurence of inherited zircon components in granitoid rocks, it is considered likely that the relict zircon was detrital. Archean ages are not expressed in the zircon data despite the indication from Nd model ages of a strong Archean component in these metasediments. Nd model ages, however, are more a commentary on the average crust/mantle separation age of the source of the sediments than anything else (McCulloch and Wasserburg, 1979). The only documented occurrence of Archean zircon in the Proterozoic rocks of the Thelon Tectonic Zone are as xenocrysts in the 1.93 Ga Campbell granite located south of the McDonald fault (Henderson and Loveridge, 1990).

No exposed terrane with an age range encompassing pre-2 Ga rocks are known in the region with the possible exception of the tonalitic gneiss at Moraine Lake. Within the Slave Province, the only similar age is the 2175 ± 7 Ma age for the alkaline phase of the Blachford Lake Intrusive Suite (Davidson, 1978; Bowring et al., 1984). South of the McDonald fault, in the eastern gneiss complex of the Taltson Magmatic Zone, ages in the range 2440 - 2270 Ma have been reported (Bostock et al., 1991; van Breemen et al., 1992). South of Great Slave Lake and in north-central Alberta, similar ages have been reported from the Buffalo Head Terrane, an early Proterozoic terrane covered by Phanerozoic sediments that consists of 2.0-2.32 Ga metaplutonic and lesser amounts of felsic metavolcanic rocks (Ross et al., 1991).

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U-Pb zircon ages from the Chantrey Inlet area, northern District of Keewatin, Northwest Territories

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Abstract

Three samples of meta-igneous basement rock from the Chantrey Inlet region of the Churchill Province yield late Archean zircon ages of 2.6 Ga, slightly younger than Sm-Nd model ages of 2.7-2.8 Ga. The protolith of a hornblende-biotite augen gneiss, representative of a major basement unit, was emplaced at 2596 +13/-10 Ma. Retrograded, porphyritic orthopyroxene granite, intrusive into gneisses of a granulite facies belt east of the head of Chantrey Inlet, is 2587 +9/-8 Ma old. Weakly gneissic granodiorite, intrusive into what appears to be the Barclay supracrustal belt, gives an age of ca. 2590-2596 Ma, consistent with the postulate that this highly deformed and disrupted belt is Archean.

Résumé

La datation sur zircon de trois échantillons de roche de socle ignée métamorphisée prélevés dans la région de l'inlet Chantrey dans la province de Churchill donne 2,6 Ga (Archéen supérieur), ce qui est légèrement plus récent que les âges modèles par Sm-Nd sont de 2,7 à 2,8 Ga. La roche d'origine provient d'un gneiss oeillé à hornblende-biotite, représentative d'une importante unité de socle; elle a été mise en place à 2596 +12/-10 Ma. Le granite à orthopyroxène porphyrique régressif, qui recoupe par intrusion les gneiss d'une zone de faciès granulitique à l'est du fond de l'inlet Chantrey, remonte à 2587 +9/-8. Le granodiorite faiblement gneissique, qui recoupe par intrusion ce qui semble être la zone supracrustale de Barclay, a été daté à environ de 2590 à 2598 Ma, ce qui est conforme au postulat énonçant que cette zone très déformée et perturbée est archéenne.

INTRODUCTION

The northern part of the District of Keewatin is characterized by northeasterly-trending granitoid gneisses and supracrustal belts belonging to the Churchill Structural Province. Much of this region has been mapped only in reconnaissance fashion and geochronological data are accordingly sparse.

In this paper we present U-Pb zircon ages from the Cape Barclay map area (NTS 56M; formerly called Lower Hayes River area), which includes much of Chantrey Inlet, into the head of which flow the Back and Hayes rivers (Fig. 1). This area was geologically mapped by Frisch in 1984 (Frisch et al., 1985; Frisch, in press) and the rocks dated were collected during that 1:250 000 scale mapping project.

GEOLOGICAL SETTING

Most of the area is underlain by a variety of biotite-bearing quartzofeldspathic gneisses, predominantly of granodioritic composition and plutonic aspect, commonly dipping moderately to steeply northwest. Migmatite or more or less massive granitic rock is present locally but overall such rocks form only a small proportion of the basement. All major granitic bodies are pretectonic with respect to the last major deformation in the area. A northeast-trending belt of granulite facies rocks, 10 km wide, with ill-defined borders, cuts across the southeastern corner of the map area. This belt consists of (garnet-) orthopyroxene-biotite (ortho?)gneiss and (cordierite-sillimanite-) garnet-biotite paragneiss, intruded by gneissic to nearly massive porphyritic garnet-orthopyroxene-biotite granite; retrogression increases towards the borders of the belt. Dips of these high grade rocks are also moderate to steep to the northwest.

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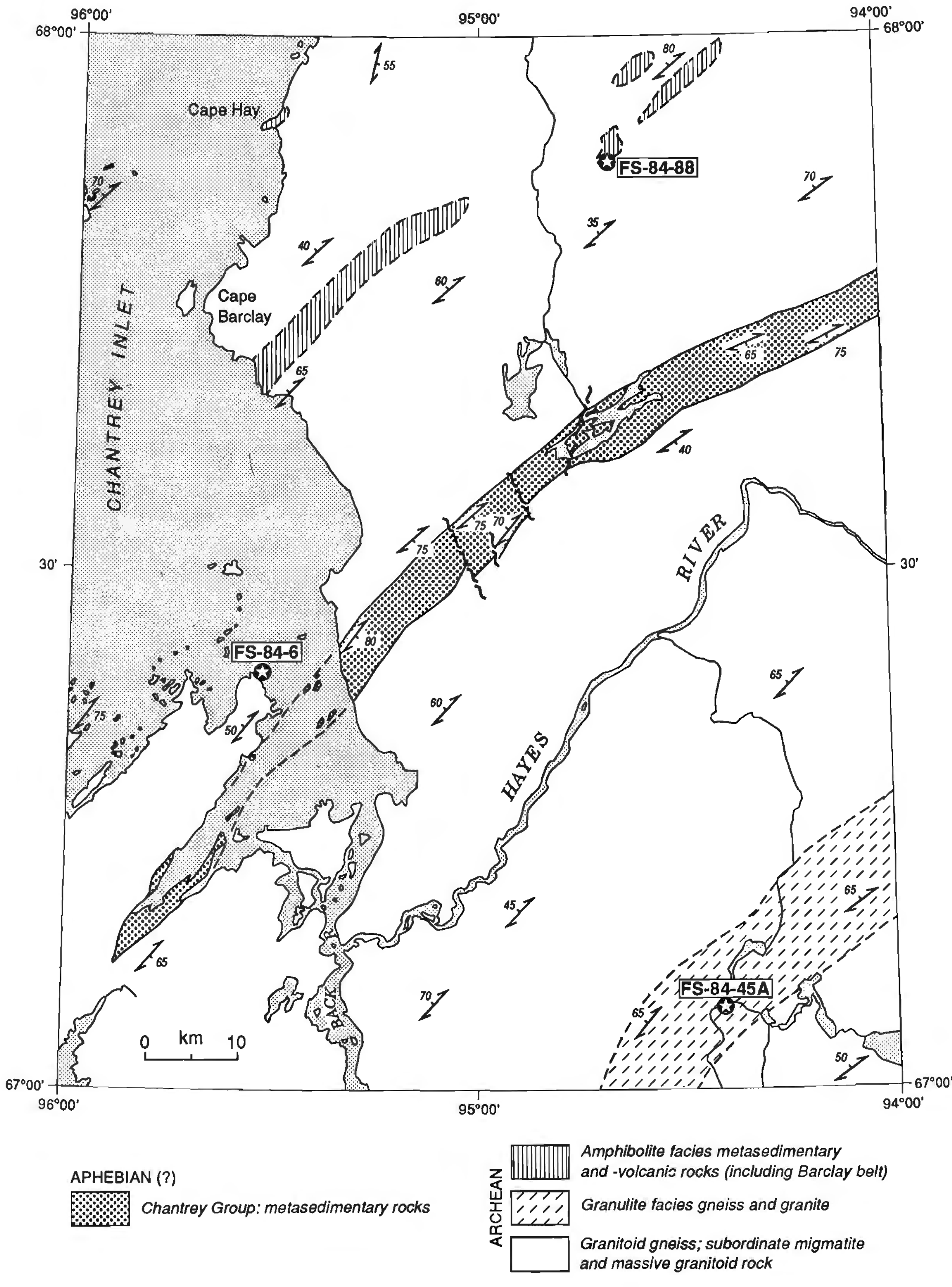

Figure 1. Geological sketch map of the Cape Barclay map area, showing locations of the three rocks dated by U-Pb methods.

Two major supracrustal belts traverse the area. The northern of the two, exposed south of Cape Barclay, is about 3 km wide, generally poorly exposed, and consists of highly deformed sedimentary and volcanic rocks metamorphosed in the middle amphibolite facies. The other belt, 3-7 km wide, consists entirely of metasedimentary rocks comprising the Chantrey Group (Heywood, 1961) and extends 175 km northeastward from west of the head of Chantrey Inlet.

The rocks of the northern supracrustal sequence, informally termed the Barclay belt, include cordierite-staurolite-andalusite-muscovite-biotite schist (much of it metaturbidite), locally with garnet and minor fibrolite, psammitic schist and quartzite, marble and calc-silicate rock, acid to intermediate metavolcanic rock, ironstone, and amphibolite and metagabbro. Pelitic and psammitic schists far outweigh the other rocks in abundance. The rocks were tightly folded during at least two episodes of deformation and are mylonitized along the margins of the belt. Contacts between basement and supracrustals are rarely exposed but appear to be largely tectonic, and outcrop patterns suggest disruption of the supracrustal belt by granitoid intrusion and deformation.

The Chantrey Group comprises shallow-marine carbonate and quartzose rocks and overlying metapelite/psammite, quartzite, dark pyritic and graphitic schist and locally, metaconglomerate. Just east of Chantrey Inlet, the bulk of the exposed Chantrey Group consists of stretched quartz-pebble metaconglomerate. Coarse andalusite and iron-rich garnet and biotite characterize the pelitic schists. Cordierite and staurolite occur sparsely throughout the belt; minor late fibrolite is confined to the western end of the belt. As in the Barclay belt, pressure and temperature of metamorphism in the Chantrey Group did not exceed ca. 0.38 GPa and ca. 550°C (T. Frisch, unpublished data).

The Chantrey Group is deformed into tight, more or less upright folds, overall forming an asymmetric synclinorium. Although the contact with basement is generally sheared and there is local northwestward overthrusting of basement, no evidence has been found of fragmentation of the Chantrey Group by remobilization of basement or granitic intrusion (except pegmatite dykes and sills). Thus the Chantrey belt is thought to be younger than the Barclay belt and to postdate granitic magmatism.

Muscovite pegmatite dykes and sills intrude the Barclay and Chantrey belts and a sill intrusive into Chantrey Group rocks east of Chantrey Inlet has yielded a K-Ar muscovite age of 1684 Ma (Heywood 1961; age recalculated according to the constants used in this report).

The youngest bedrock in the area is northwest-trending diabase dykes, which are sparsely distributed in the region and considered to be of Mackenzie (1.27 Ga) age.

The basement gneisses are all thought to be Archean, and no undeformed granitic bodies that may be Proterozoic have been recognized in the Cape Barclay map area. Prior to the work

Table 1. U-Pb analytical data for rocks from the Chantrey Inlet region[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb/^{204}Pb$*	Pbc,# (pg)	$^{208}Pb/^{206}Pb$+	$^{206}Pb/^{238}U$++	$^{206}Pb/^{238}U$ (Ma)++	$^{207}Pb/^{235}U$++	$^{207}Pb/^{235}U$ (Ma)++	$^{207}Pb/^{206}Pb$++	corr. coef.	$^{207}Pb/^{206}Pb$ (Ma)***
FS-84-6, Augen gneiss													
A,+149,c	0.024	333.0	181.2	9503	25	0.165	0.4731 ± 0.08	2497.0 ± 3.5	11.18 ± 0.10	2538.3 ± 1.8	0.1714 ± 0.03	0.96	2571.4 ± 0.9
B,+149,c	0.039	363.0	196.8	21330	20	0.174	0.4682 ± 0.09	2475.8 ± 3.6	11.04 ± 0.10	2526.5 ± 1.8	0.1710 ± 0.03	0.96	2567.5 ± 0.9
C,+149,c	0.044	382.3	205.0	10730	46	0.174	0.4635 ± 0.08	2455.1 ± 3.4	10.89 ± 0.10	2513.4 ± 1.8	0.1703 ± 0.03	0.96	2560.8 ± 0.9
FS-84-45A, Retrograded orthopyroxene granite													
A-1,sc	0.120	864.6	402.7	6401	459	0.046	0.4464 ± 0.10	2379.4 ± 4.1	10.32 ± 0.11	2464.0 ± 2.1	0.1677 ± 0.03	0.97	2534.7 ± 1.0
B-1,el	0.162	823.8	388.0	4438	859	0.041	0.4528 ± 0.12	2407.5 ± 4.7	10.52 ± 0.13	2482.1 ± 2.4	0.1686 ± 0.03	0.97	2543.8 ± 1.1
B-2,el	0.121	678.7	327.7	6932	347	0.046	0.4619 ± 0.09	2447.9 ± 3.8	10.80 ± 0.11	2505.9 ± 2.0	0.1695 ± 0.03	0.96	2553.2 ± 1.0
FS-84-88, Muscovite-biotite-granodiorite													
A-1,149,e	0.090	129.3	75.22	14030	26	0.199	0.4921 ± 0.09	2579.9 ± 3.8	11.80 ± 0.10	2588.4 ± 1.9	0.1739 ± 0.03	0.96	2595.1 ± 1.0
A-2,125,e	0.130	165.2	96.52	22600	30	0.205	0.4921 ± 0.09	2579.8 ± 3.6	11.81 ± 0.10	2589.3 ± 1.8	0.1740 ± 0.03	0.96	2596.7 ± 0.9
B-1,125,el	0.043	172.5	102.8	5397	43	0.235	0.4914 ± 0.09	2576.7 ± 3.6	11.75 ± 0.10	2584.9 ± 1.8	0.1735 ± 0.03	0.95	2591.3 ± 1.0

** Numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy

Weighing error = 0.001 mg.

+ Radiogenic Pb.

* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU

Total common Pb in analysis corrected for fractionation and spike.

++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.

*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

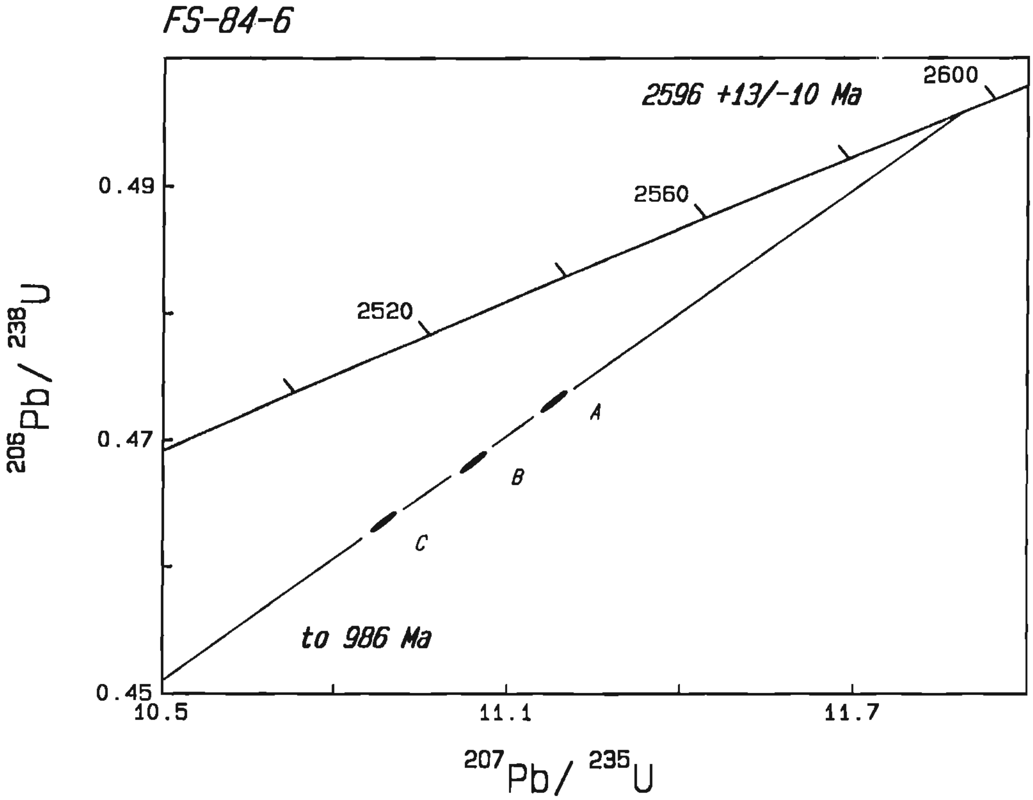

Figure 2. U-Pb concordia diagram for sample FS-84-6. Error ellipses are 2σ.

reported here, the only geochronological data on gneisses in the region were Sm-Nd "crust formation" ages (based on the depleted mantle model of DePaolo (1988)), determined by E. Hegner (pers. comm., 1991) on samples collected in 1984 and in 1982 during mapping of the Ian Calder Lake area (NTS 66I; formerly called Montresor River area), southwest of Chantrey Inlet. Sm-Nd T_{DM} model ages obtained on the three rocks for which zircon ages are reported below are: 2.68 Ga for augen gneiss from Chantrey Inlet (FS-84-6); 2.79 Ga for retrograded orthopyroxene granite from the granulite belt (FS-84-45A), and 2.70 Ga for gneissic granodioritic intrusive into Barclay belt (?) rocks (FS-84-88). Sm-Nd model ages from three other granitoid basement rocks in the region also range from 2.7 to 2.8 Ga.

U-Pb GEOCHRONOLOGY

Analytical methods

U-Pb methods follow those outlined in Parrish et al. (1987), including air abrasion of zircons (Krogh, 1982). Analytical data are summarized in Table 1 and the concordia diagrams of Figures 2, 3, and 4. Additional notes on analytical methods are given at the bottom of Table 1.

Augen gneiss, Chantrey Inlet (FS-84-6)

Medium grained biotite-rich granodioritic-granitic gneiss with pink or white augen of microcline perthite, up to several centimetres long, is the most abundant basement rock not only in the vicinity of Chantrey Inlet, but also in the Ian Calder Lake area to the southwest. Locally it appears to grade into virtually undeformed megacrystic granodiorite or granite, suggesting that the augen gneiss is derived from a magmatic rock. Further evidence of an intrusive origin is provided by the sporadic occurrence of lenses, with sharp borders, of fine grained gneiss and mafic schist.

The sample dated, FS-84-6, was collected on an island off the western shore of Chantrey Inlet, just north of the Chantrey belt (Fig. 1). Pink perthite megacrysts, up to 6 cm long, are aligned parallel to the foliation in a dark, granoblastic matrix of slightly chloritized brown biotite, fresh green hornblende, plagioclase An_{25-27}, microcline perthite and abundant quartz. Titanite is an abundant accessory mineral. The appearance of the rock in outcrop is heterogeneous: foliation ranges from swirly in places to straight in narrow shear zones, and the feldspar megacrysts are locally aggregated.

Zircons analyzed were euhedral to subhedral, with slightly rounded or resorbed outer surfaces. The grains had common inclusions, were only of moderate quality, and as seen in Figure 2, are discordant. Three fractions of zircon lie within error on a line connecting 986 Ma as the lower intercept and 2596+13/-10 Ma as the upper intercept.

The age obtained, 2596 +13/-10 Ma, is interpreted to be the time of emplacement of the megacrystic granite. This time coincides with major felsic magmatism in the central part of the District of Keewatin (LeCheminant and Roddick, 1991).

Retrograded orthopyroxene granite, south of Hayes River (FS-84-45A)

Brownish-weathering, coarsely porphyritic orthopyroxene granite ("charnockite") is intimately associated with orthopyroxene gneiss and subordinate garnet paragneiss in the granulite belt trending southwest across the southeastern part of Cape Barclay map area. The granite commonly is retrograded – orthopyroxene having been replaced by fine grained biotite – and is variably deformed, in some places being strongly gneissic and sheared, elsewhere merely showing alignment of the K-feldspar phenocrysts.

The sample dated, FS-84-45A, comes from an outcrop in a tributary stream to the Hayes River (Fig. 1). The rock consists of well-aligned, euhedral, pink K-feldspar phenocrysts, up to 2 cm long, in a medium grained matrix of plagioclase An_{25}, K-feldspar, quartz, red-brown biotite, garnet, and biotite pseudomorphs after orthopyroxene. The phenocrysts, which are of hair perthite, and smaller plagioclase crystals in the matrix are bordered by fine grained, inequigranular, recrystallized feldspar + quartz and tiny flakes of biotite. Ribbon quartz is poorly developed. Overall, the matrix texture is of mortar type. Red-brown biotite is the chief mafic mineral and garnet is partly replaced by well crystallized biotite. The former presence of orthopyroxene is indicated by fine grained, orange-coloured biotite, partly overgrown by a thin rim of blue-green hornblende.

A variety of inclusions occurs in the granite. Among the most common are lenses, 0.5-1 m long, of fine grained lineated gneiss consisting of olive-green hornblende, partly altered orthopyroxene, dark brown biotite, plagioclase An_{32}, perthite and quartz. Another type of inclusion is a dark, garnet-rich, micaceous rock (paragneiss?). Also present in the outcrop sampled is a sheet-like body of strongly flattened augen gneiss forming a septum of country rock or a large inclusion. This augen gneiss, with its brown, rather than red-brown, biotite and microcline, instead of hair perthite, is indistinguishable from much of the amphibolite facies basement. Thus the porphyritic granite not only postdates the granulite facies gneisses but may also be younger than augen gneiss of the lower grade basement.

Zircons from the gneiss were generally high in uranium, were cloudy to moderately clear, and had common inclusions. The grain morphology was of two types, round equant grains and elongate subhedral grains. Crystal faces in both types had a rounded aspect, probably indicative of granulite facies conditions. Overgrowths may have been present, but are not thought to have been substantial. The three zircon fractions analyzed were well abraded. A linear fit through the three points has an upper intercept with concordia of 2587 +9/-8 Ma (Fig. 3).

The zircon age of 2587 +9/-8 Ma is interpreted as the age of intrusion of the granite and represents a minimum age for the granulite facies gneisses. The aluminous mineralogy of

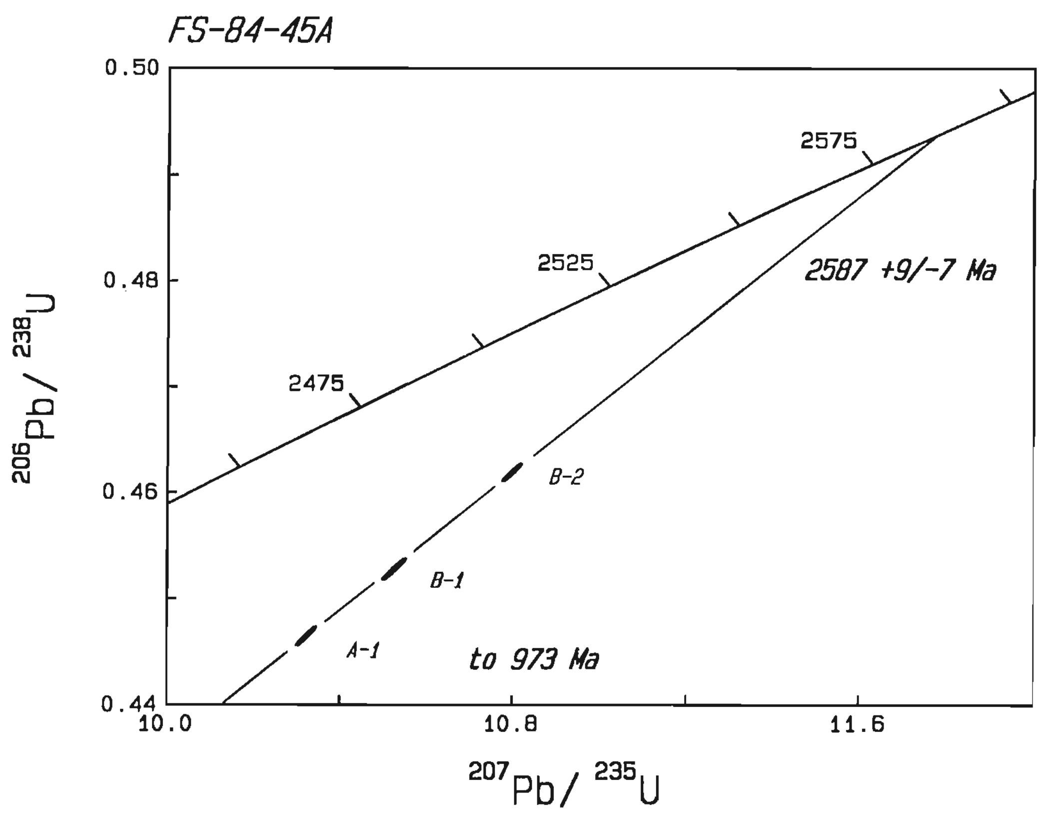

Figure 3. U-Pb concordia diagram for sample FS-84-45A. Error ellipses are 2σ.

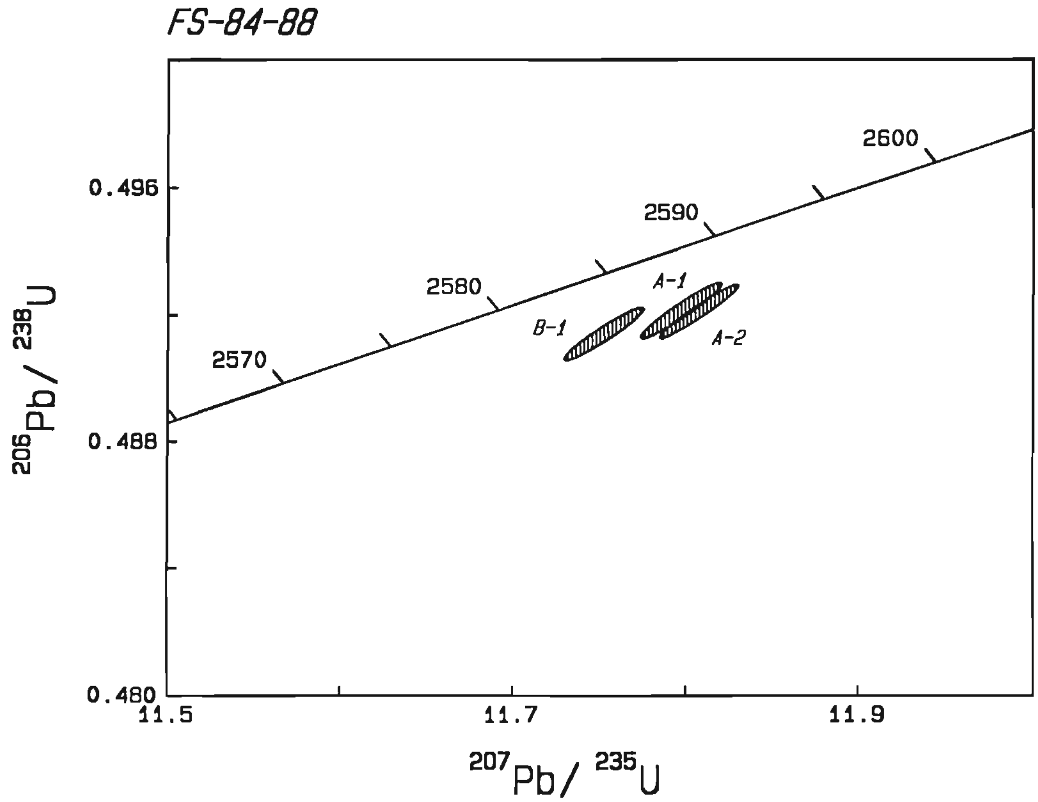

Figure 4. U-Pb concordia diagram for sample FS-84-88. Error ellipses are 2σ.

the granite suggests the rock was derived from melting of felsic material, emplaced in the crust at 2.79 Ga, according to the Sm-Nd data.

The nearest granulite facies rocks that have been dated lie 150 km to the south in Woodburn Lake map area (Fraser, 1988). "Granulite" from this complex gave a preliminary U-Pb zircon age of 2662 +78/-53 Ma (Fraser, 1988, p. 10). This rock may be equivalent to the high grade gneisses included in, and surrounding, the porphyritic granite, but the large uncertainty in the age renders this conjectural.

Muscovite-biotite granodiorite, east of Cape Hay (FS-84-88)

This rock was selected for dating because it intrudes what appears to be supracrustal rock of the Barclay belt. Well exposed on the shore of Chantrey Inlet, the belt outcrops increasingly poorly inland in an intermittent array of exposures stretching northeast to the northern border of the Cape Barclay map area. It was in this latter region that an exposure of a contact between granitic rock and mica schist similar to that of the Barclay belt was found and sampled. Exposed contacts between granitic and Barclay belt rocks are rare and where seen elsewhere are shear zones lacking any evidence of age relationships between granitic and supracrustal rocks.

At the sample locality, 39 km east of Cape Hay (Fig. 1), lenses of biotite schist, up to several metres long, are enclosed in a weakly gneissic two-mica granodiorite near a 300 m^2 outcrop of schistose rock (biotite schist and biotite-hornblende schist). Unfortunately, the schistose rocks are not diagnostic of Barclay belt supracrustals and may be unrelated to the Barclay belt. However, the fact that similar rocks occur to the northeast and southwest, all on strike with undisputed Barclay belt exposures, makes it reasonable to assume that the inclusions in the granodiorite belong to the Barclay belt.

The granodiorite, FS-84-88, is weakly gneissic in outcrop but has a well developed granoblastic (granitic) texture. Euhedral crystals of partly sericitized and epidotized plagioclase An_{26-30} lie in a matrix (grain size 1 mm) of anhedral plagioclase, microcline and quartz. Slightly chloritized brown biotite (1 mm) and minor late muscovite show no preferred orientation in thin section. Titanite, which tends to form overgrowths on biotite, is an abundant accessory mineral. The magmatic texture leaves little doubt that the granodiorite is of intrusive origin and that the schist inclusions were caught up in a melt and did not result from break-up by tectonism.

Analyzed zircons in the granodiorite were sharp euhedral grains (fraction B-1) and more equant euhedral grains which appeared moderately rounded. They were moderately clear, contained inclusions and some fractures, but are only slightly discordant. The three fractions cluster closely with ^{207}Pb-^{206}Pb ages ranging from 2591 to 2597 Ma (Fig. 4). Fractions A-1 and A-2 are of equant grains and appear a little older than the elongate euhedral ca. 2591 grains of fraction B-1. Some of the grains, particularly the more equant ones, could be xenocrysts from host gneisses, and it is possible that the elongate grains more closely date the emplacement of the granodiorite magma. Using this reasoning we interpret the granodiorite as being ca. 2591 Ma old. It is difficult to assign a meaningful uncertainty to this age estimate, but it is unlikely that the magma was emplaced prior to 2595 Ma or so.

The available analytical data do not permit calculation of a precise age but indicate ca. 2590-2595 Ma for the time of intrusion of the granodiorite. If the schistose inclusions are from the Barclay belt, the latter must be at least late Archean and very likely older than the Chantrey Group, which is not known to be intruded by granitic rock.

The foregoing results show that several varieties of granitic rock were emplaced ca. 2590-2595 Ma in the Chantrey Inlet region, providing further evidence of major 2.6 Ga felsic magmatism in the District of Keewatin.

ACKNOWLEDGMENTS

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Nd- and Pb-isotope studies of monazite, zircon and other minerals in granitoids: examples from northeastern Superior Province, Quebec and Mount Everest, Tibet

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Stern, R.A., 1992: Nd- and Pb-isotope studies of monazite, zircon and other minerals in granitoids: examples from northeastern Superior Province, Quebec and Mount Everest, Tibet; in Radiogenic Age and Isotopic Studies: Report 5, Geological Survey of Canada, Paper 91-2, p. 43-50.

Abstract

Single and multiple grains of monazite from tonalite, granodiorite, charnockite, and diatexite from the Archean Minto block, and a tourmaline leucogranite sample (Delta 33) from Mount Everest have been analyzed for Nd and Pb isotopes and rare earth elements. Monazites are shown to be useful for Nd isotopic studies because they have sufficiently high Nd concentrations (6-10 wt.%) to permit analysis of single grains, their Sm/Nd ratios tend to be lower than those of their host rocks (i.e. their $^{143}Nd/^{144}Nd$ ratios are closer to initial values), and they can be dated precisely with the U-Pb method.

Other minerals from the Minto granitoids analyzed for Nd isotopes were separates of zircon, apatite, plagioclase, orthopyroxene, and carbonate. It is evident in the mineral isochron of one sample that apatite was open to exchange of Nd for at least 100 Ma after crystallization, indicating that apatite should be used cautiously in Nd isotope studies. Similar and more severe effects were observed in the Sr isotope systematics of apatite. The other minerals remained closed to diffusion of Nd since crystallization and cooling, demonstrated by an internal mineral isochron for a Minto tonalite which gives an age of 2685 ± 40 Ma, overlapping the lower limit of the U-Pb zircon age (2734 +13/-24 Ma).

Résumé

Des grains simples et multiples de monazite provenant de tonalites, granodiorites, charnockites et diatexites du bloc archéen de Minto, et un échantillon d'un leucogranite à tourmaline (Delta 33) du mont Everest ont été analysés pour des isotopes de Nd et de Pb et des éléments des terres rares. Les monazites sont utiles dans les études d'isotopes de Nd parce qu'elles ont des teneurs de Nd suffisamment élevées (6-10 p. cent pds.) pour permettre l'analyse de grains simples, leurs rapports Sm/Nd ont tendance à être moins élevés que ceux de leurs roches encaissantes, et on peut en faire la datation précise par la méthode U-Pb.

D'autres minéraux des granitoïdes du bloc Minto analysés pour des isotopes de Nd étaient des concentrés de zircon, d'apatite, de plagioclase, d'orthopyroxène et de carbonate. Il est évident par l'isochron minéral de l'un des échantillons que l'apatite était ouverte à l'échange de Nd pendant au moins 100 Ma après la cristallisation, indiquant que l'apatite devrait être utilisée prudemment dans les études d'isotopes de Nd. Des effets semblables et plus sévères ont été observés dans la systématique d'isotopes de Sr de l'apatite. Les autres minéraux sont demeurés clos à la diffusion de Nd depuis la cristallisation et le refroidissement, tel que démontré par un isochron minéral interne d'une tonalite du bloc Minto, qui donne un âge de 2685 ± 40 Ma, chevauchant la limite inférieure de l'âge U-Pb de zircons.

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INTRODUCTION

Over the last decade, the determination of the isotopic composition of neodymium in rocks has become routine, mainly through advancements in chemical separation methods and in mass spectrometry. One of the important developments in data collection is that it is now possible to measure the ratios of Nd^+ isotopes to the required precision of at least 0.005% by simultaneous (or "static") ion counting using two or more Faraday cups. The advantages of static ion counting compared to the single-collector, peak hopping method are that the analysis times are reduced by 50% or better, and that the ion beam does not have to be stable. This means that it is possible to analyze a smaller mass of Nd (tens instead of hundreds of nanograms) using the static ion counting method, conditions under which the ion beam may decay rapidly or exhibit instability.

The small sample sizes permitted by static ion counting allow Nd isotopic compositions to be measured precisely in minerals as well as rocks. This development has special relevance at the Geological Survey of Canada because large quantities of high-purity mineral separates are routinely collected for U-Pb geochronological studies, and these can potentially be used for further petrogenetic studies. From the standpoint of Nd isotopic studies, the advantage of working with minerals as opposed to rocks is that the Nd isotopic composition can potentially be tied directly to a U-Pb age. For whole-rocks, this can usually only be inferred from U-Pb geochronology, or from the whole-rock Nd-Sm isochron method. Of the minerals used in U-Pb geochronology, monazite $\{(La,Ce,Nd,Y,Th)PO_4\}$ is highly suitable for Nd isotopic analysis due to its high concentrations of Nd (e.g. Ross et al., 1991), but there are other minerals such as zircon, apatite, sphene, and garnet which can also be valuable in petrogenetic and geochronological studies.

This report highlights some of the results of measuring Nd isotopic compositions and rare earth element concentrations of monazite, zircon, apatite, feldspar, pyroxene, and carbonate from granitoids of the 2.7 Ga Minto block of northern Quebec (Percival et al., 1990, 1991), and from a peraluminous tourmaline granite sample from Mount Everest, Tibet. The samples from the Minto block are fresh,

Table 1. Isotopic data on minerals and whole-rocks from the Minto block and from Mount Everest

	Wt. sample (mg)	Number of grains	Nd (ppm)	Sm (ppm)	147Sm/ 144Nd*	143Nd/ 144Nd**	U-Pb Age in Ma (assumed)		Epsilon Nd (T)	T(CHUR) (Ga)	Comments
MINTO BLOCK											
PBAS-89-70			10.285	1.3245	0.077815	0.510521	2688	+3.7/-3.3	-0.17	2.70	zircon U-Pb age
monazite	0.005	1	67616	5444.1	0.048646	0.509985	2676	+/- 1.6	-0.77	2.72	207/206 age
monazite	0.027	1	66870	6039	0.054628						REE's only
PBAS-89-29			21.164	2.9175	0.083299	0.510631	2725	+/- 4	0.63	2.68	zircon U-Pb age
leached w.r.			14.712	1.5817	0.064963	0.510320	(2725)		1.00	2.67	
monazite	0.835	several	79456	8891.4	0.067616	0.510364	2705	+/- 1.4	0.59	2.67	207/206 age
monazite	0.01	1	88947	10024	0.068170						REE's only
apatite	27.58		2389.6	570.09	0.144195	0.511673	(2725)		-0.41	2.78	
feldspar	290.47		0.8332	0.0812	0.058944	0.510241	(2725)		1.57	2.64	
PBAS-89-21			9.4224	1.384	0.088763	0.510737	2734	+13/-24	0.91	2.67	zircon U-Pb age
zircon	13.96	several	81.004	12.594	0.093954	0.510792	(2734)		0.15	2.72	
orthopyroxene	64.63		1.8968	0.5903	0.188158	0.512476	(2734)		-0.16	2.87	
apatite	38.2		1707.5	374.8	0.132665	0.511497	(2734)		0.28	2.70	
plag (white)	370.04		0.9959	0.0875	0.053143	0.510083	(2734)		0.69	2.70	
plag (red)	268.52		6.0275	0.5166	0.051791	0.510076	(2734)		1.03	2.68	
carbonate	14.4		105.84	24.029	0.13723	0.511602	(2734)		0.72	2.64	
PBAS-89-78			15.43	1.8769	0.073503	0.510347	2713	+/- 2.0	-1.68	2.82	zircon U-Pb age
monazite	0.017	1	80058	10055	0.075892	0.510385	2688	+/- 0.9	-2.17	2.83	207/206 age
monazite	0.019	1	83963	11115	0.080076						REE's only
PBAS-90-99			18.113	2.456	0.081949	0.510555	2724	+/- 4.5	-0.40	2.75	zircon U-Pb age
monazite	0.015	1	54861	4370	0.048123	0.509969	2704	+/- 1.5	-0.36	2.72	207/206 age
monazite	0.018	1	65596	5516	0.050866						REE's only
EVEREST											
DELTA 33			11.45	3.22	0.170111		20	+/- 1			(U-Pb; Copeland et al.)
monazite	0.015	1	83178	21107	0.153497						REE's only
monazite 12	0.016	1	104360	25948	0.150297	0.511922	21.1	+/- 0.1	-13.84	2.34	207/235 age
monazite 13	0.0204	1	98276	21845	0.134361	0.511955	20.6	+/- 0.1	-13.16	1.67	207/235 age
monazite 14	0.0156	1	86364	18850	0.131935	0.511951	20.1	+/- 0.1	-13.24	1.61	207/235 age
monazite 15	0.0082	1	104690	22576	0.130354	0.511973	20.1	+/- 0.2	-12.80	1.52	207/235 age
monazite 16	0.0181	1	84513	19665	0.140651	0.511945	20.2	+/- 0.1	-13.37	1.88	207/235 age

*147Sm/144Nd ratios have an uncertainty of 0.05% (2SD)

**143Nd/144Nd ratios have an uncertainty of 0.003% (2SD) unmetamorphosed tonalite (#21: 75°27′W, 57°31′N), granodiorite (#29: 75°22′W, 57°31′N; #99: 72°13′W, 58°10′N), charnockite (#70: 74°35′W, 57°47′N), and diatexite (#78: 74°18′W, 57°28′N) that crystallized from high-temperature (>900°C) magmas under relatively dry conditions. The mineral assemblage of these rocks includes two feldspars, quartz, orthopyroxene, biotite, magnetite, apatite, zircon, and monazite. Geological descriptions of these rock units can be found in Percival et al. (1990, 1991). Monazite grains from the 20 Ma old tourmaline-leucogranite sample "Delta 33" (Copeland et al., 1988) were kindly provided by R. Parrish. This sample was studied because some of the monazites showed an inherited Pb component of about 471 Ma (Copeland et al., 1988; Parrish, 1990), a feature which should be evident in the Nd isotope systematics.

ANALYTICAL METHODS

Monazite is sufficiently enriched in Nd that single grains of about 5 μg or greater were found to be sufficient for analysis (Table 1). Therefore, for some grains, it was possible to obtain both U-Pb ages and Nd isotopic analyses. These single grains were weighed and then spiked for U-Pb analysis only prior to

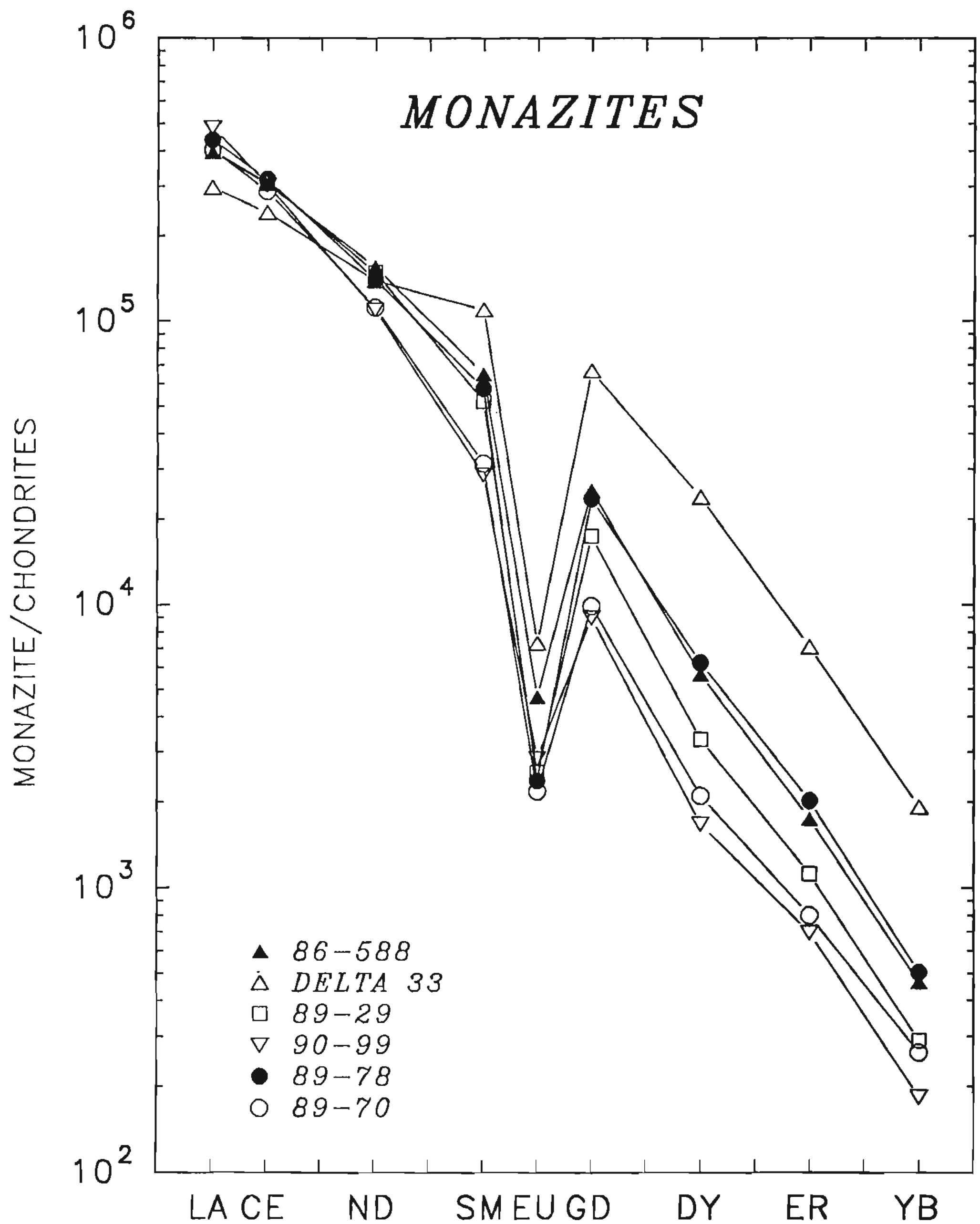

Figure 1. Chondrite-normalized rare earth element diagram of single monazite grains from granodiorite (#29, #99), charnockite (#70), and diatexite (#78) from the Minto block, tonalite from the Ashuanipi complex (#588), and leucogranite from Mount Everest (sample Delta 33).

dissolution in 12N HCl at 200°C for 48 hours in TFE Teflon containers. The procedures for chemical separation of U and Pb are outlined in Parrish et al. (1987), except that prior to the elution of Pb, Nd and Sm were collected by passing 3.1 N HCl through the column. It is uncertain whether this method gives quantitative recovery of Nd and Sm, however, this is unimportant for the Nd isotopic analysis, as a mixed ^{149}Sm and ^{148}Nd spike was added to the solution. The spiked solutions were dried down, taken up in 0.18N HCl and loaded directly onto HDEP-coated Teflon columns for separation of Nd and Sm (see Thériault, 1990).

Procedures for the dissolution of zircon for Nd-Sm chemistry are the same as for U-Pb (Parrish et al., 1987), except that the mixed Sm-Nd spike was added to the zircon separate prior to dissolution. The zircon separate was not analyzed for Pb isotopes, but prior U-Pb analyses of single zircons from the rock (#29) revealed only one age population. After dissolution, the zircon sample was loaded directly onto the Teflon columns for separation of Nd from Sm. The weighed and spiked orthopyroxene and feldspar separates were dissolved overnight in HCl-HNO_3 solutions in Savillex beakers on a hot plate. Apatite separates weighing about 30 mg were spiked for Nd-Sm and Rb-Sr and dissolved in 6N HCl in Savillex beakers on a hot plate. Carbonate present in sample #21 was analyzed by leaching carbonate-rich vugs in 1N HCl. Chemical separation of Nd and Sm from apatite and orthopyroxene followed the two column procedure outlined in Theriault (1990). The procedure for whole-rock Nd isotopic analyses is outlined in Stern and Hanson (1991).

Separate rare earth element determinations (La, Ce, Nd, Sm, Eu, Gd, Dy, Er, Yb) of single grains of monazite were also carried out by isotope dilution using the previously mentioned analytical procedures for monazite, and have a precision of ±1-2%.

Samples for isotopic analysis of Nd were loaded as phosphates onto outgassed Re side filaments. Mass spectrometry was carried out on the Finnigan Mat 261, utilizing 6 Faraday cups in the static multicollection mode. All data are corrected to give LaJolla $^{143}Nd/^{144}Nd$=0.511860, normalized to $^{146}Nd/^{144}Nd$=0.7219. Present-day chondritic values used for epsilon Nd calculations are $^{143}Nd/^{144}Nd$=0.512638 and $^{147}Sm/^{144}Nd$=0.1967. Neodymium isotope data and U-Pb ages are summarized in

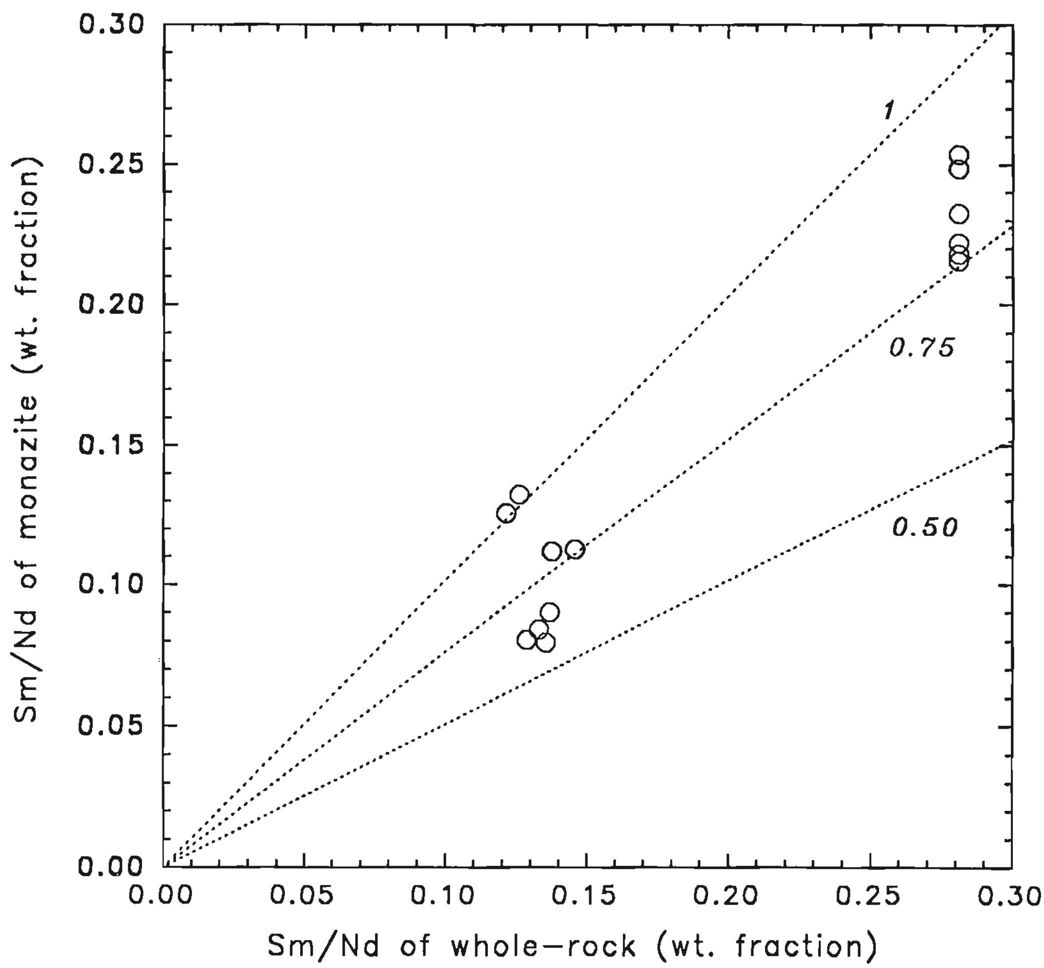

Figure 2. Sm/Nd ratios in monazites vs. the Sm/Nd ratios of whole-rock samples of the host granitoids.

Table 1. The U-Pb ages include unpublished data from this study and from Stern et al. (1991). All quoted age uncertainties are at the 2 sigma level.

RESULTS

Figure 1 shows chondrite-normalized rare earth element patterns of single monazite grains from the Minto area (70, 78, 29, 99), Mount Everest (Delta 33), and a tonalite sample from the Archean Ashuanipi complex (588). Because the light rare earth elements (REEs) are essential structural constituents in monazite, their abundances can vary only within certain stoichiometric limits, and therefore the light REE abundances are relatively similar. The large Eu anomaly indicates that Eu^{2+} (1.25 nm) is too large to be incorporated easily into the structure.

In Figure 2, the Sm/Nd ratios of monazite grains are plotted against the Sm/Nd ratios of the whole-rocks. The Sm/Nd ratios of monazite are generally lower than those of the whole-rocks, but there are two monazite analyses with similar ratios (sample #78). The tendancy for the Sm/Nd ratios of monazites to be smaller than the whole-rock means that their $^{143}Nd/^{144}Nd$ ratios will grow more slowly, which is advantageous for determination of initial Nd isotopic ratios.

In Table 1 are listed the U-Pb ages obtained for zircon and monazite from the study samples. For the Archean rocks, the monazite U-Pb ages are younger than the zircon ages by up to 20 Ma. This difference may reflect the lower closure temperatures for diffusion of U and Pb in monazite (Parrish, 1990), or it could reflect later growth of monazite. The timing of closure of the Sm-Nd system in monazite is not known for the Minto rocks; it could be at the zircon U-Pb age, at the monazite U-Pb age, or some other age. It is most likely with fresh rocks such as these that exchange of Nd between monazite and its surroundings occurred under magmatic conditions, close to the time of zircon growth. In any case, the growth in the $^{143}Nd/^{144}Nd$ ratio in monazite in less than 20 Ma will be so slight as to be undetectable with current analytical methods.

In Figure 3 are plotted the epsilon Nd values of monazites from the Archean Minto block calculated at the monazite Pb-Pb ages, vs. the epsilon Nd values of the whole-rocks

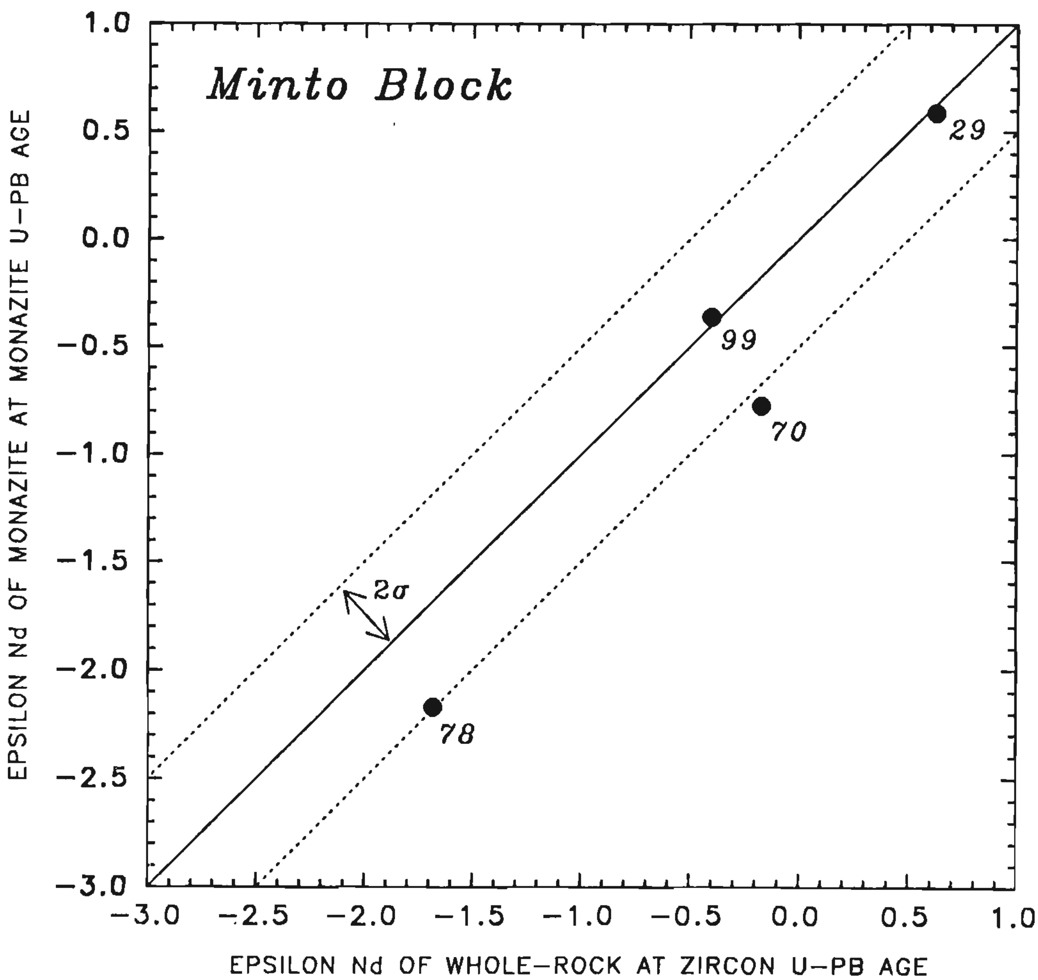

Figure 3. Comparison of epsilon Nd values of monazites and whole-rocks from the Archean Minto block.

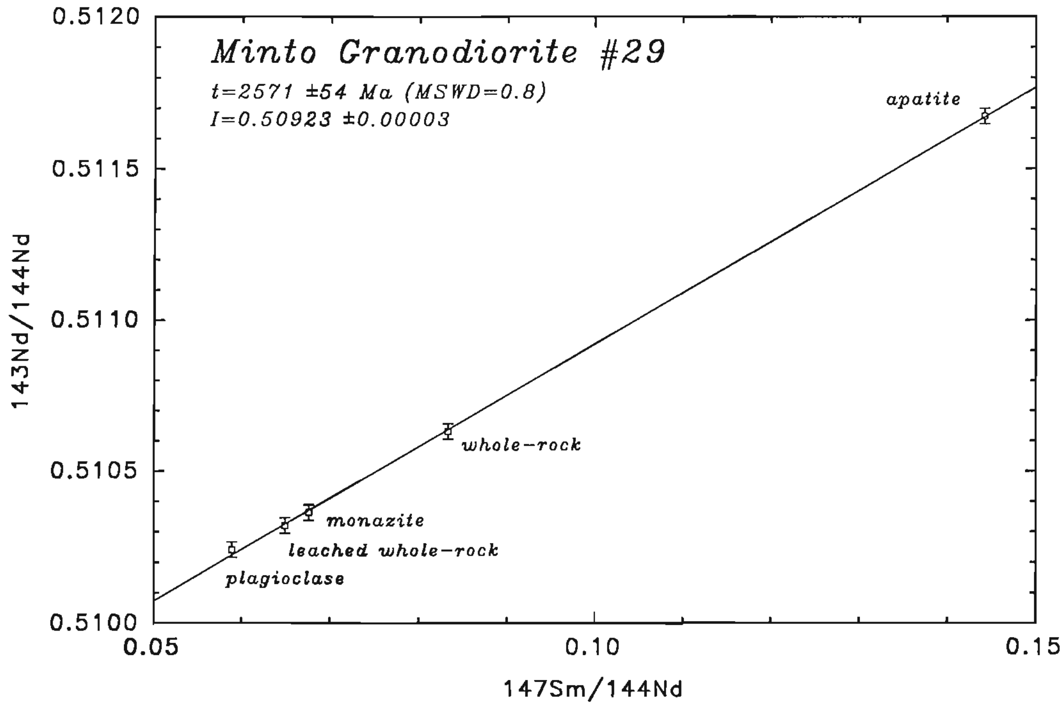

Figure 4. Mineral isochron for granodiorite #29 from the Minto block.

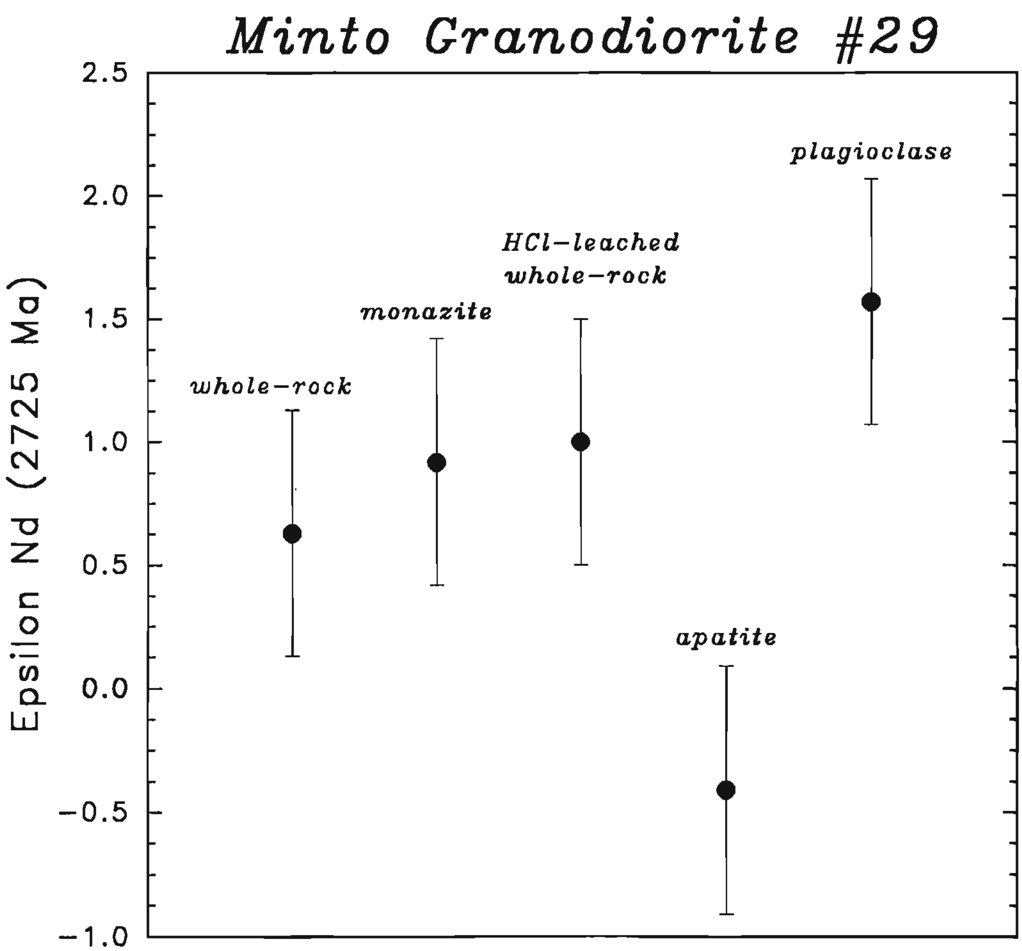

Figure 5. Comparison of epsilon Nd values of minerals from granodiorite sample #29 at the zircon crystallization age of 2725 Ma.

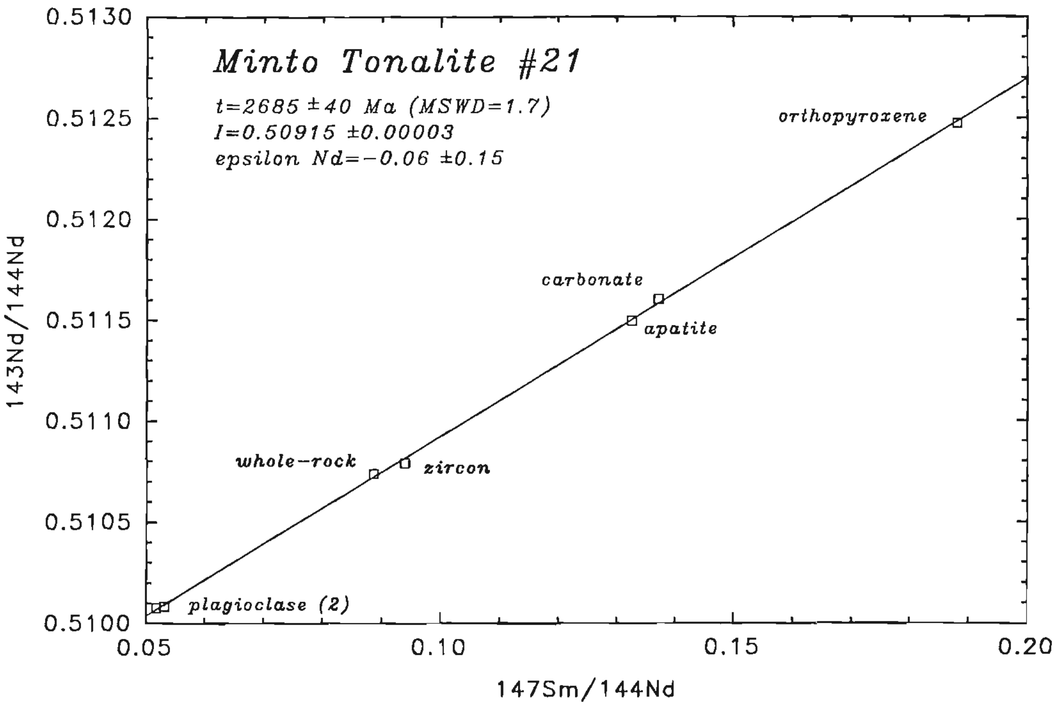

Figure 6. Mineral isochron for tonalite sample #21 from the Minto block.

calculated at the zircon U-Pb ages. The epsilon Nd values of the monazites are within error of the epsilon Nd values of the whole-rocks. Thus, the monazites from these rocks can be used as an alternative to the whole-rocks to estimate the initial isotopic compositions of the magmas. It has been observed elsewhere (e.g. the Ashuanipi Complex, southeastern Superior Province) that there can be more than 50 Ma difference in U-Pb ages between monazite and zircon (J.K. Mortensen, pers. comm.). In these suites, greater care must be taken in using monazite for inferring magmatic isotope compositions, as the monazite may have grown at a later stage from fluids, or may have recrystallized under metamorphic conditions. There is also the possibility of inherited or xenocrystic monazite (e.g. Copeland et al., 1988).

Figure 4 shows a Nd isochron diagram for monazite, plagioclase, whole-rock, the whole-rock after leaching in weak HCL, and apatite from Minto granodiorite sample #29. The Nd-isochron age of 2571 ± 54 Ma (MSWD=0.8) for this sample is significantly younger than the U-Pb zircon age of 2725 ± 4 Ma. Apatite may be the mineral responsible for the low age of the isochron, as its epsilon Nd value at 2725 Ma falls below the range of values for the other minerals (Fig. 5). Furthermore, the whole-rock/apatite Sr-isochron age is 2392 ± 8 Ma. Together, these data suggest that the apatite was open to diffusion of Nd and Sr or was recrystallized long after the magma solidified. Thus, care must be taken when using apatite for Nd (and Sr) isotopic studies.

Figure 6 shows a mineral isochron for orthopyroxene tonalite sample #21 from the Minto block, incorporating whole-rock, plagioclase, zircon, apatite, orthopyroxene, and carbonate. The regression yields an age of 2685 ± 40 Ma (MSWD=1.7), which overlaps the lower limit of uncertainty of the U-Pb zircon age of 2734 +13/-24 Ma. In contrast to sample #29, the Nd system in the apatite separate does not appear to be disturbed. Nevertheless, the Sr isotopic composition of the apatite does appear to be reset, as the apatite/whole-rock Sr-isochron age is only 1477 ± 66 Ma. Note that the zircon separate falls on the regression line, which suggests that the zircon fraction analyzed was largely magmatic and included little or no inherited component. This finding is consistent with the whole-rock isotope data which give an epsilon Nd (2734 Ma) value of +0.91, indicating that the protolith was largely juvenile. The carbonate also plots on the regression line, so that if CO_2 played a role in the origin of the rocks, it was likely magmatic.

Finally, the U-Pb analyses of monazites from the Mount Everest sample Delta 33 gave ^{207}Pb-^{235}U ages of 20-21 Ma, which are consistent with previously reported magmatic ages for the sample (Copeland et al., 1988). Unfortunately, the previously reported monazites with inherited Pb were not found in this study. Nevertheless, it is quite clear from the T_{CHUR} (1.5-2.3 Ga) ages of the magmatic monazites that the protolith of this magma contained material which separated from the mantle in or before the Proterozoic. This finding is consistent with the presence in this sample of inherited zircons with upper intercept ages of 500-2300 Ma (Copeland et al., 1988; Parrish, 1990).

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- 7 -

U-Pb zircon age for a quartz porphyry in the Thunderhill Lake area, Kisseynew gneiss belt, Manitoba

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Hunt, P.A. and Schledewitz, D.C.P., 1992: U-Pb zircon age for a quartz porphyry in the Thunderhill Lake area, Kisseynew gneiss belt, Manitoba, in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 51-54.

Abstract

The Missi Suite of the Kisseynew gneiss belt Manitoba, has been interpreted to be the more highly metamorphosed and deformed gneissic equivalent of the metasedimentary rocks of the Missi Group. A U-Pb zircon age determination of 1826 +11/-5 Ma was determined for a quartz porphyry occurring within quartzofeldspathic gneisses of the Missi Suite at Thunderhill Lake. This result will help constrain the age of volcanism and depostion of Missi Suite rocks in the Kisseynew gneiss belt at Thunderhill Lake, Manitoba.

Résumé

La suite de Missi dans la zone gneissique de Kisseynew au Manitoba, a été interprétée comme l'équivalent gneissique le plus métamorphisé et le plus déformé des roches métasédimentaires du groupe de Missi. Par la méthode U-Pb sur zircon, on a daté à 1826 +11/-5 Ma un quartz porphyrique contenu dans les gneiss quartzo-feldspathiques de la suite de Missi au lac Thunderhill. Ces données permettront de limiter l'âge du volcanisme et de la sédimentation des roches de la suite de Missi dans la zone gneissique de Kisseynew au lac Thunderhill au Manitoba.

INTRODUCTION

This paper reports on the U-Pb zircon geochronology completed on a quartz porphyry that occurs within quartzofeldspathic gneisses of the Missi Suite at Thunderhill Lake (Fig.1). These Early Proterozoic rocks are part of the Kisseynew gneiss belt that abuts the Flin Flon volcanic belt to the south. The Missi Suite in this area comprises interlayered fine grained quartzofeldspathic paragneisses, with variable quartz content, interlayered with gneissic metaconglomerate and aluminous quartzofeldspathic rocks. Aphanitic and very fine grained, granitoid quartz phyric rocks, some of which are interpreted as metarhyolite, form discontinuous tabular bodies that appear to have been deformed and metamorphosed with the surrounding Missi Suite paragneisses.

GENERAL GEOLOGY

Burntwood Suite

Graphitic garnet-biotite migmatites, derived from greywacke- mudstone turbidites, lie in the core of a recumbent fold southeast of Thunderhill Lake, and flank the Adamson Lake Dome north of Thunderhill Lake. These rocks have been assigned to the Burntwood Suite, which is probably equivalent in age to the Amisk Group.

Missi Suite

The Missi Suite in the Kisseynew gneiss belt is interpreted to be the more highly metamorphosed and deformed gneissic equivalent of the Missi Group of the Flin Flon belt (Schledewitz, 1987; Zwanzig and Lenton, 1987). The Missi

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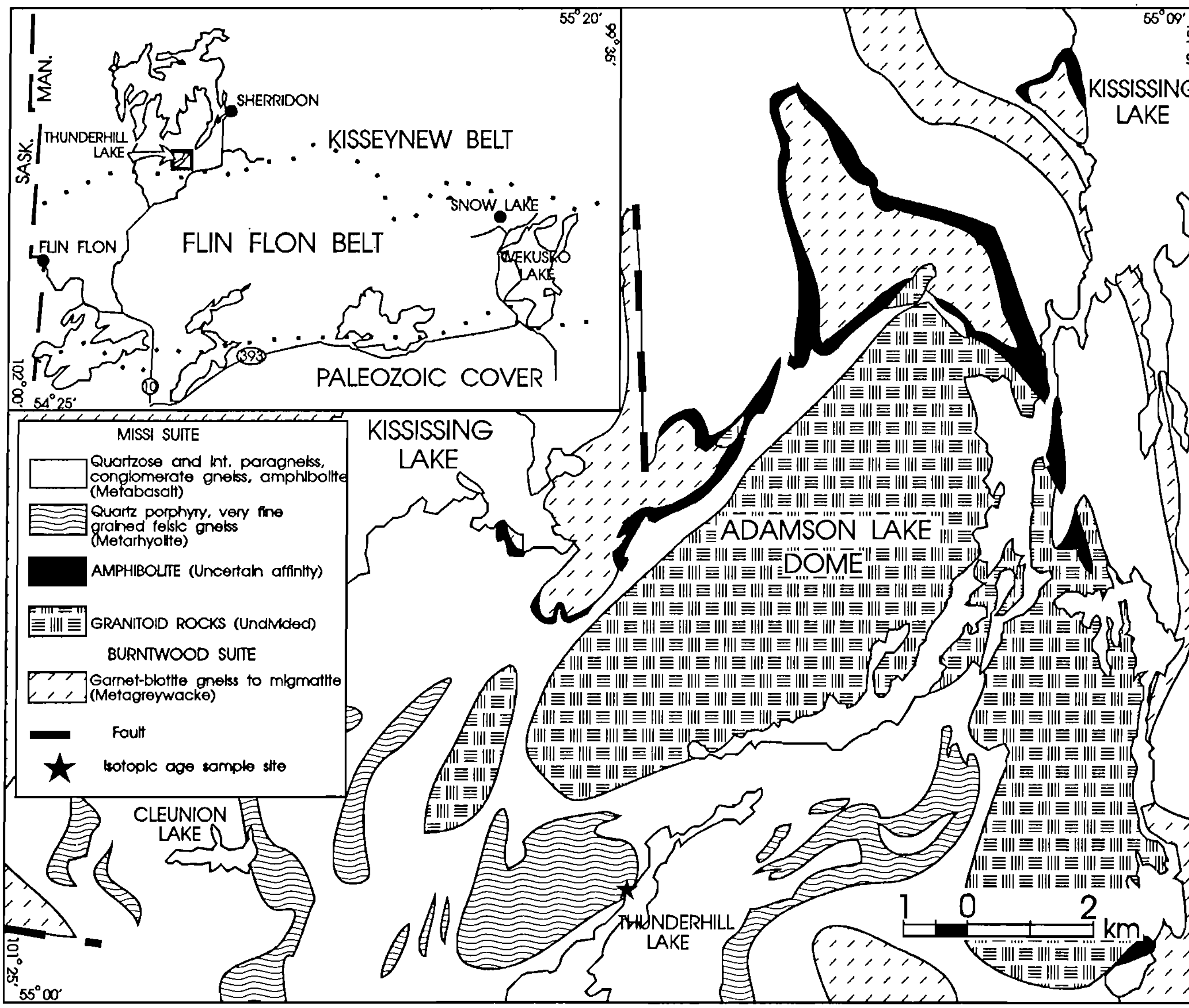

Figure 1. U-Pb zircon sample location and simplified geology of the Thunderhill Lake area, Manitoba.

Table 1. U-Pb zircon data

Fraction size[a]	Wt. mg	U ppm	Pb* ppm	$^{206}Pb^{b}/^{204}Pb$	Pb^{c} pg	^{208}Pb %	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	Corr.[d] Coeff.	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$ Age (Ma)
Thunderhill											
A -74+62 NM3 abr	0.010	375	117	2251	33	2.1	0.3196 ±0.09%	4.893 ±0.10%	.93	0.11104 ±0.04%	1816.6 ± 1.4
B -74+62 NM3 abr	0.005	519	163	1952	24	2.1	0.3218 ±0.09%	4.933 ±0.11%	.94	0.11120 ±0.04%	1819.0 ± 1.3
C -62 NM3 abr	0.008	412	129	3516	19	1.5	0.3227 ±0.09%	4.933 ±0.10%	.95	0.11087 ±0.03%	1813.7 ± 1.2
D -62 NM3 abr	0.010	263	83	2726	19	2.1	0.3209 ±0.09%	4.919 ±0.11%	.93	0.11116 ±0.04%	1818.5 ± 1.4

Errors are 1 std. error of mean in % except $^{207}Pb/^{206}Pb$ age errors which are 2 std. errors in Ma.

Pb* = Radiogenic Pb; NM = Non-magnetic; Abr = Abraded

[a] = sizes (-74+62) refer to length aspect of zircons in microns (i.e. through 74 micron sieve but not the 62 micron sieve)

[b] = Corrected for fractionation and spike Pb

[c] = Total common Pb in analysis in picograms

[d] = Correlation Coefficient of errors in $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$.

Group is a predominantly clastic suite of rocks that unconformably overlies volcanic rocks of the Amisk Group in the Flin Flon belt. This interpretation of equivalence is largely lithostratigraphic; the quartzofeldspathic nature of the gneisses and the presence of conglomeratic zones are comparable to the Missi Group stratigraphy. However, Missi Suite rocks are considerably more variable in composition. At the southwest end of the Adamson Lake dome, in the area of Cleunion Lake (Fig. 1), Missi Suite rocks have retained a high degree of stratigraphic preservation (Zwanzig, 1983, 1984). Rocks similar to the basal conglomerates and crossbedded meta-arkose of the Missi Group underlie an interlayered sequence of quartz phyric rhyolites, mafic volcanic flows (amphibolites of basaltic composition) and meta-arkoses that in turn are overlain by units of thinly interlayered variably mafic and/or calcic quartzofeldspathic gneisses.

The metarhyolites in the Missi Suite are well foliated, very fine grained to aphanitic and variably quartz phyric. They outcrop most prominantly at the south end of the Adamson Lake dome along the southern edge of the Kisseynew gneisses . These metarhyolites are similar in appearance to the 1832 ± 2 Ma Chickadee Lake rhyolite (Gordon et al., 1990). The Chickadee Lake rhyolite is interlayered with Missi Group rocks that occur east of Wekusko Lake, ca. 115 km farther to the east in the Flin Flon volcanic belt.

Quartz porphyry

Quartz porphyry sills outcrop in the southerly nose of the Adamson Lake dome (Fig.1). It is lithologically and compositionally similar to the Missi Suite metarhyolite, but it can be distinguished locally by its larger quartz phenocrysts and coarser grained matrix. The metarhyolite and quartz porphyry are mapped as part of the pink felsic gneiss of the Missi Suite. At the sample location on Thunderhill Lake the quartz porphyry is light grey, very fine grained, and contains 1 to 1.5 mm quartz phenocrysts. The rock exhibits a compositional layering defined by discontinuous biotite-rich laminations, microcline- plagioclase quartz layers and plagioclase-quartz layers. The matrix quartz in the compositional layers most commonly occurs in thin discontinuous planar lenses. Many biotite grains exhibit planar alignment at a high angle to compositional layers. Biotite in the foliation planes is finer grained than biotite grains concentrated along the laminations. Muscovite is concentrated most commonly in the microcline-bearing layers and grains are randomly oriented or interstitial. An average mode of the quartz porphyry is plagioclase (41%), quartz (34%), microcline (15%), biotite (6%), muscovite (2%), magnetite (1%), garnet(1%) and trace amounts of calcium carbonate, apatite and metamict monazite. The average grain size ranges from 0.2 to 0.6 mm.

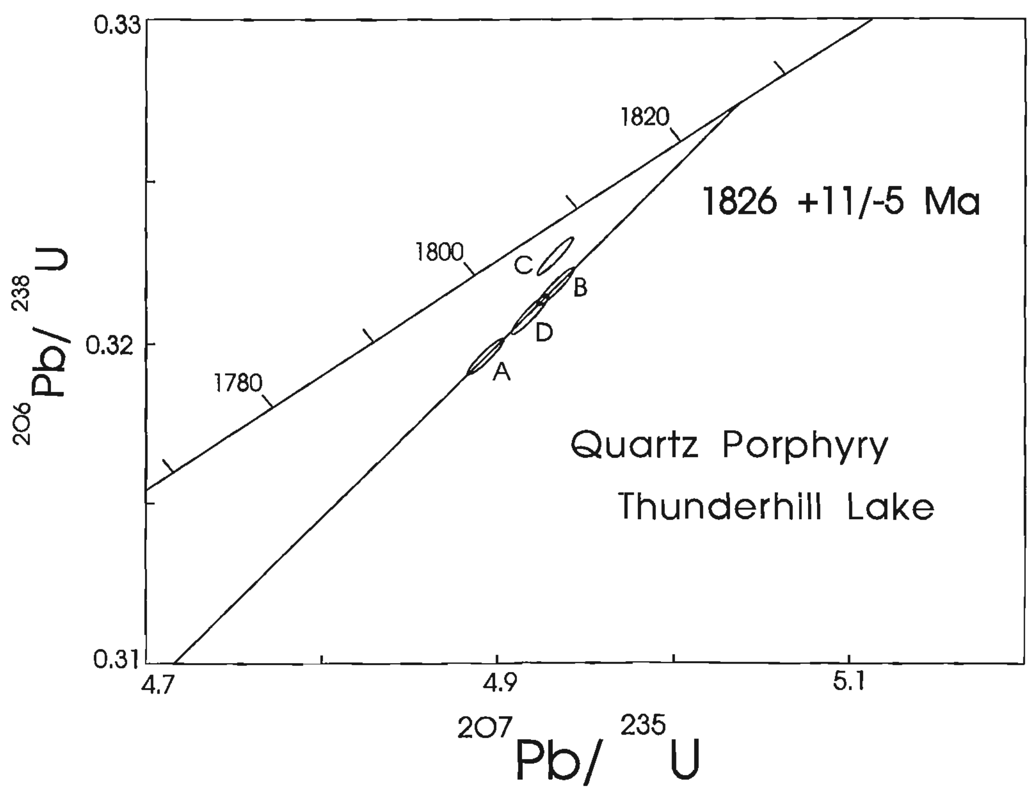

Figure 2. U-Pb concordia plot of zircon from the quartz porphyry.

U-Pb ZIRCON GEOCHRONOLOGY

Analytical methods

U-Pb analytical methods follow those outlined in Parrish et al. (1987). Techniques utilized include: strong air abrasion on all zircon fractions and crystals (Krogh, 1982), dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and assessment of errors by numerical error propagation (Roddick, 1987). Analytial results are presented in Table 1.

RESULTS

Thunderhill Lake

Zircons were separated from a fine grained, foliated, quartzofeldspathic rock, believed to be either a rhyolite ash flow or a subvolcanic intrusive rock. The grains were elongate in shape with a L:B ratio of 3:1 with rounded to blunt terminations. Most grains had black inclusions and frequent fractures, but were generally fairly clear. They ranged in size from 74 to 62μ. All fractions were moderately abraded.

Four zircon fractions were analyzed (Table 1). Three fractions are collinear and the fourth plots slightly above the array (Fig. 2). A modified York II regression of the three collinear points yields an upper intercept of 1826 +11/-5 Ma and a lower intercept of 616 Ma with a MSWD of 0.23. Fraction C is one of the smaller fractions (-62μ) and because of the size, may have had a slightly different lead loss history, and thus plots above the array. The upper intercept of 1826 Ma is regarded as a good approximation for the time of crystallization.

DISCUSSION

The age of the quartz porphyry (1826 +11/-5 Ma) constrains the age of volcanism and deposition of the Missi Suite rocks of the Kisseynew gneisses at Thunderhill Lake. The deposition of the Missi Group and the youngest volcanic activity in the Flin Flon belt at the Wekusko Lake is similarily constrained by the 1832 Ma date of the Chickadee Lake rhyolite. Both these ages correspond to an igneous event defined at 1.83 Ga (Gordon et al., 1990). This event resulted in volcanic activity in the Flin Flon belt in the vicinity of Snow Lake and in the Kisseynew gneiss belt around Thunderhill Lake. The apparent paucity of these felsic volcanic rock types in the remainder of the south flank of the Kisseynew gneiss belt may be due in part to their being obscured by intense recrystallization, or it may indicate a restricted occurrence for Missi Group/ Missi Suite volcanism.

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Age constraints on the evolution of the Flin Flon volcanic belt and Kisseynew gneiss belt, Saskatchewan and Manitoba

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Ashton, K.E., Hunt, P.A., and Froese, E., 1992: Age constraints on the evolution of the Flin Flon volcanic belt and Kisseynew gneiss belt, Saskatchewan and Manitoba; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 55-69.

Abstract

The granodioritic to dioritic Neagle Lake Pluton is part of a 1875-1835 Ma intrusive suite emplaced in 1886 ± 2 Ma Amisk Group rocks of the Flin Flon volcanic belt. An 1846 +14/-6 Ma titanite date is taken as a minimum but reasonable age for crystallization and can also be considered a maximum for D3 regional deformation and the peak of metamorphism.

A feldspar-phyric tonalite intruding the Missi Group rocks, which unconformably overlie the Amisk Group, yields a titanite age of 1830 +10/-8 Ma and is interpreted as a shallow intrusive equivalent of the Missi volcanic suite.

The southern flank of the Kisseynew gneiss belt is interpreted as the high-grade equivalent of the Flin Flon volcanic belt. Quartz-rich gneisses previously assigned to the Sherridon Group and considered sedimentary are now thought to be high-grade Amisk felsic volcanic rocks. Samples collected from the Sherridon area yield titanite dates of 1808 ± 2 and 1804 ± 3 Ma, establishing a minimum age for the peak of metamorphism in the southern flank of the Kisseynew gneiss belt and suggesting that subsequent uplift had resulted in cooling to about 550-600°C by that time.

The western boundary of the Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt is defined by the southwest-verging Sturgeon-weir Shear Zone along which the two belts have been obliquely thrust over the Hanson Lake block. Monazite and zircon dates from sheared garnetiferous pegmatites and later undeformed pegmatite dykes constrain the age of the Sturgeon-weir Shear Zone to between 1806 ± 2 and 1767 ± 1 Ma.

Résumé

Le pluton de Neagle Lake variant de granodioritique à dioritique fait partie d'une suite intrusive dont l'âge varie de 1875 à 1835 Ma mise en place en 1886 ± 2 Ma dans les roches du groupe d'Amisk de la zone volcanique de Flin Flon. La datation à 1846 +14/-6 Ma de la titanite permet d'établir l'âge minimal mais raisonnable de la cristallisation et elle peut également correspondre à l'âge maximal de la déformation régionale D3 et du métamorphisme maximal.

Une tonalite porphyrique feldspathique, recoupant par intrusion les roches du groupe de Missi qui reposent en discordance sur le groupe d'Amisk, a été datée par analyse de la titanite à 1830 +10/-8 Ma, est interprétée comme un équivalent intrusif peu profond de la suite volcanique de Missi.

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Le flanc sud de la zone gneissique de Kisseynew est interprété comme l'équivalent fortement métamorphisé de la zone volcanique de Flin Flon. Les gneiss riches en quartz, auparavant associés au Groupe de Sherridon et considérés comme sédimentaires, sont maintenant considerés comme des volcanites felsiques d'Amisk, à degré de métamorphisme élevé. Les échantillons prélevés dans la région de Sherridon contiennent des cristaux de titanite qui ont été datés à 1 808 ± 2 Ma et 1 804 ± 3 Ma; cette datation a permis d'assigner un âge minimal au métamorphisme le plus intense survenu dans le flanc sud de la zone gneissique de Kisseynew et de déduire que le soulèvement subséquent était à l'origine du refroidissement jusqu'à des températures d'environ 550-600°C à ce moment.

La bordure ouest de la zone volcanique de Flin Flon et le flanc sud de la zone gneissique de Kisseynew sont délimitées par la zone de cisaillement de Sturgeon-Weir à vergeance sud-ouest, le long de laquelle les deux zones ont subi un chevauchement oblique sur le bloc de Hanson Lake. La datation de cristaux de monazite et de zircon, provenant de pegmatites cisaillées à grenat et de dykes de pegmatite non déformés qui ont fait intrusion par la suite, a donné la possibilité de circonscrire l'âge de la zone de cisaillement de Sturgeon-Weir entre 1 806 ± 2 Ma et 1 767 ± 1 Ma.

INTRODUCTION

The Flin Flon volcanic belt of the Trans-Hudson Orogen (Fig. 1) consists of Amisk Group (Bruce, 1918) mafic to felsic volcanic and volcaniclastic rocks, shallow sub-volcanic intrusions and argillaceous sedimentary rocks. An 1886 ± 2 Ma U-Pb zircon age determination (Gordon et al., 1990) from a rhyolite crystal tuff in the Flin Flon area represents the only direct indication of the age of the Amisk Group.

The Amisk Group is intruded by an extensive suite of 1860-1835 Ma granodiorite-quartz diorite plutons (Gordon et al., 1990; Blair et al., 1988; Ansdell and Kyser, 1991) and is

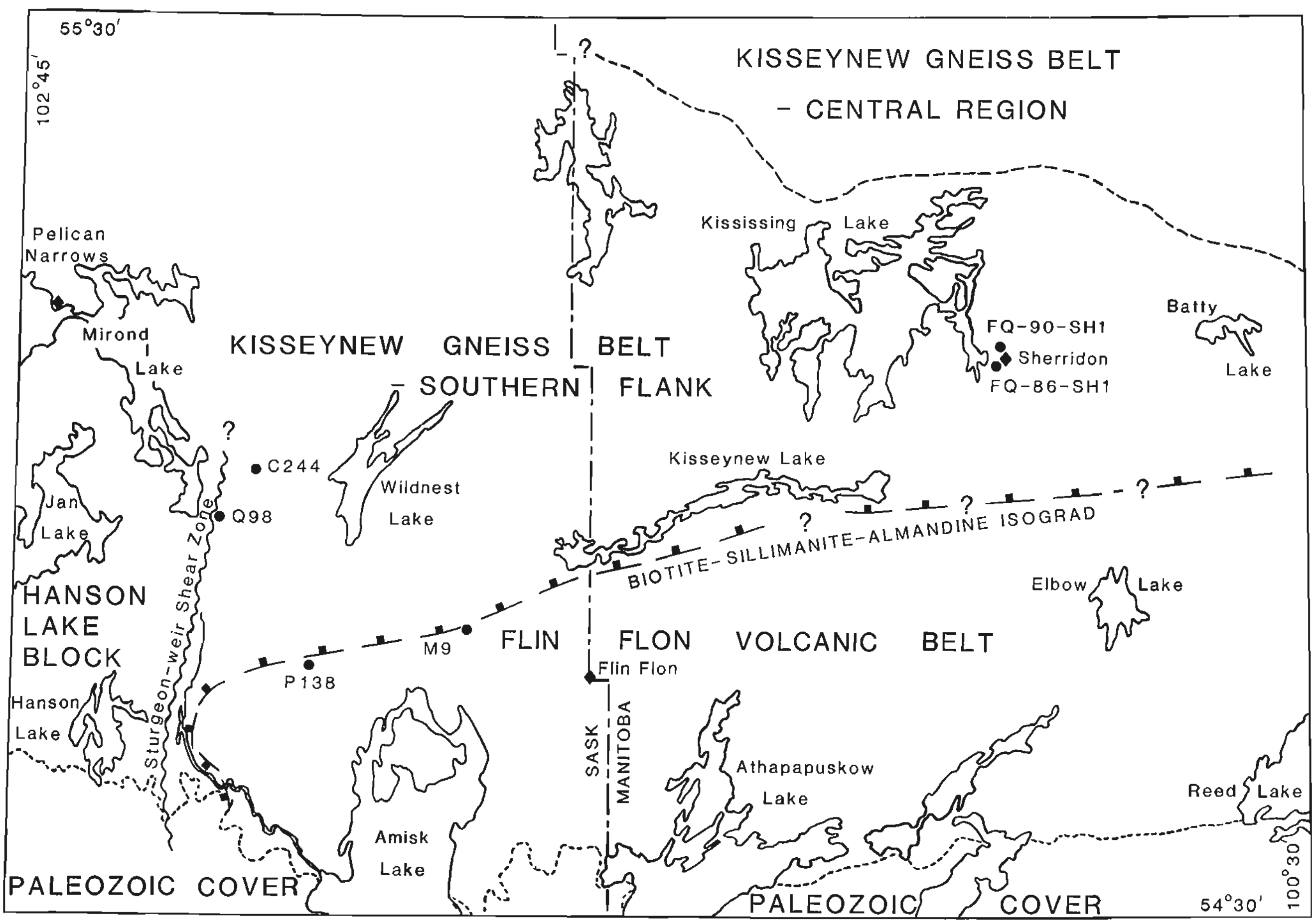

Figure 1. Map of the western Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt showing sample locations.

unconformably overlain by a sequence of polymictic conglomerates and arenaceous sedimentary rocks termed the Missi Group (Bruce, 1918; Wright and Stockwell, 1935).

A heterogeneous assemblage of high-grade rocks north of the Flin Flon volcanic belt (Fig. 1) was designated as Kisseynew gneiss by Bruce (1918). These rocks represent metamorphosed equivalents of the low-grade rocks of the Flin Flon volcanic belt (Wright and Stockwell, 1935; Byers and Dahlstrom, 1954; Bailes, 1980; Ashton et al., 1987), although they include a significantly larger sedimentary component.

North of the heterogeneous assemblage is a vast area of homogeneous gneisses derived from greywacke and shale. Harrison (1951) extended the use of the term Kisseynew to include this area. Recently, the two subdivisions of the Kisseynew gneiss belt have been referred to as the southern flank and the central region (Gordon et al., 1990) or as the south flank and the central migmatite belt (Zwanzig, 1990).

In recent work in the southern flank of the Kisseynew gneisses, the Amisk and Missi Group stratigraphic subdivisions have been used (Ashton et al., 1987; Ashton, 1989a,b; Schledewitz, 1988; Zwanzig et al., 1988; Zwanzig, 1990). A pre-Missi orthogneiss within the southern flank yielded an age of 1873 ± 4 Ma (Hunt and Zwanzig, 1990).

One way to define the boundary between the southern flank of the Kisseynew gneiss belt and the Flin Flon volcanic belt is to use an isograd (Froese and Moore, 1980; Zwanzig, 1990). A convenient mineralogical marker is the disappearance of staurolite in the presence of muscovite according to the reaction:

$$\text{staurolite} + \text{muscovite} + \text{quartz} = \text{biotite} + \text{sillimanite} + \text{almandine} + H_2O$$

Table 1. Inferred geological history of the Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt.

Age ± Error	(Ma)	Rock Analyzed	Event
			D6; Sinistral N-S faults
1767 ± 1	m T	Late pegmatite	
			D5; SwSZ & related SE-plunging folds
1804 ± 3	t T	Felsic gneiss FQ-86-SH1	D4; Titanite closure, southern flank of KGB, SW-verging
1808 ± 2	t T	Felsic gneiss FQ-90-SH1	folds & reactivation of shear zones
1804 ± 2	m G	Granodiorite	Monazite closure, central region, KGB
1806 ± 2	m G	Granite	
1806 ± 2	z T	Deformed pegmatite	
1814 +17/-11	z G	Granodiorite	Peak of metamorphism, central KGB
1816 +23/-12	z G	Granite	End of D3 deformation
			End of Missi deposition
1826 +11/-5	z C	Pink felsic gneiss	
1830 +10/-8	t T	Feldspar porphyry	
1832 ± 2	z G	Chickadee Lake rhyolite, Missi Group, Wakusko Lake	
1841 ± 18	s A	Feldspar porphyry *	
1834 ± 13	s A	Phantom Lake dyke	
1840 ± 7	s A	Phantom Lake granite	
1842 ± 13	s A	Boot Lake granodiorite	
1837 ± 5	s A	Neagle Lake granodiorite	D3; Onset of tight-isoclinal folding; development of
1846 +14/-6	t T	Neagle Lake quartz diorite	regional foliation
			D2; Isoclinal E-trending folds (no fabric)
			Onset of Missi deposition?
1845 +10/-8	z B	Cormorant pluton	
1847 ± 4	z S	Lynx Lake granodiorite	
1848 ± 11	s A	Missi Island trondhjemite	D1; Tight-isoclinal folds & onset of metamorphism
1853 ± 8	s A	Reynard Lake granite	
1860 ± 6	s A	Annabel Lake granodiorite	
1873 ± 4	z H	Puffy Lake granite	
1886 ± 2	z G	Amisk rhyolite tuff	Amisk volcanism & onset of Amisk sedimentation

(m - monazite, z - conventional U/Pb techniques on zircon, s - single-zircon Pb-evaporation technique, t - titanite; T - this study, C - Hunt and Schledewitz, this volume, A - Ansdell and Kyser, 1990, B - Blair et al., 1988, G - Gordon et al., 1990, H - Hunt and Zwanzig, 1990; * - likely affected by inheritance; N-S - north-south, SE - southeast, SW - southwest, E - east, SwSZ - Sturgeon-weir Shear Zone, KGB - Kisseynew gneiss belt).

This isograd is fairly well established in the Amisk Lake area (Ashton, 1990; Ashton, unpublished data) but is poorly constrained to the east due to the scarcity of rocks of suitable composition (Fig. 1).

The western boundary of the Flin Flon volcanic belt and parts of the Kisseynew gneiss belt is defined by the Sturgeon-weir Shear Zone (MacQuarrie, 1979; Ashton et al., 1987; Ashton and Wilcox, 1988). It is an east-dipping, southwest-verging structure along which rocks of the Flin Flon volcanic belt and Kisseynew gneiss belt have been obliquely thrust over the Hanson Lake block.

Six samples from the Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt were analyzed during this study to further constrain the ages of: 1) early deformation, 2) deposition of the Missi Group, 3) peak metamorphism and 4) the Sturgeon-weir Shear Zone. Previous and new age determinations and deformational events are given in Table 1.

ANALYTICAL TECHNIQUES

U-Pb analytical methods follow those outlined by Parrish et al. (1987). Techniques utilized include: strong air abrasion of all zircon fractions (Krogh, 1982), dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987) and assessment of errors by numerical error propagation (Roddick, 1987).

NEAGLE LAKE PLUTON (FQ-87-P138)

The Neagle Lake Pluton is a multi-phase granodiorite-gabbro stock which intrudes Amisk Group rocks northwest of Amisk Lake (Pearson, 1951; Byers and Dahlstrom, 1954; Fox, 1976). Byers and Dahlstrom (1954) distinguished the Neagle Lake Pluton from other intrusive rocks in the area because they believed that it was post-tectonic and truncated the moderately to steeply west-dipping West Channel Fault

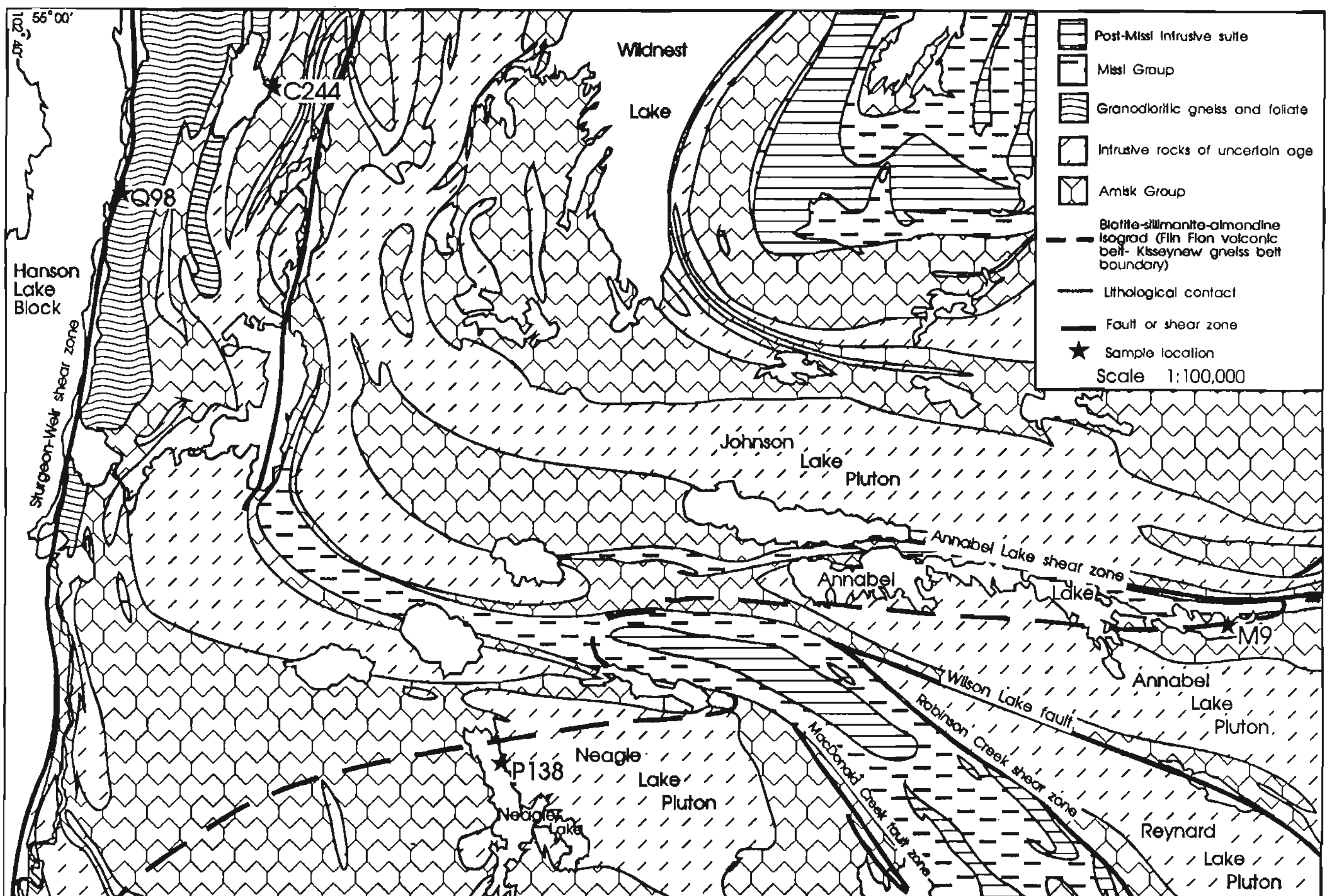

Figure 2. Simplified geological map showing the western sample locations.

Table 2. U-Pb zircon, titanite, monazite data

Fraction size[a]	Wt. mg	U ppm	Pb^* ppm	$^{206}Pb/^{204}Pb$ [b]	Pb^c pg	^{208}Pb %	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	Corr.[d] Coeff.	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$ Age (Ma)
Feldspar Porphyry (FQ-86-M9)											
A +74 NMO abr	0.024	173	55	3514	23	5.0	0.3145 ±.09%	4.896 ±.10%	.9200	0.11289 ±.04%	1846.5 ± 1.5
B +74 NMO abr	0.003	170	56	579	18	5.0	0.3292 ±.23%	5.185 ±.27%	.8195	0.11423 ±.16%	1867.8 ± 5.6
C +74 NMO abr	0.003	138	39	385	20	6.9	0.2801 ±.27%	4.003 ±.36%	.7396	0.10365 ±.24%	1690.5 ± 9.0
D -74+62 NMO abr	0.013	501	158	9160	14	3.3	0.3180 ±.08%	4.937 ±.10%	.9516	0.11262 ±.03%	1842.1 ± 1.1
TA titanite	0.144	40	15	452	248	15.9	0.3215 ±.15%	4.951 ±.26%	.7025	0.11169 ±.19%	1827.1 ± 6.9
TB titanite	0.046	36	12	252	126	13.9	0.3003 ±.39%	4.614 ±.51%	.7724	0.11144 ±.33%	1823.1 ± 11.8
TC titanite	0.205	52	17	420	443	13.8	0.2919 ±.11%	4.468 ±.30%	.6746	0.11100 ±.24%	1815.9 ± 8.7
TD titanite	0.172	52	17	390	412	14.3	0.2979 ±.13%	4.550 ±.33%	.6792	0.11078 ±.26%	1812.3 ± 9.5
Neagle Lake Pluton (FQ-87-P138)											
A -74 NMO abr	0.011	536	168	817	139	5.3	0.3100 ±.09%	4.778 ±.16%	.7410	0.11179 ±.11%	1828.7 ± 4.1
B -74 NMO abr	0.007	481	150	3141	19	6.2	0.3055 ±.09%	4.715 ±.11%	.9242	0.11194 ±.04%	1831.1 ± 1.5
C -74 NMO abr	0.009	157	45	2736	8	7.1	0.2759 ±.12%	4.280 ±.13%	.9109	0.11251 ±.05%	1840.3 ± 1.9
TA titanite	0.066	194	74	1348	185	17.6	0.3286 ±.12%	5.113 ±.15%	.8923	0.11285 ±.07%	1845.9 ± 2.4
TB titanite	0.030	110	42	337	199	16.5	0.3341 ±.13%	5.175 ±.35%	.6650	0.11236 ±.28%	1837.9 ± 10.2
TD titanite	0.037	147	53	984	108	13.9	0.3230 ±.10%	5.014 ±.15%	.8029	0.11258 ±.09%	1841.5 ± 3.4
FQ-SH90-1											
TA titanite	0.127	112	40	2289	123	13.5	0.3230 ±.09%	4.929 ±.11%	.8971	0.11069 ±.05%	1810.8 ± 1.8
TB titanite	0.121	116	43	2275	122	16.5	0.3237 ±.10%	4.933 ±.12%	.9085	0.11053 ±.05%	1808.1 ± 1.8
FQ-86-SH1											
TA titanite	0.077	175	61	185	1618	11.8	0.3221 ±.48%	4.900 ±.50%	.9633	0.11034 ±.13%	1805.1 ± 4.9
TB titanite	0.150	174	64	319	1755	15.9	0.3224 ±.29%	4.902 ±.31%	.9648	0.11028 ±.08%	1804.0 ± 2.9
Deformed Pegmatite (FQ-87-C244)											
A +140 abr	0.012	3749	1172	5510	160	.1	0.3270 ±.09%	4.986 ±.10%	.9503	0.11060 ±.03%	1809.3 ± 1.2
B -90+80 abr	0.030	4439	1352	2537	982	.2	0.3186 ±.10%	4.827 ±.12%	.9096	0.10988 ±.05%	1797.4 ± 1.8
C -170+160 abr	0.036	5806	1781	37686	104	.2	0.3207 ±.10%	4.876 ±.11%	.9700	0.11029 ±.03%	1804.1 ± 1.0
D -74+62 abr	0.008	3933	1224	6212	94	.2	0.3254 ±.09%	4.956 ±.10%	.9504	0.11046 ±.03%	1806.9 ± 1.2
E -74+62 abr	0.012	3058	941	5488	130	.2	0.3218 ±.09%	4.899 ±.10%	.9488	0.11041 ±.03%	1806.1 ± 1.2
MA monazite	0.011	11503	8946	60124	39	60.3	0.3216 ±.10%	4.879 ±.11%	.9716	0.11002 ±.03%	1799.7 ± 1.0
MB monazite	0.007	16257	7394	11834	182	32.2	0.3221 ±.10%	4.868 ±.11%	.9684	0.10961 ±.03%	1792.9 ± 1.1
Late Pegmatite (FQ-87-Q98)											
A +50 NM abr	0.007	2070	600	6232	40	5.5	0.2875 ±.08%	4.254 ±.10%	.9485	0.10733 ±.03%	1754.6 ± 1.2
B -70 NM abr	0.006	387	124	3228	13	6.7	0.3134 ±.10%	4.798 ±.11%	.9130	0.11102 ±.05%	1816.2 ± 1.7
C +50 NM abr	0.006	617	176	2842	24	5.4	0.2831 ±.09%	4.237 ±.11%	.9064	0.10856 ±.05%	1775.4 ± 1.7
D single grain abr	0.010	1758	495	11480	25	6.5	0.2763 ±.08%	4.055 ±.10%	.9568	0.10644 ±.03%	1739.3 ± 1.1
E -74+62 NM abr	0.012	912	238	10824	17	5.8	0.2590 ±.08%	3.762 ±.10%	.9568	0.10534 ±.03%	1720.3 ± 1.1
MA monazite	0.003	5857	10731	11865	26	83.4	0.3160 ±.08%	4.709 ±.10%	.9583	0.10807 ±.03%	1767.1 ± 1.1

Errors are 1 std. error of mean in % except $^{207}Pb/^{206}Pb$ age errors which are 2 std. errors in Ma.

Pb^* = Radiogenic Pb; NM = Non-magnetic; Abr = Abraded

[a] = sizes (-74+62) refer to length aspect of zircons in microns (i.e. through 74 micron sieve but not the 62 micron sieve)

[b] = Corrected for fractionation and spike Pb

[c] = Total common Pb in analysis in picograms

[d] = Correlation Coefficient of errors in $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$.

(Fig. 2). They interpreted the displacement along the West Channel Fault as oblique sinistral-reverse. However, the following observations indicate that the relationship between the Neagle Lake Pluton and the West Channel Fault is not as clear as first suggested.

1) According to the map of Byers and Dahlstrom (1954), a weak but penetrative foliation in the pluton is approximately co-planar with the main regional D3 foliation in the host supracrustal rocks (Table 1).
2) West-northwest plunging D3 and/or D4 extension and mineral lineations are also represented in the pluton.
3) Garnet porphyroblasts, which are indicative of peak metamorphic conditions in this area, overprint the D3 fabric in the host supracrustal rocks but are disaggregated in the fault zone, indicating post-D3 displacement.
4) The southeastern margin of the pluton appears offset by about 3 km in a sinistral sense along a north-striking plane. This may represent displacement along the West Channel Fault.
5) The northwestern end of the pluton appears attenuated along a west-striking plane. This is likely due to D4 deformation (Table 1) which is partly represented in this area by the MacDonald Creek Fault Zone, Robinson Creek Shear Zone, Wilson Lake Fault and Annabel Lake Shear Zone (Fig. 2). All of these display evidence of oblique sinistral> reverse shear during D4 deformation although this may represent re-activation along pre-existing structures.

These observations suggest that the Neagle Lake Pluton is not post-tectonic but was emplaced during D3 deformation.

RESULTS

The plagioclase-phyric sample was collected from the northwestern margin of the Neagle Lake Pluton (Fig. 2; UTM Zone 13, 666360E, 6073440N). It contained about 10% hornblende, 10% biotite, 15% quartz, 6% microcline and 55% plagioclase and plotted at the intersection of the granodiorite, tonalite, quartz diorite and quartz monzodiorite fields on the IUGS classification of igneous rocks (Streckeisen, 1976). Apatite , sphene, zircon and opaque minerals were present as accessory phases. Minor alteration products included epidote and carbonate.

Zircons separated from the sample consisted of small (<-74μm) stubby prisms with sharp terminations and length:base ratios of 1:1. Grains were generally clear and colourless with some internal fractures and inclusions. Internal cores were not observed and all fractions were abraded.

Three zircon and three titanite fractions were analyzed (Table 2, FQ-86-P138). A modified York (1969) regression through the three zircon fractions yields a upper intercept of 1825 ± 2 Ma and a lower intercept of -170 Ma (Fig. 3). Since a negative lower intercept is inconsistent with conventional models of Pb loss, it is likely that these zircons have some

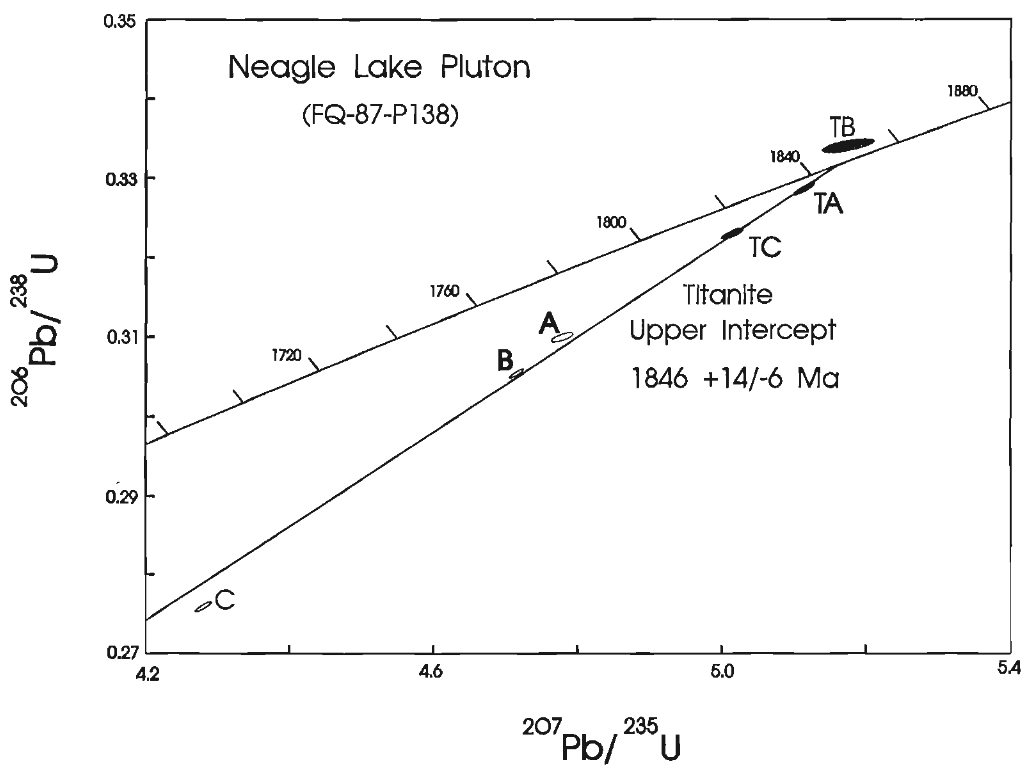

Figure 3. Concordia diagram showing U-Pb results from the Neagle Lake Pluton (FQ-87-P138).

degree of inheritance with subsequent Pb loss at different times. Therefore, the zircon regression is considered unreliable.

A modified York (1969) regression of three titanite analyses yielded a upper intercept of 1846 +14/-6 Ma, a lower intercept of 252 Ma and a MSWD of 4.4. The upper intercept age of 1846 +14/-6 Ma is interpreted as a minimum age for emplacement of the sampled phase of the Neagle Lake Pluton and as a maximum age for D3 deformation. Since minerals indicative of inferred peak metamorphic conditions overgrow D3 foliation, 1846 +14/-6 Ma can also be considered as a maximum age for the peak of metamorphism along the boundary between the Flin Flon volcanic belt and Kisseynew gneiss belt.

This result implies a somewhat older age than recent 1837 ± 5 Ma date from a granodioritic phase of the Neagle Lake Pluton (Ansdell and Kyser, 1991) derived by the single-zircon Pb-evaporation technique (Kober, 1986, 1987; Kober et al., 1989).

FELDSPAR PORPHYRY (FQ-86-M9)

Several small intrusions of a distinctive feldspar porphyry were emplaced within the Missi Group in the vicinity of the boundary between the Flin Flon volcanic belt and Kisseynew gneiss belt in the Amisk-Annabel lakes area (Wilcox, 1989, 1990a) and within Amisk wackes in the Duval Lake area of Manitoba. The porphyry generally occurs as dykes or sills less than 2 m thick but also forms small deformed plutons intruding the Missi Group north of Amisk Lake (Fig. 2). The dykes/sills are straight sided, parallel to layering within the Missi host rocks, and contain a co-planar D3 foliation (Table 1). Small feldspar porphyry plutons and large hornblende granodiorite intrusions of the 1835-1860 Ma suite have been attenuated and drag folded along the D4 Robinson Creek Shear Zone (Wilcox, 1989).

Typical samples of feldspar porphyry are homogeneous and pink with 10-15% green biotite and 10-15% white phenocrysts up to 2-4 mm. The majority are oligoclase-andesine but rare microcline phenocrysts were also noted. The estimated modal proportions of two samples indicate a tonalitic composition.

Wilcox (1990a) suggested that the feldspar porphyry was emplaced at hypabyssal depths based on: its homogeneity and lack of internal layering, the consistent abundance of phenocrysts, and the presence of Missi conglomerate xenoliths.

The feldspar porphyry was sampled to establish a minimum age for deposition of the Missi Group which has not been well constrained in the western Flin Flon volcanic belt. An intraformational Missi rhyolite from Herb Lake to the east has yielded an age of 1832 ± 2 Ma (Gordon et al., 1990).

RESULTS

The analyzed sample was taken from an irregular feldspar porphyry dike which intruded Missi conglomerate along the northeastern shore of Annabel Lake (UTM Zone 13, 689540E, 6078930N; Figs. 2, 4). Separated zircons were euhedral, with well defined terminations and length:base ratios of 2:1. Grains were strongly zoned and contained many fractures and inclusions. Most were also cracked or broken. Observed internal zoning may have been due to cores of inherited zircon.

Four zircon and four titanite fractions were analyzed (Table 2, FQ-86-M9). A modified York (1969) regression through the most concordant zircon data yields an upper intercept of 1852 Ma with a very high error, and a lower intercept of 294 Ma (not plotted). Fraction C, which contains broken tips from prismatic grains, is very discordant and has a $^{207}Pb/^{206}Pb$ age of 1691 Ma. The discordance and scatter of the zircon data have been attributed to inheritance and subsequent Pb loss making it difficult to establish a precise crystallization age for this rock.

An 1841 ± 18 Ma age derived from another occurrence of the feldspar porphyry (Ansdell and Kyser, 1990) using the single-zircon Pb-evaporation technique is similar to these results and may also be affected by a large inherited component.

A York (1969) regression of the four titanite analyses intersects concordia at 1830 +10/-8 Ma and 226 Ma, with a MSWD <1 (Fig. 5). In rocks metamorphosed to greater than 550-600°C, the estimated closure temperature of titanite (Heaman and Parrish, 1991), derived dates may reflect the age of metamorphism or cooling rather than primary crystallization. Host rocks to the feldspar porphyry occur in the vicinity of the biotite-sillimanite-almandine isograd (Fig. 1). Archibald et al. (1983) indicated that this assemblage is stable relative to staurolite + quartz in muscovite-bearing rocks at temperatures greater than 560-600°C over the 3.5-5 kbar pressure interval, making it possible that the U-Pb system in titanites of the feldspar porphyry could have been open during metamorphism. However, estimates for the peak of metamorphism elsewhere in the southern flank and core of the Kisseynew gneiss belt are significantly younger (see below) suggesting that the upper intercept age for the feldspar porphyry is related to primary crystallization.

Figure 4. Feldspar porphyry intruding the basal Missi conglomerate at the sampled outcrop.

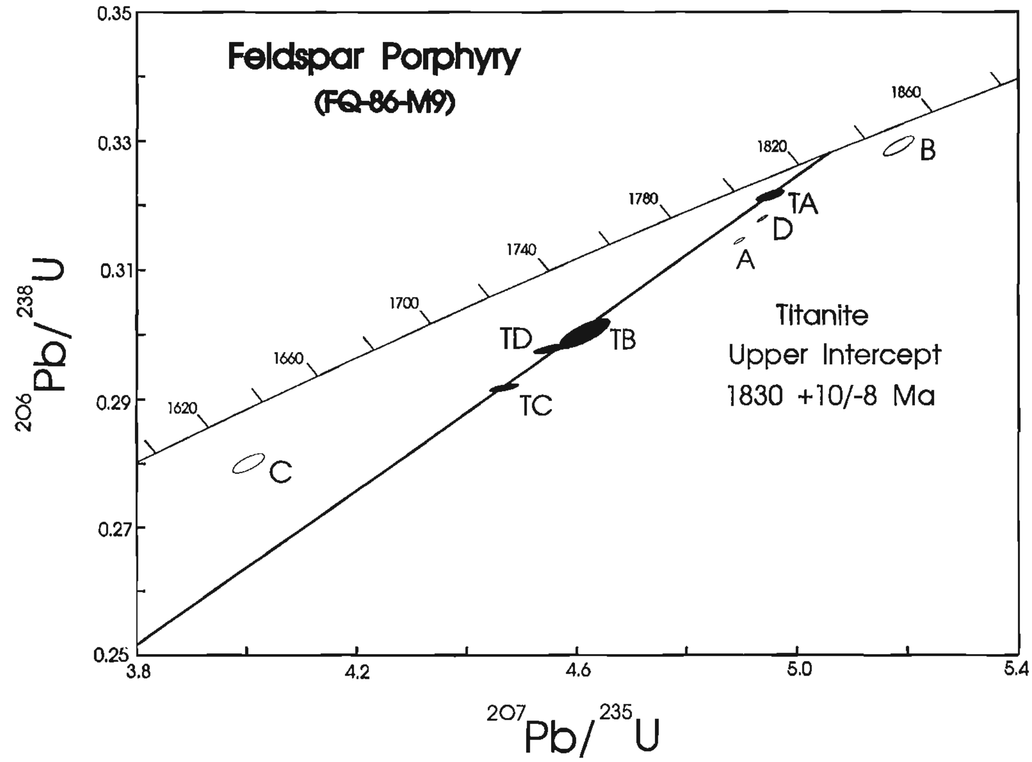

Figure 5. Concordia diagram showing U-Pb results from the feldspar porphyry (FQ-86-M9).

The 1830 +10/-8 Ma date is interpreted then, as a reasonable estimate for the emplacement of the feldspar porphyry. It also represents an additional maximum constraint on D3 deformation and a minimum age for deposition of the Missi Group in the Flin Flon area.

A unit of pink felsic gneiss, which is common in the Kisseynew Lake area (Fig. 1), is similar in appearance and setting to the feldspar porphyry. The unit was sampled for conventional U-Pb age dating at Thunderhill Lake, located 17 km south-southwest of Sherridon, where it forms a flattened lenticular body within Missi sedimentary rocks (Hunt and Schledewitz, 1992). It is described as a weakly layered rock of granodioritic composition containing minor 1-1.5 mm quartz phenocrysts and is interpreted as a Missi volcanic or volcaniclastic rock. Its upper intercept zircon age of 1826 +11/-5 Ma is very close to that of the feldspar porphyry titanite age and further indicates a widespread felsic igneous event during Missi time. This age is comparable to 1832 ± 2 Ma obtained for rhyolite interbedded with Missi sedimentary rocks at Wekusko Lake (Gordon et al., 1990).

HIGH-GRADE ROCKS FROM SHERRIDON, MANITOBA

In an attempt to subdivide the Kisseynew gneisses, a heterogeneous felsic to mafic sequence in the vicinity of Sherridon, Manitoba (Fig. 1) was assigned to the Sherridon Group (Bateman and Harrison, 1946; Froese and Goetz, 1981). Bailes (1971) suggested a correlation of the Sherridon Group with the Missi Group; however, this has been questioned in the case of the rocks at Sherridon by Ashton and Froese (1988). They proposed that the felsic gneisses were of volcanic rather than sedimentary origin and suggested that rocks of the Sherridon Group, including massive sulphide deposits and alteration zones, represented high-grade equivalents of the Amisk Group.

In order to test this model, two samples of medium grained leucocratic gneiss thought to be derived from felsic volcanic precursors were collected. Neither contained zircon but analysis of titanite has provided constraints on the timing of metamorphism. The Sherridon Group was metamorphosed to upper amphibolite facies conditions of about 660°C and 5 kbar (Froese and Goetz, 1981), well above the 550-600°C closure temperature of titanite. Therefore, U-Pb dates derived from titanite can be used to define a minimum age for the peak of metamorphism within the southern flank of the Kisseynew gneiss belt.

RESULTS

Sample FQ-86-SH1 was collected from the intersection of the road and railway in Sherridon (UTM Zone 14, 366600E, 6110150N). Sample FQ-90-SH1 was collected from an outcrop at the northern end of the tailings pond (UTM Zone 14, 365610E, 6112160N). Locations are given in Figure 6. These

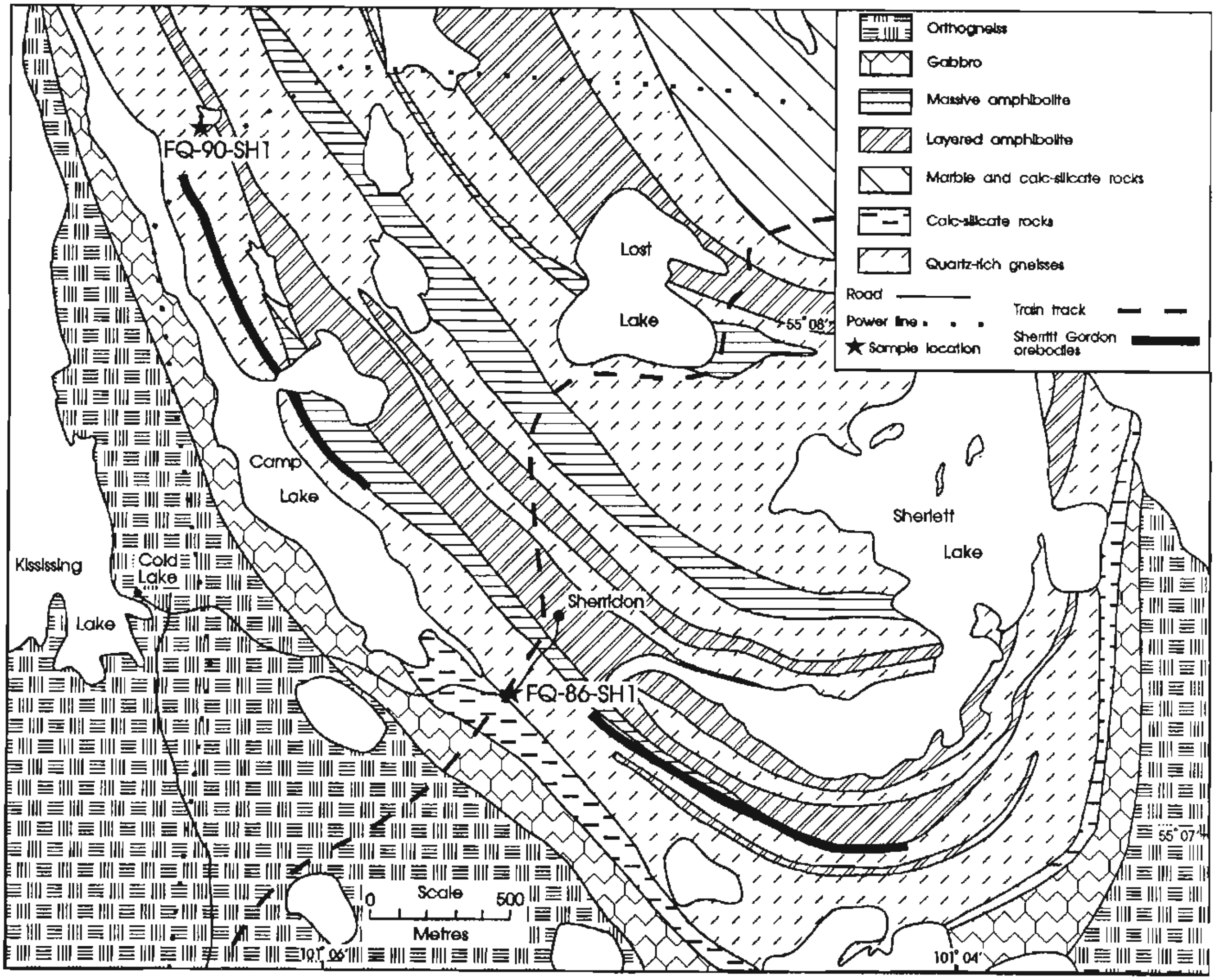

Figure 6. Simplified geological map showing sample locations in the Sherridon area.

two rocks consist mainly of quartz, plagioclase and microcline, with trace amounts of biotite, hornblende, clinopyroxene, calcite and titanite. Quartz eyes, 2 x 5 mm in size in sample FQ-90-SH1, suggest a volcanic or subvolcanic origin.

The two analyzed titanite fractions of sample FQ-86-SH1 (Table 2) yielded ^{207}Pb-^{206}Pb ages of 1804 ± 3 and 1805 ± 5 Ma, which are concordant within experimental error (Fig. 7).

Of two analyzed titanite fractions of sample FQ-90-SH1, fraction TB (Table 2) yielded a concordant ^{207}Pb-^{206}Pb age of 1808 ± 2 Ma (Fig. 8). Fraction TA was slightly discordant with a ^{207}Pb-^{206}Pb age of 1811 ± 2 Ma.

The results from both samples are taken to indicate that the Kisseynew gneisses at Sherridon cooled through the closure temperature of titanite at about 1806 Ma. This value is close to monazite ages (1806 ± 2 Ma; 1804 ± 2 Ma) obtained from intrusive rocks in the central region of the Kisseynew gneiss belt (Gordon et al. 1990).

STURGEON-WEIR SHEAR ZONE

The boundary between the Flin Flon volcanic belt and Kisseynew gneiss belt on the east and the Hanson Lake block on the west is defined by the Sturgeon-weir Shear Zone (MacQuarrie, 1979; Ashton and Wilcox, 1988), a moderately east-dipping straight zone at least 1 km wide (Fig. 1, 2). The zone occurs within locally porphyroclastic granodioritic gneiss apparently derived from deep-level Flin Flon volcanic belt intrusive and supracrustal rocks. Attenuation and rotation of all supracrustal units, plutons and pre-existing structures indicate that shearing and associated folding within the zone took place relatively late (D5) in the deformational history of the Flin Flon volcanic belt and Kisseynew gneiss belt (Table 1).

Moderately northeast-plunging stretching lineations and associated hornblende and quartz mineral lineations are best developed in the northern part of the zone. A set of mechanically independent kinematic indicators, including

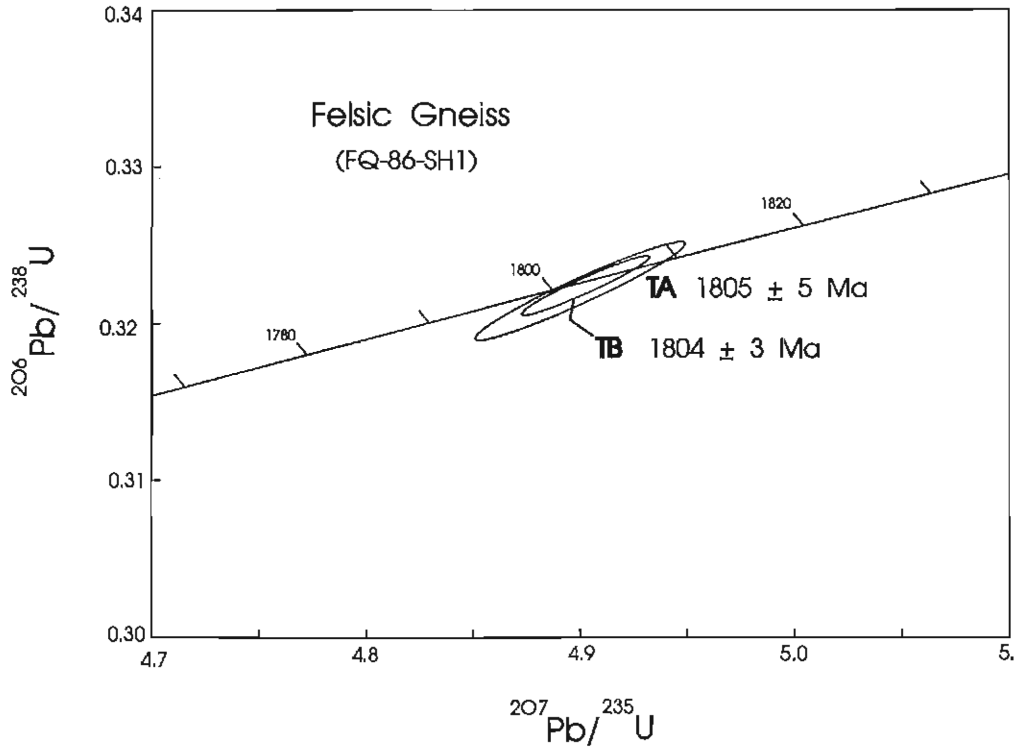

Figure 7. Concordia diagram showing results of titanite analyses from felsic gneiss FQ-86-SH1, Sherridon, Manitoba.

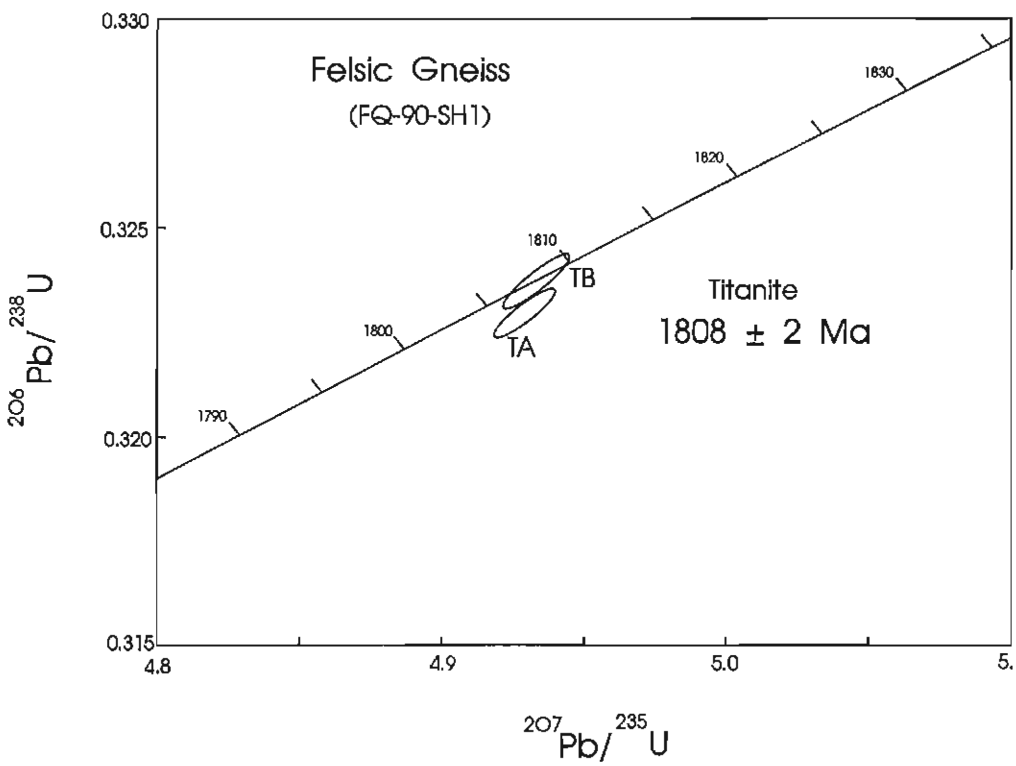

Figure 8. Concordia diagram showing results of titanite analyses from felsic gneiss FQ-90-SH1, Sherridon, Manitoba.

and s" fabrics (Berthé et al., 1979), rotated winged feldspars and granitoid segregations (Fig. 9), and asymmetric granitoid dyke pull-aparts, are consistent and indicate oblique dextral-reverse displacement. Moderately southeast-plunging open folds, developed late during the shearing event, deform the shear foliation. These fabrics confirm previous suggestions that the Flin Flon and Kisseynew belts have been thrust southwestward over the Hanson Lake block (MacQuarrie, 1979; Ashton and Wilcox, 1988).

DEFORMED PEGMATITE (FQ-87-C244)

In order to constrain the timing of the southwest-verging deformation along the Sturgeon-weir Shear Zone, two pegmatite dykes were analyzed using U-Pb techniques. A deformed pegmatite emplaced in Amisk supracrustal rocks was sampled at the eastern extent of deformation associated with the Sturgeon-weir Shear Zone (Fig. 2; UTM Zone 13, 658530E, 6095230N). The white biotite-quartz-feldspar pegmatite contains garnet up to 1 cm and minor tourmaline-quartz veining. It is about 8 m wide and is oriented slightly oblique to layering in the host intermediate volcanic and volcaniclastic rocks. The shear-induced foliation is variable in intensity, resulting in gradations from near-massive to mylonitic fabrics (Fig. 10). Coarse feldspar grains and segregations of pegmatitic material are isolated by the braided shear foliation.

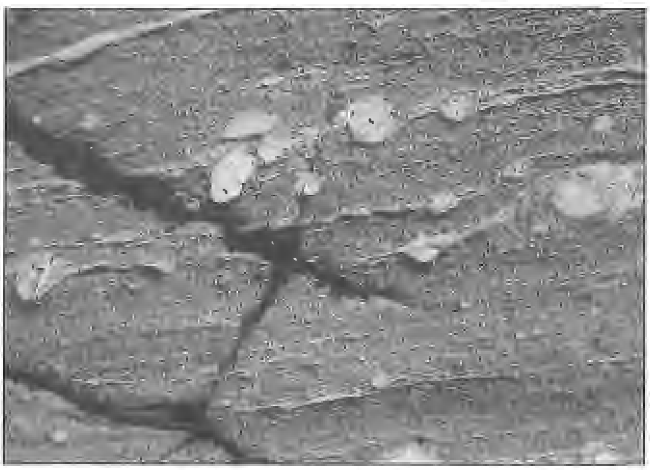

Figure 9. Rotated winged feldspars and granitoid segregations indicating a dextral component of shear in the Sturgeon-weir Shear Zone.

Figure 10. Shear-induced foliation in the deformed pegmatite sampled at the eastern extent of recognized shearing within the Sturgeon-weir Shear Zone.

The dyke contains a gently northeast-plunging stretching lineation co-linear to that observed in the main part of the Sturgeon-weir Shear Zone. A well-developed quartz foliation crosscuts the "c" shear plane and is interpreted as the "s" flattening fabric. The angular relationship between the "c and s" fabrics, together with rare shear bands, indicate southwest vergence, similar to that in the main Sturgeon-weir Shear Zone. The age of the deformed pegmatite is, therefore, thought to represent a maximum limit for the main recognized phase of deformation (D5) in the Sturgeon-weir Shear Zone.

RESULTS

Zircons separated from the deformed pegmatite consisted of euhedral, prismatic grains with slightly rounded terminations and length:base ratios of 2:1. Grains were generally clear and colourless with some having fractures and inclusions. Cores were not observed and all zircon fractions were abraded.

Five zircon and two monazite fractions were analyzed (Table 2, FQ-87-C244). Measured uranium contents in the zircons were generally very high, ranging from 3000 to 6000 ppm. Two of the zircon fractions plot above concordia and the rest are slightly discordant (Fig. 11). A modified York (1969) regression on the zircon fractions yields an upper intercept age of 1806 ± 2 Ma, a lower intercept of 543 Ma and a MSWD of 11. Two monazite grains, one slightly above concordia and one slightly below, yield ^{207}Pb-^{206}Pb ages of 1793 and 1800 Ma respectively.

The upper intercept of 1806 ± 2 Ma is believed to be a good estimate of the crystallization age of the pegmatite. Two zircon fractions plotting slightly above concordia may be due to their high U content (4000 ppm) and laboratory abrasion. As a result, there may have been a minor redistribution of radiogenic Pb from rim regions richer in U towards the core, that is subsequently removed by abrasion. The slightly younger monazite dates are thought to represent minimum ages for crystallization due to the problem of excess ^{206}Pb during monazite crystallization (Heaman and Parrish, 1991); they may have also lost Pb during subsequent metamorphism.

LATE PEGMATITE (FQ-87-Q98)

A minimum age constraint on the ductile component of shear along the Sturgeon-weir Shear Zone is represented by a late unfoliated biotite-quartz-feldspar pegmatite which intrudes well foliated grey hornblende orthogneisses and interlayered medium grained mafic rocks along the eastern side of the Sturgeon-weir River (Fig. 2; UTM Zone 13, 6091600E, 653620N). Many of

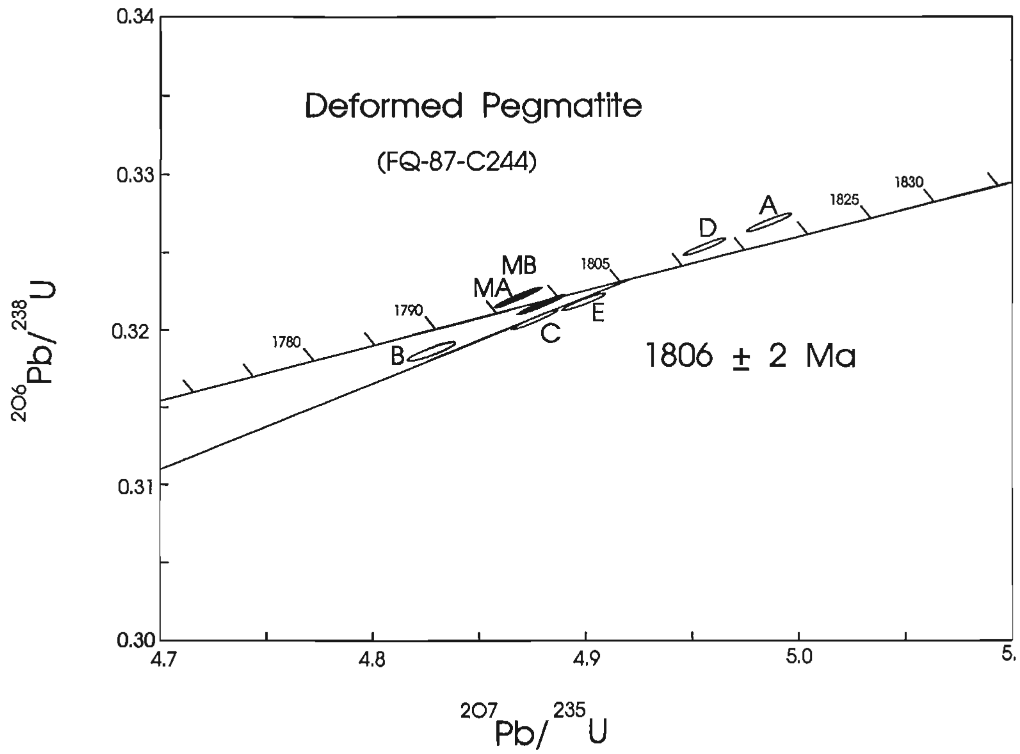

Figure 11. Concordia diagram showing U-Pb results from the deformed pegmatite (FQ-87-C244).

the host rocks contain up to 10% rounded white feldspar ± quartz porphyroclasts up to 2 cm. The leucosome component of the host gneiss tends to be finer grained and locally contains ribbon quartz.

The late pegmatite crosscuts the sheared host rocks but has been affected by late north-striking brittle fractures which elsewhere display predominantly sinistral apparent displacement (Byers and Dahlstrom, 1954; Ashton et al., 1987). The pegmatite varies from podiform to dyke-like in nature and is oriented at about 340°, parallel to numerous, locally beryliferous pegmatite dikes to the west in the Hanson Lake block (Radcliffe, 1964; Pyke, 1966; MacDougall, 1989). The orientation of the dyke set is tentatively attributed to late relaxation along the southwest-verging Sturgeon-weir Shear Zone.

RESULTS

The sampled biotite-quartz-feldspar pegmatite is pink, massive and zoned with medium grained aplitic margins and a pegmatitic core. Quartz is locally aligned perpendicular to the dyke margins.

Separated zircons were generally of very poor quality due in part to high uranium concentrations (300 to 2000 ppm). The grains were euhedral with length:base ratios between 2:1 and 3:1, and ranged in size from 105 to 62μm. All were dark with extensive fractures, inclusions and strong zoning. Cores were not recognized. All zircon fractions were abraded until completely rounded.

Five zircon fractions including a single grain were analyzed (Table 2, FQ-87-Q98). The results are quite discordant and form a scattered non-linear array (Fig. 12). A modified York (1969) regression through all five analyses yields an upper intercept age of 1823 +53/-36 Ma. Fractions B and C plot below the discordia line and most likely represent an inherited component. A York regression through the other three zircon fractions (A, D and E) yields an upper intercept age of 1783 +11/-10 Ma with a lower intercept of 462 Ma and a MSWD of 17. This is considered a better estimate for the age of crystallization of the pegmatite but the large error still suggests a component of inheritance.

A single monazite grain was analyzed and yielded a concordant ^{207}Pb-^{206}Pb date of 1767 ± 1 Ma. This represents a minimum age for the pegmatite due to the problem of excess ^{206}Pb during monazite crystallization (Heaman anmd Parrish, 1991), so the pegmatite was most likely emplaced between 1767 ± 1 and 1783 +11/-10 Ma ago. This is consistent with 1762-1773 Ma zircon and monazite ages from a number of other post-tectonic, medium-grained to pegmatitic granite dikes extending from the La Ronge Belt to the Churchill-Superior Boundary (Krogh et al., 1985; Bickford et al., 1987; Machado et al., 1987; Chiarenzelli and Lewry, 1988; Chiarenzelli, 1989).

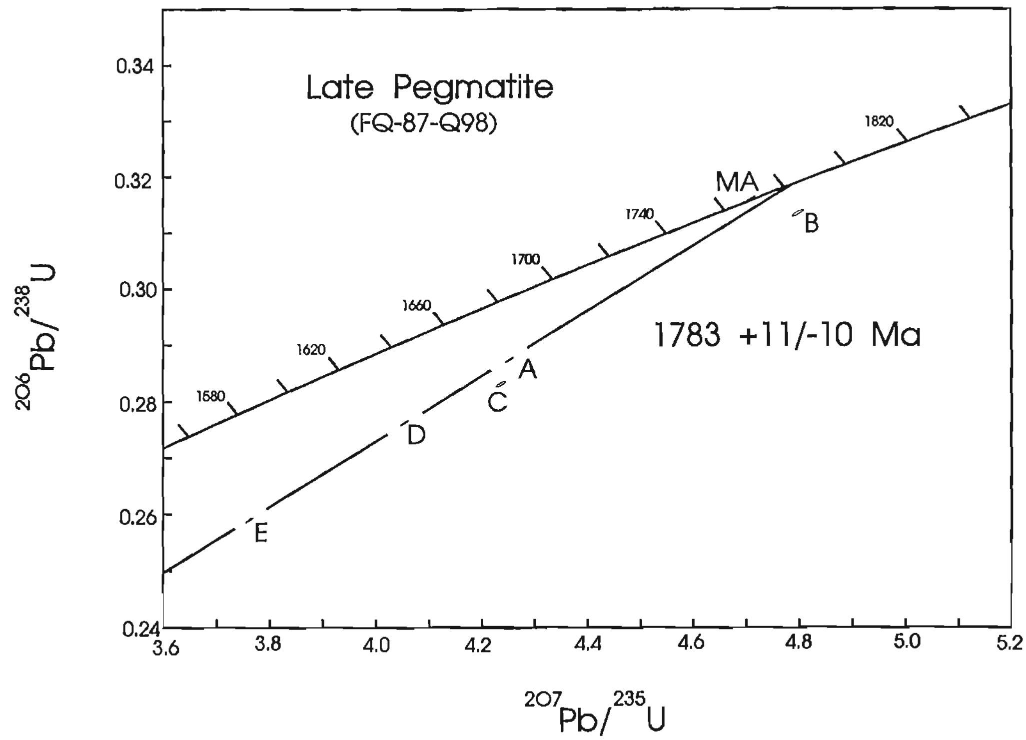

Figure 12. Concordia diagram showing U-Pb results from the late pegmatite (FQ-87-Q98).

SUMMARY AND DISCUSSION

U-Pb geochronological results obtained from six samples help to better constrain the ages of deposition, deformation and metamorphism of rocks within the Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt (Table 1). Development of the Amisk volcanic suite at ca. 1886 ± 2 Ma (Gordon et al., 1990) and Amisk sedimentary rocks was followed by major dioritic to granitic intrusive activity between about 1875 and 1835 Ma (Gordon et al., 1990; Blair et al., 1988; Hunt and Zwanzig, 1990; Ansdell and Kyser, 1991). The Neagle Lake Pluton, containing 1837 ± 5 Ma granodioritic (Ansdell and Kyser, 1991) and >1846 +14/-6 Ma quartz dioritic phases, was emplaced in Amisk supracrustal rocks towards the end of this igneous event. This massive injection of granitoid material was accompanied by early deformation (D1), the onset of regional metamorphism, and uplift, which in turn, led to the development of increased topographic relief and molasse deposition. Detritus was at first derived principally from the supracrustal rocks, forming polymictic conglomerates dominated by mafic volcanic clasts in the Flin Flon volcanic belt and a mixture of volcanic and volcaniclastic material in the southern flank of the Kisseynew gneiss belt. Continued erosion is thought to have led to unroofing of the extensive granitoid plutons, which contributed to the more arenaceous component of the Missi Group. Minor igneous activity during deposition of the Missi Group is marked by 1826 +11/-5 Ma pink felsic gneisses of the Kisseynew Lake – Thunderhill Lake area (Schledewitz and Hunt, 1992), 1830 +10/-8 Ma intrusive feldspar porphyries in the Amisk Lake – Annabel Lake area, and 1832 ± 2 Ma at Wekusko Lake (Gordon et al., 1990).

Deposition of the Missi Group is further constrained by D1 and D2 deformation (Table 1) which are interpreted as representing the early stages in the development of south- and/or southwest-verging nappes. This major thermo-tectonic event culminated during D3 with the formation of the main regional foliation. Peak metamorphic conditions in the southern flank of the Kisseynew gneiss belt were attained following D3 and did not likely exceed about 660°C (Froese and Goetz, 1981). The exception is the area adjacent to the Sturgeon-weir Shear Zone where conditions likely exceeded 7 kb and 700°C (unpublished data). Conditions had cooled to 550-600°C, the closure temperature of titanite, by about 1806 Ma.

Continued deformation resulted in the development of southwest-verging D4 and D5 structures.

Early shear zones in the Flin Flon volcanic belt continued to be active during D4-5 time and are marked by greenschist facies retrograde assemblages. In the Kisseynew gneiss belt, which remained hot and plastic for some time after the peak metamorphic conditions were attained, D4 and D5 are characterized by the development of regional fold interference patterns. Pre- to syn-D4-5 shear zones were largely recrystallized but continued northeast-southwest compression produced or re-activated the southwest-verging Sturgeon-weir Shear Zone (D5) between the Flin Flon volcanic belt and Kisseynew gneiss belt on the east and the Hanson Lake block on the west. Shearing in the zone deformed 1806 ± 2 Ma pegmatites and displaced metamorphic isograds. Undeformed northeast-dipping 1767 ± 1 Ma pegmatite dykes provide a minimum age for the plastic component of strain and may have been intruded during a period of relaxation following the southwest-directed convergence.

Minor sinistral brittle faulting postdates the late pegmatites and can also be observed in the Mari Lake and Granite Lake areas (Ashton, 1987; Ashton et al., 1987) as well as farther west along the Tabbernor Fault Zone (Wilcox, 1990b).

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U-Pb zircon age of the mid-crustal tonalite gneiss in the central Kapuskasing uplift, northern Ontario[1]

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Sullivan, R.W. and Leclair, A.D., 1992: U-Pb zircon age of the mid-crustal tonalite gneiss in the central Kapuskasing uplift, northern Ontario; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 71-78.

Abstract

Conventional U-Pb zircon geochronology yielded three concordant ages between 2722 and 2728 Ma from which an age of 2725 ±5 Ma is taken as the best estimate of the crystallization age of the Kapuskasing tonalite gneiss. These concordant ages are considered geologically real, explained by zircon formation and resorption events over a period of about 6 Ma. The U-Pb age is confirmed by a whole-grain lead evaporation ^{207}Pb-^{206}Pb result of 2724 ± 3 Ma. The dated sample also contains a small quantity of morphologically different, younger (2690 ±2 Ma) z ircons. These are interpreted as resulting from an in situ melt, producing observed white tonalitic veinlets. These veinlets are folded, implying deformation is, in part, post 2690 Ma. The 2725 Ma age for the tonalite gneiss, together with previously reported ages from the Kapuskasing uplift, suggests protracted and episodic tonalitic magmatism, possibly cogenetic with volcanic activity.

Résumé

Des datations U-Pb sur zircons ont donné trois âges concordants entre 2722 et 2728 Ma, dont un âge de 2725 ± 5 Ma est utilisé pour dater la cristallisation du gneiss tonalitique de Kapuskasing. Ces âges concordants sont considérés géologiquement réels, comme en témoignent la formation de zircon et les événements de résorption survenus pendant une période d'environ 6 Ma. L'âge U-Pb est confirmé par les données $^{207}Pb/^{206}Pb$ d'évaporation de plomb de gros grain qui donnent un âge de 2724 ± 3 Ma. L'échantillon daté contient également une petite quantité de zircons morphologiquement différents et plus récents (2690 ± 2 Ma). Ce fait est interprété comme le résultat d'une fusion in situ, ayant produit les veinules tonalitiques blanches observées. Ces veinules sont plissées, laissant supposer que la déformation est, en partie, postérieure à 2690 Ma. L'âge de 2725 Ma obtenu pour le gneiss tonalitique ainsi que les âges établis antérieurement dans le soulèvement de Kapuskasing indiquent un magmatisme tonalitique prolongé et épisodique, probablement cogénétique à l'activité volcanique.

INTRODUCTION

The Kapuskasing uplift of the south-central Superior Province includes a large amount of heterogeneous and xenolithic tonalite gneisses which are thought to represent a mid-crustal megalayer that occurs between overlying metavolcanic-metasedimentary sequences and subjacent granulite-facies gneisses (Percival and Card, 1983, 1985). Whereas U-Pb age results have been reported for high- and low-level rock types in the Kapuskasing uplift (see Leclair and Sullivan, 1991; and references therein) there is only a limited geochronological database on the vast expanse of Kapuskasing tonalite gneiss.

U-Pb zircon ages from several rock types in the Kapuskasing uplift and contiguous greenstone-granite terranes of the Wawa Subprovince have provided a regional chronological framework for the main geological events responsible for the formation of the Archean crust in this region. A recent geochronological study in the Kapuskasing

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uplift by Leclair and Sullivan (1991) added to the understanding of the ages of tonalitic magmatism, late- to post-tectonic plutonism, regional deformation, high-grade metamorphism and provided constraints on the timing of ductile shearing within the Puskuta Lake shear zone. This paper reports additional U-Pb results that complement preliminary findings of Leclair and Sullivan (1991), sample 4, and further constrain the age of tonalite gneiss in the central Kapuskasing uplift. The relevance of the age within the regional time framework is discussed.

GEOLOGICAL SETTING

The general geology of the central Kapuskasing uplift (Fig. 1) is described in Bennett et al. (1967), Thurston et al. (1977), Percival (1985), Leclair and Sullivan (1991; and references therein), and the main igneous and deformational events are outlined in a generalized chronological scheme in Leclair (1990). In general, five composite suites of intrusive rocks and at least five phases of deformation can be recognized in the Kapuskasing area. Tonalitic rocks represent the earliest intrusive igneous activity in the study area. Tonalites contain numerous mafic xenoliths which may be fragments of nearby volcanic belts or xenoliths inherited from the source. The emplacement of tonalite sheets at mid-crustal levels (0.5-0.7 GPa) and the formation of a tonalitic leucosome phase in high-grade gneisses are inferred to be roughly coeval with regional metamorphism and to precede, or coincide with, major deformation (Leclair and Poirier, 1989; Leclair 1990). The gneissossity in tonalite is part of the main foliation (S_1) recognized as the oldest structure. This foliation is folded by mesoscopic, tight to open F_2 folds and by generally broad, megascopic F_3 folds. Emplacement of a voluminous suite of massive to locally foliated granitoid plutons closely followed these three phases of deformation.

ANALYTICAL METHODS AND SAMPLE DESCRIPTION

Analytical methods

Zircon fractions were extracted from a crushed rock sample (~25 kg) by conventional Wilfley table, heavy liquid and Frantz magnetic separation techniques. Most zircon fractions were strongly air abraded (Krogh, 1982), but some were not abraded. Carefully selected single, broken single, and multigrain fractions were dissolved and analyzed using procedures outlined in Parrish et al. (1987) and Parrish (1987). Analytical blanks for zircon were typically 2 and 12 pg for U and Pb respectively. U-Pb analytical data are presented in Table 1 and displayed on a concordia diagram in Figure 3. Age calculations and regressions of data sets are as described in Parrish et al. (1987). Errors were calculated using numerical error propagation (Roddick, 1987). All quoted age errors are at the two sigma level (95% confidence interval).

Whole-grain ^{207}Pb-^{206}Pb evaporation analyses, presented in Table 2, were performed by Hazel Chapman and Chris Roddick at the Geochronology Laboratories at the Geological Survey of Canada using refined techniques, modified after Kober (1986, 1987) and Kober et al. (1989).

Kapuskasing tonalite gneiss

Tonalite gneiss is the most abundant rock type in the Kapuskasing area. It engulfs units of supracrustal rocks and appears to have concordant contacts with high-grade gneisses. Tonalite intrusions, now gneissic, predate emplacement of the voluminous suite of massive granitoid

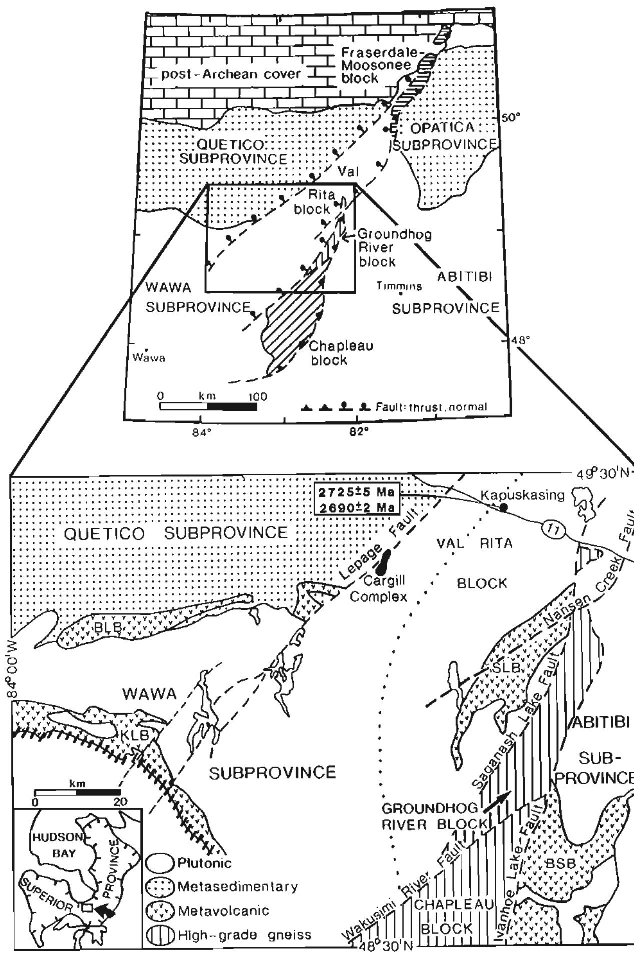

Figure 1. Generalized geological map of the Kapuskasing area showing distribution of belt-like subprovinces and major structural elements of the central Superior Province (after Card and Ciesielski, 1986; Percival and McGrath, 1986). Top map and inset show the location of the Kapuskasing area in the Superior Province. The location of the dated sample and the resulting ages are shown. The Kapuskasing uplift includes the Chapleau, Groundhog River, Fraserdale-Moosonee, and Val Rita blocks. Heavy diagonal dashes represent approximate extent of the Puskuta Lake shear zone. Dotted line is the axis of arcuate gravity and aeromagnetic anomalies of the Val Rita block. Metavolcanic belts are: BSB, Belford-Stratchan belt; SLB, Saganash Lake belt; KLB, Kabinakagami Lake belt; BLB, Buchanan Lake belt.

rocks in the region. Heterogeneous, xenolithic, and migmatitic varieties of tonalite gneisses are recognized, but do not form mappable units owing to their highly variable compositional, textural and structural character both within and between outcrops. They are generally medium to light grey, fine- to coarse-grained and consist of gneissic layers, 1 to 30 cm thick, with variable proportions of biotite, hornblende, epidote and magnetite.

The analyzed tonalite gneiss (sample 4 in Leclair and Sullivan, 1991) is from a large outcrop on the shore of the Kapuskasing River (Fig. 1), in the city park in front of the Kapuskasing town hall. The outcrop of tonalite gneiss (Fig. 2) is made up of alternating layers of grey to dark grey melanocratic hornblende-biotite tonalite, and light grey leucocratic biotite tonalite with 1-2% mafic xenoliths. It contains 15 to 20% concordant tonalitic leucosome and coarse grained hornblende-bearing tonalitic sweats which occur as veins and pods cross-cutting the gneissic layering. All these rocks are injected by late pegmatite veins. Although the melanocratic phase of this gneiss was selected for isotopic dating, the sample contained very fine veinlets which we believe were the source of younger 2690 Ma zircons. In thin section, the rock displays a poorly-developed igneous texture. It is medium grained and equigranular, with hornblende and biotite showing a preferred orientation. Paleopressures of approximately 0.8 GPa were obtained from metasedimentary

Table 1. U-Pb zircon isotopic data

Fraction size in micron	Wt. mg	U ppm	Pb* ppm	$^{206}Pb/^{204}Pb$ [1]	Pbc[2] pg	^{208}Pb* %	$^{206}Pb/^{238}U$ ±1SEM%[3]	$^{207}Pb/^{235}U$ ±1SEM%[3]	Corr. Coeff.	$^{207}Pb/^{206}Pb$ ±1SEM%[3]	$^{207}Pb/^{206}Pb$ age, error (Ma)[4]
L-727-1-88 (49•24'53"N 82•25'48")											
A +149,cl,pink,N1	.037	30	18	1412	25	9.8	.52515 ± .12%	13.5881 ± .13%	.96	.18766 ± .04%	2721.8 ± 1.2
B +149,cl,pink,N1	.025	48	28	758	51	8.4	.52646 ± .16%	13.6781 ± .17%	.94	.18843 ± .06%	2728.5 ± 2.0
C +149,cl,pink,N6	.049	31	18	2328	21	9.0	.52675 ± .10%	13.6625 ± .11%	.96	.18812 ± .03%	2725.8 ± 1.0
D -74,cl,pink,4:1,N17	.009	51	27	1156	12	1.7	.51662 ± .15%	13.1120 ± .16%	.96	.18408 ± .04%	2689.9 ± 1.5
E -74,cl,pink,1:2,N14,NA	.013	57	30	1634	13	6.0	.48222 ± .10%	12.2989 ± .11%	.92	.18498 ± .05%	2698.0 ± 1.5
F -74,cl,pink,1:2,N20	.013	53	31	2616	8	8.5	.52264 ± .13%	13.5210 ± .13%	.96	.18763 ± .04%	2721.5 ± 1.2
G +105,cl,pink,rims,N3,NA	.036	50	28	10452	5	7.8	.50786 ± .09%	13.0895 ± .10%	.96	.18693 ± .03%	2715.3 ± 1.0
S -74,transluc,5:1,N11,NA	.011	266	148	8043	10	10.3	.49100 ± .09%	12.4391 ± .10%	.95	.18374 ± .03%	2686.9 ± 1.0
X +105,dark brown,N1	.007	594	318	23081	6	.7	.51966 ± .08%	13.4848 ± .10%	.96	.18820 ± .03%	2726.5 ± 0.9
H -74,cl,pink,4:1,N20	.002	81	43	313	13	.8	.51919 ± .47%	13.1342 ± .46%	.90	.18348 ± .21%	2684.5 ± 6.9
K1 +149,broken,N1	.007	41	24	126	86	8.0	.52575 ± 1.3%	13.7243 ± 1.2%	.99	.18933 ± .18%	2736.3 ± 5.8
K2 +149,broken,N4,NA	.015	41	24	1520	13	8.2	.51723 ± .14%	13.3980 ± .15%	.97	.18787 ± .04%	2723.6 ± 1.3

Notes: all fractions were abraded except as noted, NA = non abraded; cl = clear; N = number of grains analyzed; * radiogenic lead; [1] measured ratio, corrected for fractionation and spike; [2] total common Pb in analysis corrected for fractionation and spike; [3] corrected for blank Pb and U, common Pb , errors quoted are 1 standard error of the mean percent; [4] corrected for blank and common Pb, errors are 2 sigma in Ma; initial common Pb compositions from Cumming and Richards (1975).

Figure 2. Photograph showing the outcrop of tonalite gneiss where the sample was collected (above larger hammer handle). Note the numerous small leucocratic veinlets in the outcrop. These veinlets are considered to be the source of a younger zircon population dated at 2690 ± 2 Ma.

gneisses just to the west, and the Al-in-hornblende geobarometer (Hollister et al., 1987) yielded pressures of solidification in the 0.5-0.6 GPa range in surrounding granitoid rocks (see Leclair 1990, Fig. 7).

RESULTS

A total of twelve conventional U-Pb analyses were performed. Fractions were carefully selected in an attempt to clarify preliminary results (Leclair and Sullivan, 1991) and especially to identify zircon that might fall between the two previously reported ages of 2725 and 2690 Ma. No evidence was found for ages between the two, indicating two distinct zircon populations and events. The older population is characterized by equant morphology (Fig. 4, 5), whereas the younger population occurs only as fine acicular zircon with L:B up to 5:1, (Fig. 6, 7). Of the older zircons, there is evidence for igneous growth, resorption, core-overgrowth relationships (pseudo-inheritance) and formation of zircon buds, indicating a multi-event history. These processes are interpreted to have lasted about 6 Ma (2728-2722 Ma). The younger (2690 Ma) zircons were probably formed from an in situ melt. This later event appears not to have perturbed the U-Pb systematics of the older zircons. The two zircon populations are discussed separately.

Older (ca. 2725 Ma) zircons

Zircons from this population (by far the most abundant) are clear, pale pink, and internally featureless except for minor round, clear inclusions (Fig. 4, 5). They range from euhedral to anhedral, stubby to equant grains, with L:B = 1:1 to 2:1. There is a full size distribution of this type (compared to the 2690 Ma zircons that occur only as fine crystals). In general, these zircons have a rounded appearance. Some clear, equant

Table 2. Whole-grain lead evaporation results

Step	Evaporation Temp °C	Measured $^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$ Age (Ma)
1	1400	5348	2698 ± 3
2	1420	4425	2703 ± 4
3	1440	4608	2723 ± 4
4	1450	2809	2724 ± 2
5	1470	4237	2721 ± 2
6	1485	5953	2726 ± 2
7	1510	6944	2722 ± 5

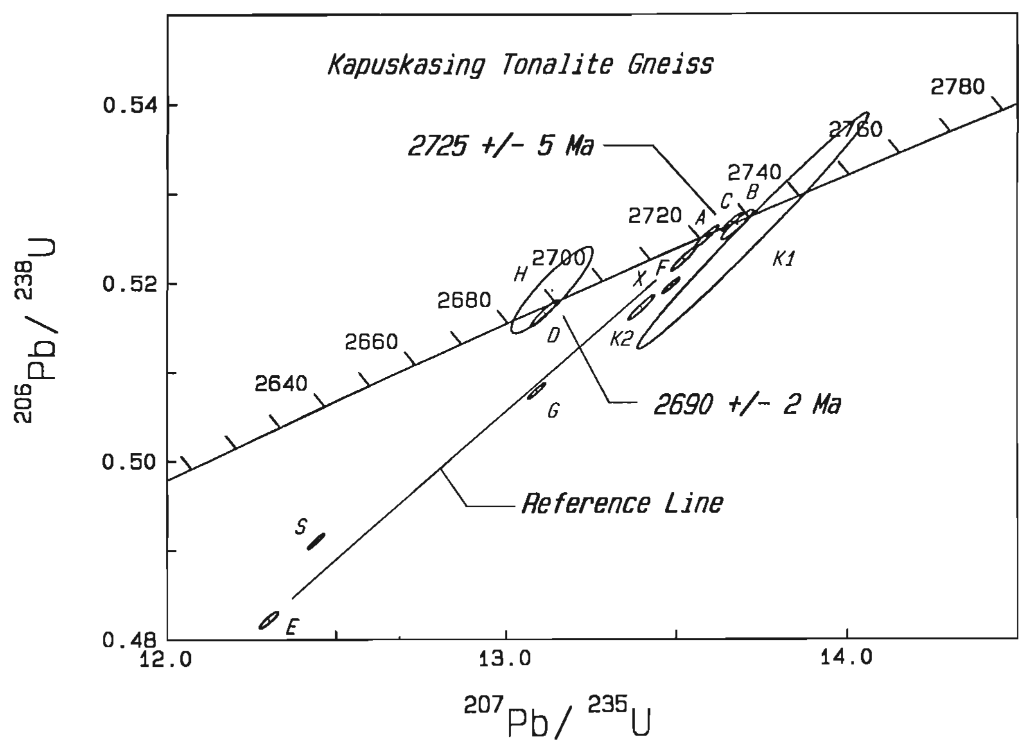

Figure 3. U-Pb concordia diagram for zircons from the Kapuskasing tonalite gneiss. There are zircons from two distinct age groups. The ca. 2725 Ma zircons have recorded a multi-event history for the tonalite (see text). The 2690 Ma zircons originated from tiny veinlets resulting from in situ melt. The reference line is drawn between fractions E and F which are a non abraded, abraded pair and is used to illustrate the two zircon populations. A regression line for fractions E, F and A yielded a 628 Ma lower intercept age.

zircons appear to have pale pink overgrowths. A small amount of dark brown zircon exists, and fragments of larger zircons are also present.

Three well abraded zircon fractions, selected from the best crystals of the coarsest size were analyzed. These equant, larger zircon fractions A, B and C, yielded concordant analyses, with ^{207}Pb-^{206}Pb ages ranging from 2721.8 ± 1.2 to 2728.5 ± 2.0 Ma (Fig. 3). Fractions A and B were single grains, whereas C was a multiple grain fraction. These different concordant ages are considered to be geologically significant.

Two finer (-74 µm) fractions were picked from the equant population, fraction F (abraded) and E (not abraded). These fine, equant zircons belong to the older (ca. 2725 Ma) population, (Fig. 3), and not the younger as one might have expected based on their morphology. A line is drawn through this pair as a reference line to separate the ca. 2725 Ma from the 2690 Ma zircons.

In an attempt to find evidence for zircon of mixed age between ca. 2725 and 2690 Ma (e.g., possible younger overgrowths), fraction G was picked but not abraded. It consisted of three curved shells, interpreted to be outer rims broken from larger zircons. This resulted in a discordant point that clearly belongs to the older (2725 Ma) population. Cores and overgrowths are present in this zircon population, but have similar ages, in the 2725 Ma range. No evidence was found for zircon of mixed age between ca. 2725 and 2690 Ma.

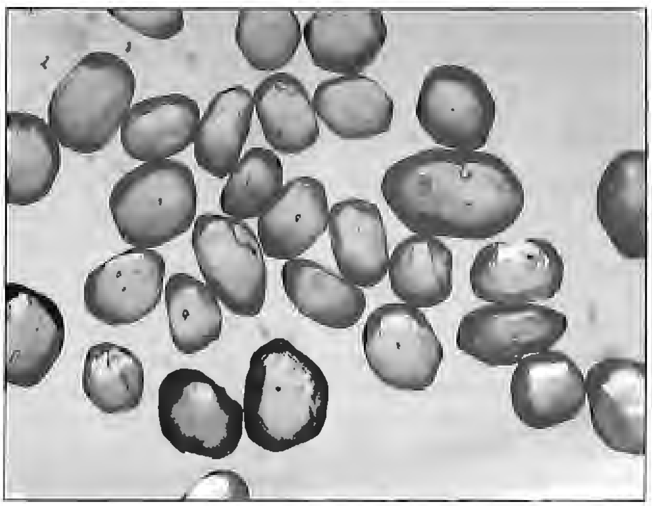

Figure 4. A photomicrograph of a picked aliquot of the large equant zircons that are typical of the bulk of the zircon from the sample. These belong to the ca. 2725 Ma age group.

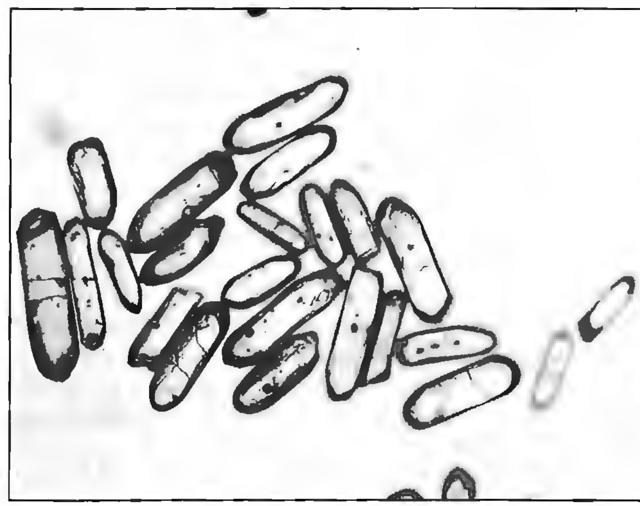

Figure 6. A photomicrograph of a picked aliquot of much less abundant, fine, -74 µm zircon. The acicular morphology characterizes the group of zircons that are 2690 Ma old.

Figure 5. A scanning electron microscope (SEM) photomicrograph from a grain mount showing a typical zircon, similar to those in Figure 4. Note the equant, generally rounded morphology.

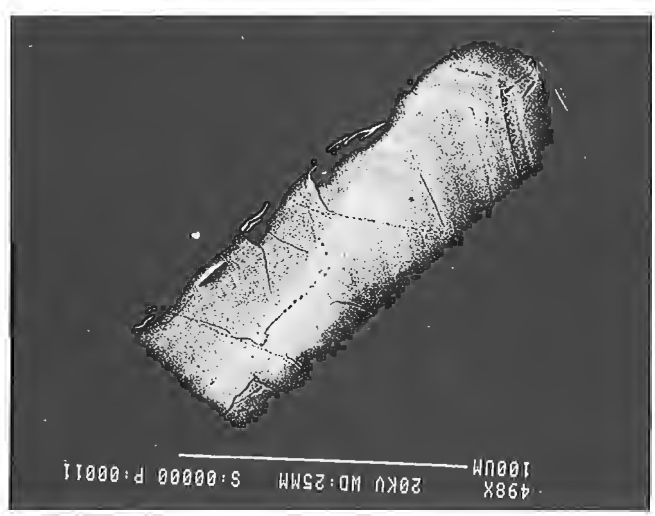

Figure 7. A SEM photomicrograph of an acicular zircon similar to those in Figure 6. This is a polished grain mount. Note the fine zoning. Similar zoning was rare or absent in the ca. 2725 Ma old zircons.

Three additional fractions were prepared to try to elucidate the age(s) of the ca. 2725 Ma zircon population. There is a very small quantity of dark brown zircon, some of which display well defined core-overgrowth relationships, with relatively thick overgrowths (Fig. 8). Fraction X represents one of these dark brown (594 ppm U) zircons. It is discordant, with a ^{207}Pb-^{206}Pb age of 2726.5 ± 0.9 Ma, and appears to belong to the ca. 2725 Ma group (Fig. 3). This result is a minimum age, however, and the true age of these very different, apparently inherited, zircons could be somewhat older.

A single large zircon, which appeared to have a thin pink overgrowth but with no visible internal features, was carefully broken into fragments to produce two more fractions, K1 (1 abraded fragment, representing the inner domain of the zircon) and K2 (4 unabraded fragments, representing the outer domain of the zircon). Fraction K1 has a large analytical error, with a ^{207}Pb-^{206}Pb age of 2736 ± 6 Ma. (Fig. 3). The other fraction from the broken zircon, K2 (not abraded), is discordant with a ^{207}Pb-^{206}Pb age of 2724 ± 2 Ma (Fig. 3). These results are consistent with there being a range of ages (at least two, i.e. ca. 2722 and ca. 2728 Ma), the zircon in this case possibly being ca. 2728 Ma.

The U-Pb results of these 9 fractions display a linear, narrow, wedge-shaped array which intersects concordia between ca. 2720 and ca. 2730 Ma (Fig. 3, ignoring K1), indicating that they belong to the same age group (ca. 2725 Ma) and are distinctly different from the 2690 Ma zircons, as illustrated by the reference line drawn between fractions E and F. The discordant linear array is attributed to normal, surface-correlated lead loss superimposed on a range of zircon crystallization ages between 2722 and 2728 Ma (i.e. concordant fractions A, B and C). The following observations and interpretations can be made. A core-overgrowth relationship, and evidence for resorption exist. These are observed in grain mounts, and polished grain mounts (Fig. 9), and are also visible in the bulk zircon population. This evidence, together with the U-Pb results, indicate a multiple-event history of zircon generation, possibly with a last crystallization event at ca. 2722 Ma. A "pseudo-inheritance" interpretation is supported by fractions A (single grain concordant at 2722 Ma), fraction B (single grain concordant at 2728 Ma) and fraction C (multiple grain concordant, with age between A and B, thus possibly representing both older and younger zircons). If inheritance is involved, the results indicate that the age difference is not great, and that the events all occurred within ca. 6 Ma. The three concordant results A, B, C could also be explained by Pb loss, associated with prolonged high temperature and slow cooling, with some zircons remaining open to Pb diffusion for ca. 6 Ma after crystallization. The former interpretation is favoured, however, as it is supported by evidence of resorption, and the existence of small zircon buds that have formed at the ends of larger zircons.

Whole-grain ^{207}Pb-^{206}Pb evaporation

A two grain aliquot (not pre-treated in any way) of the clear, equant, larger, +149 µm, zircon population was analyzed as part of the development and testing of the lead evaporation technique (Chapman and Roddick, in preparation), and also to possibly help clarify previously reported conventional U-Pb results (Leclair and Sullivan, 1991). The zircons from this sample yielded an excellent data set. Seven temperature steps were measured to give apparent ^{207}Pb-^{206}Pb ages of the 5 final steps ranging from 2721 ± 2 to 2726 ± 2 Ma, steps 3 to 7, Table 2. From these, an average of 2724 ± 3 Ma is taken as the best estimate of the age of the zircon fraction.

The measured ^{206}Pb-^{204}Pb ratios are relatively high (i.e. contain only a small common lead component), (Table 2). Corrections were made for the common lead, at the

Figure 8. A photomicrograph of a dark brown zircon with a thick, obvious overgrowth. Fraction X represents one of these zircons which are rare (high uranium), and very different from the other zircons in the sample. Energy dispersive x-ray (EDX) analyses confirmed both phases were zircon.

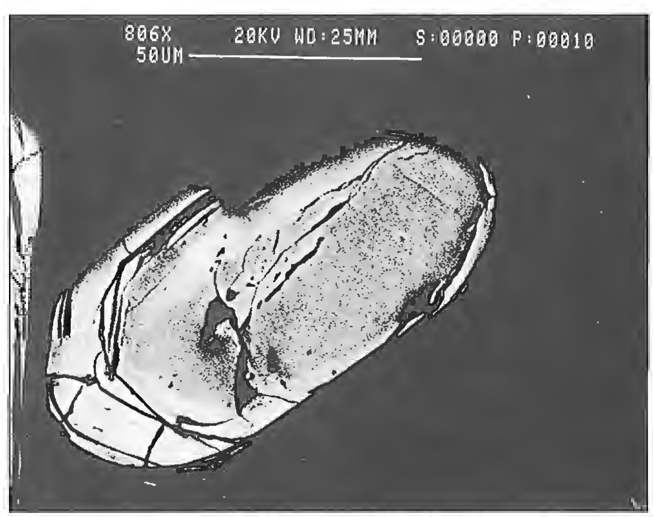

Figure 9. A SEM photomicrograph of zircon from a polished grain mount showing a core-overgrowth relationship. U-Pb analyses indicate the "core" is probably ca. 2728 Ma old, while the "overgrowth" is probably ca. 2722 Ma. The embayment is interpreted an evidence for resorption.

^{207}Pb-^{206}Pb age, based on the model for common lead composition of Cumming and Richards (1975). The lower apparent ages of steps 1 and 2 are inferred to be due to normal discordance, related to the surface-correlated lead loss, and as such are not interpreted as pointing to a younger age (e.g., a thin overgrowth). The Pb evaporated from the zircon was exhausted by the 1510°C step as indicated by the increased error.

It is interesting to note that the 2721 ± 2 and 2726 ± 2 Ma ages are just outside analytical error and might be pointing to the same range of concordant ages found from the conventional U-Pb results. This interpretation would require additional work to test.

This excellent, low-error data set, especially steps 3 to 7, is considered to be related to the clarity, homogeneity and composition of the zircons from this sample, as well as to their expected concordance based on the conventional U-Pb analyses.

The age results and other evidence indicate the ca. 2725 Ma zircon population has recorded a multi-event history including the following elements. Zircons were first crystallized at ca. 2728 Ma. This appears to have been a slow process as the zircons are clear and homogenous with negligible internal zoning. These zircons were probably colourless. Resorption produced rounded corroded surfaces that are visible in polished grain mounts. Finally, at ca. 2722 Ma, a last zircon-forming igneous event crystallized a full size distribution of pink zircons. This same event formed thin pink overgrowths on the older zircons (which abrasion would have removed). These events imparted an equant, rounded morphology to the zircons, and produced the core-overgrowth relationships (pseudo-inheritance). This scenario is supported by fractions E, F, A and possibly G, which seem to point to the 2722 Ma age. The other fractions point to the older 2728 Ma age or a mixed age between the two (Fig. 3). These events all occurred within ca. 6 Ma and are distinct from the event that occurred ca. 35 Ma later that produced the 2690 Ma zircons discussed below.

Younger (2690 Ma) zircons

This minor population consists of elongate prisms, L:B up to 5:1, with rounded terminations. This type occurs only in fine fractions. A scanning electron microscope (SEM) photomicrograph of polished grain mounts of these zircons clearly shows fine internal zoning (Fig. 7), indicating a magmatic origin. Fine zoning of this type is virtually absent in the older zircon population. Two, -74 μm, fractions (D, H) of this acicular type were analyzed. Fraction D is only slightly discordant, with a ^{207}Pb-^{206}Pb age of 2690 ± 2 Ma, taken as the best estimate of the age of this zircon population. The other fraction H plots slightly above concordia and has a larger error ellipse but is consistent with the results of fraction D. Another fine acicular fraction S was analyzed. These zircons were more translucent, and internally more complex, showing both clear and darker, cloudy domains. This fraction (266 ppm U) is very discordant, but with a ^{207}Pb-^{206}Pb age of 2686.9 ± 10 Ma. Together with the morphology, this fraction is interpreted to belong to the younger (2690 Ma) population (Fig. 3). This fraction could, however, have a component of older zircon, as indicated by its darker internal domains.

The fine, acicular, 2690 Ma zircon population is distinct and is considered to come from millimetric scale veinlets of in situ melt origin that were not entirely removed from the dated rock sample.

DISCUSSION

Three possible explanations of the preliminary data set on the Kapuskasing tonalite gneiss were proposed in Leclair and Sullivan (1991). The expanded U-Pb database and the whole-grain Pb evaporation results both clarify previous results, and establish an age for the tonalite gneiss in the Kapuskasing area at 2725 ± 5 Ma. We interpret the results to indicate a multiple event at ca. 2725 Ma lasting approximately 6 Ma, followed about 35 Ma later by another event, likely the onset of high-grade regional metamorphism, at 2690 ± 2 Ma.

The 2725 Ma age corroborates field evidence that indicates gneissic tonalite is the oldest suite of intrusive rocks in the central Kapuskasing uplift. It is worth noting also that late to post-tectonic plutons (ca. 2685-2690 Ma), which are interpreted to intrude gneissic tonalite, have an inherited zircon component, dated at ca. 2700 Ma. Within this same context, a minimum age of 2707 Ma was obtained for the Wawa tonalite gneiss (Percival and Krogh, 1983) which is contiguous and comparable in composition and structural style with the Kapuskasing tonalite gneiss. On the other hand, Moser et al. (1991) reported a ca. 2920 Ma age for highly deformed tonalite gneisses in the southern Kapuskasing uplift, with most tonalites spanning the interval from about 2720 to 2660 Ma. This wide range of ages indicates episodic and protracted tonalitic magmatism.

Kapuskasing tonalite gneisses intrude metavolcanic belts and contain numerous mafic xenoliths of hornblende-plagioclase-quartz ± clinopyroxene, assumed to be of volcanic parentage. The 2725 Ma age of the tonalite is consistent with the hypothesis that tonalites are comagmatic with volcanic activity (at 2750-2695 Ma) and represent the magma chamber that fed the overlying felsic volcanic edifices (cf. Goldie 1979; Percival and Card, 1985). Several synvolcanic plutons of 2737-2747 Ma in the Wawa subprovince have also been inferred to be comagmatic with volcanism (e.g. Turek et al., 1982; Sullivan et al., 1985).

The range of ages between 2722 and 2728 Ma is inferred to result from at least two episodes of zircon growth, possibly combined with effects of percolating fluids, partial melting and very slow cooling rates that seem to have characterized high-grade gneisses of the Kapuskasing uplift (see Krogh et al., 1988; Leclair and Sullivan 1991). These processes and their duration at lower and middle crustal levels are complex and not well understood.

The small quantity of acicular 2690 Ma zircons is interpreted as a result of an in situ tonalitic melt which occurs as thin leucocratic veinlets (Fig. 2). These veinlets are thought to have formed due to high-grade regional metamorphism. These veinlets are folded, suggesting that regional deformation took place, in part, after their emplacement.

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U-Pb age determinations from the southern Vancouver Island area, British Columbia

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Abstract

Five U-Pb age determinations from the southern Vancouver Island area are presented. Two samples of the Saltspring Intrusions intruding the Nitinat Formation of the Sicker Group have been dated at 367 ± 2 and 362 ± 2 Ma. Felsic Sicker volcanic rocks of the McLaughlin Ridge Formation are estimated to be 366 ± 4 Ma in age, though a slightly older age is possible. A mafic intrusion which is probably related to Triassic Karmutsen volcanism, is estimated to be 227 ± 3 Ma old. One of the Island Intrusions is dated at 168 ± 2 Ma, providing a minimum age for Sicker Group deformation.

Résumé

Cinq datations U-Pb faites dans le sud de l'île de Vancouver sont présentées. Deux échantillons prélevées dans les intrusions de Saltspring recoupant la formation de Nitinat du groupe de Sicker ont été datés à 367 ± 2 et 362 ± 2 Ma. L'âge des roches volcaniques felsiques du groupe de Sicker de la formation de McLaughlin Ridge a été évalué à 366 ± 4 Ma, mais elles pourraient être légèrement plus anciennes. L'âge d'une intrusion mafique probablement liée au volcanisme de Karmutsen du Trias a été évalué à 227 ± 3. L'une des intrusions de l'île a été datée à 168 ± 2 Ma, établissant un âge minimal pour la déformation du groupe de Sicker.

GEOLOGICAL FRAMEWORK

The Paleozoic Sicker Group comprises the oldest rocks of the Wrangellia terrane on Vancouver Island. The Sicker Group contains volcanic and sedimentary units ranging in age from Middle Devonian(?) to Early Permian (Massey and Friday, 1988, 1989). It is exposed mainly in three anticlinal uplifts located at Cowichan Lake, Nanoose Bay, and Buttle Lake (Fig. 1).

There have been several attempts to subdivide the Sicker Group (Yole, 1969; Muller, 1980, 1983; Massey and Friday, 1988, 1989) and a composite stratigraphic section of the Sicker in the area can be found in Massey and Friday (1989). The lower part of the Sicker Group is dominated by volcanic rocks with subordinate epiclastic sandstone, mudstone and radiolarian ribbon chert (Brandon et al., 1986; Massey and Friday, 1988, 1989). The upper part of the Sicker Group is composed of dominantly epiclastic and bioclastic sedimentary sequences (Massey and Friday, 1988, 1989; Yole, 1969; Brandon et al., 1986). It has been suggested by Sutherland Brown (reported in Massey and Friday, 1989) that the Sicker Group be redefined to include only the lower volcanic-volcaniclastic rocks, and a new Buttle Lake Group would include the younger epiclastic sediments and limestones.

The Sicker Group is associated with two suites of intrusive rocks. The Nitinat Formation, which is at a low stratigraphic level within the Sicker Group (Massey and Friday, 1989), is intruded by quartz diorite and quartz-feldspar porphyry of the Saltspring Intrusions (Muller, 1980, 1983). Mafic intrusions

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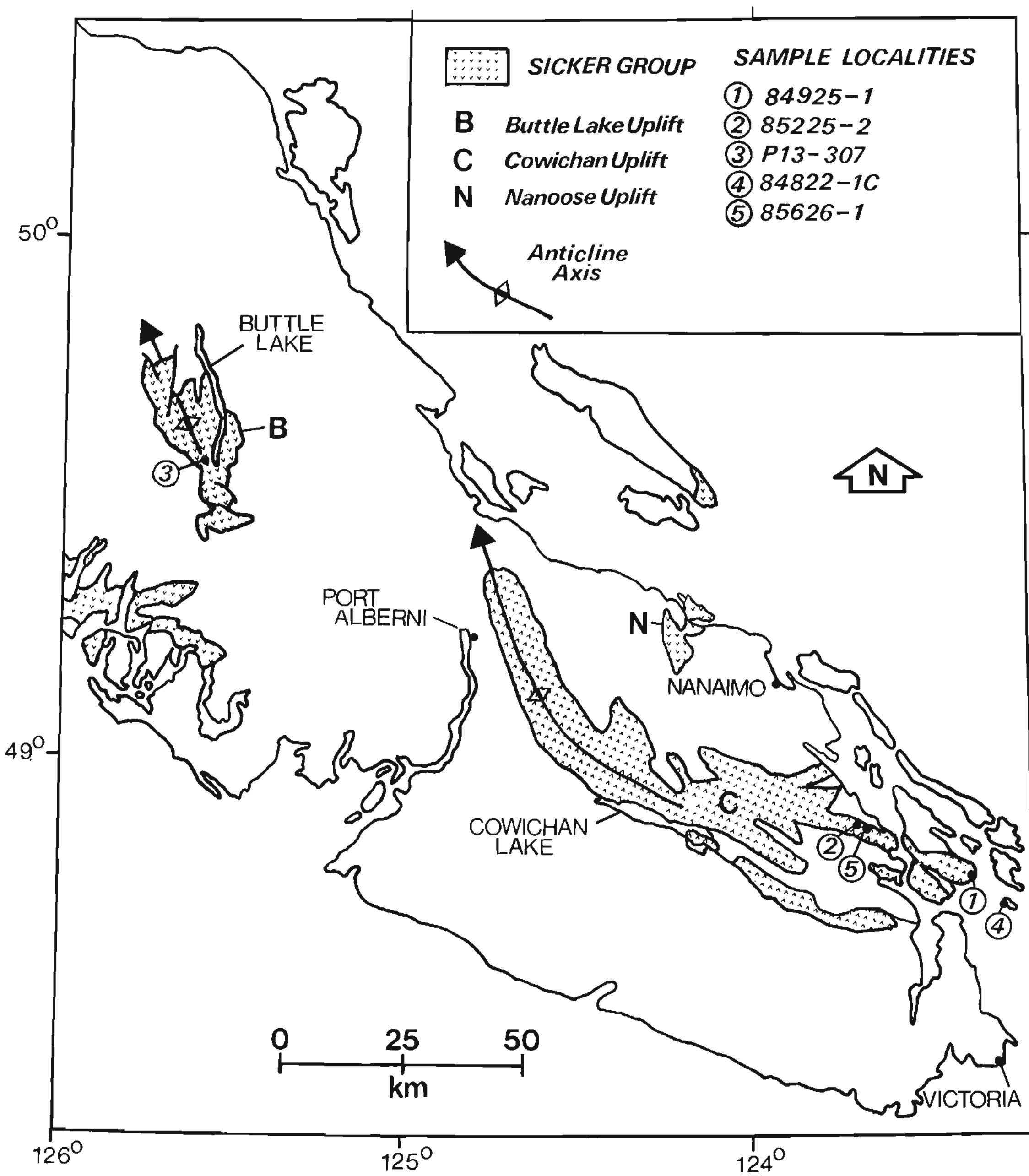

Figure 1. Map of southern Vancouver Island area with the three major anticlinal uplifts cored by Sicker Group rocks outlined (after Brandon et al., 1986; their fig. 1). Sample localities for U-Pb dating are plotted.

are widespread in the Sicker Group and are probably feeders to overlying upper Triassic basalts of the Karmutsen Formation (Brandon et al., 1986). All of the above sequences have been intruded by granodioritic stocks of the Island Intrusions (Massey and Friday, 1988, 1989). Upper Cretaceous sediments of the Nanaimo Group lie unconformably on the older sequences, including Island Intrusions.

This paper reports on five U-Pb determinations from the southern Vancouver Island area: samples 84925-1 and 85225-2 are of the Saltspring Intrusions; sample P13-307, a Sicker rhyolite from the Westmin mine at Buttle Lake; sample 84822-1C, a Karmutsen(?) mafic intrusion; and sample 85626-1, an Island Intrusion. Most of these samples formed a geochronological component of the geological investigations of M. Brandon during 1984-85.

Isotopic data and a brief outline of analytical procedures are presented in Table 1. Most, but not all, of the zircons analyzed were air abraded with pyrite (Krogh, 1982).

SAMPLE DESCRIPTIONS

Saltspring Intrusions (84925-1, 85225-2)

The Saltspring Intrusions (Muller, 1980, 1983) consist of a suite of small quartz-feldspar porphyries and some larger stocks of granodiorite. These intrusions occur on southern Saltspring Island and in the area east of Cowichan Lake. This intrusive suite includes the Mt. Maxwell stock and rocks previously called the the Tyee porphyries (Clapp and Cooke, 1917).

Sample 84925-1

This sample is a quartz-feldspar porphyry of the Saltspring Intrusions found within Sicker Group rocks. The porphyry contains large phenocrysts of quartz and feldspar up to 2 cm in size within a dark green aphanitic groundmass. At this location on southern Saltspring Island, the porphyry clearly intrudes the porphyritic Nitinat volcanic breccia of the Sicker Group (Muller, 1980; Brandon et al., 1986; Massey and Friday, 1989).

Four fractions of large, clear, euhedral zircons with minor inclusions were analyzed from this sample. The analyses plot in a short elongate cluster just below concordia at about 362 Ma (Fig. 2). The zircons have probably undergone minor

Table 1. U-Pb Analytical Data for southern Vancouver Island area [1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^*/^{204}Pb$	Pb_c,# (pg)	$^{208}Pb^+/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb/^{206}Pb$*** (Ma)
84925-1, Saltspring Intrusion, quartz-feldspar porphyry													
A,125,cl	0.332	259.8	14.99	5944	53	0.112	0.05755 ± 0.12	360.7 ± 0.8	0.4280 ± 0.14	361.7 ± 0.8	0.05393 ± 0.05	0.94	368.1 ± 2.2
B,125,sc	0.169	247.3	14.30	4631	33	0.112	0.05768 ± 0.13	361.5 ± 0.8	0.4283 ± 0.15	362.0 ± 0.8	0.05386 ± 0.06	0.91	365.2 ± 2.8
D,+149,cl	0.191	270.7	15.67	5329	35	0.111	0.05779 ± 0.09	362.2 ± 0.6	0.4294 ± 0.12	362.7 ± 0.8	0.05388 ± 0.05	0.92	366.3 ± 2.2
E,+149,cl	0.261	245.4	14.30	4658	50	0.118	0.05781 ± 0.10	362.3 ± 0.6	0.4297 ± 0.12	362.9 ± 0.8	0.05390 ± 0.05	0.92	366.9 ± 2.4
85225-2, Saltspring Intrusion, quartz-feldspar porphyry													
A,90	0.088	202.1	11.73	1670	39	0.111	0.05793 ± 0.11	363.0 ± 0.8	0.4304 ± 0.14	363.4 ± 0.8	0.05389 ± 0.08	0.86	366.2 ± 3.4
B,125	0.035	192.7	11.13	237	113	0.107	0.05786 ± 0.38	362.6 ± 2.6	0.4279 ± 0.70	361.7 ± 4.2	0.05364 ± 0.52	0.69	355.9 ± 23.4
C,90	0.055	232.1	13.37	690	68	0.108	0.05761 ± 0.19	361.1 ± 1.4	0.4268 ± 0.28	360.9 ± 1.6	0.05373 ± 0.17	0.78	359.9 ± 7.8
D,-74	0.018	287.9	16.92	1023	19	0.129	0.05777 ± 0.27	362.1 ± 1.8	0.4281 ± 0.30	361.8 ± 1.8	0.05374 ± 0.13	0.90	360.3 ± 6.0
P13-307, Quartz-feldspar porphyritic rhyolite													
A,125	0.332	144.3	8.441	1247	140	0.137	0.05713 ± 0.18	358.2 ± 1.2	0.4235 ± 0.23	358.6 ± 1.4	0.05377 ± 0.13	0.82	361.4 ± 5.8
B,90	0.299	158.5	9.331	3247	53	0.143	0.05716 ± 0.18	358.3 ± 1.2	0.4247 ± 0.20	359.4 ± 1.2	0.05389 ± 0.07	0.94	366.5 ± 3.0
C,69	0.268	175.6	10.29	2424	69	0.153	0.05646 ± 0.21	354.0 ± 1.4	0.4189 ± 0.23	355.3 ± 1.4	0.05382 ± 0.08	0.94	363.2 ± 3.4
D,90	0.976	161.7	9.530	2279	249	0.152	0.05682 ± 0.19	356.3 ± 1.4	0.4229 ± 0.22	358.2 ± 1.4	0.05399 ± 0.08	0.93	370.5 ± 3.6
E,90	0.530	160.9	9.331	1140	267	0.153	0.05586 ± 0.20	350.4 ± 1.4	0.4158 ± 0.25	353.0 ± 1.4	0.05399 ± 0.14	0.84	370.4 ± 6.2
84822-1C, Mafic Intrusion													
A,na,+62	0.183	561.7	26.82	6227	37	0.514	0.03537 ± 0.07	224.0 ± 0.2	0.2535 ± 0.10	229.4 ± 0.4	0.05200 ± 0.05	0.86	285.2 ± 2.4
B,na,53	0.183	789.9	39.22	6630	47	0.623	0.03439 ± 0.07	218.0 ± 0.2	0.2410 ± 0.10	219.3 ± 0.4	0.05083 ± 0.05	0.91	233.2 ± 2.2
BAD-1,na	0.057	265.5	8.772	522	68	0.020	0.03589 ± 0.56	227.3 ± 2.6	0.2500 ± 1.05	226.5 ± 4.2	0.05051 ± 0.89	0.53	218.6 ± 41.4
85626-1, Island Intrusion, quartz diorite													
A,125	0.373	276.7	7.518	6630	27	0.114	0.02714 ± 0.08	172.6 ± 0.2	0.1861 ± 0.11	173.3 ± 0.4	0.04975 ± 0.05	0.89	183.2 ± 2.4
B,125	0.141	297.5	7.813	2220	31	0.117	0.02619 ± 0.39	166.6 ± 1.2	0.1779 ± 0.40	166.2 ± 1.2	0.04927 ± 0.11	0.96	160.8 ± 5.0
C,125	0.111	192.0	5.116	1084	33	0.115	0.02661 ± 0.12	169.3 ± 0.4	0.1819 ± 0.20	169.7 ± 0.6	0.04958 ± 0.14	0.74	175.3 ± 6.4
D,90	0.100	155.2	4.375	774	36	0.125	0.02787 ± 0.34	177.2 ± 1.2	0.1942 ± 0.39	180.2 ± 1.2	0.05054 ± 0.30	0.65	220.0 ± 14.0
E,+149,cl	0.092	255.4	7.056	1236	33	0.115	0.02759 ± 0.25	175.4 ± 0.8	0.1896 ± 0.30	176.3 ± 1.0	0.04985 ± 0.14	0.88	187.9 ± 6.6

** BAD = baddeleyite; numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other zircon fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;

Weighing error = 0.001 mg.

\+ Radiogenic Pb.

* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU

\# Total common Pb in analysis corrected for fractionation and spike.

++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.

*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

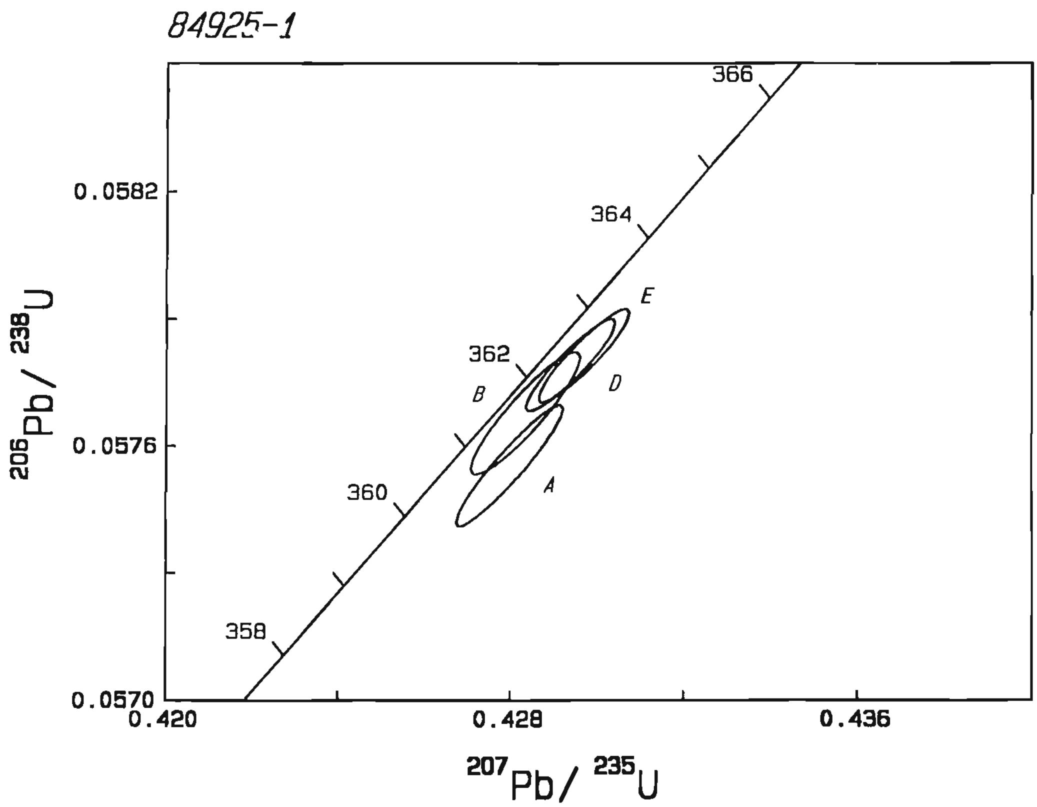

Figure 2. U-Pb concordia diagram of Sample 84925-1, Saltspring Intrusion. Error ellipses reflect 2σ uncertainty.

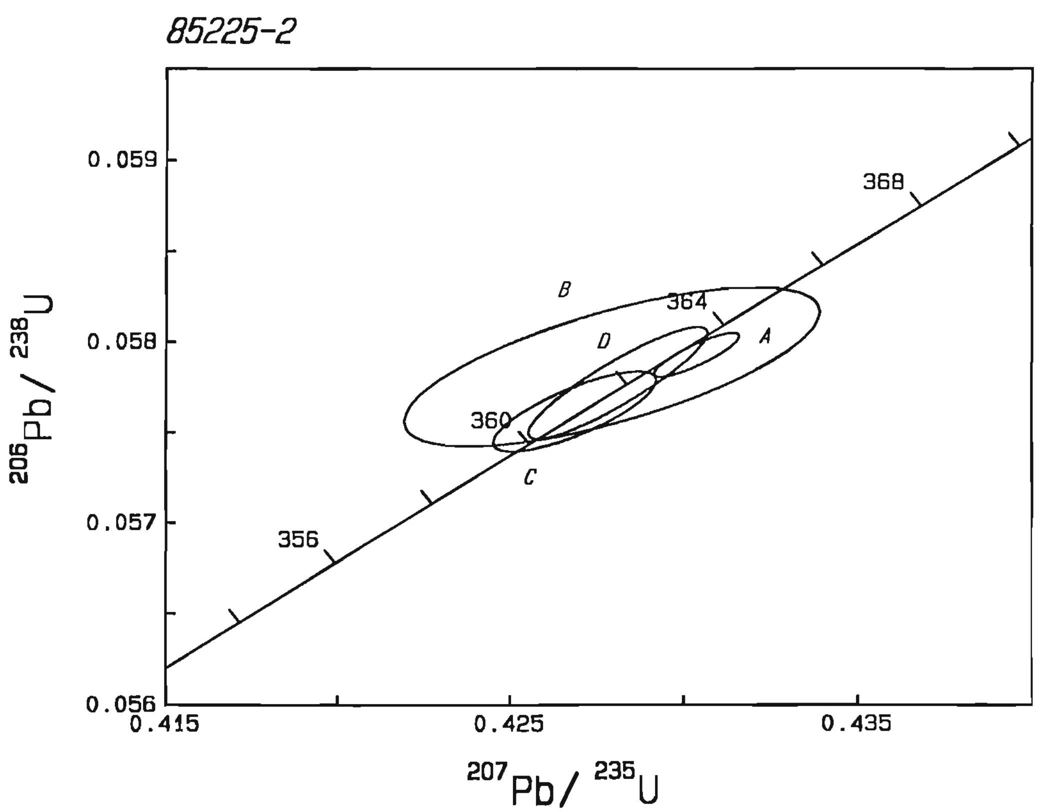

Figure 3. U-Pb concordia diagram of sample 85225-2, Saltspring Intrusion. Error ellipses reflect 2σ uncertainty.

Pb loss, and the best estimation for the age of formation of the porphyry is interpreted to be 367 ± 2 Ma, the mean of the ^{207}Pb-^{206}Pb ages.

Sample 85225-2

Sample 85225-2 is another specimen of the Saltspring Intrusions from Mt. Sicker on Vancouver Island. The sample is a quartz-feldspar porphyry with medium grained phenocrysts of rounded quartz and euhedral feldspar. The porphyry also clearly intrudes the Nitinat volcanic breccia at this locality.

Zircons from this sample are euhedral, mostly clear grains of igneous habit with some rod-like and irregular inclusions. Four zircon analyses overlap and intersect concordia, indicating an age of 362 ± 2 Ma which is interpreted as the age of crystallization of the quartz-feldspar porphyry (Fig. 3). These results were previously described in Roddick et al. (1987).

Sicker Rhyolite (P13-307)

The "Sicker Rhyolite" is a quartz-feldspar porphyritic rhyolite which is exposed in the Buttle Lake uplift (Fig. 1). Sample P13-307 is a specimen of drill core collected from the H-W mine (Westmin Resources Limited). This rock is part of a sequence of silicic pyroclastic rocks and related hypabyssal intrusions of dacite and quartz-feldspar porphyry, previously called the Myra Formation (Muller, 1980, 1983; Brandon et al., 1986), and which is now referred to as part of the McLaughlin Ridge Formation (Massey and Friday, 1988). These rocks overlie the Nitinat Formation within the Sicker Group.

Five fractions of clear, euhedral prismatic zircons with minor inclusions were analyzed from the sample. These analyses plot in a short linear array below concordia (Fig. 4). The zircons, which have only undergone very light abrasion, have experienced some Pb loss. The mean of the ^{207}Pb-^{206}Pb ages of the five analyses, 366 ± 4 Ma, is taken to be the best estimation for the *minimum* age of formation of the rhyolite. The zircons could be older, since the age of the Pb loss event need not have been recent, but the consistency of the ^{207}Pb-^{206}Pb ages of the five fractions suggests that 366 ± 4 Ma is a reasonable estimate of the crystallization age.

Mafic Intrusion (84822-1C)

Mafic intrusions which are probaby related to the Late Triassic Karmutsen Formation volcanics are widespread throughout the Sicker Group rocks in the Saltspring Island - Cowichan Lake area (Brandon et al., 1986), typically occurring as thick sills and dykes. Sample 84822-1C is a medium-grained, porphyritic gabbroic rock which intrudes

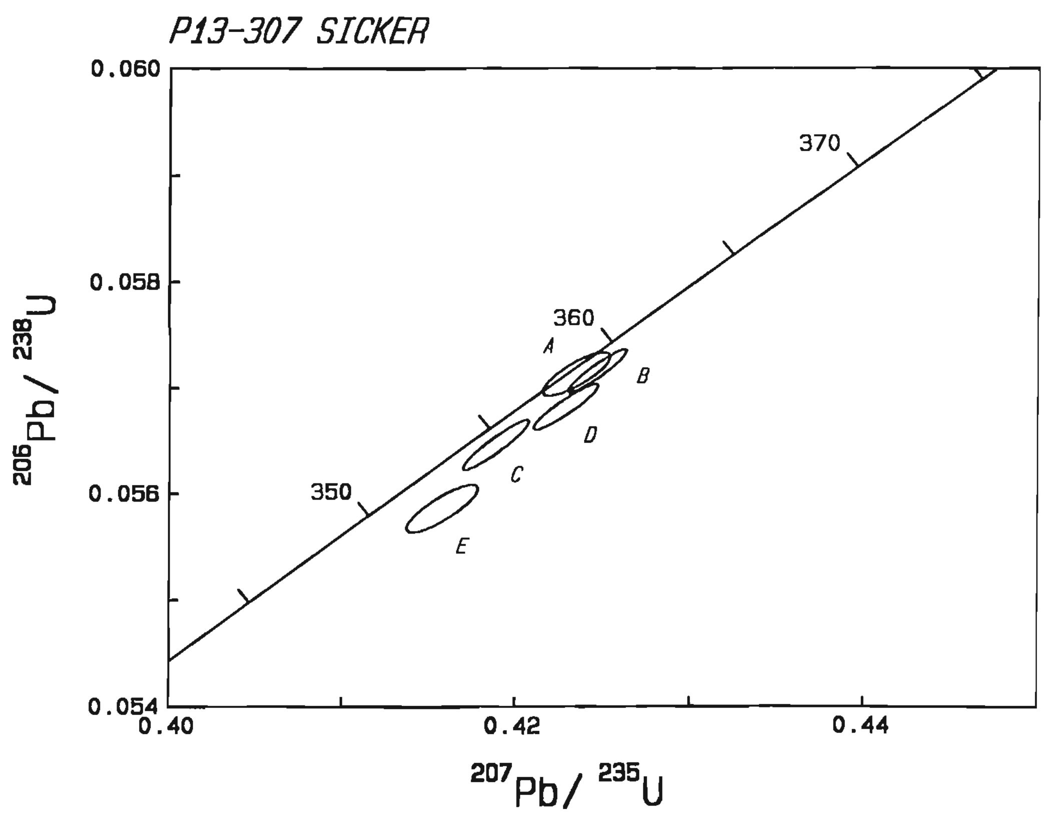

Figure 4. U-Pb concordia diagram of sample P13-307, Sicker rhyolite. Error ellipses reflect 2σ uncertainty.

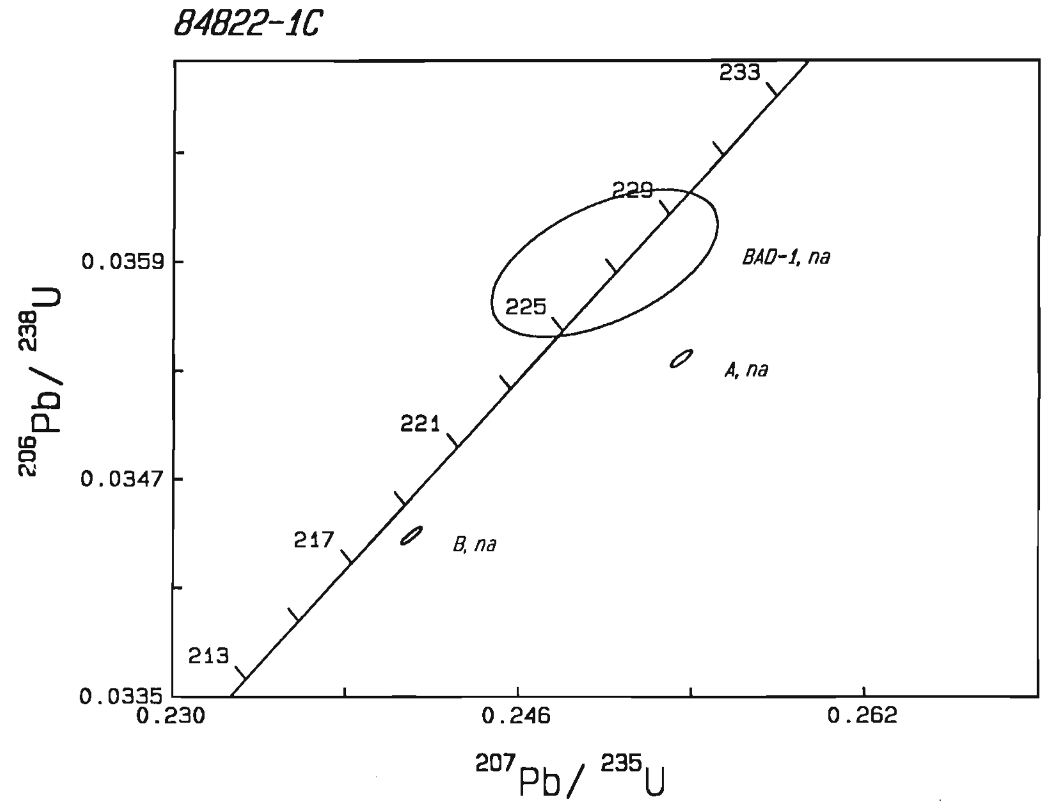

Figure 5. U-Pb concordia diagram of sample 84822-1C, mafic intrusion. Error ellipses reflect 2σ uncertainty.

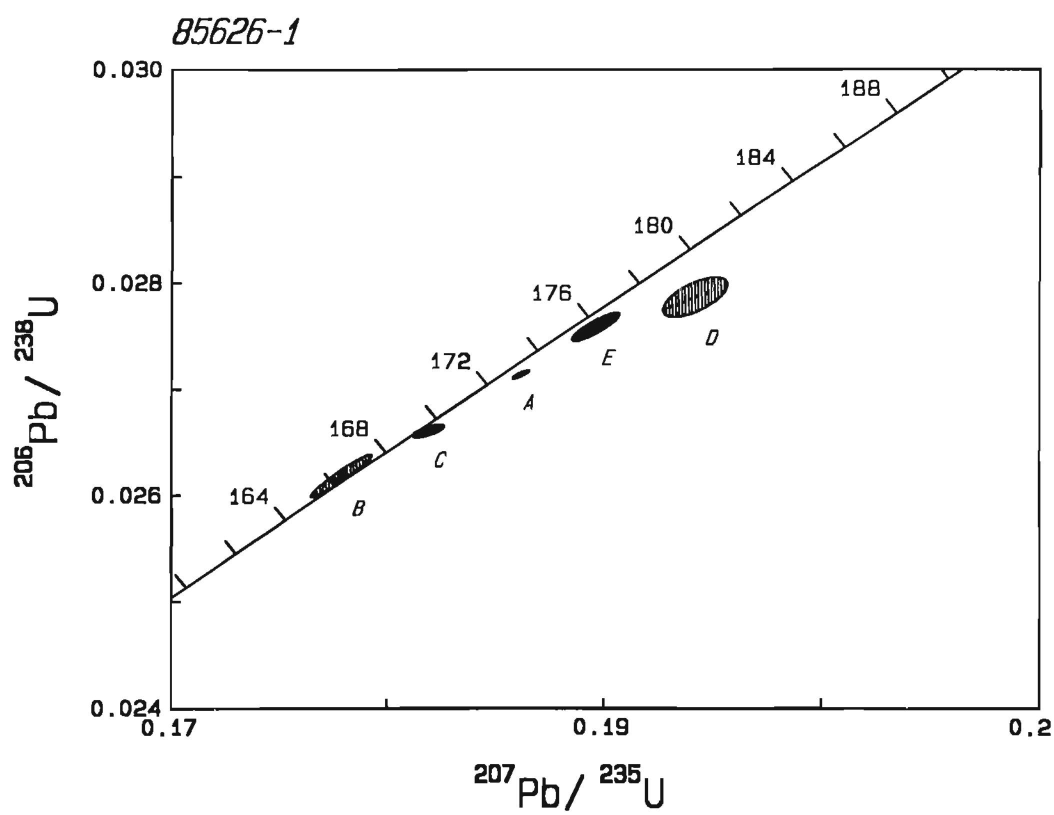

Figure 6. U-Pb concordia diagram of sample 85626-1, Island Intrusion. Error ellipses reflect 2σ uncertainty.

quartz porphyry of the Saltspring Intrusions. The gabbro is composed of plagioclase, clinopyroxene (commonly altered to uralitic amphibole) and minor skeletal ilmenite.

Two fractions composed of numerous very small zircons and one fraction of baddeleyite were analyzed from this sample; the material was generally rather poor quality, and there was sufficient baddeleyite for only one analysis.

The baddeleyite analysis overlaps concordia at 227 ± 3 Ma, which is interpreted to be the minimum age of crystallization of the mafic intrusion (Fig. 5). Since the baddeleyite may also have lost some lead, the crystallization age might have been somewhat older. The zircons, which were not abraded, may have experienced minor Pb loss and fraction A, with a ^{207}Pb-^{206}Pb age of 285 Ma, suggests the possibility that it contained xenocrysts of older age, possibly from assimilation of host rocks of Devonian age. The data from this sample are of only modest quality, but we interpret 227 ± 3 Ma as the present best estimate of age for the gabbroic magma. This age is consistent with the Carnian-Norian age span of the Karmutsen volcanic rocks, with which the gabbro is thought to be comagmatic.

Island Intrusion (85626-1)

Several granodioritic stocks of Middle Jurassic age, the Island Intrusions, occur in the Vancouver Island area. Sample 85626-1 is a medium grained, undeformed biotite-quartz diorite with minor hornblende. This pluton cuts highly deformed and metamorphosed Sicker Group rocks, and clearly postdated Sicker ductile deformation and cleavage formation.

Five fractions of clear euhedral zircons containing minor inclusions were analyzed. The analyses of discordant zircon fractions A, E, and D are interpreted to reveal components of inheritance (Fig. 6). Analyses B and C overlap concordia, but do not agree; fraction B may have lost Pb. As a result of this complex pattern we suggest an age of 168 ± 2 Ma for crystallization of this intrusion. The age of the inherited component is probably Paleozoic, possibly the Devonian Sicker volcanic and plutonic rocks which form the country rock of this intrusion.

CONCLUSIONS

Some of the main conclusions of the U-Pb age determinations are as follows:

1. The Saltspring Intrusions are late Devonian in age.
2. The felsic volcanics of the McLaughlin Ridge Formation in the Buttle Lake uplift are also late Devonian in age and could be extrusive equivalents of the Saltspring Intrusions, a possibility originally suggested by Muller (1980) (see also Brandon et al., 1986; Massey and Friday, 1988).
3. The age of a gabbroic dyke which intrudes Sicker Group is estimated to be 227 ± 3 Ma; this strengthens the correlation of the gabbro with the Late Triassic Karmutsen volcanics (Brandon et al., 1986; Massey and Friday, 1988, 1989).
4. Initial deformation of the Sicker Group is pre-Middle Jurassic. The age of an Island Intrusion (168 ± 2 Ma) provides this constraint on the minimum age of this deformation (see also Brandon et al., 1986; Massey and Friday, 1988, 1989).

ACKNOWLEDGMENTS

M. Brandon collected most of the samples analyzed in this study. Rick Walker provided the sample from the H-W mine (Westmin Resources). Nick Massey and Rod Kirkham are thanked for discussions. M. Villeneuve reviewed the manuscript.

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APPENDIX 1

Sample locations for U-Pb dating

84925-1: Saltspring Intrusion, quartz-feldspar porphyry; southern Saltspring Island, along the coast, on the south side of cove just north of Beaver Point, Victoria map area, B.C. (92B/14); 123°22′14″ W - 48°46′30″ N; UTM 10u, 472774E-5402291N.

85225-2: Saltspring Intrusion, quartz-feldspar porphyry; east side of Mt. Sicker, third outcrop (north end of outcrop) on west side of Highway 1 going south from Chemainus River (2.25 km south of bridge), Victoria map area, Vancouver Island, B.C. (92B/13); 123°42′50″ W - 48°51′38″ N; UTM 10u, 447637E-5411981N.

P13-307: Sicker Volcanics, quartz-feldspar porphyry; H-W mine (Westmin), Buttle Lake, Alberni map area, Vancouver Island, B.C. (92F/12); 125°35′20″ W - 49°34′15″ N; UTM 10u, 312820E-5493917N.

85626-1: Island Intrusion, quartz diorite; Moresby Island (southeast of Saltspring Island), location is approximately 1 km southeast of Seymour Point, Victoria map area, B.C. (92B/11); 123°19′22″ W - 48°42′45″ N; UTM 10u, 476255E-5395328N.

84822-1C: mafic intrusion; easternmost quarry on north fork of dirt road off Shasta St. (sample collected from west side of main quarry), Crofton, Victoria map area, Vancouver Island, B.C. (92B/13); 123°40′08″ W - 48°51′41″ N; UTM 10u, 450931E-5412038N.

- 11 -

New U-Pb dates from southwestern British Columbia

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Parrish, R.R. and Monger, J.W.H., 1992: New U-Pb dates from southwestern British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 87-108.

Abstract

U-Pb dates are presented for seventeen samples of igneous rocks within southern British Columbia in the Ashcroft (92I), Hope (92H), and Vancouver (92G) map sheets. Dates range in age from more than 200 Ma to 63 Ma and reflect changing patterns of plutonism in the Intermontane and Coast Belts of the southern Canadian Cordillera.

Résumé

Les datations U-Pb de 17 échantillons de roches ignées reposant dans le sud de la Colombie-Britannique correspondant aux cartes d'Ashcroft (92I), de Hope (92H) et de Vancouver (92G) sont présentées. Les âges varient de plus de 200 Ma à 63 Ma et reflètent le plutonisme changeant qu'ont connu les zones intramontagneuses et littorales dans le sud de la Cordillère canadienne.

INTRODUCTION

Sampling of rock units for dating using various isotopic systems was carried out in the last decade during 1:250 000 scale re-mapping of Hope (92H), Ashcroft (92I), and Vancouver (92G) map areas (Monger, 1989; Monger, 1991a). The new dates supplement the numerous (ca. 300), mainly K-Ar dates collected in these areas over the past three decades by many workers, which are listed on the "Isotope date location maps" of Hope and Ashcroft areas (Monger, 1989) and shown on the geological map of Vancouver, west half (Roddick and Woodsworth, 1979). Much of the isotopic dating done in Hope and Ashcroft map areas reflects interest in mineral deposits associated with relatively high level plutonic rocks of the Intermontane Belt (Fig. 1), and for many of the latter, K-Ar dates are sufficient to give good approximations of times of crystallization. For example, genesis of Highland Valley porphyry copper ores is intimately associated with crystallization history of the Late Triassic Guichon Creek Batholith (Fig. 1; McMillan, 1976), which has been dated mostly by K-Ar. By contrast, the southern Coast Belt comprises abundant granitic rocks (about 80% of all exposed lithologies). For these, several isotope systems are needed, as crystallization ages may be very different from cooling ages. Widespread dating within the Coast Belt using U-Pb, K-Ar, and other isotopic systems has been done only in the last few years, in part during re-mapping of Vancouver and Pemberton (92J) areas (Journeay, 1990; Monger, 1991a), and in part as an ancillary project related to the LITHOPROBE deep seismic reflection profile across the southern Coast Mountains (Friedman and Armstrong, 1990; J.M. Journeay and R.M. Friedman, unpublished data).

This paper presents 17 new U-Pb dates, 9 from the Intermontane Belt and 6 from the Coast Belt (Fig. 1). With one exception, all are from plutonic rocks. In addition a muscovite Rb-Sr date is reported from the Coast Belt. Sampling locations are shown in Figure 1 (localities 1-17). With two exceptions, dates from the Intermontane Belt are older, and include terrane specific granitic rocks coeval and possibly comagmatic with the Late Triassic and earliest Jurassic Nicola-Rossland arc which characterizes Quesnellia terrane (localities 1 through 4), a late Middle Jurassic granitic body (loc. 5) which post-dates accretion of Quesnellia to the ancient continental margin, and three early Tertiary granitic rocks (samples 12, 14 and 18) (Monger et al., 1982; Archibald et al., 1983). Granitic rocks from the Coast Belt include a latest Jurassic terrane specific(?) granitic rock (loc. 6) that

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pre-dates the early Late Cretaceous (92-96 Ma) major contraction (collision?) event centred in southeastern Coast Mountains, a crystal lithic tuff ejected just before that event (loc. 7), granitic and gneissic rocks recrystallized during it (loc. 8, 9, 10), and a granitic rock emplaced after initial uplift following collision (loc.11) (Journeay, 1990; Monger, 1991a).

The dates herein supplant any preliminary GSC dates reported for the same specimens on the isotope date location maps in Monger (1989). To avoid confusion where this has been done, the field specimen number (e.g. MV-86-99) is given together with the set of characters used to identify the specimen collection site on the location maps (e.g. eTg16-1; which is a combination of map unit symbol (eTg); 1:50 000 scale map sheet number (16) within map 92I; and number per unit on each map sheet (1)).

The Decade of North American Geology time scale is used where it is necessary to equate stratigraphic ages with numerical dates.

NOTES ON ANALYTICAL METHODS

The analyses presented in this paper were produced over the period of time from 1984 to 1990, a time period when analytical methods evolved rapidly at the Geological of Survey of Canada laboratories. Earlier analyses can be identified as having large sample weights (>0.5 mg for zircon) and larger amounts of total common Pb in the analyses of zircon (generally more than 100 pg). Errors have been estimated for some of these older analyses. Analytical methods are those of Parrish et al. (1987). Except where otherwise noted, all zircon fractions were air abraded with pyrite using the method of Krogh (1982). Interested readers will note that a number of samples with high precision data have zircon error ellipses which fall slightly below concordia. This is a normal feature of zircons which, when they crystallize, are deficient in ^{230}Th relative to the magma from which they crystallize, resulting in a slight deficit of ^{206}Pb; this causes the ellipses to be slightly below concordia. This is described in more detail in Coleman and Parrish (in press).

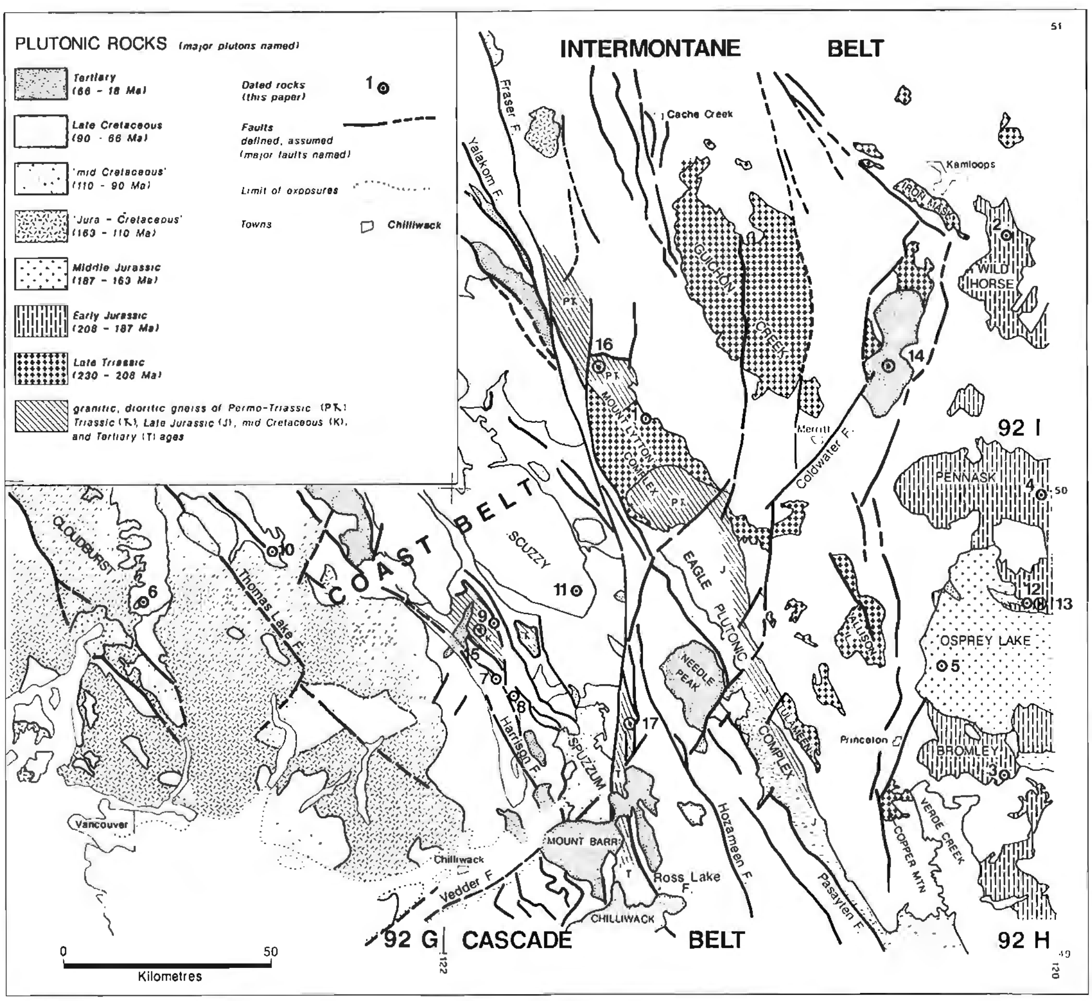

Figure 1. Map of plutonic rocks and major structures of the Intermontane and Coast Belts, southwest B.C. Sample locations discussed in this paper are plotted on the map. Exact sample localities are described in Appendix 1.

The maximum amount of this deficit is approximately 0.1 Ma, and a correction for this phenomenon will move ellipses up toward concordia by this amount. Where analytical results are sufficiently precise to discern this feature such as in Figures 4b, 5, 7, 9, 10, and/or 18, it has been taken into account when interpreting the best estimate of crystallization age. Data are presented in Table 1 and displayed in the concordia plots (Fig. 2-18). Details of localities are listed in the Appendix.

(1) MOUNT LYTTON COMPLEX

Specimen MV-86-120 (TJgd3-3) is a medium- to coarse-grained, but relatively fresh and only locally foliated, biotite granodiorite that intrudes variably foliated and mylonitic felsic and amphibolitic gneiss, marble and metadiorite. The host metamorphic rocks yield U-Pb dates of Permo-Triassic age (sample 17 below and P. van der Heyden, unpublished data). Together, both metamorphic and intrusive rocks form the Mount Lytton Complex, and with the Eagle Plutonic Complex of mainly Late Jurassic age contiguous to the south (Greig, 1988), forms a south southeast-trending welt of mid-crustal rocks uplifted in mid-Cretaceous time (Monger, 1985; 1989). The welt delineates the margin of the Intermontane Belt at these latitudes and is separated from the Cascade Belt to the west by the Pasayten Fault (Fig. 1).

Analysis

Two fractions of zircon and two of titanite were analyzed from sample MV-86-120. One of the zircon analyses and both titanite analyses intersect and overlap concordia. Based on the overlap with concordia shown in Figure 2, we interpret 212 ± 1 Ma as the age of crystallization of the granodiorite.

Interpretation and discussion

The interpreted age of 212 ± 1 Ma is very close to some dates reported from the central Guichon Creek Batholith, which is located about 40 km to the northeast, as shown by recalculated K-Ar dates (listed in Monger, 1989) and an unpublished U-Pb date of 210 Ma on a specimen collected by N. Mortimer (University of British Columbia Geochronometry File). Both Mount Lytton and Guichon Creek bodies appear to be part of the Late Triassic-earliest Jurassic magmatic arc complex that characterizes Quesnellia terrane.

However, the Mount Lytton Complex intrusion and the Guichon Creek Batholith were probably emplaced at different structural levels, as suggested by their contrasting settings and K-Ar dates. The intrusion in the Mount Lytton Complex cross-cuts deformation in enclosing gneiss and metadiorite. Concordant biotite and hornblende dates of 186 Ma (GSC K-Ar 87-193, 87-194; Hunt and Roddick 1987) obtained from a sample (MV-81-141a) located 1.25 km due

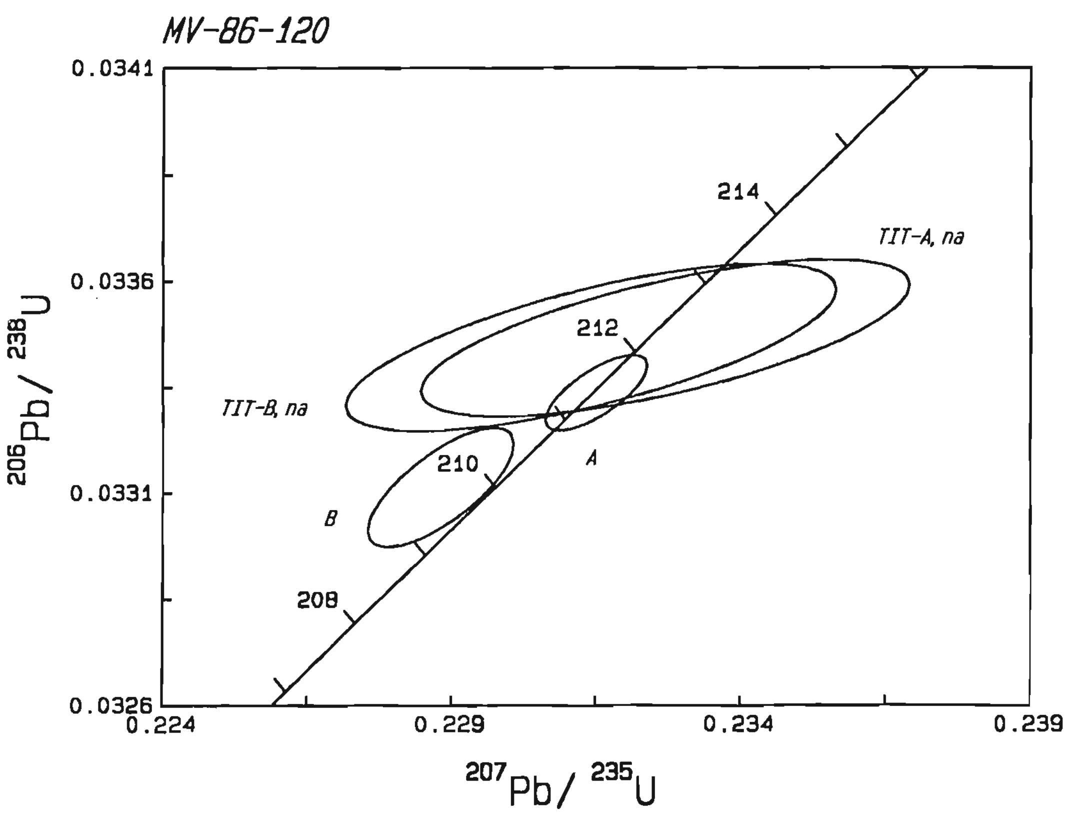

Figure 2. U-Pb concordia plot of MV-86-120, massive granodiorite of the Mount Lytton Complex. Error ellipses reflect the 2σ uncertainty.

south of MV-86-120, suggests an early Middle Jurassic cooling age for the intrusion, presumably due to rapid cooling in early-middle Jurassic time. The Complex was uplifted together with the Eagle Plutonic Complex and deeply eroded in late Early Cretaceous time, and is non-conformably overlain by the Spences Bridge Group of Albian-Cenomanian age (105 Ma; Thorkelson and Rouse, 1989). By contrast, the Guichon Creek Batholith was emplaced in Nicola volcanic and sedimentary rocks of mainly subgreenschist metamorphic facies of Late Triassic age (late Karnian-early Norian; ca. 227-215 Ma; Monger, 1989). Both were eroded prior to late Pliensbachian time (ca. 195 Ma) when the Ashcroft Formation was deposited on them. Field relationships, plus general concordance of dates from K-Ar, Rb-Sr and U-Pb systems on this pluton (see listing in Monger, 1989) shows that the Guichon Creek body was emplaced at

Table 1. U-Pb analytical data of samples from southwestern British Columbia

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb/^{204}Pb$ *	Pb_c,# (pg)	$^{208}Pb/^{206}Pb$ +	$^{206}Pb/^{238}U$ ++	$^{206}Pb/^{238}U$ (Ma) ++	$^{207}Pb/^{235}U$ ++	$^{207}Pb/^{235}U$ (Ma) ++	$^{207}Pb/^{206}Pb$ ++	corr. coef.	$^{207}Pb/^{206}Pb$ (Ma) ***
(1) MV-86-120, massive granodiorite, Mount Lytton Complex													
A,+149	0.337	111.1	3.613	1314	61	0.084	0.03334 ± 0.13	211.4 ± 0.6	0.2315 ± 0.19	211.4 ± 0.8	0.05037 ± 0.13	0.73	212.0 ± 6.2
B,125	0.245	189.0	6.084	1018	96	0.081	0.03311 ± 0.21	210.0 ± 0.8	0.2288 ± 0.28	209.2 ± 1.0	0.05012 ± 0.19	0.73	200.4 ± 8.8
TIT-1,na	0.416	79.43	4.605	213	355	0.956	0.03347 ± 0.28	212.2 ± 1.2	0.2327 ± 0.91	212.4 ± 3.5	0.05043 ± 0.75	0.68	214.8 ± 35.1
TIT-2,na	0.332	81.75	4.657	198	316	0.923	0.03344 ± 0.29	212.1 ± 1.2	0.2314 ± 0.92	211.4 ± 3.5	0.05019 ± 0.74	0.70	203.6 ± 34.9
(2) MVK-82-20, granodiorite, Wild Horse Batholith													
A,na,+149	1.820	396.0	11.88	1682	794	0.082	0.03081 ± 0.35	195.6 ± 1.4	0.2166 ± 0.40	199.0 ± 1.6	0.05097 ± 0.10	0.80	239.6 ± 4.4
B,125	5.190	324.2	9.763	3384	924	0.089	0.03075 ± 0.30	195.3 ± 1.0	0.2128 ± 0.35	195.9 ± 1.2	0.05018 ± 0.05	0.90	203.4 ± 2.2
C,90,e	3.690	339.5	10.32	1665	1399	0.095	0.03087 ± 0.35	196.0 ± 1.4	0.2138 ± 0.40	196.7 ± 1.6	0.05021 ± 0.10	0.80	204.8 ± 4.4
(3) MV-84-41, Bromley Pluton													
A,na,125	0.800	331.7	9.436	3118	154	0.102	0.02871 ± 0.18	182.5 ± 0.6	0.1989 ± 0.20	184.2 ± 0.6	0.05024 ± 0.07	0.94	206.3 ± 3.2
B,125	0.430	275.0	8.234	775	298	0.100	0.03028 ± 0.09	192.3 ± 0.4	0.2091 ± 0.24	192.8 ± 0.8	0.05010 ± 0.18	0.73	199.8 ± 8.4
C,+149	0.155	253.9	7.601	1526	50	0.100	0.03027 ± 0.20	192.2 ± 0.8	0.2089 ± 0.23	192.7 ± 0.8	0.05006 ± 0.11	0.87	197.8 ± 5.2
D,el	0.133	229.4	6.871	1565	38	0.096	0.03038 ± 0.10	192.9 ± 0.4	0.2092 ± 0.14	192.9 ± 0.5	0.04994 ± 0.09	0.80	192.3 ± 4.2
E,e	0.086	247.5	7.444	2178	19	0.094	0.03057 ± 0.09	194.1 ± 0.3	0.2106 ± 0.12	194.1 ± 0.4	0.04997 ± 0.06	0.87	193.7 ± 2.7
F,125	0.107	283.8	9.022	763	81	0.108	0.03182 ± 0.22	201.9 ± 0.8	0.2323 ± 0.28	212.1 ± 1.0	0.05296 ± 0.18	0.76	327.0 ± 4.1
G,+149	0.175	254.8	8.320	2045	45	0.099	0.03297 ± 0.10	209.1 ± 0.4	0.2391 ± 0.14	217.7 ± 0.6	0.05260 ± 0.07	0.86	311.7 ± 3.2
(4) MVM-86-9, Pennask Batholith													
A,90	0.254	236.5	7.240	2978	39	0.119	0.03045 ± 0.09	193.3 ± 0.2	0.2107 ± 0.13	194.1 ± 0.2	0.05018 ± 0.07	0.86	203.5 ± 1.6
B,el	0.120	144.0	4.424	2626	13	0.120	0.03054 ± 0.10	193.9 ± 0.4	0.2104 ± 0.12	193.9 ± 0.4	0.04997 ± 0.07	0.82	193.8 ± 3.2
C,e	0.134	151.5	4.578	3442	11	0.104	0.03046 ± 0.09	193.4 ± 0.3	0.2101 ± 0.11	193.7 ± 0.4	0.05004 ± 0.05	0.91	197.0 ± 2.1
(5) MV-84-5, Osprey Lake Batholith													
A,+149	0.800	701.5	18.34	5307	169	0.151	0.02532 ± 0.09	161.2 ± 0.2	0.1724 ± 0.11	161.5 ± 0.4	0.04938 ± 0.05	0.92	166.0 ± 2.2
B,na,125	0.461	780.4	20.25	7063	81	0.147	0.02521 ± 0.18	160.3 ± 0.6	0.1716 ± 0.20	160.8 ± 0.6	0.04936 ± 0.05	0.97	164.7 ± 2.2
C,na,68	0.764	984.8	25.57	6359	187	0.155	0.02505 ± 0.18	159.5 ± 0.6	0.1706 ± 0.20	160.0 ± 0.6	0.04939 ± 0.05	0.97	166.6 ± 2.4
(6) MV-89-157, Cloudburst quartz diorite													
A,+149,e	0.200	73.57	1.679	2141	10	0.104	0.02302 ± 0.10	146.7 ± 0.3	0.1557 ± 0.12	146.9 ± 0.3	0.04906 ± 0.062	0.85	150.6 ± 2.9
B,+149,e	0.216	75.16	1.716	2032	12	0.104	0.02302 ± 0.10	146.7 ± 0.3	0.1557 ± 0.12	146.9 ± 0.3	0.04905 ± 0.070	0.83	150.1 ± 3.3
(7) MV-85-482, Brokenback Hill Formation(?)													
A,na,cl	0.114	129.8	2.129	34	1039	0.053	0.01726 ± 3.5	110.3 ± 7.6	0.1248 ± 10.1	119.4 ± 22.8	0.05244 ± 8.1	0.69	304.8 ± 349.2
B,na	0.082	267.0	2.252	267	48	0.086	0.01600 ± 0.19	102.3 ± 0.4	0.1059 ± 0.60	102.2 ± 1.2	0.04803 ± 0.50	0.67	100.9 ± 23.7
C,na,c	0.053	106.2	14.43	2521	18	0.135	0.1300 ± 0.10	787.8 ± 1.5	1.4183 ± 0.13	896.7 ± 1.6	0.07914 ± 0.07	0.86	1175.7 ± 2.7
(8) MVH-85-92, Gneissic quartz diorite													
A,90	0.069	1241	18.29	469	180	0.092	0.01504 ± 0.12	96.2 ± 0.2	0.09937 ± 0.33	96.2 ± 0.6	0.04793 ± 0.26	0.68	96.0 ± 12.4
B,90	0.092	1265	18.88	1305	86	0.099	0.01513 ± 0.10	96.8 ± 0.2	0.1003 ± 0.16	97.0 ± 0.4	0.04807 ± 0.11	0.74	102.9 ± 5.2
C,125	0.100	1302	19.44	1440	87	0.096	0.01518 ± 0.09	97.1 ± 0.2	0.1005 ± 0.15	97.3 ± 0.2	0.04803 ± 0.10	0.76	101.0 ± 4.8
D,125	0.061	1132	16.63	896	74	0.086	0.01506 ± 0.11	96.4 ± 0.2	0.09954 ± 0.19	96.4 ± 0.4	0.04794 ± 0.14	0.68	96.1 ± 6.8
(9) MV-86-343, Mafic Breakenridge gneiss													
A,+149,cl	0.296	153.7	2.202	718	62	0.061	0.01502 ± 0.13	96.1 ± 0.2	0.09928 ± 0.26	96.1 ± 0.5	0.04795 ± 0.20	0.61	97.0 ± 9.7
B,125,cl	0.206	115.0	1.658	630	37	0.065	0.01506 ± 0.15	96.3 ± 0.3	0.1009 ± 0.35	97.7 ± 0.6	0.04863 ± 0.30	0.52	130.0 ± 14.0
C,90,cl	0.143	190.0	2.705	710	37	0.058	0.01497 ± 0.20	95.8 ± 0.4	0.09940 ± 0.37	96.2 ± 0.7	0.04816 ± 0.30	0.60	107.3 ± 14.2
(10) MVD-89-375, Snowcap Lake Pluton													
A,+149,cl	0.102	102.6	1.548	1018	10	0.145	0.01469 ± 0.16	94.0 ± 0.3	0.09742 ± 0.18	94.4 ± 0.3	0.04808 ± 0.12	0.75	103.4 ± 5.9
B,+149,cl	0.113	94.92	1.434	867	12	0.149	0.01468 ± 0.21	94.0 ± 0.4	0.09646 ± 0.26	93.5 ± 0.5	0.04765 ± 0.18	0.74	81.9 ± 8.4
C,+149,cl	0.104	110.3	1.659	862.8	13	0.142	0.01468 ± 0.13	94.0 ± 0.2	0.09678 ± 0.20	93.8 ± 0.4	0.04780 ± 0.15	0.69	89.5 ± 6.9

shallow levels, within the slightly older but probably comagmatic volcanic carapace represented by the Nicola Group.

(2, 3, 4) EARLY JURASSIC PLUTONS

Specimens MVK-82-20 (2), MV-84-41 (3), and MVM-86-9 (4), are from a north-south oriented suite of plutons located in the easternmost Hope and Ashcroft map areas and called, respectively, Wild Horse Batholith, Bromley Batholith (new name) and Pennask Batholith. They are considered by us to be comagmatic with younger parts of the Late Triassic-earliest Jurassic Nicola-Rossland arc, although no volcanic rocks younger than latest Triassic are known from this area. Locations and analyses for all three bodies are given separately.

(2) WILD HORSE BATHOLITH (MVK-82-20)(TJgd9-3)

The specimen is a medium grained hornblende biotite granodiorite with scattered large feldspar phenocrysts, from the central part of the Wild Horse Batholith. The batholith is emplaced within the subgreenschist eastern sedimentary facies of the Nicola Group, and locally is overlain by middle Eocene Kamloops Group volcanic rocks.

K-Ar dates on hornblende of 203 ± 4 Ma were reported by Preto et al. (1979) from samples near the northwestern margin of the batholith, 5 km to the northwest. Other samples 3 km west-southwest of Dardenelles Lake, 21 km farther south at the southern margin of the batholith, yielded dates

Table 1. (cont.)

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb/^{204}Pb$ *	Pbc,# (pg)	$^{208}Pb/^{206}Pb$ +	$^{206}Pb/^{238}U$ ++	$^{206}Pb/^{238}U$ ++ (Ma)	$^{207}Pb/^{235}U$ ++	$^{207}Pb/^{235}U$ ++ (Ma)	$^{207}Pb/^{206}Pb$ ++	corr. coef.	$^{207}Pb/^{206}Pb$ *** (Ma)
(11) MV-86-431, Scuzzy Pluton													
A,125,cl,e	0.096	382.4	5.029	868	38	0.045	0.01398 ± 0.13	89.5 ± 0.2	0.09513 ± 0.24	92.3 ± 0.4	0.04936 ± 0.18	0.67	164.7 ± 8.4
B,125,e	0.065	444.7	5.420	795	31	0.031	0.01315 ± 0.15	84.2 ± 0.2	0.08617 ± 0.27	83.9 ± 0.4	0.04754 ± 0.21	0.66	76.5 ± 9.9
C,+149,cl	0.133	452.2	5.506	1568	32	0.030	0.01314 ± 0.11	84.2 ± 0.2	0.08724 ± 0.17	84.9 ± 0.3	0.04815 ± 0.11	0.76	106.8 ± 5.3
D,+149,cl	0.134	436.1	5.314	1802	27	0.033	0.01311 ± 0.11	84.0 ± 0.2	0.08643 ± 0.17	84.2 ± 0.3	0.04780 ± 0.10	0.80	89.5 ± 4.9
(12) MV-86-99, Quartz-feldspar porphyry													
A,+149	0.359	313.2	3.185	1142	63	0.132	0.01002 ± 0.11	64.3 ± 0.2	0.06682 ± 0.19	65.7 ± 0.2	0.04837 ± 0.14	0.70	117.2 ± 6.4
B,125	0.247	424.4	8.041	2777	46	0.086	0.01910 ± 0.08	122.0 ± 0.2	0.1771 ± 0.11	165.6 ± 0.4	0.06727 ± 0.06	0.85	846.0 ± 2.4
(13) MV-86-98, Pennask Batholith													
TIT-A,na	0.241	103.4	3.446	74	623	0.661	0.02261 ± 0.85	144.1 ± 2.4	0.1529 ± 3.1	144.5 ± 8.2	0.04906 ± 2.5	0.69	150.6 ± 123
TIT-B,na	0.279	83.62	3.042	83	519	0.772	0.02315 ± 0.73	147.5 ± 2.1	0.1574 ± 2.7	148.4 ± 7.4	0.04931 ± 2.2	0.68	162.6 ± 109
(14) MVK-82-133A, granodiorite, Nicola batholith													
A,na,+149	0.630	556.6	5.885	2385	96	0.142	0.01034 ± 0.30	66.3 ± 0.4	0.06784 ± 0.35	66.6 ± 0.6	0.04761 ± 0.05	0.95	79.7 ± 2.8
B,na,125	5.610	658.0	6.877	1346	1672	0.160	0.01007 ± 0.32	64.6 ± 0.4	0.06597 ± 0.40	64.9 ± 0.6	0.04754 ± 0.10	0.80	76.4 ± 4.8
C,na,90	1.77	728.8	7.734	489	1592	0.178	0.01007 ± 0.35	64.6 ± 0.4	0.06559 ± 0.45	64.5 ± 0.6	0.04723 ± 0.15	0.60	61.0 ± 6.8
(16) MVM-82-176, layered quartzofeldspathic gneiss, Mt. Lytton Complex													
A,na,125,c	1.630	1697	57.09	6706	415	0.093	0.03420 ± 0.35	216.8 ± 1.4	0.2388 ± 0.35	217.5 ± 1.2	0.05065 ± 0.05	0.95	225.1 ± 2.0
B,na,+149	1.220	1644	55.85	3735	292	0.101	0.03432 ± 0.35	217.5 ± 2.0	0.2398 ± 0.35	218.2 ± 1.4	0.05067 ± 0.05	0.95	225.9 ± 2.2
C,+149	0.453	1910	64.43	5451	340	0.102	0.03403 ± 0.10	215.7 ± 0.4	0.2375 ± 0.13	216.4 ± 0.6	0.05061 ± 0.05	0.93	223.2 ± 2.2
D,na,90,c	0.070	1751	58.69	2526	106	0.085	0.03434 ± 0.10	217.6 ± 0.4	0.2399 ± 0.13	218.3 ± 0.6	0.05067 ± 0.07	0.89	225.9 ± 3.0
(17) MC-88-119, granite													
A,+105,cl	0.205	199.1	2.028	2894	10	0.062	0.01068 ± 0.09	68.5 ± 0.1	0.06985 ± 0.11	68.6 ± 0.1	0.04744 ± 0.05	0.89	71.2 ± 2.4
B,+105,cl	0.196	195.8	2.007	3074	8	0.068	0.01069 ± 0.10	68.5 ± 0.1	0.07001 ± 0.11	68.7 ± 0.1	0.04751 ± 0.07	0.79	75.0 ± 3.3
C,+105,cl	0.200	188.8	1.927	2387	11	0.064	0.1068 ± 0.17	68.5 ± 0.2	0.06990 ± 0.18	68.6 ± 0.2	0.04746 ± 0.13	0.73	72.3 ± 6.0

** TIT = titanite; Numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other zircon fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, fr = fragments;
Weighing error = 0.001 mg.
+ Radiogenic Pb.
* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU
# Total common Pb in analysis corrected for fractionation and spike.
++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987). Fractions analyzed prior to 1985 utilized different techniques and very different minimum amounts of sample. The earlier fractions were spiked with a ^{208}Pb-^{233}U-^{235}U tracer, as oppposed to the ^{205}Pb-^{233}U-^{235}U isotopic tracer. Earlier chemical and mass spectrometric procedures follow those of Parrish et al. (1987), except that those analyses preceded the direct propogation of errors and simultaneous use of Faraday cups and electron multiplier.

on hornblende and biotite of 169.3 ± 2.6 Ma and 161.8 ± 2.4 Ma, respectively, (GSC K-Ar 87-208, Hunt and Roddick 1987; Monger, 1989).

Analysis

Three analyses of zircon yield two nearly concordant overlapping points adjacent to 196 Ma and one more discordant point with the same ^{206}Pb-^{238}U age (Fig. 3). These analyses were conducted in 1984 with methods inferior to those presently employed; it is suggested that the non-abraded point A may have either analytical problems or a small amount of inheritance and it is not utilized in the age determination. The other two points are consistent and yield an age interpretation of at least 196 ± 1 Ma, and possibly as old as their ^{207}Pb-^{206}Pb ages of about 204 ± 3 Ma. We suggest an age of 196 ± 1 Ma is the best estimate because of the consistency of the ^{206}Pb-^{238}U ages of all three points.

(3) BROMLEY PLUTON (MV-84-41)(TJgd8-2)

MV-84-41 is a medium grained (biotite) hornblende granodiorite, typical of the Bromley pluton (named by Monger, 1989, for exposures at Bromley Rock Provincial Park, located 7 km northwest of the collecting locality). It is emplaced within the sedimentary facies of the Nicola Group although its western margin touches the eastern volcanic facies of the Nicola Group. From this same locality K-Ar dates on hornblende of 185.7 ± 2.8 Ma and on biotite of 173.4 ± 4.7 Ma were obtained (Monger, 1989).

Analysis

Seven zircon fractions were analyzed from sample MV-84-41 over a period of time from 1986 to 1990. Two of the zircon analyses (D and E) overlap concordia between 193-194 Ma and two others (B and C) are only slightly below concordia (Fig. 4). Analyses of zircon fractions F and G show evidence of inheritance. Fraction A, which is unabraded, has probably undergone some Pb loss. The consistency of points B, C, D, and E implies a zircon crystallization age of 193 ± 1 Ma, which we accept as the age of crystallization of the Bromley Pluton.

(4) PENNASK BATHOLITH (MVM-86-9) (TJgd16-1)

MVM-86-9, "type" Pennask Batholith, is medium grained biotite granodiorite. It is emplaced in mainly sedimentary rocks of the Nicola Group, although its westernmost part intrudes the eastern volcanic facies of that group. From within the same batholith, 6 km due north of the collecting site, 0.8 km west of Rock Lake, K-Ar dates were obtained on hornblende of 159.5 ± 2.4 Ma and on biotite of 144.4 ± 2.2 Ma (Monger, 1989).

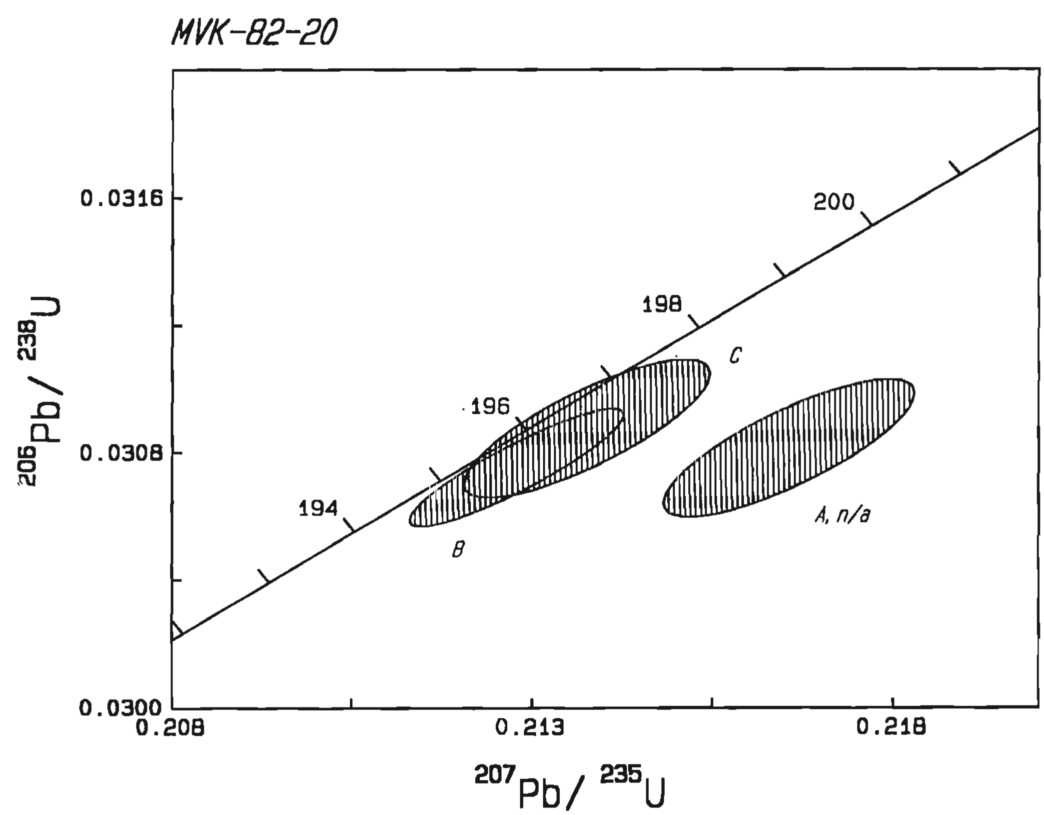

Figure 3. U-Pb concordia plot of MVK-82-20, granodiorite of the Wild Horse Batholith. Error ellipses reflect the 2σ uncertainty.

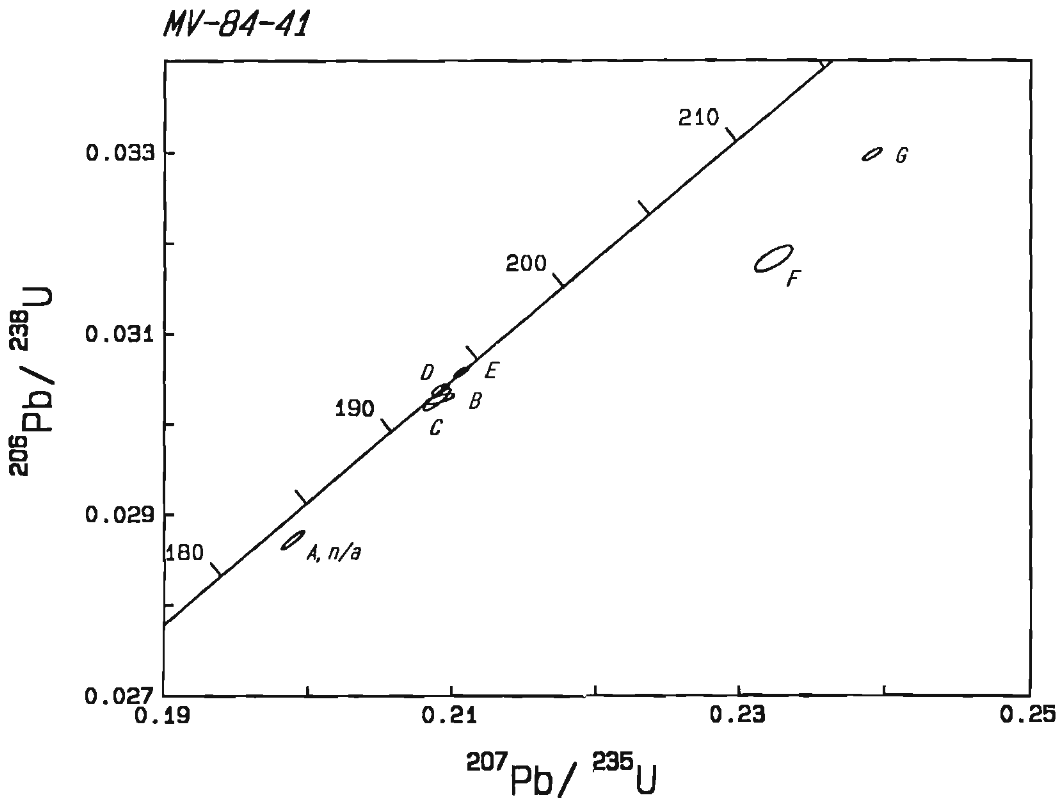

Figure 4a. U-Pb concordia plot of all of the data from sample MV-84-41, the Bromley Pluton. Error ellipses reflect the 2σ uncertainty.

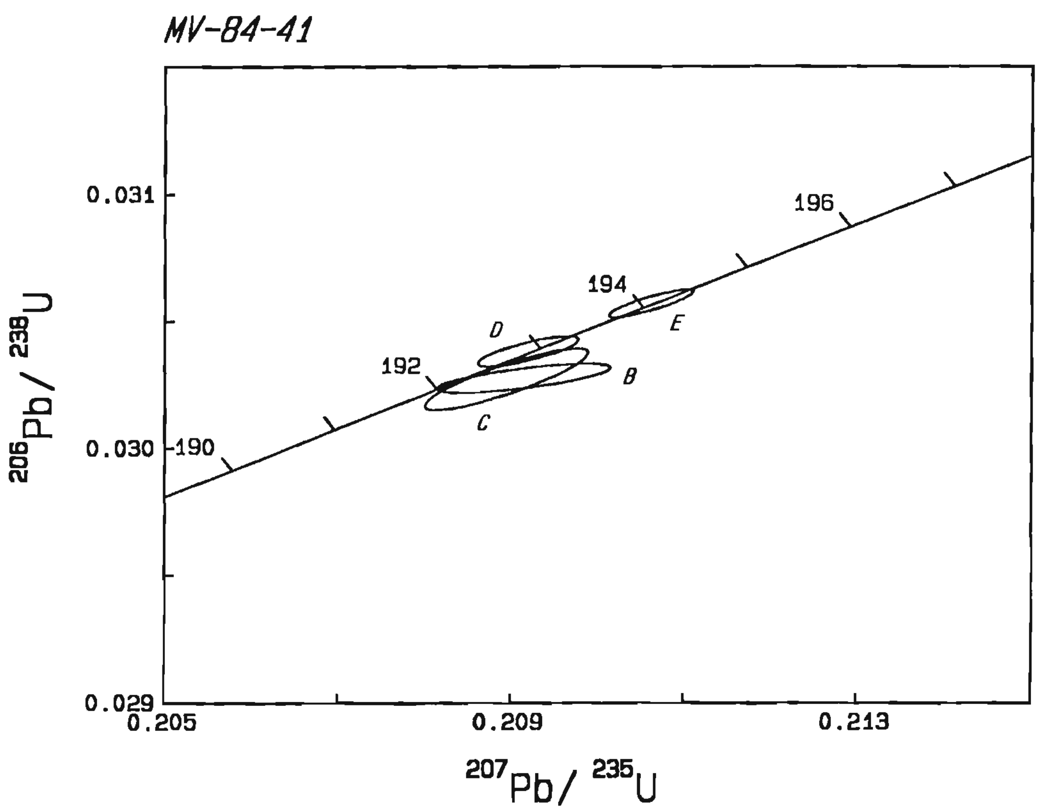

Figure 4b. U-Pb concordia plot of data points from MV-84-41 clustered near concordia. Error ellipses reflect the 2σ uncertainty.

Analysis

Three fractions of fairly clear, equant to elongate zircons were analyzed. A concordant zircon analysis at 194 ± 1 Ma, when coupled with the consistent ^{206}Pb-^{238}U ages of the two points A and C, is taken to represent the age of the granodiorite of the Pennask Batholith (Fig. 5).

Interpretation and discussion

These three Early Jurassic U-Pb dates of 196, 193, and 194 Ma are believed to represent crystallization ages for this suite of plutons. An additional body belonging to the suite is the Douglas Lake Stock, located between Pennask and Wildhorse plutons, which yields a K-Ar date on biotite of 183.3 ± 2.7 Ma (GSC Paper 87-2, p. 194; Monger, 1989). It appears that with one exception, noted above, all K-Ar dates from this suite are partly reset, perhaps in response to intrusion of the Middle Jurassic Osprey Lake Batholith discussed below, which separates Pennask and Bromley bodies, or by younger mid-Cretaceous and early Tertiary intrusions in the region (Fig. 1).

Although the Early Jurassic granitic rocks were grouped by Monger (1989) with latest Triassic plutons (e.g. Guichon Creek and Mount Lytton described above), their spatial location relative to facies of the Nicola Group, and the dates presented herein, suggest that they are a discrete suite of significantly younger intrusions. They probably belong to the same Late Triassic-earliest Jurassic magmatic episode as the Nicola-Rossland volcanic rocks. Unlike the older granitic plutons, located near the western margin of Quesnellia, these are all located within the eastern sedimentary rocks of the Nicola Group, close to its contact with the eastern volcanic facies of the Nicola. The western and eastern suites of granitic (calc-alkaline) plutons are separated by a suite of small, somewhat alkaline, ultramafic to syenitic bodies (e.g. Iron Mask, Copper Mountain, Tulameen plutons), which give (partly reset?) K-Ar dates ranging from 204 to 180 Ma, and are spatially associated with the eastern volcanic facies of the Nicola. Mortimer (1987) noted that the geochemistry of the Nicola Group, which features low potassium lavas in the west and high potassium ones in the east, indicates that these rocks formed above an east dipping subduction zone. He also noted that the volcanic chemistry is paralleled by chemical trends of the intrusive rocks, although ages and spatial distribution are not.

We propose the following hypothesis to explain the observed space-time relationships of these magmatic rocks. The Nicola arc was initiated as a west-facing arc in late Karnian time, and continued through the remainder of the Triassic in more-or-less the same configuration as a relatively narrow arc (present width 50-70 km). Latest Triassic rocks such as the Guichon Batholith are plutonic equivalents of younger, now eroded, parts of the western (low potassium) volcanic facies and the syenitic to ultramafic plutons (e.g. Iron Mask) are plutonic equivalents of the eastern (high

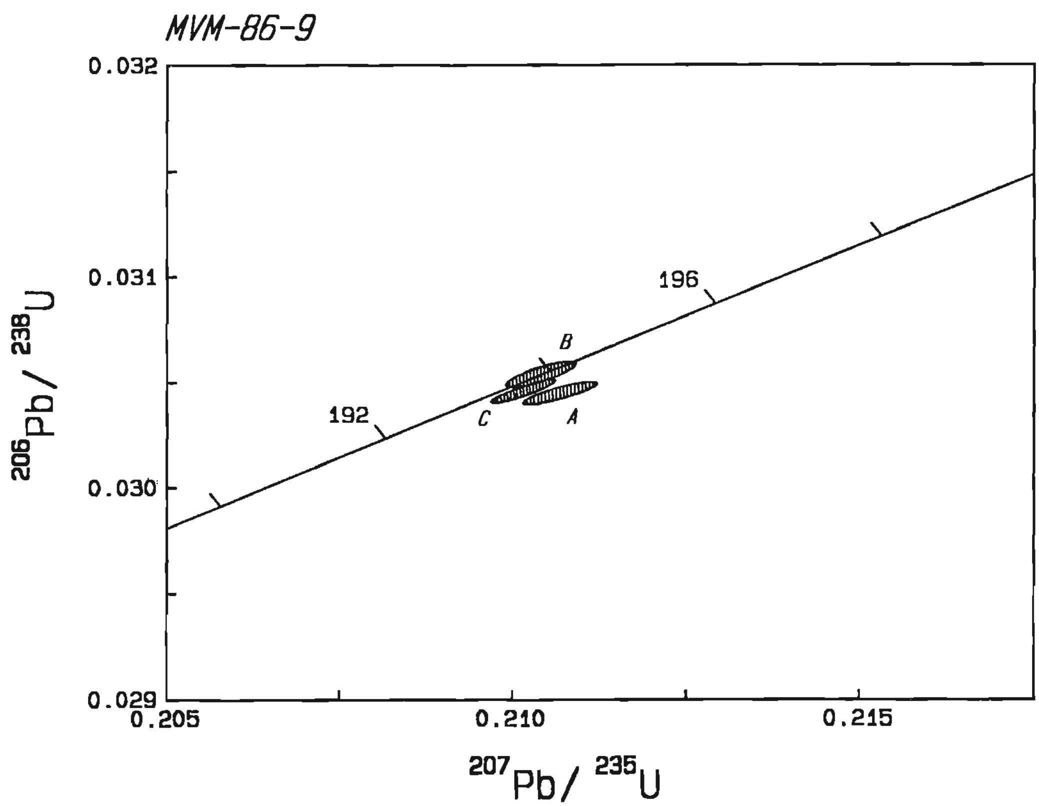

Figure 5. U-Pb concordia plot of MVM-86-9, granodiorite of the Pennask Batholith. Error ellipses reflect the 2σ uncertainty.

potassic) volcanic facies. In the Early Jurassic, calc-alkaline western magmatism, represented by the Wild Horse, Pennask and Bromley bodies, migrated to a position immediately east of the Triassic arc, and eastern magmatism, represented by Early Jurassic (Sinemurian? and older; ca. 200 Ma) Rossland volcanics, moved far to the southeast. Their present location, is 150-200 km east of the Triassic high potassium rocks. These volcanic rocks remain considerably to the east, even after removing 100 km of Eocene extension. A reason for this change may be that there was considerable flattening of the east-dipping subduction zone between Triassic and Jurassic time, causing the arc to migrate eastward and expand in width. Subsequently, pre-Late Pliensbachian (ca. 195 Ma) erosion cut deeply in many places into the magmatic arc in many places. By 185 Ma the early Mesozoic Quesnellian off-shore arc was thrust on to continental margin rocks (Archibald et al., 1983; Parrish and Wheeler, 1983).

(5) OSPREY LAKE BATHOLITH (MV-84-5)(1Jg9-4)

Specimen MV-84-5 is a coarse grained (hornblende) biotite quartz monzonite, with large pink feldpsar phenocrysts and local inclusions, which is a typical lithology of the Osprey Lake Batholith (new name: from Osprey Lake which is located 15 km northeast of sampling site; Monger, 1989). K-Ar dates of 154 ± 3 Ma on hornblende (GSC K-Ar 88-20) and 149 ± 2 on biotite (GSC K-Ar 88-21, Hunt and Roddick, 1988) were obtained from this locality.

Analysis

Three zircon fractions analyzed from this sample plot in a short linear array below concordia (Fig. 6). Two of the zircon fractions are unabraded (B,C) and the third fraction is only moderately abraded (A). The zircons have moderately high U concentrations and have lost some Pb. The discordance suggests that surface-correlated Pb loss was not entirely removed. We infer that the best estimation for the age of the Osprey Lake Batholith is from the mean of the ^{207}Pb-^{206}Pb ages, that is 166 ± 1 Ma (Table 1).

Interpretation and discussion

The batholith is surrounded on three sides by Early Jurassic granitic rocks (Pennask, Bromley bodies), and part of its northwestern margin intrudes Nicola volcanic rocks. Peto and Armstrong (1976) recognized all of these granitic bodies as a composite batholith composed of at least two genetically distinct magma suites. The ages available to these workers were not sufficiently precise to separate the two suites. We regard the U-Pb date of 166 Ma from this body as a crystallization age, and feel that the body belongs to a completely different magmatic episode, unrelated to early Mesozoic arc magmatism, and emplaced subsequent to accretion of Quesnellia to the ancient continental margin.

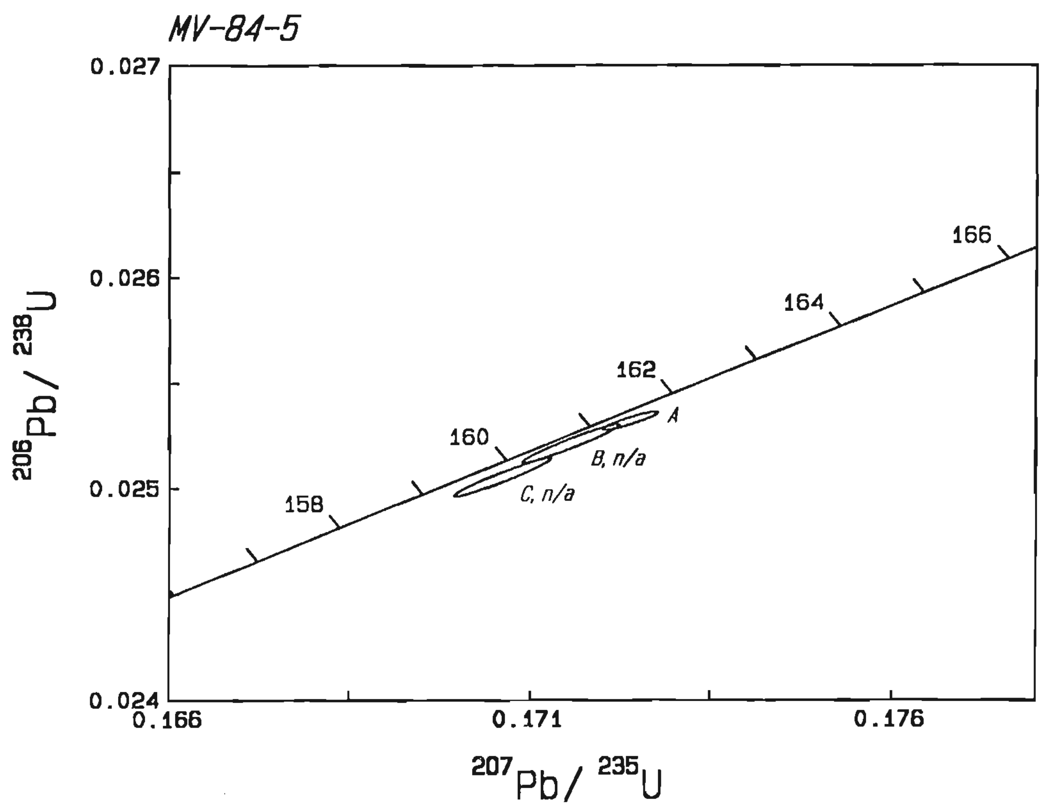

Figure 6. U-Pb concordia plot of MV-84-5, the Osprey Lake Batholith. Error ellipses reflect the 2σ uncertainty.

K-Ar and Rb-Sr dates ranging from 145 to 164 Ma determined from other parts of the batholith (listed in Monger, 1989) may reflect partial re-setting of original cooling ages by widespread mid-Cretaceous granitic rocks, common to the southwest, and by the small early Tertiary intrusions in the northern part of the batholith (loc.12).

(6) CLOUDBURST QUARTZ DIORITE (MV-89-157)

Specimen MV-89-157 is an altered, massive but locally fractured (biotite) hornblende granodiorite, which is part of an extensive Late Jurassic granitic terrane in the southwestern Coast Belt (Mathews, 1958; Friedman and Armstrong, 1990; Monger, 1991a). The granite is overlain to the north of this locality by the basal granite-bearing conglomerate of the Early Cretaceous Cheakamus Formation, a sedimentary facies of the dominantly volcanic Gambier Group. It is partly intruded by, and partly faulted against mid-Cretaceous granitic rocks (Fig. 1).

Analysis

Two fractions of clear, equant, euhedral zircons were analyzed from this sample. These data points, although slightly below concordia due to deficit of ^{206}Pb (see section on analytical methods) indicates a precise age of 147.0 ± 0.5 Ma (Fig. 7), which is interpreted to be the age of the Cloudburst quartz diorite.

Interpretation and discussion

The age of 147 Ma (latest Jurassic) is regarded as a crystallization age, and supports a preliminary U-Pb date of 145 ± 2 Ma from about 20 km to the north in type Cloudburst quartz diorite near Cheakamus Canyon (Friedman and Armstrong, 1990). Preliminary U-Pb dating by these authors shows that the Cloudburst is part of an extensive suite of Late Jurassic (ca. 173-145 Ma) plutonic rocks widespread in the southwestern Coast Belt. On its west side these granites clearly intrude Wrangellian rocks widely exposed to the west on Vancouver Island, and on the east intrude the sequence of rocks exposed on the west side of Harrison Lake (Friedman and Armstrong, 1990; Monger, 1991a). The Late Jurassic and Early Cretaceous magmatic rocks appear to be terrane specific, and restricted to a crustal block in the southwestern Coast Belt that in early Late Cretaceous time (96-92 Ma) interacted with terranes to the east, to produce the major Late Cretaceous "two-sided orogen" whose axis lies east of Harrison Lake (Journeay, 1990; Monger 1991b).

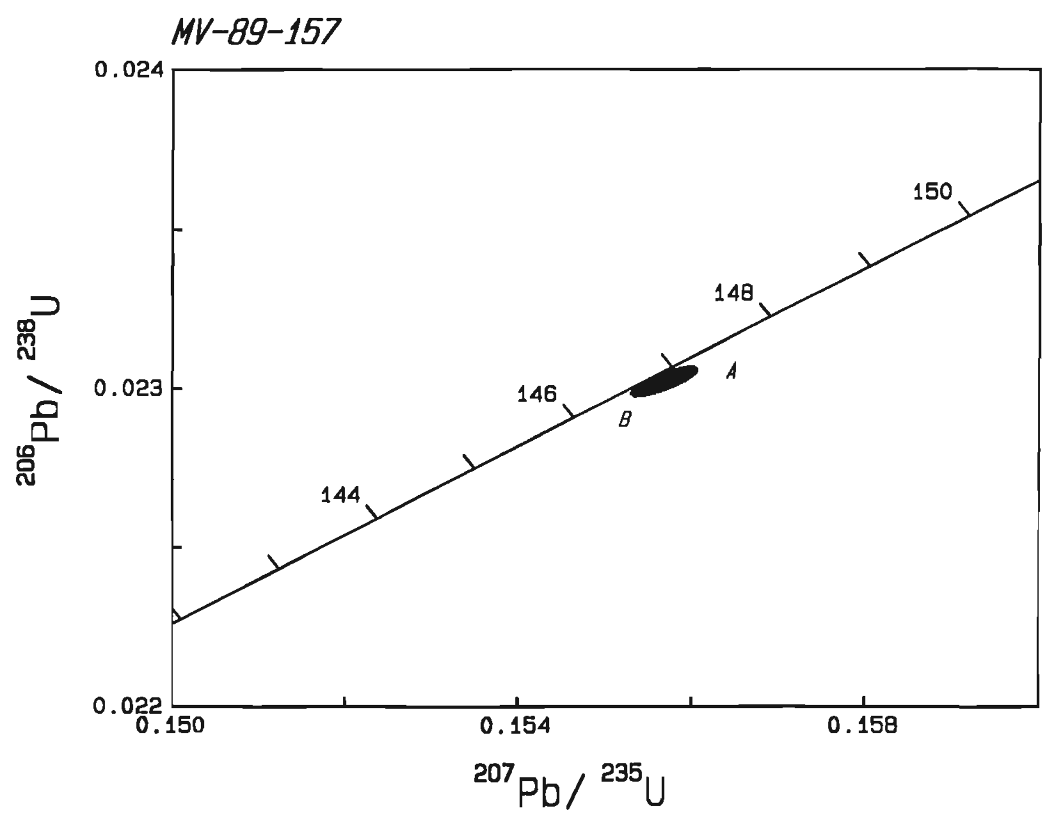

Figure 7. U-Pb concordia plot of MV-89-157, the Cloudburst quartz diorite. Error ellipses reflect the 2σ uncertainty. The data are not corrected for deficit of ^{206}Pb (see text).

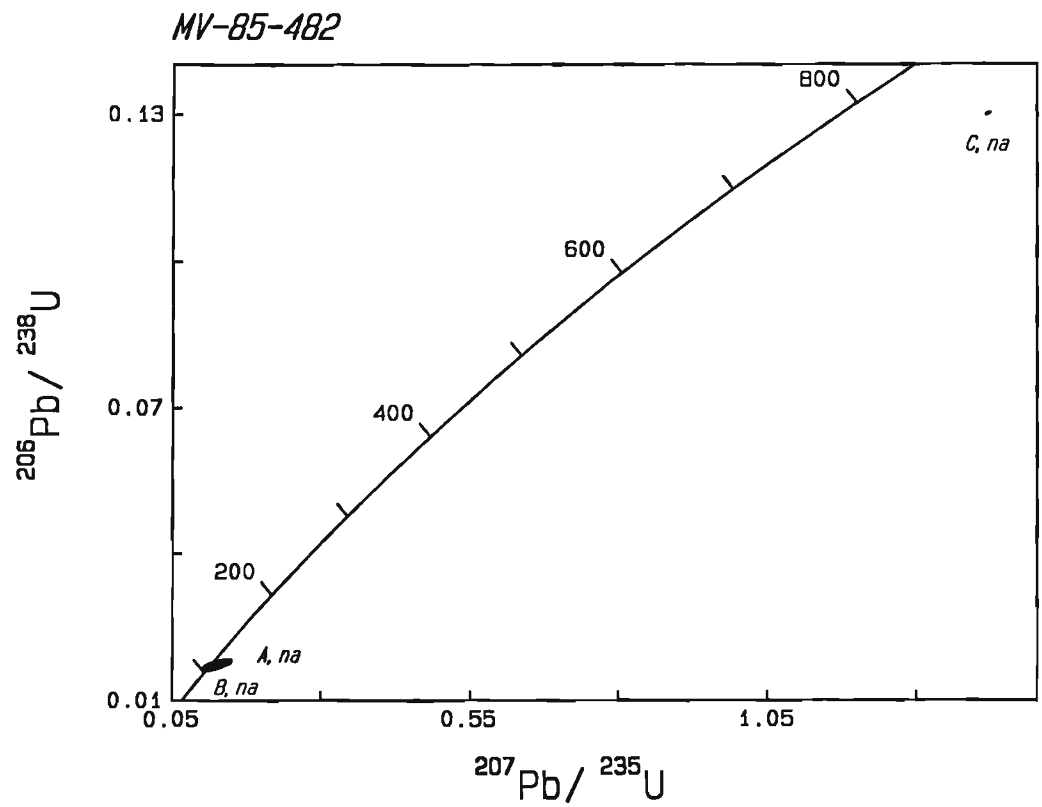

Figure 8a. U-Pb concordia plot of all the data from MV-85-482, quartz-feldspar crystal tuff of the Brokenback Hill Formation. Error ellipses reflect the 2σ uncertainty.

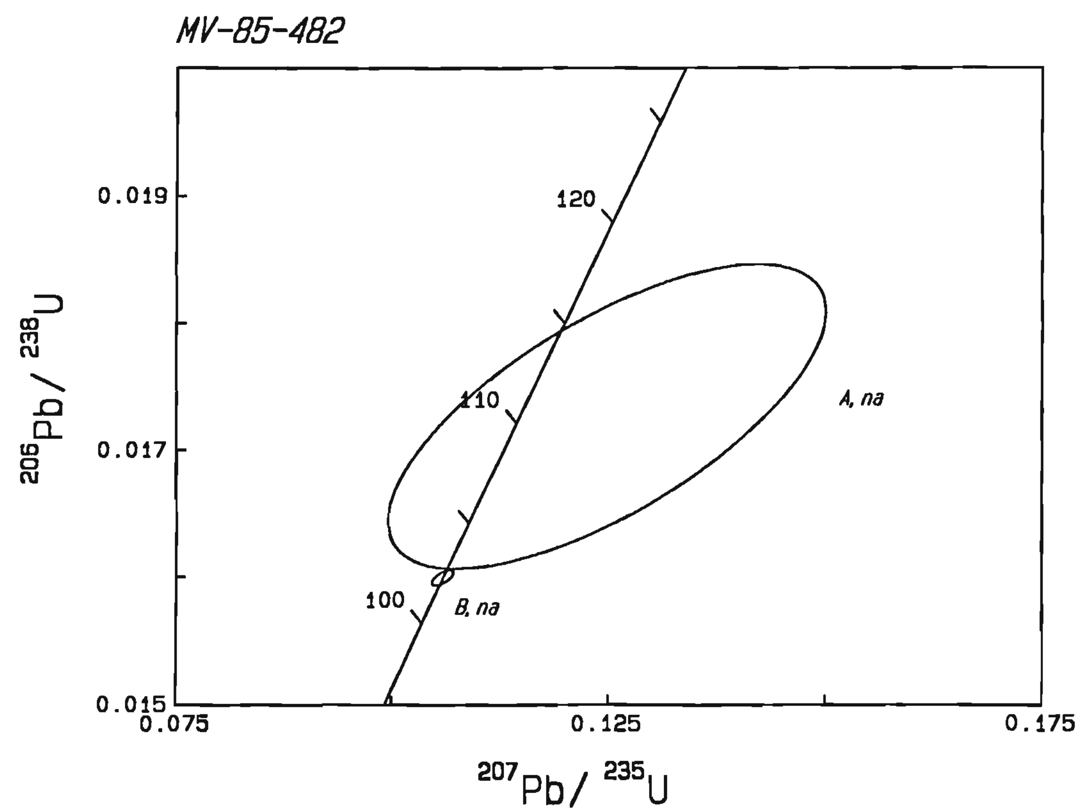

Figure 8b. U-Pb concordia plot of data points from MV-85-482 plotting on concordia. Error ellipses reflect the 2σ uncertainty.

(7) BROKENBACK HILL FORMATION(?) (MV-85-482)(JKb12-1)

MV-85-482 is a massive, weakly foliated, subgreenschist facies crystal lithic tuff, with noteable feldspar and quartz crystals, which is separated from the higher metamorphic grade Slollicum Schist to the east by a major fault, mapped by Monger (1989) as the Harrison Fault. West of Harrison Lake, Early Cretaceous (Berriasian to Albian) volcanic and sedimentary rocks are widespread and variously named as Peninsula and Brokenback Hill formations, Fire Lake and Gambier groups (Arthur, 1986; Ray, 1991; Monger, 1991a), and they have paleontological control on their ages. As a result of the U-Pb age of this sample, it is correlated with the Brokenback Hill formation.

Analysis

The data from two fractions of clear, small, equant to prismatic crystals from this sample overlap concordia (fractions A, B; Fig. 8). Fraction A had a high level of common Pb due to a high blank or a non-zircon particle in this multigrain fraction of very small crystals. The age of fraction B, 102 ± 1 Ma, is interpreted to represent the age of the tuff. Fraction C is composed of pinkish euhedral crystals with slightly rounded edges, inclusions, and fractures. The data for this fraction reveal an inherited component with a minimum age of roughly 1175 Ma, although this component is likely to be somewhat older and a mixture of various ages.

Interpretation and discussion

The age of 102 Ma is regarded as the time of crystallization and eruption of these volcanic rocks. It confirms correlation with the upper, locally dacitic, part of the Early Cretaceous succession mapped in the Doctors Point area by Ray (1991). The Precambrian zircons in the tuff suggest some interaction with older crustal material, a feature uncommon in the Coast Belt. It is notable that this older component is so far only found in some of the younger magmatic rocks.

(8, 9, 10) MID-CRETACEOUS GRANITIC AND GNEISSIC ROCKS OF THE SOUTHEASTERN COAST BELT

(8) GNEISSIC QUARTZ DIORITE (MVH-85-92)

MVH-85-92 is a foliated greenish altered hornblende quartz diorite, with oriented hornblende giving the rock a fabric. It forms a small (2x1 km) body within enclosing upper greenschist-lower amphibolite (garnet-biotite) grade

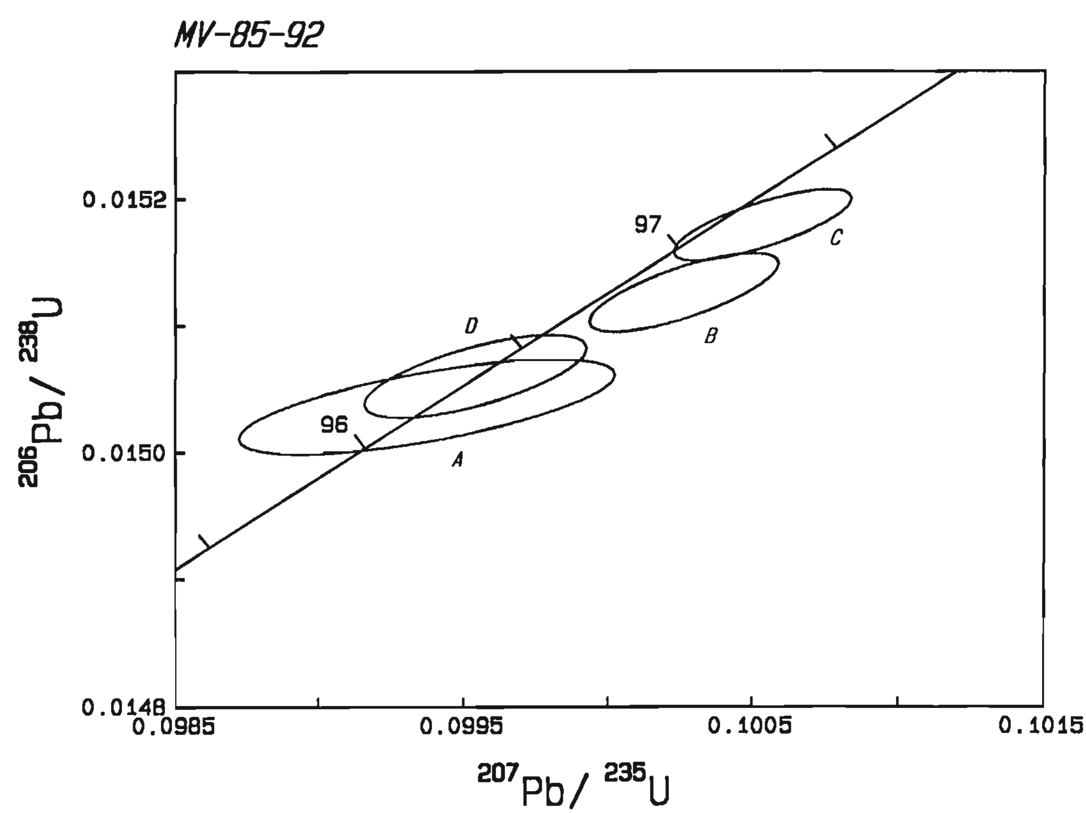

Figure 9. U-Pb concordia plot of MVH-85-92, gneissic quartz diorite. Error ellipses reflect the 2σ uncertainty.

Slollicum Schist, and immediately southwest of, and structurally below alpine-type ultramafics associated with the Cogburn Schist (Monger, 1989).

Analysis

Four zircon fractions were analyzed. Two of these fractions (A, D; Fig. 9) overlap concordia at about 96 ± 1 Ma. The other two zircon fractions may contain a small inherited component. The high uranium content of these zircons (>1000ppm) suggests the possibility of Pb loss; this sample may be slightly older, perhaps 98 Ma or so. We suggest that 97 ± 1 Ma is a good estimate of the crystallization age.

Discussion and interpretation

This body was erroneously shown by Monger (1989) as being of early Tertiary age, largely because of its location within Slollicum Schist, which contains K-Ar dated Tertiary granites 8 km south of this body. Its foliated character was recognized as distinctive, and near Harrison Lake it contains a northeast plunging linear fabric, characteristic of Slollicum Schist and probably early Late Cretaceous in age (Monger, 1986; J.M. Journeay and R.M. Friedman, unpublished data). It is probably the south-southeast continuation of the Breakenridge orthogneiss (discussed below), which is exposed 8 km to the north-northwest, but separated from it by a fault extending along Big Silver Creek.

(9) MAFIC GNEISS (MV-86-343)(Kgn12-1)

Orthogneiss mapped in the Mount Breakenridge area ranges in composition from granodiorite to diorite. It is bounded by west-vergent thrust faults which generally juxtapose it against high-grade metamorphic rocks, and forms the core of the Breakenridge Antiform, which refolds earlier thrusts (Journeay, 1990). MV-86-343 is a dark, medium grained, streaky, hornblende quartz dioritic to dioritic gneiss typical of the northeastern part of the Breakenridge Antiform. About 0.3 km to the east, it is juxtaposed on a steeply dipping, layer parallel, refolded thrust fault against biotite-garnet grade Slollicum Schist. From a locality 1.75 km north-northeast of this sampling site, mica schist with dextral shear fabrics gave a 5-point feldspar-muscovite Rb-Sr date of 93.5 ± 11.4 Ma (see sample 15 below). From a location about 8 km south of MV-86-343, Gabites (1985) obtained a U-Pb date of 105 ± 2 from a more felsic phase of the Breakenridge orthogneiss.

Analysis

Three fractions of equant, subhedral zircons were analyzed from this sample. Two of the fractions intersect concordia at 96 ± 1 Ma, which is interpreted to be the age of crystallization of the quartz diorite gneiss. The third zircon analysis shows evidence of some inheritance, which could be of Precambrian age similar to sample 7 (see above).

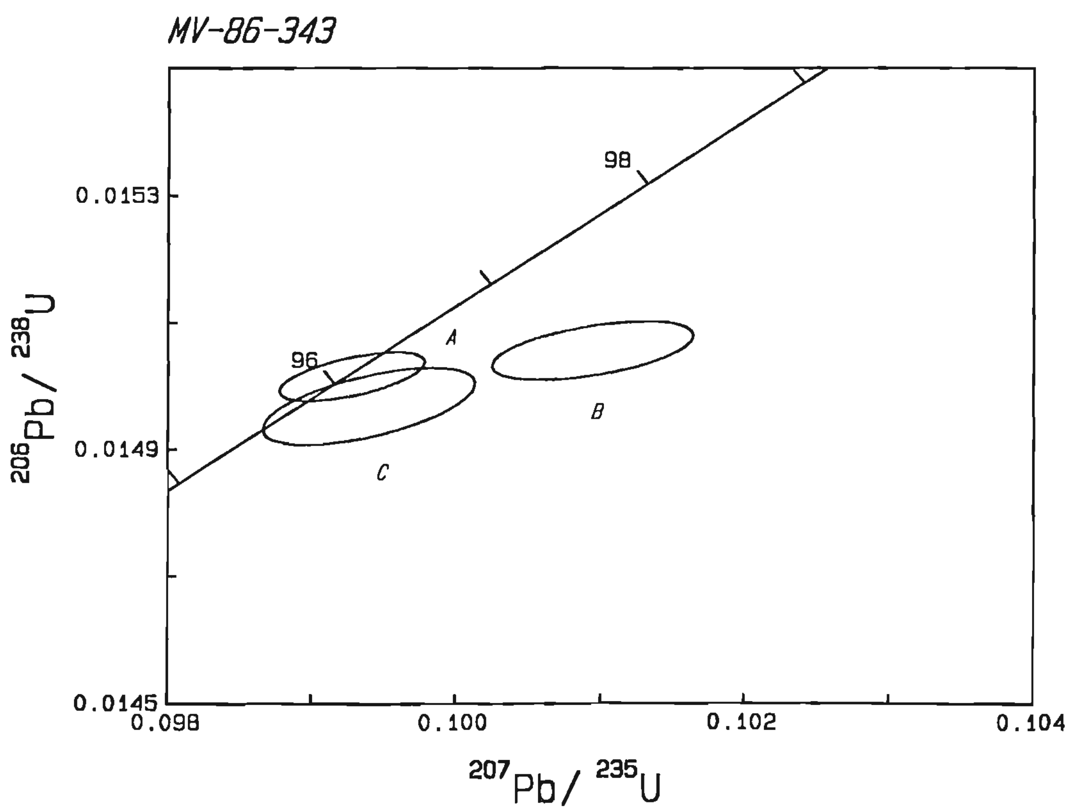

Figure 10. U-Pb concordia plot of MV-86-343, mafic Breakenridge gneiss. Error ellipses reflect the 2σ uncertainty.

Interpretation and discussion

The U-Pb dates represent time of crystallization, which took place immediately preceeding and during contractional deformation and associated metamorphism up to upper amphibolite facies. Together with less-foliated quartz diorites such as the Spuzzum quartz diorite, the Slollicum, Cogburn and Settler schists, and complex thrust faults, these rocks characterize the mid- to Early Late Cretaceous (96-92 Ma) metamorphic, plutonic and structural culmination of the southern Coast Mountains (Monger, 1989; Journeay, 1990; J.M. Journeay and R.M. Friedman, unpublished data). To the south the metamorphic core of the Cascade Belt in Washington State exhibits similar characteristics (see Tabor et al., 1989). The two are dextrally offset by the late Eocene Fraser-Straight Creek Fault (Fig. 1).

(10) SNOWCAP LAKE PLUTON (MVD-89-375)

MVD-89-375, from south of Snowcap Lake, is a medium grained biotite granodiorite-quartz diorite that at this locality has no fabric but nearby has weak gneissosity dipping northeast at 65-80°. North of Snowcap Lake its apparent continuation is locally agmatitic and it intimately intrudes granite cobble and boulder conglomerate, with variably flattened clasts, that is probably correlative with the basal Cretaceous conglomerate mentioned in (6) above. On the east it is in sharp, fault(?) contact and overlain by biotite hornblende schist, within which are metamorphosed volcanic breccias with flattened clasts locally preserved, which could either be correlated with Early Cretaceous volcanics or Jurassic volcanics preserved along Harrison Lake. The southward extent of the pluton is not known, although a preliminary U-Pb date of 94 Ma was obtained from near Thomas Lake, 35 km to the south-southeast (R.M. Friedman, reported in Monger, 1991a).

Analysis

The sample contained clear, high quality, elongate prismatic zircons. Fraction C overlaps concordia at 94.0 ± 0.5 Ma and all three fractions have the same 94 Ma ^{206}Pb-^{238}U age (Fig. 11), which we interpret as the age of the Snowcap Lake Pluton. The reason for the dispersion of points both to the left and right of concordia probably results from some difficulties encountered in accurate common Pb corrections when this sample was analyzed. These problems were limited to samples with less than 300pg of Pb. The consistency of ^{206}Pb-^{238}U ages, however, strongly suggests 94 Ma is the correct crystallization age.

Interpretation and discussion

The U-Pb date records the time of crystallization of this body. It intrudes probable Lower Cretaceous clastic rocks, and is involved in deformation that is congruent with 92-96 Ma contraction reported from elsewhere in the region (Monger, 1991; J.M. Journeay and R.M. Friedman, unpublished data). Intrusion must have immediately preceeded deformation.

(11) SCUZZY PLUTON (MV-86-431)(lKgd13-4)

MV-86-431 is white weathering, non-foliated biotite quarz monzonite, with minor amounts of muscovite, typical of the Scuzzy Pluton. From elswhere in the same body, K-Ar dates on biotite range from 72 to 79 Ma (Monger, 1989). To the south, the Scuzzy Pluton intrudes locally foliated, more mafic rocks that have been correlated with the Spuzzum Pluton, that in turn both intrudes and is faulted against Settler Schist (Monger, 1989). To the northeast, near Kwoiek Creek and Natatlatch River, it intrudes amphibolite facies schists and greenschist grade rocks of the Bridge River Complex.

Analysis

Four fractions of large, clear, euhedral zircons were analyzed from this sample. Two of the fractions intersect concordia at about 84 ± 1 Ma and the two other fractions contain inherited components (Fig. 12).

Interpretation and discussion

The 84 Ma date represents a crystallization age, and the younger K-Ar dates represent cooling ages. Journeay (1990) noted that P-T conditions recorded in schist on the northeast side of the Scuzzy Pluton indicate that emplacement of the pluton was at 10-12 km depth, which contrasts with mid-early Late Cretaceous (ca.96-92 Ma) "peak" metamorphism at depths of 20-30 km in the region east of Harrison Lake. Therefore, significant (10-20 km) uplift occurred in the region in between 96 and 84 Ma. Also the cross-cutting nature of the Scuzzy pluton indicates that all west-vergent deformation was over by 84 Ma. The presence of considerably older zircon inheritance is notable, and again, common only in the younger plutons of the Coast Belt.

EARLY TERTIARY GRANITIC ROCKS

(12) (MV-86-99)(eTg16-1)

MV-86-99 is a distinctive porphyry with K-spar phenocrysts up to 5 cm long, rounded quartz eyes, up to 1 cm in diameter and minor fresh biotite in a fine grained matrix. It forms a series of small east-northeast trending elongate intrusions (ranging from a maximium size of 7x3 km down to dykes a few metres in width) near and across the contact of Pennask and Osprey Lake batholiths. This specimen is from a large dyke emplaced within Pennask biotite granodiorite. Another body, located near Siwash Creek, 18 km to the west, gave K-Ar ages on biotite of 52-53 Ma (in Monger, 1989).

Analysis

Two zircon fractions show Precambrian inheritance with a lower intercept of the two-point array of 62.5 ± 0.2 Ma. The least discordant point A is quite close to concordia, although a more realistic estimate of uncertainty is perhaps 62 ± 2 Ma.

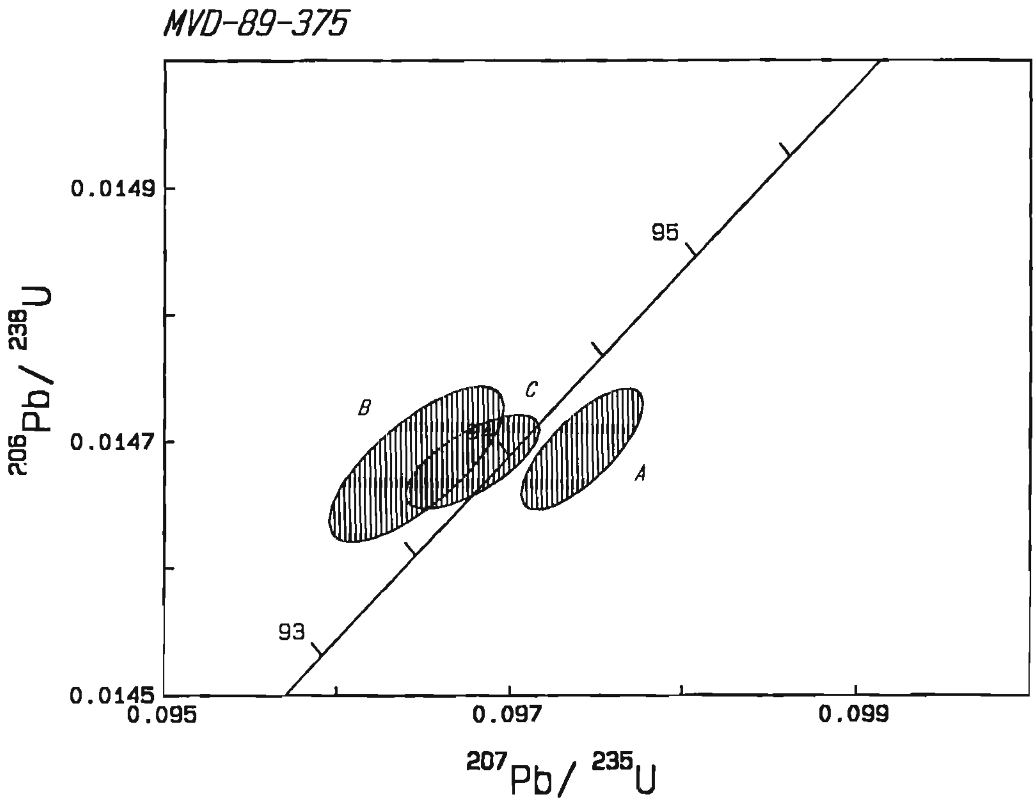

Figure 11. U-Pb concordia plot of MVD-89-375, the Snowcap Lake Pluton. Error ellipses reflect the 2σ uncertainty.

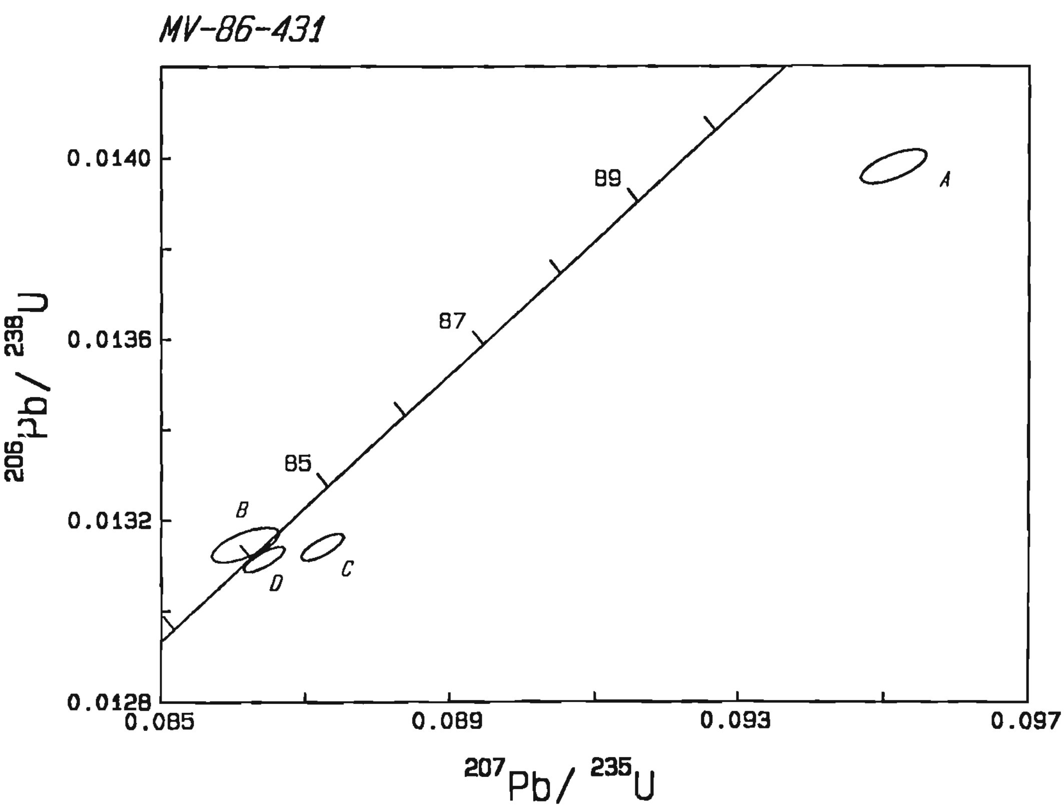

Figure 12. U-Pb concordia plot of MV-86-431, biotite quartz-monzonite of the Scuzzy Pluton. Error ellipses reflect the 2σ uncertainty.

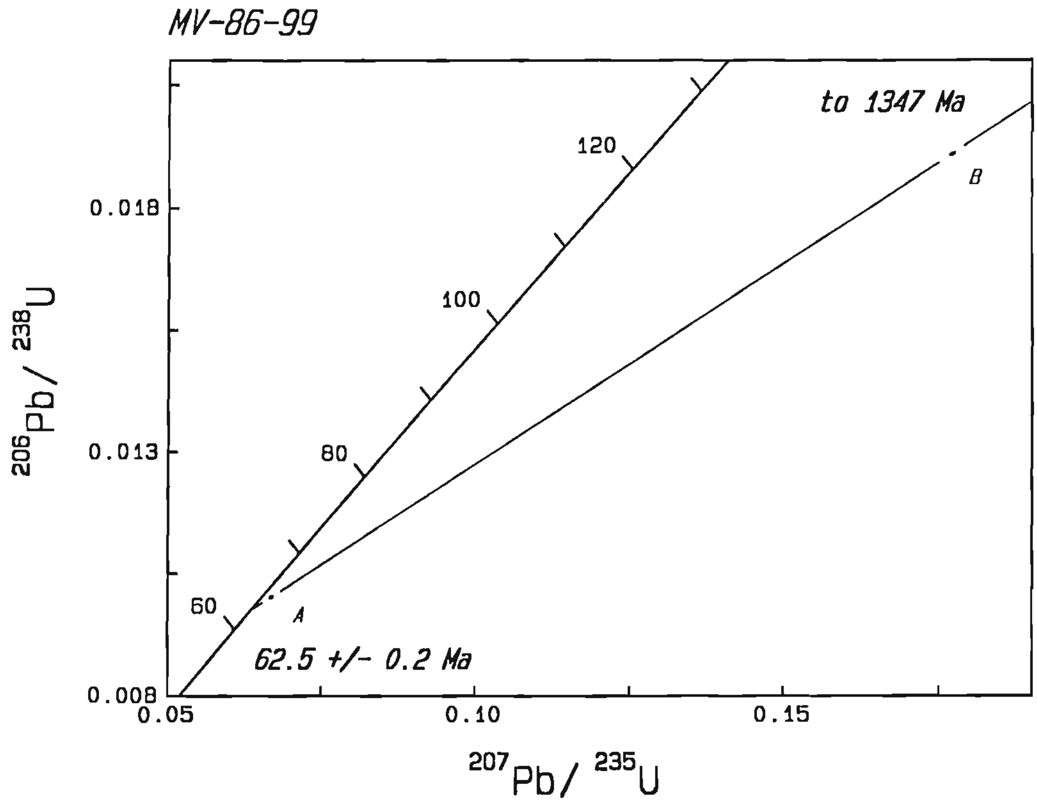

Figure 13. U-Pb concordia plot of MV-86-99, quartz-felspar porphyry. Error ellipses reflect the 2σ uncertainty.

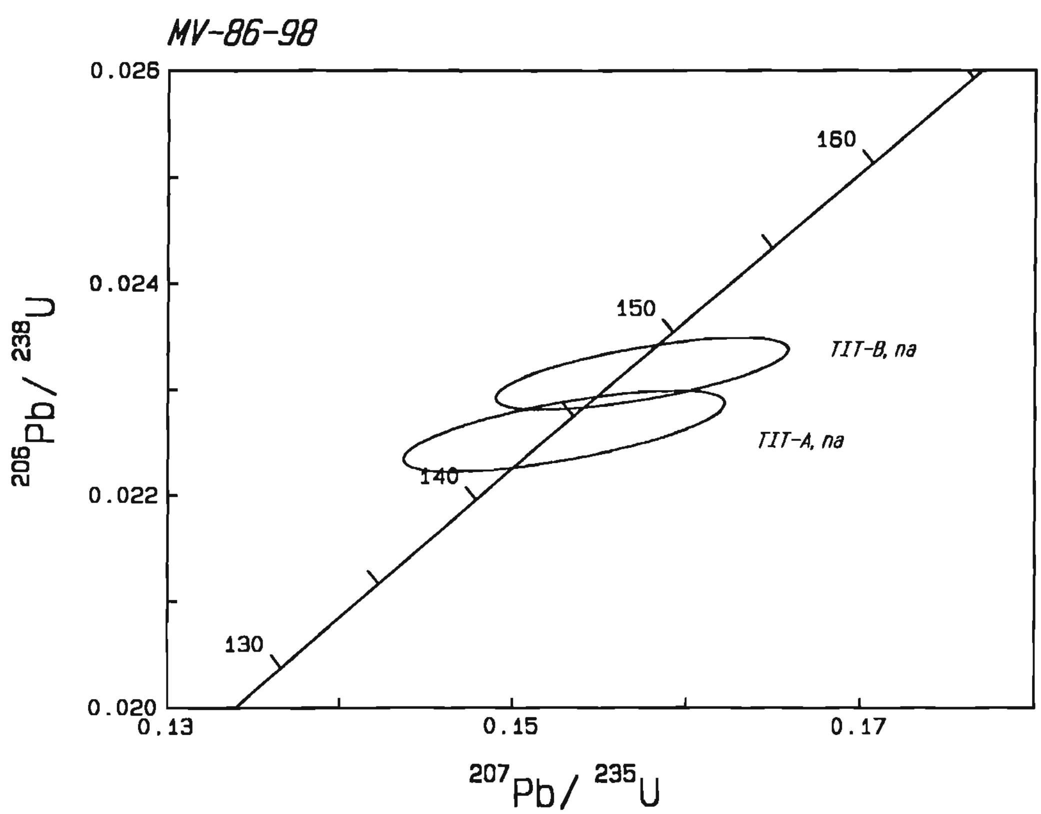

Figure 14. U-Pb concordia plot of MV-86-98, granodiorite of the Pennask Batholith. Error ellipses reflect the 2σ uncertainty.

ADDITIONAL SAMPLES

(13) PENNASK BATHOLITH (MV-86-98)

MV-86-98 is from typical Pennask batholith, composed of medium grained biotite granodiorite. It is about 900m east of sample 12 (63 Ma, MV-86-99) and about 2km north of the contact with the Osprey Lake pluton (sample 5, MV-84-5) dated at 165 Ma. Only titanite was dated from this sample.

Analysis and Interpretation

Two titanite determinations overlap concordia between 140 and 150 Ma but only barely agree within analytical uncertainty. Since this rock is from the typical Pennask batholith similar to sample 4, its crystallization age is inferred to be ca.195 Ma. The younger titanite ages can be explained by the fact that this sample is about 900m east of the younger 63 Ma phase within the Pennask (sample 12; Fig. 1) and only 2 km north of the ca. 165 Ma Osprey Lake phase. Both are suggested to have partially reset the titanite U-Pb system, the closure temperature of which is about 600°C (Heaman and Parrish, 1991). Thus the ca. 145 Ma ages have no particularly useful geological significance. The errors on titanite are sufficiently large as to preclude a more definite interpretation of this sample's thermal history.

(14) NICOLA BATHOLITH (MVK-82-133A)

MVK-82-133A is a sample of biotite granodiorite of the Nicola Batholith which lies east of the Coldwater fault northeast of Merritt. Most of the batholith is composed of this Tertiary age body, whereas a minor part, mainly in the north, is of Late Triassic age (Fig. 1). These rocks form the core of the Nicola "horst", an uplifted block of granitoid and metamorphic rocks with extensional faults bounding it on both sides.

Analysis

Clear, euhedral unabraded zircons were analyzed from this sample. A concordant zircon analysis (C, Fig. 15) at 64.3 ± 0.3 Ma is interpreted to be the age of the biotite granodiorite of the Nicola batholith. Fraction B is nearly concordant and in agreement with fraction C, and fraction A probably contains a small inherited component of probably earlier Mesozoic age.

Interpretation

This date of 64 Ma indicates significant Paleocene plutonism within the southern Intermontane Belt, and provides an explanation of younger K-Ar ages in the metamorphic rocks of the Nicola "horst".

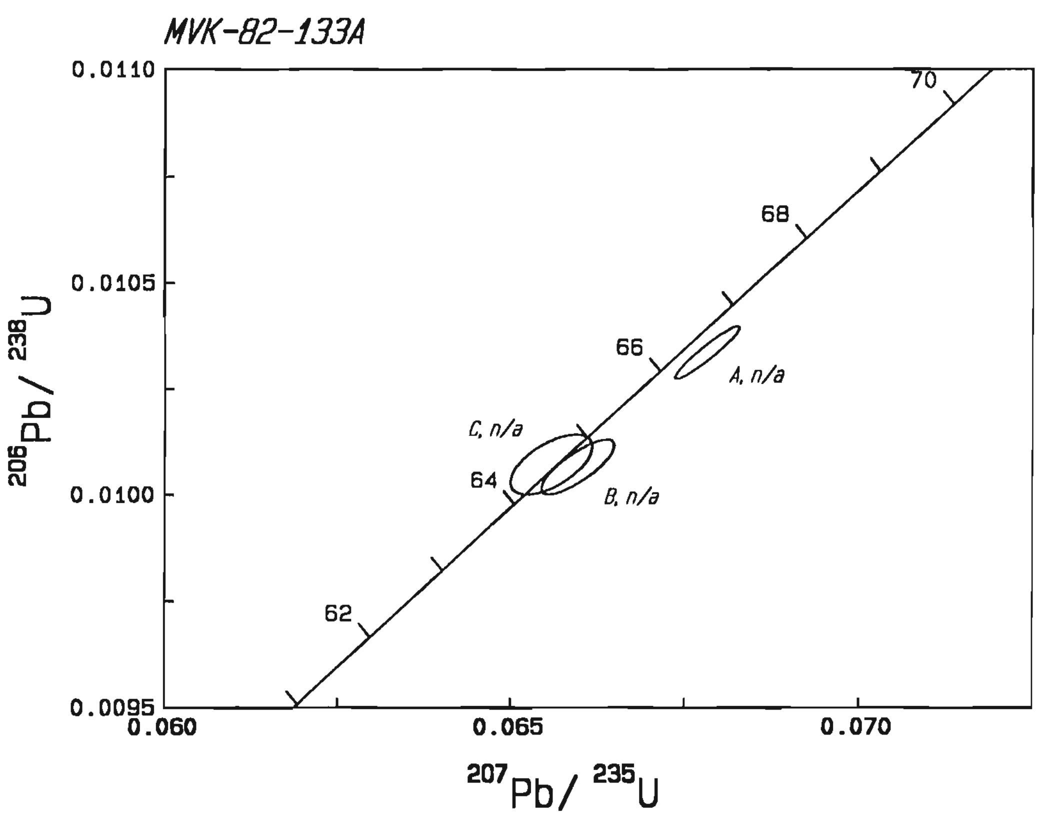

Figure 15. U-Pb concordia plot of MVK-82-133A, granodiorite of the Nicola Batholith. Error ellipses reflect the 2σ uncertainty.

(15) MUSCOVITE FROM SHEARED SLOLLICUM SCHIST (MV-86-344)

MV-86-344 is a sample of muscovite-rich quartzofeldspathic schist at Big Silver Creek on the east side of Harrison Lake. The sample is in the footwall of the Harrison fault, and has a distinctive sheared structural fabric. Excellent mesoscopic S-C fabric is developed with muscovite concentrated in the zones of shear. Muscovite is interpreted to have crystallized synkinematically with the dextral shear, and not just rotated by the shear, during greenschist-facies metamorphic conditions. The rock resembles muscovite-bearing mylonitic rocks described and dated by Parrish et al. (1988) on the east side of Valhalla Complex. The objective of dating the sample was to determine the age of the muscovite and by inference, the dextral shear zone. The dextral shear fabric is developed in metre-size zones of shear which dip >80° and trend north-northeast. It is suggested that these dextral shear zones are related to other dextral shear zones in the Harrison Lake area.

Analysis

Feldspar and muscovite were separated from this rock, and multiple analyses of both minerals produced only minor dispersion in Rb-Sr ratio, with feldspar having the higher ratio. A linear regression of these 5 data points has an MSWD of 1.62, an age of 93.5 ± 11.4 Ma (2 standard errors of the mean) and an initial ratio of 0.70346 ± 44 (Table 2). Analytical methods for Rb-Sr are the same as those described in Parrish et al. (1988) and were performed by R. J. Thériault at the Geological Survey of Canada.

Interpretation

The Rb-Sr age is interpreted as the age of development of metamorphic muscovite developed within the shear zone, and we interpret this as the age of shearing. Unfortunately the limited dispersion in Rb-Sr leads to a large error of 11 Ma, allowing this shearing to have occurred from 104-82 Ma. Since the shear zone appears to cut other metamorphic fabrics and postdate the peak thermal conditions, it is interpreted as occurring late in the ca. 96-92 transpressional event recorded in the Harrison Lake area. Therefore, it is likely that the age constraints on the formation of the muscovite are ca. 94-82 Ma, since the main period of peak thermal conditions was in the range from 96-92 Ma (J.M. Journeay and R.M. Friedman, unpublished data).

Table 2. Rb-Sr data on muscovite from sample 15 (MV-86-344)

Analysis	ppm Rb	ppm Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$, error
feldspar-1	7.045	6.290	3.241	0.70778 (9)
feldspar-2	6.519	5.730	3.292	0.70778 (15)
muscovite-1	144.6	158.1	2.647	0.70695 (4)
muscovite-2	144.8	158.1	2.648	0.70699 (7)
muscovite-3	144.3	158.1	2.642	0.70700 (5)

feldspar-muscovite age (n = 5): 93.5 ± 11.4 (2 sigma), MSWD = 1.62
2-sigma error on $^{87}Rb/^{86}Sr$ = 0.5%, 2-sigma error on $^{87}Sr/^{86}Sr$ refers to the last digit(s).

(16) GNEISSIC ROCKS, MT. LYTTON COMPLEX (MV-82-176) (TJm5-1)

Sample MV-82-176 is a layered quartzofeldspathic gneiss which is part of a rock unit composed of layered and quite deformed dykes and intrusions within the Mount Lytton metasediments. The gneiss is medium grained and composed of an aggregate of strained quartz, feldspar, and fine grained epidote.

Analysis

Zircons from this sample were of relatively poor quality, being fractured and cloudy. Their uranium contents are very high (>1600 ppm), and it is likely that Pb-loss has occurred in all zircons. Analysis of four fractions of cloudy fragments of zircon resulted in a quasi-linear cluster of points below concordia (Fig. 16). Three of the four zircon fractions are unabraded and the fourth fraction is only moderately abraded. Perhaps the best estimation for the age of this layered quartzofeldspathic gneiss is the mean of the ^{207}Pb-^{206}Pb ages for all four fractions, approximately 225 Ma. It is difficult to properly assign an uncertainty to this age but we feel the age is highly likely to lie in the range 225 ± 5 Ma.

Interpretation

This complex of metamorphic and meta-igneous rocks contains rocks of ca 225 Ma (this sample) but also some older rocks of possible Permian age (van der Heyden, unpublished data). These deformed rocks are intruded by massive intrusions 212 Ma old, represented by sample 1 of this paper (Fig. 1). It seems that there was magmatic activity of Permo-Triassic age in part coeval with magmatism of the Nicola volcanic belt. The older gneissic rocks may in part represent a basement to part of the Nicola arc in the western Intermontane region. If extrapolated northward, rocks of the Mt. Lytton Complex would lie west of Cache Creek Group. Since the Nicola Belt is regarded as a west-facing arc, in part subducting material of the Cache Creek Group, an explanation for this spatial pattern may be that Cache Creek rocks were thrust eastwards over the Mt. Lytton Complex in the Jurassic, prior to arching of the Mt. Lytton-Eagle Plutonic Complex belt in the Jurassic and early Cretaceous.

(17) MC-88-119, MYLONITIC GRANITE WITHIN CUSTER GNEISS

Sample MC-88-119 is a coarse-grained, foliated and mylonitic, equigranular hornblende-biotite granite. This granite intrudes metamorphic rocks of the Custer gneiss in the footwall of a normal fault interpreted by Coleman and Parrish (in press) to be the northernmost extension of the Ross Lake Fault. This normal fault was interpreted by them to be in part an Eocene extension fault. The granite was dated in

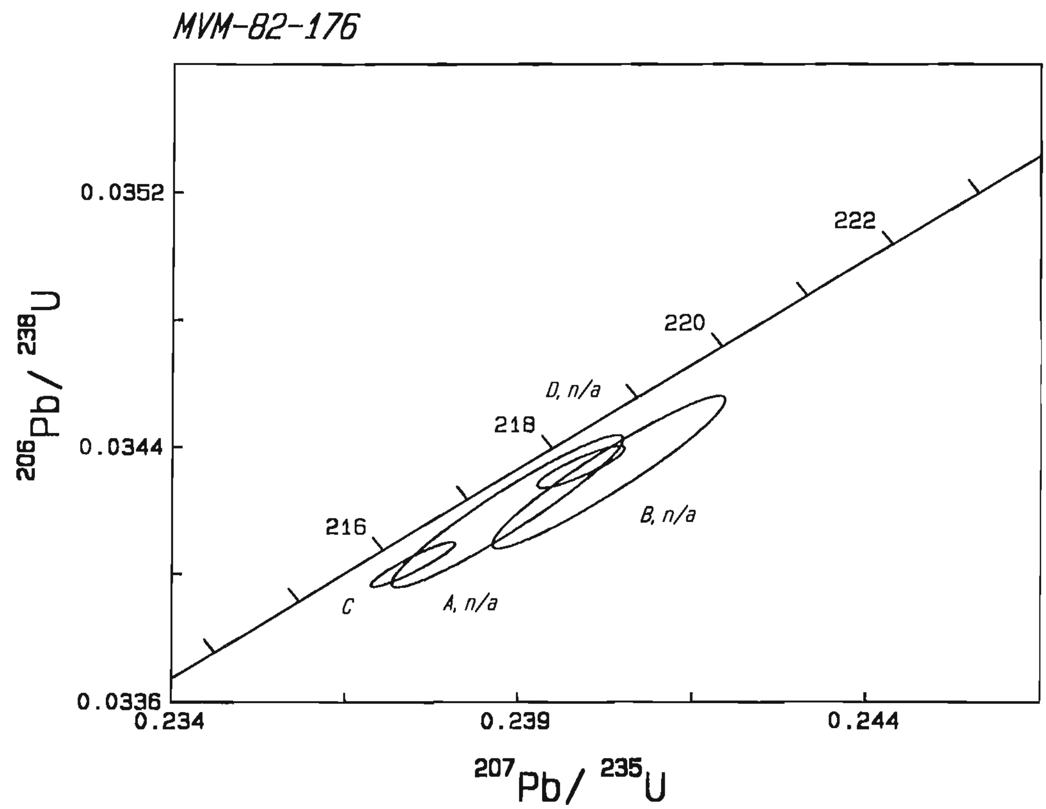

Figure 16. U-Pb concordia plot of MV-82-176, layered quartzofeldspathic gneiss of the Mt. Lytton Complex. Error ellipses reflect the 2σ uncertainty.

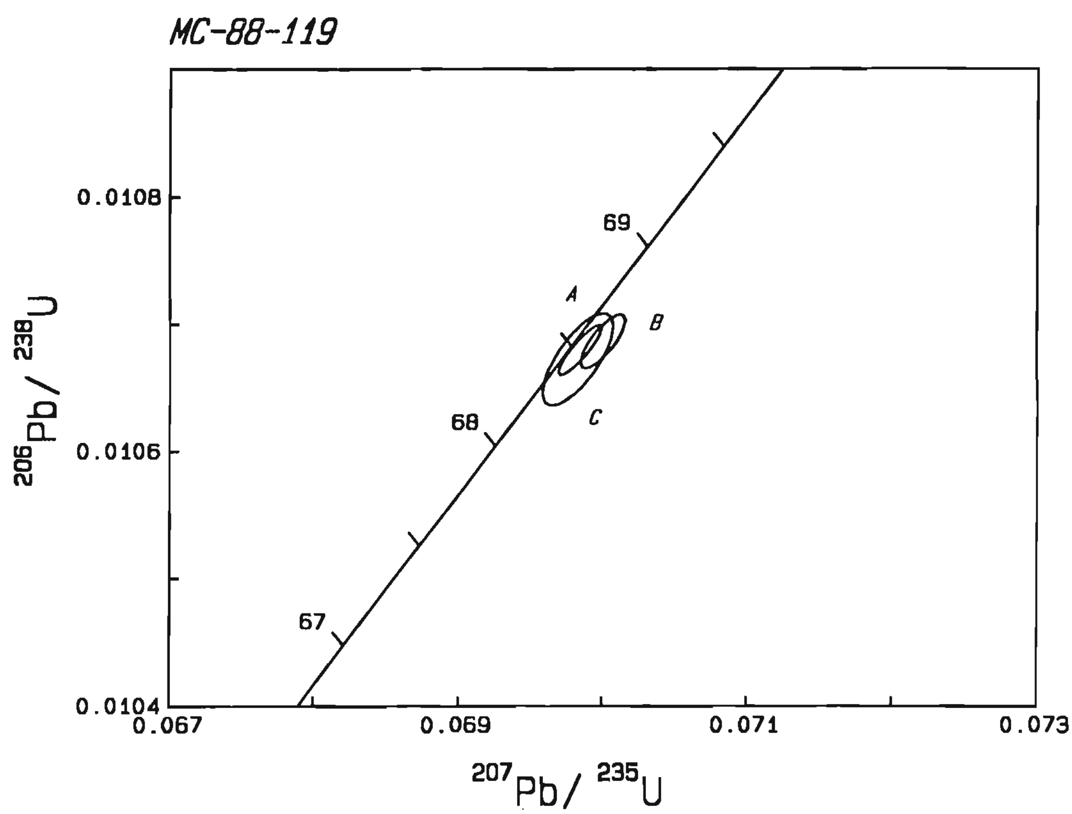

Figure 17. U-Pb concordia plot for MC-88-119, foliated hornblende-biotite granite. Error ellipses reflect the 2σ uncertainty.

order to determine if this intrusion correlated with the ca. 46 Ma Mission Ridge Pluton northwest of Lillooet on the opposite side of the north-striking Fraser River-Straight Creek fault (Coleman and Parrish, in press).

Analysis

Three fractions of strongly abraded, euhedral clear zircons with a few inclusions were analyzed. Three overlapping analyses indicate an age of crystallization of 68.5 ± 0.4 Ma (Fig. 17).

Interpretation

The 68 Ma age is older than the Mission Ridge Pluton (46-47 Ma) to the northwest near Lillooet; thus they cannot be correlated, although the host rocks of the Custer gneiss are similar to the Bridge River schist near Lillooet. The deformed nature of the granite, which has subhorizontal stretching lineations and evidence for dextral shear, indicates considerable dextral and extensional displacement in the vicinity of the Custer gneiss and Fraser River Fault in the period after 68 Ma.

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APPENDIX 1

Sample localities and descriptions

Localities of the samples for U-Pb dating are as follows:

(1) MV-86-120: massive (hornblende) biotite granodiorite of the Mt. Lytton Complex; on logging road, at an elevation of 4475 feet, in the uppermost part of a tributary of the Nicoamen River, 4.0 km on a bearing of 280° from the summit of Zakwaski Mountain, and approximately 18 km on a bearing of 118° from the town of Lytton, B.C. (92I/3); 121°21′21″ W - 50°09′39″ N; UTM 10u, 617440E-5557575N.

(2) MVK-82-20: medium to coarse grained granodiorite of the Wild Horse Batholith; located on the northern shore of Scuttio Lake, approximately 1 km on a bearing of 190° from the summit of Mt. Vicars, B.C. (92I/9); 120°08′35″ W - 50°33′15″ N; UTM 10u, 702376E-5603927N.

(3) MV-84-41: granodiorite of the Bromley Pluton; collected on the north side of the old highway on the north side of the Similkameen River, 1.5 km WNW of the bridge carrying Highway 3 over the river, and approximately 7.5 km WNW of Hedley, B.C. (92H/8); 120°10′15″ W - 49°22′55″ N; UTM 10u, 705339E-5473547N.

(4) MVM-86-9: hornblende biotite granodiorite of the Pennask Batholith; 1 km from the southern edge of Hatheume Lake, B.C. at the end of the road in lot L4243 (92H/16); 120°02′29″ W - 49°59′14″ N; UTM 10u, 712075E-5541185N.

(5) MV-84-5: coarse grained biotite quartz monzonite of the Osprey Lake Batholith; collected from a roadcut approximately 20 km NW of Princeton, B.C. on the west side of the Hayes Creek Road, 1 km SSW of the confluence of Hayes and Grant Creeks (92H/9); 120°21′31″ W - 49°37′36″ N; UTM 10u, 690510E-5550020N.

(6) MV-89-157: massive (biotite) hornblende granodiorite from the Cloudburst Quartz Diorite; located at an elevation of 2550 ft., 4.15 km from the junction of Ring Creek and Mamquam River, 9 km northeast of Squamish in the southwesternmost part of Garibaldi Provincial Park, B.C. (92G/11); 123°03′00″ W - 49°44′40″ N; UTM 10u, 496398E-5509900N.

(7) MV-85-482: low grade, altered quartz-feldspar crystal tuff; west bank of Big Silver Creek, 150 m from shore, and 1.5 km north of outflow of creek into the east side of Harrison Lake, Mount Urquhart map-area, southwestern B.C. (92H/12); 121°50′25″ W - 49°35′20″ N; UTM 10u, 583800E-5493120N.

(8) MVH-85-92: gneissic quartz diorite; located on northwest side of logging road along Cogburn Creek, 0.5 km above bridge on Cogburn Creek east of Harrison Lake, Coast Mountains, B.C. (92H/12); 121°45′03″ W - 49°32′50″ N; UTM 10u, 590364E-5488821N.

(9) MV-86-343: mafic gneiss; on west side of logging road on west side of Big Silver Creek, immediately west of first bridge over creek, approximately 6 km on a bearing of 107° from the summit of Mount Breakenridge, B.C. (92H/12); 121°51′16″ W - 49°42′12″ N; UTM 10u, 582615E-5506050N.

(10) MVD-89-375: medium grained biotite granodiorite - quartz diorite of the Snowcap Lake Pluton; about 0.3 km from the south shore of Snowcap Lake, approximately 1.3 km west of the western end of Cutoff Lake, Garibaldi Provincial Park, Coast Mountains, B.C. (92G/15); 122°36′35″ W - 49°51′40″ N; UTM 10u, 528051E-5523100N.

(11) MV-86-431: white biotite quartz monzonite of the Scuzzy Pluton; in southeastern part of Scuzzy Pluton, on logging road south of Scuzzy Creek at an elevation of 3700 feet, approximately 8.5 km ESE from the summit of Mount Nesbitt, Scuzzy Mountain map area, southwestern B.C. (92H/13); 121°34′40″ W - 49°47′25″ N; UTM 10u, 602474E-5515675N.

(12) MV-86-99: quartz - K-feldspar porphyry; located on logging road at an elevation of approximately 5700 feet, on top of rounded ridge 1.7 km on a bearing of 233° from the fire lookout on Mount Kathleen, Okanagan Provincial Forest, B.C. (92H/16); 120°04′33″ W - 49°45′27″ N; UTM 10u, 710620E-5515550N.

(13) MV-86-98: medium to fine-grained biotite granodiorite of the Pennask Batholith; located 1.27 km southwest of the fire lookout at about 5500 feet on Mount Kathleen, Okanagan Provincial Forest, B.C. (92H/16); 120°04′0″ W - 49°45′26″ N; UTM 10u, 711275E-5515545N.

(14) MVK-82-133a: medium to coarse grained granodiorite of the Nicola Batholith; elevation of 4450 ft. approximately 1.2 km southeast of Conant Lake, B.C. (92I/7); 120°32′35″ W - 50°16′08″ N; UTM 10u, 675094E-5571197N.

(15) MV-86-344: sheared muscovite-quartz-feldspar schist; on creek north of upper bridge over Big Silver Creek, just below where it joins with west-flowing creek, 6 km east of Mount Breakenridge, B.C. (92H/12); 121°51′01″ W - 49°42′58″ N; UTM 10u, 582875E-5507470N.

(16) MVM-82-176: medium grained, layered quartzofeldspathic gneiss of the Mt. Lytton Complex; approximately 4.5 km east of Lytton, B.C. on the north side of the Thompson River, on the C.N. railroad track located midway between the first and second train tunnels heading east (92I/5); 121°32′12″ W - 50°15′56″ N; UTM 10u, 604295E-5568962N.

(17) MC-88-119: foliated hornblende-biotite granite; located northeast of Hope at an elevation of approximately 2600 feet on the ridge just east of the Fraser River, B.C. (92H/6); 121°21′40″ N - 49°28′50″ W; UTM 10u, 618666E-5481750N.

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U-Pb ages for Cretaceous plutons in the eastern Coast Belt, southern British Columbia

Randall R. Parrish[1]

Parrish, R.R., 1992: U-Pb ages for Cretaceous plutons in the eastern Coast Belt, southern British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 109-113.

Abstract

Three ages of plutons in the eastern Coast Mountains of southern British Columbia are presented. Two of these have specific structural relationships to Late Cretaceous structures which are east-verging, and they bracket these structures to the period of time between 87 and 68 Ma. The third sample intrudes Bridge River Group near Mt. Dickson and is 92 Ma old, placing constraints on the age of the Bralorne fault zone.

Résumé

Trois datations de plutons reposant dans l'est de la chaîne Côtière du sud de la Colombie-Britannique sont présentées. Deux d'entre eux ont des liens structuraux spécifiques avec des structures du Crétacé supérieur à vergence est et permettent de corréler ces structures à la période allant de 87 à 68 Ma. Le troisième échantillon provient d'une intrusion de pluton dans le groupe de Bridge River près du mont Dickson dont l'âge, 92 Ma, limite celui de la zone faillée de Bralorne.

INTRODUCTION

Three samples were dated as part of investigations of the geology on the eastern side of the Coast Mountains of southern British Columbia in the vicinity of Mount Waddington by Rusmore and Woodsworth (1989, 1991). Between Homathko River and Mosley Creek (Fig. 1) a series of imbricate faults, which verge east and carry a variety of sedimentary, volcanic, metasedimentary, and meta-igneous rocks, are exposed. These east-verging structures are part of a much larger belt of east-verging tectonic features which lie on the eastern side of the Coast Belt nearly the full length of British Columbia (Rusmore and Woodsworth, 1991). A recent study by Evenchick (1991) outlined these structures in northern British Columbia. Near Mount Waddington, these structures involve Triassic to Cretaceous rocks. In order to bracket the age of these structures, a sample of orthogneiss which was transported and deformed, and a sample of cross-cutting granite were collected by Woodsworth and Rusmore for U-Pb age determinations.

A third sample (86-WV-2), which bears on the age of the Bralorne fault zone and related structures, was also collected by Rusmore from near Mount Dickson farther southeast near Bralorne. Isotopic data are summarized in Table 1. U-Pb analytical methods follow those outlined by Parrish et al. (1987).

U-Pb AGE DETERMINATIONS

Orthogneiss west of Nude Creek (88-WV-100)

A well foliated tonalitic orthogneiss, locally containing epidote, was collected on the west side of Nude Creek within a map unit of orthogneiss that is juxtaposed structurally with overlying sheets of metasedimentary rocks. The locality of the sample is 7.5 km east-southeast of Rusty Peak, Nuit Range at an elevation of 7700 feet (UTM zone 10u, 374000E, 5708500N). This rock is clearly involved in the east-directed deformation.

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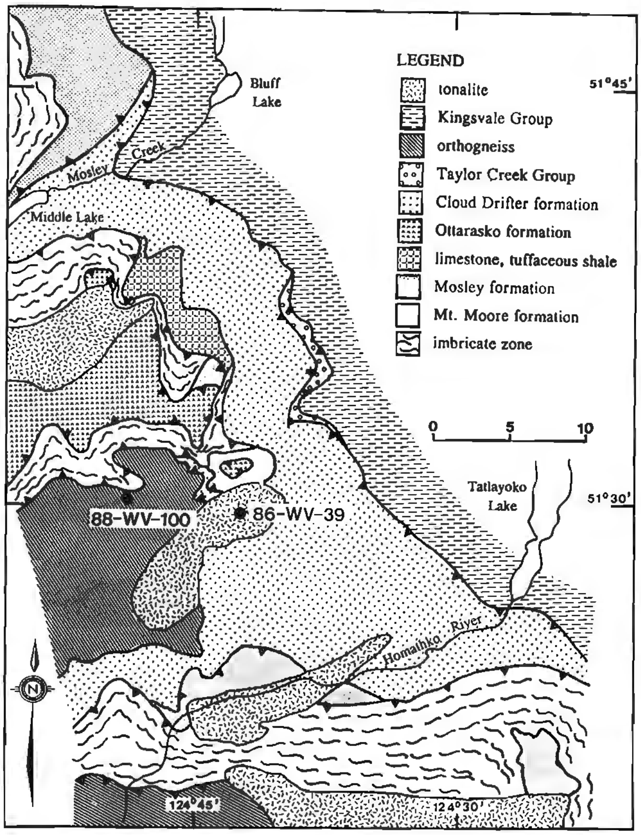

Figure 1. Geological reference map for the area east of Mount Waddington. Sample localities for 86-WV-39 and 88-WV-100 are shown (after Rusmore and Woodsworth, 1991; their fig. 2).

Zircons were of excellent quality and were strongly abraded by the method of Krogh (1982). Three overlapping analyses are concordant to slightly below concordia at 87.3 ± 0.3 Ma (Fig. 2). The reason that these high precision analyses fall slightly below concordia is because of an initial deficiency in ^{230}Th during crystallization of zircon; further details on this phenomena are outlined in Coleman and Parrish (in press). A correction for this phenomena would cause the ^{206}Pb-^{238}U ages to increase by 0.05-0.1 Ma, in which case they would more centrally overlap concordia. The interpreted age of 87.3 ± 0.3 Ma takes this into account. This age is interpreted as the crystallization age of the tonalite, and its deformation therefore postdates 87 Ma.

K-Ar dates from hornblende and biotite from this sample are 77.4 ± 1.0 and 57.2 ± 0.9 Ma, respectively (GSC K-Ar 91-20,21; Hunt and Roddick, 1992), indicating slow cooling following the emplacement, deformation and metamorphism of this originally 87 Ma pluton.

Unfoliated quartz diorite east of Nude Creek (86-WV-39)

An unfoliated quartz diorite forms an elongate (15 x 4 km) northeast-striking pluton south of Ottarasko Mountain in the Nuit Range. The sample comes from the central part of this body on the east side of Nude Creek. The pluton is unfoliated except for a local magmatic layering within a few 100 m of the pluton margins. The layering is steep and strikes northeast.

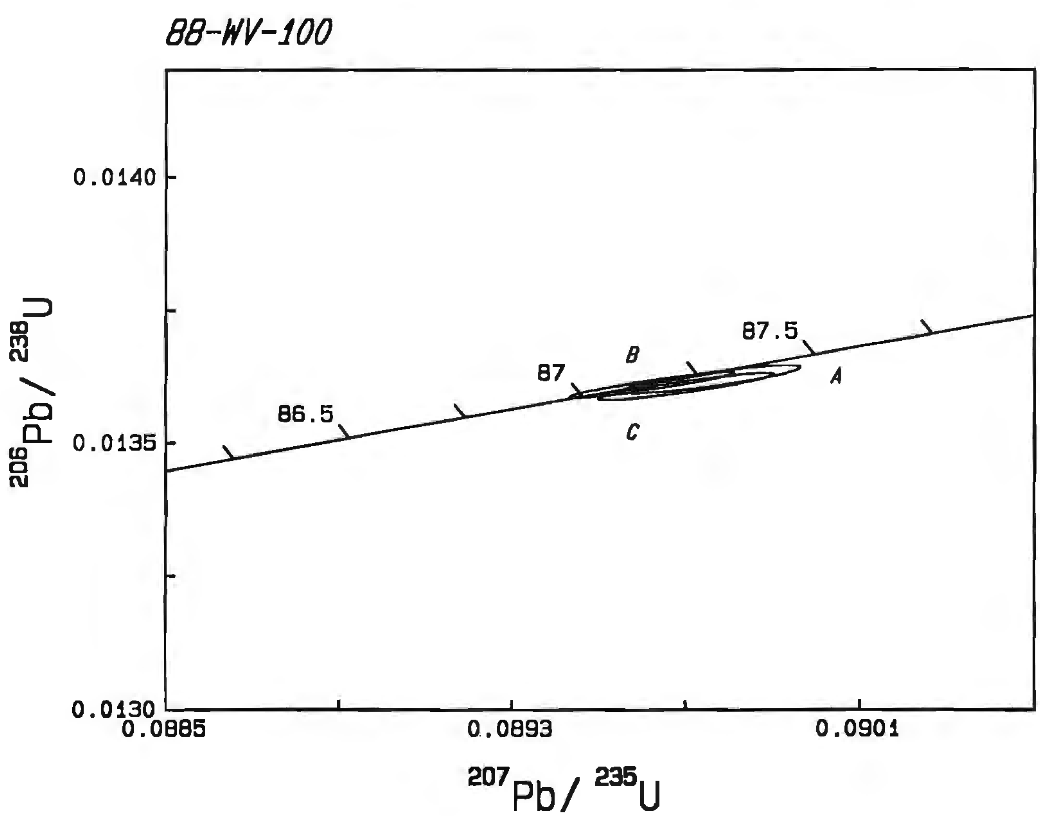

Figure 2. Concordia plot for zircons from sample 88-WV-100. Error ellipses reflect 2σ uncertainty.

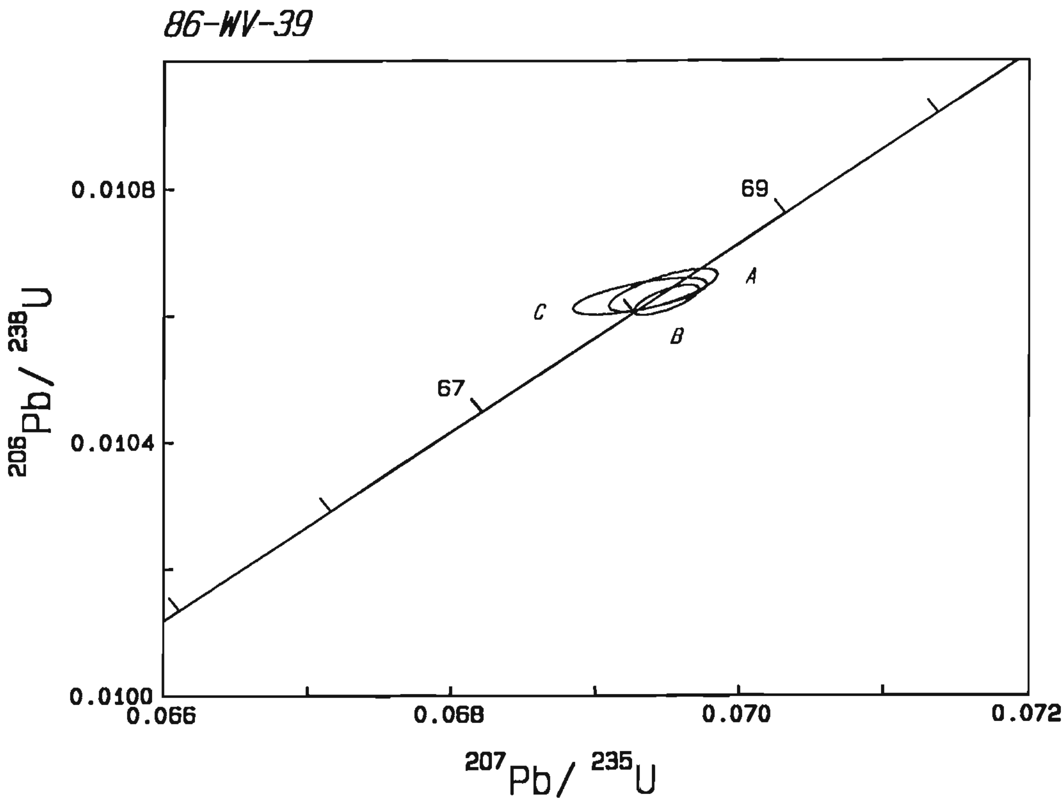

Figure 3. Concordia plot for zircons from sample 86-WV-39. Error ellipses reflect 2σ uncertainty.

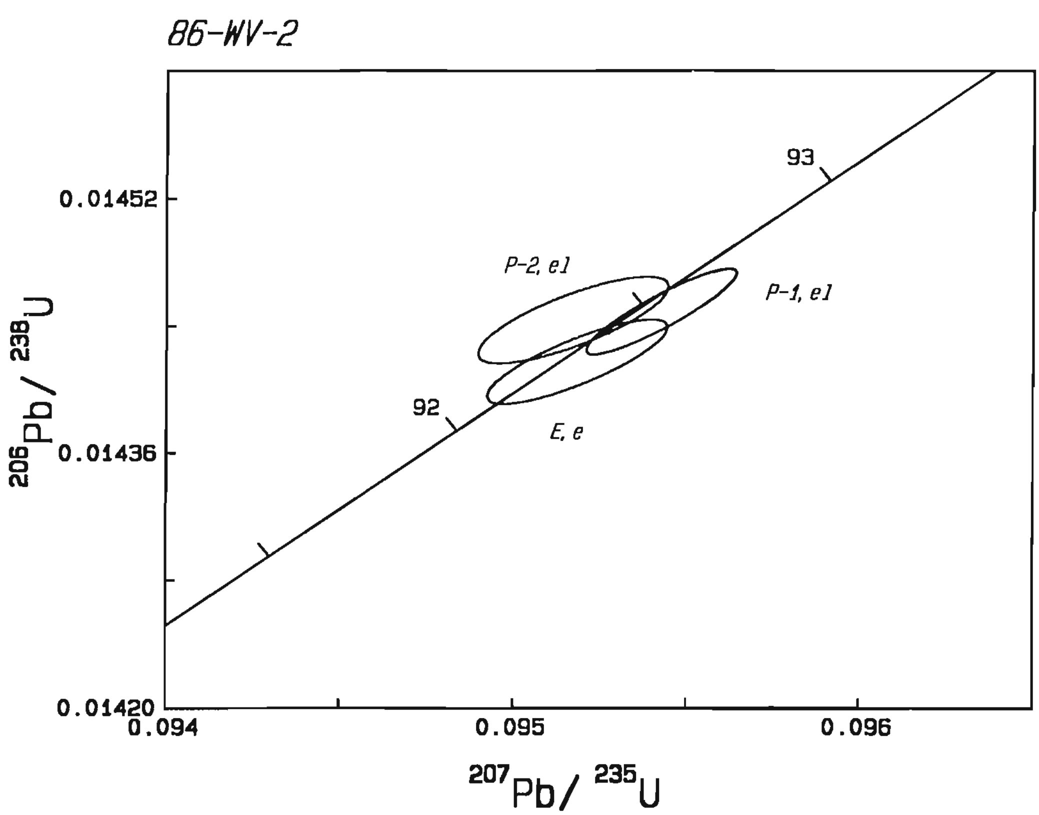

Figure 4. Concordia plot for zircons from sample 86-WV-2. Error ellipses reflect 2σ uncertainty.

Table 1. U-Pb Analytical Data for Cretaceous plutons in the eastern Coast Belt, southern British Columbia[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^*/^{204}Pb$	Pb_c,# (pg)	$^{208}Pb^+/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb^{***}/^{206}Pb$ (Ma)
88-WV-100, tonalitic orthogneiss													
A,+149,el	0.469	145.3	2.039	2527	23	0.150	0.01362 ± 0.10	87.2 ± 0.2	0.08976 ± 0.12	87.3 ± 0.2	0.04780 ± 0.06	0.86	89.2 ± 2.8
B,+149	0.572	166.8	2.321	4128	20	0.141	0.01361 ± 0.09	87.1 ± 0.2	0.08962 ± 0.11	87.2 ± 0.2	0.04776 ± 0.04	0.94	87.4 ± 1.8
C,+149	0.852	156.0	2.156	3249	35	0.134	0.01361 ± 0.09	87.1 ± 0.2	0.08970 ± 0.11	87.2 ± 0.2	0.04781 ± 0.05	0.92	90.1 ± 2.2
86-WV-39, unfoliated quartz diorite													
A,sc,+149	0.085	233.9	2.463	487	28	0.104	0.01064 ± 0.16	68.2 ± 0.2	0.06947 ± 0.27	68.2 ± 0.4	0.04735 ± 0.19	0.75	67.1 ± 8.9
B,cl,125	0.107	269.1	2.871	812	24	0.120	0.01062 ± 0.11	68.1 ± 0.2	0.06949 ± 0.16	68.2 ± 0.2	0.04744 ± 0.11	0.72	71.2 ± 5.4
C,sc,125	0.100	171.7	1.799	340	36	0.099	0.01063 ± 0.13	68.2 ± 0.2	0.06930 ± 0.34	68.0 ± 0.4	0.04729 ± 0.27	0.67	63.7 ± 12.7
86-WV-2, granodiorite													
P-1,el,+149	0.078	499.6	7.371	2180	16	0.139	0.01445 ± 0.09	92.5 ± 0.2	0.09543 ± 0.11	92.5 ± 0.2	0.04790 ± 0.05	0.90	94.4 ± 2.3
P-2,el,+149	0.226	210.7	3.105	1329	33	0.138	0.01444 ± 0.09	92.4 ± 0.2	0.09517 ± 0.15	92.3 ± 0.3	0.04779 ± 0.09	0.78	92.3 ± 0.3
E,e,cl,+149	0.192	138.4	2.007	1123	22	0.122	0.01442 ± 0.09	92.3 ± 0.2	0.09519 ± 0.14	92.3 ± 0.2	0.04788 ± 0.08	0.82	93.5 ± 3.8

** Numbers (i.e. +149,125) refer to average size of zircons in microns; n/a = not abraded (all other fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;
Weighing error = 0.001 mg.
+ Radiogenic Pb.
* Measured ratio, corrected for spike, and Pb fractionation of 0.09%·+/- 0.03%/AMU
# Total common Pb in analysis corrected for fractionation and spike.
++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion, mineral dissolution in microcapsules, a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer, multicollector mass spectrometry, and estimation of errors by numerical error propagation.

The pluton intrudes and cross-cuts the northeast-verging thrust belt (Fig. 1), and so provides an upper limit on the age of this thrusting (Rusmore and Woodsworth, 1988, 1989, 1991; Rusmore et al., 1989, 1990).

Zircons from this sample were similarly of excellent quality and were also abraded. Three analyses are all concordant and overlapping with an estimated age and uncertainty of 68.2 ± 0.2 Ma (Fig. 3). This age is interpreted as the age of crystallization, and indicates that the thrusting predates 68 Ma; these structures are thus bracketed between 68 and 87 Ma. These conclusions are emphasized in a recent paper which cites these ages (Rusmore and Woodsworth, 1991).

Granodiorite on Mount Dickson (86-WV-2)

This sample is from the south side of Mount Dickson and is part of a large pluton of uncertain extent that underlies the peak. Its location is 0.75 km southwest of Dickson Peak, just south of a small lake at an elevation of 7700 feet, (UTM zone 10u, 500275E, 5636850 N). This locality is about 140 km southeast of the samples described above, and is not shown in Figure 1. The sample was collected by M. Rusmore, who noted the following field observations.

The stock is cut by several steep, north- to northeast-trending faults which are narrow and marked by slickensides and minor chloritization. Other than these features, the granodiorite appears structureless and is unfoliated. It intrudes the Bridge River Complex and a large serpentinite body that underlies the eastern part of Mount Penrose (see Roddick and Woodsworth, 1977). The serpentinite is lightly metamorphosed near the contact, and the stock and related dykes cross-cut foliation in the serpentinite.

Zircons from the granodiorite were of excellent quality and three abraded zircon analyses are concordant and overlapping with an interpreted crystallization age of 92.4 ± 0.3 Ma (Fig. 4). Based on this date and the above field relations, the Mount Dickson stock is younger than the faulting that bounds the serpentinite. Rusmore interpreted the exhumation of the serpentinite as due to movement on the Bralorne fault zone which is therefore inferred to be older than 92 Ma. The Albian-Aptian Taylor Creek Group is included in the fault zone farther south, so the main phase of motion along the Bralorne Fault zone probably occurred in late Early Cretaceous time, synchronous with major west-directed thrusting in southwestern British Columbia and northern Washington within the western Cascade Belt. Shallow level faults affected the Bralorne area since mid-Cretaceous time.

CONCLUSIONS

The age determinations presented in this paper, in conjuction with the summarized field relations of these three samples derived from observations of Rusmore and Woodsworth (1989), have implications for the age of structures in the eastern Coast Mountains of southern British Columbia.

The Bralorne Fault zone can be inferred to have been active in the late Early Cretaceous, possibly overlapping in age with west-directed major thrusting in the Cascade Belt of Washington and British Columbia.

The major belt of east-vergent thrusts, which extends from northern Washington to northern British Columbia, is clearly bracketed in age between 87 and 68 Ma in the vicinity of Mount Waddington. This age range is consistent with observation elsewhere farther north and south from this area. It is quite likely that the west-verging thrusting recorded to the west in the Cascade Belt (which ranges mainly from ca. 96-92 Ma; J.M. Journeay and R.M. Friedman, unpublished data) is distinctly older than the east-verging belt described here.

ACKNOWLEDGMENTS

I appreciate the information on these samples supplied by M. Rusmore and G.J. Woodsworth. Discussion with M. Rusmore was most useful in writing this brief manuscript. M. Rusmore and G.J. Woodsworth collected the samples. V. McNicoll drafted Figure 1 and reviewed the manuscript.

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U-Pb ages of Jurassic-Eocene plutonic rocks in the vicinity of Valhalla Complex, southeast British Columbia

Randall R. Parrish[1]

Parrish, R.R., 1992: U-Pb ages of Jurassic-Eocene plutonic rocks in the vicinity of Valhalla Complex, southeast British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 115-134.

Abstract

U-Pb ages are presented for 15 samples of plutonic rocks from the hangingwall of the Eocene extensional Slocan Lake fault and Valkyr shear zones, which form the roof fault system to the Valhalla metamorphic core complex of southeast British Columbia. The Nelson batholith, a major geological unit in the hangingwall of the complex, is composite with early phases as old as 170 Ma and younger intrusions perhaps as young as 163 Ma.

Résumé

Les datations U-Pb de 15 échantillons de roches plutoniques provenant de la lèvre supérieure de la faille de distension du lac Slocan de l'Éocène et des zones de cisaillement de Valkyr qui forment le réseau de failles supérieure du complexe de noyau métamorphique de Valhalla dans le sud-est de la Colombie-Britannique sont présentées. Le batholite de Nelson, unité géologique importante de la partie supérieure du complexe, est composite étant formé d'intrusions anciennes datant de 170 Ma et d'intrusions peut-être aussi récentes que 163 Ma.

INTRODUCTION

The Valhalla Complex of southeastern British Columbia is a metamorphic core complex with extensional shear and fault zones separating high grade metamorphic and metaplutonic rocks in the footwall from lower grade plutonic and subordinate metamorphic rocks in the hangingwall. These hangingwall rocks include: an Early Jurassic pluton; the Middle Jurassic Nelson Batholith, Mackie Pluton, and Ruby Range stock; the Late Cretaceous Snowslide Creek stock and Whatshan batholiths; and other younger Eocene intrusions. U-Pb ages are presented for 14 of these plutonic rocks and discussed in order of decreasing age. Zircon inheritance is nearly ubiquitous and especially common in rocks younger than 170 Ma, rendering precise ages difficult to obtain.

During geological investigations of the Valhalla Complex (Parrish, 1984; Parrish et al., 1985a; Carr, 1985; Carr et al., 1987; Parrish et al., 1988), it was noted that a number of plutonic rock units in the hangingwall of the extensional Slocan Lake-Valkyr fault system were very similar and it was suggested that they formed parts of the Nelson composite batholith, subsequently broken and separated during Eocene extensional faulting. Many of the rocks presented in this study were dated to test this prediction. During this reconaissance dating program, other magmatic events were indentified, notably of Early Jurassic and Late Cretaceous age. The U-Pb analytical data presented in this study show Precambrian zircon inheritance to be common, and although considerable work has been done on most of the samples, precise ages were obtained on only a few of them. Despite

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Table 1. U-Pb Analytical Data for Jurassic – Eocene plutonic rocks, Valhalla Complex[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^*/^{204}Pb$	Pb_c,# (pg)	$^{208}Pb^+/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb^{***}/^{206}Pb$ (Ma)
(1) PCA-64-85, Diorite													
A,+149	0.280	198.0	5.916	1472	73	0.098	0.03027 ± 0.10	192.2 ± 0.4	0.2092 ± 0.17	192.9 ± 0.6	0.05014 ± 0.13	0.63	201.2 ± 6.2
B,125	0.424	168.8	5.122	2110	66	0.108	0.03046 ± 0.07	193.4 ± 0.2	0.2102 ± 0.13	193.8 ± 0.4	0.05007 ± 0.09	0.76	198.0 ± 4.2
C,90	0.289	219.4	6.611	4484	27	0.100	0.03046 ± 0.07	193.4 ± 0.2	0.2103 ± 0.10	193.8 ± 0.4	0.05007 ± 0.07	0.78	198.1 ± 3.0
G,sc	0.209	168.5	5.119	1459	47	0.108	0.03050 ± 0.09	193.7 ± 0.4	0.2104 ± 0.12	193.8 ± 0.4	0.05002 ± 0.06	0.88	196.0 ± 2.7
H,sc	0.150	163.0	4.918	443	110	0.105	0.03038 ± 0.11	192.9 ± 0.4	0.2090 ± 0.34	192.7 ± 1.2	0.04991 ± 0.28	0.61	190.8 ± 13.3
J,sc	0.254	175.4	5.293	1240	70	0.104	0.03039 ± 0.09	193.0 ± 0.3	0.2096 ± 0.12	193.2 ± 0.4	0.05002 ± 0.07	0.86	196.0 ± 3.0
(2) PCA-452-83-SY, Nelson Batholith quartz syenite													
A,na,cl	0.990	56.80	1.634	483	1232	0.233	0.02762 ± 0.30	175.6 ± 1.1	0.1931 ± 0.40	179.3 ± 1.2	0.05072 ± 0.10	0.80	227.9 ± 5.0
B,na,cl	3.140	216.1	5.961	27830	264	0.148	0.02678 ± 0.15	170.4 ± 0.4	0.1833 ± 0.20	170.9 ± 0.6	0.04965 ± 0.05	0.90	178.7 ± 2.2
C,na,cl	8.530	221.5	6.153	1331	887	0.180	0.02684 ± 0.30	170.8 ± 0.9	0.1844 ± 0.40	171.8 ± 1.3	0.04981 ± 0.10	0.80	186.2 ± 4.6
D,sc,+177	1.182	376.8	10.48	6007	1256	0.154	0.02686 ± 0.07	170.9 ± 0.2	0.1836 ± 0.10	171.2 ± 0.4	0.04959 ± 0.05	0.89	175.7 ± 2.2
E,+177	1.687	237.0	6.602	1594	427	0.160	0.02677 ± 0.18	170.3 ± 0.6	0.1829 ± 0.22	170.5 ± 0.6	0.04954 ± 0.12	0.84	173.6 ± 5.6
F,sc,163	1.041	207.0	5.782	7092	52	0.158	0.02689 ± 0.07	171.1 ± 0.2	0.1839 ± 0.10	171.4 ± 0.4	0.04959 ± 0.05	0.86	175.8 ± 2.4
TIT-1,na	0.141	51.39	2.249	89	171	0.871	0.02643 ± 0.56	168.1 ± 1.9	0.1774 ± 1.6	165.8 ± 5.0	0.04869 ± 1.4	0.60	132.8 ± 66.0
TIT-2,na	0.166	50.65	2.153	73	257	0.816	0.02642 ± 0.80	168.1 ± 2.6	0.1788 ± 2.5	167.0 ± 7.8	0.04909 ± 2.1	0.66	152.2 ± 101
TIT-3,na	1.345	48.58	2.157	81	1719	0.902	0.02636 ± 0.84	167.7 ± 2.8	0.1800 ± 2.7	168.1 ± 8.4	0.04954 ± 2.2	0.69	173.2 ± 107
(3) PCA-34A-85, Nelson Batholith quartz syenite													
A,+149	0.335	305.0	7.715	3433	49	0.087	0.02589 ± 0.18	164.8 ± 0.6	0.1765 ± 0.19	165.1 ± 0.6	0.04945 ± 0.07	0.93	169.4 ± 3.2
B,125	0.508	341.2	8.493	4019	70	0.081	0.02557 ± 0.17	162.7 ± 0.6	0.1807 ± 0.19	168.7 ± 0.6	0.05128 ± 0.07	0.94	253.3 ± 3.0
C,90	0.465	338.0	8.260	3882	64	0.082	0.02512 ± 0.17	160.0 ± 0.6	0.1708 ± 0.18	160.1 ± 0.6	0.04929 ± 0.07	0.92	161.7 ± 3.2
D,68	0.030	383.1	9.009	684	26	0.079	0.02424 ± 0.23	154.4 ± 0.8	0.1658 ± 0.32	155.8 ± 1.0	0.04962 ± 0.25	0.64	177.3 ± 11.6
E,+149,sc	0.332	314.4	8.276	2390	74	0.101	0.02660 ± 0.17	169.3 ± 0.6	0.1822 ± 0.21	169.9 ± 0.6	0.04967 ± 0.11	0.84	179.5 ± 5.2
F,90	0.083	323.3	8.458	3042	15	0.098	0.02650 ± 0.08	168.6 ± 0.2	0.1828 ± 0.13	170.5 ± 0.4	0.05005 ± 0.09	0.71	197.2 ± 4.2
G,125	0.556	290.7	7.566	6056	44	0.101	0.02629 ± 0.06	167.3 ± 0.2	0.1815 ± 0.10	169.3 ± 0.4	0.05008 ± 0.05	0.89	198.5 ± 2.4
(4) NB-7, Nelson Batholith granodiorite													
A,125,cl	0.157	948.5	24.94	6781	37	0.080	0.02703 ± 0.19	171.9 ± 0.6	0.1900 ± 0.21	176.6 ± 0.6	0.05098 ± 0.05	0.97	239.7 ± 2.4
B,90,cl	0.250	945.8	24.57	9993	39	0.088	0.02653 ± 0.18	168.8 ± 0.6	0.1853 ± 0.20	172.6 ± 0.6	0.05066 ± 0.04	0.97	225.2 ± 2.0
C,na,co	0.254	1317	38.22	5798	104	0.129	0.02845 ± 0.18	180.8 ± 0.6	0.2236 ± 0.20	204.9 ± 0.8	0.05702 ± 0.05	0.97	492.2 ± 2.2
D,68	0.020	1542	35.82	1158	40	0.097	0.02355 ± 3.7	150.0 ± 11.0	0.2236 ± 0.20	152.2 ± 10.4	0.04981 ± 0.15	0.99	186.1 ± 7.0
(5) PCA-251-85, Nelson Batholith foliated granodiorite													
A,68	0.868	460.5	12.47	6875	100	0.097	0.02738 ± 0.18	174.1 ± 0.6	0.1989 ± 0.19	184.2 ± 0.6	0.05269 ± 0.05	0.97	315.3 ± 2.2
B,na,105	0.845	419.2	11.58	2769	226	0.096	0.02796 ± 0.18	177.8 ± 0.6	0.2018 ± 0.20	186.6 ± 0.6	0.05233 ± 0.07	0.94	299.8 ± 3.2
C,-62	0.123	350.6	9.711	1728	43	0.124	0.02739 ± 0.08	174.2 ± 0.2	0.1950 ± 0.14	180.9 ± 0.4	0.05164 ± 0.09	0.79	269.7 ± 4.2
D,90	0.251	370.3	10.65	4748	35	0.114	0.02865 ± 0.06	182.1 ± 0.2	0.2102 ± 0.10	193.7 ± 0.4	0.05323 ± 0.05	0.88	338.5 ± 2.4
TIT-1,na	0.569	128.3	3.239	152	782	0.243	0.02269 ± 0.40	144.6 ± 1.1	0.1541 ± 1.3	145.5 ± 3.4	0.04926 ± 1.0	0.70	160.4 ± 48.2
(6,7) Ruby Range Stock East (E, PCA-RRE-83) and West (W, PCA-RRE-83)													
A,na,(E)	0.950	370.5	11.23	511	1315	0.103	0.03078 ± 0.30	195.4 ± 1.2	0.2356 ± 0.50	214.8 ± 2.0	0.05552 ± 0.20	0.80	433.3 ± 8.8
B,na,(E)	1.250	771.8	20.34	630	1727	0.130	0.02704 ± 0.30	172.0 ± 1.0	0.1946 ± 0.5	180.6 ± 1.5	0.05219 ± 0.20	0.80	293.8 ± 9.0
C,na,(W)	4.470	525.4	13.12	4966	776	0.065	0.02704 ± 0.30	172.0 ± 1.0	0.1950 ± 0.35	180.9 ± 1.1	0.05230 ± 0.10	0.90	298.5 ± 4.6
D,na,(W)	2.040	510.1	13.12	5089	331	0.069	0.02680 ± 0.30	170.5 ± 1.0	0.1926 ± 0.35	178.9 ± 1.0	0.05212 ± 0.10	0.90	290.6 ± 4.7
TIT-1,(W)	1.168	126.8	4.411	208	1286	0.496	0.02611 ± 0.31	166.2 ± 1.0	0.1788 ± 0.91	167.0 ± 2.8	0.04965 ± 0.73	0.70	178.7 ± 34.3
TIT-2,(W)	0.477	55.39	2.391	75	780	0.849	0.02635 ± 0.87	167.7 ± 2.9	0.1805 ± 2.8	168.5 ± 8.7	0.04967 ± 2.3	0.69	179.4 ± 111
TIT-3,(W)	0.669	73.82	2.885	126	749	0.691	0.02604 ± 0.46	165.7 ± 1.5	0.1766 ± 1.5	165.1 ± 4.5	0.04918 ± 1.2	0.69	156.2 ± 57.4
TIT-4,(W)	0.651	64.21	2.803	79	1149	0.864	0.02643 ± 0.83	168.2 ± 2.8	0.1826 ± 2.7	170.3 ± 8.3	0.05012 ± 2.2	0.69	200.3 ± 104
TIT-5,(W)	0.658	59.65	2.471	119	648	0.974	0.02629 ± 0.53	167.3 ± 1.8	0.1789 ± 1.7	167.1 ± 5.2	0.04935 ± 1.4	0.70	164.6 ± 66.1
TIT-6 (W)	0.498	60.25	2.270	135	423	0.620	0.02619 ± 0.45	166.7 ± 1.5	0.1771 ± 1.5	165.6 ± 4.5	0.04904 ± 1.2	0.70	149.9 ± 57.0
(8) PCA-27-85, Mackie Pluton													
A,+105,cl	0.084	440.0	12.27	1750	38	0.093	0.02829 ± 0.12	179.8 ± 0.4	0.2073 ± 0.16	191.3 ± 0.6	0.05314 ± 0.09	0.83	334.9 ± 4.2
B,+105,sc	0.145	514.6	12.95	2971	41	0.083	0.02583 ± 0.14	164.4 ± 0.4	0.1770 ± 0.16	165.5 ± 0.5	0.04971 ± 0.06	0.92	181.4 ± 3.0
C,na,90,cl	0.068	629.8	15.46	2751	25	0.076	0.02537 ± 0.11	161.5 ± 0.3	0.1726 ± 0.14	161.7 ± 0.4	0.04936 ± 0.09	0.77	165.2 ± 4.0
D,na,90,sc	0.045	670.7	16.38	1506	32	0.072	0.02534 ± 0.14	161.3 ± 0.4	0.1731 ± 0.18	162.1 ± 0.6	0.04954 ± 0.13	0.72	173.6 ± 6.0

Table 1 (cont.)

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^*/^{204}Pb$	Pb_c,# (pg)	$^{208}Pb^+/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb^{***}/^{206}Pb$ (Ma)
(9) PCA-35-84, Diorite													
A,125	0.500	625.1	15.24	6788	74	0.055	0.02558 ± 0.19	162.8 ± 0.6	0.1883 ± 0.20	175.2 ± 0.6	0.05340 ± 0.05	0.97	345.8 ± 2.2
B,90	0.179	677.6	15.83	5064	37	0.048	0.02471 ± 0.16	157.4 ± 0.6	0.1768 ± 0.18	165.3 ± 0.6	0.05187 ± 0.06	0.93	279.6 ± 3.0
C,na,+149	0.224	616.9	15.11	6106	37	0.058	0.02572 ± 0.16	163.7 ± 0.4	0.1783 ± 0.17	166.6 ± 0.6	0.05029 ± 0.05	0.96	208.3 ± 2.4
D,na,125	0.662	625.0	14.91	15001	43	0.053	0.02504 ± 0.15	159.4 ± 0.4	0.1921 ± 0.17	178.4 ± 0.6	0.05564 ± 0.04	0.97	438.1 ± 1.8
E,na,90	0.249	699.1	15.76	4593	57	0.051	0.02379 ± 0.15	151.6 ± 0.4	0.1682 ± 0.17	157.9 ± 0.6	0.05129 ± 0.05	0.96	253.9 ± 2.4
F,na,68	0.298	840.0	17.17	4326	79	0.047	0.02163 ± 0.16	138.0 ± 0.4	0.1536 ± 0.18	145.1 ± 0.4	0.05149 ± 0.07	0.92	262.8 ± 3.2
G,125	0.003	791.0	17.80	201	19	0.057	0.02368 ± 0.27	150.9 ± 0.8	0.1569 ± 1.2	148.0 ± 3.4	0.04805 ± 1.1	0.51	101.5 ± 53.2
(10) PCA-SCS-83, Snowslide Creek Stock													
A,na,cl	0.910	767.5	9.606	1149	478	0.092	0.01285 ± 0.30	82.3 ± 0.5	0.08962 ± 0.40	87.2 ± 0.6	0.05059 ± 0.15	0.80	222.0 ± 7.3
B,na,sc	0.880	684.9	8.685	605	787	0.088	0.01310 ± 0.30	83.9 ± 0.5	0.09172 ± 0.50	89.1 ± 0.8	0.05078 ± 0.10	0.80	230.6 ± 5.2
C,na,90	1.650	4644	55.19	10327	542	0.097	0.01209 ± 0.30	77.5 ± 0.4	0.08005 ± 0.35	78.2 ± 0.5	0.04802 ± 0.05	0.90	100.0 ± 2.8
ALL-1	0.900	97.78	29.88	32	5621	24.842	0.01361 ± 4.1	87.1 ± 7.1	0.09340 ± 14.8	90.7 ± 25.7	0.04978 ± 12.6	0.65	184.5 ± 499
ALL-2	0.970	104.0	31.31	32	6406	25.115	0.01327 ± 4.1	85.0 ± 6.9	0.08644 ± 13.2	84.2 ± 21.3	0.04726 ± 10.8	0.69	62.5 ± 446
ALL-3	1.009	109.0	31.44	30	7732	24.14	0.01320 ± 4.6	84.5 ± 7.7	0.1014 ± 12.5	98.1 ± 23.5	0.05573 ± 10.0	0.69	441.6 ± 442
ALL-4	0.141	94.36	28.48	30	980	25.54	0.01309 ± 4.8	83.9 ± 8.0	0.08207 ± 15.8	80.1 ± 24.3	0.04546 ± 12.9	0.69	-30.9 ± 500
ALL-5	0.388	99.76	27.01	32	2487	22.40	0.01332 ± 4.2	85.3 ± 7.1	0.08928 ± 12.8	86.8 ± 21.3	0.04863 ± 10.4	0.69	129.8 ± 426
(11) PCA-264-85, Whatshan Batholith													
A,+149,cl	0.300	905.6	10.33	5407	38	0.055	0.01203 ± 0.14	77.1 ± 0.2	0.07912 ± 0.16	77.3 ± 0.2	0.04771 ± 0.05	0.95	84.6 ± 2.5
C,125,cl	0.189	932.3	10.47	3825	35	0.051	0.01189 ± 0.11	76.2 ± 0.2	0.07815 ± 0.14	76.4 ± 0.2	0.04766 ± 0.07	0.89	82.4 ± 3.1
D,125,sc	0.126	706.0	7.967	1724	39	0.055	0.01190 ± 0.13	76.3 ± 0.2	0.07815 ± 0.17	76.4 ± 0.3	0.04763 ± 0.10	0.83	80.9 ± 4.7
E,na,90	0.358	996.8	11.14	6491	41	0.054	0.01180 ± 0.16	75.6 ± 0.2	0.07761 ± 0.17	75.9 ± 0.3	0.04769 ± 0.05	0.96	83.9 ± 2.3
TIT-1,na	0.975	137.0	1.918	116	977	0.386	0.01133 ± 0.27	72.7 ± 0.4	0.07466 ± 0.89	73.1 ± 1.3	0.04778 ± 0.73	0.68	88.1 ± 35.1
TIT-2,na	0.867	184.6	2.518	111	1222	0.365	0.01122 ± 0.40	71.9 ± 1.2	0.07234 ± 1.3	70.9 ± 1.8	0.04678 ± 1.1	0.77	38.1 ± 38.1
(12) PCA-WHAT-83, Whatshan Batholith													
A,na,125,cl	3.740	917.2	10.86	3608	703	0.068	0.01237 ± 0.30	79.3 ± 0.4	0.08161 ± 0.35	79.7 ± 0.5	0.04784 ± 0.15	0.90	91.2 ± 7.4
B,na,90,cl	3.730	964.7	11.48	3243	828	0.075	0.01237 ± 0.30	79.2 ± 0.4	0.08172 ± 0.35	79.8 ± 0.4	0.04791 ± 0.15	0.90	94.7 ± 7.4
TIT-1,na	3.146	167.2	2.554	102	4808	0.418	0.01208 ± 0.66	77.4 ± 1.0	0.08051 ± 2.2	78.6 ± 3.4	0.04832 ± 1.8	0.65	115.2 ± 91.2
(13) PCA-56-84, Late granite													
A,na	2.418	705.5	44.96	7157	845	0.159	0.05889 ± 0.15	368.9 ± 1.0	0.8245 ± 0.20	610.6 ± 1.6	0.1015 ± 0.05	0.90	1652.4 ± 0.5
B,na,90	0.949	933.3	31.92	3292	527	0.137	0.03251 ± 0.20	206.3 ± 0.6	0.4070 ± 0.25	346.7 ± 1.5	0.09079 ± 0.10	0.85	1442.4 ± 3.4
C,na,+149	0.421	775.7	25.15	3301	194	0.120	0.03102 ± 0.19	197.0 ± 0.8	0.4072 ± 0.21	346.8 ± 1.2	0.09518 ± 0.05	0.97	1531.7 ± 1.8
D,na,+149	0.566	781.6	51.99	7141	239	0.161	0.06116 ± 0.16	382.7 ± 1.2	0.8713 ± 0.17	636.3 ± 1.6	0.1033 ± 0.04	0.97	1684.9 ± 1.6
E,na,68	0.249	1106	30.10	5001	90	0.135	0.02586 ± 0.18	164.6 ± 0.6	0.3145 ± 0.19	277.6 ± 1.0	0.08818 ± 0.05	0.96	1386.4 ± 2.0
F,na,-62	0.363	1314	23.21	3587	142	0.141	0.01688 ± 0.17	107.9 ± 0.4	0.1759 ± 0.19	164.5 ± 0.6	0.07557 ± 0.07	0.93	1083.8 ± 2.8
(14) PCA-104-85, Quartz monzonite													
A,+149	0.175	635.2	7.147	3374	24	0.084	0.01156 ± 0.16	74.1 ± 0.2	0.07634 ± 0.19	74.7 ± 0.2	0.04790 ± 0.16	0.58	94.5 ± 7.2
B,na,125	0.172	1130	12.11	3872	36	0.060	0.01124 ± 0.08	72.1 ± 0.2	0.07429 ± 0.11	72.8 ± 0.2	0.04792 ± 0.06	0.85	95.4 ± 3.0
(15) PCA-21-84D, Metadiorite dyke													
A,na,149	0.244	612.9	48.91	13412	54	0.111	0.07639 ± 0.21	474.5 ± 1.8	1.084 ± 0.22	745.8 ± 2.2	0.1030 ± 0.04	0.98	1678.2 ± 1.4
B,na,90	0.473	571.4	30.70	6436	135	0.118	0.05102 ± 0.17	320.8 ± 1.0	0.7439 ± 0.18	564.7 ± 1.6	0.1058 ± 0.04	0.97	1727.4 ± 1.6
C,na,68	0.358	540.0	25.26	6432	83	0.134	0.04398 ± 0.17	277.5 ± 0.8	0.6171 ± 0.18	488.0 ± 1.4	0.1018 ± 0.04	0.97	1656.4 ± 1.6

** TIT = titanite, M = monazite; numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;

Weighing error = 0.001 mg.

\+ Radiogenic Pb.

* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU

\# Total common Pb in analysis corrected for fractionation and spike.

++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.

*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987). Fractions analyzed prior to 1985 utilized different techniques and very different minimum amounts of sample. The earlier fractions were spiked with a ^{208}Pb-^{233}U-^{235}U tracer, as opposed to the ^{205}Pb-^{233}U-^{235}U isotopic tracer. Earlier chemical and mass spectrometric procedures follow those of Parrish et al. (1987), except that those analyses preceded the direct propagation of errors and simultaneous use of Faraday cups and electron multiplier.

this, the ages have proven very useful in identifying different magmatic units that have been incorporated into maps of Carr et al. (1987) and Parrish et al. (1988).

U-Pb dating of metaplutonic rocks in the footwall of the Slocan Lake-Valkyr fault system was also undertaken and those rocks which are Eocene, in particular the Ladybird suite (Carr et al. 1987), have had their U-Pb ages presented in Parrish et al. (1988). U-Pb dates on older metaplutonic and metamorphic rocks within the complex are either presented in Parrish (1984), Parrish (1990), Heaman and Parrish (1991), or are presently being prepared for publication.

ANALYTICAL NOTES

U-Pb dating for rocks presented in this paper was done from 1985 to 1990. Older analyses had higher Pb blanks and higher weights of zircons, and were not all characterized by zircon abrasion. Where relevant, comments are made in the text which qualify some of the data. Otherwise, analytical methods were similar to those of Parrish et al. (1987) and are detailed at the bottom of Table 1. Titanite and allanite U-Pb ages use methods of Parrish et al. (1992); for common Pb corrections the model of Stacey and Kramers (1975) was used with ages equivalent to the crystallization age of the rock. Analytical data are presented in Table 1, and in the concordia plot figures. Error ellipses reflect the 2σ uncertainty in the U-Pb concordia diagrams. Exact sample localities are presented in the Appendix. K-Ar, and in two cases, ^{40}Ar-^{39}Ar age spectra on hornblende or biotite are referenced if published, or presented here. The morphology of the zircons analyzed in this study were all relatively typical of igneous zircons, with subhedral to euhedral shape, and relatively high clarity. Many of the samples contain inherited zircons, many of which were not particularly visible. In some of the strongly metamorphosed samples, there is a distinctive rounding of the euhedral zircon crystals, but otherwise, the morphologies of zircons from individual samples is not described. Where titanite and allanite are dated, their morphologies are euhedral to subhedral and they have an igneous origin as accessory minerals.

(1) EARLY JURASSIC DIORITE (PCA-64-85)

On the southwest side of the Valhalla complex in the headwaters of Ladybird Creek, there occurs a massive to strongly foliated hornblende-biotite diorite. It is well exposed along upper Ladybird Creek in the immediate hangingwall of Valkyr shear zone. It has a strong foliation adjacent to the shear zone, but gradually loses this fabric structurally upwards to the west, becoming massive at the locality where this sample was collected. It was thought that this diorite was part of the Nelson suite, but U-Pb dating demonstrates that it is considerably older. This unit appears to intrude metasedimentary rocks thought to be correlative with the Mt. Roberts formation just to the south. This diorite, as well as the Mt. Roberts rocks are truncated at the Valkyr shear zone and are not found in the Valhalla Complex footwall.

Analysis

Six analyses of abraded high quality zircons plot in a cluster on or slightly below concordia at 193-194 Ma (Fig. 2). The consistency of the data indicate that the age of zircon crystallization is best estimated at 193 ± 1 Ma.

Interpretation

The 193 Ma age for this pluton indicates that it is coeval with part of the Early Jurassic Rossland Group, extensively exposed at higher structural levels to the south and southwest. It is also coeval with other plutonic rocks of the southern Intermontane Belt, namely the Wild Horse, Bromley , and the Pennask batholiths west of the Okanagan Valley, about 200 km west (Parrish and Monger, 1992). Part of the explanation for the great distance separating these coeval plutons is that up to 100 km of east-west extension occurred during the Eocene and separated this belt of Early Jurassic magmatism.

K-Ar dating from this sample yielded a hornblende age of 107 ± 4 Ma (GSC K-Ar 89-65, Hunt and Roddick, 1990). A number of other samples of this lithology between this locality and the Valkyr shear zone were collected on an east-west traverse to assess the thermal impact of the Valkyr shear zone on mineral ages. From this sample eastwards, the biotite ages are all uniformly 48 ± 1 Ma, excepting one altered sample (GSC K-Ar 90-4, 5, 6, 7, and 8; Hunt and Roddick, 1991). Hornblende was analyzed from all of these same samples using the ^{40}Ar-^{39}Ar methods, and they have age spectra characteristic of extensive argon loss during an event about 60 Ma ago, with no steps indicating ages older than 120 Ma or so. The thermal event related to the Valkyr shear zone was sufficiently strong to completely reset biotites and cause >50% loss of argon from hornblende. These data will be published elsewhere.

(2) NELSON BATHOLITH, QUARTZ SYENITE (PCA-452-83-SY)

A narrow belt of hornblende (biotite) leucocratic quartz-bearing syenitic rocks is exposed in a north-south zone on the immediate east side of the Slocan Lake fault in the vicinity of Winlaw (Fig. 1). It is also exposed in the immediate hangingwall of the Valkyr shear zone on Robson Ridge southwest of Castlegar (sample 3). At locality 2 of Figure 1 (this sample), it is massive to foliated, fractured and somewhat hydrothermally altered, being within 100-200 m above the Slocan Lake fault. The foliation in this unit, however, is not tectonic but magmatic, with igneous textures in thin section and alignment of inclusions and mafic clots on outcrop scale exposures. In this respect, it is very similar to the Kuskanax Batholith described by Read (1973) and dated at 173 ± 5 Ma by Parrish and Wheeler (1983). This unit was recognized by Reesor (1965), but due to its undeformed nature, he considered it as one of the younger intrusions of the complex.

U-Pb analyses

Six zircon and three titanite fractions were analyzed; three of the zircon fractions were not abraded. Fractions A and C appear to have inheritance, but the other four zircons lie on or slightly to the right of concordia (Fig. 3). The consistency of the data indicates that a good estimate of the age of zircon crystallization is 170.5 ± 0.5 Ma, which is interpreted as the age of this rock. Titanite fractions mutually overlap concordia at 168 ± 2 Ma, indicating rapid cooling and confirming the 170.5 Ma zircon date. This age is the most precise obtained for the Nelson Batholith. Its age overlaps with that obtained for the Kuskanax Batholith to the north and could be the same alkalic subsuite of the Middle Jurassic magmatic episode.

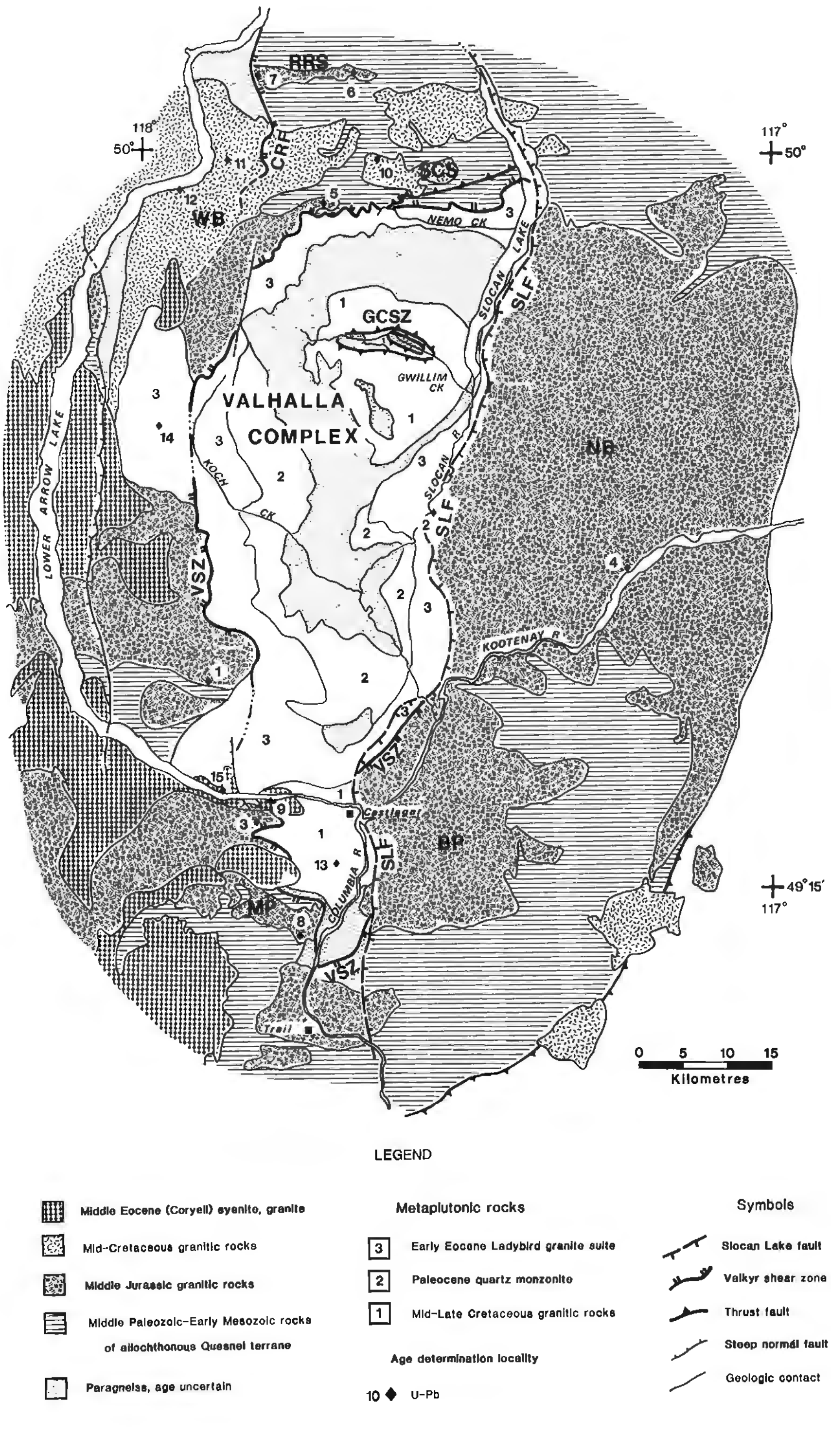

Figure 1. Geological map of Valhalla Complex and vicinity showing plutonic suites in the hangingwall of Slocan Lake fault zone and Valkyr shear zone, and localities of samples dated in this study. Abbreviations are as follows: SLF, Slocan Lake fault; VSZ, Valkyr shear zone; GCSZ, Gwillim Creek Shear Zone; NB, Nelson Batholith; MP, Mackie Pluton; BP, Bonnington Pluton; SCS, Snowslide Creek stock; RRS, Ruby Range stock; WB, Whatshan Batholith.

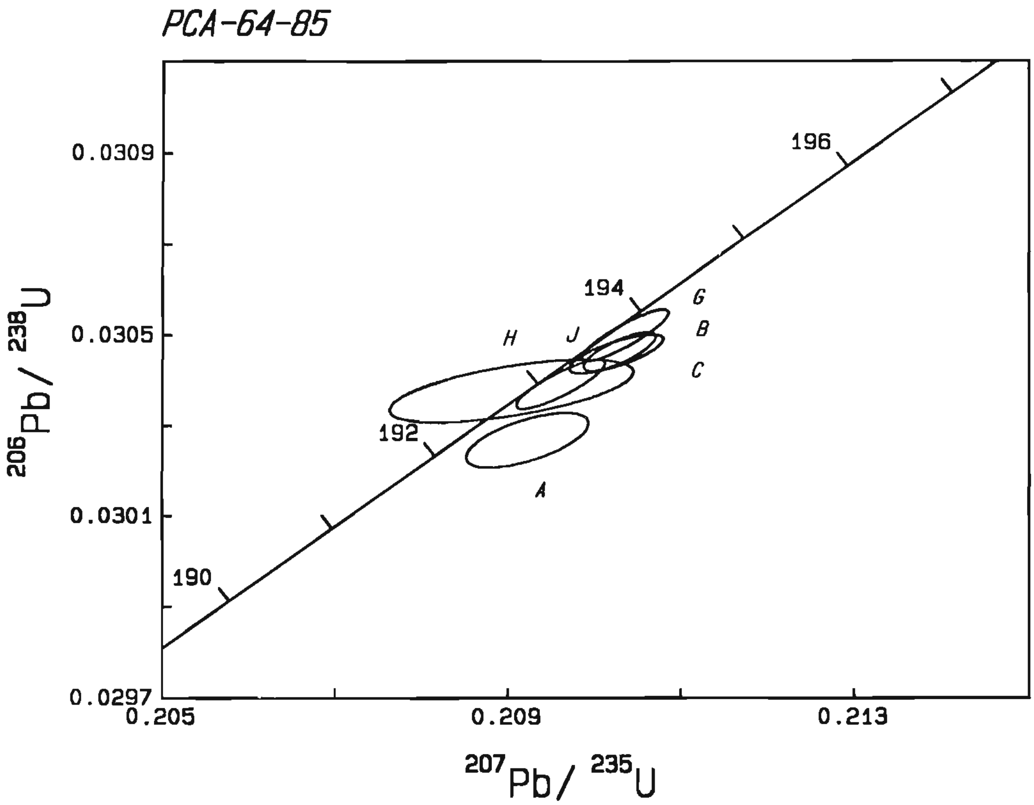

Figure 2. U-Pb concordia diagram for zircon from sample 1, PCA-64-85.

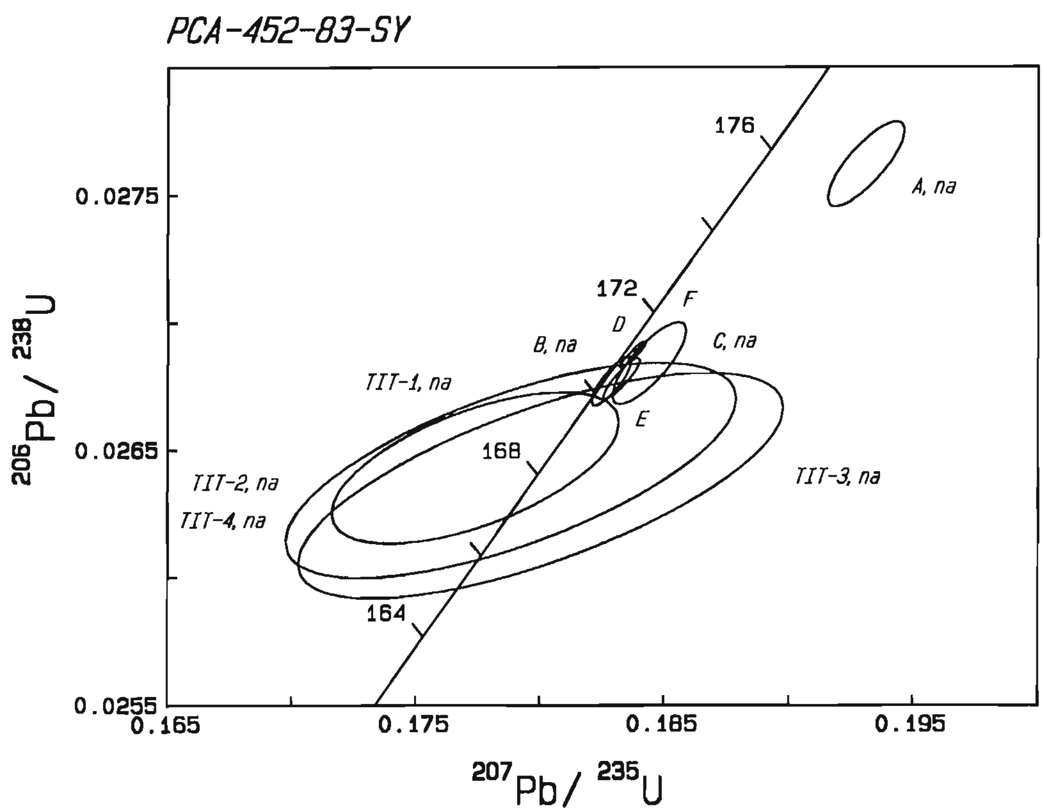

Figure 3. U-Pb concordia diagram for zircon and titanite from sample 2, PCA-452-83-SY; na, not abraded.

(3) NELSON BATHOLITH, QUARTZ SYENITE (PCA-34A-85)

This sample of foliated hornblende leucocratic quartz syenite is identical to the previous sample 2, except that it is well-foliated. It occurs in the hangingwall of the Valkyr shear zone in a structural position similar to the Early Jurassic diorite described above (sample 1). It has been metamorphosed, like the diorite, during the Eocene as a result of shearing and heating from below. The reason for dating this sample was to test correlation of these two leucocratic syenitic rocks. Other leucocratic syenitic rocks to the northwest appear similar, but are 52 Ma old (Parrish et al., 1988), and they also are locally sheared. The foliation in this sample dips west about 30°.

U-Pb analyses

Seven fractions of zircon were analyzed with results spread along and adjacent to concordia between 155 and 170 Ma (Fig. 4). This is a complex pattern given the high quality abraded zircons and the pattern requires an explanation in addition to zircon inheritance. Analyses B, G, and F have ^{207}Pb-^{206}Pb ages exceeding 190 Ma, and they are interpreted as having a significant component of inheritance, in addition to some Pb loss. Analyses A and C, and to a lesser extent D and E, are nearly concordant but they do not agree. The explanation for this dispersion parallel to concordia is suggested to be metamorphically-induced Pb loss from zircons which contained little or no inheritance, but which originally crystallized at least 169 Ma ago, the ^{206}Pb-^{238}U age of fraction E. A precise age cannot be determined, but it is felt that all fractions have experienced some Pb loss. If compared to the U-Pb data of the previous sample 2 (Fig. 3) which is massive and not metamorphosed, it is inferred that the two samples had much the same premetamorphic configuration of zircon analyses and an age of ca. 170-171 Ma. Subsequent metamorphism in this sample obscured and dispersed the originally simple pattern of zircon U-Pb systematics. This sample is thus inferred to be ca. 170 Ma old.

(4) NELSON BATHOLITH, GRANODIORITE (NB-7)

Hornblende-biotite megacrystic granodiorite typical of much of the Nelson Batholith is exposed in the vicinity of Nelson. This sample was collected by T.M. Harrison to confirm a Middle Jurassic crystallization age, despite the K-Ar and ^{40}Ar-^{39}Ar dates being considerably younger (Duncan and Parrish, 1979; T.M. Harrison, unpublished data).

U-Pb analysis

No great attempt was made to precisely date this sample, merely to confirm its Middle Jurassic crystallization age. Four zircon fractions are all discordant (Fig. 5), with 3 of the 4 indicating significant inheritance (fractions A, B, and C). Fraction D may have had only slight inheritance, but it, as

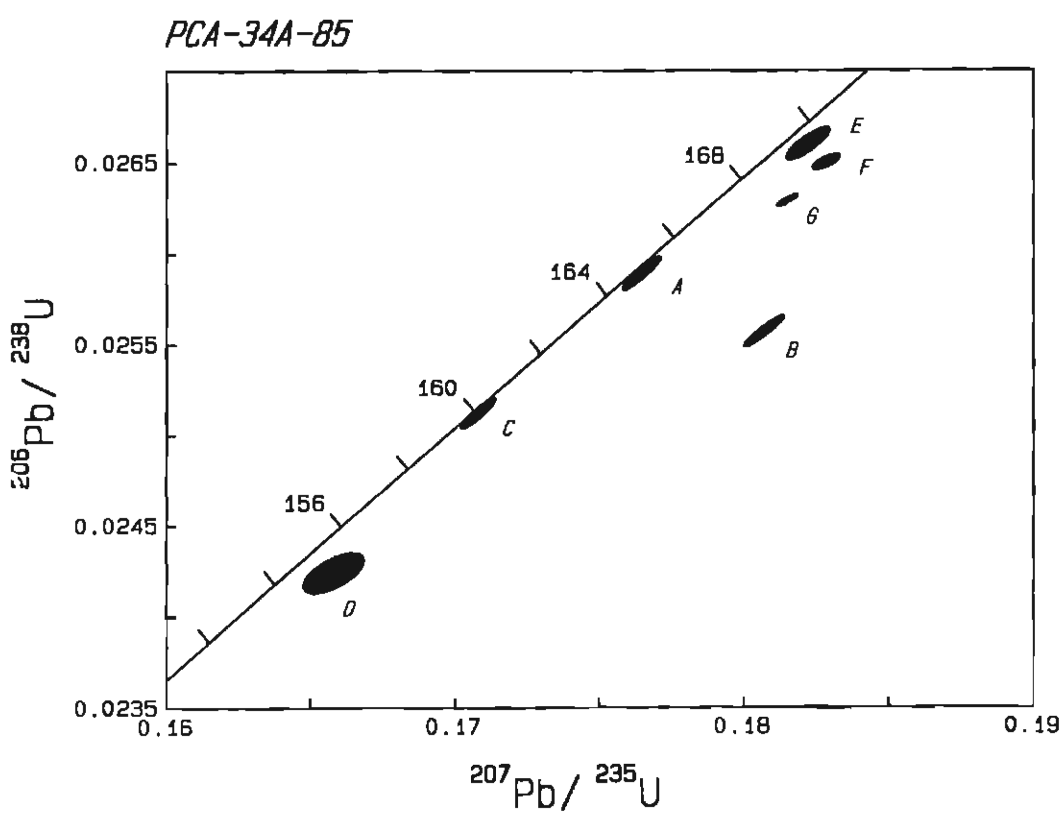

Figure 4. U-Pb concordia diagram for zircon from sample 3, PCA-34A-85.

well as the other fractions, has probably also lost some Pb. It is inferred that the age of zircon crystallization from this sample alone is between 160 and 170 Ma; the available data are insufficient to more clearly define its age.

(5) NELSON BATHOLITH, FOLIATED GRANODIORITE (PCA-251-85)

On the northwest side of Valhalla complex, hornblende-biotite-bearing megacrystic granodiorite is moderately to strongly foliated near the Valkyr shear zone. This pattern of foliation is similar to samples 2 and 4 above, and each of these three samples is in a similar structural position. This small hand sample was collected while traversing a ridge which crossed from the Valkyr shear zone into the hangingwall. This granodiorite forms a foliated, slightly discordant body intruding metasedimentary rocks on strike with the Paleozoic Nemo Lakes belt (Parrish, 1981) a few kilometres to the east (Fig. 1). The objective of dating this sample was to assess whether these megacrystic granitic rocks were correlative with the Nelson suite or the Late Cretaceous Whatshan Batholith, which is widespread just to the northwest (Fig. 1). The Whatshan can have a very similar appearance lithologically to the Nelson suite.

U-Pb and ^{40}Ar-^{39}Ar analyses

Four zircon and one titanite fractions were analyzed from this sample, in addition to hornblende for K-Ar and ^{40}Ar-^{39}Ar dating. The zircons have significant inheritance, like the other granodiorites of the Nelson Batholith, and plot in a scattered array with ^{206}Pb-^{238}U ages older than 170 Ma (Fig. 6). It is suggested that this pattern is consistent with this rock being of Nelson age, that is, 160-170 Ma old. A precise age cannot be inferred. Titanite yields an age of about 145 Ma. Although one could infer that the titanite yields the correct crystallization age, it is also consistent with it having lost Pb during metamorphism related to Eocene shearing on the Valkyr shear zone.

To clarify this problem, hornblende was analyzed first by the K-Ar method, and an age of 136 Ma was obtained (GSC K-Ar 88-6, Hunt and Roddick, 1988). It was further suspected that this age represented a partially reset age and ^{40}Ar-^{39}Ar dating was then obtained on the same hornblende at Queen's University with the assistance of S. Heinrich and D. Archibald using analytical methods of McBride et al. (1987). Figure 7 shows the age spectra; a table of the numerical data is available from the author upon request. Excess Ar is present in the first 5% of gas release and this is followed by a steady rise from ca. 70 Ma to 120 Ma. The remainder of gas release hovers around 120 Ma but does not satisfy criteria for a plateau. We suggest that this sample has experienced significant degassing at about 60 Ma due to heating and shearing from below during extension faulting. No particular significance is attributed to other age spectra steps shown in Figure 7.

(6, 7) RUBY RANGE STOCK (PCA-RRE-83 [6] AND PCA-RRW-83 [7])

North of the Valhalla complex in the Nakusp area mapped by Hyndman (1968), an elongate E-W oriented body termed the Ruby Range stock intrudes sedimentary and volcanic strata of the Slocan and Rossland(?) groups (Fig. 1). The plutonic

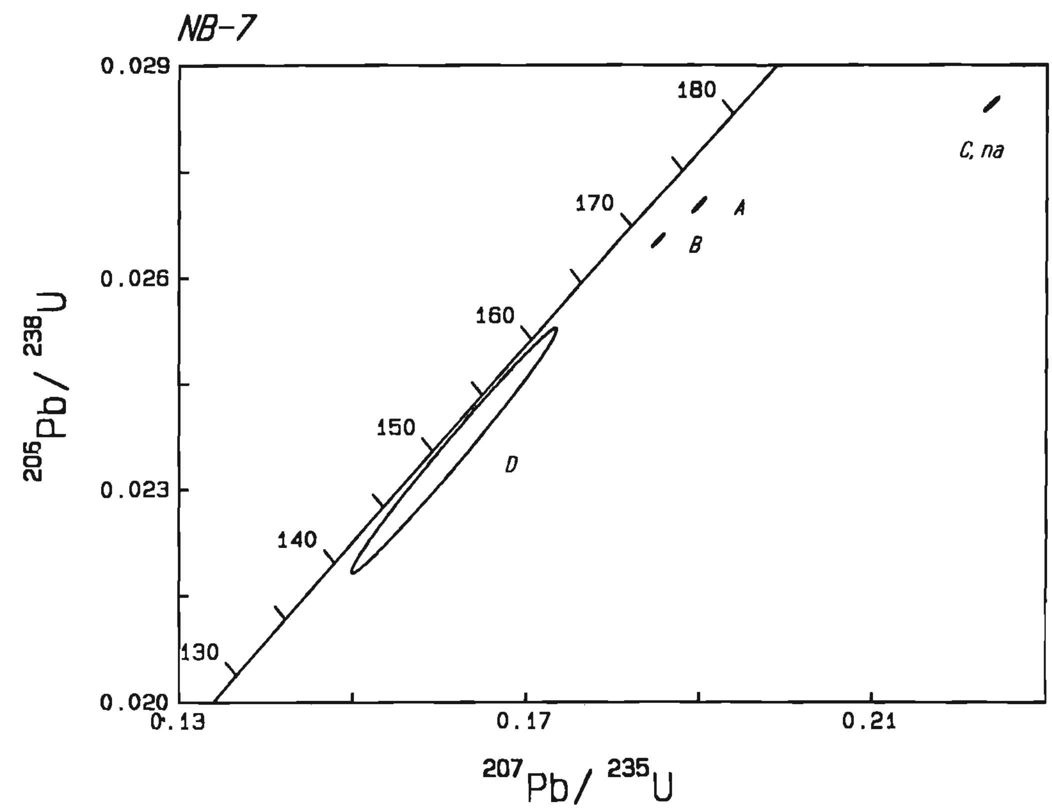

Figure 5. U-Pb concordia diagram for zircon from sample 4, NB-7.

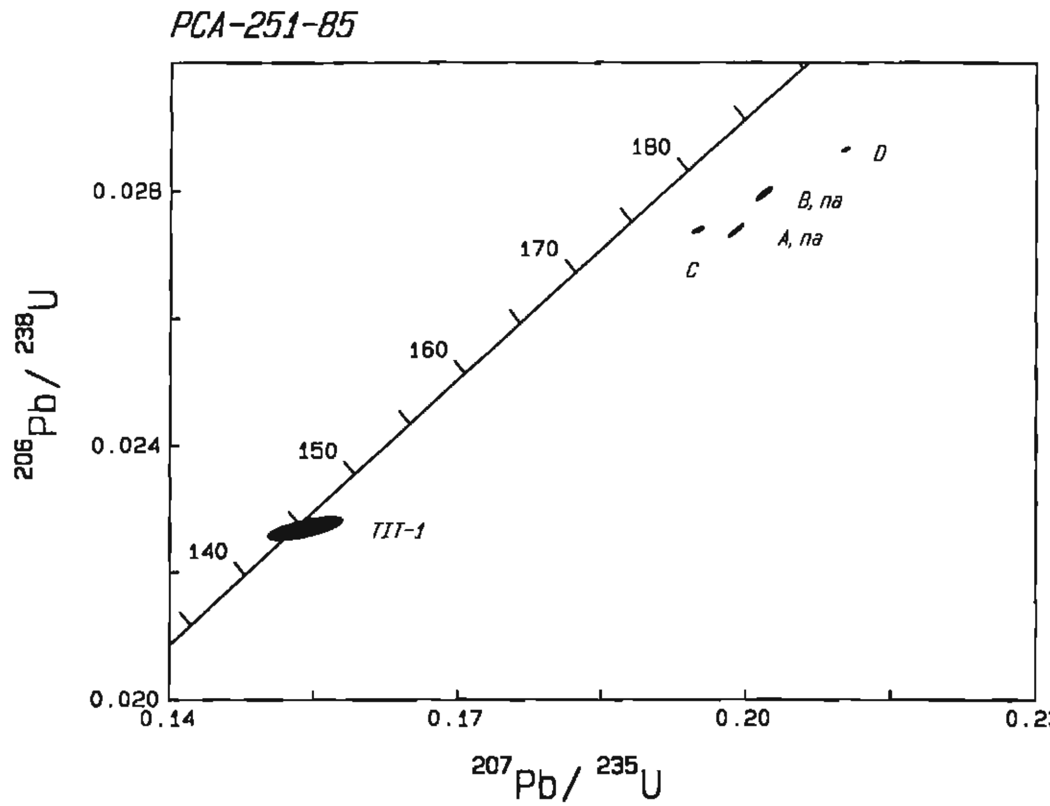

Figure 6. U-Pb concordia diagram for zircon and titanite from sample 5, PCA-251-85.

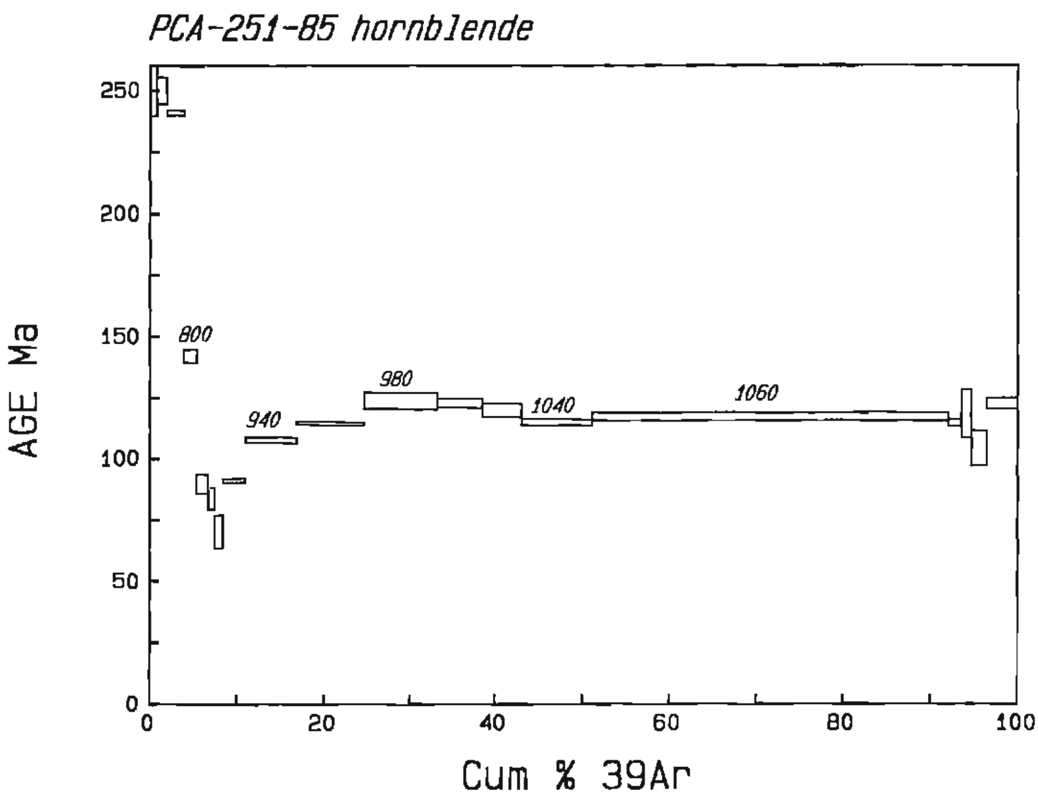

Figure 7. $^{40}Ar/^{39}Ar$ age spectra for hornblende from sample 5, PCA-251-85.

rocks range from quartz diorite to monzonite and syenite (Hyndman, 1968), and the samples dated are distinctly biotite-rich and porphyritic. Titanite is a common accessory mineral. The plutonic rocks are not metamorphosed, and they intrude folded and metamorphosed phyllite of the Slocan Group. Two samples were collected and dated, and their results are discussed together.

U-Pb analyses

Two zircon fractions were analyzed from each sample, and in addition, six titanite fractions were analyzed from the western sample. Like other Middle Jurassic Nelson suite rocks, the zircons have significant zircon inheritance, yielding an array which intersects concordia at about 155 Ma (Fig. 8). None of these zircons were abraded, as these analyses date from 1984-1985, before the author was routinely abrading Mesozoic and younger zircons. Six titanite fractions all yield self-consistent concordant data with a collective age of 167 ± 3 Ma, which is taken as the crystallization age. The lower intercept of the zircon array yield an erroneous age, and demonstrates that precise ages can only be obtained with confidence if analyses are very near the concordia curve or concordant. This sample is coeval with the Nelson suite of granitic rocks, although it is an entirely separate body, unlike the previous samples which were probably contiguous prior to Eocene extensional faulting.

(8) MACKIE PLUTON (PCA-27-85)

The Mackie Pluton (Simony, 1979) is an intrusion of hornblende-biotite granodioritic rocks west of the Columbia River Valley southwest of Castlegar (Fig. 1). It is compositionally similar to the Nelson suite and was mapped as such by Little (1960). It is generally confined to the hangingwall of the Valkyr shear zone as shown by Parrish et al. (1988) and in Figure 1. At its easternmost extent low in the Columbia Valley, it is thoroughly involved in ductile deformation in a wide zone of east-directed shearing interpreted by Parrish et al. (1988) as extensional in origin (i.e. the Valkyr shear zone). As one follows this body to the west, structurally upwards, it becomes less foliated in a manner identical to the Early Jurassic diorite (sample 1), and other Nelson suite samples 3 and 5. The Mackie Pluton intrudes mainly Mt. Roberts formation (Simony, 1979). This sample was collected for dating to confirm its suspected correlation with the Nelson suite.

U-Pb analyses

Four zircon fractions were analyzed, two of which were not abraded. Fractions A and B clearly have inheritance, and fractions C and D may have experienced some Pb loss (Fig. 9). It is difficult to define a precise age as a result of these problems, but the pluton is interpreted to be slightly older than the nearly concordant fractions C and D, ca.

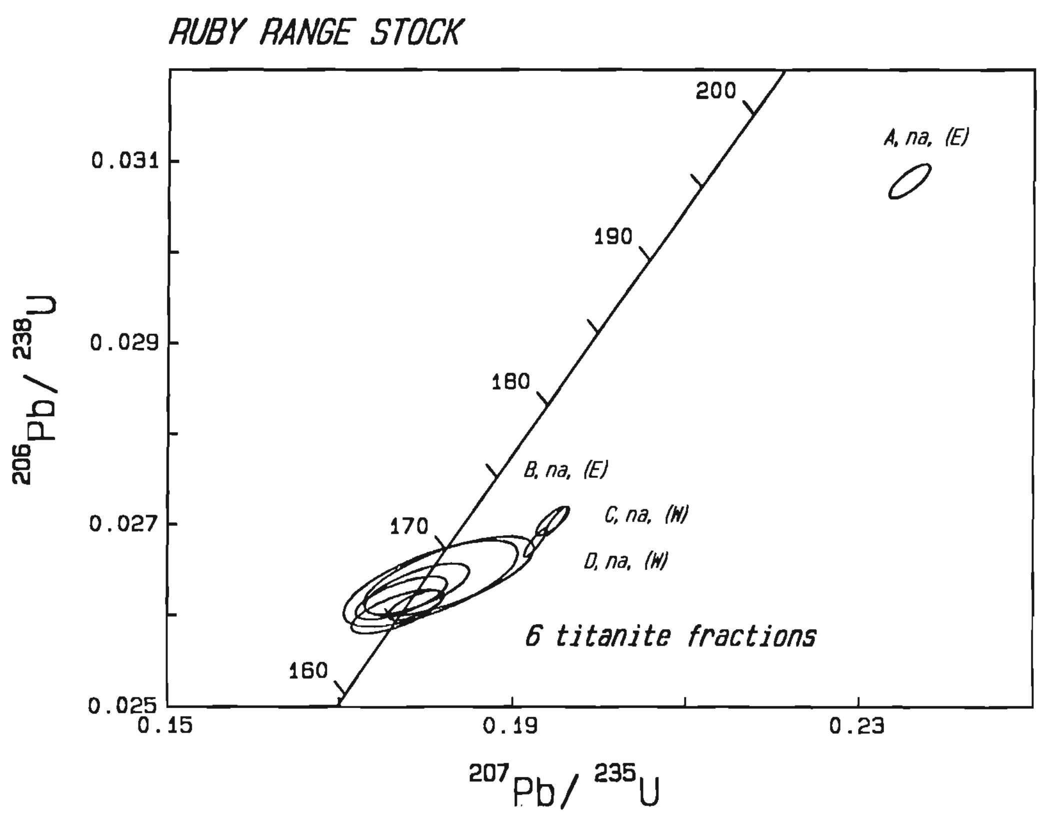

Figure 8. U-Pb concordia diagram for zircon and titanite from samples 6 and 7 of the Ruby Range stock, PCA-RRE-83 (6) and PCA-RRW-83 (7).

163 Ma, perhaps 163-166 Ma. Hornblende from this sample yielded a K-Ar age of 107 ± 2 Ma (GSC K-Ar 89-56, Hunt and Roddick, 1990); this age is interpreted as a partially reset age resulting from metamorphism and foliation development related to Eocene shearing on the Valkyr shear zone.

(9) DIORITE WEST OF KEENLEYSIDE DAM (PCA-35-84)

A complex of massive to foliated dioritic rocks veined with granitic veins occurs in the vicinity of Keenleyside Dam on the Columbia River west of Castlegar. The local geology is very complex, and too complicated to show in Figure 1. On the north side of the river west of the dam, slightly foliated Eocene Ladybird granite overlies dioritic rocks. On the south side, much less Ladybird granite is present, and it is found as concordant southerly and easterly tapering sheared sheets amongst several varieties of dioritic rocks, the coarser grained massive varieties of which are probably Eocene Coryell intrusions. The foliated and recrystallized diorite dated in this study is well exposed on the railway, west of the pumping station one kilometre west of the dam.

The exact location of the Valkyr shear zone in this area is uncertain due to the extensive intrusion of Coryell syenitic rocks (Eocene) and Ladybird granite, both of which occur in both hangingwall and footwall of the shear zone. This general pattern of rocks was shown in Parrish et al. (1988) and is reproduced with slight changes in Figure 1. Five kilometres east of locality 8 occurs the Castlegar gneiss which is entirely within the footwall of the shear zone at this latitude. The diorite is interpreted to have been sheared and metamorphosed within the Valkyr shear zone, and later intruded extensively by Eocene Ladybird granite and Coryell syenite.

The objective of dating this sample was to ascertain whether this diorite complex was Eocene (i.e. Coryell syenite), Jurassic (i.e. Nelson suite), or alternatively, part of the Valhalla Complex and more similar in age to the ca. 110 Ma Castlegar gneiss (Parrish et al., 1988).

U-Pb analyses

Seven zircon analyses were undertaken, but the data are complex and difficult to interpret (Fig. 10). The analyses are composed of abraded and unabraded fractions, and fraction G was strongly abraded but consisted of only a few grains and yielded an analysis with a large error. The pattern can be rationalized as consistent with an initial crystallization age of 160-170 Ma with an initial endowment of much older zircon inheritance, followed by considerable Pb loss during heating and/or metamorphism associated with Eocene magmatism and shearing on the Valkyr shear zone. Little else can be said, other than a correlation with the Nelson suite is the most plausible interpretation of the data.

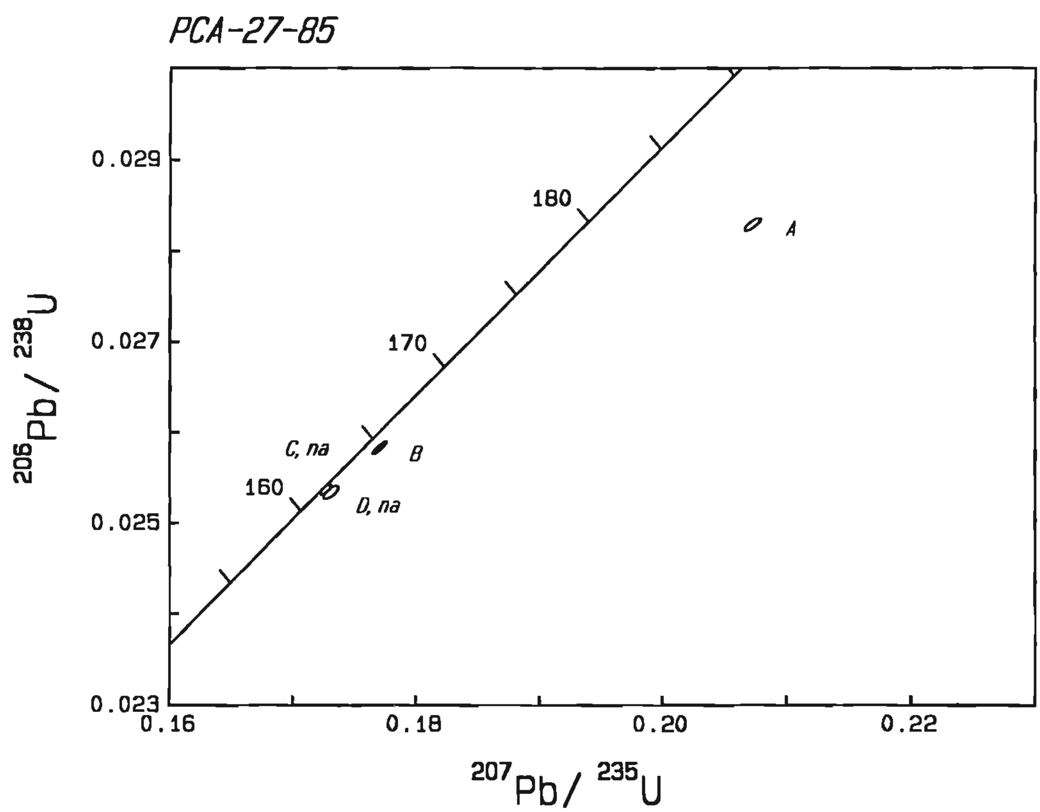

Figure 9. U-Pb concordia diagram for zircon from sample 8, PCA-27-85.

INTERPRETATION OF U-Pb AGES FROM THE NELSON SUITE

Samples 2-8 encircle the Valhalla Complex and indicate that granitoid rocks between 160-170 Ma old form most of the hangingwall of the Valkyr-Slocan Lake fault system. Restoration of the Eocene extensional faulting, such as was done in restored cross-sections in Parrish et al. (1988), strongly implies that most of these plutonic rocks, excepting the Ruby Range stock and perhaps the Mackie Pluton, were a contiguous plutonic body perhaps 50% larger than the presently exposed Nelson Batholith. The batholith was faulted and separated by Eocene displacement on the extensional fault system.

Samples very close to the Valkyr shear zone are strongly foliated and metamorphosed to amphibolite facies, affecting the zircon U-Pb and argon systematics by loss of daughter isotopes. These samples proximal to the shear zone were, in all likelihood, heated from below as a result of heat transfer from a hot footwall into a cooler hangingwall during juxtaposition of middle with upper crustal rocks.

Zircon inheritance presents major difficulties in obtaining precise ages of crystallization of Nelson suite magmas. Only when the magmas are alkalic in nature (sample 2 for example) is inheritance relatively minor; in this sample a precise age of 170.5 Ma was obtained. When the rocks have escaped subsequent metamorphism, i.e. the Ruby Range stock, titanite yielded an age of 167 ± 3 Ma. Extrapolations of linear arrays of inherited zircons to the lower intercept with the concordia curve, even when the linear fit is good, will not likely lead to a confident estimation of age because of Pb loss and variable age of inherited zircons. On this basis, it is concluded that the Nelson suite is mainly 165-171 Ma old. Additional work on zircons without inheritance will be required to refine this age span for the Nelson suite.

(10) SNOWSLIDE CREEK STOCK (PCA-SCS-83)

The Snowslide Creek stock is an intrusion of biotite quartz monzonite within Paleozoic-Mesozoic metasedimentary rocks in the hangingwall of Valkyr shear zone on the north side of Valhalla Complex. It was mapped by Little (1960), Hyndman (1968), and named by Parrish (1981). It crosscuts a cryptic Jurassic(?) terrane boundary fault marked by ultramafic rocks and described by Parrish (1981). It was suggested that the southern part of the pluton truncates the Mt. Meers fold, a major E-W structure on the north side of the complex near the head of Nemo Creek, but it may be deformed by the fold instead (S. Carr, personal communication, 1990). The pluton is restricted to the hangingwall of the Valkyr shear zone, although it lies in the footwall of the northern part of the Slocan Lake fault. It was dated to place constraints on the age of these structures. The sample dated contains biotite as the mafic mineral and allanite, zircon and apatite as accessory minerals.

U-Pb analysis

Three zircon fractions were analyzed in 1984-1985 prior to routine abrasion on young zircons, and yield slightly discordant points near 80 Ma on the concordia curve (Fig. 11).

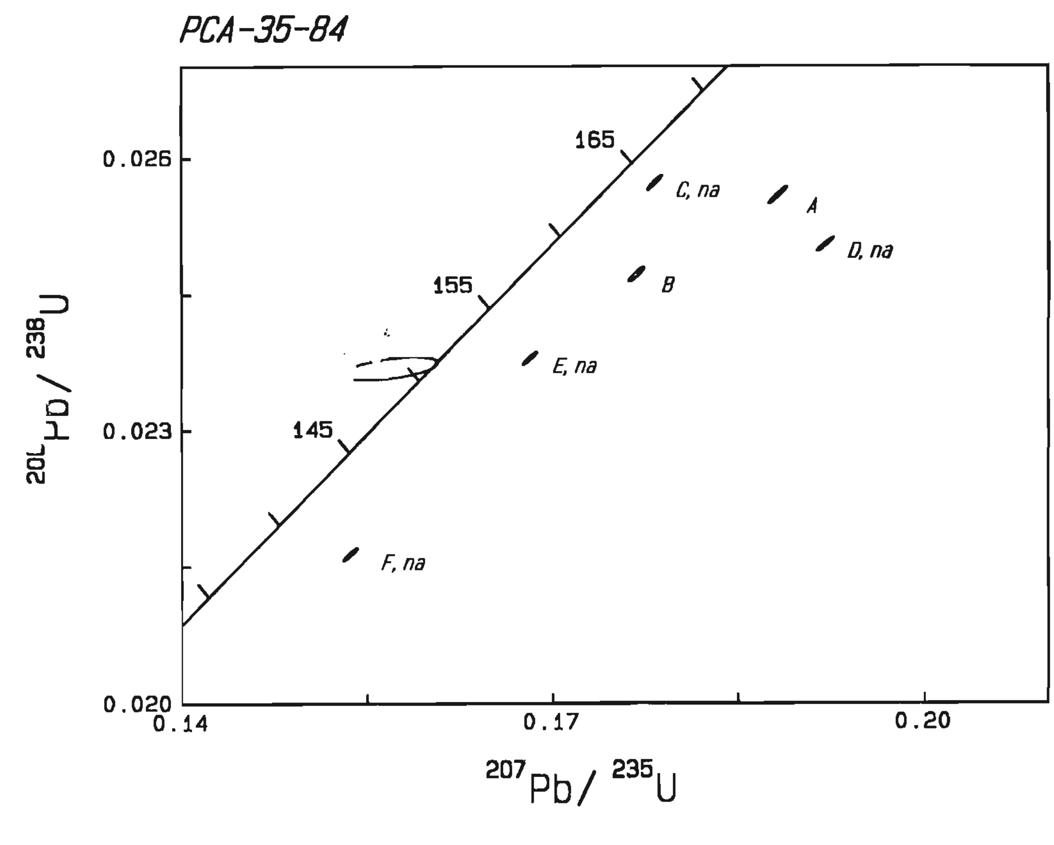

Figure 10. U-Pb concordia diagram for zircon from sample 9, PCA-35-84.

Inheritance is present in fractions A and B, and probably in fraction C to a lesser degree. Five allanite analyses plot as overlapping concordant ellipses centred on 85-86 Ma. Because allanite is a mineral strongly enriched in thorium like monazite, its U-Pb systematics can be characterized by significant amounts of excess ^{206}Pb and ^{206}Pb-^{238}U ages which are too old. Referring to Figure 4 of Parrish (1990), a reasonable estimation of the correction for this excess ^{206}Pb may be obtained by assuming that the Th/U for the rock is in the range 2-5, similar to many crustal granitoid rocks. The allanite fractions are characterized by radiogenic ^{208}Pb-^{206}Pb ratios of about 25 (Table 1), leading to excess ^{206}Pb equivalent to approximately 1.5-4.5 Ma of excess time on the ^{206}Pb-^{238}U age. The ellipses shown in Figure 11 are not corrected for this, but would move down about 2.0 ± 1.5 Ma using this correction and its assumed value for Th/U of the magma. Based on this reasoning, and the fact that the rock is not metamorphosed and should not have experienced Pb loss in allanite, we suggest that the best estimate of age for the Snowslide Creek stock is 84 ± 4 Ma. Although zircon fraction C is nearest to concordia, it is very high in uranium (4644 ppm) and is therefore suspected of having significant Pb loss.

(11, 12) WHATSHAN BATHOLITH (PCA-264-85 [11] and PCA-WHAT-83 [12])

The Whatshan Batholith is a composite batholith of hornblende granodiorite, quartz monzonite, and subordinate quartz diorite. Minor amounts of biotite are usually present and K-feldspar megacrysts are common and often abundant and very coarse. Accessory minerals are zircon, titanite and apatite. The batholith underlies an area northwest of the Valhalla Complex (Fig. 1) in the vicinity of the town of Burton, and farther west in the area of Whatshan Lake on the south side of the Pinnacles area (Parrish et al., 1988; Carr, in press). It intrudes other older igneous and metamorphic rocks, and is locally strongly foliated where adjacent extensional faults and shear zones of Eocene age are present. Much of it was originally termed the Caribou Creek stock of Hyndman (1968). It is cut by the Eocene Columbia River fault (Fig. 1; Parrish et al., 1988) near the fault's southern termination, but it also occurs in the hangingwall of the fault south of the Ruby Range stock. Except for its low content of biotite, it strongly resembles the Nelson Batholith, and prior to this study its age was not known. Samples 10 and 11 were collected in two different localities. Both are megacrystic hornblende (biotite) granodiorites with weak foliation.

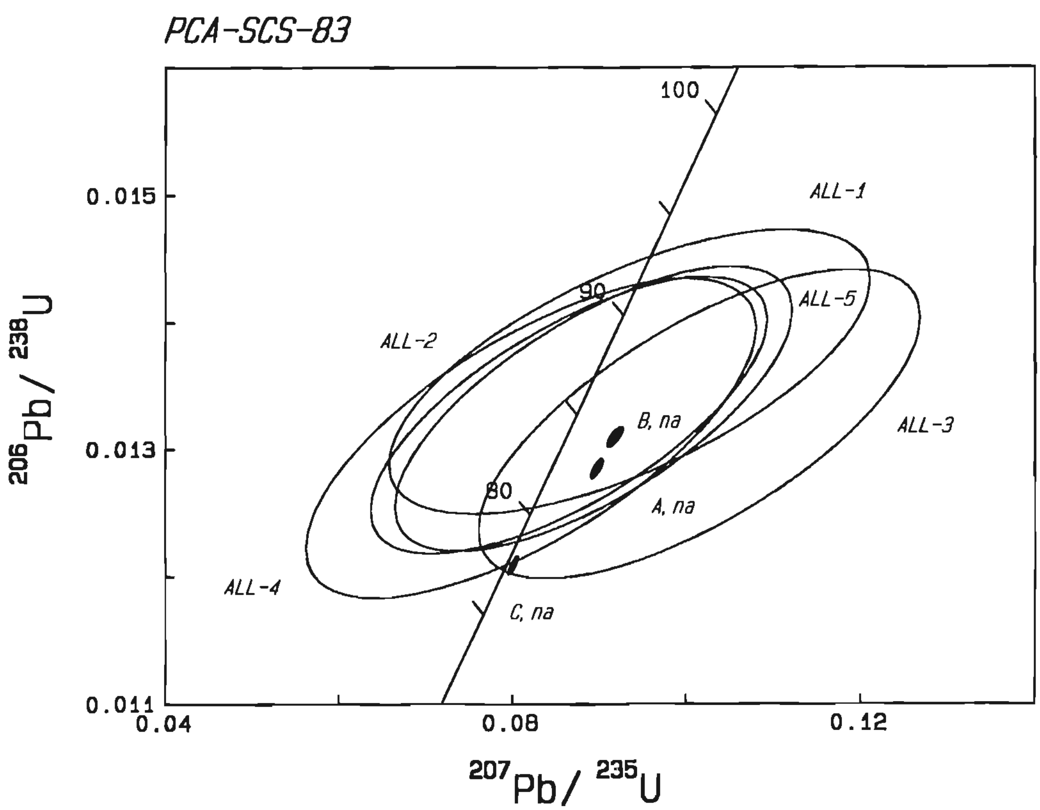

Figure 11. U-Pb concordia diagram for zircon and allanite from sample 10, PCA-SCS-83.

U-Pb analysis of PCA-264-85 (11)

Sample 11, from north of Burton (Fig. 1), has 2 titanite and 4 zircon analyses. The zircons are nearly concordant at 76-77 Ma with ^{207}Pb-^{206}Pb ages of 81-84 Ma, but are dispersed in a short linear array (Fig. 12). Correction for ^{206}Pb defiency due to original deficiency of ^{230}Th will decrease the ^{207}Pb-^{206}Pb ages by about 3-4 Ma (see Coleman and Parrish, in press), and we interpret ca. 79 ± 2 Ma as a reasonable estimate of the zircon crystallization age. There is no evidence for zircon inheritance in these fractions. Titanites have ages of 72 ± 1 Ma, which, with reasoning given below, is suggested to be either a cooling age or a partially reset age as a result of heating during Eocene extension. Hornblende from this sample had a K-Ar age of 65 ± 1 Ma (GSC K-Ar 89-57; Hunt and Roddick, 1990), which is also attributed to partial resetting or slow cooling prior to more rapid cooling during Eocene extension.

U-Pb analysis of PCA-WHAT-83 (12)

Sample 12 is from a locality southwest of sample 11, farther from the Columbia River fault. Its U-Pb analyses of zircon were done in 1984 and were not abraded. Its two zircons are essentially duplicates of each other just to the right of concordia at 79 Ma (Fig. 13). It is not certain whether there is either zircon inheritance or alternatively, an identical amount of Pb loss in these two fractions to explain their similarity. There are insufficient data to suggest a precise age, but it is inferred that the consistency of U-Pb ages indicates an age of ca. 79-80 Ma. A lone titanite analysis overlaps concordia at 77 ± 1 Ma; this could be either the crystallization age, an age of post-magmatic cooling, or it could be slightly reset. In any case the minimum age of the Whatshan Batholith is 77 Ma, derived from analysis of data from both samples 10 and 11.

The K-Ar age of hornblende from this sample is 74 ± 3 Ma (GSC K-Ar 88-7; Hunt and Roddick, 1988). In an attempt to clarify the interpretation of titanite and K-Ar dates, hornblende from this sample was analyzed by the ^{40}Ar-^{39}Ar method at the Geological Survey of Canada by J. C. Roddick using analytical methods outlined in Roddick (1990); a detailed table of isotopic data is available upon request. The age spectra is presented in Figure 14. This 11-step spectra has several initial steps ranging in age from 68-70 Ma which rise to steps of 75-80 Ma for most of the gas released. There is no proper plateau, and the spectra is interpreted as indicating cooling below hornblende Ar closure temperature 75-80 Ma ago, followed by minor Ar loss during reheating(?) during events (magmatic and tectonic) associated with extension during the Eocene.

Comments on the Whatshan Batholith

Both samples of the Whatshan Batholith indicate crystallization of the magma 77-81 Ma ago, consistent with zircon, titanite, and hornblende Ar data, and we suggest that 79 ± 2 Ma is a reasonable estimate of crystallization age. The rock probably took a few million years to cool below closure temperatures of about 600°C for titanite (Heaman and Parrish, 1991) and about 500-550°C for argon in hornblende. During the Eocene, when magmatic rocks of the 52 Ma Coryell suite

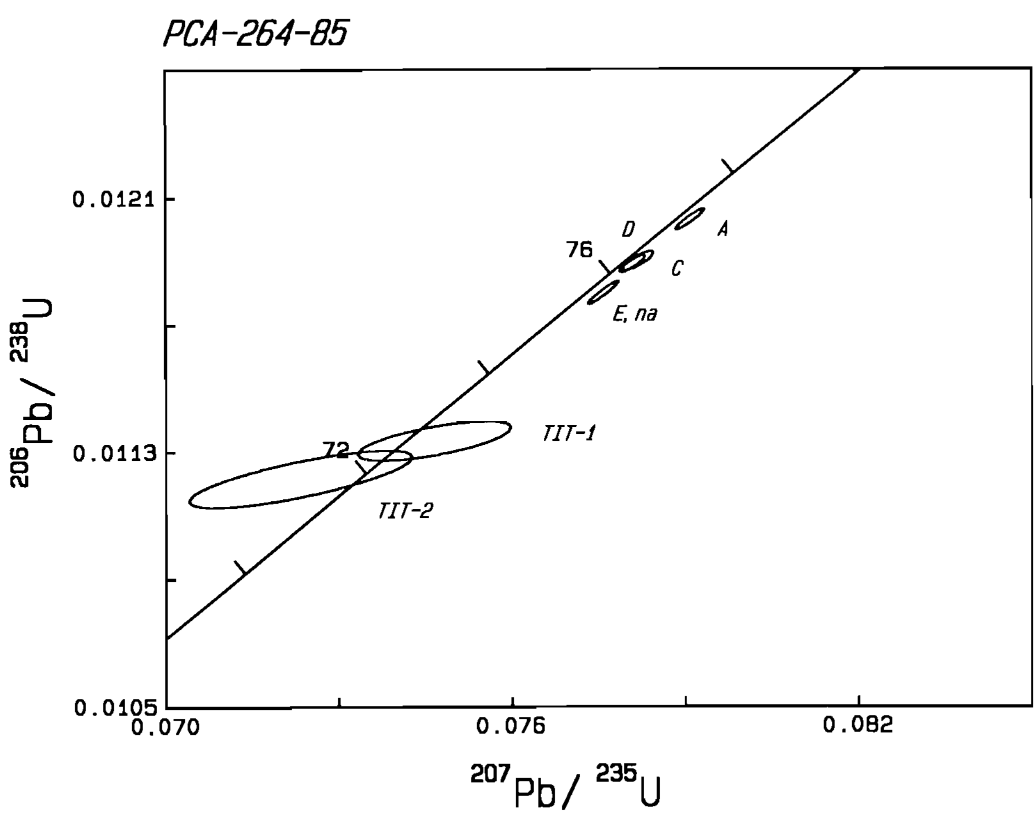

Figure 12. U-Pb concordia diagram for zircon and titanite from sample 11, PCA-264-85.

were intruded and when faulting on the Columbia River fault was active (Parrish et al., 1988), there was minor radiogenic daughter loss.

(13) LATE GRANITE SOUTHWEST OF CASTLEGAR (PCA-56-84)

A small granite body intrudes the strongly deformed Castlegar gneiss about 4-5 km southwest of Castlegar, but is too small to show separately in Figure 1. It is a biotite (hornblende) quartz monzonite with abundant inclusions of country rocks, and it is unfoliated, postdating the strong deformation in rocks of the Valhalla Complex. This locality was field trip stop 2-5 of Parrish et al. (1985b). It is similar to other post-tectonic granitic rocks mapped on the south side of Valhalla Complex, another of which has been dated at 47 Ma (Parrish et al., 1988). It was collected to estimate the youngest age of ductile deformation of the gneisses of southern Valhalla Complex.

U-Pb analysis

Six zircons were analyzed from this sample and all of them show large amounts of zircon inheritance. The six analyses do not fit a line within uncertainty, but they do define a good linear array with lower intercept of 62 ± 8 Ma, which is an approximate age for the body (Fig. 15). The assignment of age with this kind of discordant data, however, is not sufficiently confident to rule out an age similar to the other dated 47 Ma granite to the south which has essentially concordant data (Parrish et al., 1988). It is unlikely that the rock is older than 62-63 Ma, the age of the Airy quartz monzonite gneiss (unit 2 of figure 1; Parrish, 1984), since the Airy gneiss is strongly foliated and ductilely deformed. Therefore it is suggested that this granite sample 13 is at most 60 Ma and could be as young as ca. 47 Ma, the age of other similar post-tectonic granite intrusions in southern Valhalla Complex.

(14) QUARTZ MONZONITE, WESTERN VALHALLA RANGE (PCA-104-85)

A hornblende-biotite quartz monzonite outcrops on the road paralleling the power line west of upper Koch Creek west of the Valhalla Complex in an area intruded extensively by Ladybird granite (Fig. 1). The intent of dating this sample was to assess whether this rock was more similar to the Whatshan Batholith or the Nelson suite; it shares lithological similarities to both, and differs from the biotite granite of the Eocene Ladybird suite. This rock was sampled and analyses were produced in 1987 prior to completion of mapping on the west side of the complex. During this mapping it was noted just how extensive Ladybird granite really was, and it is likely that the hornblende quartz monzonite of sample 14 is a large raft within Ladybird granite. Sample 14 is not foliated and its contact relationships were not investigated in detail.

U-Pb analysis

Only two zircon fractions were analyzed, and they are discordant and lie beside the concordia curve adjacent to 72-74 Ma (Fig. 16). No further work was done to more

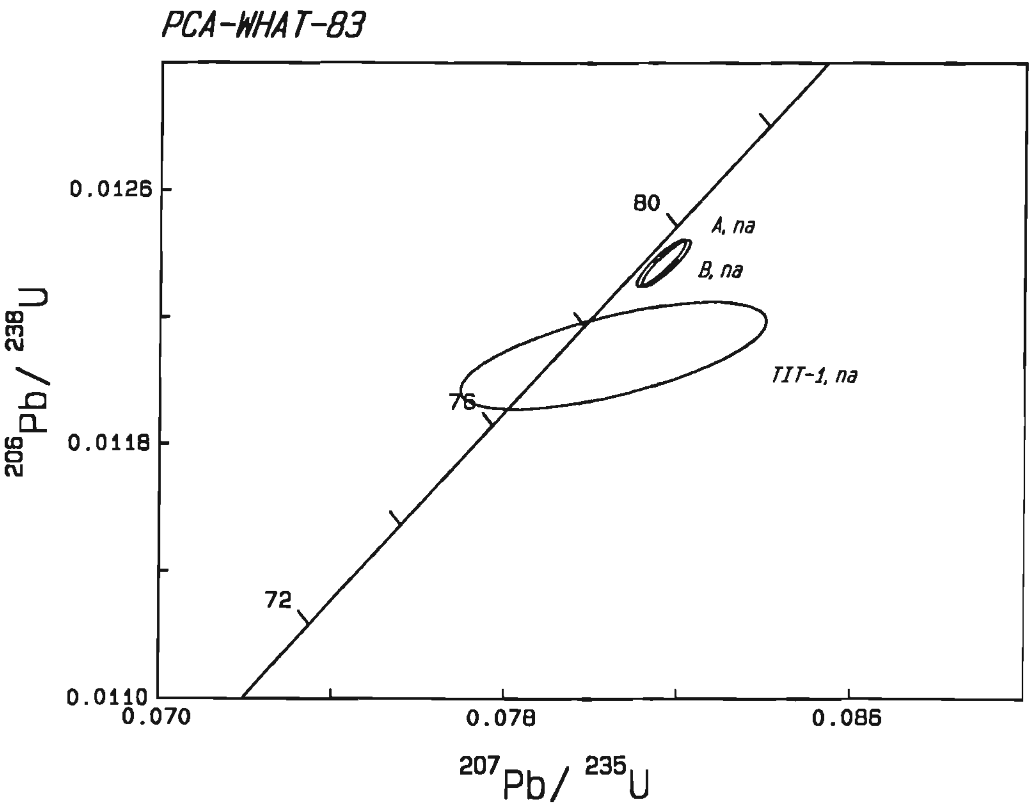

Figure 13. U-Pb concordia diagram for zircon and allanite from sample 12, PCA-WHAT-83.

precisely assess this rock's age, since its field relationships are only vaguely known. However, comparison of the U-Pb pattern of zircons with those of both Nelson (samples 2-8) and Whatshan (samples 10 and 11) suites indicates a strong similarity with Whatshan zircons and suggests a correlation with it. It would be very difficult to correlate this body with the Nelson suite because one would be required to have 80-90% Pb loss during the Paleocene-Eocene, and this degree of Pb loss has not been documented in zircons (Heaman and Parrish, 1991). We therefore suggest based on analogy to samples 10 and 11 that this sample is also ca. 79 Ma old. Its hornblende K-Ar age is 56 ± 2 Ma (GSC K-Ar 89-63; Hunt and Roddick, 1990), indicating essentially total degassing by being engulfed within the Early Eocene Ladybird granite.

(15) METADIORITE DYKE, SOUTHWEST VALHALLA COMPLEX (PCA-21-84D)

A suite of deformed and recrystallized mafic dykes intrude the Coryell syenite near Allandale Creek in the southern Valhalla Complex (Fig. 1). These dykes are sheared and have a mineralogy characteristic of lower amphibolite facies. When this sample was collected, the age of the host syenite was suspected, but not known to be of Coryell (i.e. 52 Ma) age. The dykes intrude the syenite, yet were deformed, and it was thought that a U-Pb age on them might refine the younger limit of mafic magmatism and late ductile deformation of the complex. These dykes were sampled by Marquis and Irving (1990) in a paleomagnetic study which documented significant westward tilt during the Eocene.

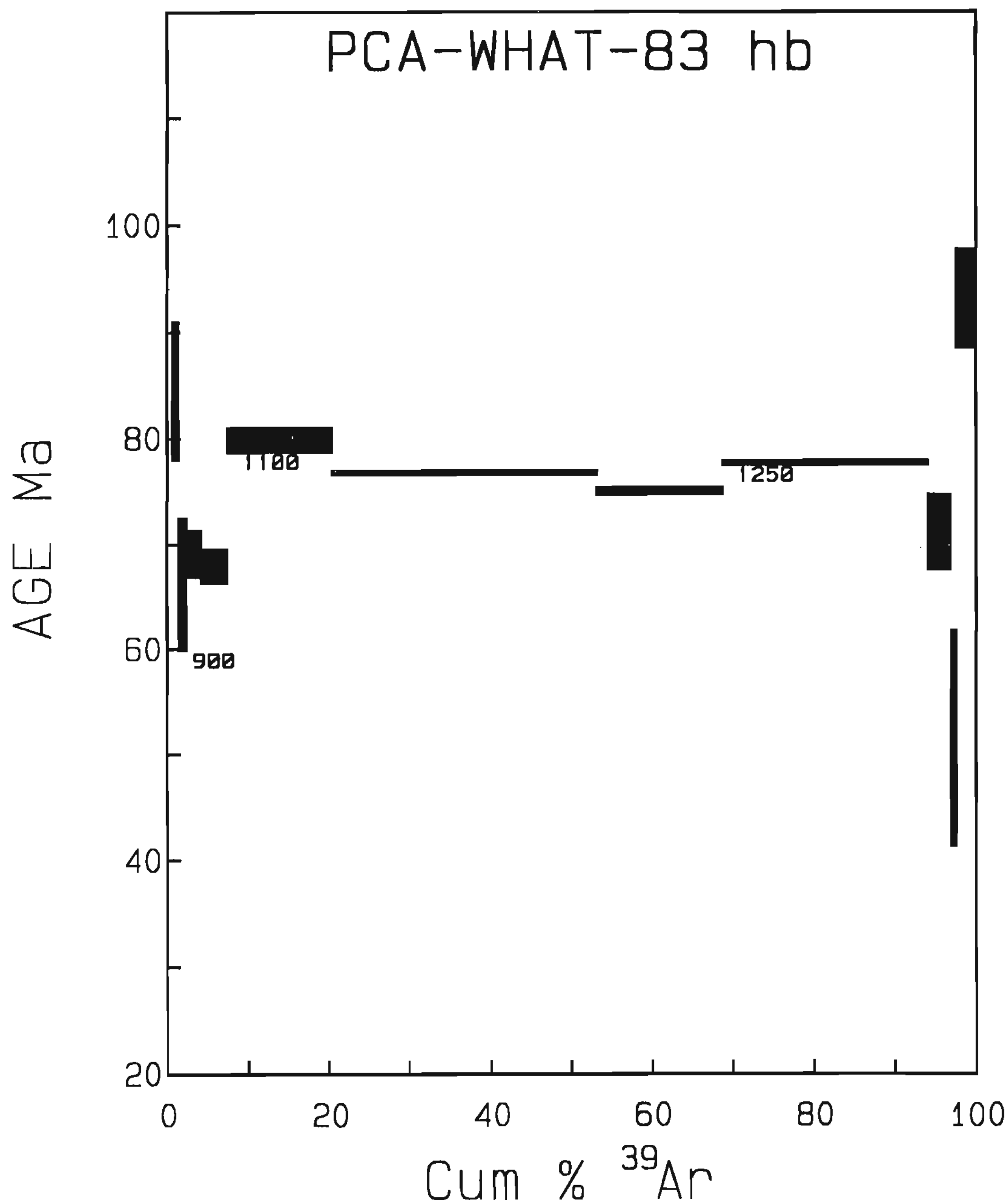

Figure 14. $^{40}Ar/^{39}Ar$ age spectra for hornblende from sample 12, PCA-WHAT-83..

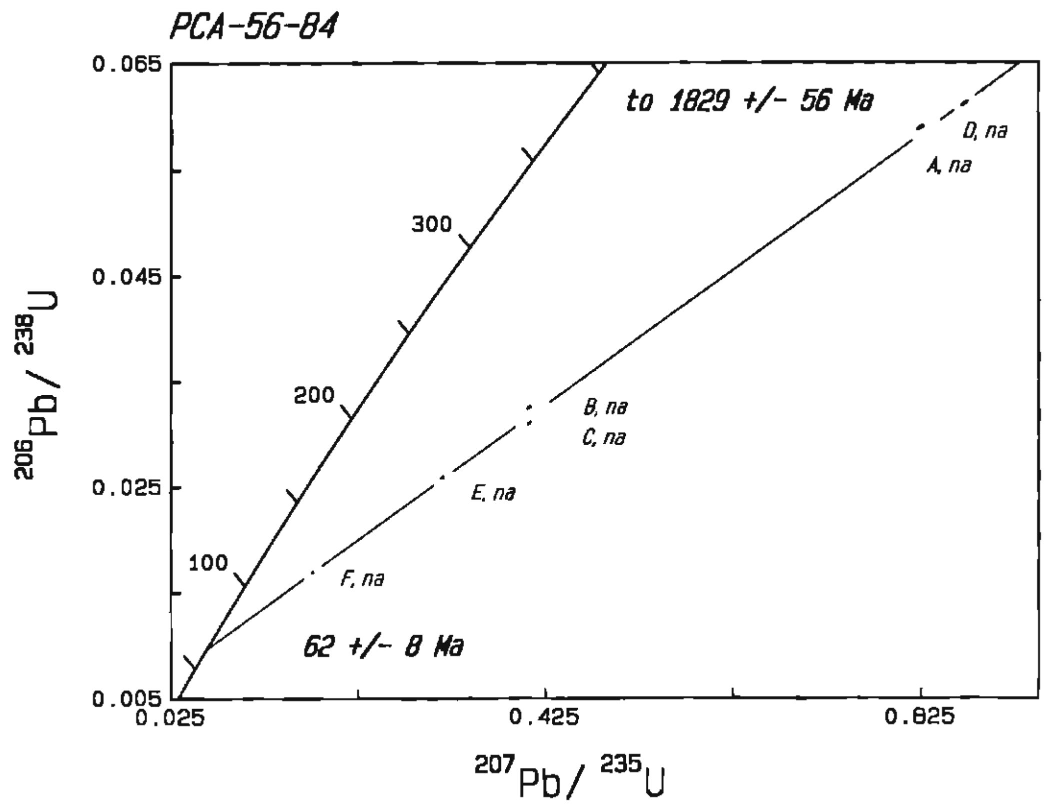

Figure 15. U-Pb concordia diagram for zircon from sample 13, PCA-56-84.

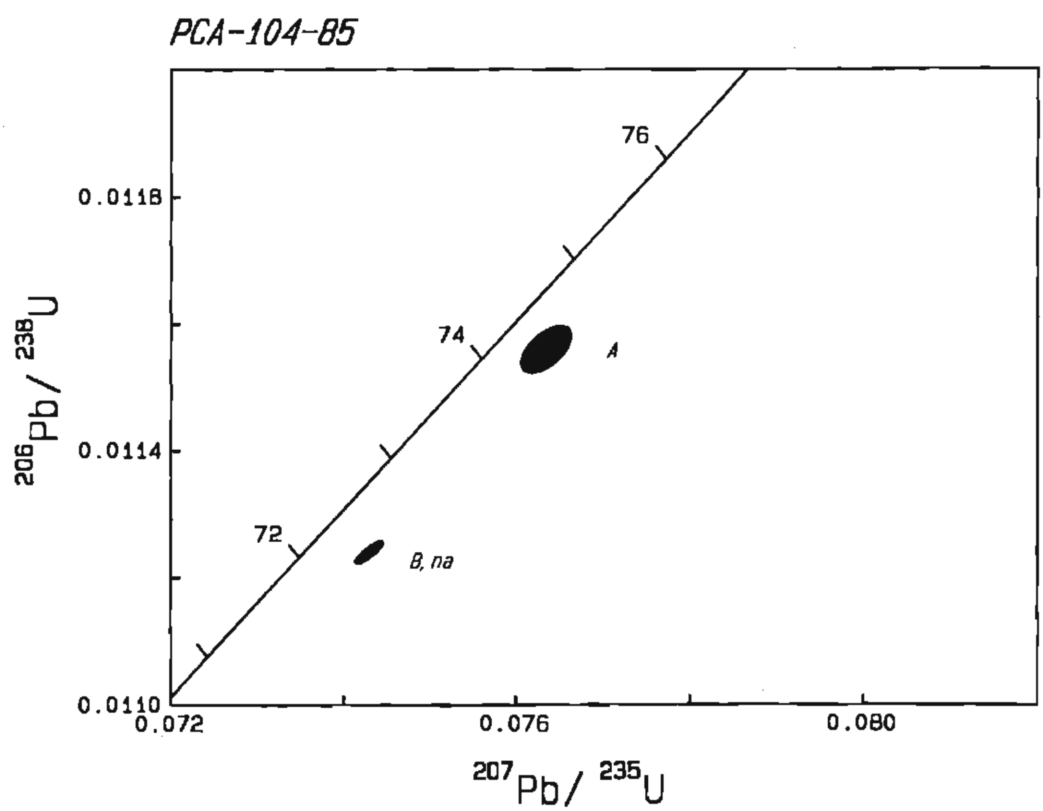

Figure 16. U-Pb concordia diagram for zircon from sample 14, PCA-104-85.

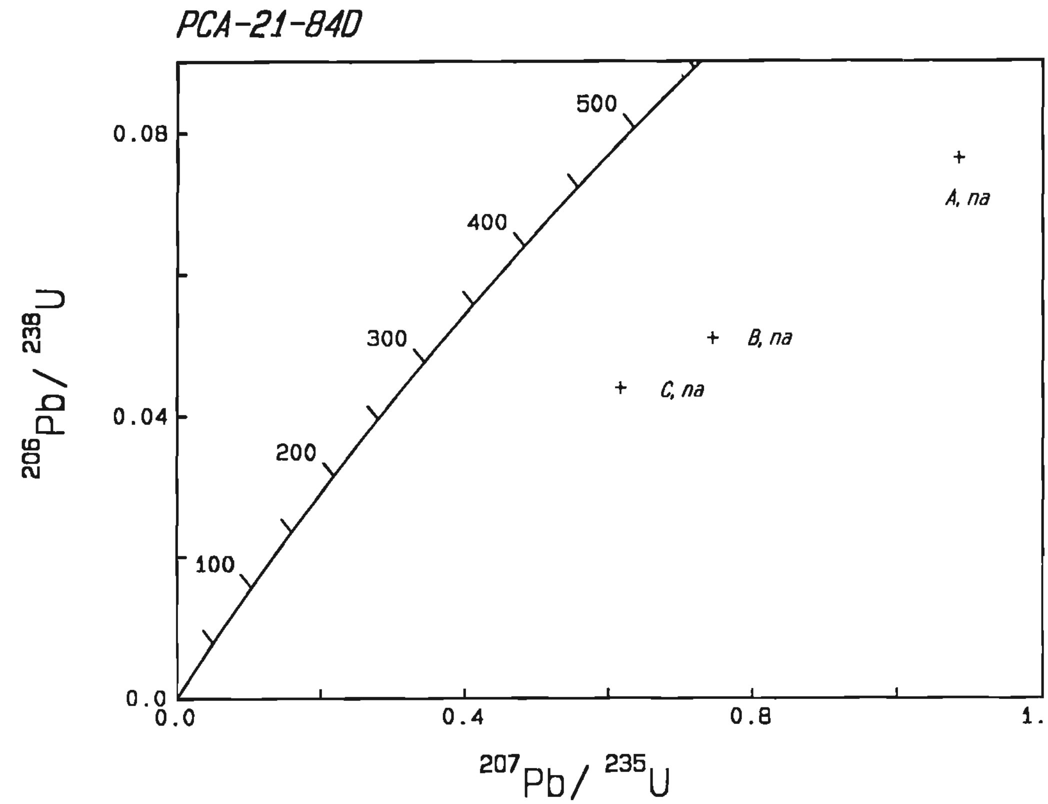

Figure 17. U-Pb concordia diagram for zircon from sample 15, PCA-21-84D.

U-Pb analysis

Three zircons were analyzed and are plotted in Figure 17. All three show huge amounts of Precambrian zircon inheritance, and do not allow an age interpretation other than Mesozoic or Tertiary to be made. The geological interpretation of age is much better: these dykes cut the 52 Ma Coryell syenite and are younger than it. They are ductilely deformed and therefore probably predate the intrusion of ca. 47 Ma granites which are unfoliated and post-tectonic (Parrish et al., 1988). Therefore this sample is interpreted to be 47-52 Ma old. The zircons are almost entirely xenocrysts of extraneous origin.

DISCUSSION

This paper has presented U-Pb data for fourteen samples of magmatic rocks, summarized hornblende K-Ar data available for many of these, and presented two new hornblende ^{40}Ar-^{39}Ar age spectra.

The oldest rock is a quartz diorite 194 Ma old, which is coeval with Rossland volcanic rocks and correlative with other ca. 194 Ma intrusions of the southern Intermontane Belt west of the Okanagan Valley (Parrish and Monger, 1992). It is the only plutonic rock of this age known to extend this far east. It intrudes rocks thought to be correlative with the Upper Paleozoic Mt. Roberts Group.

The extensive Nelson batholithic suite contains alkalic to calc-alkalic plutonic rocks 165-170 Ma old; the best ages are 168-171 Ma, and are derived from either alkaline intrusions or titanite ages from unmetamorphosed parts of the batholith. Zircon inheritance is ubiquitous in the calc-alkaline varieties, and precise age determinations are difficult to obtain. The Nelson Batholith was more extensive prior to Eocene extensional faulting, and the rocks of the batholith presently are found encircling the Valhalla Complex where they occur entirely above the Valkyr-Slocan Lake extensional fault system. This geometry of crustal rocks and faults is outlined in Figure 18, an east-west cross-section modified slightly from Parrish et al. (1988). In the section, the Nelson, Whatshan, and Ladybird plutonic suites are identified. Plutonic rocks within Valhalla Complex below this fault system are entirely younger than the Nelson suite. This intriguing pattern has been discussed and explained in Parrish et al. (1988). To summarize briefly, the Nelson suite and their enclosing Paleozoic-Mesozoic host rocks have been significantly transported eastwards on Cretaceous-Paleocene thrust faults during the development of the foreland fold and thrust belt of the Rocky Mountains.

Plutonism continued in the late Cretaceous with the emplacement of the extensive 79 Ma Whatshan Batholith, represented by three samples dated in this study. It is found in both footwall and hangingwall of the southern part of the Columbia River normal fault northwest of the Valhalla Complex.

Within Valhalla Complex, magmatic rocks have been strongly deformed, and range in age from about 110 Ma to 56 Ma, with pulses at ca. 100-110 Ma (Mulvey and Castlegar gneisses), ca. 70-80 Ma (pegmatites), 63 Ma (Airy quartz monzonite gneiss), and 59-56 Ma (Ladybird granite).

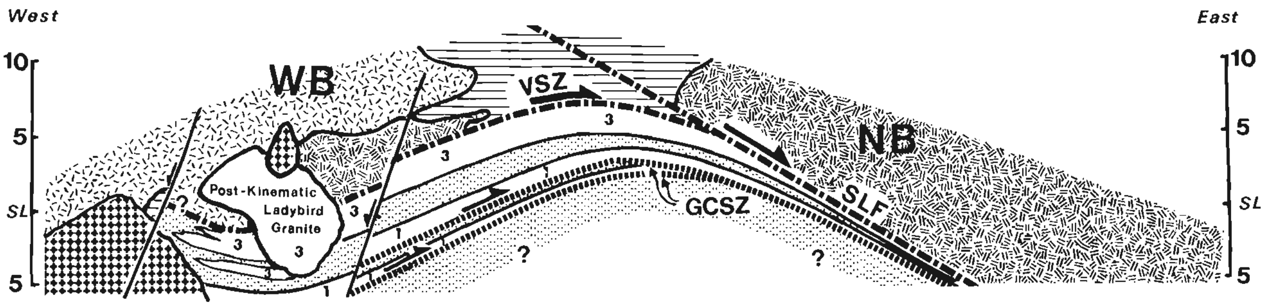

Figure 18. Schematic east-west cross-section of Valhalla Complex, modified from Parrish et al. (1988). The location of the Nelson Batholith (NB), Whatshan Batholith (WB), Coryell and Ladybird plutonic suites, Slocan Lake fault (SLF), Gwillim Creek shear zone (GCSZ),

These ages derive from Parrish (1984), Parrish et al. (1988), Heaman and Parrish (1991) and unpublished data presently being prepared for publication.

Just prior to and during the Eocene period of crustal extension, magmatic rocks of the Ladybird granite suite were intruded in both hangingwall and footwalls of the normal faults, and finally, later middle Eocene intrusion of syenite (Coryell suite, 52 Ma) and granite (post-tectonic intrusions, ca. 47 Ma) complete the plutonic history of the region.

ACKNOWLEDGMENTS

I thank the staff of the Geochronology Laboratory for assistance in generating these data; in addition Vicki McNicoll helped prepare the U-Pb table and figures, J.C. Roddick kindly provided the age spectra shown in Figure 14, and S. Heinrich, formerly of Queen's University, provided the age spectra shown in Figure 7. S. Carr of Carleton University helped in fieldwork during mapping of Valhalla Complex, and developed some of the ideas elaborated in this paper. The paper was reviewed by R. Stern.

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APPENDIX 1
Sample locations

All of these samples were collected by the author, in part with the assistance of S. Carr, except as noted.

Sample 1, PCA-64-85. Hornblende quartz diorite from roadcut along logging road in upper Ladybird Creek drainage, B.C., elevation 5000 ft., UTM zone 11u, 436150E, 5478050N, NTS 82F.

Sample 2, PCA-452-83-SY. Hornblende leucocratic quartz syenite from small roadside quarry along Highway 6, 3 km northeast of Winlaw, B.C., elevation 1800 ft., UTM zone 11u, 461200E, 5497400N, NTS 82F.

Sample 3, PCA-34A-85. Hornblende leucocratic quartz syenite from logging road on Robson Ridge, elevation 4650 ft., 2.5 km south of Arrow Lake, UTM zone 11u, 438300E, 54622500N, 13.5 km west of Castlegar, B.C., NTS 82F.

Sample 4, NB-7. Hornblende-biotite granodiorite on the north shore of the west arm of Kootenay Lake near the mouth of Kokanee Creek, B.C., approximately 49°37′N, 117°7′W, NTS 82F. Collected by T.M. Harrison

Sample 5, PCA-251-85. Hornblende-biotite granodiorite from a small peak on a ridge at elevation 7300 ft., 3km NNW of Avis Lakes, B.C., UTM zone 11u, 449100E, 5531700N, NTS 82F.

Sample 6, PCA-RRE-83. Hornblende granodiorite from roadcut in logging road, on ridge south of McDonald Creek, B.C., elevation 6300 ft., UTM zone 11u, 453700E, 5547200N, NTS 82K.

Sample 7, PCA-RRW-83. Hornblende granodiorite from roadcut in logging road, 4 km east of Scalping Knife Mountain, B.C., UTM zone 11u, 441600E, 5547300N, NTS 82K.

Sample 8, PCA-27-85. Hornblende biotite quartz diorite on roadcut on Sullivan Creek logging road at elevation 3000 ft., 3 km southwest of Genelle, B.C., UTM zone 11u, 446150E, 5449250 N, NTS 82F.

Sample 9, PCA-35-84. Diorite along railway cuts on south side of Arrow Lake near the pumping station, elevation 1500 ft., 1 km west of Keenleyside dam, B.C., UTM zone 11u, 443000E, 5464700N, NTS 82F.

Sample 10, PCA-SCS-83. Biotite quartz monzonite from logging road cut in upper Caribou Creek, elevation 5200 ft., UTM zone 11u, 455100E, 5538800N, NTS 82K.

Sample 11, PCA-264-85. Hornblende quartz monzonite from logging roadcut 1.8 km east of Burton, B.C., elevation 1950 ft., UTM zone 11u, 438900E, 5537300N, NTS 82F.

Sample 12, PCA-WHAT-83. Hornblende-biotite granodiorite from roadcut on highway 6, 5 km east of Caribou Point along the shore of Arrow Lake, UTM zone 11u, 433100E, 5534600N, NTS 82F.

Sample 13, PCA-56-84. Biotite granite-quartz monzonite, from highway 3 roadcut 4 km west of Kinnaird B.C., 200 m north of junction of Blueberry and Judkin Creeks, elevation 2400 ft., UTM zone 11u, 449900E, 5457700N, NTS 82F.

Sample 14, PCA-104-85. Hornblende-biotite quartz monzonite from roadcut along powerline access logging road, elevation 5600 ft., west of Koch Creek, B.C., UTM zone 11u, 430330E, 5507500N, NTS 82F.

Sample 15, PCA-21-84D. Recrystallized diorite dyke intruding Coryell syenite, roadcut on Deer Creek Road at point of land 700 m west of the mouth of Allandale Creek, B.C., elevation 1500 ft., UTM zone 11u, 437600E, 5465750N, NTS 82F.

- 14 -

Middle Jurassic plutonism in the Kootenay Terrane, northern Selkirk Mountains, British Columbia

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Brown, R.L., McNicoll, V.J., Parrish, R.R., and Scammell, R.J., 1992: Middle Jurassic plutonism in the Kootenay Terrane, northern Selkirk Mountains, British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 135-141.

Abstract

U-Pb age determinations of the Fang and Pass Creek plutons, 168 ± 2 (zircon) and 168 ± 3 Ma (titanite) respectively, establish the importance of pre-Late Jurassic strain within the Kootenay terrane of the northern Selkirk Mountains, British Columbia.

Résumé

Les datations U-Pb des plutons de Fang et Pass Creek à 168 ± 2 Ma (zircon) et à 168 ± 3 Ma (titanite) respectivement, confirment l'importance de la déformation antérieure au Jurassique supérieur au sein du terrane de Kootenay dans le nord des monts en Selkirk Colombie-Britannique.

INTRODUCTION

Published data pertaining to the timing of deformation events and regional metamorphisms within the Kootenay terrane of the southern Omineca belt suggest a complex history. In the southern part of the terrane (Kootenay arc) there is evidence of Paleozoic deformation (Wheeler, 1968; Read, 1976; Klepacki, 1985) with superimposed events that occurred in the Middle Jurassic and Late Cretaceous (Archibald et al., 1983). In the northern part of the terrane (Selkirk Mountains and Cariboo Mountains) evidence of major deformation in the Paleozoic has not been found. Plutonism occurred in the Devonian within the terrane (see Wheeler and McFeely, 1987) but the orogenic significance of this event remains unclear. Major deformation in the northern part of the Kootenay terrane has generally been ascribed to Late Jurassic and Cretaceous activity (Price and Mountjoy, 1970; Price, 1986), but Brown et al. (1986) have proposed a model which assumes a period of major deformation and regional metamorphism in the Middle Jurassic.

The location of the area and a simplified geological map are shown in Figure 1. The reader is referred to Brown (1991) and Wheeler (1963, 1965) for stratigraphic and structural details, and to Wheeler and McFeely (1987) for the location and tectonic setting of the Kootenay terrane.

The purpose of this paper is to present the results of U-Pb geochronology obtained from samples of the Fang and Pass Creek plutons which intrude Lower Paleozoic strata of the Hamill Group, Badshot Formation, and Lardeau Group in the Kootenay terrane in the northern Selkirk Mountains (Fig. 1). The geological setting and local field relationships of the plutons are outlined and the significance of the data is discussed. These data firmly establish the importance of pre-Late Jurassic strain within the Kootenay terrane.

GEOLOGICAL SETTING

The plutons and surrounding geology were first mapped by Wheeler (1963, 1965) as part of the 1:250 000 scale Rodgers Pass and Big Bend map sheets. In the accompanying report

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to the Rodgers Pass sheet, Wheeler suggested that the Fang pluton is Late Jurassic in age and post-tectonic. The pluton has yielded a biotite K-Ar date of 172 Ma (recalculated with new constants, Leech et al., 1963) and a hornblende K-Ar dates of 164 Ma (Wanless et al., 1972) and 185 ± 3 Ma (GSC-K-Ar-89-69; Hunt and Roddick, 1990). The oldest of these dates was ascribed in part to excess argon.

Brown and Tippett (1978) reported results of additional mapping in the Big Bend area, and referred to the Fang pluton to support their contention that the major deformation and associated regional metamorphism occurred before the Late Jurassic. Price (1979) pointed out that structures on Wheeler's map appear to wrap around the pluton and therefore may post-date its emplacement. In later publications Price (1981,1986) suggested an Early Cretaceous age for the regional deformation and associated metamorphism. With these uncertainties in mind, geological relationships of the Fang pluton and similar Pass Creek pluton have been re-examined.

In the summer of 1986 the Fang pluton was sampled for U-Pb geochronology and its contact relationships were examined. The Pass Creek pluton and surrounding geology have been mapped at a scale of 1:50 000 by Höy (1979) and Brown (1991), and the structural setting of the area has recently been discussed by Brown and Lane (1988). Based on regional structural and metamorphic relationships, Brown and Lane (1988) suggested that the major deformation and metamorphism of the region occurred in the Middle Jurassic. In particular a major westerly verging anticlinal fold nappe, the Carnes Nappe, is considered to have formed in the Middle Jurassic during the accretion of oceanic and island arc terranes (Superterrane 1, see Monger et al., 1982) to the margin of North America. Reliable geochronologic data confirming this model have not been available. The Pass Creek pluton has intruded the overturned limb of the Carnes Nappe, and the time of emplacement of this pluton therefore places an upper limit on the age of the nappe.

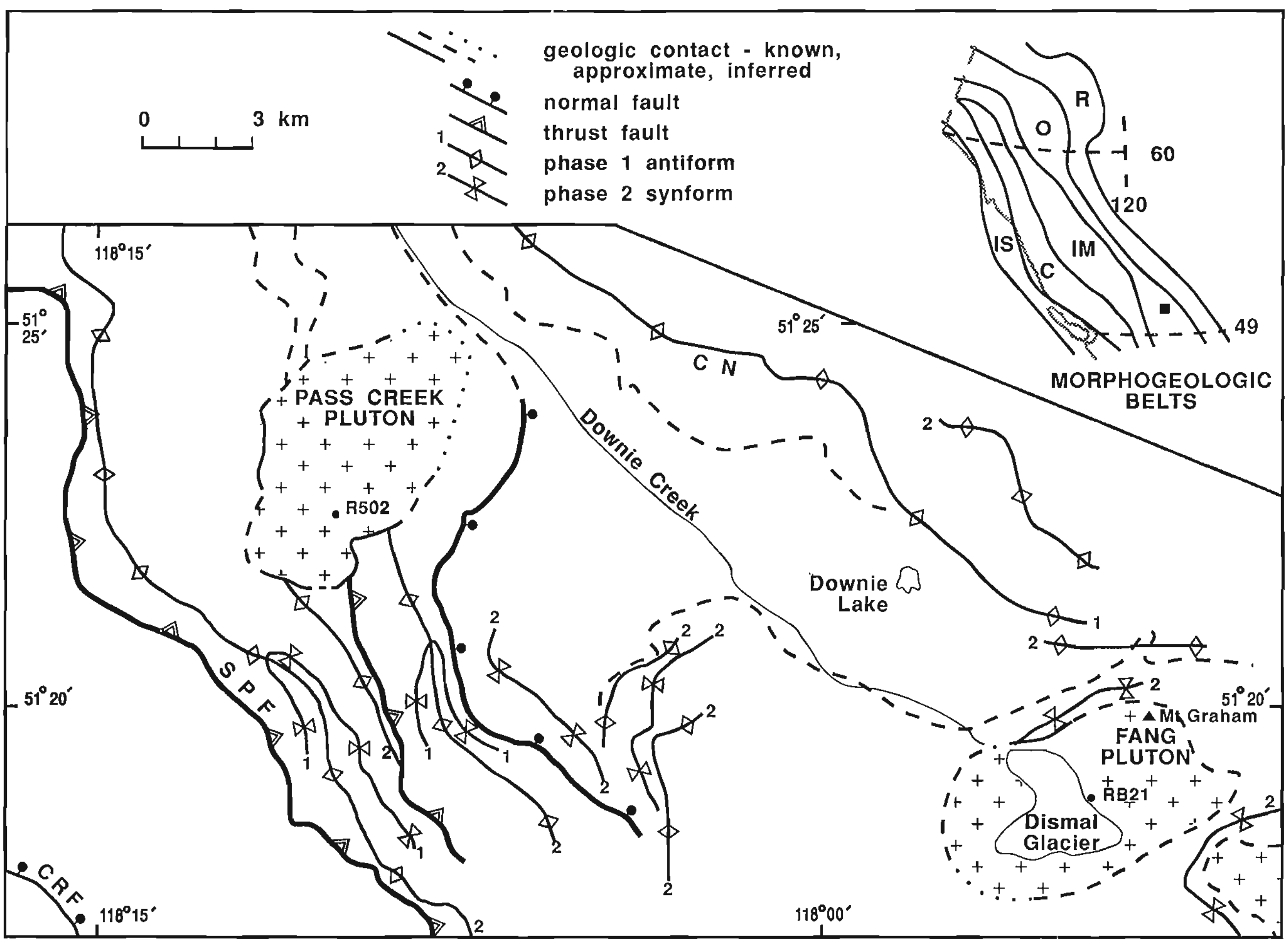


Figure 1. Location and geological setting of Pass Creek and Fang plutons. Black square in insert of Morphological Belts of the Canadian Cordillera locates the area. R – Rocky Mountain Belt, O – Omineca Belt, IM – Intermontane Belt, C – Coast Belt, IS – Insular Belt. CRF – Columbia River Fault, SPF – Standard Peak Fault, CN – Carnes Nappe. Plutons (+) are intrusive into greenschist grade Lower Paleozoic rocks of the Hamill Group, Badshot Formation, and Lardeau Group (unpatterned). Map is simplified from Brown (1991), Wheeler (1963), and unpublished data (RLB). See text for further explanation.

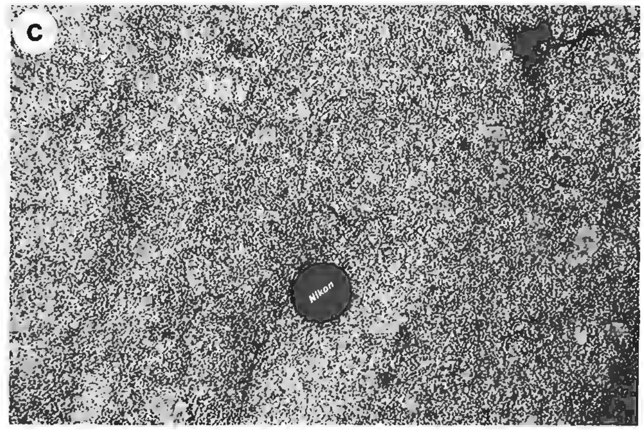


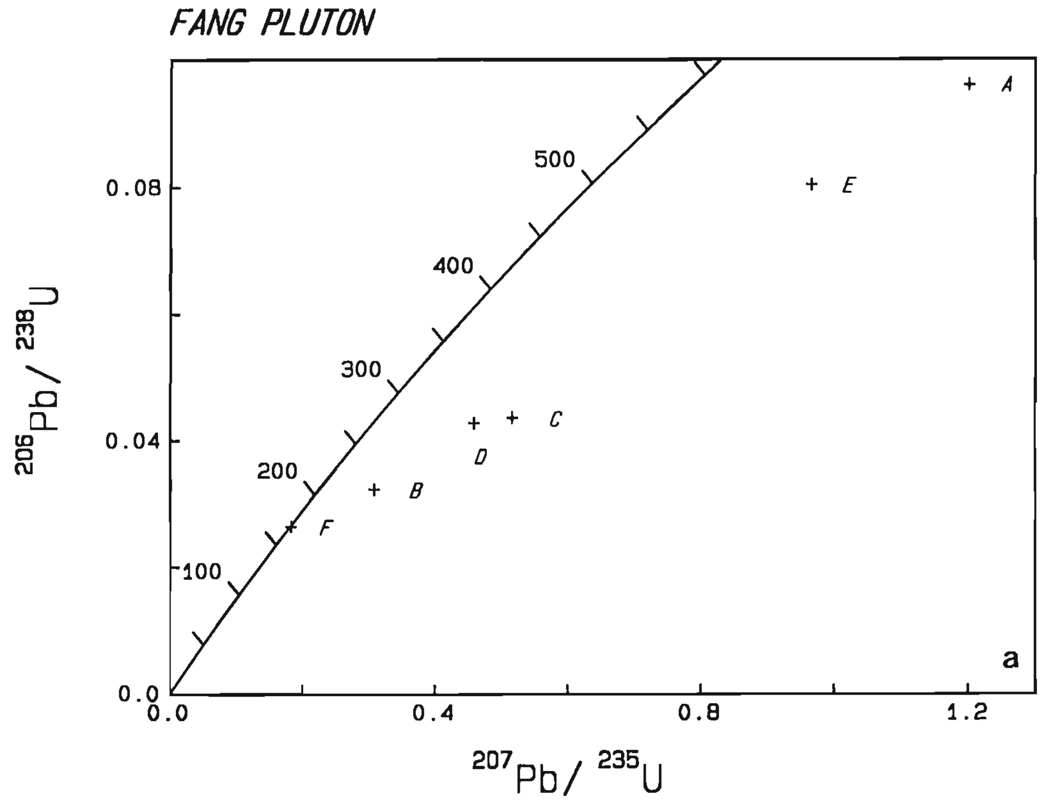
Figure 2. a. View looking northward from the Fang Pluton (light grey) near the snout of Dismal Glacier towards Carnes Nappe, a major westerly verging anticlinal fold nappe. GSC photo 1991-403-B. **b.** Photo of the intrusive contact between the Fang Pluton (light grey) and phyllitic rocks of the Lower Paleozoic Lardeau Group. Foliated xenoliths of phyllite occur within the pluton adjacent to the contact. The foliation within the metasedimentary rocks is oblique to and terminates at the intrusive contact. GSC photo 1991-403-D. **c.** Outcrop photo of the Pass Creek pluton showing the well preserved igneous textures of the K-feldspar megacrystic, biotite- and hornblende-bearing quartz monzonite. GSC photo 1991-399.

Table 1. U-Pb Analytical Data for Middle Jurassic plutons, northern Selkirk Mountains, British Columbia[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb/^{204}Pb$ *	Pbc,# (pg)	$^{208}Pb/^{206}Pb$ +	$^{206}Pb/^{238}U$ ++	$^{206}Pb/^{238}U$ ++ (Ma)	$^{207}Pb/^{235}U$ ++	$^{207}Pb/^{235}U$ ++ (Ma)	$^{207}Pb/^{206}Pb$ ++	corr. coef.	$^{207}Pb/^{206}Pb$ *** (Ma)
Fang Pluton, RB-21													
A,+149,cl	0.186	314.5	30.82	1016	356	0.095	0.09608 ± 0.11	591.4 ± 1.2	1.201 ± 0.14	801.2 ± 1.6	0.09069 ± 0.08	0.84	1440.2 ± 3.0
B,125,cl	0.099	338.7	10.82	1132	61	0.081	0.03232 ± 0.11	205.1 ± 0.4	0.3068 ± 0.15	271.7 ± 0.7	0.06884 ± 0.09	0.80	894.0 ± 3.8
C,90,cl	0.167	361.7	15.99	4068	41	0.092	0.4367 ± 0.09	275.5 ± 0.5	0.5157 ± 0.11	422.3 ± 0.7	0.08565 ± 0.04	0.95	1330.2 ± 1.3
D,68,cl	0.220	388.7	16.74	4098	57	0.091	0.4283 ± 0.09	270.4 ± 0.5	0.4577 ± 0.10	382.7 ± 0.7	0.07751 ± 0.04	0.94	1134.3 ± 1.4
E,125	0.032	320.2	26.90	4578	11	0.127	0.08042 ± 0.09	498.6 ± 0.8	0.9664 ± 0.10	686.6 ± 1.0	0.08716 ± 0.04	0.90	1364.0 ± 1.6
F,125	0.040	332.8	8.869	1791	12	0.123	0.02643 ± 0.10	168.2 ± 0.3	0.1814 ± 0.12	169.3 ± 0.4	0.04979 ± 0.07	0.83	185.3 ± 3.0
Pass Creek Pluton, R502													
A,105,cl	0.037	419.7	16.01	2657	14	0.109	0.03769 ± 0.09	238.5 ± 0.4	0.3556 ± 0.12	308.9 ± 0.7	0.06843 ± 0.07	0.82	881.6 ± 2.9
B,105,cl	0.040	309.6	39.38	7696	12	0.107	0.1205 ± 0.08	733.7 ± 1.1	1.983 ± 0.10	1109.9 ± 1.3	0.1193 ± 0.03	0.95	1946.3 ± 1.1
T-1,na	0.275	67.26	2.322	81	478	0.481	0.02616 ± 0.81	166.5 ± 2.6	0.1806 ± 2.6	168.6 ± 7.9	0.05008 ± 2.1	0.69	198.5 ± 100
T-2,na	0.215	64.06	2.231	70	443	0.471	0.02657 ± 0.98	169.1 ± 3.3	0.1838 ± 3.2	171.3 ± 10.1	0.05017 ± 2.6	0.69	202.8 ± 126

** T=titanite; numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;
Weighing error = 0.001 mg.
\+ Radiogenic Pb.
* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU
\# Total common Pb in analysis corrected for fractionation and spike.
++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).



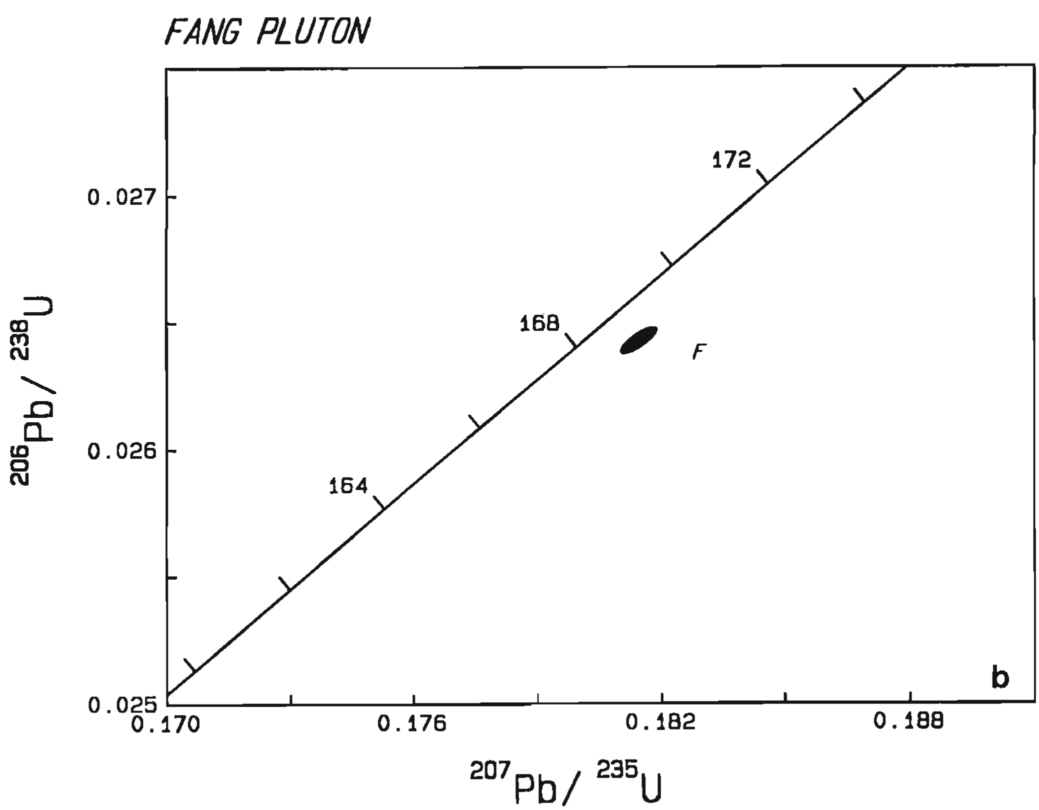
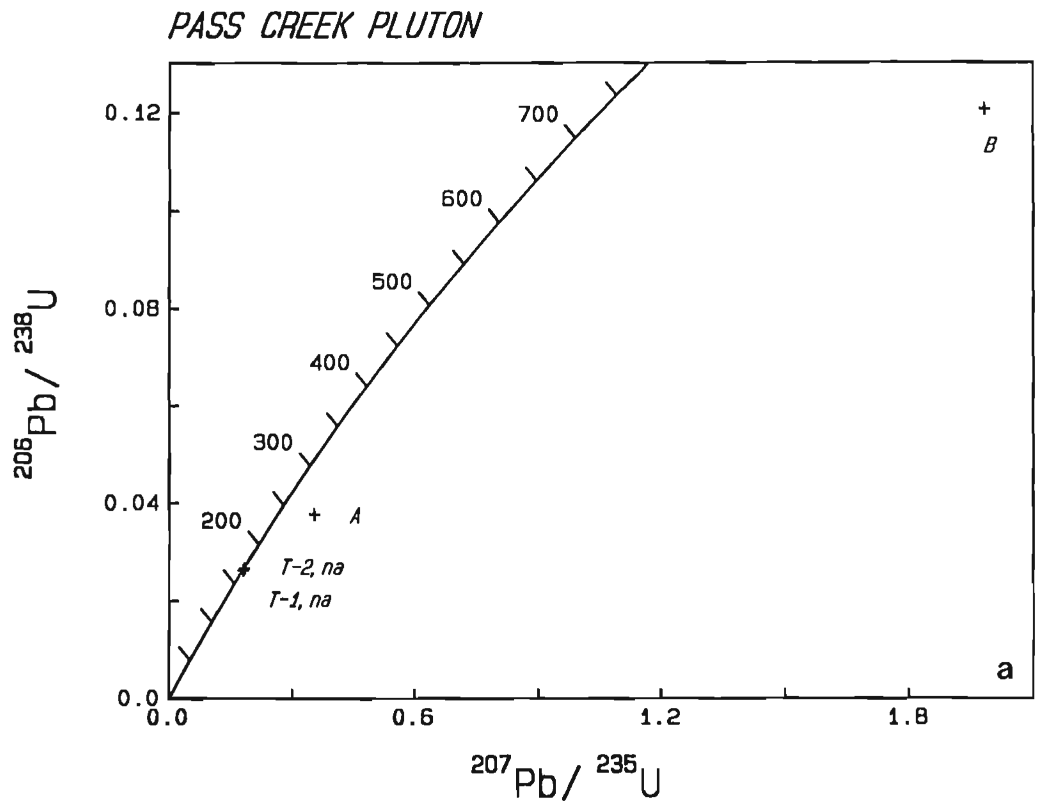


Figure 3. U-Pb concordia diagram of zircon data from the Fang Pluton. Error ellipses are 2σ.



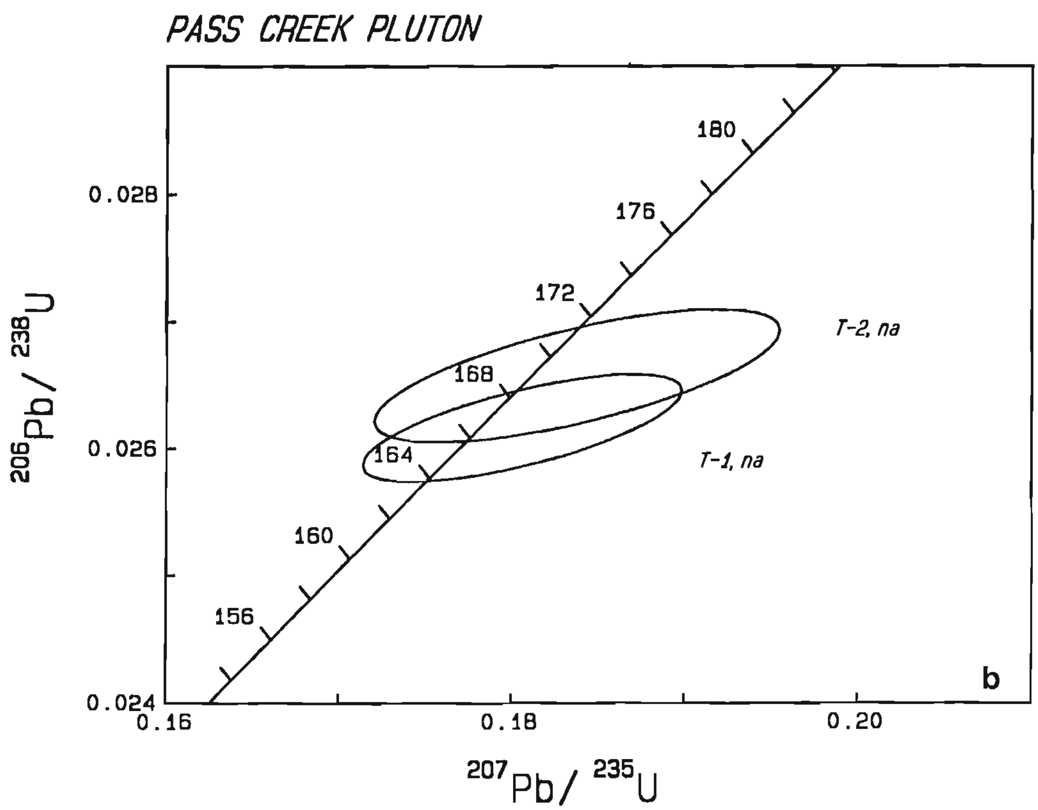


Figure 4. U-Pb concordia diagram of zircon and titanite data from the Pass Creek Pluton. Error ellipses are 2σ.

FIELD RELATIONSHIPS

Fang Pluton (RB-21)

At the sample site, the pluton is a coarse grained K-feldspar-megacrystic, biotite- and hornblende-bearing quartz monzonite with an igneous texture. The rock has no evident foliation; fractures are greater than 1 m apart and display no displacement. The pluton is continuously exposed above the northern floor of Dismal glacier (Fig. 1) and does not change in character along the distance of 3 km between the sample site (RB-21) and the eastern boundary at the snout of Dismal glacier other than to decrease its fracture spacing as the boundary is approached. At the well exposed contact the plutonic rock is clearly intrusive into phyllitic rocks of the Lower Paleozoic Lardeau Group. Dykes of the pluton extend into the metasedimentary rocks without disruption, foliation within the metasedimentary rocks is oblique to and terminates at the intrusive contact, and foliated xenoliths of phyllite occur within the pluton adjacent to the contact (Fig. 2b). The penetrative fabrics within the adjacent phyllites of the Lardeau Group all appear to pre-date the emplacement of the Fang pluton.

Regional trends of axial surfaces to the north and south of tha pluton are generally northwesterly but east-west trends occur locally and are particularly well developed over an area that extends from the northern boundary of the pluton northward for 5 km along the ridges north of Downie Creek (Fig. 1). This change in trend may predate emplacement of the pluton but is more likely a result of post-emplacement strain and associated rigid body rotation of the pluton.

Pass Creek Pluton (R502)

This body is petrologically and structurally very similar to the Fang pluton. It is a K-feldspar megacrystic, biotite- and hornblende-bearing quartz monzonite (Fig. 2c). No penetrative foliation is developed and igneous textures are well preserved. Locally the rock is fractured and altered with chlorite commonly found adjacent to the fracture surfaces. Figure 1 illustrates the high discordance of the pluton boundary with stratigraphic and structural trends within the adjacent metasedimentary rocks. Stratigraphic and structural features within the metasedimentary rocks are truncated at the contact.

The region has experienced complex strain which includes southwesterly verging isoclinal folding (F1 and F2) and thrust faulting. One of the thrust faults is cut by the pluton at its southwestern boundary and major F2 closures are also truncated. The plutonic rock adjacent to these structural discordances is unstrained and retains its igneous texture. Adjacent to the southeastern boundary of the pluton there is a major change in regional trend of the metasedimentary rocks from northwesterly to northeasterly. This change in trend affects stratigraphic contacts and the trend of normal faults that parallel these boundaries. These deflections also occur at a larger distance from the pluton (see Brown, 1991) and are interpreted to be features that could have developed after its emplacement, in a similar manner to the regional structural trends around the Fang Pluton.

U-Pb GEOCHRONOLOGY

U-Pb analytical methods utilized are those outlined in Parrish et al. (1987). All of the zircon fractions analyzed were air abraded with pyrite (Krogh, 1982). Isotopic data and a brief outline of analytical procedures are summarized in Table 1.

Fang Pluton (RB-21)

Six fractions of clear, euhedral, prismatic zircons with minor inclusions were analyzed from the Fang Pluton. All of the zircon analyses reveal evidence of inheritance (Fig. 3a). Fraction F is very slightly discordant, plotting just below concordia at about 168 Ma (Fig. 3b). The ^{206}Pb-^{238}U age is 168.2 ± 0.3 Ma, which is interpreted to be the age of this zircon fraction. Therefore an age of 168 ± 2 Ma (a more reasonable estimation of uncertainty) is interpreted to be the age of crystallization of the pluton.

Pass Creek Pluton (R502)

Zircons from the Pass Creek pluton are very similar in appearance to those from the Fang – clear, euhedral prismatic crystals with minor inclusions. The grains are crack-free and visible cores are not apparent. Analysis of two fractions of zircon reveals the presence of an inherited component as least as old as 1950 Ma (B; Fig. 4a). Two fractions of clear, light yellowish brown, flat disc-shaped titanite were also analyzed from the sample. The two fractions intersect and overlap concordia at 168 ± 3 Ma (Fig. 4a,b). This age is interpreted to be a crystallization age of the post-tectonic, undeformed Pass Creek Pluton which was intruded following the peak of greenschist facies metamorphism within the host metasedimentary rocks (see below).

DISCUSSION

The U-Pb data establish a Middle Jurassic age for crystallization of the Fang and Pass Creek plutons. The field relationships clearly require development of penetrative strain (F1 and F2) and associated metamorphism prior to emplacement of these plutons. Phase one deformation includes formation of the Carnes Nappe and westerly directed thrust faulting (Brown and Lane, 1988; Brown, 1991). F2 folding gave rise to map scale westerly verging folds that are superimposed on the right-way-up and overturned limbs of the Carnes Nappe. Peak metamorphism occurred during F2 deformation (Brown and Tippett, 1978; Lane, 1977). Within the area under consideration this peak did not exceed upper greenschist grade.

A lower age limit for the deformation and metamorphism cannot be established from the above relationships, but it is clear that post Middle Jurassic strain in this part of the Kootenay terrane is restricted to post-peak metamorphic deformation that caused local folding (F3, see Brown and Tippett, 1978; and Brown, 1991) and possible rigid body rotation of the plutons.

ACKNOWLEDGMENTS

Field work was supported by Geological Survey of Canada Research Agreements and NSERC Operating Grant (A2693) to R.L. Brown. We thank Lois Hardy for assistance in the preparation of Figure 1. The staff of the geochronology laboratory are thanked for their assistance in generating the U-Pb data. M. Villeneuve reviewed the manuscript.

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APPENDIX 1
Sample localities for U-Pb dating

Fang Pluton (RB21): K-feldspar megacrystic quartz monzonite; collected from a rock point on the east side of Dismal Glacier (locality is now the site of a climbing hut) at an elevation of about 7600 ft., Glacier map sheet (82N/5), southeastern British Columbia; 117°54'29" W - 51°18'40" N; UTM 10u, 436700E-5684600N.

Pass Creek Pluton (R502): K-feldspar megacrystic quartz monzonite; collected near a small round lake east of the headwaters of Pass Creek at an elevation of 6700 ft., Downie Creek map sheet (82M/8), southeastern British Columbia; 118°09'37" W - 51°22'39" N; UTM 10u, 419250E-5692230N.

- 15 -

Miscellaneous U-Pb zircon dates from southeast British Columbia

Randall R. Parrish[1]

Parrish, R.R., 1992: Miscellaneous U-Pb zircon dates from southeast British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 143-153.

Abstract

U-Pb zircon dates are reported for six granitic rocks of the Omineca Belt of southeastern British Columbia. Three range in age from 56-70 Ma and are in part correlative with the Ladybird granite suite, which is widespread in southeastern British Columbia; one is the Shaw Creek intrusion west of Creston, about 76 Ma old; one is from the Clachnacudainn gneiss east of Revelstoke, which is ca. 358 Ma old; and the last is a gneiss in the Seymour Range northwest of Revelstoke which is ca. 359 Ma old. The two Devonian orthogneisses are part of a major magmatic association along the western side of the Omineca Belt of the eastern Cordillera.

Résumé

Les datations U-Pb de six roches granitiques de la zone d'Omenica dans le sud-est de la Colombie-Britannique sont présentées. Trois ont un intervalle d'âge variant entre 56 et 70 Ma et sont en partie corrélées à la suite granitique de Ladybird qui est largement présente dans le sud-est de la Colombie-Britannique; une correspond à l'intrusion de Shaw Creek à l'ouest de Creston et âgée d'environ 75 Ma; une autre correspond au gneiss de Clachnacudainn à l'est de Revelstoke qui date d'environ 358 Ma; et la dernière correspond à un gneiss reposant dans la chaîne Seymour au nord-ouest de Revelstoke et datant d'environ 359 Ma. Les deux orthogneiss dévoniens font partie d'une importante association magmatique longeant la partie occidentale de la zone d'Omineca dans l'est de la Cordillère.

INTRODUCTION

During regional geological investigations of southeastern British Columbia, the author, in conjunction with colleagues, dated numerous samples of granitic rock to assist in the compilation of a regional tectonic map which has been published in Parrish et al. (1988). Some of these samples were biotite leucogranites suspected of belonging to the previously defined Ladybird granite suite (Carr et al., 1987); they were dated to help define the extent of Ladybird magmatism. Another was the Clachnacudainn gneiss near Revelstoke whose age was thought to be Ordovician, based on the work of Okulitch (1985).

This paper presents the descriptions and analytical data from these and two other samples and comments on their tectonic significance. The samples are plotted on a simplified tectonic map, Figure 1. U-Pb analytical procedures are those outlined in Parrish et al. (1987). Techniques included air abrasion for most, but not all, of the zircon fractions (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987) and estimation of errors by numerical error propogation (Roddick, 1987). Monazite analyses are interpreted using criteria described in Parrish (1990). Details of sample locations are given in Appendix 1.

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LADYBIRD GRANITE SUITE AND POTENTIALLY CORRELATIVE ROCKS

The Ladybird granite suite is named for extensive exposures of leucogranite in Valhalla complex near Castlegar (Carr et al., 1987). This suite was termed the Valhalla granite suite by Little (1960) who mapped these granitic rocks by their distinctive smoky quartz, mica phenocrysts, abundance of pegmatite, lack of hornblende, and occasional presence of garnet and metasedimentary inclusions. They constitute a major suite of leucocratic granitic rocks which may have been generated in the crust. Much U-Pb analytical data from these

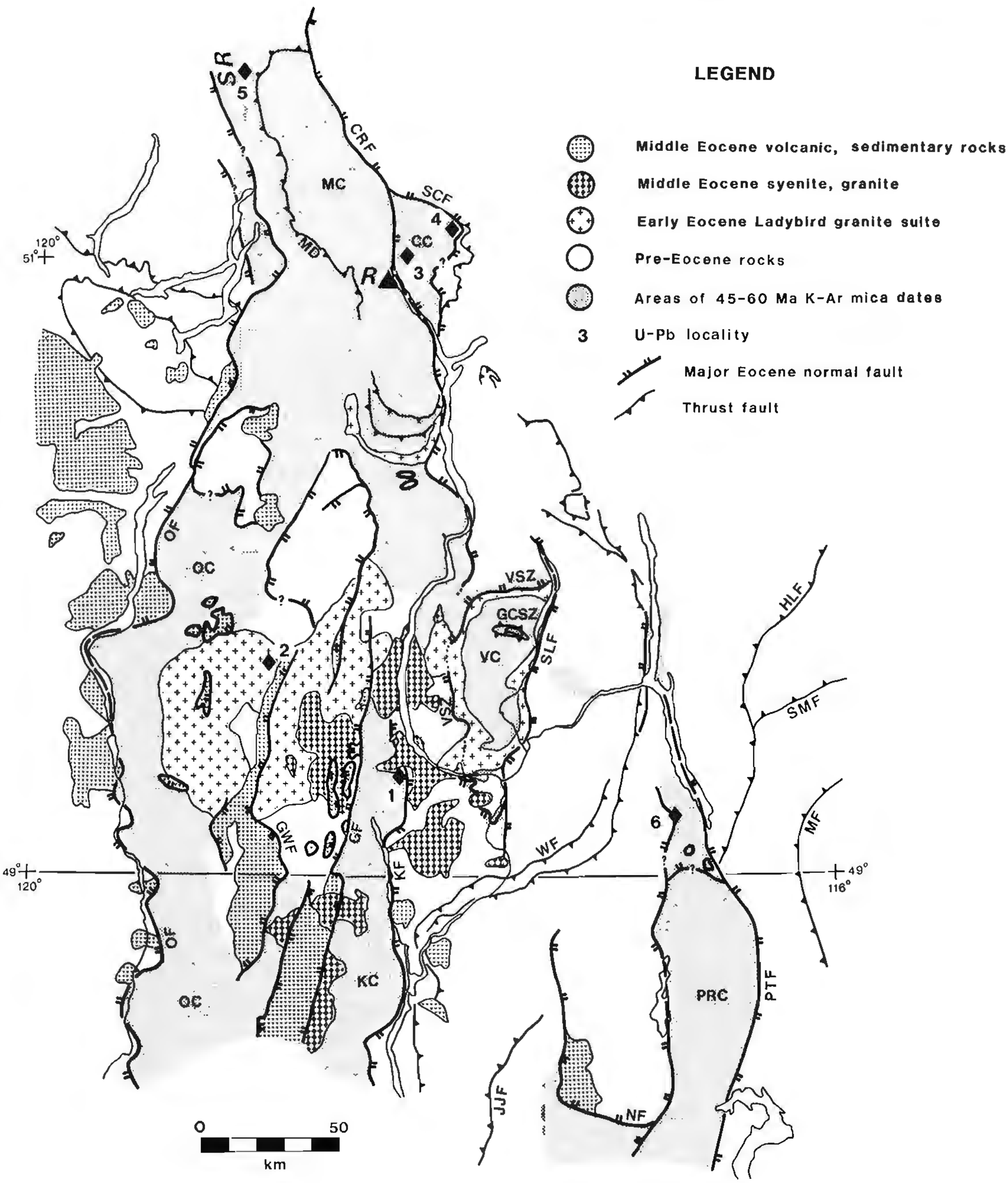

Figure 1. Simplified tectonic map of southern British Columbia showing major geological features and sample localities referred to in the text, modified from Parrish et al. (1988). Abbreviations are as follows: CC – Clachnacudainn complex; CRF – Columbia River fault; GCSZ – Gwillim Creek shear zone; GF – Granby fault; GWF – Greenwood fault system; HLF – Hall Lake fault; JJF – Jumpoff Joe fault; KC – Kettle-Grand Forks complex; KF – Kettle fault; MC – Monashee complex; MD – Monashee decollement; MF – Moyie fault; NF – Newport fault; OC – Okanagan complex; OF – Okanagan Valley fault; PRC – Priest River complex; PTF – Purcell Trench fault; SCF – Standfast Creek fault; SLF – Slocan Lake fault; SMF – St. Mary fault; VC – Valhalla complex; VSZ – Valkyr shear zone; WF – Waneta fault; R – town of Revelstoke; SR – Seymour Range.

rocks in Valhalla Range have already been published in Parrish et al. (1988), but three additional samples were collected, in part from areas which Little (1957, 1960, 1961) mapped as Valhalla granite. Carr et al. (1987) recognized the distinctive aspects of this suite in the Valhalla Range and renamed it the Ladybird granite after exposures in Ladybird Creek. The question addressed for these three samples was to confirm or refute the correlation with Early Eocene granites of Valhalla Range.

(1) Granite from near Renata (PCA-141-86)

This sample is biotite leucogranite typical of the Ladybird suite. As mapped by Little (1957) and Parrish et al. (1988), this granite occurs on the northeast extension of the Kettle-Grand Forks complex, and is probably faulted against hornblende granitic rocks to the east. Both granitic rocks are intruded by Coryell syenite.

U-Pb data

Three fractions of zircon show modest amounts of inheritance and define a wedge-shaped array of points, with the most concordant being adjacent to 57 Ma, but still slightly discordant (Fig. 2). In a manner similar to other Ladybird granite U-Pb data, this pattern is interpreted as a mixture of igneous grains ca. 56 Ma old and Precambrian inherited zircons, mostly as cores within otherwise magmatic grains. The proximity of point C to concordia gives reasonable confidence that the age of crystallization of this granite is 56 ± 1 Ma.

(2) Granite from upper Kettle River (PCA-155-86)

Little (1957) mapped granite in fault contact with Eocene sedimentary rocks in the upper Kettle River valley. These granitic rocks closely resemble Ladybird granite and a sample was collected to determine if it also was Eocene in age. Sample PCA-155-86 is a biotite leucogranite with a modest foliation.

U-Pb data

Three zircon fractions from the sample are equant to elongate and prismatic euhedral in habit. There were faintly visible cores in many grains, in addition to small inclusions. These zircons were not abraded. The analyses lie on a linear array within analytical uncertainty; the lower and upper intercepts with concordia are 48 ± 5 Ma and 157 ± 3 Ma, respectively (Fig. 3). This distinctive pattern strongly implies zircon inheritance of Middle to Late Jurassic age in an Eocene magma. It is difficult to be completely confident that the values and uncertainties of the concordia intercepts are accurate because of potential problems of Pb loss and/or variable age of zircon inheritance. Both of these are probable; this general age is in agreement with other ages of the Ladybird suite to the east (Parrish et al., 1988).

The Jurassic inheritance contrasts with almost all other dated Ladybird granites which have mainly Precambrian inheritance. This is probably because this sample locality is one of the farthest west and its host rocks include Paleozoic eugeoclinal rocks and Jurassic intrusions.

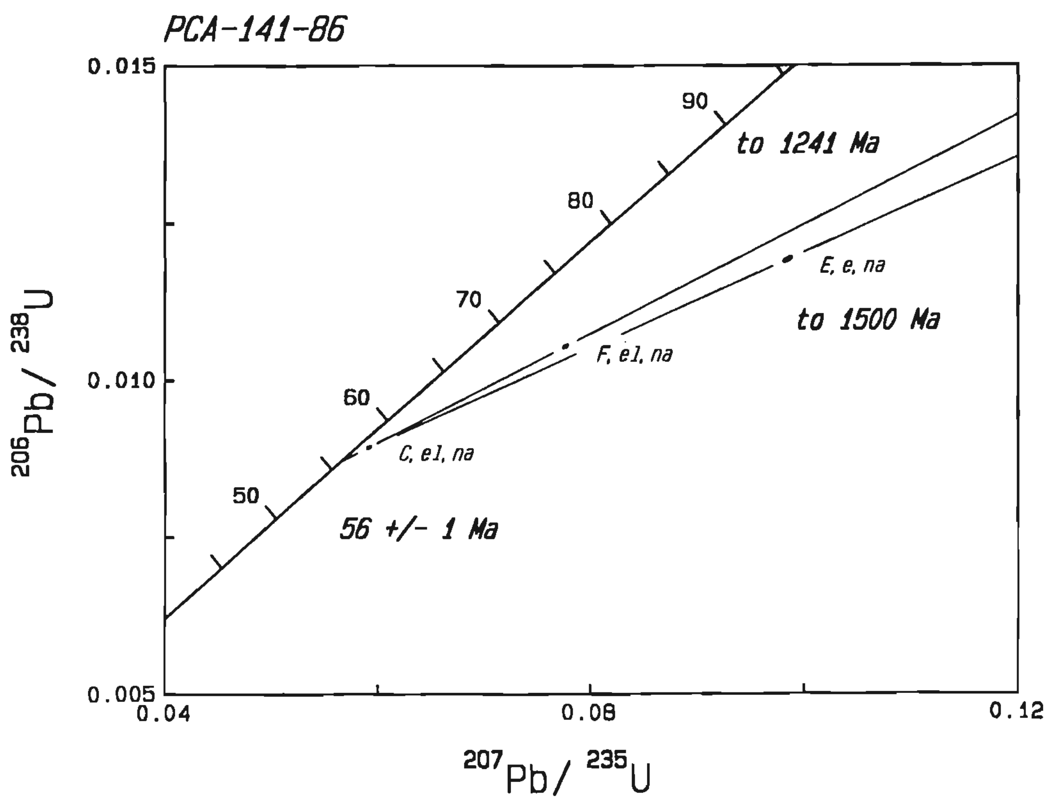

Figure 2. Concordia diagram for U-Pb data from sample PCA-141-86. Error ellipses reflect 2σ uncertainty.

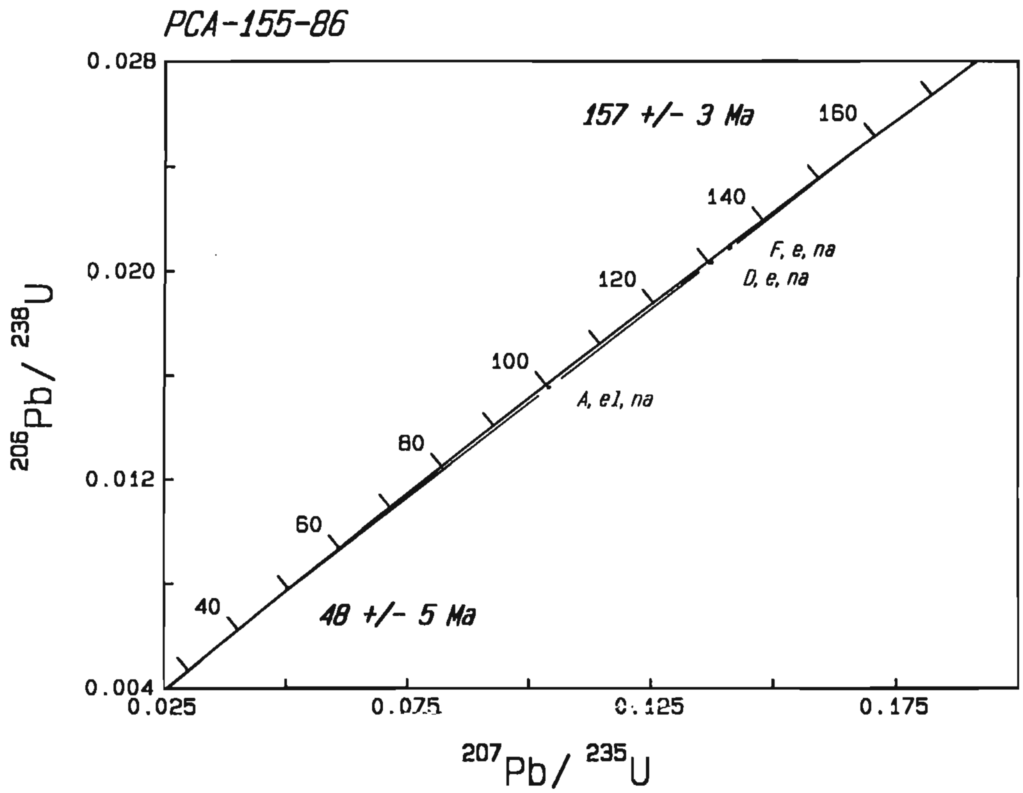

Figure 3. Concordia diagram for U-Pb data from sample PCA-155-86. Error ellipses reflect 2σ uncertainty.

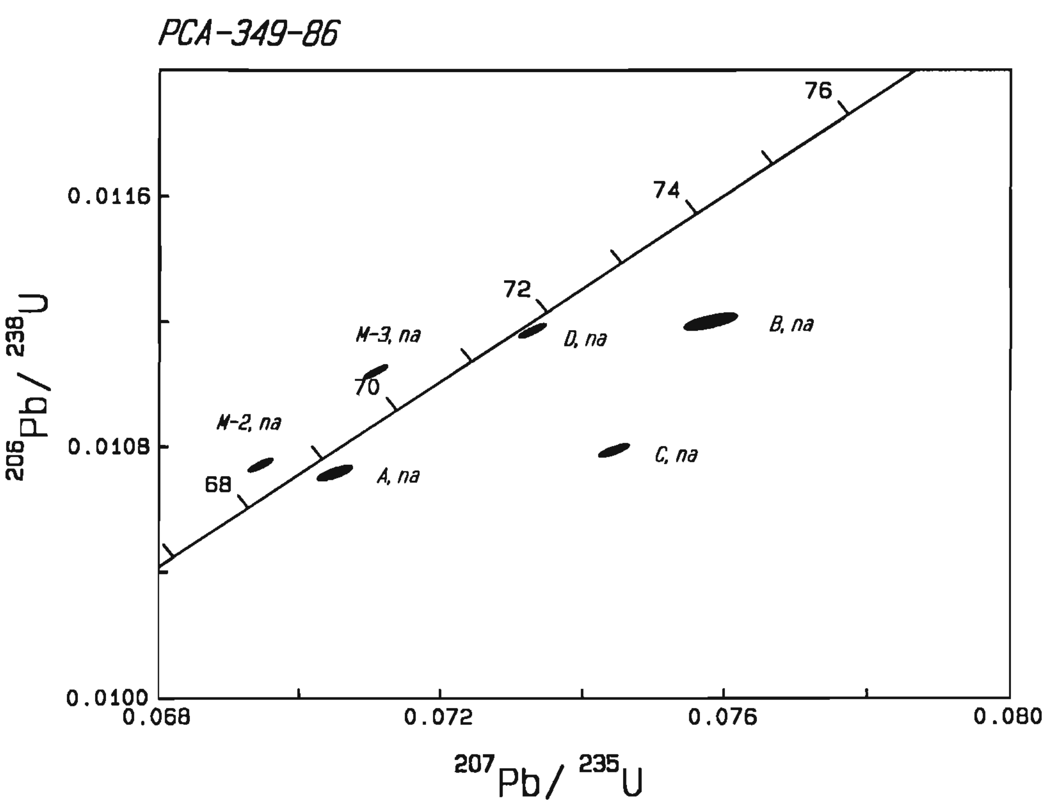

Figure 4. Concordia diagram for U-Pb data from sample PCA-349-86. Error ellipses reflect 2σ uncertainty.

(3) Granite from Mt. Revelstoke (PCA-349-86)

East of Revelstoke lies a group of meta-igneous and high grade metamorphic rocks known collectively as the Clachnacudainn salient or slice. It is sandwiched between the Selkirk allochthon above and the Monashee complex below by the Standfast Creek fault and Columbia River faults, respectively. Both of these faults have been interpreted as normal faults by Parrish et al. (1988), but the Columbia River fault in part coincides with an earlier thrust fault, the Monashee décollement (Read and Brown, 1981). The rocks of the Clachnacudainn slice contain a variety of granitic rocks including Devonian (see below), Cretaceous (R.L. Armstrong, unpublished data 1989), and latest Cretaceous-Paleocene (this sample), and may also contain intrusions of even younger age.

On the west side of the Monashee complex in the Seymour Range, a similar suite of intrusive rocks exists which includes Devonian rocks (see below, and Okulitch, 1985), the Cretaceous Anstey pluton, and younger leucogranites of ca. 63 Ma and older age (R. Parrish, unpublished data; Scammell, 1991). Prior to the determination of these ages, it was suggested that the rocks in the Seymour Range and in Clachnacudainn slice were the same structural and lithological assemblage which was transported eastwards by the Monashee décollement and then separated from each other by arching and normal faulting in Eocene time (Parrish et al., 1988). It was also suggested that ductile deformation within, and particularly near the top of the slice was of, in part, Tertiary age.

There is abundant leucogranite exposed on the paved road up Mt. Revelstoke in Mt. Revelstoke National Park. A small sample of foliated leucogranite was sampled in 1986, and an age of this sample would place constraints on the latter part of ductile deformation, as well as elucidating the magmatic history.

U-Pb data

Sample PCA-349-86 is a foliated biotite leucogranite which intrudes other metamorphic and meta-igneous rocks. It has a weak to moderate developed foliation, but is not mylonitic. The origin of this foliation was not investigated in detail. The granite contains zircon and monazite; zircons were moderately clear euhedral crystals or fragments of larger grains, and in fractions A, B, and C, there were visible areas which were suspected of being inherited cores. Fraction D was composed of 10 larger clear grains without obvious core material. Monazites were relatively cloudy, and the two fractions consisted of two grains each.

Fractions A, B, and C are slightly discordant and scattered whereas fraction D is essentially concordant (Fig. 4), particularly if a correction for deficiency of ^{230}Th-derived ^{206}Pb is made (not corrected in Fig. 4; see Coleman and Parrish, in press). Fraction D has an interpreted age of 72 Ma. Monazite is reversely discordant and has an interpreted crystallization age of at least 70 Ma. The age of the inherited component in the zircons is difficult to assess but could include Mesozoic or Paleozoic zircons, in addition to

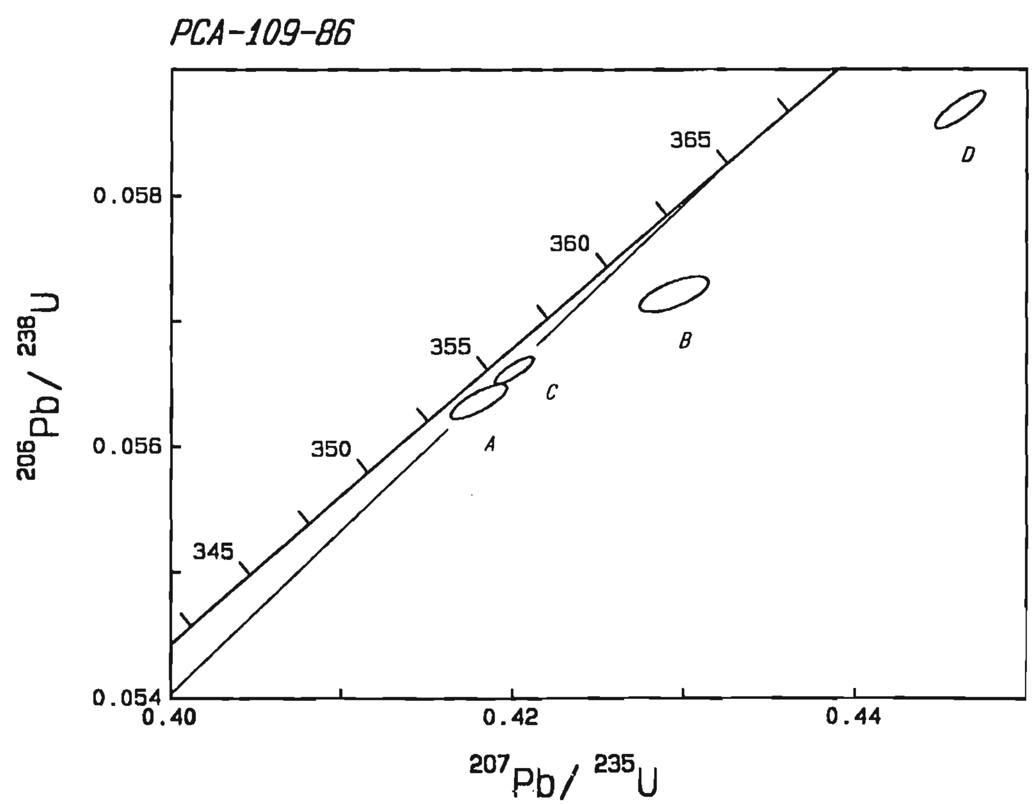

Figure 5. Concordia diagram for U-Pb data from sample PCA-109-86. Error ellipses reflect 2σ uncertainty.

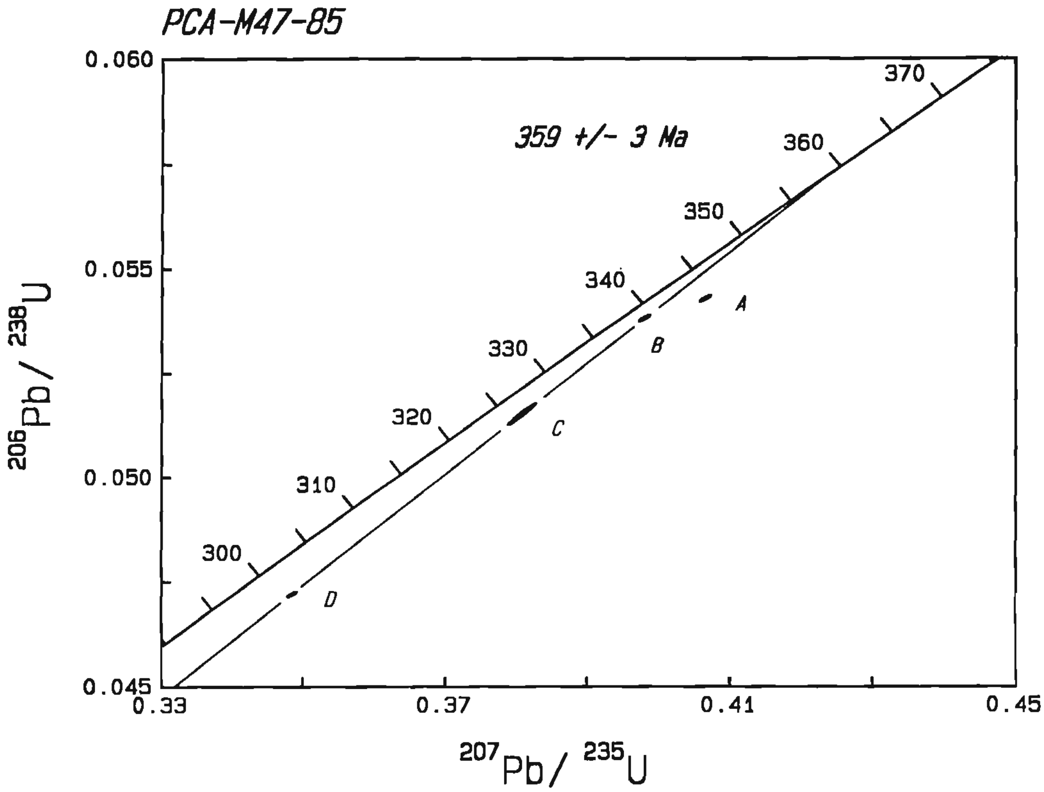

Figure 6. Concordia diagram for zircon U-Pb data from sample PCA-M47-85. Error ellipses reflect 2σ uncertainty.

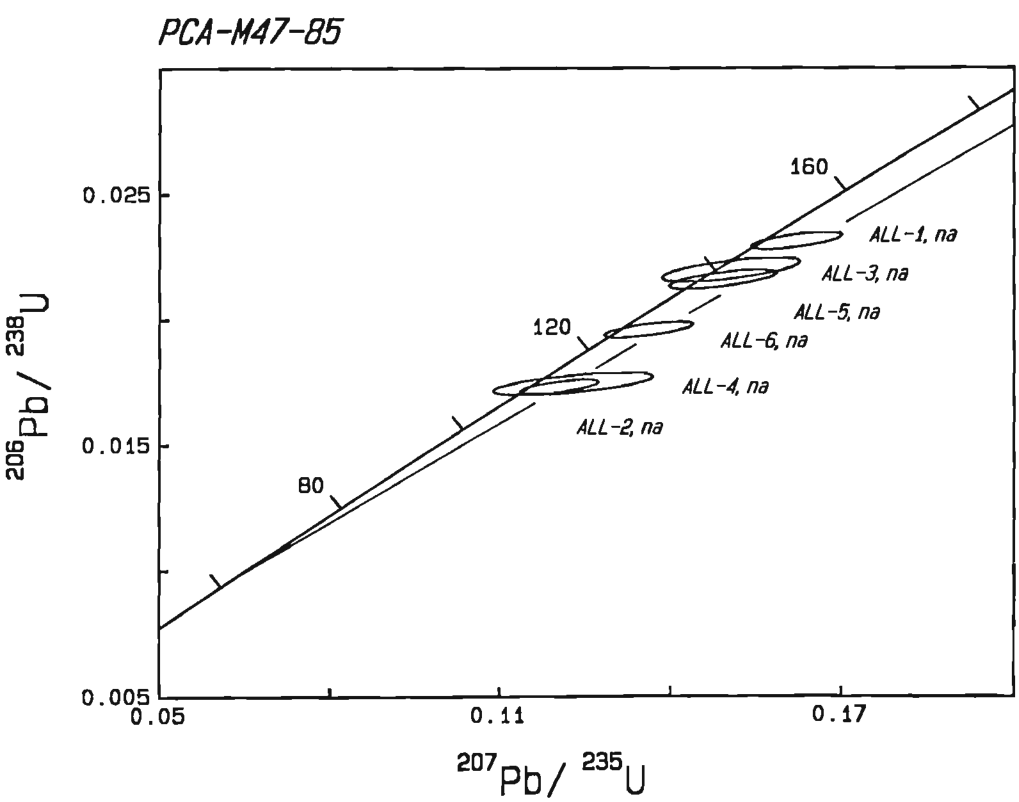

Figure 7. Concordia diagram for allanite U-Pb data from sample PCA-M47-85. A reference line from 63 to 359 Ma is included. Error ellipses reflect 2σ uncertainty.

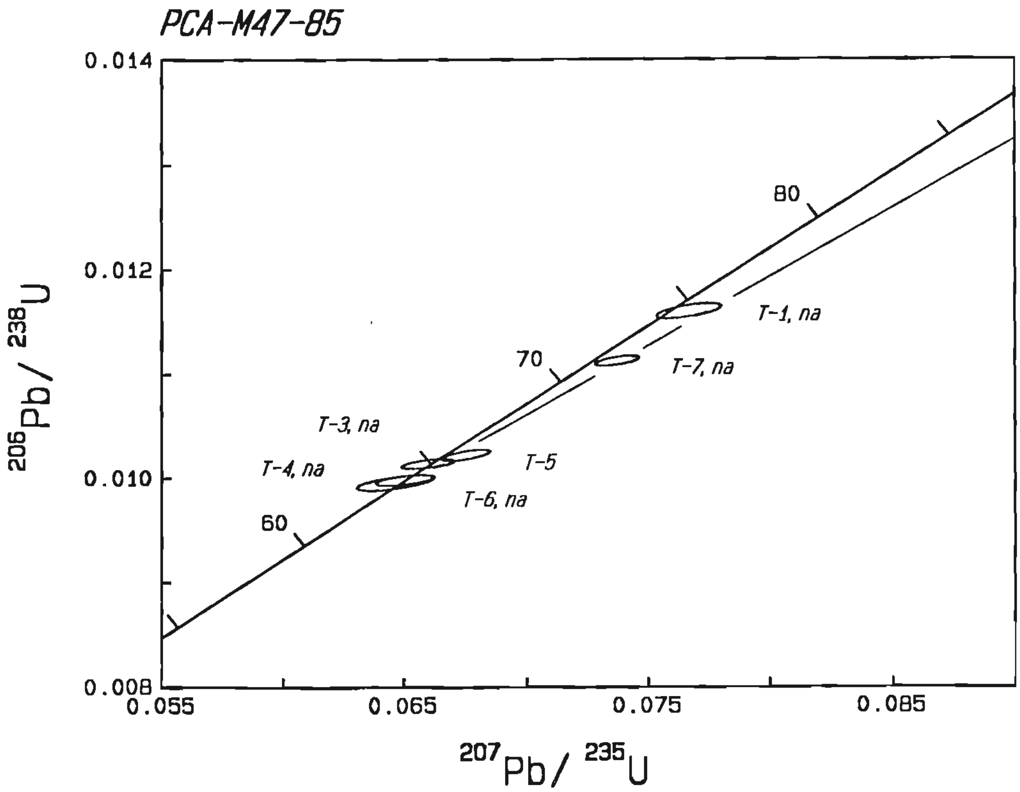

Figure 8. Concordia diagram for titanite U-Pb data from sample PCA-M47-85. A reference line from 63 to 359 Ma is included. Error ellipses reflect 2σ uncertainty.

Precambrian. The general consistency of zircon and monazite ages suggests that an age of 71 ± 1 Ma is a reasonable estimate for the crystallization of this body.

Scammell (1991) reported an age of 70-71 Ma for weakly foliated leucogranite in the northern Monashee Mountains. These rocks lie in the same structural position as the rocks of the Seymour Range from which a leucogranite age of 63 Ma has been determined (R. Parrish, unpublished data). This similarity in age strengthens the correlation between the rocks of the hangingwall of the Monashee décollement north and west of the Monashee complex and those of the Clachnacudainn slice. This correlation is further supported by ages for Paleozoic rocks presented below.

PALEOZOIC GNEISS EAST AND NORTHWEST OF REVELSTOKE

Paleozoic orthogneiss has been described by Okulitch (1985) in the vicinity of Shuswap Lake near Sicamous and in the Clachnacudainn salient east of Revelstoke. The U-Pb data from that study were imprecise and it is desirable to know more clearly the age of these rocks.

Two rock samples were selected for this study, one being the Clachnacudainn gneiss, a sheet of orthogneiss along the eastern (upper) part of the slice; the other sample was a gneissic granodiorite in the Seymour Range collected by R.L. Brown and J.M. Journeay during geological mapping.

(4) Clachnacudainn gneiss (PCA-109-86)

This sample was collected from the strongly deformed sheet of orthogneiss well exposed in a quarry on the south side of the Illecilliwaet River near Albert Canyon. The gneiss has straight foliation and is an annealed mylonite. It contains biotite, hornblende, quartz and feldspar with an annealed recrystallized fabric, and has zircon, titanite, and apatite as accessory minerals.

U-Pb data

Four fractions of well abraded, euhedral zircons with some inclusions were analyzed from this sample (Fig. 5). Fractions A and C plot in a quasi-linear array below concordia and have ^{207}Pb-^{206}Pb ages of about 364 Ma, which may represent the age of the zircons that have undergone some Pb loss. The analyses of fractions B and D reveal evidence of older inheritance combined with younger components, perhaps as young as 352 Ma. The best estimate for the age of the Clachnacudainn gneiss is 358 ± 6 Ma.

(5) Seymour Range orthogneiss (PCA-M47-85)

The Seymour Range sample is a foliated hornblende-biotite granodiorite engulfed in leucogranite veins and dykes. The foliation is cut by a stockwork of leucogranite dykes which are 63 Ma old (unpublished data of R. Parrish). Within the gneiss, titanite and allanite U-Pb systematics show substantial resetting and have been described by Heaman and Parrish (1991) as a good example of partial resetting of these minerals. The U-Pb data are shown in Figures 6, 7, and 8 and were previously published in Heaman and Parrish (1991); however, the analytical data were not described there. The morphology of zircon, titanite, and allanite is subhedral to euhedral, and there is every reason to think that the minerals dated are relict igneous grains, not newly grown during metamorphism. Zircons were strongly abraded and four fractions were analyzed. One of these plots to the right of a linear array connecting the other fractions, and is suspected of having a small amount of older inheritance (Fig. 6). The other three fractions form a good fit and their upper intercept with concordia is 359 ± 3 Ma, which is taken as the crystallization age of the granodiorite. Six allanite and seven titanite fractions were also analyzed. Allanites have large uncertainties due to significant corrections for common Pb, and they lie in an array parallel to concordia between 110 and 150 Ma (Fig. 7), whereas titanites are concordant to slightly discordant in a shorter array between 63 and 92 Ma (Fig. 8). The separation of the allanite and titanite data was used by Heaman and Parrish (1991) to conclude that allanite was significantly more resistant to Pb loss than titanite, and zircon considerably more so than allanite.

The pattern of titanite and allanite data implies that a heating event caused partial to complete Pb loss in allanite or titanite at 63 Ma. The leucogranite, which intrudes this locality of orthogneiss, has been dated at 63 Ma using monazite and zircon (R. Parrish, unpublished data), but these data are not presented here. Hornblende and biotite dates from the orthogneiss are 55 ± 2 and 55 ± 3 Ma, respectively (GSC K-Ar 89-67, 68; Hunt and Roddick, 1990). Obviously a substantial thermal event is implied which probably was caused by the intrusion of leucogranite.

Discussion

These two zircons ages indicate that an important igneous episode about 360 Ma old exists in rocks within the hangingwall of Monashee décollement near the latitude of Revelstoke. The data herein supercede analyses from the Clachnacudainn gneiss presented in Okulitch (1985) which were interpreted to indicate an Ordovician age for this gneiss. Other Devonian ages come from the Mt. Fowler gneiss near Shuswap Lake (Okulitch et al., 1975), felsic tuff from part of the Eagle Bay Group (387 ± 4 Ma, R.L. Armstrong, unpublished data cited in Okulitch, 1985), and ca. 350 Ma ages of the Quesnel Lake and related gneisses to the northwest (Okulitch, 1985; Mortensen et al., 1987). The samples described here were, with their host rock, transported large

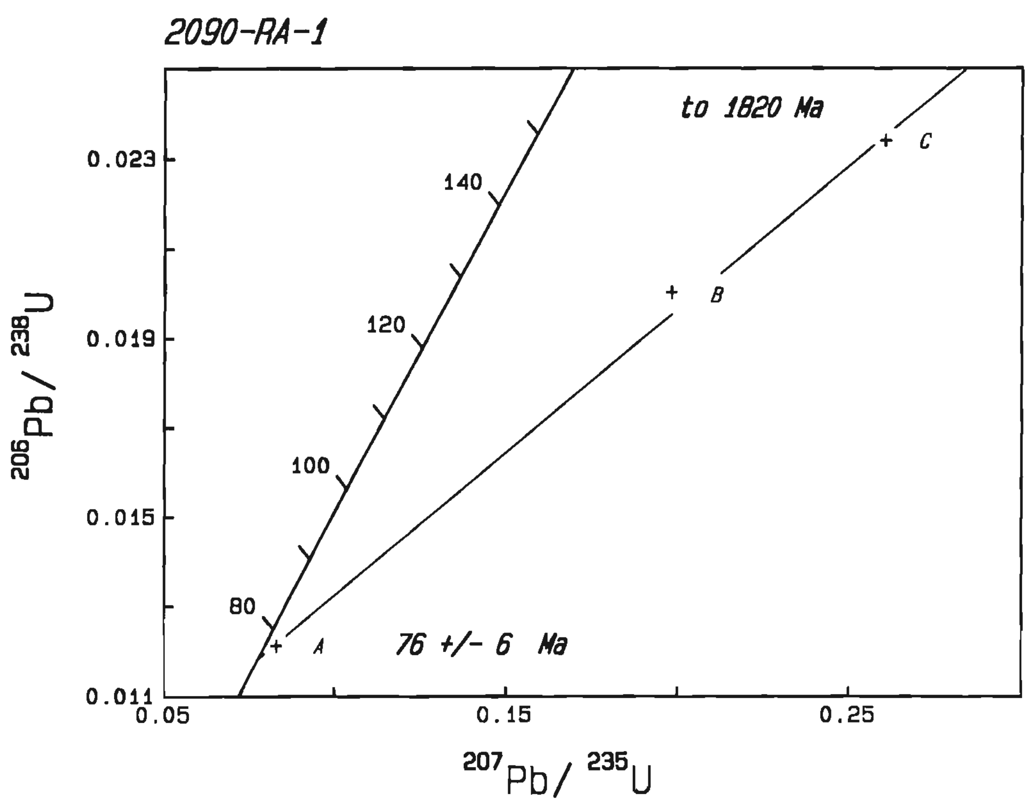

Figure 9. Concordia diagram for U-Pb data from sample 2090-RA-1. Error ellipses reflect 2σ uncertainty.

Table 1. U-Pb Analytical Data for granites from southeast British Columbia[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^{*}/^{204}Pb$	Pbc,# (pg)	$^{208}Pb^{+}/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb/^{206}Pb$*** (Ma)
(1) PCA-141-86, Ladybird Granite, Renata area													
C,el,na,125	0.095	728.3	6.146	1659	24	0.052	0.008924 ± 0.12	57.3 ± 0.1	0.05917 ± 0.16	58.4 ± 0.2	0.04809 ± 0.12	0.69	103.4 ± 5.4
E,e,na,90	0.069	522.9	6.214	919	30	0.100	0.01192 ± 0.19	76.4 ± 0.3	0.09846 ± 0.20	95.3 ± 0.4	0.05993 ± 0.14	0.76	601.0 ± 6.0
F,el,na,68	0.114	817.9	8.428	2162	29	0.084	0.01053 ± 0.14	67.6 ± 0.2	0.07766 ± 0.15	75.9 ± 0.2	0.05347 ± 0.10	0.78	348.6 ± 4.4
(2) PCA-155-86, Ladybird Granite, upper Kettle River													
A,el,na,90	0.115	659.2	9.893	5884	13	0.075	0.01553 ± 0.10	99.3 ± 0.2	0.1041 ± 0.11	100.6 ± 0.2	0.04863 ± 0.04	0.93	129.8 ± 1.9
D,e,na,125	0.075	668.1	12.83	4142	16	0.050	0.02032 ± 0.09	129.7 ± 0.2	0.1373 ± 0.12	130.6 ± 0.3	0.04899 ± 0.05	0.89	147.3 ± 2.4
F,e,na	0.109	511.8	10.52	6548	11	0.095	0.02089 ± 0.09	133.3 ± 0.2	0.1412 ± 0.10	134.1 ± 0.3	0.04901 ± 0.04	0.93	148.4 ± 1.8
(3) PCA-349-86, Deformed leucogranite, Mt. Revelstoke													
A,na,el,-62	0.065	558.7	6.106	560	47	0.085	0.01120 ± 0.11	71.8 ± 0.2	0.07580 ± 0.25	74.2 ± 0.4	0.04908 ± 0.18	0.72	151.5 ± 8.5
B,na,el,68	0.082	577.3	6.044	905	36	0.088	0.01072 ± 0.11	68.7 ± 0.2	0.07050 ± 0.18	69.2 ± 0.2	0.04771 ± 0.11	0.78	84.8 ± 5.4
C,na,el,90	0.103	528.4	5.513	1076	35	0.075	0.01079 ± 0.10	69.2 ± 0.1	0.07444 ± 0.14	72.9 ± 0.2	0.05003 ± 0.09	0.81	196.3 ± 4.0
D,na,el	0.121	423.7	4.813	1519	24	0.118	0.01117 ± 0.10	71.6 ± 0.1	0.07328 ± 0.13	71.8 ± 0.2	0.04756 ± 0.06	0.88	77.5 ± 2.9
M-2,na	0.058	1931	141.0	1906	40	6.802	0.01074 ± 0.10	68.9 ± 0.1	0.06943 ± 0.12	68.2 ± 0.2	0.04688 ± 0.06	0.87	43.2 ± 2.9
M-3,na	0.059	2226	167.4	2525	37	6.815	0.01104 ± 0.10	70.8 ± 0.1	0.07107 ± 0.12	69.7 ± 0.2	0.04668 ± 0.05	0.89	33.0 ± 2.5
(4) PCA-109-86, Clachnacudainn gneiss													
A,+149,cl	0.067	270.2	15.44	811	81	0.125	0.05636 ± 0.12	353.5 ± 0.8	0.4180 ± 0.20	354.6 ± 1.2	0.05379 ± 0.12	0.81	362.4 ± 5.5
B,+149	0.043	306.9	18.03	529	94	0.139	0.05722 ± 0.12	358.7 ± 0.9	0.4295 ± 0.24	362.8 ± 1.4	0.05444 ± 0.18	0.68	389.3 ± 8.0
C,125,sc	0.150	285.9	16.45	1602	97	0.128	0.05661 ± 0.10	355.0 ± 0.7	0.4201 ± 0.14	356.1 ± 0.8	0.05382 ± 0.08	0.85	363.6 ± 3.4
D,90,sc	0.044	395.4	23.98	1240	53	0.146	0.05868 ± 0.13	367.6 ± 0.9	0.4461 ± 0.16	374.5 ± 1.0	0.05513 ± 0.09	0.84	417.7 ± 4.0
(5) PCA-M47-85, Seymour Range orthogneiss													
A,+149	0.145	593.2	34.00	8522	35	0.173	0.05428 ± 0.07	340.7 ± 0.4	0.4065 ± 0.10	346.3 ± 0.6	0.05431 ± 0.04	0.92	384.0 ± 2.0
B,125,cl	0.089	513.6	29.77	4151	37	0.198	0.05381 ± 0.07	337.8 ± 0.4	0.3981 ± 0.10	340.2 ± 0.6	0.05366 ± 0.05	0.90	356.8 ± 2.2
C,90,cl	0.091	630.6	34.60	4472	42	0.184	0.05151 ± 0.25	323.8 ± 1.6	0.3809 ± 0.26	327.6 ± 1.4	0.05361 ± 0.05	0.98	354.7 ± 2.2
D,68,cl	0.086	889.9	44.11	5782	39	0.167	0.04721 ± 0.07	297.3 ± 0.4	0.3483 ± 0.10	303.5 ± 0.6	0.05353 ± 0.05	0.92	351.3 ± 2.0
ALL-1,na	0.217	198.4	111.8	83	971	27.042	0.02314 ± 0.73	147.5 ± 2.1	0.1621 ± 2.5	152.5 ± 7.0	0.05080 ± 2.0	0.68	231.5 ± 97
ALL-2,na	0.204	365.3	90.46	78	1368	15.435	0.01733 ± 0.82	110.7 ± 1.8	0.1204 ± 2.9	115.4 ± 6.2	0.05040 ± 2.4	0.68	213.4 ± 114
ALL-3,na	0.596	141.1	114.3	68	2350	41.437	0.02199 ± 1.1	140.2 ± 3.0	0.1505 ± 4.0	142.4 ± 10.7	0.04966 ± 3.4	0.67	178.9 ± 167
ALL-4,na	0.557	177.2	100.5	58	2705	36.440	0.01745 ± 1.3	111.5 ± 2.8	0.1229 ± 5.7	117.7 ± 12.7	0.05108 ± 5.0	0.64	244.4 ± 244
ALL-5,na	0.538	191.9	124.1	74	2492	33.467	0.02160 ± 0.89	137.8 ± 2.4	0.1491 ± 3.2	141.2 ± 8.4	0.05007 ± 2.6	0.70	198.2 ± 128
ALL-6,na	0.741	273.1	108.3	81	3981	22.286	0.01961 ± 0.81	125.2 ± 2.0	0.1361 ± 2.9	129.6 ± 7.0	0.05036 ± 2.4	0.69	211.5 ± 116
T-1,na	0.100	612.8	7.139	195	253	0.120	0.01160 ± 0.31	74.3 ± 0.5	0.07670 ± 0.88	75.0 ± 1.3	0.04796 ± 0.71	0.67	97.4 ± 34
T-2,na	0.147	813.1	12.52	397	289	0.179	0.01458 ± 0.16	93.3 ± 0.3	0.09890 ± 0.42	95.8 ± 0.8	0.04921 ± 0.32	0.70	157.9 ± 15
T-3,na	0.136	369.3	3.424	208	169	0.017	0.01014 ± 0.23	65.0 ± 0.3	0.06598 ± 0.83	64.9 ± 1.0	0.04721 ± 0.70	0.63	59.9 ± 34
T-4,na	0.108	530.2	4.894	139	295	0.032	0.009955 ± 0.37	63.9 ± 0.5	0.06464 ± 1.2	63.6 ± 1.5	0.04709 ± 1.0	0.67	53.7 ± 50
T-5	0.922	478.6	4.527	257	1187	0.029	0.01022 ± 0.26	65.5 ± 0.3	0.06752 ± 0.74	66.3 ± 1.0	0.04792 ± 0.59	0.69	95.3 ± 28
T-6,na	0.495	502.0	4.630	208	821	0.028	0.009980 ± 0.25	64.0 ± 0.3	0.06505 ± 0.94	64.0 ± 1.2	0.04727 ± 0.81	0.60	63.0 ± 39
T-7,na	1.022	479.1	5.201	256	1442	0.087	0.01112 ± 0.22	71.3 ± 0.3	0.07373 ± 0.62	72.2 ± 0.9	0.04808 ± 0.51	0.65	103.1 ± 24
(6) 2090-RA-1, Shaw Creek Intrusion, Bayonne Batholith													
A,+149	0.177	756.4	8.717	3319	31	0.056	0.01212 ± 0.08	77.7 ± 0.1	0.08289 ± 0.12	80.9 ± 0.1	0.04961 ± 0.08	0.76	176.6 ± 1.9
B,125	0.165	648.0	13.03	4267	32	0.096	0.02002 ± 0.09	127.8 ± 0.1	0.1983 ± 0.12	183.7 ± 0.2	0.07184 ± 0.07	0.77	981.3 ± 1.5
C,90	0.222	702.1	16.91	5772	40	0.116	0.02339 ± 0.08	149.1 ± 0.1	0.2605 ± 0.11	235.0 ± 0.2	0.08076 ± 0.05	0.93	1215.6 ± 0.9

** M = monazite; ALL = allanite; T = titanite; numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other zircon fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;
Weighing error = 0.001 mg.
\+ Radiogenic Pb.
* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU
\# Total common Pb in analysis corrected for fractionation and spike.
++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

distances eastwards during the Cretaceous-Paleocene displacement on the Monashee décollement. Because of this, these Devonian gneisses have no counterpart in the footwall.

(6) SHAW CREEK INTRUSION, BAYONNE BATHOLITH (2090-RA-1)

During field studies in Nelson east-half map area, John Reesor of the Geological Survey of Canada collected a sample of the Shaw Creek stock within the composite Bayonne Batholith of the southern Kootenay Arc near Creston, British Columbia (Fig. 1). This area has had extensive K-Ar geochronology (Archibald et al., 1983, 1984), and plutons of ca. 100 Ma and ca. 165 Ma have been identified. The sample is from upper Shaw Creek and is near the eastern edge of what is shown as Shaw Creek intrusion (Fig. 3 of Archibald et al., 1983). The intrusion has been described as being apparently gradational into the Mine Stock to the west which is older than 166 Ma (Archibald et al., 1983), but also gradational into the southern Bayonne Batholith of presumed Cretaceous age. As indicated in Figure 3 of Archibald et al. (1983), the Shaw Creek stock was inferred by Archibald to be Jurassic in age. The objective of dating this sample was to determine its age and attempt to determine to which plutonic suite it belonged. The stock intruded Purcell Supergroup and is in uncertain contact with other intrusions on its east and west sides.

The sample is a biotite granodiorite with about 10% K-feldspar megacrysts. The sample contains zircon, titanite, and apatite as accessory minerals.

Three zircon fractions were abraded and analyzed and they are shown in Figure 9. All fractions were discordant but fraction A is sufficiently close to concordia to infer that the sample is less than 80 Ma old. A regression of the three points has a lower intercept age of 76 ± 6 Ma, which is interpreted as the age of the intrusion. These data demonstrate that the intrusion is younger than the generally accepted age of ca. 100 Ma for the majority of the Bayonne Batholith (Archibald et al., 1984). In the immediate vicinity of this sample, K-Ar dates on micas are 50-60 Ma, and elsewhere in the Bayonne Batholith they range from 70-100 Ma.

Few reliable indications of emplacement age are available for the composite Bayonne Batholith, and since the area has seen significant thermal perturbations, K-Ar dates are not a particularly reliable means of providing crystallization ages. The ca. 76 Ma age presented here indicates that, in addition to the 100 Ma estimate for the composite Bayonne Batholith and the ca. 165-170 Ma ages for the Mine and related stocks, Late Cretaceous plutonism was also present.

SUMMARY

This paper has presented U-Pb data for six samples of granitic rocks from southeast British Columbia. These ages help define the extent of granitic rocks referred to as the Eocene Ladybird granite suite; they also identify ca. 360 Ma granitic rocks which occur on both sides of the Monashee complex; finally, a sample of ca.76 Ma granodiorite has been identified in the composite Bayonne Batholith.

ACKNOWLEDGMENTS

I thank Richard Brown, Murray Journeay, and John Reesor for collecting some of the samples presented in this paper. The staff of the geochronology laboratory are thanked for assistance in generating U-Pb data. Vicki McNicoll helped prepare the tables and concordia diagrams and assisted in the preparation of this manuscript. R. Theriault reviewed the manuscript.

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APPENDIX 1

Sample locations

(1) PCA-141-86: weak to moderately foliated biotite leucogranite; located on a logging roadcut at an elevation of 5250 ft., near the headwaters of Renata Creek, B.C.; NTS 82E, UTM zone 11u, 409400E-5480150N.

(2) PCA-155-86: Fractured, coarse grained biotite-granite; collected from a small roadcut beside the Kettle River, 10.6 km north of bridge over Rendell Creek (between Nevertouch and Damfino Creeks), B.C.; NTS 82E, UTM zone 11u, 472600E-5509200N.

(3) PCA-349-86: Leucogranite; collected from Mt. Revelstoke highway roadcut at elevation 6050 ft., about 1 km south of fire lookout on Mt. Revelstoke, B.C.; NTS 82N, UTM zone 11u, 419800E-5654400N.

(4) PCA-109-86: Strongly deformed orthogneiss; located in the CP rail yard quarry, 1.2 km west of Albert Canyon village, B.C.; NTS 82N, UTM zone 11u, 438750E-5664800N.

(5) PCA-M47-85: Foliated hornblende-biotite granodiorite gneiss; collected due east of small lake in saddle of ridge, eastern Seymour Range, B.C., NTS 82M, UTM zone 11u, 363700E-5716700N.

(6) 2090-RA-1: Biotite granodiorite of the Shaw Creek intrusion; collected at an elevation of 5750 ft. on logging road above and south of Shaw Creek, B. C.; NTS 82F, UTM zone 11, 515870E-5451100N.

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Geology and age of the Naver and Ste Marie plutons, central British Columbia

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Struik, L.C., Parrish, R.R., and Gerasimoff, M.D., 1992: Geology and age of the Naver and Ste Marie plutons, central British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 155-162.

Abstract

In the central Canadian Cordillera, the Eureka Thrust, a suture between upper Paleozoic-lower Mesozoic oceanic and arc rocks and North American continental rocks, is pierced by the ca. 110 Ma (U-Pb) (mid-Cretaceous) Naver Pluton, an orthoclase megacrystic biotite granite. The ca. 167 Ma (U-Pb) (Middle Jurassic) Ste Marie Pluton, a microcline megacrystic hornblende granite, intrudes rocks above the thrust but is inferred to cut the fault at depth. K-Ar ages of each of these plutons are only slightly younger than their U-Pb ages. Both the Naver and Ste Marie plutons are unfoliated except along discrete faults.

Résumé

Dans le centre de la Cordillère canadienne, la zone de chevauchement d'Eureka, zone de suture entre les roches océaniques et d'arc du Paléozoïque supérieur et du Mésozoïque inférieur et les roches continentales nord-américaines, est percée par le pluton de Naver du Crétacé moyen âgé d'environ 110 Ma (U-Pb) composé d'un granite à biotite mégacristallin et à orthoclase. Le pluton de Ste Marie du Jurassique moyen dont l'âge est d'environ 167 Ma (U-Pb) et composé d'un granite à hornblende mégacristallin et à microcline, recoupe par intrusion les roches situées au-dessus de la zone de chevauchement; il devrait recouper la faille en profondeur. Les datations K-Ar de chacun de ces plutons donnent des âges légèrement plus récents que les datations U-Pb. Les plutons de Naver et de Ste Marie ne sont pas foliés sauf le long de failles distinctes.

INTRODUCTION

Southeast of Prince George, British Columbia (Fig. 1), feldspar megacrystic biotite granite forming the Naver Pluton intrudes the suture between "obducted" oceanic rocks and continental sedimentary rocks of the ancient western North American margin. Throughout central British Columbia this suture is called the Eureka Thrust or, locally, the Quesnel Lake shear zone (Rees, 1987). Wanless et al. (1967) dated the Naver Pluton as mid-Cretaceous from K-Ar on biotite (104 Ma and 107 Ma), and that age has been confirmed by new K-Ar isotopic ages of biotite and U-Pb isotopic ages of zircon, each of which are reported in this paper. The Eureka Thrust is therefore older than mid-Cretaceous. Directly to the north of the Naver Pluton the mainly feldspar megacrystic hornblende granite of the Ste Marie Pluton intrudes arc basalt and greywacke above the Eureka Thrust, and possibly the Eureka Thrust at depth. Wanless et al. (1967) dated the Ste Marie Pluton as Middle Jurassic using K-Ar on hornblende (176 Ma), and we report a slightly younger age using the same technique and also U-Pb dating of zircons.

TECTONIC SETTING

In central British Columbia, as throughout the strike length of the Cordillera, "oceanic rocks" from the west have overthrust the Precambrian and Paleozoic sedimentary rocks interpreted to be part of the ancient western North American continental margin. Published tectonic models require that oceanic basaltic crust developed in the late Paleozoic and that

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this crust was obducted onto the ancient western margin of North America. In the Canadian Cordillera the obduction is thought to be Jurassic (for some examples and further references see Tempelman-Kluit, 1979; Erdmer, 1982; Nelson and Bradford, 1989; Struik, 1988; Schiarizza and Preto, 1987; and Klepacki, 1985), although in the oceanic thrust sheet thrust faults can be dated as Permian or older (Harms, 1985; Klepacki, 1985). In the United States Cordillera the time of obduction is controversial; many authors consider it to be Late Permian or Early Triassic and part of a prolonged upper Paleozoic subduction-generated accretionary wedge imbrication of the oceanic sequence, (for recent examples and further references see Miller et al., 1984; Tomlinson et al., 1987; and Babaie, 1987), however some suggest the obduction to be Jurassic or Cretaceous (see Ketner, 1984). Throughout British Columbia the outer margin continental rocks are included in Selkirk Terrane (Struik, 1986) or the nearly equivalent Kootenay Terrane (Wheeler et al., 1988). Within central British Columbia these outer margin clastic sedimentary rocks are included in the Precambrian and Paleozoic Snowshoe Group of Selkirk Terrane. The oceanic rocks are part of the Slide Mountain Terrane (Monger and Berg, 1984), and in central British Columbia consist of the Crooked Amphibolite, which has been considered part of the Slide Mountain Group (Struik, 1988). Thrust onto the Slide Mountain Terrane are the lower Mesozoic volcanic and sedimentary arc rocks of Quesnel Terrane. To the north rocks of similar age and composition are mapped as Takla Group (Muller and Tipper, 1969) and to the south as Nicola Group (Campbell and Tipper, 1971) and informally as Quesnel River Group (Campbell, 1978). For the purpose of this paper these Quesnel Terrane rocks will be informally included in the Nicola Group, following recent usage of the term Nicola Group for rocks immediately to the south (Bailey, 1989).

In central British Columbia these ocean basin and continental rocks were imbricated along the Eureka Thrust (Fig. 1). Southeast of Prince George the mid-Cretaceous Naver Pluton intruded the continental rocks and pierced the overlying Eureka thrust and allochthonous oceanic sequence and higher thrust imbricates of volcanic arc rocks. The upper Middle Jurassic Ste Marie Pluton intruded the allochthonous Slide Mountain and Quesnel terranes and may also have intruded the Eureka Thrust at depth. This paper describes the Naver and Ste Marie plutons and several U-Pb and K-Ar dates obtained from them.

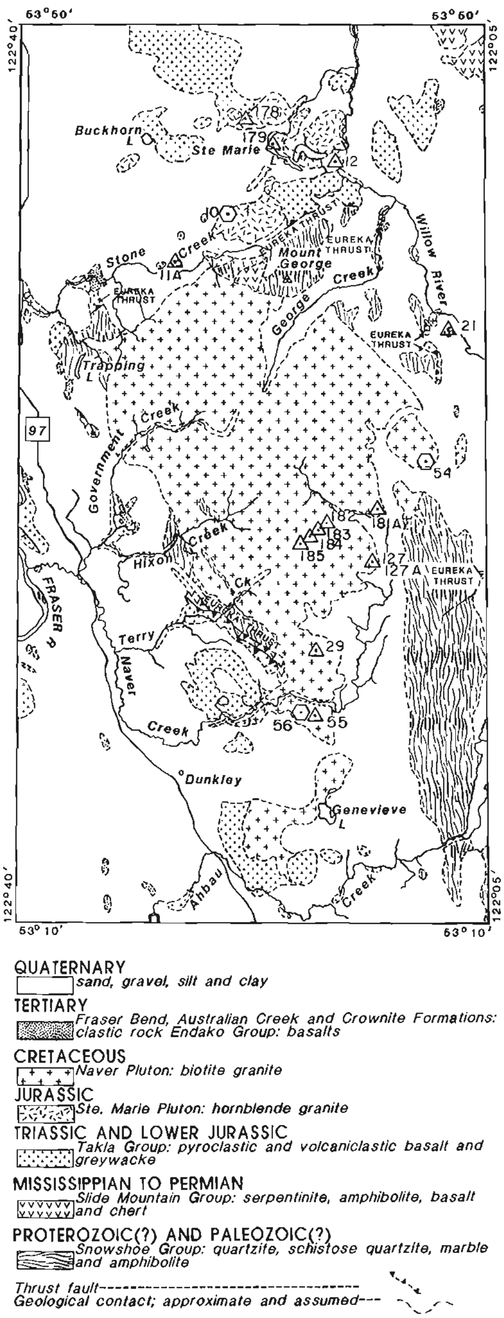

Figure 1. Regional geology around the Naver and Ste Marie plutons in central British Columbia. Sample sites for point counted specimens (triangles) and isotopically dated specimens (hexagons) are labelled to correspond with Table 1.

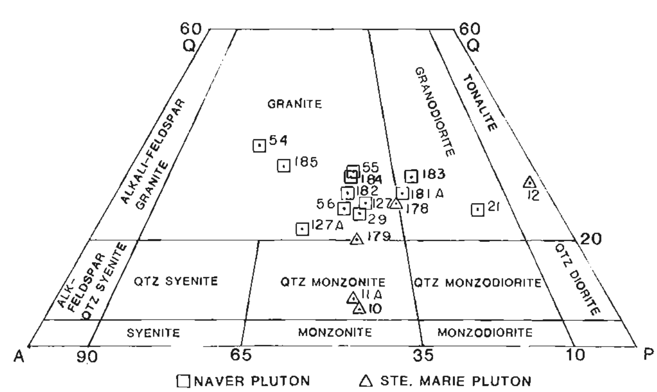

Figure 2. Classification of the point counted specimens from the Naver and Ste Marie plutons, where A = Alkali Feldspar, Q = Quartz, and P = Plagioclase. Table 1 has the percentage relative concentrations of the macroscopic rock forming minerals.

NATURE OF THE PLUTONS

The Naver Pluton

The Naver Pluton consists of mainly granite and granodiorite (Fig. 2, Table 1). It underlies the area of Mount George and southward to Genevieve Lake (Fig. 1). Small intrusions similar lithologically to the Naver Pluton are found scattered throughout the area to its south and west. Tipper (1960) included the pluton as part of the Early Jurassic Topley intrusions, and noted that part of it may be younger than the Topley suite because it appeared to intrude Jurassic rocks. Isotopic age determinations (Wanless et al., 1967) subsequently demonstrated a Cretaceous age.

The pluton is distinguished macroscopically from the Ste Marie Pluton directly to the north by its general lack of hornblende, except in areas of assimilation of mafic xenoliths. The texture of the pluton varies from equigranular, coarsely crystalline to feldspar megacrystic. The granite and granodiorite contain biotite and locally muscovite except in some of the border phases where they may have hornblende or be leucocratic. White to light pink alkali feldspar phenocrysts (1-10 cm long; Fig. 3) are usually multiply zoned and are perthitic in some places. The feldspar is mostly orthoclase, although in some samples there is some microcline. Accessory minerals include sphene, zircon, rutile and in one case allanite. Secondary chlorite, epidote and sausserite have grown at the expense of hornblende, biotite and plagioclase.

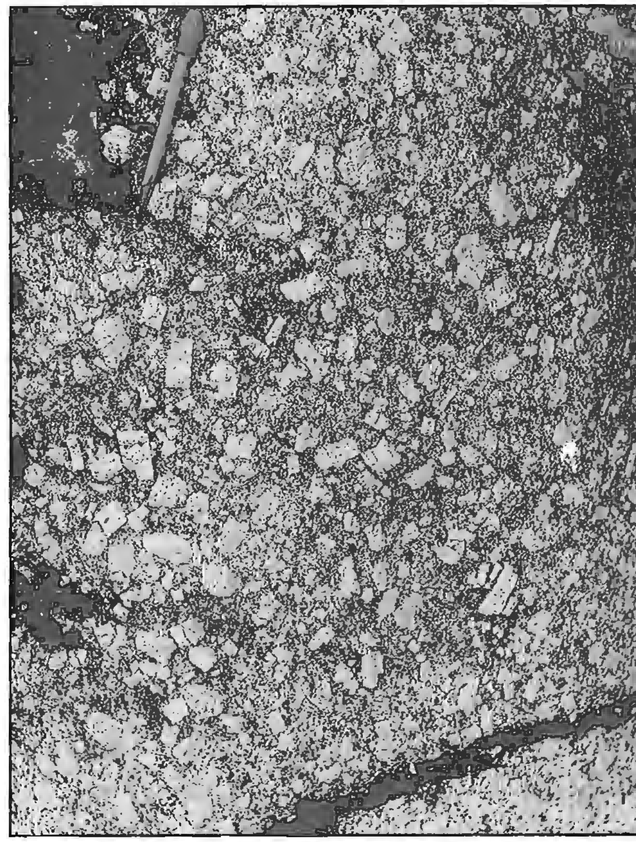

Figure 3. Texture of the Naver Pluton granite where the orthoclase phenocrysts are exceptionaly well developed.

Table 1. Sample locations for modal counts and isotopic age determinations

Sample number	UTM coordinates		%plag	%Kspar	%qtz	%mafics	Felsic minerals only		
	Easting	Northing					%plag	%Kspar	%qtz
10	539050	5949050	42	33	7	18(H)	51	41	7
11A	534140	5945450	44	37	8	11(E,B)	49	42	9
12	548000	5953490	57	2	26	16(E,B)	67	2	31
21	557430	5940050	58	12	25	5(E)	61	13	26
29	546900	5913310	39	31	23	7(B)	42	33	25
54	555810	5928950	19	43	36	2(B)	19	43	38
55	546610	5908250	35	29	32	4(B)	37	30	33
56	545650	5908065	38	33	25	4(B)	39	34	26
127	551800	5920520	41	29	27	3(B)	42	30	27
127A	551800	5920520	32	42	22	4(B)	34	44	22
178	540680	5957450	43	23	24	10(E)	47	26	27
179	542750	5955150	41	33	18	8(H)	44	36	20
181A	551190	5924760	44	23	28	6(B)	47	24	29
182	547540	5923890	37	31	29	3(B)	38	32	29
183	547000	5923190	45	20	30	5(B)	47	21	32
184	546430	5922600	36	29	31	3(B)	37	29	32
185	545475	5922100	25	40	34	2(B)	25	40	34

B, biotite; E, epidote; H, hornblende

Table 2. U-Pb Analytical Data for the Naver and Ste Marie plutons, central British Columbia[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb^*/^{204}Pb$	Pb_c,# (pg)	$^{208}Pb^+/^{206}Pb$	$^{206}Pb^{++}/^{238}U$	$^{206}Pb^{++}/^{238}U$ (Ma)	$^{207}Pb^{++}/^{235}U$	$^{207}Pb^{++}/^{235}U$ (Ma)	$^{207}Pb^{++}/^{206}Pb$	corr. coef.	$^{207}Pb/^{206}Pb$*** (Ma)
SCB-84-56, biotite granite, Naver pluton													
A,68	0.276	1208	21.47	3465	106	0.140	0.01739 ± 0.09	111.2 ± 0.2	0.1162 ± 0.12	111.6 ± 0.2	0.04844 ± 0.06	0.88	120.7 ± 2.8
B,90	0.370	1475	26.17	2588	231	0.143	0.01731 ± 0.13	110.6 ± 0.2	0.1173 ± 0.16	112.6 ± 0.4	0.04915 ± 0.07	0.91	154.9 ± 3.2
C,na,-62	0.438	1565	27.58	4987	150	0.136	0.01728 ± 0.19	110.5 ± 0.4	0.1165 ± 0.20	111.9 ± 0.4	0.04888 ± 0.05	0.97	141.9 ± 2.4
D,na,68	0.429	1537	26.71	4331	166	0.124	0.01723 ± 0.18	110.1 ± 0.4	0.1155 ± 0.20	110.9 ± 0.4	0.04859 ± 0.06	0.96	128.2 ± 2.6
SCB-84-54, biotite granite, Naver pluton													
M-1,na,+74	0.298	1435	351.3	837	590	14.662	0.01798 ± 0.20	114.9 ± 0.4	0.1180 ± 0.29	113.3 ± 0.6	0.04761 ± 0.21	0.68	79.8 ± 10.0
M-2,na,+74	0.096	1417	329.5	961	164	13.876	0.01797 ± 0.10	114.8 ± 0.2	0.1176 ± 0.20	112.9 ± 0.4	0.04747 ± 0.14	0.76	72.8 ± 6.6
A,90	0.459.	4950	81.78	2120	1141	0.095	0.01679 ± 0.53	107.4 ± 1.2	0.1128 ± 0.54	108.5 ± 1.2	0.04872 ± 0.08	0.99	134.5 ± 3.8
B,na,90	0.522	3359	55.56	1184	1574	0.107	0.01664 ± 0.23	106.4 ± 0.4	0.1119 ± 0.28	107.7 ± 0.6	0.04875 ± 0.16	0.81	135.7 ± 7.8
C,na,90	0.258	4396	72.08	1598	740	0.114	0.01641 ± 0.20	104.9 ± 0.4	0.1098 ± 0.24	105.8 ± 0.4	0.04854 ± 0.12	0.86	125.6 ± 5.6
D,na,68	0.824	3160	52.25	1794	1523	0.115	0.01653 ± 0.20	105.7 ± 0.4	0.1107 ± 0.24	106.6 ± 0.4	0.04857 ± 0.11	0.88	127.1 ± 5.4
SCB-84-10, hornblende quartz monzonite, Ste Marie pluton													
A,90,cl	0.684	337.2	9.002	2541	153	0.113	0.02670 ± 0.13	169.8 ± 0.4	0.1832 ± 0.16	170.8 ± 0.4	0.04976 ± 0.07	0.91	183.8 ± 3.0
B,,na,90	0.828	321.9	8.570	4128	109	0.112	0.02665 ± 0.18	169.6 ± 0.6	0.1827 ± 0.20	170.4 ± 0.6	0.04972 ± 0.06	0.96	181.9 ± 2.6
C,na,125	0.683	306.0	8.682	937	407	0.110	0.02841 ± 0.21	180.6 ± 0.8	0.2003 ± 0.28	185.4 ± 1.0	0.05113 ± 0.19	0.73	246.5 ± 8.8

** M = monazite; numbers (i.e. +149,125) refer to average size of zircons in microns; na = not abraded (all other zircon fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;

Weighing error = 0.001 mg.

\+ Radiogenic Pb.

* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU

\# Total common Pb in analysis corrected for fractionation and spike.

++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.

*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

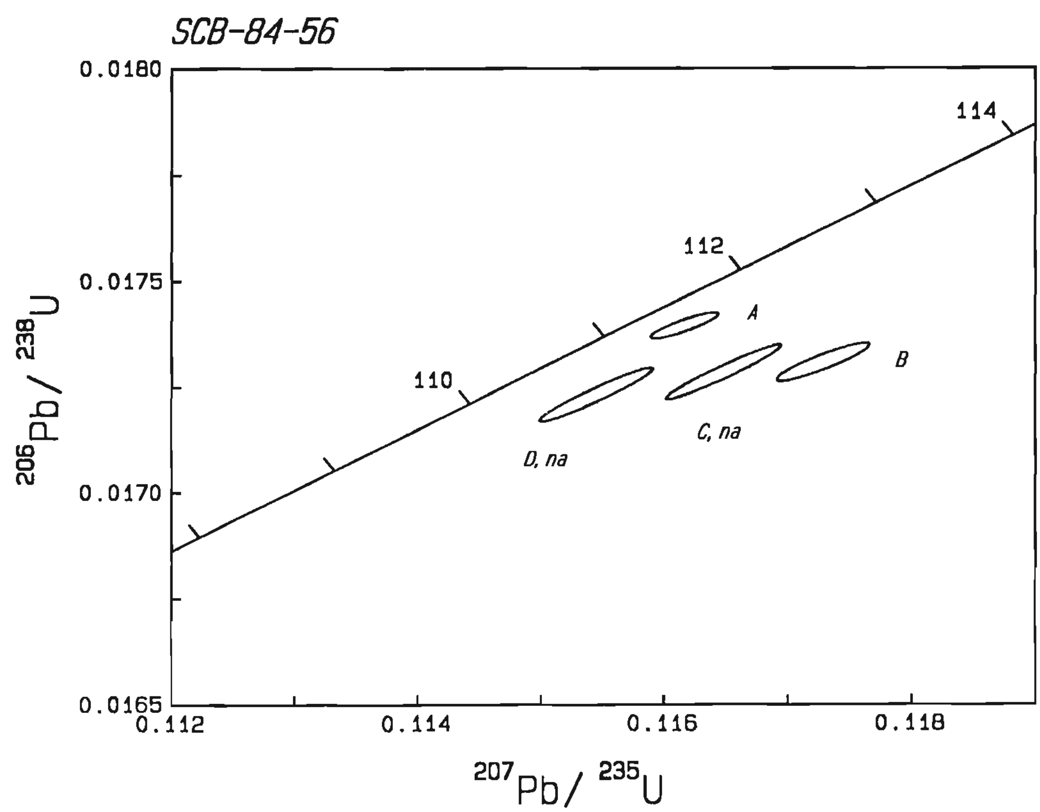

Figure 4. U-Pb concordia diagram for the Naver Pluton, sample SCB-84-56. Error ellipses reflect 2σ uncertainty.

The borders of the pluton are generally narrow (1-10 m) and contacts with the country rock gradational over 0.5 m or less (texturally). The pluton contains lenses of country rock near Hixon and Terry creeks, including serpentinite of the Crooked Amphibolite north of Naver Creek and southeast of the head of Terry Creek. Those lenses are in some places up to a kilometre long and 200 m wide. North of Naver Creek hornblende is the predominate mafic mineral. In this locality the granite has xenoliths of serpentinite and sheared ultramafic rock of the Crooked Amphibolite. Diorite and quartz diorite xenoliths (up to 2 m across) are prominent at the headwaters of Naver Creek along the eastern margin of the pluton. The diorite is fine to medium crystalline (0.5-4 mm crystals), generally equigranular, but locally has small plagioclase phenocrysts. The hornblende and lesser amounts of biotite are partly altered to chlorite in places.

Aplitic and pegmatitic dykes have intruded much of the pluton, but they form a volumetrically minor part of the suite. The pegmatite generally contains muscovite but in places has both muscovite and biotite along with the megacrystic feldspars and quartz. Parts of the pluton are altered along faults and contain sausseritized feldspar and epidote. One such fault that follows Penefather Creek (along the string of samples 182-185, Figure 1) has also induced a weak foliation throughout the granite up to 300 to 400 m from the fault.

Tipper (1960) suggested that emplacement of the Naver pluton metamorphosed the Snowshoe Group sedimentary rocks (called the Cariboo Group at that time). However, the Snowshoe Group is metamorphosed to garnet grade over a large region; it extends southeastward from the Naver Pluton approximately 50 km into the McBride map area to Tregillus Creek. Within the Snowshoe Group the metamorphic garnet grew prior to and during the regional foliation defined by biotite, muscovite and trains of quartz crystals. The Naver pluton intrudes and cross-cuts the mica foliation and locally overprints it with mica-quartz hornfels. The metamorphic conditions of the country rock during intrusion of the Naver Pluton are not known, and therefore the reason for the limited hornfels aureole is uncertain.

Age of the Naver Pluton (SCB-84-56, SCB-84-54)

U-Pb age determinations were done on samples of the Naver Pluton using analytical methods described by Parrish et al. (1987), including zircon abrasion (Krogh, 1982). Additional analytical notes and references are provided at the bottom of Table 2. K-Ar ages on biotite are 98 ± 3 and 101 ± 2Ma (GSC K-Ar 88-30, 31; Hunt and Roddick, 1988) for samples SCB-84-56 and SCB-84-54, respectively.

Sample SCB-84-56

Four zircon fractions were analyzed and are slightly discordant (Fig. 4). The zircons were abraded, but not strongly so, and their uranium contents were high, between 1200 and 1600 ppm, making Pb loss a probable explanation for their discordance. Fractions B, C, and D lie on a short linear array and indicate a small amount of zircon inheritance. Fraction A is just barely below concordia, and indicates an age of of approximately 111 Ma or slightly older. The ^{207}Pb-^{206}Pb age of this fraction is 121 Ma, probably a maximum age for this zircon.

Sample SCB-84-54

From this monazite-bearing biotite granite, four zircon and two monazite fractions were analyzed; zircon fraction A was abraded but the others were not. The zircons were cloudy and very high in uranium, and consequently Pb loss is no doubt present, serving to render an age interpretation based solely on zircon data questionable. The limited spread of the four zircon analyses suggests a slight amount of zircon inheritance in addition to Pb loss (Fig. 5). Two monazites are reversely discordant and overlap each other above 113 Ma. They are reversely discordant due to excess ^{206}Pb from the decay of ^{230}Th (Parrish, 1990). Based on criteria outlined in Parrish (1990) for monazite U-Pb systematics, it is suggested that the igneous monazite crystallization age is 113 ± 1 Ma, which we infer is the crystallization age of the sample.

Taken together, the zircon and monazite data for these two samples of the Naver pluton suggest that 113 ± 1 Ma is its crystallization age. The ca. 100 Ma biotite K-Ar dates indicate relatively slow cooling after crystallization.

Ste Marie Pluton

Samples from the Ste Marie Pluton are mainly granite and quartz monzonite; one is a tonalite (Fig. 2, Table 1). The pluton underlies the area north and south of the Buckhorn Lake road, west of Ste Marie Lake (Fig. 1). Tipper (1960) included these granitoids with the Topley intrusions. Later, Tipper et al. (1979) differentiated the Ste Marie and Naver plutons on the basis of their isotopic ages (Wanless et al., 1967).

The pluton is distinguished macroscopically from the Naver Pluton by hornblende as the predominant mafic mineral. Satellite plugs that contain hornblende are interpreted to be part of the Ste Marie Pluton. The Ste Marie Pluton has microcline that forms white subhedral to mainly euhedral phenocrysts (up to 5 cm long). The plagioclase (mostly oligoclase, some albite and andesine) forms euhedral and subhedral microphenocrysts and subhedral to anhedral interstitial crystals. Rarely the interstitial plagioclase is myrmekitic. Quartz forms irregular crystals within the interstices of the phenocrysts and as inclusions within the feldspars and hornblende. Hornblende, minor biotite and epidote form the macroscopic subordinate constituents. Euhedral hornblende needles can be up to 5 mm long. Accessory minerals include titanite, zircon, and rutile. The epidote appears to be secondary; it occurs as irregular rims to and patches within hornblende, as dustings and small crystals within plagioclase and as anhedral crystals throughout parts of the late plagioclase and quartz. Along the north-northwesterly trending faults parallel to the Willow River, a sample examined showed early subhedral 1-2 mm crystals of epidote and later irregular patches of polycrystalline aggregates. Presumably the epidote in that sample has completely replaced hornblende which is absent.

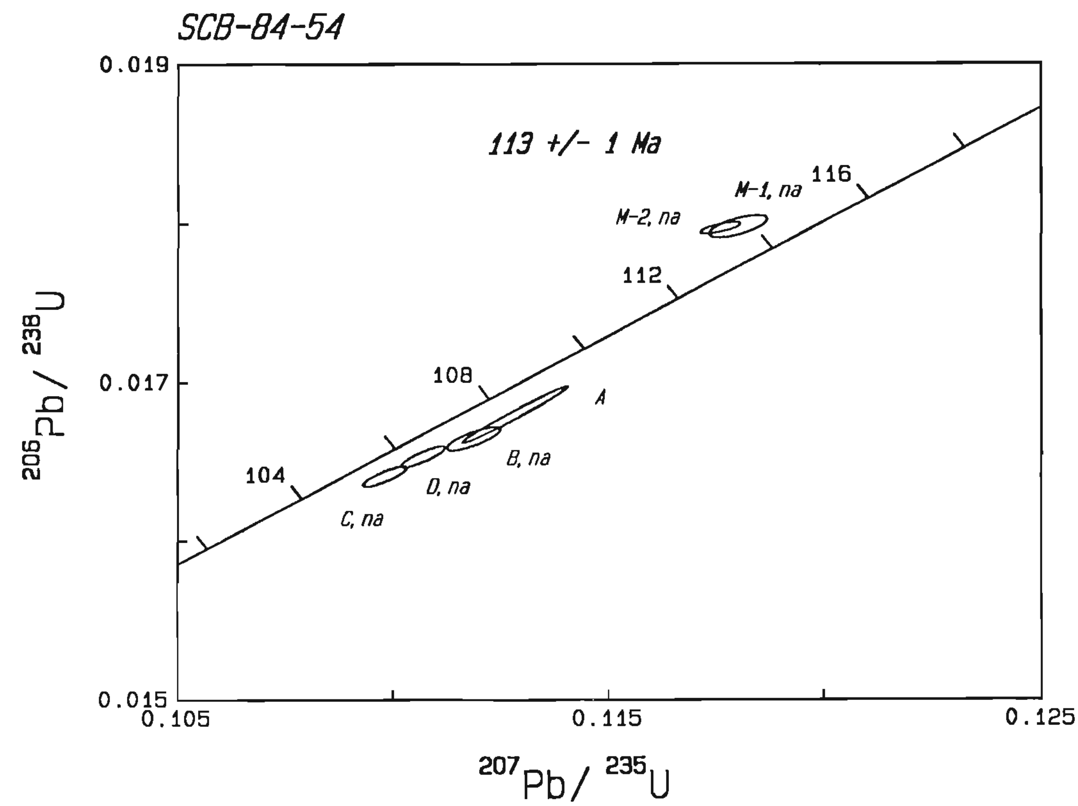

Figure 5. U-Pb concordia diagram for the Naver Pluton, sample SCB-84-54. Error ellipses reflect 2σ uncertainty.

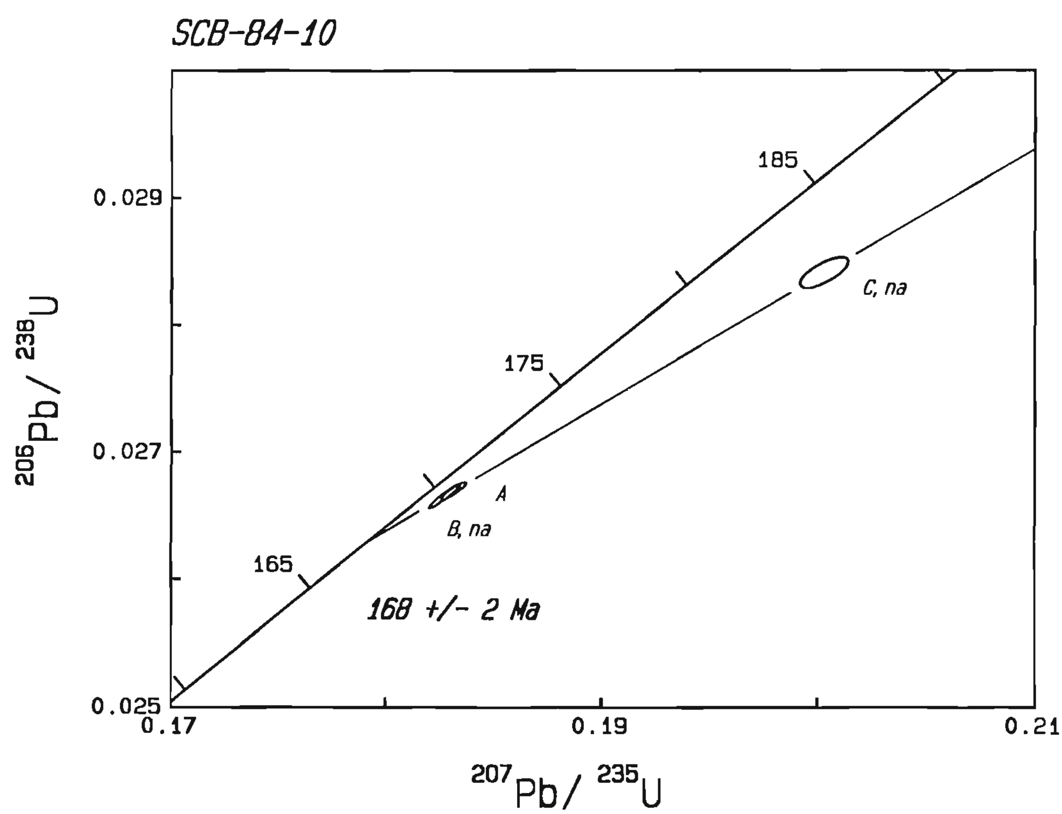

Figure 6. U-Pb concordia diagram for the Ste Marie Pluton, sample SCB-84-10. Error ellipses reflect 2σ uncertainty.

At that same locality minor muscovite may also be a consequence of alteration. Minor chlorite has crystallized around and within the hornblende. The amount of chlorite is proportional to the amount of epidote. Sausseritic alteration affects the cores of the plagioclase phenocrysts to about 5-25%. A portion of the pluton's contact with "Nicola" Group black phyllite of Quesnel Terrane is exposed north of Stone Creek. Although the phyllite has been partly recrystallized at the contact with the pluton, it retains its regional foliation. Aplitic veins extend from the finely crystalline border of the pluton into the black phyllite. Approximately 150 m into the pluton from the contact the granite has a 4 mm equigranular texture.

Just north of the headwaters of Stone Creek, at the site where the sample was collected for isotopic dating (sample 10, Table 1), the granite has xenoliths (less than 25 cm across) of darker granitoids (less than 1%) and a single xenolith of quartzite (20 cm long) with a dioritic halo.

Age of the Ste Marie Pluton (SCB-84-10)

The dated sample is a hornblende quartz monzonite with zircon, titanite and apatite as accessory minerals. A K-Ar date on hornblende from the pluton is 165 ± 3Ma (GSC K-Ar 88-29; Hunt and Roddick, 1988). Three zircon fractions were analyzed, one of which was abraded (Fig. 6), and they are all discordant, though in fractions A and B the discordance is slight. This discordance is mainly caused by older zircon inheritance, possibly of Precambrian age. We suggest that the lower intercept of a chord drawn through the analyses, 168 ± 2 Ma, is a reasonable age of zircon crystallization, and this is interpreted as the age of the sample. The excellent agreement between the hornblende K-Ar and U-Pb ages indicates rapid cooling to below about 500°C following crystallization.

Tectonic history of the Ste Marie Pluton

The Ste Marie Pluton overprints the foliation in the black phyllite of Quesnel Terrane and therefore that foliation is older than 167 Ma. Because the pluton itself is not regionally foliated in any way the pre-167 Ma foliation in the phyllite may be the last intense and regional deformation to have affected this level of the crust. The pluton does not cut the trace of the Eureka Thrust. The dated sample has evidence of inherited zircon; it may have intruded and incorporated the Snowshoe Group and older rocks of the Selkirk Terrane at depth. The one quartzite xenolith found at site 10 (Table 1) may be part of the underlying Precambrian Snowshoe Group. If that were true, then the pluton should have intruded the Eureka thrust at depth. The Eureka thrust to the south is cut by, and therefore older than, the Raft Batholith, inferred by Jung (1986) to be of Jurassic age. As in the case of the Naver Pluton, the scattered satelites around the Ste Marie Pluton may reflect a larger subsurface extent to the pluton.

SUMMARY

The Naver Pluton, which clearly cuts major faults in the map area, is interpreted to be 113 ± 1 Ma old, whereas the Ste Marie Pluton, which is in the hangingwall of the Eureka thrust, is 168 ± 2 Ma old. It may intrude the fault at depth, but this cannot be determined using exposed geological relationships.

ACKNOWLEDGMENTS

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U-Pb zircon age of a clast from the Uslika Formation, north-central British Columbia, and implications for the age of the Uslika Formation

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Parrish, R. R. and Tipper, H. W., 1992: U-Pb zircon age of a clast from the Uslika Formation, north-central British Columbia, and implications for the age of the Uslika Formation; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 163-166.

Abstract

A syenitic clast collected from the Uslika Formation of north-central British Columbia is 168 ± 1 Ma old, and indicates that the Uslika Formation is Callovian or younger, and not Early Jurassic in age.

Résumé

Un claste syénitique prélevé dans la formation d'Uslika dans le centre-nord de la Colombie-Britannique a été daté à 168 ± 1 Ma, indiquant que la formation d'Uslika remonte au Callovien ou à un étage plus récent et non pas au Jurassique inférieur.

INTRODUCTION AND GEOLOGICAL PROBLEM

Spanning the boundary of the Fort St. James and Aiken Lake map areas in north central British Columbia, a narrow composite belt of sedimentary rocks is downfaulted and preserved near the western margin of the Omineca Belt.

The largest exposure of sediments, a faulted sliver of coarse boulder conglomerate and micaceous conglomerate/sandstone in the southernmost part of Aiken Lake map area, has been termed the Uslika Formation (Fig. 1; Roots, 1954). This belt of conglomerate is up to 3 km wide and about 12 km long and occurs east and southeast of Uslika Lake. The Uslika conglomerate is juxtaposed on both sides against Upper Paleozoic sedimentary and volcanic rocks. It is described by Roots (1954, p. 187-190) to consist of conglomerate with clasts of volcanic and plutonic rocks, quartzite, vein quartz, chert, and minor schist and limestone. Roots considered it to be probable Early Cretaceous in age based on correlation with other dated beds nearby, but no fossils were found in the Uslika Formation.

A second belt of sedimentary rocks occurs in another thin fault sliver southwest of Uslika Lake in a narrow 0.5 km by 3 km belt (labelled 2 in Fig. 1). Roots (1954, p.191-192) described this section as consisting of conglomerate with well-rounded clasts of granitic rock, chert, volcanic rocks, sandstone, and minor clasts of shale, schists and argillite, all within a sandy matrix. The conglomerate is overlain by shale and micaceous sandstone containing abundant woody fragments (up to 22 inches in diameter) and coaly beds. The total section was thought to be about 400 feet thick. Fossils were identified as Albian or later, and are likely mainly Late Cretaceous to early Tertiary, and as a result, these rocks were correlated with the Sustut Group. Roots (1954, p. 192) suggested that these beds correlated with the Uslika conglomerate and were separated from it by downfaulting and erosion. The base of the Uslika is likely faulted, as described by Roots.

On strike with the Uslika Formation to the south in the Fort St. James map area, but separated from the Uslika by about 10 km of limited or no exposure, lies a third belt, termed the Discovery Creek sediments, which outcrop in the banks of Discovery Creek (Armstrong, 1949; Tipper, unpublished

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mapping; labelled 3 in Fig. 1). The Discovery Creek beds consist of micaceous sandstone, siltstone, shale, conglomerate, and minor coal. One of the authors (Tipper) discovered unequivocal marine fossils of early Toarcian to late Pliensbachian(?) age; these rocks are thus thought to be late Early Jurassic in age. These sedimentary rocks are described on p.55 of Armstrong (1949), and granitic clasts were not noted, unlike within the Uslika Formation. These rocks lie between Triassic Takla Group volcanic rocks to the west, and Upper Paleozoic sedimentary and volcanic rocks to the east. Armstrong (1949) assigned these beds to the Takla Group. There is probably a fault contact, unexposed, on one or more sides. These beds are equivalent in age to the Hall Formation of the Rossland Group in southeast British Columbia.

A fourth exposure of sediments occurs north of Uslika Lake in the lower part of Vega Creek (labelled 4 in Fig. 1). These strata contain fossils of Barremian (the oldest possible) or Cenomanian age (at the youngest; Roots 1954), thus, a middle Cretaceous age is inescapable for these beds.

The peculiar coincidence of these four conglomeratic bodies with similar lithological characteristics but in part different fossils ages prompted this study. It was thought that dating granitic clast(s) in the Uslika Formation, the only one of the three lacking fossils, might allow a resolution to the correlation or lack thereof between these three belts of sediments.

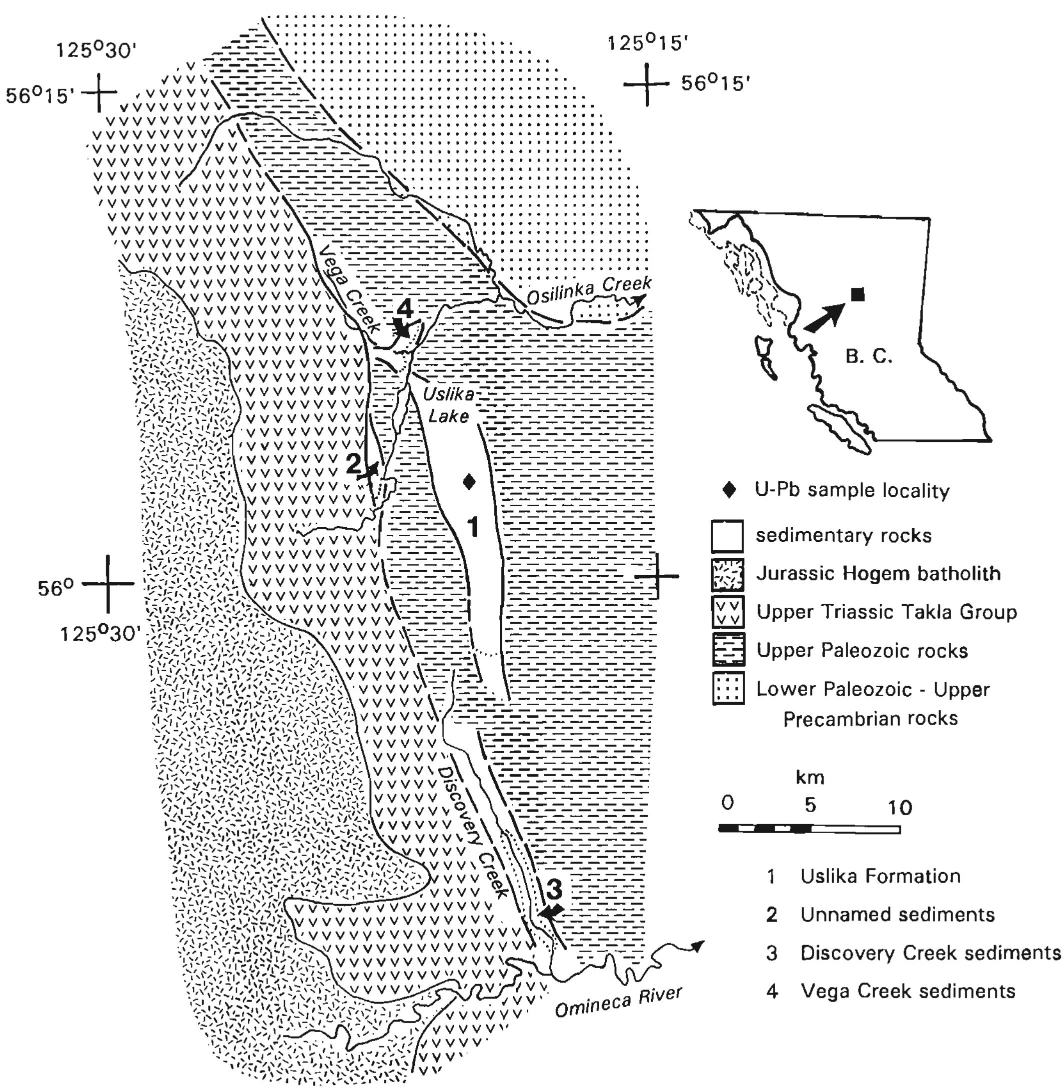

Figure 1. Geological map of the vicinity of Uslika Lake, north-central British Columbia, showing features discussed in the text and the locality of the dated boulder (after Roots, 1954 and Armstrong, 1949).

Table 1. U-Pb Analytical data for clast from the Uslika Formation conglomerate[1]

sample,** analysis	wt.## (mg)	U, (ppm)	Pb,+ (ppm)	$^{206}Pb/^{204}Pb$ *	Pb_c,# (pg)	$^{208}Pb/^{206}Pb$ +	$^{206}Pb/^{238}U$ ++	$^{206}Pb/^{238}U$ (Ma) ++	$^{207}Pb/^{235}U$ ++	$^{207}Pb/^{235}U$ (Ma) ++	$^{207}Pb/^{206}Pb$ ++	corr. coef.	$^{207}Pb/^{206}Pb$ (Ma) ***
85-TD-16-C14, Uslika Formation conglomerate[2]													
B,cl,+105	0.061	221.7	5.907	1231	17	0.195	0.02489 ± 0.10	158.5 ± 0.3	0.1695 ± 0.14	159.0 ± 0.4	0.04939 ± 0.09	0.76	166.3 ± 4.1
C,e,+105	0.068	199.9	5.520	941	25	0.167	0.02637 ± 0.10	167.8 ± 0.3	0.1797 ± 0.17	167.8 ± 0.5	0.04943 ± 0.11	0.77	168.2 ± 5.3
D,e,+105	0.082	203.0	6.412	1574	20	0.177	0.02989 ± 0.10	189.9 ± 0.4	0.2092 ± 0.13	192.9 ± 0.4	0.05075 ± 0.07	0.82	229.7 ± 3.3

** Numbers (i.e. +149,125) refer to average size of zircons in microns; n/a = not abraded (all other zircon fractions are abraded), e = equant, el = elongate, c = cloudy, cl = clear, sc = slightly cloudy, cz = color zoned, co = having visible cores, frags = fragments;
Weighing error = 0.001 mg.
+ Radiogenic Pb.
* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU
# Total common Pb in analysis corrected for fractionation and spike.
++ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock) ; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
*** Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

1 U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

2 Sample locality: Conglomerate Mountain, southeast of Uslika Lake in the Aiken Lake area (94C/3); 125°09'58" W - 56°02'35" N; UTM 10v, 365062E-6212774N.

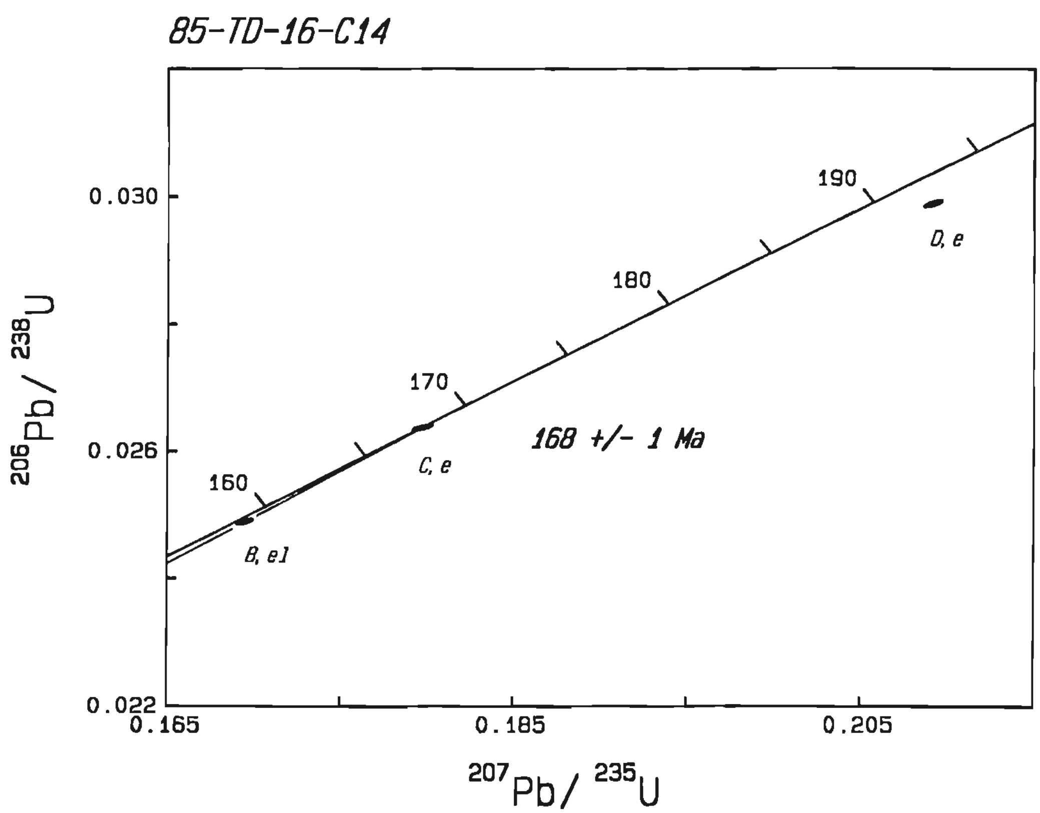

Figure 2. U-Pb concordia diagram for sample 85-TD-16-C14, a syenite boulder from the Uslika conglomerate. Error ellipses reflect 2σ uncertainty.

U-Pb GEOCHRONOLOGY

Several samples of the Uslika Formation were collected for dating; these were detrital mica-bearing sandstones, and several granitic boulders. Upon close examination, it was found that mica was not abundant and was sufficiently altered and difficult to separate as to be unusable. One of the granitic boulders was deemed suitable for dating.

The selected boulder is a syenite clast, with medium grained massive texture. Its location is listed at the bottom of Table 1. It was processed for zircon separation using conventional methods summarized in Parrish et al. (1987), and further zircon U-Pb procedures also follow the Parrish et al. (1987) paper, including abrasion of all zircon fractions. Analytical results are summarized in Table 1 and in Figure 2.

The zircons were of excellent quality, being elongate as well as equant, euhedral crystals without obvious evidence of cores or inheritance, and having a high degree of clarity. Fraction C, of equant habit, is concordant and indicates an age of 168 ± 1 Ma. Fraction B is more discordant, and is composed of crystals of elongate habit with a much greater surface to volume ratio than fraction C, and this is the most likely reason for its Pb loss. A regression line between them intersects concordia at 168 Ma, and it is suggested that these two fractions are inheritance-free and cogenetic. Fraction D is of the equant type of crystals similar to fraction C, but is discordant due to older zircon inheritance. Upon inspection of the photograph of the crystals which make up fraction D, it was noted that there was one grain in the total of 29 which appears different and which was probably a xenocryst. The mixture of this grain with the rest offers the best explanation for this discordant analysis.

We suggest that the source for the boulder is probably the Jurassic syenitic rocks of the composite Hogem batholith.

DISCUSSION

We interpret the 168 Ma zircon age as the age of syenite crystallization. Since the boulder comes from the conglomerate, the age of the conglomerate must postdate 168 Ma, that is, it is Callovian or younger using the DNAG time scale. It therefore cannot be correlative with the Discovery Creek beds which have Toarcian fossils of Early Jurassic age.

Since the Discovery Creek beds are poor in granitic clasts whereas the Uslika Formation and the other sedimentary rocks southwest of Uslika Lake contain notable granitic material, it is plausible that these three latter sequences are correlative and of potential Cretaceous or Early Tertiary age. It is apparent that they must postdate the Discovery Creek sedimentary strata which are of Early Jurassic age.

The exact age of the Uslika Formation is still not known, but at least it can be excluded from the Early Jurassic.

ACKNOWLEDGMENTS

We have relied on the descriptions of rock units from Roots (1954) and Armstrong (1949) in writing this paper, and the authors are appreciative of these descriptions. We thank Anne Kinsman for assistance in the analytical work on this project. V. McNicoll prepared the table and concordia plot, and reviewed the manuscript.

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New U-Pb ages for the Slide Mountain Terrane in southeastern Yukon Territory

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J.K. Mortensen, 1992: New U-Pb zircon ages for the Slide Mountain Terrane in southeastern Yukon Territory; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 167-173.

Abstract

A tectonic block of plagiogranite within serpentinite matrix mélange of the Slide Mountain Terrane within the Finlayson Lake fault zone in southeastern Yukon gives a U-Pb zircon age of 274.3 ± 0.5 Ma. Quartz-feldspar porphyry from a composite plug that occurs within a thrust fault-bounded body of massive greenstone of the Slide Mountain Terrane that is tectonically interleaved with metamorphic rocks of the Yukon-Tanana Terrane gives a U-Pb zircon age of 360.5 ± 1.9 Ma. These data, together with fossil ages reported by other workers, indicate that rocks of the Slide Mountain Terrane in southeastern Yukon range in age from latest Devonian through Early Permian.

Résumé

La datation U-Pb sur zircon d'un bloc tectonique de plagiogranite au sein d'une matrice complexe à serpentinite du terrain du mont Slide dans la zone faillée du lac Finlayson dans le sud-est du Yukon donne un âge de 274,3 ± 0,5 Ma. Le porphyre quartzo-feldspathique d'un bouchon composite situé dans un massif de roches vertes du terrain du mont Slide, limité par des failles chevauchantes et tectoniquement interfolié avec les roches métamorphiques du terrain de Yukon-Tanana, a été daté par U-Pb sur zircon à 360,5 ± 1,9 Ma. Ces données, combinées aux datations de fossiles publiées par d'autres chercheurs, indiquent que les roches du terrain du mont Slide dans le sud-est du chaînon Yukon varient en âge de la toute fin du Dévonien au Permien inférieur.

INTRODUCTION

Massive greenstone, mafic and ultramafic plutonic rocks, and associated sedimentary rocks that together comprise the Slide Mountain Terrane (SMT) in the northern Cordillera occur sporadically throughout large areas of northeastern British Columbia, Yukon and east-central Alaska (Fig. 1). Exposures of Slide Mountain Terrane are largest and most abundant near the boundary between the western edge of the North American miogeocline and the innermost of the allochthonous terranes that abut it on the west (Monger and Berg, 1987; Wheeler et al., 1988; Mortensen, in press). Slide Mountain Terrane forms klippen resting on miogeoclinal strata east of this boundary in northern British Columbia and southeastern Yukon, and on metamorphic rocks of the Yukon-Tanana Terrane (YTT) in Yukon and east-central Alaska (Fig. 1). Bodies of Slide Mountain Terrane are also widespread within the Yukon-Tanana Terrane in Yukon and east-central Alaska as thrust panels imbricated with the Yukon-Tanana Terrane assemblages (e.g. Mortensen and Jilson, 1985; Foster et al., 1985; Mortensen, 1988, in press). Similar fault-bounded bodies also occur rarely within the westernmost edge of the miogeocline (e.g. Tempelman-Kluit, 1977, 1979; Gordey, 1981; Mortensen, 1982).

The Slide Mountain Terrane has been studied in some detail in east-central and northern British Columbia (e.g. Struik and Orchard, 1985; Ferri and Melville, 1988, 1989, 1990; Harms et al., 1988; Nelson and Bradford, 1989), but has received less attention farther north. Correlation between the various bodies of the Slide Mountain in the northern Cordillera are hampered both by the scarcity of fossil and isotopic age constraints, and the complex internal thrust

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imbrication displayed by many of these bodies. Such correlations must be established before any plausible tectonic reconstruction of the Slide Mountain Terrane can be made.

In this study, U-Pb ages were determined for igneous units within two separate bodies of Slide Mountain Terrane in southeastern Yukon. The ages provide the basis for preliminary correlations between the Slide Mountain Terrane in this area and better studied sections in British Columbia.

REGIONAL GEOLOGY

Fault-bounded bodies of Slide Mountain Terrane are widespread and abundant in an area northeast of the Tintina Fault Zone in southeastern Yukon (Fig. 2). These bodies are largest and most abundant within and adjacent to the Finlayson Lake fault zone, a complex structure which separates the miogeocline on the east from the Yukon-Tanana Terrane on the west (Fig. 2), and has been interpreted as a dextral transpressive suture (Mortensen and Jilson, 1985). The miogeocline east of the Finlayson Lake fault zone consists of a nearly complete sequence of mainly sedimentary strata which range in age from Upper Proterozoic through Upper Triassic. West of the Finlayson Lake fault zone, the Yukon-Tanana Terrane comprises a medium to high grade metamorphic assemblage derived largely from meta-igneous and metasedimentary rocks of dominantly Paleozoic age (Mortensen, 1983, in press).

In most of the study area the Finlayson Lake fault zone is overlapped by an elongate synform of stacked thrust sheets, most of which are considered to be part of the Slide Mountain Terrane, that is referred to by Mortensen and Jilson (1985) as the Campbell Range belt (Fig. 2).

Slide Mountain Terrane in southeastern Yukon consists mainly of massive, typically aphyric and non-vesicular greenstone, a variety of mafic to ultramafic plutonic rocks, red, green and grey ribbon chert, and minor marble.

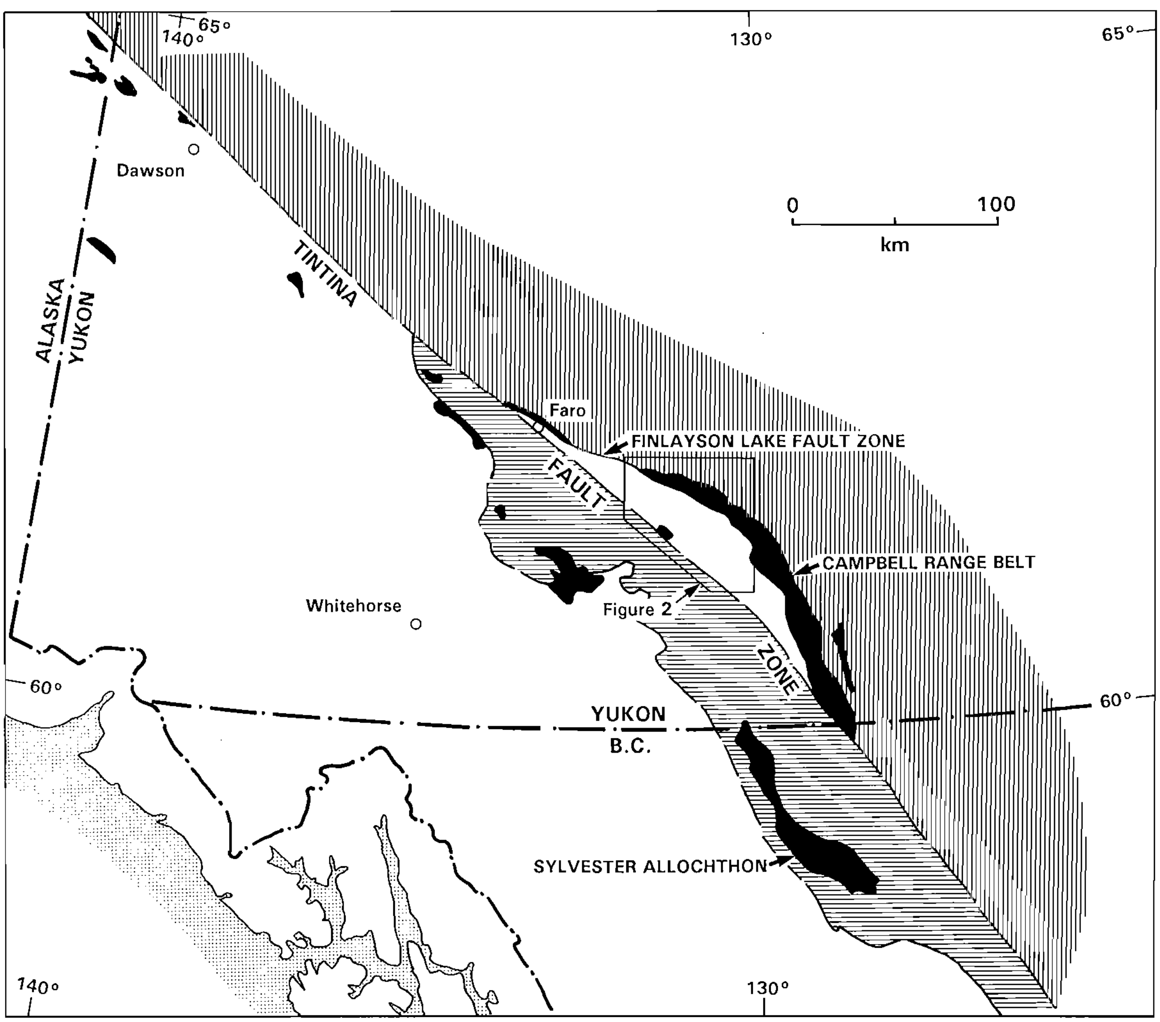

Figure 1. Map of the northern Cordillera, showing the distribution of major bodies of Slide Mountain Terrane (in black). The western edge of the North American miogeocline is shown in vertical hatchure, and the Cassiar Terrane (displaced margin of North America) is shown in horizontal hatchure. Modified from Monger and Berg (1987) and Wheeler et al. (1988).

Metavolcanic rocks of this package were previously mapped as the Anvil Range Group by Tempelman-Kluit (1972, 1977), and the assemblage was collectively referred to as the Anvil Allochthon by Tempelman-Kluit (1979). For convenience it will here be termed the Anvil assemblage.

A major period of post-metamorphic thrust imbrication affected the Yukon-Tanana Terrane west of the Finlayson Lake fault zone in Jurassic time. Discontinuous bodies of sheared greenstone and mafic and ultramafic plutonic rocks of the Slide Mountain Terrane were structurally emplaced along many of the regional-scale thrust faults (Fig. 2). A package of thinly bedded, fine- to coarse-grained clastic rocks, which commonly contain abundant detrital mica and have yielded Upper Triassic conodonts (Tempelman-Kluit, 1979; and personal communication, 1982), occur along one of the thrust surfaces southwest of the Campbell Range belt in the central part of the study area (Fig. 2). The basal contact of this unit is poorly exposed, and it is uncertain whether the sedimentary rocks unconformably overlie or are in thrust contact with the underlying metamorphic rocks.

The Campbell Range belt and Finlayson Lake fault zone (Fig. 1, 2) are poorly understood structures. The Campbell Range belt consists predominantly of interleaved thrust panels of grey chert and metachert, and a mafic to ultramafic

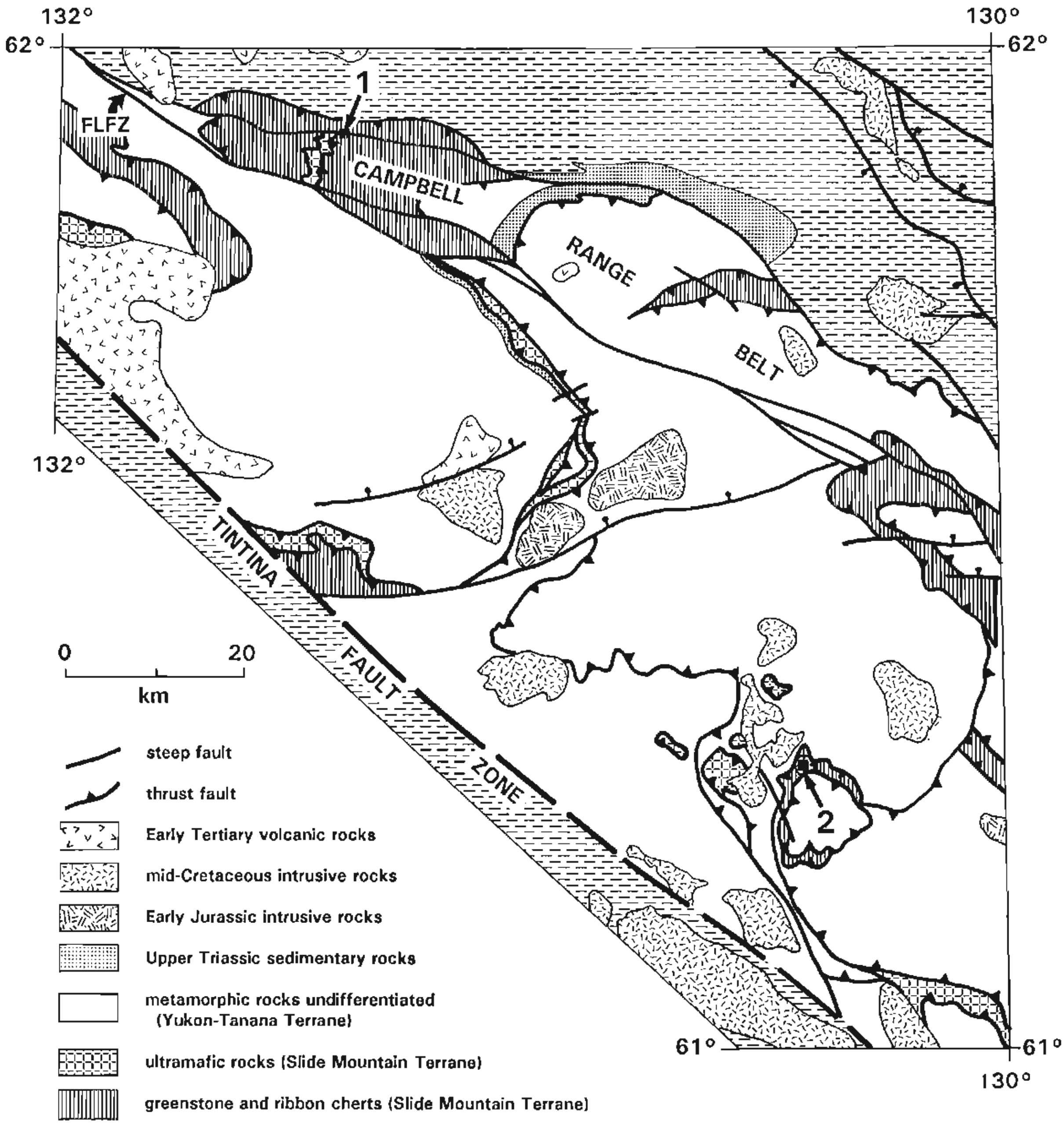

Figure 2. Geological map of part of southeastern Yukon Territory (modified from Mortensen and Jilson, 1985). FLFZ = Finlayson Lake fault zone. Sample localities are shown by numbered squares.

Table 1. U-Pb analytical data

Fraction, Size[1]	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb^3/^{204}Pb$	$^{208}Pb^2$ (%)	$^{206}Pb^4/^{238}U$	$^{207}Pb^4/^{235}U$	$^{207}Pb^4/^{206}Pb$	$^{207}Pb/^{206}Pb$ age[5]
	Sample 1 (plagiogranite; 61°54.8'N, 131°24.1'W)								
A N1,b	0.306	169	7.5	166	11.3	0.04357(.34)	0.3124(1.1)	0.05200(.91)	285.3(20.7)
B N1,+62-74	0.107	187	8.2	1682	10.7	0.04348(.10)	0.3104(.13)	0.05177(.08)	275.5(3.5)
C N1,+44-62	0.085	200	17.1	3908	54.0	0.04343(.09)	0.3107(.11)	0.05188(.05)	280.3(2.1)
	Sample 2 (quartz-feldspar porphyry; 61°15.3'N, 130°24.8'W)								
B N1,↓,+105,a	0.160	368	22.4	2163	12.1	0.05897(.25)	0.4556(.25)	0.05603(.05)	453.6(3.5)
C N1,+74,a	0.107	421	25.3	1194	12.5	0.05785(.19)	0.4301(.26)	0.05393(.11)	368.0(5.0)
D N2,+74,a	0.031	397	24.2	2332	11.4	0.05938(.09)	0.4599(.11)	0.05617(.05)	459.0(2.3)
E N1,+74,a	0.012	302	19.6	957	11.6	0.06284(.33)	0.5227(.61)	0.06032(.53)	615.0(22.8)
F M2,+74,a	0.046	288	21.0	3205	12.7	0.06966(.09)	0.6476(.10)	0.06743(.04)	851.1(1.6)
G N3,+74,c,a	0.065	521	36.9	2694	10.6	0.06884(.09)	0.6721(.11)	0.07082(.05)	952.1(1.8)

[1]sizes (+62-74) refer to size of zircons in microns; N1,2=non-magnetic cut with frantz at 1, 2 or 5 degrees side slope; b=bulk fraction, c=with visible cores, ↓=non-magnetic with magnetic "pin", a=abraded
[2]radiogenic Pb
[3]measured ratio, corrected for spike and fractionation
[4]corrected for blank Pb and U, common Pb, errors quoted are 1σ in percent
[5]corrected for blank and common Pb, errors are 2σ in Ma
Initial common Pb compositions from Stacey and Kramers (1975)

volcanic and plutonic complex that is lithologically similar to and is correlated with the Anvil assemblage, but here includes abundant ribbon chert. Minor amounts of mafic and felsic metavolcanic rocks occur within the chert and metachert package. This package shows some similarities to both the Devonian-Mississippian Earn Group in the miogeocline east of the Finlayson Lake fault zone, and mid-Paleozoic units in the Yukon-Tanana Terrane to the west. It may, however, be in part Triassic in age. Massive bodies of marble and lenses of undeformed clastic sediments which resemble the Upper Triassic unit farther west are also present. Tempelman-Kluit (1972, 1979) obtained Late Triassic conodonts from this clastic sedimentary package in the Faro area, 80 km northwest along the Finlayson Lake fault zone from the study area (Fig. 1), as well as late Pennsylvanian to early Permian fusilinid and conodont ages for massive carbonate interbedded with Anvil assemblage greenstones in the same area. Pennsylvanian or early Permian conodonts have been recovered from a massive marble lens in the Campbell Range belt 10 km east of the study area (M. Orchard, personal communication, 1984). Radiolaria are present in ribbon cherts of the Anvil assemblage in two localities in the Campbell Range belt but are too recrystallized to permit a definite identification (D.L. Jones, personal communication, 1982).

Small bodies of eclogite occur within and immediately east of the Campbell Range belt (Fig. 2) (Mortensen and Jilson, 1985; Erdmer, 1987), and in the Finlayson Lake fault zone in the Faro area (Tempelman-Kluit, 1970; Erdmer and Helmstaedt, 1983; Erdmer, 1987). K-Ar and Rb-Sr muscovite ages which range from 267 to 243 Ma (Middle Permian to earliest Triassic) were obtained from eclogite and blueschist bodies at three separate localities near Faro (Wanless et al., 1978; Erdmer and Armstrong, 1988).

The Finlayson Lake fault zone is generally not well exposed, but appears to consist of anastomosing vertical to steeply dipping faults that are interpreted to be largely strike-slip in nature (Mortensen and Jilson,1985).

ANALYTICAL TECHNIQUES

Mineral concentrates were prepared using standard Wilfley table and heavy liquid techniques. Analytical procedures have been descibed by Parrish et al. (1987). Measured blank levels ranged from 0.040 to 0.006 ng for Pb, and from 0.003 to 0.001 ng for U. Isotopic measurements were done on a Finnegan MAT 261 solid source mass spectrometer equipped with a fully adjustable multiple collector, electron multiplier, and operating software modified to permit simultaneous measurement of all five Pb masses.

Ages were calculated using the constants recommended by Steiger and Jäger (1977). A numerical error propagation technique was used to calculate errors associated with individual analyses (Roddick, 1987). Discordia line fitting and calculation of concordia intercept ages and associated errors employed a modified York-II model (Parrish et al., 1987), and the algorithm of Ludwig (1980). All errors are quoted at the 2σ level. Age assignments follow the time scale of Harland et al. (1989).

U-Pb RESULTS

Plagiogranite

A tectonic block of fine- to medium-grained, unfoliated plagiogranite approximately 3 m by 8 m in surface area occurs within a narrow, steeply dipping, zone of sheared,

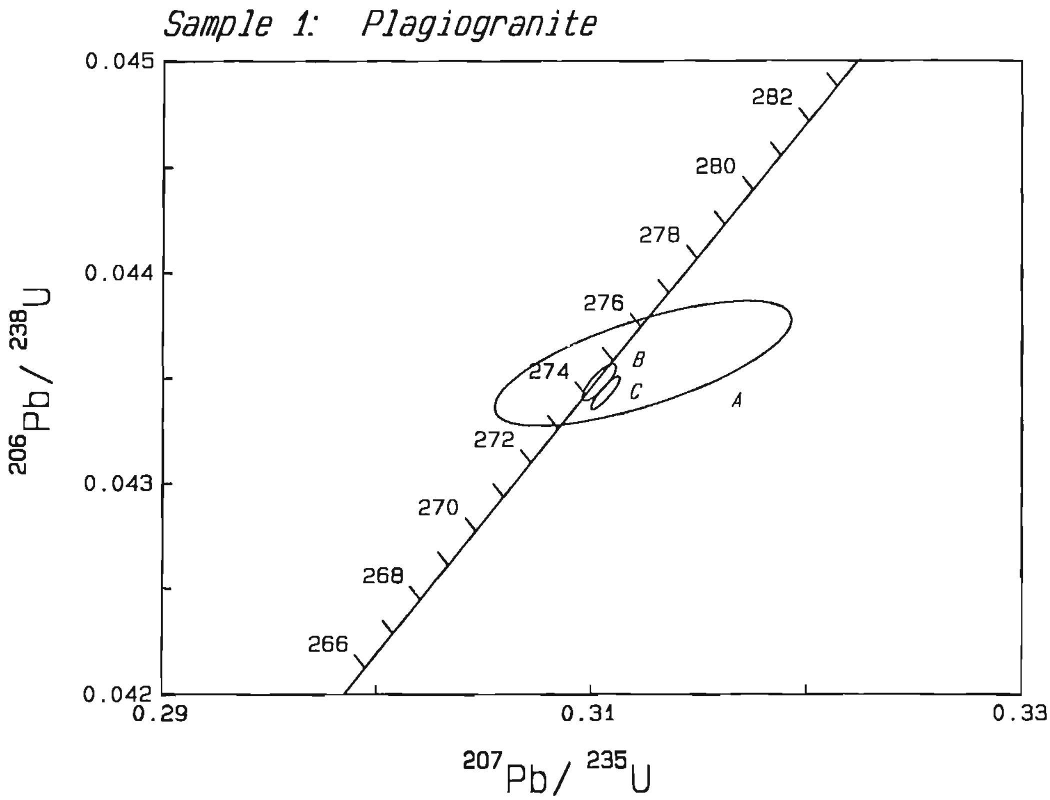

Figure 3a. U-Pb concordia plot for sample 1 (plagiogranite). Error ellipses are shown at 2σ level.

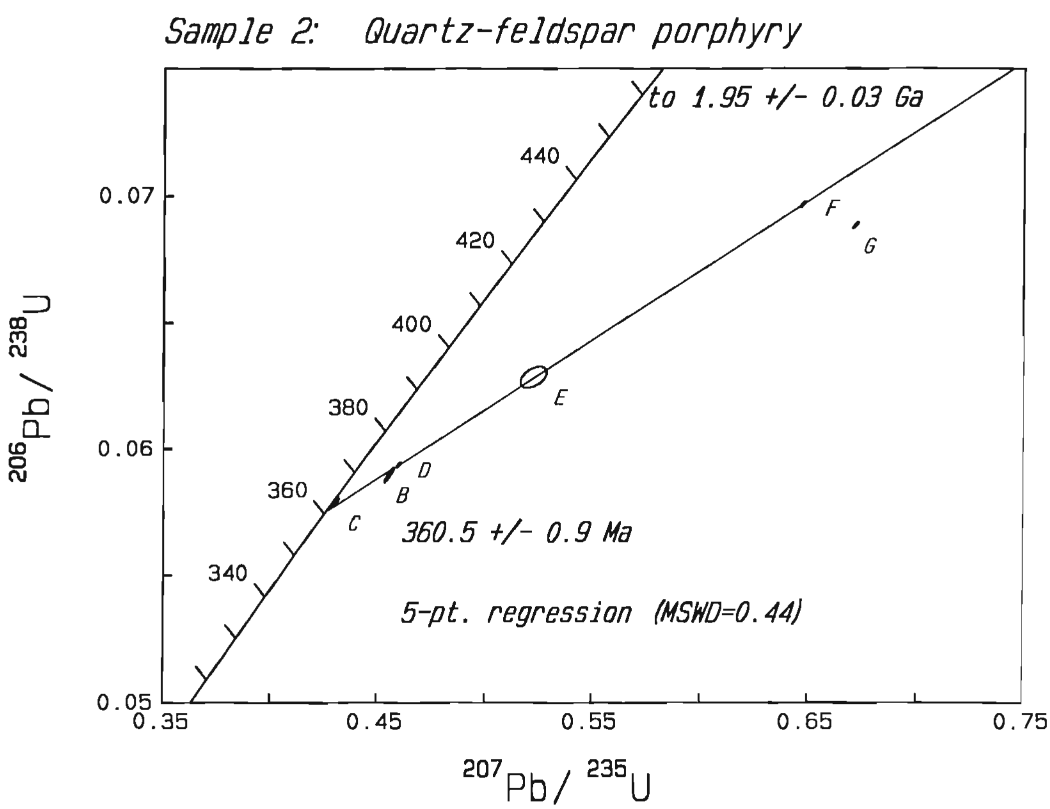

Figure 3b. U-Pb concordia plot for sample 2 (quartz-feldspar porphyry). Error ellipses are shown at 2σ level.

"fish-scale", serpentinite matrix melange ("ophiolitic melange") of the Anvil assemblage near the northern edge of the Campbell Range belt (sample 1 in Fig. 2). The plagiogranite is one of a large number of such blocks that occur in this immediate area. All of the blocks, which include massive greenstone, gabbro, partially serpentinized harzburgite, radiolarian chert and rodingite, are represented elsewhere in the Anvil assemblage, and the plagiogranite is also thought to be part of the assemblage.

A small amount of fine grained, euhedral, prismatic zircon was recovered from a sample of the plagiogranite. Three fractions of unabraded zircon from the sample range from concordant to very slightly discordant (Fig. 3a). The concordant analysis yields a ^{206}Pb-^{238}U age of 274.3 ± 0.5 Ma, which is interpreted as the crystallization age of the plagiogranite.

Quartz Porphyry

Massive greenstone of the Anvil assemblage which comprises a large, thrust-fault-bounded panel within the Money Klippe is intruded by a small, composite, high-level plug (Fig. 2). The plug is massive and unfoliated, and was originally considered by Tempelman-Kluit (1977, 1979) and Erdmer (1981) to be Cretaceous or Tertiary in age. A K-Ar whole-rock age of 251 ± 10 Ma obtained for a sample of andesite porphyry from the body (Tempelman-Kluit, unpublished data) and field evidence which indicates that the plug is truncated at its base by a thrust fault (Mortensen and Jilson, 1985) casts doubt on this age assignment. The data suggest rather that the plug was intruded into the greenstones prior to thrust imbrication of the Slide Mountain Terrane in this area.

Zircons recovered from a sample of quartz porphyry from the plug (sample 2, Fig. 2) are euhedral, stubby to elongate, and pale pink in colour. Cloudy, rounded cores are visible in some of the grains. Five analyses of abraded zircon from the sample (excluding one fraction of fractured zircon which contained cloudy cores) define a discordia line with lower and upper intercept ages of 360.5 ± 1.9 Ma and 1.95 ± 0.33 Ga (Fig. 3b). The lower intercept age is the best estimate of the emplacement age of the plug, and the upper intercept indicates the presence of a significant component of inherited zircon with an Early Proterozoic average age.

DISCUSSION

The nature and origin of the Anvil assemblage, and its original relationship (if any) to the much more strongly deformed and metamorphosed Yukon-Tanana Terrane is problematical. The new age data reported here indicate that the Slide Mountain Terrane, including the Anvil assemblage, comprises rocks which range in age from latest Devonian to Early Permian, and thus overlaps in age with Yukon-Tanana Terrane (Mortensen, in press). Tempelman-Kluit (1979) concluded that the Anvil assemblage represents oceanic crust that was imbricated with the Yukon-Tanana Terrane in an accretionary complex. Hansen (1988), however, argued that Slide Mountain Terrane includes fragments of both ocean floor and the same magmatic arc represented by the deformed and metamorphosed, middle to late Paleozoic portions of the Yukon-Tanana Terrane. There is little evidence from the study area to support a close link between the Anvil assemblage and Yukon-Tanana Terrane other than the similarities in age. The presence of inherited zircon in the quartz-feldspar porphyry unit in the study area, however, indicates that the Slide Mountain does not consist entirely of juvenile oceanic crust, but must also include a source of Precambrian zircons. This source may be either distributed bodies of continental crust that have not yet been identified through field mapping, or sediments derived from continental crust.

On a regional scale, the Anvil assemblage resembles both the Sylvester Allochthon in northern British Columbia and southern Yukon (e.g. Harms et al., 1988; Nelson and Bradford, 1989), and the Fennell Formation of south-central British Columbia, both of which have been assigned to Slide Mountain Terrane (Wheeler and McFeely, 1987; Wheeler et al., 1988). The Fennell Formation consists predominantly of massive and pillowed basalt, gabbro and chert, and has yielded Mississippian through Permian conodont ages, and a Devonian U-Pb zircon age for a quartz-feldspar porphyry flow or sill (Schiarizza and Preto, 1987). Additional work to further geochemically and isotopically characterize the Slide Mountain in southeastern Yukon is presently underway by the author.

CONCLUSIONS

A U-Pb zircon age of 274.3 ± 0.5 Ma has been obtained for a tectonic block of plagiogranite within serpentinite matrix melange of the Slide Mountain Terrane within the Finlayson Lake fault zone in southeastern Yukon. Quartz-feldspar porphyry from a composite plug that intrudes massive greenstone of the Slide Mountain that is tectonically interleaved with metamorphic rocks of the Yukon-Tanana Terrane gives a U-Pb zircon age of 360.5 ± 1.9 Ma. These data, together with fossil ages reported by other workers, indicate that Slide Mountain Terrane in southeastern Yukon ranges in age from latest Devonian through Early Permian. The presence of significant inherited zircon of Proterozoic average age in the quartz-feldspar porphyry indicates that the Slide Mountain Terrane does not consist entirely of juvenile oceanic crust.

ACKNOWLEDGMENTS

The mapping phase of this study was part of a regional mapping project lead by G.A. Jilson. I thank the staff of the Geochronology Laboratory at the Geological Survey of Canada (Ottawa) for assistance in the analytical procedures. The manuscript was critically reviewed by R. Parrish and M.L. Bevier.

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A U-Pb zircon age for a Maple Creek gabbro sill, Tatamagouche Creek area, southwest Yukon Territory

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Mortensen, J.K. and Hulbert, L., 1992: A U-Pb zircon age for a Maple Creek gabbro sill, Tatamagouche Creek area, southwest Yukon Territory; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 175-179.

Abstract

We have obtained a U-Pb zircon age of 232.2 ± 1.0 Ma for a sill of Maple Creek gabbro in the Tatamagouche Creek area of southwestern Yukon Territory. This unit is a subvolcanic feeder to the Nikolai Group volcanic rocks in the Kluane Lake area. This date indicates that Nikolai Group volcanism began in the Carnian, rather than Norian stage, and spanned a period of at least 9-10 Ma.

Résumé

Une datation par U-Pb sur zircon de 232,2 ± 1,0 Ma a été obtenue pour un filon-couche du gabbro de Maple Creek dans la zone du ruisseau Tatamagouche dans le sud-ouest du Yukon. Cette unité est une cheminée hypovolcanique qui a alimenté les roches volcaniques du groupe de Nikolai dans la zone du lac Kluane. Cette datation indique que le volcanisme du groupe de Nikolai a débuté au Carnien, plutôt qu'au Norien, et qu'il a duré pendant une période d'au moins 9 à 10 millions d'années.

INTRODUCTION

Mafic and ultramafic intrusive rocks occur sporadically in a belt adjacent to western margin of the Denali Fault in southwestern Yukon (Fig. 1). In the Yukon these bodies typically occur as thick sills, and have been collectively termed the Kluane ultramafic belt (KUB). This belt extends along strike into northwestern British Columbia and eastern Alaska, and defines a semi-continuous band of intrusions with a total strike length in excess of 600 km. Mafic-ultramafic intrusions in the Kluane ultramafic belt have attracted considerable attention recently because of their Ni-Cu-PGE potential (Hulbert et al., 1988).

Although generally considered to be broadly comagmatic with proximal mafic volcanic rocks of Triassic age that are assigned to the Nikolai Group, the only direct age information for the Kluane ultramafic belt mafic and ultramafic intrusions consists of a small number of relatively imprecise K-Ar ages. These data, together with field observations, have been interpreted to indicate that the main Kluane ultramafic intrusive event occurred prior to most of the volcanism. The present study was undertaken to resolve two important questions: 1) what is the age of initial rifting and volcanism that gave rise to the Nikolai Group volcanic rocks?, and 2) was there a significant hiatus between the inferred subvolcanic Kluane mafic-ultramafic sill emplacement event and the main Nikolai extrusive event?

Campbell (1981) recognized a suite of gabbroic sills that she called the Maple Creek gabbros, and considered to be younger than the main Klunae ultramafic belt intrusions. During a regional investigation of the Kluane ultramafic belt (Hulbert, unpublished data), it was observed that the Maple Creek gabbro sills occur only within or very near the Kluane intrusions. The gabbros crosscut the earlier mafic-ultramafic bodies, but are clearly feeders for the Nikolai volcanic rocks in the area. Therefore a precise age for the Maple Creek gabbros would establish a reliable age for the onset of Nikolai volcanism, and constrain the possible relationships between the Kluane belt intrusions and the volcanism.

In this study we report a U-Pb zircon age for a sample of Maple Creek gabbro. The date indicates that Nikolai Group volcanism began somewhat earlier than was previously thought. It also constrains the age of the main phase of plutonic rocks that host Ni-Cu-PGE mineralization in this area.

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

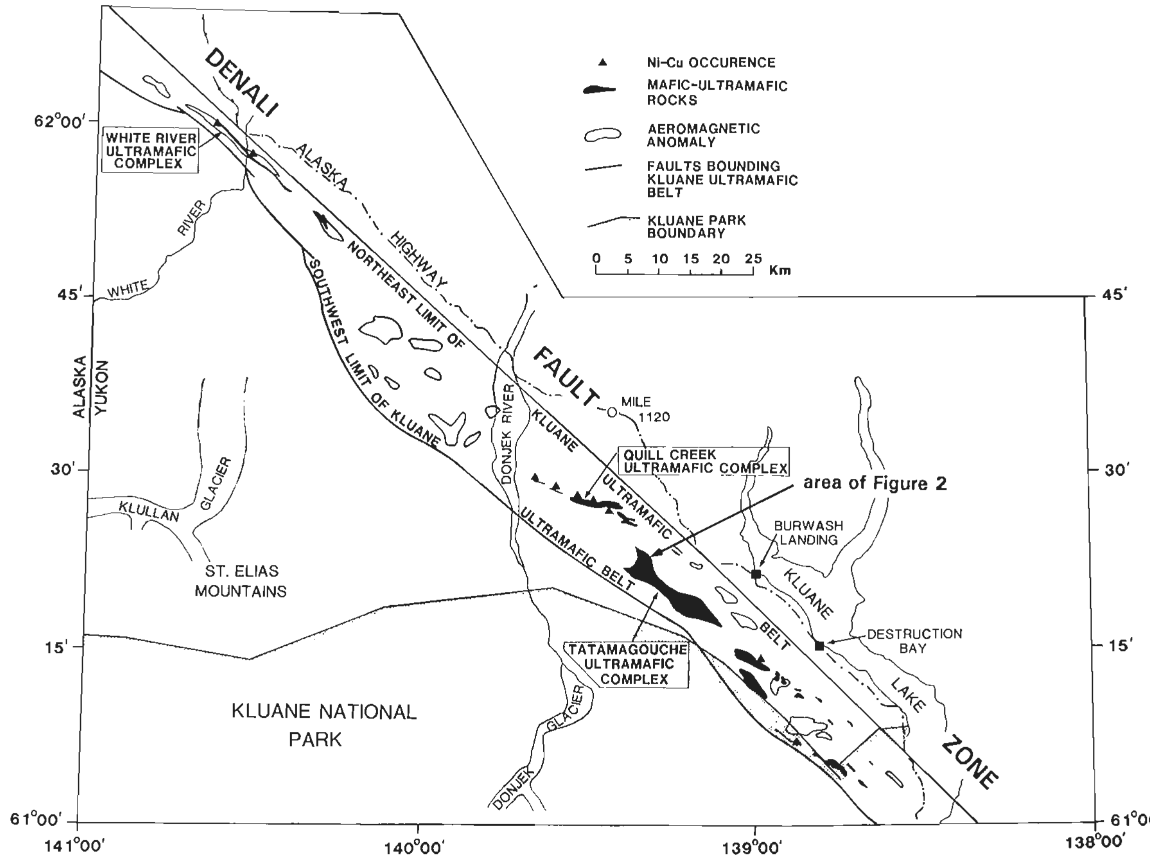

Figure 1. Map showing the regional setting and distribution of mafic and ultramafic intrusive bodies of the Kluane ultramafic belt in southwestern Yukon.

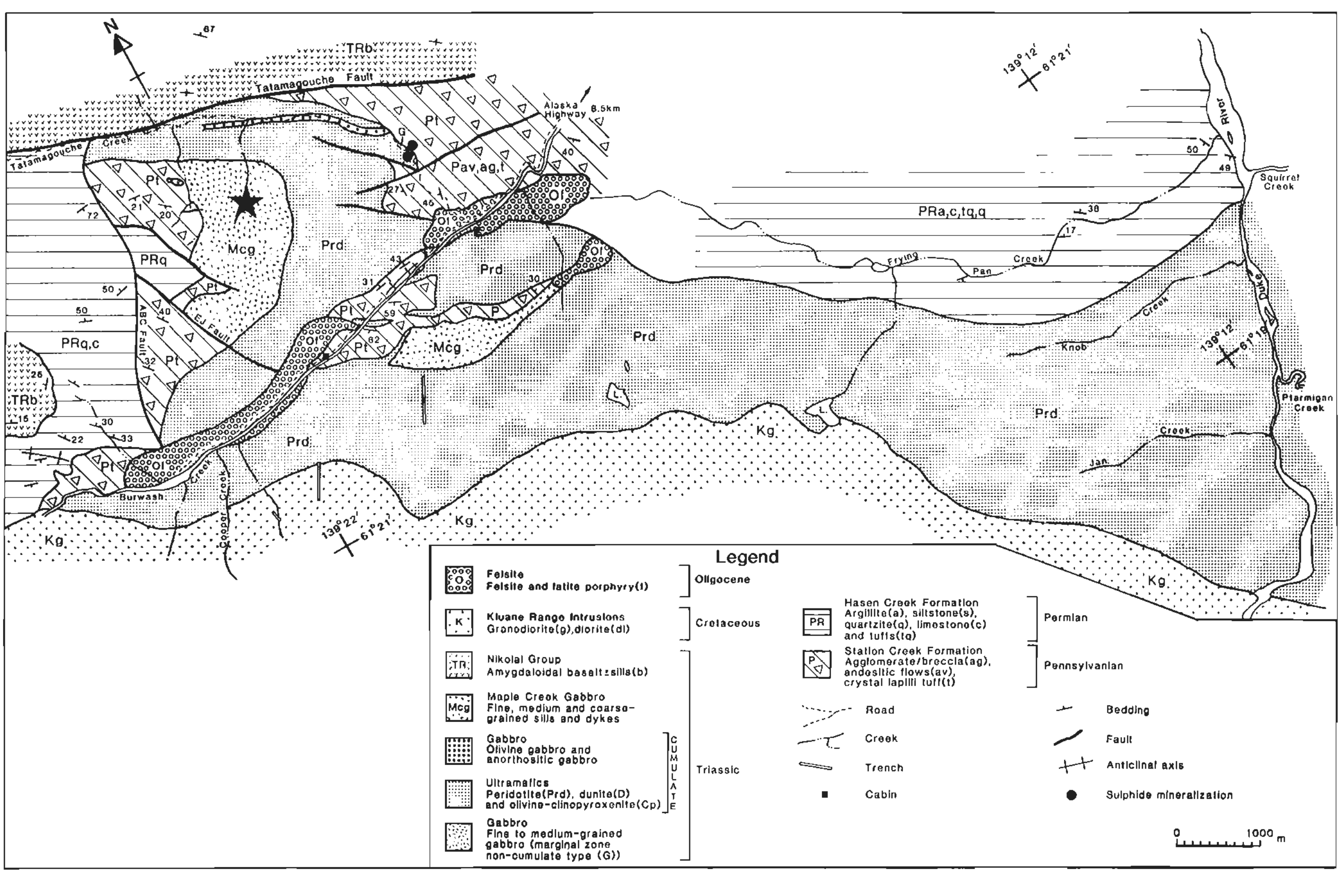

Figure 2. Detailed geology of the western part of the Tatamagouche Creek intrusive complex. Star shows sampling locality of the Maple Creek gabbro sample dated in this study.

REGIONAL SETTING AND SAMPLE DESCRIPTION

Dykes and sills of Maple Creek gabbro are common in the area, and intrude Pennsylvanian and Permian volcanic and sedimentary strata as well as intrusions belonging to the Kluane ultramafic belt (Muller, 1967; Campbell, 1981). Field, petrographic and geochemical data, and stratigraphic confinement of these bodies suggest that they are comagmatic with the tholeiitic Nikolai Group volcanic rocks in the area. Dykes of Maple Creek gabbro have been observed to intrude the base of the volcanic section and sill-out to become texturally, mineralogically and geochemically indistinguishable from other Nikolai volcanic rocks in the area.

The sample locality is approximately 17 km west of Burwash Landing (Fig. 2), and occurs on the south side of Tetagamagouche Creek, 4.2 km upstream from the confluence with Burwash Creek. The sample is from a gabbro sill emplaced into the western end of the Tatamagouche Creek mafic-ultramafic complex (Fig. 2). The sill has a maximimum strike length of 2 km, and attains an average thickness of 120 m (Fig. 3). It consists primarily of variably textured, greyish-black oxide-rich gabbro. Grain size varies considerably within the body; however most rock types present are fine- to medium-grained with a well

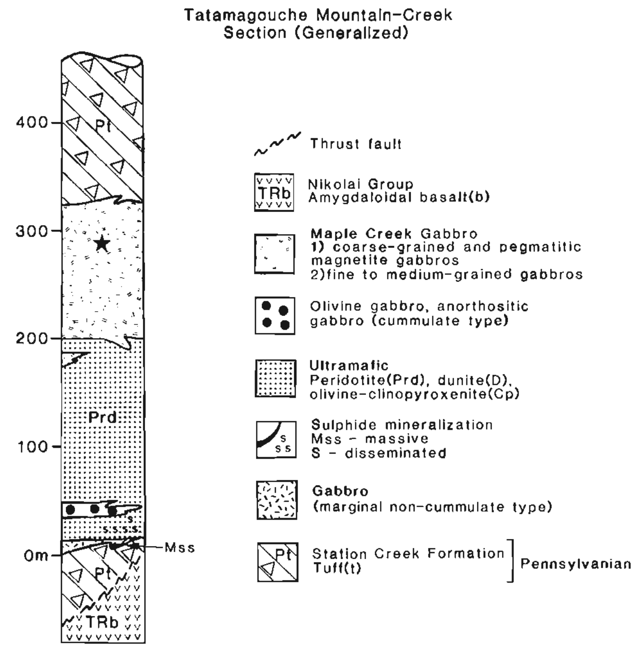

Figure 3. Schematic section through the western part of the Tatamagouche Creek intrusive complex. Star shows the relative position of the dated sample of Maple Creek gabbro.

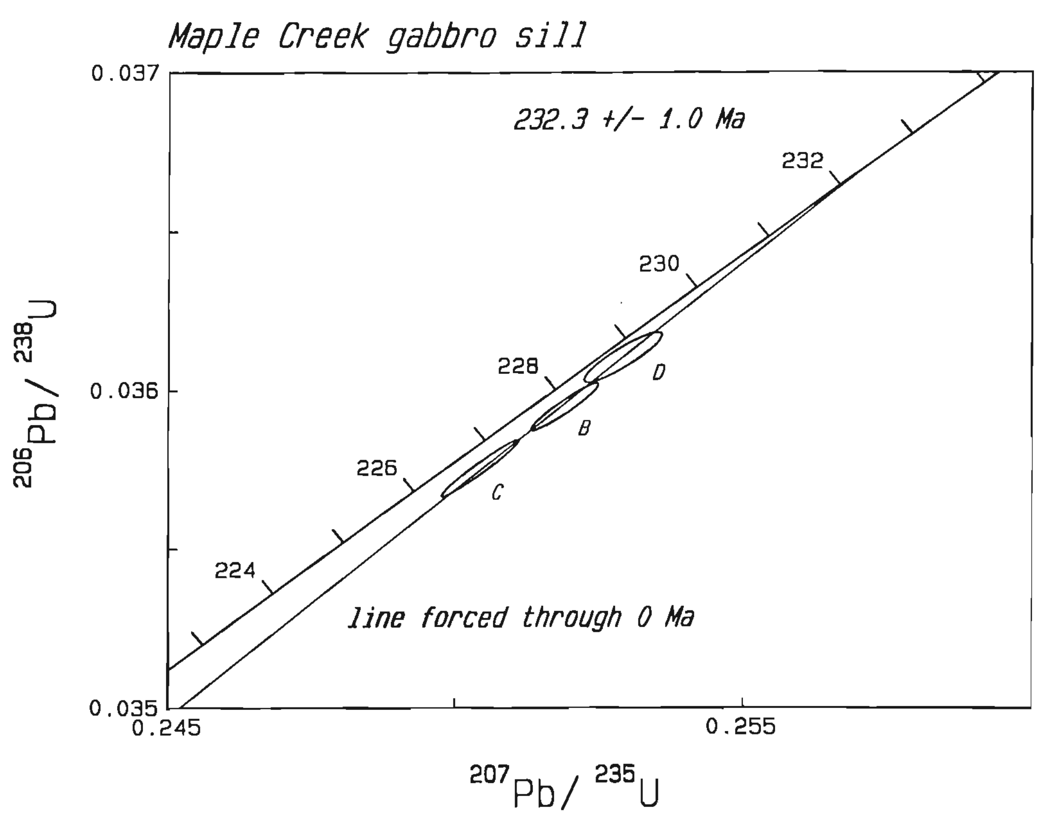

Figure 4. U-Pb concordia plot for zircons from Maple Creek gabbro sample.

Table 1. U-Pb analytical data

Fraction, Size[1]	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$ [3]	^{208}Pb[2] (%)	$^{206}Pb/^{238}U$ [4]	$^{207}Pb/^{235}U$ [4]	$^{207}Pb/^{206}Pb$ [4]	$^{207}Pb/^{206}Pb$ age[5]
Sample HDB-88-TAT-22 (61°23.5'N, 139°21.0'W)									
A N5,+62-74,clr	0.079	2196	103.2	2676	30.4	0.03610(.11)	0.2529(.14)	0.05081(.06)	232.4(2.8)
B N5,-105,clr	0.103	1774	83.1	6297	30.5	0.03595(.10)	0.2519(.12)	0.05083(.04)	233.0(1.7)
C N5,+62-74,brn	0.110	2751	126.0	14210	29.3	0.03575(.13)	0.25044(.14)	0.05080(.03)	231.9(1.4)

[1]sizes (+62-74 refers to size of zircons in microns; N5=non-magnetic cut with Frantz at 5 degrees side slope; clr=clear, brn=brown
[2]radiogenic Pb
[3]measured ratio, corrected for spike and fractionation
[4]corrected for blank Pb and U and common Pb (errors quoted are 1σ in percent)
[5]corrected for blank and common Pb (errors are 2σ in Ma)
Initial common Pb compositions from Stacey and Kramers (1975)

developed ophitic texture. Fine grained chilled margins have been observed near the top of the intrusion where the body intrudes Station Creek Formation tuffs. Irregular, very coarse grained pockets and segregations of oxide-rich gabbro are scattered throughout the sill. These pegmatitic areas are up to several metres thick and tens of metres in length. Within the coarser grained facies, augite crystals up to 2-4 cm x 0.5-1.0 cm are not uncommon. Large skeletal Fe-Ti oxide patches of similar dimensions are also present. Apart from their coarse grain size the pegmatitic material resembles the enveloping host rocks. Mineralogically the rocks consists of plagioclase (49%), augite (38%), ilmenite (6%), magnetite (2%), quartz (1%), apatite (0.5%), and pyrite, biotite, titanite, leucoxene, and zircon constitute the remainder. Compositionally the Mg# of the clinopyroxene varies from 0.64-0.67, and associated plagioclase has an average composition of An_{60}.

ANALYTICAL TECNNIQUES

Zircon was separated from a 3 kg sample of pegmatitic gabbro using conventional Wilfley table and heavy liquid methods. Analytical techniques employed in this study are summarized by Parrish et al. (1987). All zircon fractions were abraded prior to dissolution (Krogh, 1982). Errors in ages are quoted at the 2σ level, and age assignments follow Harland et al. (1989).

ANALYTICAL RESULTS

A very small amount of zircon was recovered from the sample. It consisted mainly of stubby to elongate, clear, colourless to medium brown prisms. Three fractions of zircon were selected based on colour and morphology. The analyses (Table 1, Fig. 4) range from 1.6 to 2.4% discordant (calculated for individual fractions through the origin) and define a short linear array. There is insufficient spread in the data to calculate meaningful concordia intercept ages. However the ^{207}Pb-^{206}Pb ages of the three fractions are identical within error, and, as the rocks have not been strongly altered or metamorphosed, Pb-loss is likely related to relatively recent uplift. Thus the average ^{207}Pb-^{206}Pb age of 232.2 ± 1.0 Ma is considered to be the best estimate for the crystallization age of the body.

DISCUSSION AND CONCLUSIONS

Prior to this study two age determinations were made on selected mafic-ultramafic intrusions in the Kluane belt. Campbell (1981) reported K-Ar ages of 224 ± 16 and 225 ± 14 Ma for intercumulus phlogopite separated from the Tatamagouche Creek intrusive complex and White River intrusive complex (Fig. 1), respectively. The U-Pb zircon age of 232 ± 1 Ma age determined in this study for Maple Creek gabbro from Tatamagouche Creek provides a minimum age for the Kluane ultramafic belt mafic-ultramafic sill phase in this area. The K-Ar ages only represent cooling ages through the closure temperature of the K-Ar system in phlogopite (~280°C). However if these intrusions are subvolcanic, then they should have cooled relatively rapidly, and the K-Ar ages may closely approximate the actual emplacement age. Thus intrusion of the Kluane belt mafic-ultramafic bodies may have begun in Early Triassic time.

Our date for subvolcanic Maple Creek gabbro indicates that Nikolai Group volcanism began in the Carnian, not Norian stage of the Triassic as previously believed. Conodonts from limestone beds within the Nikolai Group indicate that Nikolai volcanism continued into at least the Early Norian stage (Campbell, 1981). These age brackets indicate that volcanism spanned a period of at least 9-10 Ma.

ACKNOWLEDGMENTS

We thank the staff of the Geochronology Laboratory at the Geological Survey of Canada (Ottawa) for assistance in the analytical procedures. The manuscript was critically reviewed by O.R. Eckstrand and R. Parrish.

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A U-Pb zircon age for host rocks of a syngenetic strontium(-zinc) occurrence in the Kitsault Lake area, west-central British Columbia

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Mortensen, J.K. and Kirkham, R.V., 1992: A U-Pb zircon age for host rocks of a syngenetic strontium(-zinc) occurrence in the Kitsault Lake area, west-central British Columbia; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 181-185.

Abstract

An unusual type of syngenetic, volcanic exhalative strontium(-zinc) occurrence is present within the upper part of the Hazelton Group in the Kitsault Lake area of west-central British Columbia. A U-Pb zircon age of 193.5 ± 0.4 Ma was obtained for a felsic volcanic unit that underlies this occurrence. This dates the deposit, and provides a constraint for regional correlation within the Hazelton Group in the northern Coast Mountains of British Columbia.

Résumé

Une venue de strontium(-zinc) exhalatif volcanique syngénétique d'un type inhabituel est présente dans la partie supérieure du groupe de Hazelton dans la zone du lac Kitsault du centre-ouest de la Colombie-Britannique. On a obtenu une datation U-Pb sur zircon de 193,5 ± 0,4 Ma pour une unité volcanique felsique qui repose au-dessous de cette venue. On peut ainsi dater le gisement et affiner la corrélation régionale au sein du groupe de Hazelton dans le nord de la chaîne Côtière en Colombie-Britannique.

INTRODUCTION

The Kitsault River area (NTS 103 P/11, 12) is located on the eastern side of the Coast Mountains about 42 km southeast of Stewart, British Columbia (Fig. 1). This area has been the site of intermittent mineral exploration and limited mining since the early 1900s. The Torbrit mine (Fig. 1) produced approximately one million tonnes of silver ore between 1949 and 1957 (Campbell, 1959), and minor production also came from the Dolly Varden and North Star properties (Black, 1952). In recent years exploration for silver, gold, copper, lead and zinc has been carried out on a number of mineral occurrences in this area (Fig. 1). A variety of occurrences are present. Most are epigenetic; however some syngenetic ones have also been recognized south of Kitsault Lake at the head of Kitsault River (Fig. 1). These occurrences consist of bedded celestite and minor sphalerite within a sequence of volcanic and volcaniclastic rocks of the Hazelton Group.

Syngenetic, volcanic exhalative mineral deposits are known in several places in the Coast Mountains of British Columbia. Metamorphism, complex structural relationships, and lack of diagnostic fossils in some areas has precluded accurate dating of many of these deposits. Interest in syngenetic deposits within Hazelton Group strata in northwestern British Columbia has recently been revived, largely because of the discovery of the rich Eskay Creek deposit 120 km to the northwest (Britton et al., 1990). In this paper we report a U-Pb zircon age for the volcanic rocks that underlie the Kitsault Lake syngenetic occurrence. The age provides an important constraint both

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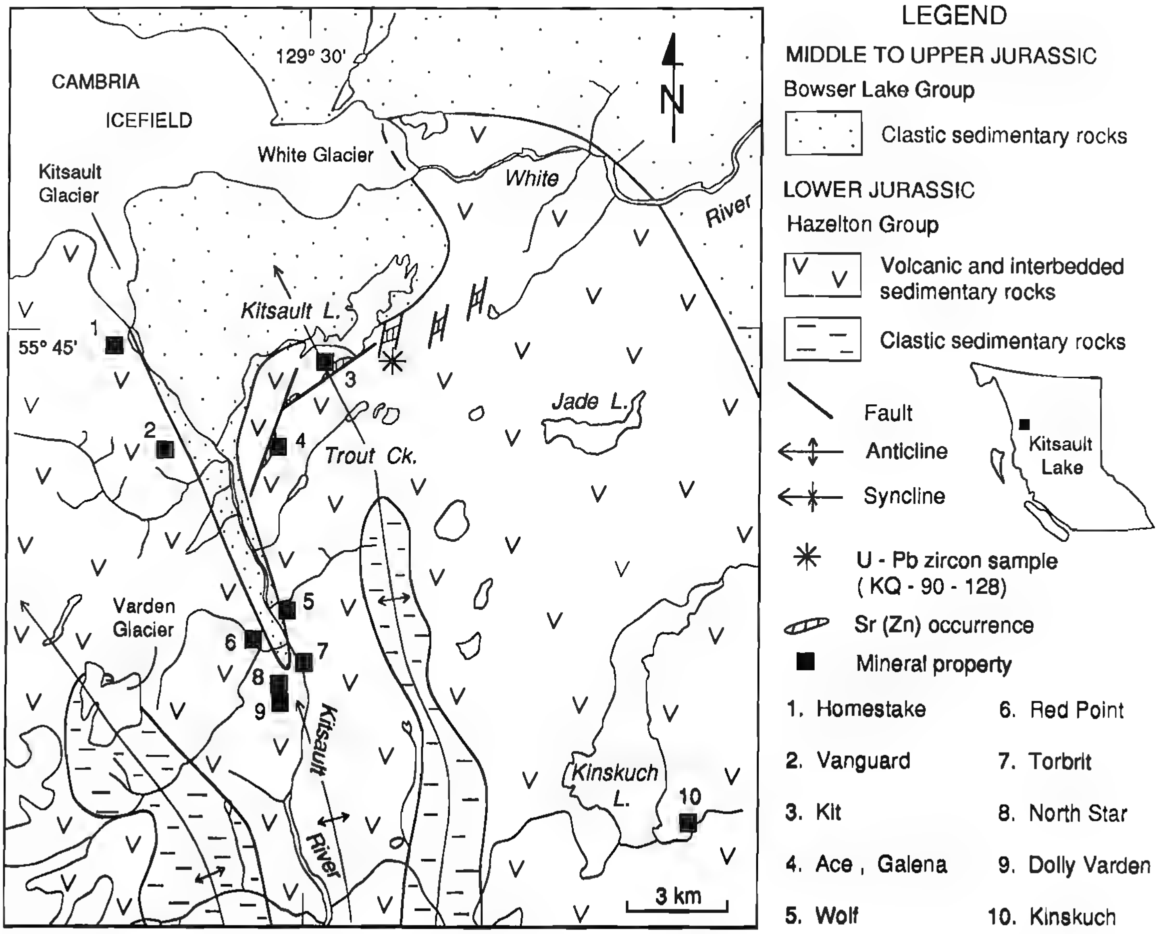

Figure 1. Generalized geology of the Kitsault Lake area (adapted from Dawson and Alldrick, 1986), showing the location of the strontium(-zinc) occurrence and the U-Pb dating sample.

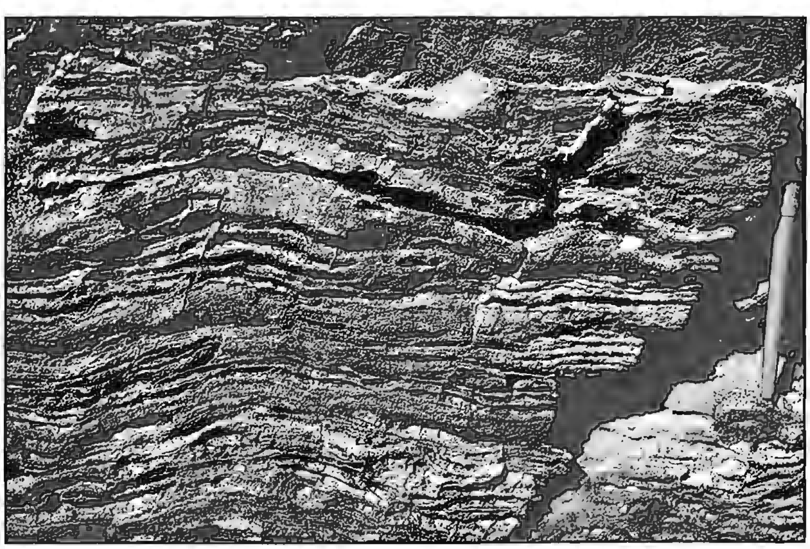

Figure 2. Typical well-bedded celestite (barite) from the Main showing, Kit property. Photograph is about 25 cm high (GSC 1991-390A).

Table 1. U-Pb analytical data

	Fraction, Size[1]	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb^{3}/^{204}Pb$	$^{208}Pb^{2}$ (%)	$^{206}Pb^{4}/^{238}U$	$^{207}Pb^{4}/^{235}U$	$^{207}Pb^{4}/^{206}Pb$	$^{207}Pb/^{206}Pb$ age[5]
	Sample KQ-90-128									
A	N2,+177	0.168	291	8.8	6453	8.7	0.03049(.09)	0.21005(.10)	0.04997(.04)	193.7(1.6)
B	N2,+177	0.110	379	11.8	2599	9.1	0.03047(.10)	0.21018(.12)	0.05002(.08)	196.0(3.8)
C	N2,+105-149,u	0.274	322	9.6	4301	8.8	0.03021(.10)	0.20828(.11)	0.05000(.05)	195.1(2.1)

[1]sizes (+62-74 refers to size of zircons in microns; N5=non-magnetic cut with Frantz at 5 degrees side slope; u = unabraded
[2]radiogenic Pb
[3]measured ratio, corrected for spike and fractionation
[4]corrected for blank Pb and U and common Pb (errors quoted are 1σ in percent)
[5]corrected for blank and common Pb (errors are 2σ in Ma)
Initial common Pb compositions from Stacey and Kramers (1975)

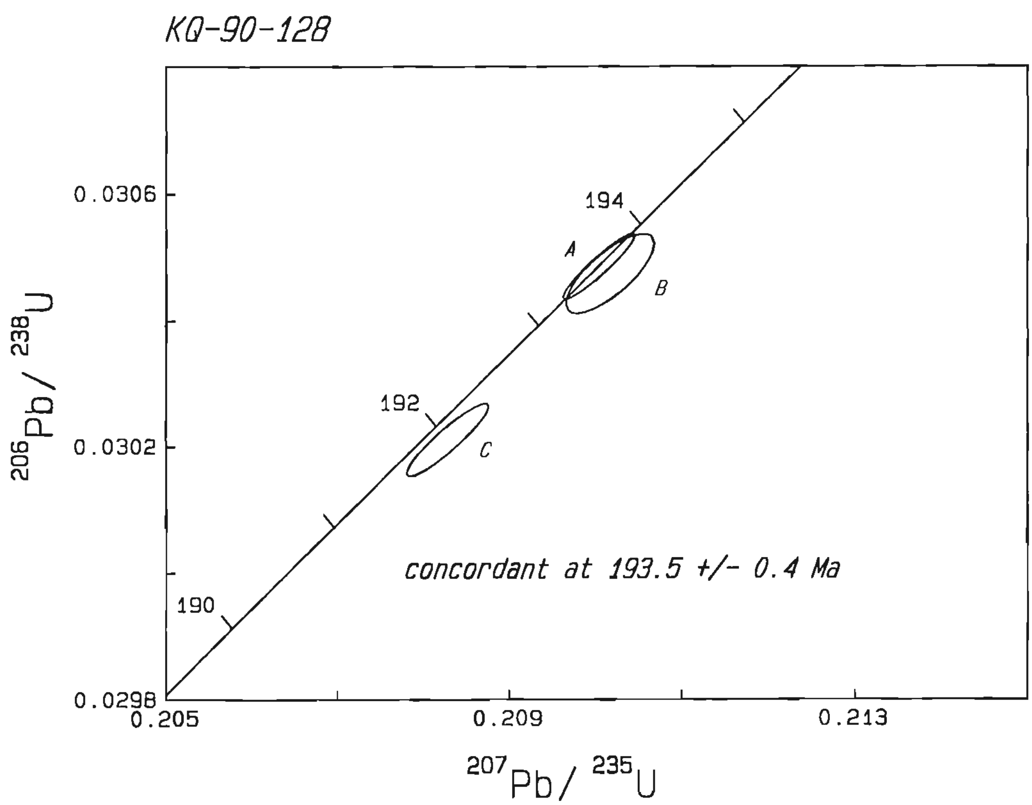

Figure 3. U-Pb concordia plot for sample KQ-90-128.

for chronostratigraphic correlation within the Hazelton Group of northwestern British Columbia, and for regional metallogenic models.

REGIONAL AND LOCAL GEOLOGY

The Kitsault Lake area (Fig. 1) is underlain by Lower Jurassic volcanic and sedimentary rocks of the Hazelton Group and Middle and Upper Jurassic clastic sedimentary rocks of the Bowser Lake Group (Alldrick et al., 1986; Dawson and Alldrick, 1986). Drab green- and maroon-grey, massive andesitic tuff-breccia predominates in the upper part of the Hazelton Group volcanic succession in the area. Some dust and ash tuff occurs along the east side of Kitsault Lake below dark graphitic sedimentary rocks of the Bowser Lake Group. In diamond-drill core thin accretionary or armoured(?) lapilli units are also present locally. Greig (1991) has documented tight, westerly vergent folds in the Bowser Lake Group in the area.

On the Kit and Ace/Galena properties south of Kitsault Lake (Fig. 1) are scattered outcrops in small fault blocks of dark, carbonaceous, pebbly mudstone (diamictite) with lithic, muscovite, and highly altered volcanic and pyrite clasts, as well as calcareous, pyritic siltstone, and buff-weathering, well-bedded celestite (Fig. 2). Minor barite, strontianite, pyrite, sphalerite, galena, arsenopyrite, and traces of orpiment occur with the celestite in gash veins. These units, although not well exposed, form a distinctive sequence that can be correlated in outcrops and diamond-drill holes from one fault block to another. They are both underlain and overlain by massive, fragmental andesitic units. Most of the sulphides are fine grained and disseminated or occur in wispy veinlets. Framboidal pyrite and soft-sediment deformation features are common in these units. Arsenopyrite characteristically occurs as scattered, minute lath-shaped crystals in dark carbonaceous siltstone or pebbly mudstone. In old (1968) drill core from the Ace/Galena area (Fig. 1), 1-2 cm thick conformable(?) layers of pale, fine grained sphalerite are also present. Carter (1969) described the detailed geology and mineral occurrences on the Ace/Galena property.

Four main lines of evidence indicate a syngenetic, volcanogenic exhalative origin for the strontium(-zinc) occurrence. These are: 1) the well-bedded nature of the celestite; 2) the distinctive bedded, carbonaceous rock units with sulphide and highly altered volcanic clasts; 3) the occurrence of thin, relatively massive bedded sphalerite units on the Ace/Galena property; and 4) the relatively massive andesitic tuff-breccia units in both the hanging wall and footwall of the strontium(-zinc) occurrence.

The U-Pb dating sample (KQ-90-128) is from a medium maroon-grey, feldspar-(quartz-)phyric lapilli tuff unit on the southeast side of "Quartz-Eye lake" (local informal name) about 400 m south of the Discovery showing (Fig. 1). This unit is probably a welded dacitic ash-flow tuff(?), and the maroon colouration is probably the result of subaerial oxidation at the time of eruption. The sampled unit occurs about 100 to 200 m(?) stratigraphically below the strontium(-zinc) occurrence.

ANALYTICAL TECHNIQUES

Zircons were separated from a 25 kg sample using conventional Wilfley table and heavy liquid methods. Analytical techniques employed in this study are summarized by Parrish et al. (1987). All analyses reported are from strongly abraded single grains. Errors in ages are quoted at the 2σ level.

ANALYTICAL RESULTS

Zircons recovered from the sample form stubby euhedral prisms that are clear and unzoned, pale pink in colour, and contain abundant clear rod- and bubble-shaped inclusions, but no visible cores. Two strongly abraded fractions and one unabraded fraction were analyzed (Table 1). The two abraded fractions are concordant with overlapping ^{206}Pb-^{238}U ages of 193.5 ± 0.4 Ma (Fig. 3), which gives the crystallization age of the sample. The unabraded fraction is slightly discordant with younger Pb-U ages, reflecting recent Pb-loss.

DISCUSSION AND CONCLUSIONS

The 193.5 ± 0.4 Ma (Pleinsbachian) age for the feldspar-phyric unit that stratigraphically underlies the Discovery showing is interpreted as the age of formation of the syngenetic strontium(-zinc) deposit in the Kitsault Lake area. Although the stratigraphic section is different and the Kitsault Lake occurrence, as explored thus far, does not contain significant gold or silver, the setting near the top of the Hazelton Group volcanic pile is analogous to that in the Eskay Creek area, about 120 km to the northwest, with its important precious and base metal deposits.

ACKNOWLEDGMENTS

R.V. Kirkham thanks the many geologists who have shared freely their geological knowledge of the Kitsault area and information on its mineral occurrences, particularly J.D. Blackwell, P. A. MacRobbie, J.R. Woodcock, D. Tupper, C.J. Greig, P.J. McGuigan, B.D. Devlin and W.N. Pearson. We also thank the staff of the Geochronology Laboratory at the Geological Survey of Canada (Ottawa) for assistance in the analytical procedures. The manuscript was improved by critical reviews by I.R. Jonasson and R. Parrish.

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U-Pb chemical procedures for titanite and allanite in the Geochronology Laboratory, Geological Survey of Canada

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Parrish, R.R., Bellerive, D., and Sullivan, R.W., 1992: U-Pb chemical procedures for titanite and allanite in the Geochronology Laboratory, Geological Survey of Canada; in Radiogenic Age and Isotopic studies: Part 5; Geological Survey of Canada, Paper 91-2, p. 187-190.

Abstract

A method of chemical separation for U and Pb from minerals with significant iron is presented. The method can be applied to titanite, allanite, rutile, apatite, feldspar, or to whole rocks. We infer that iron significantly reduces the ionization efficiency of uranium, and the method presented in this paper leads to fully satisfactory mass spectrometric analyses of uranium for these minerals.

Résumé

Une méthode de séparation chimique de U et de Pb contenus dans des minéraux ferrifères est présentée. Cette méthode peut être appliquée à la titanite, l'allanite, le rutile, l'apatite, le feldspath ou à des roches totales. Par inférence, il est conclu que le fer réduit considérablement l'ionisation de l'uranium, et que la méthode présentée dans le présent document permet des analyses de l'uranium par spectrométrie de masse entièrement satifaisantes dans le cas de ces minéraux.

INTRODUCTION

The dating of titanite and rutile is becoming commonplace in U-Pb geochronological studies and the accurate mass spectrometric measurement of U and Pb isotopes is essential in this effort. Various chemical recipes exist from a number of laboratories, but few are published. In testing and adopting procedures at the geochronology laboratory, we found that modest changes to procedures have an important impact on the quality of uranium ionization during mass spectrometry. We have also found some methods unsatisfactory when using a multicollector mass spectrometer to rapidly measure the isotopic composition of uranium; this method requires a nearly immediate, strong, high quality uranium signal. Methods currently in use in geochronological laboratories involve the separation of Pb and U isotopes using HCl, HBr, and/or HNO_3 acids and anion exchange resins.

THE PROBLEM OF LOW URANIUM IONIZATION EFFICIENCY

Zircon and monazite dissolution and purification procedures for U and Pb are well established (Krogh, 1973; Parrish et al., 1987) and owe their simplicity and success to the absence of iron and other cations in the mineral chemistry. We encountered difficulties in obtaining good uranium isotopic composition measurements in our early work on titanite; we attributed this to incomplete separation of iron from uranium when using HCl chemistry. The consequence of this is that, for a given amount of uranium, the mass spectrometric signal for a titanite or allanite sample is much weaker than for zircon, a uranium blank, or a uranium standard.

Mass spectrometric uranium and lead procedures at the geochronology laboratory utilize multicollector methods on a MAT 261 instrument (Roddick et al., 1987; Parrish et al.,

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1987), typically requiring 10-15 minutes per element per sample for zircons, blanks and standards. Inferior chemical purification of uranium required us to spend much more time per sample to get acceptable uranium isotopic measurements. The penalty for this was inconvenience, inefficiency, and loss of a significant number of samples, even though the lead analyses were good.

We tried about six variations of chemical procedures for uranium, ranging from single and double passes through anion exchange columns in HCl, acetone-HCl mixtures on anion exchange columns, HCl chemistry on cation exchange columns, and single and double HNO_3 anion exchange columns. Although these experiements were not particularly rigorous in that we did not specifically measure the elemental composition of the column washes or quantify the quality of mass spectrometric analyses, we found that only the double pass through anion exchange columns in HNO_3 for uranium derived from allanite was as satisfactory as zircon-derived uranium using the multicollector mode of mass spectrometry.

The purpose of this research note is to present our preferred method for the benefit of other laboratories. Visitors to our laboratory commonly request these details, so we felt it would be useful to publish them.

MINERAL CHEMISTRY AND ION EXCHANGE BEHAVIOR OF RELEVANT ELEMENTS

Of the minerals of interest in U-Pb geochronology and Pb isotopic studies, zircon and monazite have only trace concentrations (<0.1% by weight of the oxide) of the cations Ca, Ti, and Fe. However, there are major amounts of some of these elements in allanite (major Ca, Fe, minor to trace Ti), titanite (major Ca, Ti, minor to trace Fe), rutile (major Ti, minor Fe), apatite (major Ca, trace Ti and Fe) and feldspar (major to minor Ca, trace Ti, trace to minor Fe). When analyzed with the same procedures, the uranium ionization is much lower for allanite than titanite and rutile, suggesting that iron is the main cause of the problem.

Standard ion exchange charts (Marhol, 1982) for the elements in HCl on an anion exchange resin (Dowex 1, for instance) indicate weak to no absorption of Ca or Ti for normalities of 0-6M HCl, indicating that in an HCl wash using anion exchange media, essentially all Ti and Ca can be easily separated from U. Fe, however, closely follows the behavior of uranium, being absorbed strongly for HCl concentrations of 2-12M. Thus these two elements are difficult or impossible to separate using HCl alone with anion exchange resin. Alternatively, U and Fe have different behavior using HNO_3 with the anion exchange media, with uranium preferentially retained when the HNO_3 concentration lies between 5-10M. This difference forms the basis for separation of U from Fe. Fe can also be separated from U using a mixed HCl-acetone acid on anion exchange media (Korkisch, 1969; Strelow and van der Walt, 1981), but Fe-tailing was indicated as a problem by these authors. These observations form the background to the discussion which follows.

CHEMICAL PROCESSING

The steps in analyzing a mineral such as titanite consist of (1) mineral selection; (2) weighing, addition of isotope tracer, and dissolution; (3) chemical separation and purification of U and Pb; and (4) mass spectrometric analysis. This research note is concerned with chemical purification, particulary for uranium. For background information on procedures, the reader is referred to Parrish et al. (1987); methods presented here for titanite, allanite, and rutile supercede those previously published methods.

Mineral selection, weighing, spiking and dissolution

Silicate minerals are selected, weighed, spiked and dissolved as generally outlined in Parrish et al. (1987). Some minerals can be dissolved in Savillex capsules on a hotplate (apatite, feldspars, etc.) but in general, for titanite, allanite and rutile, we are conservative and dissolve them in the same apparatus as for zircons, in an oven for 30-60 hours at about 180°C. This generally guarantees equilibration of spike and sample, even though there may be a fluoride precipitate during cooling.

Chemical procedures

The preferred column procedure is summarized in Table 1, which is a detailed work sheet we use during U-Pb chemistry for titanite, allanite, or rutile. It uses two columns, and is completed in two days. The sample, which is dissolved in HF, is evaporated and taken up in 6.2N HCl to dissolve the sample, and repeated, if necessary, to achieve a clear solution (this repetition seems necessary most of the time). It is evaporated again and dissolved in 3.1N HCl, ready for anion exchange column chemistry.

The columns which we presently use have 0.25ml of 200-400 mesh BioRad Laboratories™ AG1X8 anion exchange resin, forming a column of settled resin about 1.5-2.0 cm high.

It is essential to prewash the anion exchange resin in the column very thoroughly with 1M HCl and H_20 to remove all Pb and U which are inherently present. We found that, although H_2O should be efficient at removing uranium at the cleaning stage, 1N HCl is much better, and 1N HCl is necessary to achieve uranium blanks which are routinely less than 1pg. Prior to adding sample to the columns, the resin in the columns is equilibrated with 3.1N HCl.

The sample is then added and washed with 3.1N HCl; U and Pb are retained on the column, but all of the Fe is also retained. Uranium (and Fe) is then collected in the eluant when 1N HBr is added, which purifies the Pb on the column. Using a separate collection beaker, Pb is then collected with 6.2N HCl, a drop of 0.3N H_3PO_4 is added and after evaporation, it is ready for mass spectrometry. Adding the 0.3N H_3PO_4 at this stage aids in loading the sample later and preventing the sample from becoming too dry.

Table 1. U-Pb chemical procedures for titanite and allanite

Acid Wash:

_____20 minutes in 1N HNO_3, low setting hot plate

Dissolution:

_____0.5 ml HF + a few drops HNO_3, into Parr bombs in oven 2 days @180°C

_____evaporate to dryness, add 0.5 ml 6N HCl, into Parr bombs in oven overnight @140°C

_____evaporate to dryness, add 0.5 ml 3.1N HCl, into Parr bombs in oven overnight @140°C

Precondition and wash:

_____0.25 ml BioRad™ AG1X8 resin to columns
_____1ml H_2O
_____1ml H_2O
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml H_2O
_____1ml 6.2N HCl
_____1ml 6.2N HCl
_____1ml 6.2N HCl
_____1ml H_2O
_____1ml 3.1HCl

Add and wash sample:

_____add sample
_____wash with 10 drops 3.1HCl drop by drop (0.1ml)
_____0.5 ml 3.1HCl
_____0.5 ml 3.1HCl
_____0.5 ml 3.1HCl

Collect U:

_____replace beakers with U beakers
_____10 drops 1N HBr, drop by drop
_____0.5 ml 1N HBr
_____0.5 ml 1N HBr
_____replace with waste beakers; begin evaporating U beakers
_____1ml 1N HBr (discard)
_____1ml 1N HBr (discard)

Collect Pb:

_____replace with clean Pb beakers
_____0.25 ml 6.2 HCl, drop by drop
_____0.5 ml 6.2 HCl
_____0.5 ml 6.2 HCl
_____add 1 drop H_3PO_4 and evaporate these to dryness

Uranium:
Evaporation...

_____evaporate U solution to dryness or a small drop
_____add 0.25 ml 8N HNO_3, dissolve covered

Uranium:
Column washing...
(using a new column)

_____add 0.50 ml of BioRad™ AG1X8 resin
_____1ml H_2O
_____1ml H_2O
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml H_2O
_____1ml 8N HNO_3
_____1ml 8N HNO_3
_____1ml 8N HNO_3

Add sample and collect U...

_____add sample to column
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____replace with U collect beakers
_____1ml H_2O
_____1ml H_2O
_____evaporate to dryness
_____redissolve in 0.25ml 8N HNO_3, covered

Prepare 2nd nitric column..
To the same column, wash with..

_____1ml H_2O
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml 1N HCl
_____1ml H_2O
_____1ml H_2O
_____1ml 8N HNO_3
_____1ml 8N HNO_3
_____1ml 8N HNO_3

Add sample and collect U...

_____add sample to column
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____0.5ml 8N HNO_3
_____replace with U collect beakers
_____1ml H_2O
_____1ml H_2O
_____1 drop H_3PO_4
_____evaporate to dryness

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A second column is prepared using 0.5 ml of the same type of resin and precleaned in a manner similar to the first column. The uranium-bearing HBr solution from the first column is evaporated to dryness, converted to a 8N HNO_3 solution, and added to the uranium column which is then extensively washed with 8N HNO_3 (Table 1). Thorough washing of the sample with HNO_3 is critical to remove traces of Fe. Uranium collection beakers are then inserted beneath the columns, and the uranium is eluted with water. This water solution (which contains the U and traces of Fe) is evaporated and dissolved again with 8N HNO_3, and the column procedure is repeated using the same resin after cleaning, usually on the next day. One drop of 0.3N H_3PO_4 is added to the resulting final uranium eluant, it is then evaporated, and ready for mass spectrometry.

DISCUSSION

The procedure presented above and detailed in Table 1 results in very good uranium and lead mass spectrometric analyses, equivalent in quality to zircon-derived uranium analyses of the same quantity. This allows us to achieve strong uranium signals quickly on our MAT 261 instrument without additional attention being paid to time-consuming preheating or, when the signal is too weak for Faraday cup analysis, to tedius measurement of isotopic composition using peak-switching on an electron multiplier.

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Precise calibration of tracer compositions for Sm-Nd isotopic studies

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Roddick, J. C., Sullivan, R. W., and Dudás, F. Ö., 1992: Precise calibration of tracer compositions for Sm-Nd isotopic studies; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 191-200.

Abstract

In Sm-Nd studies a sample's $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ are generally determined by mixing an isotopic tracer of known composition with the sample. To ensure accurate sample ratios, the tracer isotopic composition should be determined to the highest possible accuracy. However, tracer accuracy is limited by mass fractionation and lack of an internal reference normalization ratio. A modified calibration technique, which corrects for mass fractionation, is presented, as well as calibrations of Nd and Sm used to demonstrate that tracer isotopic compositions can be determined to better than 0.01%/amu. Improved blank determinations can be made using tracers calibrated in this manner. The technique may also be applied to absolute abundance determinations of many elements with 3 or more isotopes.

Résumé

Dans les études Sm-Nd, les rapports $^{147}Sm/^{144}Nd$ et $^{143}Nd/^{144}Nd$ sont généralement déterminés en mélangeant un traceur isotopique de composition connue avec l'échantillon. Pour s'assurer que les rapports des échantillons sont exacts, la composition isotopique du traceur doit être déterminée avec la plus grande précision. Cependant, l'exactitude du traceur est limitée par le fractionnement de masse et l'absence d'un rapport de normalisation de référence interne. On présente ici une technique d'étalonnage modifiée qui permet de corriger le fractionnement, ainsi que des étalonnages de Nd et de Sm qui servent à montrer que les compositions isotopiques de traceurs peuvent être déterminées à plus de 0,01%/uma de précision. Il est possible d'améliorer la détermination des blancs en utilisant des traceurs étalonnés de cette façon. Cette technique peut également servir à déterminer l'abondance absolue de nombreux éléments à trois isotopes ou plus.

INTRODUCTION

Application of the ^{147}Sm-^{143}Nd radioactive decay scheme is dependent on accurate characterization of the isotopic ratios $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ in geological materials. This usually involves isotope dilution, a technique in which an enriched tracer of known composition is added to the sample. The precise and accurate determination of tracer isotopic compositions is especially important where the $^{143}Nd/^{144}Nd$ of a sample is derived from analysis of material spiked with a Nd tracer and corrections for the tracer $^{143}Nd/^{144}Nd$ must be made. This is a typical procedure where limited sample material is available (e.g., analyses of mineral separates). Currently available rare earth element tracers are not of high enrichment in single isotopes so that corrections to sample data for other isotopes in the tracer are significant. In determining tracer composition, mass fractionation during mass spectrometric analyses introduces uncertainties. For normal Nd, this is not a problem as all isotopic ratios may be normalized to an internal reference ratio known to be constant in all samples; however, for enriched tracers, no such ratio exists. In calibrating enriched tracers, a typical procedure to minimize the uncertainties of mass fractionation of Nd or Sm involves the measurement of isotopic ratios in exactly the

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same manner as normal Nd or Sm, with the same filament loads, filament heating schedule and ion beam intensity. The observed fractionation of normal Nd or Sm from the reference normalized composition is then used to normalize the tracer compositions. Using these techniques, Wasserburg et al. (1981) estimated that their Nd tracer compositions were known to about ±0.02 to ±0.03 % per mass unit, with an absolute maximum uncertainty due to mass fractionation of 0.2 % per mass unit.

Alternate and more precise methods of correcting for mass fractionation have been utilized for more than 20 years but usually have been applied to elements with no known internal reference ratio. Dietz et al. (1962) first described a double spike technique while Dodson (1963) independently developed and detailed the mathematics of the method. The technique was first applied by Dietz et al. (1963) to determine the half lives of artificially produced Cs isotopes and by Wetherill (1964) to the isotopic composition of Mo in iron meteorites. It has been used extensively ever since, in particular for Pb evolution studies (see Hamelin et al., 1985 for references), and, more recently, to study Cd in meteorites (Rosman and De Laeter, 1988). With this method, an accurate tracer composition can be determined for elements with four or more isotopes, provided that the isotopic composition of the normal element is known. Hofmann (1971) specifically addressed the problem of correcting mass fractionation in isotopic tracers of a variety of elements. Starting from a solution developed by Krogh and Hurley (1968) to correct spiked Sr mixtures for mass fractionation he showed that the double spike technique is unnecessary for many elements with two internal reference ratios. He developed an elegant but simple method for tracer calibration using a three isotope procedure which eliminated uncertainties in tracer fractionation from single tracer isotope dilution calculations. While the method also improved the accuracy and precision of tracer composition determinations, significant uncertainties remained. The purpose of this paper is to demonstrate that, with some modification, Hofmann's three isotope procedure can be used to provide accurately normalized compositions for tracers of many elements. In particular, the calibration of a mixed Sm and Nd isotopic tracer used in the GSC Laboratory is detailed and it is shown that the precision of sample $^{143}Nd/^{144}Nd$ in spiked analyses can have accuracy and precision comparable to unspiked analyses. A companion paper (Sullivan and Roddick, 1992) describes the preparation and concentration calibration of this tracer.

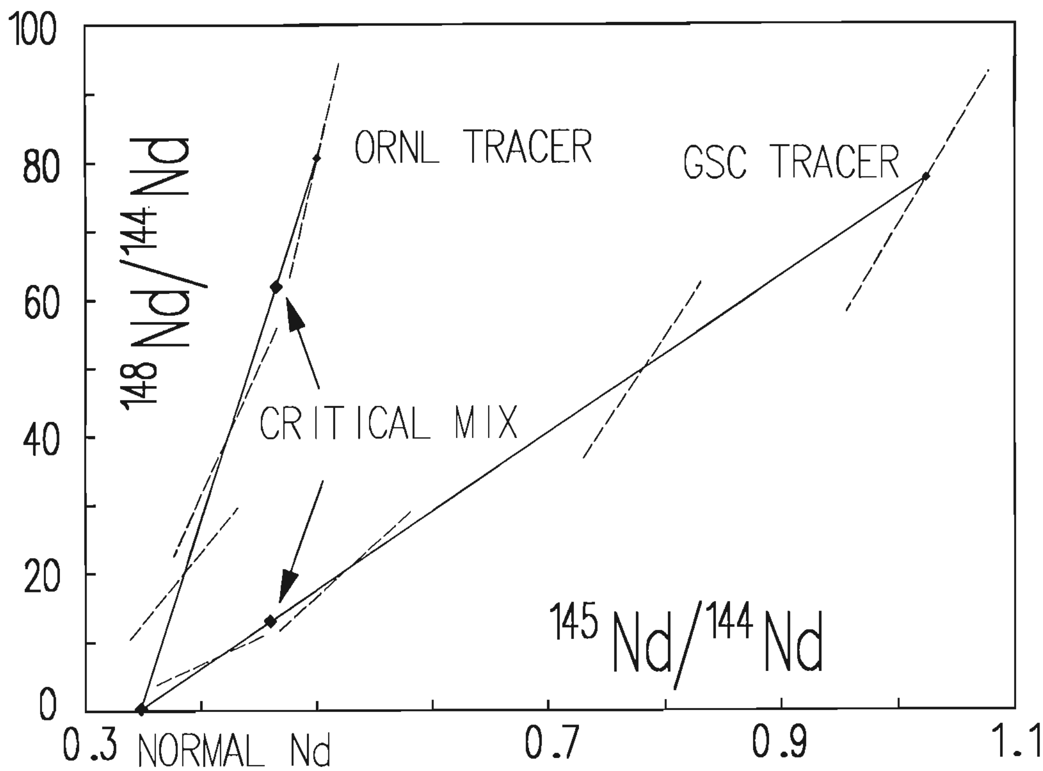

Figure 1. Mixing and isotopic mass fractionation relationships in three isotope ratio plots. Two solid lines define mixing between normal Nd and two ^{148}Nd tracers differing in $^{145}Nd/^{144}Nd$. Short dashed lines are isotopic mass fractionation trends for various normal-tracer compositions. At the critical mixtures, mixing and fractionation lines coincide. Critical mixture and normal Nd define the mixing line independent of fractionation. Tracer isotopic composition is determined by intersection of the mixing line and tracer fractionation line. A larger angle of intersection for the GSC tracer permits tracer composition to be determined more accurately than for the ORNL tracer.

PRINCIPLES OF THE TRACER CALIBRATION TECHNIQUE

Figure 1 presents the principles of the calibration technique for selected Nd isotopes. For two isotopic ratios with a common mass in the denominator, a mixing line between normal Nd and tracer Nd (i.e. ORNL, Oak Ridge National Laboratory) defines a locus of compositions with zero mass fractionation in the ratios. Superimposed on this mixing line is a series of lines showing paths of isotopic mass fractionation. With changes in composition along the mixing line, slopes of fractionation paths change such that at a particular composition, termed the critical mixture, slopes of mixing and fractionation lines coincide. This critical mixture, along with the normal Nd composition, precisely defines the mixing line and is insensitive to mass fractionation during measurement. Intersection of this mixing line with the tracer fractionation line determines tracer composition relative to the reference normal Nd composition. With the normalized composition of these two tracer ratios defined, all remaining Nd tracer ratios can be normalized. Hofmann (1971) used this technique for Sr tracer calibration and detailed the mathematical equations required to determine the critical mixture.

Figure 1 displays two mixing lines, one for a Nd tracer as supplied by ORNL and this same tracer modified by addition of a small amount of ^{145}Nd tracer. Precision in determining normalized tracer composition is dependent on the angle of intersection of the mixing line with the tracer fractionation line. In the case of the ORNL tracer, this angle is small and uncertainties in the critical mixture composition will be propagated into and magnify errors on the normalized tracer composition. Hofmann (1971) pointed out that this is not a constraint in determining fractionation of spiked samples which have compositions located between normal and tracer values as the critical mixture composition effectively defines the mixing line. This uncertainty, however, will be propagated into the tracer ^{143}Nd/^{144}Nd and, for samples spiked with this tracer, will result in greater uncertainty in a sample's measured ^{143}Nd/^{144}Nd, the ratio of interest in geological studies. To improve the precision in determining normalized tracer composition, it is necessary to increase the angle of intersection between the mixing line and the tracer fractionation line. The second mixing line in Figure 1 shows that by doubling the abundance of ^{145}Nd in the ORNL ^{148}Nd tracer a greater angle of intersection can be achieved. This addition of ^{145}Nd results in a different critical mixture with a composition closer to normal Nd and a greater extrapolation of the mixing line to the tracer composition. This extrapolation increases uncertainty in the intersection point with the tracer, but this propagation of critical mixture uncertainty is more than compensated for by improved precision resulting from the increased intersection angle with the tracer.

For a ^{145}Nd doped ^{148}Nd tracer, the use of these particular tracer ratios (148/144, 145/144) for routine isotope dilution measurements of Nd concentrations in samples is usually limited to compositions between normal Nd and the critical mixture. This is because fractionation becomes indeterminate as the critical mixture composition is approached. However, isotopic ratios not involving ^{145}Nd do not have this restriction and exhibit mixing and fractionation relationships similar to that of the original ORNL tracer (Fig. 1). Therefore ^{148}Nd/^{144}Nd and ^{146}Nd/^{144}Nd still have a critical mixture near the tracer composition and nearly the full range of compositions between the tracer and normal Nd is available for isotope dilution concentration determinations.

The following sections detail the steps involved in determining precisely normalized Nd and Sm tracer compositions. Briefly, the steps required are:

1. Selection of the major enriched tracer isotope and determination of the most appropriate ratios to precisely define the normalized tracer composition.
2. Preparation of a tracer with the required isotopic composition and measurement of its approximate composition by assuming mass fractionation factors derived from measurements of normal Nd (or Sm).
3. Preparation of appropriate critical mixtures by mixing the tracer with normal Nd (or Sm) and measurement of their isotopic compositions.
4. Using the measured compositions, solving for the intersection of the tracer fractionation line with the mixing line defined by normal Nd (or Sm) and the critical mixture.

SELECTION OF Nd AND Sm TRACER COMPOSITIONS

Figure 1 presents mixing-fractionation relationships for normal Nd mixed with both an ORNL ^{148}Nd tracer and the Geological Survey of Canada (GSC) ^{148}Nd tracer ultimately prepared for calibration and use. Selection of this Nd tracer required assessment of a number of criteria. Initially it was planned to employ ^{150}Nd tracer but at the time of preparation only a 68% enriched tracer was available from ORNL. As an alternative, ^{148}Nd has a similarly low abundance as ^{150}Nd in natural samples and was available in a more enriched form (94%). It was therefore selected as the Nd tracer. For a Sm tracer, both ^{147}Sm and ^{149}Sm have similarly low abundances in natural samples. The ^{149}Sm was selected as it was available in enriched form (98%), analyses of extra-terrestrial samples were not planned (see Wasserburg et al., 1981) and ^{144}Sm/^{147}Sm near the normal composition was required in Nd analyses so that ^{147}Sm could be used to monitor and correct for isobaric interference of ^{144}Sm on ^{144}Nd.

It is necessary, as described above, to have a tracer with three isotopes that will yield a critical mixture when mixed with normal Nd (or Sm). Both Nd and Sm have seven isotopes and 35 different three isotope sets are available. To minimize measurement errors, the enriched tracer should be one of the masses in the isotope set. A critical mixture may exist for single enriched tracers (as for the ORNL tracer, Fig. 1) or may be achieved by appropriate addition of a small quantity of another tracer isotope. In assessing this requirement for the ^{148}Nd tracer, critical mixtures were calculated for 10 possible combinations of ^{148}Nd with five of the six other Nd isotopes.

Since ^{143}Nd is variable in nature and is determined following corrections for tracer composition, it is not suitable for tracer calibration. In six cases, critical mixtures exist with the ORNL 94% enriched ^{148}Nd tracer either with the composition as supplied or with addition of other tracers. Tracer compositions which are suitable are ^{148}Nd with 145-144 (as in Fig. 1), 145-142, 145-146, 142-146, 144-142, and 144-146. The first composition produces a critical mixture without addition of another tracer but control on tracer fractionation is poor (i.e. ORNL in Fig. 1). The other compositions are even more unsuitable, with critical mixtures approximately coinciding with tracer compositions or, because fractionation trends never coincide with the normal-tracer mixing line, beyond the limits of this line. All examined tracer compositions require the addition of ^{146}Nd, ^{145}Nd or ^{144}Nd to separate the critical mixture and tracer compositions.

Following this identification of tracers from which critical mixtures might be prepared, the optimum tracer compositions were examined through numerical error analysis (Roddick, 1987). The requirement was to minimize uncertainties on the final normalized tracer compositions with the minimum addition of another enriched tracer. Realistic error estimates were assigned to hypothetical 'measured' tracer and critical mixture compositions used in determining normalized spike compositions. Tracer compositions were numerically varied to simulate the addition of enriched ^{146}Nd, ^{145}Nd or ^{144}Nd. For each new composition, a critical mixture was calculated and the intersection of each mixing line and tracer fractionation line was used to determine a normalized tracer composition.

The desire to normalize sample analyses to $^{146}Nd/^{144}Nd$ indicated that the addition of ^{146}Nd or ^{144}Nd would not be appropriate, and this was confirmed by the error propagation for determination of a sample's $^{143}Nd/^{144}Nd$. This limited the addition of another tracer to ^{145}Nd. Analysis of error propagation to the tracer composition determined that modification of the tracer $^{145}Nd/^{144}Nd$ from 0.5, as in the ORNL ^{148}Nd tracer, to 1.0 would provide a critical mixture necessary to significantly improve uncertainties in the normalized tracer composition for two three isotope sets: 148-145-144 and 148-145-142. While only one mixture is required for calibration, the option of a second set of ratios could provide an additional check. As the 148-145-144 set appeared to provide slightly better precision, calibration was pursued using this ratio set. The effect of the ^{145}Nd modified ORNL ^{148}Nd tracer on calibration is seen in Figure 1 as an increased angle of intersection of the mixing and mass fractionation lines after addition of ^{145}Nd. Addition of ^{145}Nd was limited to providing a $^{145}Nd/^{144}Nd$ of about 1.0, as a larger $^{145}Nd/^{144}Nd$ results in only a minor improvement of precision in tracer calibration. It is also feasible to measure $^{145}Nd/^{144}Nd$ in spiked samples with no loss of precision, with this limited addition of ^{145}Nd. As the enrichment of the ^{145}Nd tracer is 92% and only a small

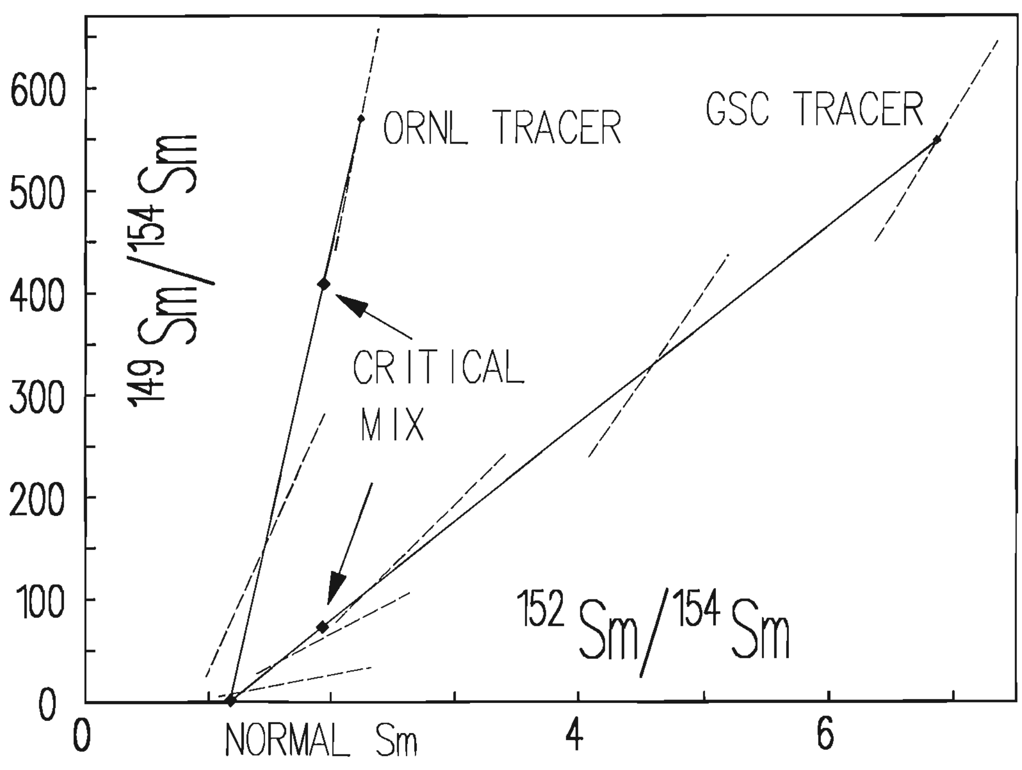

Figure 2. Mixing and fractionation relationships for normal Sm and two ^{149}Sm tracers. See Figure 1 caption for details.

amount is added to the ^{148}Nd tracer, there is very small addition of other isotopes (about 2% of 144 and 146) and no degradation of precision in normalization using the ^{144}Nd/^{146}Nd ratio as this ratio changes by 0.5% in the modified spike.

For the ^{149}Sm tracer, a similar analysis indicated 15 possible combinations of the six other Sm isotopes with ^{149}Sm but only 6 ratio sets yield possible critical mixtures. The optimum combination yielding the highest precision was the set 149-152-154. The ORNL tracer, as received, results in a critical mixture near the tracer composition (see Fig. 2) but the addition of ^{152}Sm greatly improves the control of fractionation on tracer calibration. It should be noted that this ^{149}Sm ORNL tracer, as received, is adequate for precise isotope dilution determinations provided the critical mixture is used to calibrate the tracer and the same three isotope set is used for normalization and isotope dilution calculation for samples. Isotope dilution mixtures would be confined between the limits of normal Sm and the critical mixture and, although the tracer ratios would still have a significant uncertainty due to fractionation corrections, the critical mixture effectively becomes the reference limit and fixes the position of the mixing line. To maintain the option of using other isotopic ratios for isotope dilution calculations and to demonstrate the precision of this method of calibration, enriched ^{152}Sm was added to the ^{149}Sm tracer. In this case it was necessary to increase ^{152}Sm by a factor of about three to optimize tracer calibration.

There is one additional advantage to these modified Sm and Nd tracers. Blank determinations usually involve measurement of isotopic ratios near the tracer composition because of the small quantity of normal element mixed with the tracer. For unmodified tracers of the ORNL composition the near co-linearity of mixing and fractionation lines near the tracer composition results in small errors propagating into large uncertainties in fractionation corrections and ultimately the amounts of blank determined. The modified tracers minimize these problems and the same ratios used for tracer calibrations can be used for blank measurements. The high angle of intersection of the fractionation and mixing line near the tracer composition results in precise control of mass fractionation and thus more reliable blank determinations.

Examination of the combinations of isotopes that yield possible critical mixtures for Nd and Sm indicates three relationships that may apply generally for other elements:

1. Critical mixtures are only possible for three isotope sets where the major or enriched tracer is the heaviest or lightest isotope in the set. No critical mixtures exist if the enriched isotope is of intermediate mass.
2. The ratio of tracer minor isotopes for critical mixtures (X-axis in Figs. 1 and 2) is a nearly constant value for a range of tracer compositions. Therefore, if a critical mixture is calculated for any tracer ratio set, a graphical plot such as Figures 1 or 2 can be used to predict critical mixtures for other tracer compositions.
3. If, in plots with the tracer minor isotopes as the X-axis, the mixing line has a positive slope, a critical mixture exists only if the mixing line intersects the Y-axis at a negative value. Hofmann (1971) noted this relation for a Sr mixing line.

CHOICE OF FRACTIONATION MODEL

Hofmann's original treatment (1971) employed a linear mass fractionation model for calibration of a Sr tracer. This resulted in simple algebraic expressions for calculating the critical mixture and tracer compositions. Wasserburg et al. (1981) compared and evaluated a number of fractionation models and, on the basis of their Nd data and previous work of Russell et al. (1978) on Ca mass fractionation, concluded that an exponential model may more closely approximate observed mass fractionation in Sm and Nd. They also noted that when fractionation is small (±0.1 %/amu; amu = atomic mass unit), there is little difference in the models and therefore they retained the power model of fractionation which they had used previously.

These fractionation models are basically empirical but other studies have attempted to relate mass fractionation to physical processes taking place during thermal evaporation and ionization. Moore et al. (1978) outlined a multiple filament thermal ionization model based on the vaporization of molecules and atoms while Habfast (1983) considered a similar model involving the dependence of mass fractionation on the chemical form of the sample and Rayleigh evaporation. The Habfast model could also account for the apparent exponential fractionation observed by Russell et al. (1978) for Ca isotopes with large degrees of fractionation (up to 1%/amu) by postulating a change in the molecular species evaporating from the hot filament.

Corrections for mass fractionation in this laboratory utilize a procedure written by K. Habfast and provided in Finnigan-MAT software for their thermal ionization mass spectrometers. This is a simplified single species evaporation version of the model proposed by Habfast (1983; Appendix 1). Comparison of this normalization procedure with previous empirical models yields similar results for small fractionations (±0.2 %/amu) of the ratios used in the tracer calibrations. In the present study, fractionation during analyses ranged from <0.05 %/amu to about 0.24 %/amu with typical values of 0.10 %/amu. This is the range of fractionation during analysis, and must be distinguished from the fractionation correction applied to normalize the ratios to a reference value for Nd or Sm. In the case of normal Sm, the isotopic reference value is that of Wasserburg et al. (1981) for ^{149}Sm/^{154}Sm (= 0.6075, Table 1). The resultant fractionation corrections during analysis of normal Sm were typically 0.35 to 0.25 %/amu. For Nd the reference ratio used is ^{146}Nd/^{144}Nd, with a value of 0.7219 (O'Nions et al., 1977) and fractionation corrections range from 0.18 to 0.0 %/amu.

Table 1. Isotopic composition of normal Nd and Sm measured at the Geological Survey of Canada

NEODYMIUM 142/144	143/144	145/144	146/144»	148/144	150/144	
1.141829	0.511846	0.348407	0.7219	0.241581	0.236453	LaJolla, Single
±19	±23	±13		±3	±9	collector
-	0.511862	0.348407	0.7219	0.241572	0.236439	LaJolla,
	±8	±5		±5	±8	Multicollector
1.141835	0.512147	0.348407	0.7219	0.241565	0.236433	Ames/GSC,
±9	±7	±7		±5	±10	Multicollector
1.141827	0.511862#	0.348417	0.7219	0.241578	0.236418	Wasserburg
±9	±22	±6		±6	±6	et al.(1981)
1.141844	-	0.348408	0.7219	0.241571	0.236434	UCLA DePaolo
±14		±7		±8	±16	(1981)
SAMARIUM						
144/154	147/154	148/154	149/154	150/154	152/154	
0.13513	0.65911	0.49414	0.6075»	0.32438	1.17529	Ames/GSC, Single
±2	±3	±3		±3	±2	Collector
-	0.65917	0.49426	0.6075»	0.32444	1.17538	Ames/GSC,
	±2	±1		±1	±4	Multicollector
0.13516	0.65918	0.49419»	0.60750	0.32440	1.17537	Wasserburg
±1	±2		±2	±2	±4	et al.(1981)

» Normalizing value; ± errors are 2σ of the mean in last digits for GSC data
2 analyses of LaJolla standard, other ratios from Ames metals

MEASUREMENT OF STANDARD AND TRACER ISOTOPIC COMPOSITIONS

Appropriate tracer and standard reference solutions were prepared by dissolving salts of the tracers from ORNL and ultrapure Sm and Nd metals from Ames Laboratory (Beaudry and Gschneidner, 1978) in dilute acids (Sullivan and Roddick, 1992). A mixed Sm and Nd tracer solution was prepared as a mixture improves the precision that can be attained on Sm/Nd ratios in samples. Similarly a mixed Ames standard reference solution was prepared.

As the tracer and standard solutions are mixtures of Sm and Nd, they must be chemically separated before isotopic analysis. Two distinct separation procedures were employed prior to isotopic measurements of the tracer and critical mixture compositions. The first separation method employed high performance liquid chromatography (HPLC), a technique which involves a bonded silica gel column and solutions of alpha-hydroxyisobutyric acid (HIBA) as the separation medium, as described in Sullivan (1988) and is similar to procedures of Cassidy and Chauvel (1989). The second method uses di (2-ethylhexyl) orthophosphoric acid (HDEHP) coated TFE powder columns and HCl as the separation medium, after Richard et al. (1976) and as described in Thériault (1990).

Sm and Nd isotopic compositions were measured on a Finnigan-MAT 261 mass spectrometer equipped with an adjustable seven Faraday cup multicollector assembly, of which six cups can be utilized for Sm and Nd static multicollector analyses. The MAT 261 employs a high-precision current source to calibrate relative gains of amplifiers associated with the Faraday cups. Linearity is within ±10 ppm and relative gains are stable to 100 ppm per day and 30 ppm (2 standard deviations) over the one to two hours required for an analysis.

Samples were loaded in 0.3 N H_3PO_4 on one filament of a double Re filament assembly and filament currents adjusted to 1.5-2.2 amps on the evaporation filament and 4.5-5.0 amperes on the ionization filament. These settings provide beam currents of 2-8 x 10^{-11} amps of the major isotope of Sm or Nd. Both Sm and Nd were measured as metal ions in static multicollector mode for a limited number of the isotopes of interest. Normal Nd was also measured in single collector mode as a comparison with the multicollector data. Isobaric interference of Sm on Nd was monitored using one of the cups in static mode and corrections were applied to ^{144}Nd and ^{148}Nd. Corrections were typically <0.007% in the tracer calibrations. Similarly Nd interferences on ^{144}Sm, ^{148}Sm and ^{150}Sm were monitored and only analyses free of interference as monitored by ^{146}Nd were accepted for ratios of ^{144}Sm and ^{148}Sm. In several runs from the HPLC chemistry interference as large as 0.5% was observed on ^{148}Sm because of ^{148}Nd tracer contamination. Data were collected in blocks of 10 to 12 separate integrations of 8 seconds. Baselines were monitored between data blocks at mass positions well above the Sm and Nd spectra and which were previously checked to be free of any interference. The fixed central cup of the multicollector assembly, located behind the focal plane where the other cups are positioned, had an apparent transmission bias of -60 ppm relative to the other cups. Masses collected in this cup (usually ^{145}Nd and ^{149}Sm) have been corrected for this bias.

Table 1 presents the Nd and Sm isotopic compositions measured in the Ames metals and LaJolla Nd standard (Lugmair and Carlson, 1978). Also listed are the ratios of Wasserburg et al. (1981) and DePaolo (1981). Ratios are given for both single collector and multicollector analyses. All ratios show good agreement with those of Wasserburg et al. (1981) except for ^{148}Sm/^{154}Sm in multicollector mode, which appears to be 0.02 % higher than the single collector result and the Wasserburg et al. (1981) value.

Table 2 lists two isotopic compositions of the Nd and Sm tracers. One set represents ratios measured before making corrections for fractionation using the critical mixtures. Most ratios are averages of nine (Nd) and seven (Sm) analyses of 20 data blocks each, and have been normalized to the initial composition of ^{148}Nd/^{146}Nd or ^{152}Sm/^{149}Sm in one of the analyses. Ratios for ^{150}Nd, ^{142}Nd, ^{144}Sm and ^{148}Sm represent two analyses. The normalizing composition represents the initial composition measured in an analysis and is the most

Table 2. Isotopic composition of GSC tracer Nd and Sm

NEODYMIUM 142/144	143/144	145/144	146/144	148/144	150/144	Normalization
0.95850	0.47164	1.02126	1.48080	77.094	0.57016	Initial«
±7	±3	±4	±9	±6	±7	Composition
0.95577	0.47097	1.02270	1.48498	77.529	0.57499	Reference»
±11	±4	±8	±15	±14	±16	Normal
SAMARIUM						
144/154	147/154	148/154	149/154	150/154	152/154	Normalization
0.16050	2.1315	4.7530	558.60	3.1946	6.9100	Initial«
±40	±3	±15	±13	±3	±14	Composition
0.15474	2.0782	4.6510	548.69	3.1494	6.8614	Reference»
±40	±4	±16	±15	±4	±15	Normal

« Normalized to initial (first) measured isotopic composition of ^{148}Nd/^{146}Nd (=52.06) or ^{152}Sm/^{149}Sm (=0.01237) in analyses (see text)
» Normalized to reference normal using critical mixtures in Tables 3, 4
± errors are 2σ of the mean in last digits; errors on reference normal are compounded from errors on initial composition and normalized tracer of Tables 3 and 4.

positively fractionated composition (see Appendix 1) of the analyses. These compositions are thus comparable to the initial fractionation observed in normal Nd and Sm analyses. Therefore they are expected to be fractionated from the standard reference composition by about 0.35%/amu for Sm and 0.18 %/amu for Nd, fractionations typically observed for initial isotopic ratios of normal Nd and Sm analyses. Fractionation corrections applied during analyses by normalization to the initial composition range from 0.0 to -0.21%/amu (typically <-0.1%/amu) with no detectable changes in normalized ratios during analyses. The second ratio set lists the final normalized tracer ratios after fractionation corrections were determined using critical mixtures as detailed below.

It is not possible to utilize the isotopic compositions of the Nd and Sm tracers from analyses of the tracers prior to mixing. Comparison of measured ratios of the Sm tracer before and after mixing reveals that ^{148}Sm in the mixed tracer increased by 8% after mixing. This indicates that the ^{148}Nd tracer contained a 0.14% impurity of ^{148}Sm. In a similar comparison employing ^{150}Nd, the minor isotope in the Nd tracer closest to ^{149}Sm, a significant increase was not detected.

PREPARATION AND MEASUREMENT OF CRITICAL MIXTURE COMPOSITIONS

Preparation of critical mixtures requires addition of a small amount of normal Nd or Sm to the related tracer to produce the appropriate isotopic composition. Portions of the standard reference Nd and Sm solutions were aliquoted into a teflon dropping bottle and diluted to obtain a mixed Nd-Sm normal solution which, when mixed with the tracer solution in a one-to-one ratio would produce critical mixtures for both Nd and Sm. After mixing with the tracer, chemical separation of the elements was the same as detailed previously, with two aliquots separated on HCl columns and two on HPLC columns. Blanks are insignificant, but even if contamination were present in small amounts, it would not be important to the critical mixture calibration because the isotopic compositions would remain along collinear mixing and fractionation lines (Fig. 1, 2). This is another advantage of the critical mixture calibration technique.

Measurement of the isotopic compositions of critical mixtures was similar to that for the tracers, with normalization to the first block mean (i.e. the mean initial measured composition). Tables 3 and 4 present results for Nd and Sm tracers, respectively. Two sets of analyses were performed, one in November 1988, and another in April 1989. In both cases, the analyses numbered 1 and 2 were separated using HPLC columns and analyses 3 and 4 using HCl columns. For the Nd analyses, the ^{143}Nd/^{144}Nd of GSC Ames metal as calculated from the critical mixtures is also given and further discussed below. The second set was performed for three reasons: 1) The first set of critical mixtures were not at precisely the correct values because of the uncertainty in the initial tracer composition used to calculate them, and the limitations of dispensing precise quantities of the mixtures from dropping bottles – the second set of analyses were on compositions closer to the correct values. 2) Data from the two chemical separation techniques apparently produce slightly different measured ratios. This could be caused by a different amount of initial fractionation from different chemical solutions loaded on filaments, or contamination, though neither should affect the application of the critical mixtures to tracer normalization. 3) The poor agreement of the final normalized tracer compositions for the November/88 data set as seen in Tables 3 and 4 and discussed below.

CALCULATION OF NORMALIZED TRACER COMPOSITIONS

A normalized tracer composition is determined by the intersection of the tracer fractionation line with the mixing line defined by the normal composition and a critical mixture (Fig. 1, 2). Tables 3 and 4 list the solutions of this calculation for each of 8 critical mixtures combined with ratios for the measured

Table 3. Nd isotopic compositions of critical mixtures and normalized tracer

	Measured Critical Mixtures				Normalized Tracer⁺		Ames Metal	Crit. Mix
	146/144«	148/144	145/144	143/144	148/144	145/144	$143/144_c$#	% Frac/amu
Nov/88								
1	0.8499	13.3596	0.462881	0.505785	77.5078	1.02263	0.512153	0.129
2*	0.8501	13.3635	0.462933	0.505763	77.4880	1.02256	0.512209	0.115
3	0.8489	13.3273	0.462590	0.506065	77.5184	1.02267	0.512133	0.187
4	0.8488	13.3268	0.462560	0.506089	77.5464	1.02276	0.512114	0.195
Apr/89								
1	0.8464	13.0435	0.460071	0.506101	77.5512	1.02277	0.512147	0.161
2	0.8461	13.0364	0.460029	0.506209	77.5282	1.02270	0.512159	0.180
3	0.8469	13.0690	0.460322	0.505985	77.5235	1.02268	0.512146	0.140
4	0.8465	13.0568	0.460209	0.506121	77.5290	1.02270	0.512164	0.163
MEAN (of 7):					77.529	1.02270	0.512145	(154 last 4)
2σ of mean:					±11	±4	±13	±9

⁺ Measured composition = 77.094±6, 1.02126±4 (Table 2)
« Normalized to initial ^{146}Nd/^{144}Nd measured
Unspiked value = 0.512147±7 (Table 1)
* Omitted from calculation of mean

Table 4. Sm isotopic compositions of critical mixtures and normalized tracer

	Measured Critical Mixtures 152/149«	149/154	152/154	Normalized Tracer' 149/154	152/154
Nov/88					
1	0.02542	77.678	1.97457	548.600	6.86092
2	0.02544	77.570	1.97339	548.723	6.86153
3*	0.02537	77.962	1.97791	548.027	6.85808
4	0.02537	77.935	1.97721	548.605	6.86095
Apr/89					
1	0.02572	76.167	1.95901	548.695	6.86139
2	0.02571	76.217	1.95953	548.680	6.86132
3	0.02572	76.164	1.95893	548.762	6.86173
4	0.02571	76.212	1.95941	548.774	6.86179
MEAN (of 7):				548.69	6.8614
2σ of mean:				±6	±3

' Measured composition = 558.60±13, 6.9100±14 (Table 2)
« Normalized to initial $^{152}Sm/^{149}Sm$ measured
* Omitted from calculation of mean

Nd and Sm tracer compositions given in Table 2. In carrying out this calculation, the effects of fractionation of the measured critical mixtures were minimized by normalizing, prior to calculation, the measured critical mixtures by typical fractionation corrections observed in common Sm and Nd analyses (0.35, 0.18 %/amu respectively) to a common composition for a given mixing (Nov/88, Apr/89). This helps reduce a further source of uncertainty as the prepared critical mixtures are not at the precisely correct values because of the limitations in preparation of critical mixtures as noted above. This normalization had little effect on the final normalized tracer compositions because of the co-linear relationship of fractionation and mixing at the critical mixture composition. For example, for the final normalized Sm tracer composition, the correct Sm critical mixture has a composition of 149/154 = 72.7 and 152/154 = 1.92, significantly different from the measured values (i.e. 77.7 and 1.97 for Nov/88). However the effects of this deviation from the ideal composition on the final normalized Sm tracer composition are small. Normalization of the November/88 critical mixtures by 0.35 %/amu changes normalized tracer compositions by about 0.020 %/amu. If this critical mixture normalization procedure has an uncertainty of ±0.05 %/amu, it will result in an additional uncertainty in the normalized Sm tracer composition of ±0.003 %/amu. For the Nd critical mixture, the two preparations more closely bracket the correct mixture ($^{148}Nd/^{144}Nd$ = 13.23; $^{145}Nd/^{144}Nd$ = 0.4617); arbitrary normalization of the critical mixture contributes an insignificant uncertainty (<0.001%/amu) to the calculated normalized Nd tracer composition.

The mean normalized Nd and Sm tracer compositions are derived from seven of the eight analyses in Tables 3 and 4. The rejected analyses (one each for Nd and Sm) are both from the November/88 data. The remaining three normalized tracer compositions have values similar to those of April/89, even though the critical mixture ratios vary by up to 2.5%. These normalized tracer isotopic ratios have been used to determine fractionation of the measured tracer composition (Table 2) and have been applied to the other isotopic ratios to determine the complete normalized tracer composition given in Table 2. The tracer fractionations are 0.14 %/amu for Nd and 0.35 %/amu for Sm, within the range observed in analyses of normal Nd and Sm. Final uncertainties in the normalized tracer compositions represent errors compounded from measured tracer ratios and uncertainty of mass fractionation corrections derived from the mean critical mixtures. These calculations result in tracer uncertainties typically of 0.006 %/amu on most ratios of both Nd and Sm.

The isotopic ratios of the Nd critical mixtures may be used to demonstrate the accuracy of the normalized tracer composition and the application of this tracer to conventional sample measurements. The critical mixtures in Table 3 can be considered to be a sample, in this case Ames metal, excessively spiked with Nd tracer. The ratio pairs $^{146}Nd/^{144}Nd$ and $^{148}Nd/^{144}Nd$ are used for routine fractionation corrections and isotope dilution calculations of samples spiked with this tracer. Table 3 includes these ratios, the calculated fractionation per amu and the $^{143}Nd/^{144}Nd$ measured in the critical mixtures. Using the normalized Nd tracer composition, corrections for fractionation and tracer ratios were made, and $^{143}Nd/^{144}Nd$ in the GSC Ames metal calculated (Table 3). The results range from 0.512114 to 0.512164 with a mean of 0.512145 ± 13 on seven analyses and may be compared with a value of 0.512147 ± 7 determined from unspiked analyses (Table 1). The agreement is excellent, despite significant scatter in the November/88 analyses. Measurements of LaJolla standard showed similar scatter at this time and were ultimately traced to a dirty Faraday collector for ^{143}Nd. Considering only the April/89 critical mixture data, taken after the collector was repaired, the mean of four analyses is 0.512154 ± 9, a value slightly higher than the reference value but well within the uncertainty limits. This is an excellent check on the correct normalized tracer values as the critical mixtures represent grossly over-spiked analyses ($^{148}Nd/^{144}Nd$ = 13.0) in which significant corrections must be made for fractionation and tracer Nd isotopic composition and, in particular, to $^{143}Nd/^{144}Nd$, to obtain the $^{143}Nd/^{144}Nd$ in GSC Ames metal. Thus, for routine analyses using significantly less tracer (spiked $^{148}Nd/^{144}Nd$ = 0.5 to 0.8) $^{143}Nd/^{144}Nd$ ratios may be determined with precision comparable to that of unspiked measurements.

OTHER APPLICATIONS

The isotopic tracer calibration proposed and demonstrated for Nd and Sm shows that tracer isotopic compositions can be determined to better than 0.01%/amu. This technique is applicable to many elements with at least three isotopes with constant ratios in natural materials. The great advantage of the technique is that tracers are normalized to the normal compositions of the natural element, and other isotopic ratios in the element can be precisely determined from analyses spiked with the tracer. Thus the approach is particularly applicable to elements such as Nd in which the concentration of one isotope (143) varies because of the addition of a small radiogenic component. Other radioactive decay schemes currently used in geological studies that have sufficient isotopes in the daughter element to apply this technique are K-Ca, Rb-Sr, Ce-La, Lu-Hf and Re-Os.

The technique could also be applied to determination of the absolute isotopic abundances of many multi-isotope elements. The National Bureau of Standards and Technology, Gaithersburg, Maryland, has determined the absolute composition of a number of elements (see Moore et al., 1982 for details). This requires the preparation of high purity tracers and their subsequent mixing to produce known compositions similar to normal isotopic compositions of elements. Measurement of these prepared mixtures requires careful control of fractionation, as discussed in the introduction, and is one of the uncertainties in the final absolute isotopic determinations. In applying the critical mixture procedure, the availability of high purity tracers would still be required, but they could be mixed in an appropriate manner to produce critical mixtures when mixed with the normal element. In essence, the inverse of the tracer calibrations for Nd and Sm would be applied using a tracer with known composition prepared from high purity isotopes. A mixing line between this tracer and a critical mixture would define the absolute abundance of the normal element. The main advantage would be that rigorous and time consuming control of fractionation would not be required. One disadvantage of this technique is that the linearity of the measuring system would have to be known to high precision and uncertainties propagated into the final abundance determinations. With linearities of 10 ppm claimed for present amplifiers this may not be a serious problem. Given the various current normalization procedures and reference values for Sm-Nd isotopic analyses, determination of the absolute abundance of these elements could lead to standard isotopic measurement procedures.

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APPENDIX 1

MAT Mass Fractionation Correction

The correction for mass fractionation used in this paper is a modified version of that proposed by Habfast (1983) and used in Finnigan-MAT software for their thermal ionization mass spectrometers. It is based on the principle that Langmuir-type evaporation takes place in a multiple filament thermal ionization source and the ratio r of the light (M_l) to heavy (M_h) isotopic masses evaporating is given by

$$r = \tau R \tag{1}$$

$$\text{where } \tau = \sqrt{(M_h/M_l)} \tag{2}$$

is the fractionation factor, and M_h and M_l are molecular weights of the evaporating species. R is the isotope ratio of the species in the solid phase. The resulting fractionation model is quite general but requires additional thermodynamic data and knowledge of the evaporating species. If evaporation is limited to a single chemical species and isotopic effects of dissociation are assumed negligible then a simplified fractionation model results.

Following the notation of Wasserburg et al. (1981), let R_{ij} be the isotopic ratio of isotopes i and j with masses m_i and m_j. In MAT normalization, the reference isotope for any ratio is always the heavier mass. Therefore define

$$m_{ij} = | m_i - m_j | \text{ and } m_j > m_i \tag{3}$$

with m_j the isotopic mass in the denominator of the ratio. Ratios are inverted if the lighter mass is in the denominator to always meet this requirement. Assuming a 'standard state' of isotopic reference values R^C_{ij} is known, a measured (M) set of ratios R^M_{ij} may be corrected for mass fractionation by reference to a known 'standard state' or correct (C) value for a ratio R^C_{uv} where $m_{uv} = | m_u - m_v |$ and $m_v > m_u$ as in Eq. 3. The MAT normalization relationship is

$$R^C_{ij} = R^M_{ij} / (1 + \phi (R^M_{uv} / R^C_{uv} - 1)) \tag{4}$$

$$\text{where } \phi = \frac{\tau_{ij} - 1}{\tau_{uv} - 1} \tag{5}$$

Using a linear approximation for the square root of in Eq. 2 results in

$$\phi = \frac{m_{ij}\,(m_u + S)}{m_{uv}\,(m_i + S)} \tag{6}$$

where (m_u+S) and (m_i+S) are masses of the molecular species evaporating from a hot filament prior to ionization. For the Nd and Sm data used here the metal species are assumed to evaporate and therefore S=0.

The mass fractionation per atomic mass unit, α_{ij} may be calculated as

$$\alpha_{ij} = (R^M_{ij} / R^C_{ij} - 1)/m_{ij} = \alpha_{uv}(m_u + S)/(m_i + S) \tag{7}$$

and Eq. 4 re-written as

$$R^C_{ij} = R^M_{ij} / (1 + \alpha_{ij} m_{ij}) \tag{8}$$

As in the case of different fractionation models given by Wasserburg et al. (1981), $\alpha > 0$ if light isotopes are enhanced in the evaporation-ionization process relative to the 'standard state' and $\alpha < 0$ for heavy isotope enhancement. Note that Eq. 7 is similar to that of Russell et al. (1978) but the sign of α is reversed since m_{ij} is always positive for MAT normalization.

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Preparation and concentration calibration of a mixed ^{149}Sm-^{148}Nd tracer solution used for Sm-Nd geochronology and tracer studies in the Geochronology Laboratory, Geological Survey of Canada

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Sullivan, R.W. and Roddick, J.C., 1992: Preparation and concentration calibration of a mixed ^{149}Sm-^{148}Nd tracer solution used for Sm-Nd geochronology and tracer studies in the Geochronology Laboratory, Geological Survey of Canada; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 201-205

Abstract

This paper documents the preparation and concentration calibration details for a mixed ^{149}Sm-^{148}Nd enriched isotopic tracer solution used in the GSC laboratory for isotope dilution determination of $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ in geological materials. To permit more accurate and precise determinations of its isotopic composition this mixed tracer solution was doped with small amounts of enriched ^{145}Nd and enriched ^{148}Sm. The ^{145}Nd and ^{152}Sm tracers were also used to prepare a mixed tracer for reagent blank determinations. Concentration calibration of the tracer was carried out using mixed standard reference solutions of normal Nd and Sm prepared from small pieces of pure metal supplied by Ames Laboratory. Final tracer concentrations were determined to be 6.95608 µg ^{148}Nd and 14.0700 µg ^{149}Sm. The $^{149}Sm/^{148}Nd$ of the tracer was determined with a precision of 0.005%. Measurement of Sm/Nd in the CIT standard solution agrees to within 0.025%.

Résumé

Le présent document porte sur la préparation et l'étalonnage des concentrations détaillés d'une solution mixte de traceurs isotopiques enrichie en ^{149}Sm-^{148}Nd utilisée à la CGC pour déterminer par dilution isotopique les rapports $^{147}Sm/^{144}Nd$ et $^{143}Nd/^{144}Nd$ dans des matériaux géologiques. Pour déterminer avec plus d'exactitude et plus de précision la composition isotopique, cette solution mixte de traceurs a été dopée avec de petites quantités de ^{145}Nd enrichi et de ^{148}Sm enrichi également. On utilise également les traceurs ^{145}Nd et ^{152}Sm pour préparer un traceur mixte servant à la détermination des blancs de réactifs. L'étalonnage des concentrations du traceur a été réalisé en utilisant des solutions mixtes d'étalonnage standard de Nd et Sm ordinaires préparées à partir de petits morceaux de métal pur fourni par le laboratoire Ames. Les concentrations finales des traceurs étaient 6,95608 µg pour ^{148}Nd et 14,0700 µg pour ^{149}Sm. Le rapport $^{149}Sm/^{148}Nd$ du traceur a été déterminé avec une précision de 0,005 %. La mesure de Sm/Nd dans la solution étalon CIT concorde à 0,025 % près.

INTRODUCTION

Application of the ^{147}Sm-^{143}Nd radioactive decay scheme in geochronology and isotopic tracer studies has made significant contributions to the understanding of earth evolution. The success of this method is dependent on accurate characterization of isotopic tracers used in determining $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ in geological materials. Considerable effort has been expended in developing procedures to prepare suitable standards and to calibrate tracers against these standards (Wasserburg et al., 1981).

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The characterization of any isotopic tracer requires the accurate determination of its isotopic composition and, for a solution prepared using the tracer, the concentration of the major tracer isotope. This paper documents the preparation of a mixed ^{148}Nd-^{149}Sm enriched isotopic tracer solution used in the GSC laboratory for the determination of $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ ratios in geological materials. The Nd and Sm concentrations of this tracer were calibrated using a reference solution prepared from pure metals of Nd and Sm. A companion paper (Roddick et al., 1992) describes a technique for the accurate determination of the isotopic composition of the Sm and Nd in the mixed tracer.

A mixed Sm and Nd tracer provides better precision in the determination of $^{147}Sm/^{144}Nd$ in unknowns compared to separate tracers of Sm and Nd. Whereas weights of sample and tracer are required for Sm and Nd concentrations, these weights need not enter into the calculation of ratios of Sm/Nd and therefore any weighing errors do not add uncertainty to the Sm/Nd of a sample. Similarly a mixed standard reference solution provides calibration of the tracer's $^{149}Sm/^{148}Nd$ free of weighing errors.

PREPARATION OF TRACER AND STANDARD SOLUTIONS

The selection of Sm and Nd tracers used in the GSC spike was based on availability of material and the requirements of the isotopic composition calibrations (see Roddick et al. 1992). The selected major tracers were ^{148}Nd and ^{149}Sm but with small additions of ^{145}Nd and ^{152}Sm to optimize the isotopic composition calibrations. For precise concentration calibrations ultrapure Nd and Sm metals were used as reference materials.

A final $^{149}Sm/^{148}Nd$ ratio of ca. 2 was targeted for the tracer to optimize isotope dilution calculations when mixed with rock samples with Sm/Nd, typically ca. 0.3 to ca. 0.5. Separate, more concentrated stock solutions of each were made for subsequent aliquoting and combining in the proportions required for the final mixed spike. A requirement was the ability to confidently weigh a dose of 0.1 g of spike (about 10 drops from our dispensing bottles). Consequently, nominal concentrations of 7 μg/g ^{148}Nd and 14 μg/g ^{149}Sm were targeted. At these concentrations, an aliquot of 0.1 g delivers sufficient mixed tracer for whole rock analyses of samples (0.2 g) containing typically 25 ppm Nd and 8 ppm Sm.

Tracer solutions

Four stock tracer solutions, two major tracers (^{148}Nd and ^{149}Sm) and two minor modifying tracers (^{145}Nd and ^{152}Sm) were prepared from oxide salts (eg. $^{148}Nd_2O_3$) obtained from Oak Ridge National Laboratories (ORNL), Tennessee.

Rare earth oxides are prone to absorbtion of CO_2 and H_2O when exposed to air. In order to remove these volatiles and to make the salts as stoichiometrically correct as possible it is necessary to heat the salts and all weighing containers to high temperature. A temperature of 950°C was used based on a recommendation of Cassidy (per. comm., 1986). The procedure first involved heating empty quartz weighing vials, including lids, to 950°C in a muffle furnace to bring them to constant weight and to pretreat them using similar conditions that would be used later to weight the salts. The furnace was cold when loaded then brought up to temperature. The actual time at 950°C was about 3 hours. The vials were placed on a clean flat quartz plate and each covered with a clean quartz crucible which was moved off the edge of the plate slightly so as to allow air circulation around and into the vials while in the furnace. Prior to cooling and removal from the furnace the crucibles were moved back fully onto the quartz plate so as to provide an essentially sealed atmosphere around the vials. The plate with vials and crucibles was transferred while still hot into a desiccator for final cooling under dry conditions.

All weighings were done using a Sartorius R160P analytical research balance. This is a electronic single pan balance capable of giving reproducible weights to ± 20 μg. This balance has a built-in calibrated weight which was verified to be accurate. A Mettler PE3600, high capacity pan balance was used to weigh solutions to ± 0.01 g. Ambient conditions and details of weighing were carefully controlled to eliminate or minimize possible problems associated with moisture gain or loss, static, etc. Clean cotton gloves or special anti-static clean room gloves were used when weighing or handling the vials. The process of heating, desiccation and weighing was repeated over a period of several days until the weights of the vials were constant.

Approximate amounts of each salt (in excess of what would finally be required) were weighed into their respective quartz vials and heated to 950°C in the furnace until they attained constant weight. The total weight losses of the salts varied from 3% to 21%. The salts were then dissolved with 6 N HNO_3 and quantitatively transferred with washings into their respective clean, pre-weighed FEP teflon bottles, resulting in four primary enriched tracer solutions (Table 1). Mass spectrometric analyses were performed on each primary stock solution to confirm ORNL isotopic compositions.

Table 1. Preparation details for the GSC ^{149}Sm-^{148}Nd mixed tracer

	Primary Solutions			Mixed Tracers
Isotope	Salts* mg	Enrichment % (ORNL)	Conc.† μg/g	Final conc.† μg/g
Mixed tracer components				
^{148}Nd	9.09	94.07	14.71	7.105
^{145}Nd	1.74	89.67	2.676	
^{149}Sm	21.00	97.72	35.34	14.21
^{152}Sm	2.59	99.06	4.430	
Mixed reagent blank tracer components				
^{145}Nd		89.67	2.676	0.2614
^{152}Sm		99.06	4.430	0.5018

* Enriched tracer salts were oxides, i.e. Nd_2O_3 and Sm_2O_3
† Calculated concentrations

A predetermined series of aliquoting, transfers and dilutions with 0.1 N HNO_3 were performed (all by weight) to arrive at the intermediate and final targeted concentrations and amounts of the required mixtures. Thus, the two primary Nd solutions, ^{148}Nd and ^{145}Nd, were first combined into one; then the two primary ^{149}Sm and ^{152}Sm solutions were combined and the concentration diluted. Finally these intermediate solutions were aliquoted and combined to produce the final mixed tracer solution. Details of these operations are presented below and summarized in Table 1.

A calculated amount (17.75 g), of minor tracer ^{145}Nd stock solution, was added by weight to the entire amount (500.3 g) of the major tracer ^{148}Nd stock solution, to yield a $^{145}Nd/^{144}Nd$ of about one. Similarly an aliquot (34.30 g) of the minor tracer ^{152}Sm solution was added to the entire amount (500.3 g) of the ^{149}Sm stock solution, to yield a $^{152}Sm/^{154}Sm$ of about seven. This ^{149}Sm-^{152}Sm solution was then diluted by weight (to 622.06 g) to give the required concentration so that when it was later mixed 1:1 with the ^{148}Nd-^{145}Nd solution, both would be at the correct target concentrations. Further isotopic analyses confirmed that target isotopic compositions were achieved for the intermediate Sm and Nd solutions.

Finally, weighed aliquots ($\approx$ 480 g) of the intermediate solutions of ^{148}Nd-^{145}Nd and ^{149}Sm-^{152}Sm were combined so that a final 960 g solution of the mixed Sm-Nd tracer had nominal concentrations of 7 μg/g ^{148}Nd and 14 μg/g ^{149}Sm (see Table 1 for calculated concentrations). This solution was stored in a 1 L FEP teflon bottle. After thorough mixing over ca. 10 days a 60 ml FEP teflon dispensing bottle was filled with the tracer for further characterization. This specially prepared dispensing bottle (Nalge 1600-0002), has a length of FEP teflon spaghetti tubing (Chemplast, AWG-24, 0.008″ wall thickness) inserted through a hole drilled in its cap and can deliver a 10 mg drop of tracer with good control. The protruding end of the tubing is protected by a cover made from a plastic pipet tip.

At this stage calibration of the isotopic composition of the tracer was carried out (Roddick et al., 1992). For this procedure and subsequent concentration calibrations Sm and Nd must be chemically separated before isotopic analyses. Separation procedures employed high performance liquid chromatography (HPLC), a technique which involves a bonded silica gel column and solutions of alpha-hydroxyisobutyric acid (HIBA) as the separation medium, as described in Sullivan (1988) and is similar to procedures of Cassidy and Chauvel (1989).

Primary and mixed standard reference solutions

Primary Nd and Sm reference solutions were prepared from small pieces (1 to 2 g) of pure, specially refined metal, supplied by Ames Laboratory (Beaudry and Gschneidner, 1978). The metals were received in evacuated pyrex vials, sealed by fusion. The metal pieces were bright and shiny indicating no significant oxidation had taken place since preparation by Ames Laboratory. The vials were carefully broken one at a time and the metal pieces transferred to pyrex weighing vials and weighed as quickly as possible. The resulting weights, namely 1.63256 g Sm and 1.17887 g Nd were used. These weights were both slightly less than the weights provided by the Ames Laboratory but within the weight differences expected between inter-laboratory weighings. The metals were carefully transferred into separate, clean, pre-weighed 1 L FEP teflon bottles containing 100 mL of water. High purity 8 N HNO_3 was added to the respective weighing vials and then carefully added, with swirling, to the bottles to ensure complete transfer and to dissolve the metals. The metals dissolved readily with moderate bubbling, resulting in a blue solution for Nd and pale yellow for Sm. The Nd metal dissolved about three times faster than the Sm. These solutions were then adjusted by weight to the desired concentration with 0.5 N HNO_3, to produce two primary reference solutions, thus resulting in 452.72 g of Sm solution at 3606.11 μg Sm/g and 412.10 g of Nd solution at 2860.64 μg Nd/g. About one half of these primary unmixed solutions were used to prepare mixed Sm/Nd solutions for calibrating the Sm-Nd spike as described below. About 250 mL of each of the remaining unmixed solutions were transferred by weight and stored in teflon stoppered quartz bottles for future use. The weight of these and other stored solutions are monitored on a regular basis for weight loss.

Three mixed Sm-Nd independent and different reference solutions A, B and C were prepared by removing aliquots from the primary solutions, mixing and diluting them by weight to obtain the desired concentrations and Sm/Nd ratios. A range of concentrations and Sm/Nd ratios were targeted to bracket the spike. Different details of preparation also challenged the calibration process and incorporated weighing and other errors associated with calibration procedures.

Solution A was prepared by aliquoting into a clean pre-weighed pyrex beaker, covered with parafilm, 7.37058 g of the primary Nd solution and 2.05073 g of the primary Sm solution. Clean 10 mL pyrex pipets were used to estimate and dispense the amounts required. The beakers were weighed before and after to obtain the weights. After weighing the aliquots the mixture was quantitatively transferred into a clean pre-weighed 2L FEP teflon bottle and diluted to 1000.45 g with 0.5 N HNO_3. Solution B was prepared by aliquoting by weight into a clean 60 mL pre-weighed FEP teflon bottle, 10.91283 g Nd solution and 3.82000 g of the Sm solution. In this case plastic syringes with a stainless steel needle were employed to withdraw the solutions. The syringes, fitted with Luer caps to prevent evaporation, were weighed. This mixture was then quantitatively transferred to a clean pre-weighed 500 mL FEP teflon bottle and diluted to 500.24 g. Solution C was prepared by aliquoting 3.89844 g and 1.02050 g of the Nd and Sm solutions respectively into a parafilm covered pyrex beaker. In this case an Eppendorf pipetter with clean tips was utilized to dispense the solutions into the beaker. The mixture was then quantitatively transferred into a clean pre-weighed 500 mL FEP teflon bottle and diluted to 501.35 g. The resulting concentrations and Sm/Nd ratios are presented in Table 2. These three reference solutions were used to prepare calibration mixtures to determine the concentration of the tracer solution.

Table 2. Concentration calibration of the GSC ^{149}Sm-^{148}Nd tracer

Ident.	Solution Sm µg/g	Solution Nd µg/g	Solution Sm/Nd	Tracer ^{149}Sm µg/g	Tracer ^{148}Nd µg/g	Tracer $^{149}Sm/^{148}Nd$
A1	7.39183	21.0751	0.351	14.0705	6.95666	2.02259
B1	27.5375	62.4054	0.441	14.0696	6.95594	2.02267
C1	7.34025	22.2440	0.330	14.0698	6.95563	2.02279
Mean				14.0700	6.95608	2.02269
% Standard Dev.				0.003	0.008	0.005
CIT[a]	7.5124	23.104	0.325	13.9739	6.90688	2.02319
% Deviation from GSC Mean						0.025

[a] Diluted mixed Sm-Nd solution, California Institute of Technology, (Wasserburg et. al., 1981)

Concentration calibrations were carried out subsequent to tracer composition determinations (Roddick et al., 1992) using both the above three GSC solutions and an aliquot of the California Institute of Technology (CIT) normal Sm-Nd standard (Wasserburg et al., 1981).

CONCENTRATION CALIBRATION OF TRACER SOLUTION

Calibration mixtures were prepared, in duplicate, with the spike using each of the three mixed reference solutions A, B and C. Procedure blanks were also prepared. The Sm and Nd from the calibration mixtures and blanks were separated using HPLC and mass analyzed as summarized in Roddick et al., (1992). Blanks were 80 ng Sm and 300 ng Nd and represent corrections of about 0.001% on both elements. As the results of the first three mixtures (A1, B1 and C1) were in excellent agreement (Table 2), the other three preparations were not analyzed. The weights of the spike and reference solution for the calibration mixtures were A1: 0.61794, 1.00388 g; B1: 2.23939, 1.00360 g and C1: 0.44488, 1.00182 g. Comparison of the concentrations of tracer Sm and Nd determined using the Ames standard with concentrations calculated from the dispensed tracer oxides (Table 1) indicate a difference of 1.0% (Sm) and 2.1% (Nd). These differences indicate that the actual weights of tracers in the oxide powders were less than the weighed amounts. The heating to 950°C may have been inadequate to remove all volatiles from the powders or additional non-volatile impurities may be present in the powders. The spectroscopic detection limits of the ORNL specifications could account for just over 1.0% impurities. The errors associated with weighing the salts (0.3% Sm; 0.6% Nd) may also have contributed to the differences.

As a further check on the calibration of the tracer an aliquot of CIT normal Sm-Nd standard, XXVII was used (Wasserburg et al., 1981). The bottle of CIT solution had been wrapped in plastic film and stored in a refrigerator since it was received in 1982. Upon weighing in 1987 however, the bottle weighed 0.1301 g less, which represented a 2.962% weight loss of the original solution, assumed to be attributed to slow vapour loss during storage. The original quoted CIT concentrations were thus adjusted upward for this aliquot.

The CIT Sm and Nd concentrations were too high to reliably dispense the solution so an aliquot was diluted with pure water by a factor of 20 by weight. New concentrations were calculated resulting in 7.5124 µg Sm/g and 23.104 µg Nd/g. This adjusted CIT solution was used to prepare a calibration mixture with the GSC spike. The results of isotope dilution analyses are presented in Table 2.Both concentrations of Sm and Nd in the tracer are lower than the GSC calibrations with Ames solutions by about 0.6% but the $^{149}Sm/^{148}Nd$ ratio is in good agreement with a difference of 0.025%. The reason for the concentrations differences is not known but it is probably due to incorrect assumptions concerning the weight loss that occurred during storage of the original CIT solution.

Preparation of $^{152}Sm/^{145}Nd$ spike for reagent blanks

Aliquots of the original unmixed ^{152}Sm and ^{145}Nd tracer solutions were combined by weight into a drop dispensing bottle and diluted with 0.5 N HNO_3 to nominally 50 g. This resulted in concentrations of 0.26142 µg^{145}Nd/g and 0.50184 µg^{152}Sm/g. Sm and Nd reagent blanks prepared using this dilute tracer can be mass analyzed together with one filament loading without chemical separation. Ratios $^{146}Nd/^{145}Nd$ and $^{147}Sm/^{152}Sm$ are measured during the analyses and utilized in the blank calculation equations.

RESULTS AND DISCUSSION

An enriched tracer solution was prepared and calibrated for use in the Geochronology Laboratory at The Geological Survey of Canada. Three different mixed normal Sm-Nd reference solutions were prepared from primary unmixed solutions of dissolved pieces of pure metal. Calibration mixtures were prepared and analyzed resulting in excellent agreement between the three preparations. The final concentration results for the spike are 14.0700 µg ^{149}Sm and 6.95608 µg ^{148}Nd with standard deviations of ± 0.003% and ± 0.008% respectively. The $^{149}Sm/^{148}Nd$ ratio is 2.02269 with a standard deviation of ± 0.005%. The tracer $^{149}Sm/^{148}Nd$ ratio derived from the CIT reference solution is in excellent agreement (± 0.025%) with that from the GSC solution.

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A COMPILATION OF K-Ar AGES REPORT 21

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Hunt, P.A. and Roddick, J.C., 1992: A compilation of K-Ar ages, Report 21; in Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 207-261.

Abstract

One hundred and eight-seven potassium-argon age determinations carried out by the Geological Survey of Canada are reported. Each age determination is accompanied by a description of the rock and mineral concentrate used; brief interpretative comments regarding the geological significance of each age are also provided where possible. The experimental procedures employed are described in outline. An index of all Geological Survey of Canada K-Ar age determinations published in this format has been prepared using NTS quadrangles as the primary reference.

Résumé

Les auteurs présentent 187 datations au potassium-argon effectuées par la Commission géologique du Canada. Chaque datation est accompagnée d'une description de la roche ou du concentré mineral utilisé ainsi que d'une bréve interprétation touchant l'aspect géologique. Les méthodes expérimentales qui ont servi aux datations sont aussi résumées. De plus, l'aide de quadrilatéres du SRCN, un index de toutes les datations au potassium-argon a été publié par la Commission géologique du Canada.

INTRODUCTION

This compilation of K-Ar ages determined in the Geochronological Laboratories of the Geological Survey of Canada is the latest in a series of reports, the last of which was published in 1991 (Hunt and Roddick, 1991). In this new contribution 187 determinations are reported. The format of this compilation is similar to the previous reports, with data ordered by province or territory and subdivided by map sheet number. In addition to the GSC numbers, laboratory numbers (K-Ar xxxx) are included for internal reference.

EXPERIMENTAL PROCEDURES

The data compiled here represent analyses in 1989 and 1990, as well as a number from British Columbia that were analyzed in 1972. Potassium was analyzed by atomic absorption spectrometry on duplicate dissolutions of the samples. Argon extractions were carried out using an RF vacuum furnace with a multi-sample loading system capable of holding six samples. The extraction system is on-line to a modified A.E.I. MS-10 with a 0.18 tesla magnet. An atmospheric Ar aliquot system is also incorporated to provide routine monitoring of mass spectrometer mass discrimination. Details of computer acquisition and processing of data are given in Roddick and Souther (1987). Decay constants recommended by Steiger and Jäger (1977) are used in the age calculations and errors are quoted at the 2 sigma level. Analytical details for the 1972 samples are given in Wanless et al. (1973).

The complete series of reports including the present one is as follows:

		Determinations
GSC Paper 60-17,	Report 1	59-1 to 59-98
GSC Paper 61-17,	Report 2	60-1 to 60-152
GSC Paper 62-17,	Report 3	61-1 to 61-204
GSC Paper 63-17,	Report 4	62-1 to 62-190
GSC Paper 64-17,	Report 5	63-1 to 63-184
GSC Paper 65-17,	Report 6	64-1 to 64-165
GSC Paper 66-17,	Report 7	65-1 to 65-153
GSC Paper 67-2A,	Report 8	66-1 to 66-176
GSC Paper 69-2A,	Report 9	67-1 to 67-146
GSC Paper 71-2,	Report 10	70-1 to 70-156

GSC Paper 73-2,	Report 11	72-1 to 72-163
GSC Paper 74-2,	Report 12	73-1 to 73-198
GSC Paper 77-2,	Report 13	76-1 to 76-248
GSC Paper 79-2,	Report 14	78-1 to 78-230
GSC Paper 81-2,	Report 15	80-1 to 80-208
GSC Paper 82-2,	Report 16	81-1 to 81-226
GSC Paper 87-2,	Report 17	87-1 to 87-245
GSC Paper 88-2,	Report 18	88-1 to 88-105
GSC Paper 89-2,	Report 19	89-1 to 89-135
GSC Paper 90-2,	Report 20	90-1 to 90-113
GSC Paper 91-2,	Report 21	91-1 to 91-187

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BRITISH COLUMBIA
(GSC 91-1 to GSC 91-67)

GSC 91-1 Hornblende
99.3 ± 1.9 Ma

Wt % K = 0.335
Rad. Ar = 1.329×10^{-6} cm^3/g
K-Ar 4156 % Atmos. Ar = 22.4

From a garnetiferous biotite-hornblende-chlorite schist of the metavolcanic unit (Gareau, in press, 1991).

(103 H/13) On a ridge north of the Ecstall River in the central region of the Scotia-Quaal metamorphic belt, approximately 5.5 km east of the Ecstall pluton, B.C.; 53°53.82'N, 129°31.02'W; UTM zone 9 466025E, 5971975N; sample Sg-88-73-2. Collected and interpreted by S. A. Gareau.

This sample is one of four collected on a west-east transect across the central region of the Scotia-Quaal metamorphic belt. Dating of hornblende from these samples aimed at obtaining a lower estimate of the age of the last regional metamorphic event in the area and at evaluating the thermal impact of the bounding plutons (early Late Cretaceous Ecstall to the west and Paleocene-Eocene Quottoon to the east) on the metamorphic belt.

The 99.3 ± 1.9 Ma date for this sample is interpreted as a minimum estimate of the date of cooling below the hornblende K-Ar closure temperature after the last regional metamorphic event in that region. Minor diffusion of argon may have occurred during late retrograde metamorphism postdating the last regional metamorphic event. See GSC 91-5 for references.

GSC 91-2 Hornblende
115.2 ± 1.9 Ma

Wt % K = 0.393
Rad. Ar = 1.817×10^{-6} cm^3/g
K-Ar 4157 % Atmos. Ar = 19.3

From a coarse grained rusty weathering hornblendite of the mafic-ultramafic unit (Gareau, in press, 1991).

(103 H/13) Near the western margin of the metamorphic belt on a ridge south of the Ecstall river in the central region of the Scotia-Quaal area, B.C.; 53°48.34'N, 129°31.89'W; UTM zone 9 465000E, 5961825N; sample Sg-88-147-1. Collected and interpreted by S.A. Gareau.

This 115.2 ± 1.9 Ma date is interpreted as the minimum age of cooling of the mafic-ultramafic body below closure temperature of hornblende for K-Ar after emplacement in metamorphic rocks of the Scotia-Quaal belt. Such an age is consistent with inference of Jurassic or Cretaceous age for mafic-ultramafic bodies in the Scotia-Quaal metamorphic belt based on the relatively undeformed state of the small bodies and their relationship to the surrounding country rock. The date is interpreted as a minimum estimate because the K-Ar system may have been partially reset at the time of intrusion of the Ecstall pluton to which the mafic-ultramafic body is proximal. See GSC 91-5 for references.

GSC 91-3 Hornblende
56.5 ± 1.4 Ma

Wt % K = 0.725
Rad. Ar = 1.617×10^{-6} cm^3/g
K-Ar 4158 % Atmos. Ar = 54.5

From a biotite-clinopyroxene-hornblende gneiss in the clastic metasedimentary unit (Gareau, in press, 1991).
(103 H/14) On a ridge north of the Ecstall River in the central region of the Scotia-Quaal metamorphic belt, approximately 10 km east of the Ecstall pluton and 900 m west of the Quottoon pluton, B.C.; 53°53.85'N, 129°27.12'W; UTM zone 9 470300E, 5972000N; sample Sg-88-84-2. Collected and interpreted by S.A. Gareau.

This sample is one of four collected on a west-east transect across the central region of the Scotia-Quaal metamorphic belt. Dating of hornblende from these samples aimed at obtaining a lower estimate of the age of the last regional metamorphic event in the area and at evaluating the thermal impact of the bounding plutons (early Late Cretaceous Ecstall to the west and Paleocene-Eocene Quottoon to the east) on the metamorphic belt.

The sample was collected within a 2.5 km wide zone at the eastern margin of the Scotia-Quaal belt characterized by intrusion of numerous Paleogene dykes interpreted to be associated with the Quottoon pluton. The 56.5 ± 1.4 Ma date is comparable to the age of the Quottoon pluton and the associated dykes (Gareau, 1988 and unpublished data) and is interpreted to represent resetting of the K-Ar system in the sample at the time of intrusion of the Quottoon pluton and related dykes. See GSC 91-5 for references.

GSC 91-4 Hornblende
99.6 ± 1.8 Ma

Wt % K = 0.358
Rad. Ar = 1.425×10^{-6} cm^3/g
K-Ar 4159 % Atmos. Ar = 18.0

From an amphibolite in the quartzite unit (Gareau, in press, 1991).
(103 H/13) On a ridge north of the Ecstall river in the central region of the Scotia-Quaal metamorphic belt, less than 100 m east of the Ecstall pluton, B.C.; 53°54.25'N, 129°35.68'W; UTM zone 9 460924E, 5972800N; sample Sg-88-67-2. Collected and interpreted by S.A. Gareau.

This sample is one of four collected on a west-east transect across the central region of the Scotia-Quaal metamorphic belt. Dating of hornblende from these samples aimed at obtaining a lower estimate of the age of the last regional metamorphic event in the area and at evaluating the thermal impact of the bounding plutons (early Late Cretaceous Ecstall to the west and Paleocene-Eocene Quottoon to the east) on the metamorphic belt.

The 99.6 ± 1.8 Ma date is younger than the inferred 110 Ma minimum age of the last regional metamorphic event in the area and is interpreted to result from Ar loss from hornblende at the time of intrusion of the Ecstall pluton. This interpretation is supported by evidence for a thermal aureole in metamorphic rock proximal to Ecstall pluton (Gareau, unpublished data). See GSC 91-5 for references.

GSC 91-5 Hornblende
110.0 ± 4.3 Ma

Wt % K = 0.174
Rad. Ar = 7.670×10^{-7} cm^3/g
K-Ar 4160 % Atmos. Ar = 28.7

From a hornblende gneiss in the metavolcanic unit (Gareau, in press, 1991).
(103 H/13) On a ridge north of the Ecstall river in the central region of the Scotia-Quaal metamorphic belt, approximately 3.7 km east of the Ecstall pluton, B.C.; 53°53.88'N, 129°32.14'W; UTM zone 9 4648004E, 5972100N; sample Sg-88-80-2. Collected and interpreted by S.A. Gareau.

This sample is one of four collected on a west-east transect across the central region of the Scotia-Quaal metamorphic belt. Dating of hornblende from these samples aimed at obtaining a lower estimate of the age of the last regional metamorphic event in the area and at evaluating the thermal impact of the bounding plutons (early Late Cretaceous Ecstall to the west and Paleocene-Eocene Quottoon to the east) on the metamorphic belt.

The 110 ± 4.3 Ma date is interpreted as the age of cooling of the sample below hornblende K-Ar closure temperature after the last regional metamorphic event in the area. This date is the closest available estimate of the age of the last regional metamorphic event in the Scotia-Quaal metamorphic belt.

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GSC 91-6 Biotite
64.2 ± 2.7 Ma

Wt % K = 7.350
Rad. Ar = 1.868 x 10^{-5} cm^3/g
K-Ar 4113 % Atmos. Ar = 71.7

From a porphyritic biotite granite.
(93 L/14) Sampled at the Hudson Bay Mountain stock (Kirkham and Sinclair, 1988), diamond drillhole #70-2471 at 2491 feet, B.C.; 54°49'05" N, 127°17'50" W; UTM zone 9, 609415E 6075658N; sample 70-2471-91. Collected and interpreted by R.V. Kirkham.

The Hudson Bay Mountain stock is a large blind intrusion in the core of Hudson Bay Mountain. This date is consistent with previously-reported K-Ar dates for the area (Kirkham, 1970) but at present it is uncertain whether this is an emplacement age of the intrusion or a cooling age. Further geochronological studies are underway for rocks in the area.

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GSC 91-7 Biotite
173.4 ± 2.2 Ma

Wt % K= 6.868
Rad. Ar= 4.859x10^{-5} cm^3/g
K-Ar 4149 % Atmos. Ar= 2.8

From a serpentinite shear zone.
(93 J/14) Exposed along the McLeod River, elevation 777 m, 3.1 km 273° from the campsite at Warhorse Lake, B.C.; 54°56'33.3"N, 123°09'37.1"W; UTM zone 10, 489730E, 6088200N; sample SCB88-575b. Collected and interpreted by L.C. Struik.

This sample was collected from a northwesterly trending shear zone through Triassic-Jurassic Takla and upper Paleozoic Slide Mountain group greywacke and basalt exposed along the McLeod River in northern McLeod Lake map area (93J) in central British Columbia. Rocks of these two groups are caught up in a series of northwest and north trending faults, most of which appear to have dextral strike-slip motion. The biotite (0.5-2mm) is from a narrow vertical zone through a serpentine-chlorite rock invaded with calcite-quartz veins. The pod lies within a northwest trending shear zone, and the serpentine and biotite were interpreted as a hydrothermal reaction product generated during the formation of the shear zone. The shear zone was interpreted to be part of the family of northwest trending dextral strike-slip faults of the region; although the sense of motion of the shear zone was not determined.

The K-Ar date from the biotite was intended to date the hydrothermal activity along the shear zone and therefore the minimum age of motion along the zone. Should the biotite have formed during the formation of the fault, as interpreted, then the fault would have been active during or prior to the Middle Jurassic. The Takla and Slide Mountain groups were probably thrust onto the western North American continental margin rocks at that time (Struik, 1988), and the fault with the serpentinite may have been active then.

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GSC 91-8 Biotite
127.9 ± 2.4 Ma

Wt % K= 6.647
Rad. Ar=3.424×10^{-5} cm^3/g
K-Ar 4150 % Atmos. Ar=6.4

Sample SCB-85-376B. For location, interpretation and reference see GSC 91-9.

GSC 91-9 Muscovite
71.5 ± 2.1 Ma

Wt % K= 5.120
Rad. Ar= 1.452×10^{-5} cm^3/g
K-Ar 4151 % Atmos. Ar= 10.0

From a garnet-biotite-muscovite-feldspar-quartz schist.

(83 D/5) Outcrop is at elevation of 7000' (2133m), 2.85 km to 330° from a small tarn at the head of Manteau Creek, and 3.15 km to 116° from the northeast end of lake at head of Angus Horne Creek, B.C.; 52°25'48.8"N, 119°33'19"W; UTM zone 11, 326275E, 5811750N; sample SCB85-376b. Collected and interpreted by L.C. Struik.

These two determinations (GSC 91-8,9) are from a single schist sample from low in the Hadrynian Kaza Group in the southern Cariboo Mountains of British Columbia. The regional grade of metamorphism is staurolite. The biotite and muscovite of this sample were analyzed to provide a control on the cooling history of the area in comparison to the Eocene cooling history of the southern Omineca Belt (Parrish et al., 1988, Sevigny et al., 1990), the mid Cretaceous thermal event of the Monashee Mountains (Sevigny et al., 1990), and the Jurassic cooling history of the low-grade rocks of the central Omineca Belt. The sample is from the rocks in the hanging wall of the North Thompson River extension fault.

The disparity in ages between the biotite and muscovite is opposite to that expected from a simple cooling history. The Cretaceous ages of the two determinations are consistent with the hypothesis of cooling of the host rock below the 250-300°C blocking temperature of biotite before the Eocene cooling of the Shuswap Metamorphic Complex to the south and after the Jurassic cooling of the central Omineca Belt. The older age of the biotite is probably due to excess argon and the muscovite age of ca. 72 Ma is probably a better estimate of the cooling age for these metamorphic rocks.

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GSC 91-10 Hornblende
93.1 ± 1.4 Ma

Wt % K= 0.684
Rad. Ar= 2.540×10^{-6} cm^3/g
K-Ar 4152 % Atmos. Ar= 8.2

From an amphibolite.

(93 A/8) Outcrop on Mount Buchanan along the north shore of Azure Lake, elevation 7100' (2164m), 0.66km to 16° from Buchanan Peak, B.C.; 52°26'19.5"N, 120°04'31.5"W; UTM zone 10, 698790E, 5813650N; sample SCB85-512. Collected and interpreted by L.C. Struik.

This sample was collected from an amphibolite layer that is part of a series of calc-silicate, marble, gneiss, schist and amphibolite of the possibly Paleozoic Snowshoe Group along the ridge north of Azure Lake, at the north end of the Shuswap Metamorphic Complex. The rocks are in regional sillimanite grade of metamorphism. The amphibolite consists primarily of hornblende and plagioclase, and locally some biotite. The age of the hornblende was determined to compare its K-Ar retention age with that of kyanite grade hornblende to the north at Quesnel Lake, where the northern most sillimanite grade rocks of the Shuswap Complex are exposed (sample GSC 91-11). The Cretaceous age matches that of some granitic bodies to the north and the lower intercepts of U-Pb ages for the Quesnel Lake Gneiss of the region (Getsinger, 1985; Mortenson et al., 1987). This may indicate a regional thermal and uplift event in the mid Cretaceous as shown for the Monashee Mountains to the south (Sevigny et al., 1990).

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GSC 91-11 Hornblende
147.7 ± 3.1 Ma

Wt % K= 0.346
Rad. Ar=2.070×10^{-6} cm^3/g
K-Ar 4153 % Atmos. Ar= 11.9

From an amphibolite.
(93 A/11) Outcrop on Mount Stevenson, B.C., elevation 7230' (2516m); 52°40'23.9"N, 121°08'23.4"W; UTM zone 10, 625775E, 5837340N; sample SCB82-1018. Collected and interpreted by L.C. Struik.

This amphibolite comes from a sequence of quartzo-feldspathic paragneiss, and schist of the probably Precambrian part of the Snowshoe Group in regional kyanite grade of metamorphism. The amphibolite consists of hornblende with some biotite. The unit is approximately 3 km north of schists with sillimanite and there appears to be no structural break between the two metamorphic zones (Getsinger, 1985). The Late Jurassic age of the K-Ar retention within the hornblende is older than the mid Cretaceous Rb-Sr whole rock age from some of the local pegmatite granitic bodies (Getsinger, 1985), which Struik (1988) argued may signal the peak of sillimanite grade metamorphism in the region.

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1988: Structural geology of the Cariboo gold belt, central British Columbia; Geological Survey of Canada, Memoir 421, 100 p.

GSC 91-12 Whole Rock
152.9 ± 2.6 Ma

Wt % K= 3.278
Rad. Ar= 2.032×10^{-5} cm^3/g
K-Ar 4010 % Atmos. Ar= 69.0

From a rusty weathering, weakly crenulated, medium grey phyllite of Upper Miette Group.
(93 H/15) Exposed on ridge 1.6 km east of Holy Cross Mountain, B.C.; 53°46'41"N, 120°46'25"W; UTM zone 10, 646700E, 5960900N; sample MBB-82-111; 100 m stratigraphically below Cunningham Formation, illite crystallinity 1.5. Collected and interpreted by M. McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-13 Whole Rock
226.0 ± 3.0 Ma

Wt % K= 2.645
Rad. Ar= 2.475×10^{-5} cm^3/g
K-Ar 4011 % Atmos. Ar= 45.2

From a rusty weathering, cleaved, grey silty argillite.
(93 H/16) From the Middle Miette Group exposed on ridge 4.4 km north-northwest of Wallop Mountain, B.C.; 53°49'43"N, 120°29'37"W; UTM zone 10, 664950E, 5967150N; sample MBB-81-89; illite crystallinity 3.2-3.5. Collected and interpreted by M. McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-14 Whole Rock
140.9 ± 2.4 Ma

Wt % K= 3.109
Rad. Ar= 1.770 x 10^{-5} cm^3/g
K-Ar 4012 % Atmos. Ar= 37.4

From a weakly crenulated, grey, laminated phyllite.
(93 H/8) From the Upper Miette Group exposed on ridge southeast of East Twin Creek, B.C.; 53°28'30"N, 120°15'06"W; UTM zone 10, 682400E, 5928400N; sample A-83-5; illite crystallinity 2.6. Collected by A. Carey, interpreted by M. McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-15 Whole Rock
178.3 ± 4.6 Ma

Wt % K= 3.418
Rad. Ar= 2.490 x 10^{-5} cm^3/g
K-Ar 4129 % Atmos. Ar= 84.5

From a moderately to poorly cleaved, grey, silty argillite.
(93 H/15) From the Middle Miette Group exposed 3.6 km northwest of Wallop Mountain, B.C.; 53°49'08"N, 120°30'04"W; UTM zone 10, 664500E, 5966050N; sample MBB-81-66; illite crystallinity 3.0-3.2. Collected and interpreted by M. McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-16 Whole Rock
261.4 ± 5.7 Ma

Wt % K= 2.748
Rad. Ar= 3.004 x 10^{-5} cm^3/g
K-Ar 4130 % Atmos. Ar= 17.9

From a rusty weathering, cleaved, laminated, phyllitic, blue-grey argillite.
(93H/16) From the Upper Miette Group approximately 150 m below Cambrian Gog Group, in footwall of Back Ranges fault exposed on ridge between Bastille and Idol creeks, B.C.; 53°49'09"N, 120°14'51"W; UTM zone 10, 681200E, 5966700N; sample MBB-81-90; illite crystallinity 3.2. Collected and interpreted by M.McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-17 Whole Rock
182.0 ± 4.8 Ma

Wt % K= 4.801
Rad. Ar= 3.574 x 10^{-5} cm^3/g
K-Ar 4131 % Atmos. Ar= 1.9

From a rusty weathering, cleaved, blue-grey, silty, phyllitic argillite with some larger (0.5 mm diameter) muscovite flakes on the cleavage faces.
(93 H/15) From the Cambrian McNaugton Formation exposed on ridge 6 km east of Holy Cross Mountain, B.C.; 53°46'17"N, 120°42'29"W; UTM zone 10, 651050E, 5960300N; sample MBB-82-117; 500 m from top of formation, illite crystallinity 3.8. Collected and interpreted by M.McMechan.

See GSC 91-18 for interpretation and references.

GSC 91-18 Whole Rock
231.9 ± 3.7 Ma

Wt % K= 6.335
Rad. Ar= 6.089 x 10^{-5} cm^3/g
K-Ar 4132 % Atmos. Ar= 1.1

From a phyllite with a fine crenulation cleavage.
(93 H/9) From the lower part of the Middle Miette exposed on the ridge southeast of East Twin Creek, B.C.; 53°30'50"N, 120°01'49"W; UTM zone 10, 696900E, 5933350N; sample A-83-157; illite crystallinity 2.8. Collected by A. Carey and interpreted by M. McMechan.

These seven samples (GSC 91-12 to 18) were submitted in an attempt to determine the age of low grade metamorphism in the western Rocky Mountains near 54°N. Structural relationships in this area suggest that the main cleavage predates movement on thrust faults of probable mid to late Cretaceous age. Samples with a range of illite crystallinities[1] (3.8-1.5) were selected in the hopes that the whole rock K-Ar ages would converge towards a lower limit that would define the age of regional metamorphism and penetrative deformation. Similar

1 Illite crystallinity was measured on the 1µm size fraction of clay minerals. Illite crystallinity values less than approximately 4 indicate the onset of low grade metamorphism. Lower values indicate a higher degree of low grade metamorphism.

structural relationships occur in the Rocky Mountains 250 km to the north (McMechan, 1987). There, K-Ar dating of greenschist facies schists from the Upper Proterozoic Misinchinka Group exposed along the John Hart Highway near the Rocky Mountain Trench have yielded latest Jurassic to Lower Cretaceous ages (143-116 Ma; Muller, 1963; Dodds, 1979). In the present study the two samples with illite crystallinities of 2.6 or less (GSC 91-12,14) have much lower K-Ar whole rock ages than those with higher crystallinities. With the other samples there was no obvious correlation between illite crystallinity and whole rock age suggesting there may be a problem with excess Ar and/or the presence of unrecrystallized detrital muscovite. Thin section examination showed sample GSC 91-17 contained coarser, oriented muscovite than the other samples and was the only one with some bent muscovite flakes. These observations are consistent with its higher illite crystallinity (3.8), yet it produced a lower whole rock age (182 Ma) than three of the samples (GSC 91-13,16,18) with lower crystallinities and recrystallized-looking muscovite textures. In the two samples (GSC 91-12,14) with younger whole rock ages (152, 141 Ma) over 95% of the muscovite is oriented in the cleavage, the rest has random orientation. Sample GSC 91-16 had similar muscovite textures with 90-95% oriented in the cleavage, yet the whole rock age was 261 Ma. These comparisons support the contention that the older ages are largely 'memory' (excess Ar) problems that disappear with slightly higher levels of metamorphism.

This set of K-Ar whole rock data shows that metamorphism and associated penetrative deformation of the western Rocky Mountains near 54°N is no older than earliest Cretaceous-latest Jurassic. However, because of potential excess Ar problems new samples will need to be collected and run to determine whether this represents the actual age of metamorphism or not.

REFERENCES

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GSC 91-19 Hornblende
148.9 ± 6.3 Ma

Wt. % K= 0.479
Rad. Ar= 2.890 x 10^{-6} cm^3/g
K-Ar 4217 % Atmos. Ar= 37.0

From a foliated hornblende quartz diorite.

(93 C/4) Outcrop on SE trending spur, elevation 6350', 3.2 km at 192° from Glacier Mountain, B.C.; 52°11'6"N, 125°54'4"W; UTM zone 10u, 301675E, 5785375N; sample V89-62-4. Collected and interpreted by P. van der Heyden.

This sample was collected from the western margin of the Atnarko Complex (van der Heyden, 1990, 1991), a metamorphic tectonite complex in the eastern margin of the Coast Belt. The quartz diorite forms a small, sheet-like, late-kinematic intrusion in strongly sheared, greenschist facies metavolcanic rocks with an inferred Early Jurassic protolith age. U-Pb zircon geochronometry of identical material, collected approximately 22 km to the south (sample V89-112-3, van der Heyden 1991), indicates early Late Jurassic emplacement and deformation ages (156 ± 2 Ma) for similar intrusions. At least one of these intrusions is nonconformably overlain by Hauterivian volcanic and sedimentary rocks, indicating cooling and uplift of this part of the Atnarko Complex by Early Cretaceous time. The 148.9 Ma hornblende date for this sample confirms these inferences: the western margin of the Atnarko Complex was cooled, uplifted, and eroded in Late Jurassic-Early Cretaceous time, following Middle-Late Jurassic plutonism, metamorphism and deformation.

REFERENCES

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GSC 91-20 Hornblende
77.4 ± 1.0 Ma

Wt % K= 0.597
Rad. Ar= 1.834×10^{-6} cm^3/g
K-Ar 4142 % Atmos. Ar= 16.4

(92 N/10) See GSC 91-21 for this samples biotite pair, location, rock type, and interpretation.

GSC 91-21 Biotite
57.2 ± 0.9 Ma

Wt % K= 7.351
Rad. Ar= 1.660×10^{-5} cm^3/g
K-Ar 4143 % Atmos. Ar= 5.1

From a tonalite.
(92 N/10) Outcrop at 7700 ft, 7.5 km east southeast of Rusty Peak, Niut Range, Coast Mountains, B.C.; 51°30.5'N, 124°50.5'W; UTM zone 10, 372250E, 5707800N; sample 88WV-100. Collected and interpreted by M.E. Rusmore and G.J. Woodsworth.

The sample is a medium grained, moderately foliated hornblende> biotite tonalite containing conspicuous primary edidote from an orthogneiss body in an extensive fold-and-thrust belt near the eastern margin of the Coast Plutonic Complex (Rusmore and Woodsworth, in press). The highly discordant K-Ar dates from this sample reflect cooling below the hornblende and biotite blocking temperatures, rather than the age of deformation of the tonalite. See GSC 91-22 for references.

GSC 91-22 Muscovite
84.3 ± 1.5 Ma

Wt % K= 8.366
Rad. Ar= 2.807×10^{-5} cm^3/g
K-Ar 4211 % Atmos. Ar= 3.5

From a vein.
(92 N/10) Outcrop at 7700 ft, 6 km southwest of Razorback Mt., Niut Range, Coast Mountains, B.C.; 51°34.4'N, 124°43.1W; UTM zone 10, 380900E, 5714800N; sample 87WV-190. Collected and interpreted by M.E. Rusmore and G.J. Woodsworth.

The sample consists of coarse grained muscovite extracted from a vein-filled boudin neck. The boudinage reflects deformation during an extensive fold-and-thrust event along the east side of the Coast Plutonic Complex (Rusmore and Woodsworth, in press). The muscovite is undeformed and grew after cessation of boudin formation; the deformation is thus older than 84.3 ± 1.5 Ma.

REFERENCE

Rusmore, M.E. and Woodsworth, G.J.
in press: The Coast Plutonic Complex: a mid-Cretaceous contractional orogen; Geology, in press.

GSC 91-23 Hornblende
79.5 ± 2.6 Ma

Wt % K=0.493
Rad. Ar=1.557×10^{-6} cm^3/g
K-Ar 4161 % Atmos. Ar=17.1

From a foliated, protomylonitic leucocratic hornblende-biotite quartz monzonite.
(92 H/6) From a ridge outcrop, elevation 2600 feet, northeast of Hope, B.C.; 49°27.96'N, 121°23.32'W; UTM zone 10U, 616750E, 5480300N; sample MC-89-119. Collected by R. Parrish and M. Coleman and interpreted by R. Parrish.

This sample intrudes metamorphic rocks of the Custer gneiss. It is from the footwall of the moderately to steeply east-dipping Petch Creek fault northeast of Hope, B.C. The Custer gneiss is juxtaposed against the Hozameen group by this fault. This sample's U-Pb zircon crystallization age is 68.5 ± 0.2 Ma. See GSC 91-28 for interpretation.

GSC 91-24 Biotite
40.1 ± 0.6 Ma

Wt % K=7.066
Rad. Ar=$1.113x10^{-5}$ cm^3/g
K-Ar 4162 % Atmos. Ar=7.2

From a foliated, protomylonitic leucocratic hornblende-biotite quartz monzonite.
(92 H/6) From a ridge outcrop, elevation 2600 feet northeast of Hope, B.C.; 49°27.96'N, 121°23.32'W; UTM zone 10U, 616750E, 5480300N; sample MC-89-119. Collected by R. Parrish and M. Coleman and interpreted by R. Parrish.

This sample intrudes metamorphic rocks of the Custer gneiss. It is from the footwall of the moderately to steeply east-dipping Petch Creek fault northeast of Hope, B.C. The Custer gneiss is juxtaposed against the Hozameen group by the Ross Lake fault. This sample's U-Pb zircon crystallization age is 68.5 ± 0.2 Ma. See GSC 91-28 for interpretation.

GSC 91-25 Biotite
66.9 ± 1.2 Ma

Wt % K= 6.423
Rad. Ar= $1.701x10^{-5}$ cm^3/g
K-Ar 4212 % Atmos. Ar= 9.4

From the Mission Ridge Pluton, here a weakly foliated, medium to coarse-grained quartz monzonite, which intrudes the Bridge River schist.
(92 J/16) From a roadcut in Bridge River Canyon, B.C.; 50°47'30" N, 122°13'30" W; UTM zone 10U, 554750E, 5625250N; sample MC-89-1. Collected by M. Coleman and interpreted by R. Parrish.

This sample has a U-Pb zircon age of 47.5 ± 0.5 Ma. See GSC 91-28 for interpretation.

GSC 91-26 Hornblende
53.8 ± 2.8 Ma

Wt % K=0.833
Rad Ar=$1.769x10^{-6}$ cm^3/g
K-Ar 4213 % Atmos. Ar=17.5

From the Mission Ridge Pluton, here a weakly foliated, medium to coarse-grained quartz monzonite, which intrudes the Bridge River schist.
(92 J/16) Roadcut in Bridge River Canyon, B.C.; 50°47'30" N, 122°13'30" W; UTM zone 10U, 554750E, 5625250N; sample MC-89-1. Collected by M. Coleman and interpreted by R. Parrish.

This sample has a U-Pb zircon age of 47.5 ± 0.5 Ma. See GSC 91-28 for interpretation.

GSC 91-27 Biotite
45.5 ± 0.8 Ma

Wt % K=5.575
Rad Ar=$9.993x10^{-6}$ cm^3/g
K-Ar 4214 % Atmos. Ar=7.1

From a quartz monzonite.
(92 J/16) From a roadcut in Bridge River Canyon, B.C.; 50°47'10" N, 122°11'00" W; UTM zone 10U, 557500E, 5627000N; sample MC-89-3. Collected by M. Coleman and interpreted by R. Parrish.

From the Mission Ridge Pluton, here a strongly foliated, mylonitized medium to coarse-grained quartz monzonite, which intrudes the Bridge River schist. This sample has a U-Pb zircon age of 47.5 ± 0.5 Ma. See GSC 91-28 for interpretation.

GSC 91-28 Hornblende
50.4 ± 1.4 Ma

Wt % K= 1.343
Rad. Ar= $2.668x10^{-6}$ cm^3/g
K-Ar 4215 % Atmos. Ar= 13.1

From a quartz monzonite.
(92J/16) From a roadcut in Bridge River Canyon, B.C.; 50°47'10" N, 122°11'00" W; UTM zone 10U, 557500E, 5627000N; sample MC-89-3. Collected by M. Coleman and interpreted by R. Parrish.

This sample was from the Mission Ridge Pluton, here a strongly foliated, mylonitized medium to coarse-grained quartz monzonite, which intrudes the Bridge River schist. This sample has a U-Pb zircon age of 47.5 ± 0.5 Ma.

These six K-Ar dates (GSC 91-23 to 28) are from variably deformed granitic rocks which intrude upper greenschist to lower amphibolite facies metamorphic rocks of either the Bridge River schist (GSC 91-25,26,27,28) or the Custer gneiss (GSC 91-23,24). Coleman (1990) has interpreted these metamorphic rock units to be the same, displaced dextrally about 100 km by the younger Fraser River fault. Granitic rocks have U-Pb zircon ages of either 47.5 Ma (MC-89-1,3; Coleman 1990) or 68.5 Ma (sample MC-89-119 unpublished data, R. Parrish). The K-Ar ages of hornblende and biotites from these rocks are, with two exceptions, older than crystallization ages and the explanation for this must be excess argon. It is difficult to ascribe any significance to the dates because of this. The biotite ages of 45.5 Ma for GSC 91-27 and 40.1 Ma for GSC 91-24 could represent cooling ages which would be consistent with the known geological and age relationships.

REFERENCE

Coleman, M.E.
1990: Eocene dextral strike-slip and extensional faulting in the Bridge River terrane, southwest British Columbia; M.Sc. thesis, Carleton University, Ottawa, 87 p.

GSC 91-29 Biotite
20.6 ± 0.6 Ma

Wt % K= 4.841
Rad. Ar= 3.90×10^{-6} cm^3/g
K-Ar 3955 % Atmos. Ar= 83.1

From a fresh, medium grained hornblende biotite granodiorite intrusion.
(92H/12) Located on the shore of the east side of the north end of Harrison Lake, 1.3 km on a bearing of 53° from Doctors Point, on the west side of the lake, B.C.; 49°40'03", 121°57'50"; UTM zone 10U 574890E, 5502020N; sample MV85-492. Collected and interpreted by J.W.H. Monger.

This sample is from the northeastern extension of the Doctors Point pluton on the west side of Harrison Lake, K-Ar dates by Ray (in press) range from 19.3 Ma (Bi) to 20.4 Ma (Hb). It is younger than the Chilliwack Batholith and satellitic stocks (ca. 24 Ma) and slightly older than the Mount Barr Batholith (ca. 18 Ma) located about 50 km to the south-southeast. The intrusion is elongate in a northeast direction, which is the orientation of numerous faults and linears in the region. On other evidence, these structures are of late Tertiary age (Monger, 1989).

REFERENCES

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GSC 91-30 Biotite
100.7 ± 1.4 Ma

Wt % K= 7.096
Rad. Ar 2.855×10^{-5} cm^3/g
K-Ar 3956 % Atmos. Ar= 4.4

From a muscovite-biotite granodiorite.
(92 H/11) Located on the northwest side of Coquihalla River valley, elevation 3900 feet, 1.5 km southeast of southeast end of Falls Lake, B.C.; 49°36'10"N, 121°03'50"W; UTM zone 10U 639960E, 5496120N; sample MV85-881a. Collected by J.W.H. Monger.

This sample is from a suite of muscovite-bearing mid-Cretaceous plutons that occur on the west side of the Eagle Plutonic Complex (and its northerly structural continuation the Mount Lytton Complex (Monger, 1989)). It is flanked on the west-southwest by the Pasayten Fault, across which are Triassic through early Late Cretaceous strata of the Methow trough. It is the youngest, westernmost, component of the Eagle Plutonic Complex, most components of which are latest Jurassic (155-148 Ma) in age based on extensive U-Pb dating (Greig, 1989). The K-Ar date herein probably represents a cooling date as Greig has U-Pb dates of ca. 110 Ma from this body.

REFERENCES

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1989: Geology and geochronometry of the Eagle Plutonic complex, Coquihalla area, southwestern British Columbia; M.Sc. thesis, University of British Columbia, 423 p.

Monger, J.W.H.
1989: Geology of Hope and Ashcroft map-areas, British Columbia; Geological Survey of Canada, Maps 41-1989 and 42-1989; scale 1:250 000.

GSC 91-31 Hornblende
106.5 ± 2.2 Ma

Wt % K= 0.831
Rad. Ar= 3.545×10^{-6} cm^3/g
K-Ar 4144 % Atmos. Ar= 8.0

From a quartz diorite-diorite complex.
(92G/15) 100 m from its northeastern margin; elevation 5100 feet, 1.5 km due west of the northern tip of Misty Lake, southwestern Coast Mountains, B.C.; 49°49'07"N, 122°36'10"; UTM zone 10U, 528600E, 5518350N; sample MVD89-394. Collected by K. Dixon for J.W.H. Monger and interpreted by Monger.

The apparent (along strike, not traced physically because of ice-cover) continuation of this mafic granitic complex, 12 km to the north, on the north side of Snowcap Lake, intrudes probable Jura-Cretaceous sedimentary rocks, which contain distinctive granitic cobble-boulder conglomerate, a characteristic of the Lower Cretaceous Cheakamus Formation farther west, near the settlement of Whistler. A possible continuation of this belt, still farther north, may be the Pemberton Diorite, from which a preliminary U-Pb date of 113 ± 2 Ma has been obtained by Friedman and Armstrong (1990). These rocks form part of the Squamish River tract of southwestern Coast Belt (Monger, 1991), which features mainly greenschist grade Middle Triassic to mid Cretaceous stratified rocks, and Late Jurassic and Early Cretaceous granitic intrusions, all cut by northeast-dipping, northeast-side-up reverse and thrust faults of early Late Cretaceous age.

REFERENCES

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Monger, J.W.H.
1991: Georgia Basin Project: structural evolution of parts of southern Insular and southwestern Coast belts, British Columbia; in Current Research, Part A; Geological Survey of Canada, Paper 91-1A, p. 219-228.

GSC 91-32 Hornblende
96.0 ± 1.4 Ma

Wt % K= 0.794
Rad. Ar= 3.043×10^{-6} cm^3/g
K-Ar 4145 % Atmos. Ar= 7.1

From a quartz diorite-diorite complex.
(92 G/15) Elevation 5300 feet, on the south side of Snowcap Lake, 0.3 km south of shore, and 1.3 km west of west end of Cutoff Lake, B.C.; 49°51'40"N, 122°36'35"W; UTM zone 10U 528150E, 5523100N; sample MVD89-375. Collected by K. Dixon for J.W.H. Monger, and intepreted by Monger.

See discussion for GSC 91-31. From this locality, a preliminary U-Pb date of 94 ± 1 Ma was obtained (R.R. Parrish, pers comm. 1990). The agreement between the two ages, argues for rapid cooling of this body.

GSC 91-33 Hornblende
145.5 ± 2.2 Ma

Wt % K= 0.532
Rad. Ar= 3.134×10^{-6} cm^3/g
K-Ar 4146 % Atmos. Ar=8.5

From a medium grained, greenish, partly chloritized hornblende quartz diorite.
(92 G/10) On logging road, elevation 3200 feet, 1.3 km east of fork in upper part of Mamquam River, 4.6 km south of confluence of Mamquam River and Crawford Creek, B.C.; 49°38'40"N, 122°53'55"W; UTM 10U 507275E 5498950N; sample MV89-216. Collected and interpreted by J.W.H. Monger.

The sample is from the SSE extension of the Cloudburst quartz diorite. Fifteen km to the northwest, on the north side of Mamquam River, a preliminary U-Pb date of 147 ± 1 Ma (R.R. Parrish pers comm., 1990) has been obtained from this body, and about 20 km north of this again, a preliminary U-Pb date of 145 ± 2 Ma has been determined from it in the Cheakamus Canyon area (Friedman and Armstrong, 1990). The lithological similarity between these areas and the dates suggest that the Cloudburst quarz diorite occupies an area at least 25 km wide across the regional strike, and perhaps 100 km along it. In addition, there is concordance between U-Pb (crystallization) dates and K-Ar (cooling) dates.

REFERENCE

Friedman, R.M. and Armstrong, R.L.
1990: U-Pb dating, southern Coast Belt, British Columbia; Lithoprobe Cordilleran Workshop 1990; University of Calgary, p. 146-155.

GSC 91-34 Hornblende
39.1 ± 0.9 Ma

Wt % K= 0.265
Rad. Ar= 4.069×10^{-7} cm^3/g
K-Ar 4147 % Atmos. Ar= 29.6

From an andesite dyke containing hornblende needle phenocrysts, intruding probable Jura- Cretaceous Cloudburst quartz diorite.
(92G/10) Located on logging road at an elevation of 2700 feet, 0.75 km southeast of fork in upper part of Mamquam River, and 4.6 km south of confluence of Mamquam River and Crawford Creek, B.C.; 49°30'30"N, 122°54'15"W; UTM zone 10U 5106950E 5498575N; sample MV89-213. Collected by J.W.H. Monger.

The dyke represents a late Eocene magmatic episode, of which little other evidence has been found so far in this part of the Coast Mountains, which is dominated by Jura-Cretaceous, Early to mid Cretaceous and latest Tertiary and Recent magmatism.

GSC 91-35 Hornblende
93.6 ± 1.8 Ma

Wt % K= 0.834
Rad. Ar= 3.115×10^{-6} cm^3/g
K-Ar 4123 % Atmos. Ar= 66.3

From a foliated, lineated amphibolite gneiss.
(82 L/8) Located on Goat Mountain, B.C. Elevation 7220', 50°25'50.2"N, 118°18'19.1"W; UTM zone 11, 407260E-5587120N; sample SC1367H. Collected and interpreted by S. Carr.

See GSC 91-39 for interpretation and references.

GSC 91-36 Hornblende
174.3 ± 4.3 Ma

Wt % K= 0.820
Rad. Ar= 5.833×10^{-6} cm^3/g
K-Ar 4124 % Atmos. Ar= 10.9

From a medium grained, hornblende-bearing granodiorite.
(82 L/2) West of Spruce Grove, B.C. on the Monashee Highway. Elevation 3990'; 50°4'12.4"N, 118°32'36.2"W; UTM zone 11, 389550E-5547350N; sample SC919. Collected and interpreted by S. Carr.

See GSC 91-39 for interpretation and references.

GSC 91-37 Hornblende
64.7 ± 1.2 Ma

Wt % K= 0.754
Rad. Ar= 1.929×10^{-6} cm^3/g
K-Ar 4125 % Atmos. Ar= 14.1

From a foliated, lineated biotite and hornblende-bearing amphibolite schist.
(82 L/1) From Pinnacles Peaks, B. C. Elevation 7000'; 50°612'56.7"N, 118°14'28.0"W; UTM zone 11, 411450E-5563140N; sample SC857. Collected and interpreted by S. Carr.

See GSC 91-39 for interpretation and references.

GSC 91-38 Hornblende
57.3 ± 1.1 Ma

Wt % K= 0.431
Rad. Ar= 9.750×10^{-7} cm^3/g
K-Ar 4126 % Atmos. Ar= 16.6

From a foliated, lineated amphibolite gneiss.
(82 K/4) From Saddle Mountain, B.C. in the footwall of the Columbia River fault. Elevation 7500'; 50°10'15.2"N, 117°53'55.1"W; UTM zone 11, 435825E-5557800N; sample SC928. Collected and interpreted by S. Carr.

See GSC 91-39 for interpretation and references.

GSC 91-39 Hornblende
58.3 ± 1.2 Ma

Wt % K= 0.846
Rad. Ar= 1.948×10^{-6} cm^3/g
K-Ar 4127 % Atmos. Ar= 22.8

	Foliated, lineated hornblende plagioclase gneiss.
(82 L/9)	From Cariboo Alp, Thor-Odin culmination B. C. in the hanging wall of the Monashee décollement and the footwall of the Columbia River fault. Elevation 7200'; 50°31'4.9"N, 118°11'48.6"W; UTM zone 11, 415150E-5596700N; sample SC569. Collected and interpreted by S. Carr.

Five K-Ar hornblende samples GSC 91-35 to 39 were collected from the Thor-Odin — Pinnacles area in the southern Omineca Belt in British Columbia to elucidate the cooling history of the region. Amphibolite-facies rocks, exposed as a gneiss complex within the Shuswap metamorphic complex, were buried during Mesozoic — Late Paleocene compression and were exhumed in the lower plates of Eocene low to moderate angle normal faults. U-Pb zircon and monazite dates from syn-metamorphic igneous rocks and U-Pb ages of metamorphic zircons and titanites indicate that the thermal peak of metamorphism occurred in the Late Cretaceous — Paleocene (Carr, 1990) and that rapid cooling occurred in the Eocene (Parrish et al., 1988). These rocks, in the footwall of the Columbia River and Okanagan Valley normal faults, were juxtaposed against hanging wall rocks which preserve a pre-Middle Jurassic penetrative deformation and metamorphic history and a Jurassic — Cretaceous cooling history (Parrish et al., 1988).

Samples GSC 91-35,37,39 are from the lower plate of the normal faults. GSC 91-37,38,39 which range from 64.7 to 57.3 Ma represent cooling ages of the amphibolite-grade Thor-Odin — Pinnacles area. These ages are slightly younger than metamorphic zircon and titanites from amphibolites in the region. The 93.6 hornblende of sample GSC 91-35 is much older than the others. Field relations indicate that the age is too old; the hornblende probably contained excess argon. Paragneisses and amphibolites at this locality were intruded by Paleocene — Eocene Ladybird granite pegmatites and granites sills and it is unlikely that this happened at temperatures below the 540°C closure temperature of Ar in hornblende.

Sample GSC 91-36 is from granodiorite of the Middle Jurassic Spruce Grove batholith that is inferred to be in the hanging wall of the Okanagan Valley fault system. The U-Pb zircon and titanite ages from the same sample are 175 Ma. The 174.3 Ma K-Ar hornblende age indicates that the batholith cooled rapidly and was thermally undisturbed by the metamorphism that was ongoing at lower structural levels during the Mesozoic and Paleocene.

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GSC 91-40	Biotite **141.1 ± 2.6 Ma**
	Wt % K= 7.108 Rad. Ar= 4.054 x 10^{-5} cm^3/g
K-Ar 4271	% Atmos. Ar= 2.4
	From a dump sample of altered metagabbro cut by a drusy quartz-sulphide vein.
(82 F/14)	Mountain Con deposit, headwater of Carpenter Creek, B.C.; 117°08'12.8"W, 49°55'56.5"N; UTM zone 11u, 5530900N, 490175E; sample 81-3. Collected and interpreted by Georges Beaudoin.

This metagabbro dyke has a biotite K-Ar age similar to two other lamprophyre dykes farther west in Quesnellia terrane. Beaudoin et al. (in press) suggested they represent an Early Cretaceous alkaline event, although this event has yet to be adequately documented. See GSC 91-47 for references.

GSC 91-41	Whole rock **150 ± 7 Ma**
	Wt% K= 1.38 Rad. Ar= 0.838 x 10^{-5} cm^3/g
K-Ar 1939	% Atmos. Ar= 23.0
	From a lamprophyre.
(82 F/15)	Amazon deposit, north of Ainsworth, B.C.; 116°53'35.0"W, 49°46'41.2"N, UTM zone 11u, 5513750N, 506500E; sample SP-1726. Collected by D.F. Sangster, interpreted by Georges Beaudoin.

This lamprophyre cuts metasedimentary rocks with Eocene whole rock K-Ar ages (Matthews, 1983). The Jurassic age therefore indicates significant excess Ar (Beaudoin et al., in press). See GSC 91-47 for references.

GSC 91-42 Whole rock
60 ± 3 Ma

Wt % K= 3.01
Rad. Ar= 0.719 x 10^{-5} cm^3/g
K-Ar 1938 % Atmos. Ar= 18.0

From a lamprophyre.
(82 F/3) Vicinity of the Jersey deposit, south of Salmo, B.C.; 117°13'27.5"W, 49°05'43.7"N; UTM zone 11u, 5437875N, 483625E; sample SP-1724. Collected by D.F. Sangster, interpreted by Georges Beaudoin.

This lamprophyre cuts Jersey deposit stratabound massive sulphides. The whole rock K-Ar age is older than a biotite K-Ar age of 49.5 ± 1.5 Ma from a similar dyke from the Jersey (Macdonald, 1974), therefore indicating that excess Ar is present in this sample (Beaudoin et al., in press). See GSC 91-47 for references.

GSC 91-43 Biotite
39.9 ± 0.7 Ma

Wt % K= 7.506
Rad. Ar= 1.177 x 10^{-5} cm^3/g
K-Ar 4270 % Atmos. Ar= 11.3

From a biotite porphyric kersantite dyke.
(82 F/14) This sample cuts the Nelson batholith on N bank of Enterprise Creek logging road, B.C.; 117°22'47.2" W, 49°51'19.2" N; UTM zone 11u, 5522400N, 472700E; sample ENT-2. Collected and interpreted by Georges Beaudoin.

See GSC 91-47 for discussion and references.

GSC 91-44 Biotite
44.5 ± 0.8 Ma

Wt % K= 6.160
Rad. Ar= 1.080 x 10^{-5} cm^3/g
K-Ar 4272 % Atmos. Ar= 14.0

From a biotite porphyritic kersantite dyke.
(82 F/10) Near the Spokane-Trinket deposit, near Ainsworth, B.C.; 116°55'10.3" W, 49°43'57.7" N; UTM zone 11u, 5508700N, 505800E; sample 28-8. Collected and interpreted by Georges Beaudoin.

See GSC 91-47 for discussion and references.

GSC 91-45 Biotite
49.4 ± 1.0 Ma

Wt % K= 7.320
Rad. Ar= 1.425 x 10^{-5} cm^3/g
K-Ar 4273 % Atmos. Ar= 14.0

From a porphyritic biotite kersantite dyke.
(82 K/3) Cutting Slocan Group metapelites, Whitewater Creek, near Retallack, B.C.; 117°07'48.0" W, 50°03'11.7" N; UTM zone 11u, 5544340N, 490693E; sample 33-9. Collected and interpreted by Georges Beaudoin.

See GSC 91-47 for discussion and references.

GSC 91-46 Biotite
48.8 ± 1.0 Ma

Wt % K= 7.667
Rad. Ar= 1.473 x 10^{-5} cm^3/g
K-Ar 4274 % Atmos. Ar= 38.2

From a porphyritic biotite kersantite dyke.
(82 F/14) Cutting the Nelson batholith, Chapleau Creek logging road, B.C.; 117°23'48.5"W, 49°43'32.7"N; UTM zone 11u, 5508000N, 471400E; sample 131-1. Collected and interpreted by Georges Beaudoin.

See GSC 91-47 for discussion and references.

GSC 91-47 Hornblende
45 ± 3 Ma

Wt % K= 0.924
Rad. Ar= 0.164 x 10^{-5} cm^3/g
K-Ar 1926 % Atmos. Ar= 37.0

From a lamprophyre.
(82 F/15) Vicinity of the Star deposit, near Ainsworth, B.C.; 116°51'53.8" W, 49°44'39.6" N; UTM zone 11u, 5510000N, 509730E; sample F-70-42. Collected by J.T. Fyles, interpreted by Georges Beaudoin.

These kersantite and lamprophyre dykes (GSC 91-43,44,45,46,47) are from a suite of similar dykes outcropping in SE B.C.. Biotite and hornblende K-Ar ages from these dykes range from 40 to 52 Ma (Beaudoin et al., in press). These dykes are coeval, on a structural basis, with Ag-Pb-Zn-Au vein and replacement deposits in Kokanee Range.

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GSC 91-48 Whole rock
26 ± 5 Ma

Wt % K= 1.98
Rad. Ar= 0.203 x 10^{-5} cm^3/g
K-Ar 1918 % Atmos. Ar= 40.0

From an altered gabbro dyke.
(82 F/15) Stope 2Q, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-143. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

See GSC 91-51 for discussion and references.

GSC 91-49 Whole rock
28 ± 2 Ma

Wt % K= 2.03
Rad. Ar= 0.222 x 10^{-5} cm^3/g
K-Ar 1919 % Atmos. Ar= 37.0

From an altered gabbro dyke.
(82 F/15) Stope 3Z, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-144. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

See GSC 91-51 for discussion and references.

GSC 91-50 Whole rock
30 ± 2 Ma

Wt % K= 3.18
Rad. Ar= 0.369 x 10^{-5} cm^3/g
K-Ar 1920 % Atmos. Ar= 16.0

From an altered gabbro dyke.
(82 F/15) Stope 3W, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-145. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

See GSC 91-51 for discussion and references.

GSC 91-51 Whole rock
30 ± 3 Ma

Wt % K= 1.96
Rad. Ar= 0.227 x 10^{-5} cm^3/g
K-Ar 1921 % Atmos. Ar= 45.0

From an altered gabbro dyke.
(82 F/15) Stope 5S, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-147. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

These altered gabbro dykes (GSC 91-48,49,50,51) are cut by sulphide veins but have younger K-Ar ages than crosscutting vein micas. Dyke whole rock ages are interpreted as intermediate ages between age of intrusion and age of alteration (Beaudoin et al., in press). See GSC 91-56 for references.

GSC 91-52 Biotite
50 ± 3 Ma

Wt % K= 6.45
Rad. Ar= 1.276 x 10^{-5} cm^3/g
K-Ar 1932 % Atmos. Ar= 52.0

From a dacite dyke.
(82 F/15) Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample F-70-29. Collected by J.T. Fyles, interpreted by Georges Beaudoin.

This dacite dyke was emplaced before Middle Jurassic regional deformation. The Eocene K-Ar age represent either a metamorphic cooling age (Archibald et al., 1984) or thermal resetting (Matthews, 1983). This age is identical to the ages of other country rocks from the Bluebell deposit (Beaudoin et al., in press). See GSC 91-56 for references.

GSC 91-53 Phlogopite
76 ± 4 Ma

Wt % K= 7.09
Rad. Ar= 2.128 x 10^{-5} cm^3/g
K-Ar 1944 % Atmos. Ar= 13.0

From a marble.
(82 F/15) Phologopite from mica-rich laminations within Badshot Formation marble, stope 8M, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" W, UTM zone 11u, 5512000N, 510000E; sample 67-0-35. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

This phlogopite is older than K-Ar ages from country rocks. This is likely the result of incomplete resetting of older phlogopite during the Eocene thermal event or possibly from incorporation of extraneous radiogenic argon during metamorphic crystallisation of phlogopite in Eocene time (Beaudoin et al., in press). See GSC 91-56 for references.

GSC 91-54 Muscovite
59 ± 3 Ma

Wt % K= 7.01
Rad. Ar= 1.626 x 10^{-5} cm^3/g
K-Ar 1948 % Atmos. Ar= 26.0

From a quartz vein.
(82 F/15) With disseminated muscovite, galena, and sphalerite cutting an altered gabbro dyke, stope 516, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N, UTM zone 11u, 5512000N, 510000E; sample 67-0-44. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

This muscovite K-Ar age dates Ag-Pb-Zn mineralisation at the Bluebell deposit. This 59 ± 3 Ma age is identical to a $^{40}Ar/^{39}Ar$ step heating age of 58.2 ± 0.7 Ma for a hydrothermal alteration muscovite from the Enterprise deposit (Beaudoin et al. in press). See GSC 91-56 for references.

GSC 91-55 Biotite
50 ± 3 Ma

Wt% K= 7.63
Rad. Ar= 1.509 x 10^{-5} cm^3/g
K-Ar 1945 % Atmos. Ar= 19.0

From a Hangingwall mica schist.
(82 F/15) Bluebell deposit adit tunnel, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-40B. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

See GSC 91-56 for discussion and references.

GSC 91-56 Muscovite
55 ± 3 Ma

Wt % K= 5.56
Rad. Ar= 1.213 x 10^{-5} cm^3/g
K-Ar 1949 % Atmos. Ar= 22.0

From a hangingwall mica schist.
(82 F/15) Stope 2Z1, Bluebell deposit, Riondel, B.C.; 116°51'40.2" W, 49°45'44.4" N; UTM zone 11u, 5512000N, 510000E; sample 67-0-103. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

These Eocene K-Ar age for hangingwall muscovite (GSC 91-55,56) represent a metamorphic cooling age (Archibald et al., 1984) or thermal resetting (Matthews, 1983). Whole rock ages increase east and west of a metamorphic culmination centred on Kootenay Lake (Beaudoin et al., in press).

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GSC 91-57 Whole rock
82 ± 7 Ma

Wt % K= 2.70
Rad. Ar= 8.840 x 10^{-4} cm^3/g
K-Ar 1922 % Atmos. Ar= 6.0

From a green dyke intruded into metasedimentary rocks.
(82 F/11) From a road cut on highway 3A, about 16.9 km south of Nelson, B.C.; 40°20'N, 117°14.5'W; UTM zone 11 482442E 5464321N; sample 67-0-137. Collected by H. Ohmoto, interpreted by Georges Beaudoin.

This Cretaceous age likely represents a metamorphic cooling age for this dyke which is described by Ohmoto as older than Middle Jurassic (?) regional deformation.

GSC 91-58 Biotite
56.3 ± 2.5 Ma

Wt % K= 7.615
Rad Ar= 1.692x10^{-5} cm^3/g
K-Ar 3879 % Atmos. Ar= 6.6

From a biotite granite.
(82 E/10) West of Copperkettle Creek, about 10 km west of upper Kettle River, B.C.; 49°40.1'N, 118°51.7'E, UTM zone 11, 363800E, 5503200N; sample TO-83-44-13. Collected by D. Templeman-Kluit; interpreted by R. Parrish.

This biotite age of 56 Ma comes from Eocene biotite granite of the Ladybird granite suite. The sample is from a locality about 10km SW of another granite which has a U-Pb zircon age of Early-Middle Eocene, and the K-Ar age is probably a cooling age.

GSC 91-59 Biotite
47.8 ± 0.8 Ma

Wt % K= 4.399
Rad Ar= 8.289x10^{-6} cm^3/g
K-Ar 3880 % Atmos. Ar= 9.6

(82E/9) About 3km SE of the summit of Mt. O'Leary and about 8km W of Lower Arrow Lake, B.C.; 49°42.0'N, 118°15.3'E, UTM zone 11, 409000E, 5505000N; sample TO-83-52-12. Collected by D. Templeman-Kluit; interpreted by R. Parrish.

This sample is either from the Coryell syenite or the biotite granite of the Ladybird granitic suite of the southern Omineca belt. No information on the exact rock type collected is available. The localitiy is very near the western margin of a large pluton of Coryell syenite centred on Lower Arrow Lake, and the 48 Ma age likely reflects cooling following the intrusion of the Middle Eocene body.

GSC 91-60 Biotite
52.4 ± 0.8 Ma

Wt % K= 7.353
Rad. Ar= 1.518 x 10^{-5} cm^3/g
K-Ar 4188 % Atmos. Ar= 21.8

From a foliated biotite leucogranite of the Hugh Allan gneiss.
(83 D/7) Outcrop on a steep hillside, elevation about 2800', about 150m from the east shore of Kinbasket Lake, B.C.; 52°22.5'N, 118°38.58'W; UTM zone 11u, 388150E, 5803800N. Location given by M.R. McDonough (unpublished data, 1991). Sample GQ67-351.360, collected by C. Giovanella and interpreted by M.R. McDonough.

This sample is a weakly foliated pale buff gneissic biotite leucogranite that forms a 30-50 cm thick layer within the Hugh Allan gneiss, part of the Malton Complex. The protolith is a 736 Ma (U-Pb zircon age; McDonough and Parrish, in press) biotite granite that appears to intrude older (Early Proterozoic) layered gneisses similar to those of the rest of the Malton Complex. The Hugh Allan gneiss occurs on the east side of the Southern Rocky Mountain Trench, in the hanging wall of the post-metamorphic Purcell thrust (M.R. McDonough, unpublished data, 1991). These data are from a re-analysis of a sample that yielded an age of 57 ± 3 Ma (GSC 70-18, Wanless et al., 1972). The 52 Ma age is interpreted as the age at which the Hugh Allan gneiss cooled through the closure temperature of Ar in biotite (ca. 250-300°C). This may be related to uplift associated with the Purcell thrust, or possibly to west-side-down normal faulting in the Southern Rocky Mountain Trench (McDonough and Simony, 1988), which was coincident with the widespread crustal extension in the southern Omineca belt during the Eocene (Parrish et al., 1988). Alternatively, Sevigny et al. (1990) reported muscovite K-Ar ages that are slightly younger than biotite ages from the same suite of rocks from the cover to the Malton gneiss, which suggests that the biotites may have a slight amount of excess Ar. See GSC 91-63 for references.

GSC 91-61 Biotite
118.3 ± 1.7 Ma

Wt % K= 6.947
Rad. Ar= 3.302 x 10^{-5} cm^3/g
K-Ar 4189 % Atmos. Ar= 28.8

From a coarse-grained biotite-muscovite schist.
(83 D/10) Outcrop adjacent to an alpine lake, elevation about 7100'; 52°32'34"N, 118°41'56"W; UTM zone 11u, 384730E, 5822180N. Location given by McDonough and Morrison (1990). Sample 65PF-284, collected by R.A. Price and interpreted by M.R. McDonough.

This sample is a coarse-grained muscovite-biotite schist of the lower Miette Group that forms the cover to Early Proterozoic basement gneisses of the Malton Complex on the east side of the Southern Rocky Mountain Trench. Basement and cover rocks were transported toward the northeast on the Early Cretaceous dextral-oblique Bear Foot thrust (Currie, 1989; McDonough and Simony 1988, 1989). The new age data are from a re-analysis of a sample that yielded a date of 111 ± 5 Ma (GSC 66-47, Wanless et al., 1968). The 111 Ma age was interpreted by Price and Mountjoy (1970) as the age of thrust-generated uplift of these rocks, forming the source terrane for Aptian-Albian sediments of the foreland basin to the east. However, the 118 Ma age is 50-65 Ma older than all other biotite K-Ar ages for the Canoe River (83 D) map-sheet. Additionally, the fact that a muscovite separate from this rock yielded a K-Ar age of 68 Ma (D.A. Archibald, pers. comm., 1990) strongly suggests that the 118 Ma age is more correctly attributed to excess Ar retention in biotite. See GSC 91-63 for references.

GSC 91-62 Hornblende
156.1 ± 2.5 Ma

Wt % K= 0.748
Rad. Ar= 4.742 x 10^{-6} cm^3/g
K-Ar 4190 % Atmos. Ar= 17.2

From medium-grained foliated and lineated biotite-hornblende mafic gneiss.
(83D/11) Outcrop in alpine valley, elevation about 6800', B.C.; 52°35'41"N, 119°01'20"W; UTM zone 11u, 362980N, 5828880. Location given by M.R. McDonough (unpublished data, 1991). Sample 282CACC-1, collected by R.B. Campbell and interpreted by M.R. McDonough.

This sample is a strongly foliated, lineated biotite-hornblende amphibolitic gneiss that forms layers and boudins within the main leucogranitic gneiss of the Malton Complex (McDonough and Morrison, 1990; McDonough and Parrish, in press; M.R. McDonough unpublished data, 1991). The data are from a re-analysis of a sample that yielded an age of 114 ± 12 Ma (GSC 67-44, Wanless et al., 1970). The new date of 156 Ma is similar to Late Jurassic ages for hornblendes of the adjacent Cariboo Mountains (Currie, 1989, also this report, GSC 91-67 (155.6 Ma hornblende collected and interp. by L. Currie)). The 156 Ma age may represent the time of cooling through the closure temperature of Ar in hornblende (ca. 500°C) following Middle Jurassic deformation and metamorphism, which predated the 174 Ma Hobson Lake pluton in the western Cariboo Mountains (Gerasimov, 1988). This age supercedes the previous determination. See GSC 91-63 for references.

GSC 91-63 Biotite
67.4 ± 1.0 Ma

Wt. % K= 7.625
Rad. Ar= 2.036 x 10^{-5} cm^3/g
K-Ar 4191 % Atmos. Ar= 15.4

From a foliated muscovite-biotite gneiss.
(83 D/7) Outcrop along road now submerged under Kinbasket Lake, about 1 km southeast of Hugh Allan Bay, B.C.; 52°25'N, 118°42'W, UTM zone 11u, 384800E, 5810350N. Location given by M.R. McDonough (unpublished data, 1991). Sample GQ67-400-1, collected by C.A. Giovanella and interpreted by M.R. McDonough.

The sample consists of chips of granitic(?) foliated muscovite-biotite gneiss of the Hugh Allan gneiss, part of the Malton Complex. The Hugh Allan gneiss occurs on the east side of the Southern Rocky Mountain Trench, in the hanging wall of the post-metamorphic Purcell thrust (M.R. McDonough, unpublished data, 1991). These data are from a re-analysis of a sample that yielded an age of 66 ± 3 Ma (GSC 70-17, Wanless et al., 1972). The 67 Ma age is essentially the same as the earlier date. It is probably slightly older than the cooling age of the Hugh Allan gneiss since biotites of this area generally cooled through their closure temperature of 250-300°C between 52-57 Ma (see Hunt and Roddick, 1990; Sevigny et al. 1990; also this report GSC 91-60; and GSC 91-65 (55.4 Ma biotite collected and interp. by L. Currie). This suggests that the 67 Ma age probably reflects a component of excess Ar.

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GSC 91-64	Muscovite **63.1 ± 1.1 Ma**
	Wt % K= 8.166 Rad. Ar= 2.038 x 10^{-5} cm^3/g
K-Ar 4098	% Atmos. Ar= 7.5
	From a pegmatite.
(83 D/11)	Outcrop located in cirque at headwaters of Allan Creek, southern Cariboo Mountains, B.C., elevation 7700'; 52°37'20"N, 119°21'42"W; UTM zone 11u, 340134E, 5832648N; sample PCA-PRE. Collected and interpreted by L. Currie.

This pegmatite (156 +6.6/-6.9 Ma; U-Pb zircon, Currie, 1989) intrudes granule conglomerate layers of the lower grit unit of the Mica Creek Succession, and is exposed west of the west-side down, post-metamorphic North Thompson River normal fault and north of the south-side down, post-metamorphic Allan Creek normal fault (Currie and Simony, 1986; M.R. McDonough, pers. comm., 1991). K-Ar analysis of muscovite from this sample was undertaken to provide additional control for the cooling history for this area, which could then be compared with cooling histories for areas on the other sides of the normal faults. See also GSC 91-65, 66, and 67.

The main body of the pegmatite is 2 to 4 m wide, can be followed for at least 300 m, and its margins are concordant with the pervasive penetrative fabric observed in the rocks it intrudes. The apophyses of the pegmatite are folded by mesoscopic folds that also fold the penetrative fabric. It contains muscovite, biotite, quartz, plagioclase and accessory zircon, monazite, rutile and tourmaline. Some of the micas form books up to 3x5x0.5 cm, which in some places outline folds, but in thin section the micas show no evidence of deformation.

The lack of evidence for deformation in the micas that outline folds is interpreted to indicate that the micas were recrystallized during a Middle Jurassic metamorphic event (136 +5.6/-1.8 Ma; U-Pb monazite; Currie 1989) that post-dated folding. The age of GSC 91-64, 63.1 ± 1.1 Ma, is interpreted reflect the cooling of these rocks through the closure temperature of muscovite (about 350°C). This age is consistent with a simple cooling history for this area that is based on other isotopic and geothermometric constraints (Currie, 1989).

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GSC 91-65 Biotite
55.4 ± 0.9 Ma

Wt % K= 6.988
Rad. Ar= 1.529 x 10^{-5} cm^3/g
K-Ar 4100 % Atmos. Ar= 2.9

From a graphitic garnet-kyanite-muscovite schist with plagioclase (An33) accessory rutile and allanite, and post-deformational chloritoid, and sericite.

(83D/11) Outcrop located on logging road that leaves B.C. Forestry North Thompson River Road about 20 km west of highway 5, about 60 km @ 210° from Valemount, B.C., in the southern Cariboo Mountains, B.C., elevation 3500', 56°36'42"N, 119°13'06"W, UTM zone 11u, 363849E, 6276148N; sample PCA-SOF. Collected and interpreted by L. Currie.

The pelite sample from south of the Allan Creek fault is from the black pelite subdivision of the lower Kaza Group (Currie and Simony, 1986). The pelite is exposed west of the west-side down, post-metamorphic North Thompson River normal fault and south of the south-side down, post-metamorphic Allan Creek normal fault (Currie, 1989; M.R. McDonough, pers. comm., 1991). K-Ar analysis of biotite from this sample was undertaken to provide additional control for the cooling history for this area, which could then be compared with cooling histories for areas on the other sides of the normal faults.

The sample is from a 1 m thick pelite layer that is interlayered with 1 m thick impure carbonate layers.

The age of GSC 91-66, 55.4 ± 0.9 Ma, is interpreted as a cooling age that reflects the cooling of these rocks through the closure temperature of biotite (about 280°C). This age is consistent with a simple cooling history for this area that is based on other isotopic and geothermometric constraints for the cooling history of this area (Currie 1989). See GSC 91-64 for references.

GSC 91-66 Biotite
65.5 ± 1.1 Ma

Wt % K= 6.774
Rad. Ar= 1.757 x 10^{-5} cm^3/g
K-Ar 4101 % Atmos. Ar= 2.8

From a garnet-kyanite-muscovite schist with minor biotite, quartz, plagioclase (An5), and accessory rutile, zircon, and monazite.

(83 D/11) Outcrop located on north side of ridge just north of Allan Creek, west of Mt. Milton, and about 50 km @ 190° from Valemount, B.C., in the southern Cariboo Mountains, B.C., elevation 6900', 52°31'00"N, 119°24'40"W; UTM zone 11u, 336395E, 5821020N; sample PCA-NOF. Collected and interpreted by L. Currie.

The pelite sample from north of the Allan Creek Fault was collected from a 30 cm thick schist layer in the semipelite-amphibolite unit of the Mica Creek succession. The pelite is exposed west of the west-side down, post-metamorphic North Thompson River normal fault and north of the south-side down, post-metamorphic Allan Creek normal fault (Currie, 1989; M.R. McDonough, pers. comm., 1991). K-Ar analysis of biotite from this sample was undertaken to provide additional control for the cooling history for this area, which could then be compared with cooling histories for areas on the other sides of the normal faults.

K-Ar analysis of biotite from this sample gives an age of 65.5 ± 1.1 Ma, which is considered to be older than the time that the biotite cooled through its closure temperature (280 ± 40°C). A possible explanation for the anomalously old K-Ar biotite date is the presence of excess argon in the pelite sample. See GSC 91-64 for references.

GSC 91-67 Hornblende
155.6 ± 5.2 Ma

Wt % K= 0.662
Rad. Ar= 4.181 x 10^{-6} cm^3/g
K-Ar 4102 % Atmos. Ar= 9.8

From an amphibolite.
(83D/11) Outcrop located on logging road on east flank of Mt. Milton, 2.2 km west of highway 5, about 50 km @ 185° from Valemount, B.C., in the southern Cariboo Mountains, B.C., elevation 3700', 52°32'20"N, 119°08'30"W, UTM zone 11u, 354749E, 5822915N; sample PCA-AMPH. Collected and interpreted by L. Currie.

The amphibolite sample is from a 1.5 m thick amphibolite layer in the lower Kaza Group (Currie and Simony, 1986). It comprises hornblende, plagioclase, with minor biotite and sphene, is concordant with the overlying and underlying semipelitic layers and concordant with the pervasive metamorphic foliation within the surrounding rocks, and is interpreted to be a metamorphosed basalt flow, basaltic sill or basaltic dyke that has been transposed. The amphibolite is exposed west of the west-side down, post-metamorphic North Thompson River normal fault and south of the south-side down, post-metamorphic Allan Creek normal fault (Currie, 1989; M.R. McDonough, pers. comm., 1991). K-Ar analysis of hornblende from this sample was undertaken to provide additional control for the cooling history for this area, which could then be compared with cooling histories for areas on the other sides of the normal faults.

K-Ar analysis of hornblende for GSC 91-67 gives an age of 155 ± 5.2 Ma. This age is considered to be anomalously old because the closure temperature for K-Ar in hornblende (530 ± 40°C) is within error of the temperatures attained this area during peak metamorphic conditions (540 ± 50°C) 136 +5.6/-1.8 Ma (Currie, 1989). The hornblende may preserve a K/Ar ratio that reflects the age of an earlier metamorphism, or the ratio may have been only partially reset during peak metamorphic conditions, or the ratio may have been completely reset, but due to excess Ar in the sample records an age that is older than the age of peak metamorphic conditions. See GSC 91-64 for references.

YUKON
(GSC 91-68 to GSC 91-119)

GSC 91-68 Biotite
115.3 ± 3.6 Ma

Wt % K= 6.665
Rad. Ar= 3.083 x 10^{-5} cm^3/g
K-Ar 4185 % Atmos. Ar= 88.4

From a quartz-biotite schist.
(116 C/2) Near the headwaters of Browns Creek, 1.8 km north of the Top-of-the-World Highway, 7.7 km due west of Sixtymile Road turnoff, Yukon; 64°06.9'N, 140°46.5'W, UTM zone 7, 510960E, 7109652N; sample M-721. Collected and interpreted by J.K. Mortensen.

This sample is from a band of quartz-biotite schist within grey, carbonaceous schists of the Nasina Series (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-69 Biotite
179.2 ± 4.1 Ma

Wt % K= 7.144
Rad. Ar= 5.232 x 10^{-5} cm^3/g
K-Ar 4116 % Atmos. Ar= 1.6

From a granodioritic gneiss.
(116 C/1) In a small saddle in a ridge, 5.5 km west of California Creek, 3.7 km north of Sixtymile River, Yukon; 64°04.4'N, 140°28.3'W, UTM zone 7, 525775E, 7105096N; sample M-737. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-70 Biotite
183.0 ± 2.3 Ma

Wt % K= 6.868
Rad. Ar= 5.141 x 10^{-5} cm^3/g
K-Ar 4187 % Atmos. Ar= 26.7

From a quartz monzonitic gneiss.
(116 C/2) On the left limit (northeast side) of Miller Creek, 2.0 km above its mouth, Yukon; 64°00.0'N, 140°49.3'W; UTM zone 7, 5508723E, 7096830N; sample M-902. Collected and interpreted by J.K. Mortensen.

The sample is from a biotite quartz monzonitic orthogneiss that probably represents a faulted extension of the Early Mississippian (U-Pb zircon age) Fiftymile Batholith (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-71 Biotite
125.3 ± 2.1 Ma

Wt % K= 6.512
Rad. Ar= 3.285 x 10^{-5} cm^3/g
K-Ar 4114 % Atmos. Ar= 9.3

From a granodioritic gneiss.
(116 C/2) On the south side of the Sixtymile River, 2.3 km above the mouth of Five Mile Creek, Yukon; 64°01.1'N, 140°40.7'W; UTM zone 7, 515724E, 7098900N; sample M-905. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-72 Biotite
179.5 ± 2.8 Ma

Wt % K= 7.090
Rad. Ar= 5.200 x 10^{-5} cm^3/g
K-Ar 4115 % Atmos. Ar= 2.4

From a granodioritic gneiss.
(116 C/2) On the south side of the Sixtymile River, 0.75 km above the mouth of Five Mile Creek, Yukon; 64°01.7'N, 140°39.3'W; UTM zone 7, 515858E, 7100020N; sample M-909. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-73 Biotite
189.4 ± 5.0 Ma

Wt % K= 6.321
Rad. Ar= 4.906 x 10^{-5} cm^3/g
K-Ar 4118 % Atmos. Ar= 1.2

From a granodioritic gneiss.
(116 C/2) On the south side of the Sixtymile River, 1.6 km downstream from the mouth of Twelve Mile Creek, Yukon; 64°02.5'N, 140°30.7'W; UTM zone 7, 523851E, 7101552N; sample M-918. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-74 Biotite
182.0 ± 2.6 Ma

Wt % K= 6.724
Rad. Ar= 5.006 x 10^{-5} cm^3/g
K-Ar 4117 % Atmos. Ar= 1.1

From a granodioritic gneiss.
(116 C/1) On the north side of the Sixtymile River, 6.1 km above the mouth of California Creek, Yukon; 64°02.4'N, 140°29.1'W; UTM zone 7, 525155E, 7101377N; sample M-921. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-75 Biotite
164.7 ± 2.7 Ma

Wt % K= 7.149
Rad. Ar= 4.793 x 10^{-5} cm^3/g
K-Ar 4120 % Atmos. Ar= 64.2

See GSC 91-76 for location and interpretation.

GSC 91-76 Hornblende
172.2 ± 4.3 Ma

Wt % K= 1.014
Rad. Ar= 7.123 x 10^{-6} cm^3/g
K-Ar 4119 % Atmos. Ar= 5.5

From a granodioritic gneiss.
(116 C/1) On the north side of the Sixtymile River, 3.8 km below the mouth of California Creek, Yukon; 64°00.9'N, 140°17.4'W; UTM zone 7, 534710E, 7098682N; sample M-929. Collected and interpreted by J.K. Mortensen.

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See also K-Ar biotite age for same sample (GSC 91-75). See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-77 Biotite
184.0 ± 2.9 Ma

Wt % K= 7.589
Rad. Ar= 5.714 x 10^{-5} cm^3/g
K-Ar 4179 % Atmos. Ar= 1.2

From a biotite-amphibole schist.
(116 C/2) In the west fork of California Creek, 5.1 km above forks, Yukon; 64°08.0'N, 140°26.2'W; UTM zone 7, 527423E, 7111797N; sample M-295. Collected and interpreted by J.K. Mortensen.

This sample is from a band of biotite-amphibole schist (metabasite) within grey, carbonaceous schist and quartzite of the Nasina Series (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-78 Muscovite
155.6 ± 2.5 Ma

Wt % K= 8.108
Rad. Ar= 5.121 x 10^{-5} cm^3/g
K-Ar 4183 % Atmos. Ar= 2.1

From a quartz-muscovite schist.
(116 C/2) In the bed of Hall Creek, 0.3 km upstream from the Yukon-Alaska boundary, Yukon; 64°06.4'N, 140°59.6'W; UTM zone 7, 500325E, 7108704N; sample MLB-88-149. Collected and interpreted by J.K. Mortensen.

This sample is from pyritic quartz-muscovite schist of the Klondike Schist that hosts the BALDY Zn-Pb-Cu sulphide occurrence (Mortensen, 1988). The date represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-79 Muscovite
154.7 ± 3.5 Ma

Wt % K= 8.931
Rad. Ar= 5.608 x 10^{-5} cm^3/g
K-Ar 4108 % Atmos. Ar= 13.2

From a quartz-muscovite schist.
(116 C/2) From a roadcut on the Top-of-the-World Highway, 5 km west of the Sixtymile Road turnoff, Yukon; 64°07.0'N, 140°48.0'W; UTM zone 7, 509742E, 7109834N; sample XXI. Collected by J.E. Muller and interpreted by J.K. Mortensen.

The sample is from quartz-muscovite schist of the Klondike Schist (Mortensen, 1988). This is a re-analysis of GSC 61-40 (Lowdon et al., 1968), and supercedes the previously reported date of 178 Ma. The date obtained represents a cooling age for this unit following regional metamorphism. See GSC 91-80 for references and for further discussion of this and other cooling ages in this area.

GSC 91-80	Biotite **149.7 ± 3.4 Ma**	
	Wt % K= 7.401 Rad. Ar= 4.489 x 10^{-5} cm^3/g	
K-Ar 4109	% Atmos. Ar= 80.3	
	From a granodioritic gneiss.	
(116 C/1)	On the north side of the Sixtymile River, 2 km above the mouth of California Creek, Yukon; 64°02.0'N, 140°23.3'W; UTM zone 7, 529856E, 7100675N; sample Rd 61-475a. Collected by J.A. Roddick and interpreted by J.K. Mortensen.	

The sample is from near the northern margin of the Fiftymile Batholith, a Early Mississippian (U-Pb zircon age) biotite-hornblende granodioritic orthogneiss (Mortensen, 1988). This is a re-analysis of GSC 62-82 (Leach et al., 1963), and supercedes the previously reported age of 203 Ma. The date represents a cooling age for this unit following regional metamorphism.

This series of thirteen K-Ar age determinations (GSC 91-68 to 80) was undertaken to evaluate the cooling history of several thrust fault-bounded structural blocks in part of the Yukon-Tanana terrane in southwestern Dawson map area. K-Ar and Ar-Ar geochronological studies in the Tanacross and Eagle quadrangles immediately to the west (Hansen et al., 1991) suggest that different subterranes that comprise the Yukon-Tanana terrane in this region cooled at markedly different times. Although the database for the Dawson map area is somewhat limited, cooling ages from a thrust panel of Klondike Schist in the northern part of the Sixtymile District (Mortensen, 1988) are consistently 10-35 Ma younger than those obtained from underlying and overlying structural units which consist of metasedimentary rocks of the Nasina Series and/or orthogneisses of the Fiftymile Batholith. Three biotite dates (GSC 91-68,71,80) are anomalously young, and probably reflect local thermal influence of Late Cretaceous plutons and volcanic rocks in this area (Mortensen, 1988).

Cooling ages from the Fiftymile Batholith are only slightly younger than a U-Pb zircon age of 192 Ma determined for a late to post-metamorphic pegmatite dyke that was emplaced prior to thrust imbrication, suggesting relatively rapid cooling following regional deformation and lower amphibolite facies metamorphism. The structural juxtaposition of large bodies of metamorphic rocks with distinct ranges of cooling ages as young as 155 Ma may indicate that thrust faulting in this area was post-155 Ma.

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GSC 91-81	Biotite **253.4 ± 5.8 Ma**	
	Wt % K= 6.231 Rad. Ar.= 6.588 x 10^{-5} cm^3/g	
K-Ar 4220	% Atmos. Ar= 3.4	
	From a medium grained, equigranular biotite hornblende quartz monzonite.	
(115 P/12)	Sampled from an outcrop along Stewart River 250 m downstream of Pique Creek, el. 457 m, Yukon Territory; 63°35.7'N, 137°44.4'W; UTM zone 8 364070E, 7054599N; sample 23689-14b-1. Collected by B. Ward for L.E. Jackson, Jr., and interpreted by L.E. Jackson, Jr. and J.K. Mortensen.	

See GSC 91-82 for discussion and references.

GSC 91-82	Biotite **201.9 ± 3.7 Ma**	
	Wt % K= 5.502 Rad. Ar= 4.568 10^{-5} cm^3/g	
K-Ar 4221	% Atmos. Ar= 1.5	

From a heterogeneous biotite hornblende quartz monzonite (contains coarse aggregates of biotite).

(115 P/12) Sampled from an outcrop along west bank of Stewart River, elevation 457 m, 0.5 km upstream from the mouth of independence Creek, Yukon; 63°30.8'N, 137°48.9'W; UTM zone 8 359951E 7045667N; sample JJ 23689-14a-1. Collected by B. Ward for L.E. Jackson, Jr., and interpreted by L.E. Jackson, Jr. and J.K. Mortensen.

These two dates (GSC 91-81,82) are from a large body of plutonic rocks on the immediate southwest side of the Tintina Fault Zone. This pluton is shown on many compilation maps to cross the fault zone, and therefore could potentially provide a minimum age for displacement on the zone. The granitic rocks that form the apparent extension of this body northeast of the fault zone, however, give a K-Ar biotite age of 87.4 ± 7.0 Ma (GSC 65-50, in Wanless et al., 1967), and are likely related to either the Selwyn or Tombstone plutonic suites. The dates reported here clearly show that the intrusive rocks southwest of the fault zone are a much older, and unrelated body. The two dates are considerably different, however, and it is uncertain whether these are actually cooling ages for middle or late Paleozoic plutonic rocks, or whether some Early Jurassic phases are present. U-Pb dating of the two samples will be employed to resolve this problem.

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GSC 91-83 Muscovite
162.8 ± 3.6 Ma

Wt % K= 8.621
Rad. Ar= 5.708 x 10^{-5} cm^3/g
K-Ar 4110 % Atmos. Ar= 10.0

From a quartz-muscovite schist.

(115 O/15) Roadcut on the Hunker Creek Road near Hunker Summit, Yukon; 63°54.0'N, 138°52.0'W; UTM zone 7, 604708E, 7087425N; sample XXIII. Collected by J.E. Muller and interpreted by J.K. Mortensen.

The sample is from a quartz-muscovite schist phase of the Klondike Schist that is derived from a mid-Permian felsic metavolcanic rock (Mortensen, 1990). The date is a re-analysis of GSC 60-33 (Lowdon, 1961) and supercedes the previously reported age of 141 Ma for this sample. The date is interpreted as a cooling age after regional metamorphism. See GSC 91-95 for references and further discussion of this and related ages for the Klondike District.

GSC 91-84 Muscovite
104.9 ± 1.5 Ma

Wt % K= 8.455
Rad. Ar= 3.548 x 10^{-5} cm^3/g
K-Ar 4177 % Atmos. Ar= 2.3

See GSC 91-85 for location and interpretation.

GSC 91-85 Biotite
99.1 ± 2.8 Ma

Wt % K= 6.859
Rad. Ar= 2.716 x 10^{-5} cm^3/g
K-Ar 4176 % Atmos. Ar= 7.4

From a granitic augen orthogneiss.

(115 O/9) Summit of Mount Burnham, Yukon; 63°41.6'N, 138°17.0'W; UTM zone 7, 634307E, 7065500N; sample R-35. Collected by R. Debicki and interpreted by J.K. Mortensen.

The sample is from the strongly foliated, Early Mississippian, Mt. Burnham granitic orthogneiss (Mortensen, 1990). See GSC 91-84 for muscovite pair. The dates are interpreted as cooling ages after regional metamorphism. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-86 Biotite
117.3 ± 2.3 Ma

Wt % K= 7.335
Rad. Ar= 3.455 x 10^{-5} cm^3/g
K-Ar 4181 % Atmos. Ar= 1.2

Sample MLB-88-177b. See GSC 91-87 for location and interpretation.

GSC 91-87 Muscovite
113.9 ± 3.0 Ma

Wt % K= 7.928
Rad. Ar= 3.621 x 10^{-5} cm^3/g
K-Ar 4180 % Atmos. Ar= 3.6

From a garnetiferous schist.
(115 O/15) From bedrock exposed in a placer mine in the bed of Dominion Creek, 2.5 km above the mouth of Jensen Creek, Yukon; 63°46.8'N, 138°31.0'W; UTM zone 7, 622401E, 7074682N; sample MLB-88-177a. Collected and interpreted by J.K. Mortensen.

The sample is from garnetiferous quartz-muscovite-calcite schist interlayered with quartz-biotite schist. Both units form part of the Klondike Schist (Mortensen, 1990). See GSC 91-86 for biotite pair. The dates are in good agreement, and are interpreted as cooling ages after regional metamorphism. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-88 Hornblende
221.5 ± 5.4 Ma

Wt % K= 0.293
Rad. Ar= 2.684 x 10^{-6} cm^3/g
K-Ar 4122 % Atmos. Ar= 16.4

From an amphibolite.
(115 O/15) From a roadcut on Hunker Creek Road, 0.5 km above the mouth of Gold Bottom Creek; 63°57.7'N, 138°57.3'W; UTM zone 7, 600153E, 7094152N; sample MLB-88-103. Collected and interpreted by J.K. Mortensen.

The sample is from massive to moderately foliated, fine to medium grained amphibolite of the "chloritic schist unit" of Mortensen (1990). This date is considerably older than the K-Ar hornblende age of 159 Ma that was obtained previously for a sample from the same outcrop by Debicki (personal communication, 1985), and for samples from elsewhere in the Klondike District. The date is thought to reflect the presence of excess Ar in the analysis, and no age significance is therefore attached to this analysis. See GSC 91-95 for references and further discussion of this and related ages for the Klondike District.

GSC 91-89 Muscovite
54.1 ± 4.7 Ma

Wt % K= 7.354
Rad. Ar= 1.569 x 10^{-5} cm^3/g
K-Ar 4186 % Atmos. Ar= 95.6

From a quartz-muscovite schist.
(115 O/15) From bedrock exposed in a placer mine in the bed of Dominion Creek at the mouth of Robinson Pup, Yukon; 63°48.8'N, 138°39.4'W; UTM zone 7, 615367E, 7078134N; sample MLB-88-116. Collected and interpreted by J.K. Mortensen.

The sample is from a sillimanite-bearing band of quartz-muscovite schist interlayered with garnet amphibolite and carbonaceous quartz-muscovite-biotite schist (Mortensen, 1990). The date is anomalously young; porphyry dykes that yield ages from 50-55 Ma are widespread but volumetrically minor in the Klondike District, and none are known from near this sample locality. It is therefore unlikely that the date represents a thermal overprinting of Eocene age. The poor quality of the analysis (very high % atmospheric Ar) makes the interpretation of this age somewhat dubious. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-90 Biotite
180.8 ± 3.4 Ma

Wt % K= 7.158
Rad. Ar= 5.290 x 10^{-5} cm^3/g
K-Ar 4248 % Atmos. Ar= 4.3

From a biotite-quartz-feldspar gneiss.
(115 O/13) On a ridge crest 6.3 km due north of the mouth of Bertha Creek, Yukon; 63°50.4'N, 138°31.0'W; UTM zone 7, 572966E, 7079836N; sample MLB-89-296. Collected and interpreted by J.K. Mortensen.

The sample is from a narrow band of biotite-bearing quartz-feldspar gneiss within biotite (± garnet) grade Nasina Series carbonaceous metasediments (Mortensen, 1990). Sample locality is 2.6 km from the contact with the Jim Creek pluton, which yields a similar K-Ar biotite age (see GSC 91-96). The date is interpreted as a cooling age following regional metamorphism, although there may have been some thermal influence associated with the intrusion of the pluton. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-91 Muscovite
181.7 ± 3.6 Ma

Wt % K= 8.866
Rad. Ar= 6.587 x 10^{-5} cm^3/g
K-Ar 4197 % Atmos. Ar= 1.9

From a quartz-muscovite schist.
(115 O/14) From a block in dredge tailings on the northwest side of Quartz Creek, 1.9 km upstream from the mouth, Yukon; 63°46.2'N, 139°07.0'W; UTM zone 7, 592868E, 7072557N; sample MLB-89-284. Collected and interpreted by J.K. Mortensen.

The sample is from an angular block dredged from underlying bedrock, and consists of pyritic quartz-muscovite schist derived from mid-Permian felsic metavolcanic rocks of the Klondike Schist (Mortensen, 1990). The date is interpreted as a cooling age following regional metamorphism, and is consistent with other ages in the area. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-92 Muscovite
175.7 ± 2.9 Ma

Wt % K= 7.364
Rad. Ar= 5.281 x 10^{-5} cm^3/g
K-Ar 4222 % Atmos. Ar= 0.8

From a muscovite schist.
(115 O/14) From the 210'-211' in Arbor Resources diamond drill hole LS-5 adjacent to the abandoned open cut at the Lone Star gold mine at the head of Victoria Gulch, Yukon; 63°53.4'N, 139°13.5'W; UTM zone 7, 587155E, 7085772N; sample LS-5-210-211. Collected by S. Tomlinson and interpreted by J.K. Mortensen.

The sample is from a very muscovite-rich phase of the Klondike Schist (Mortensen, 1990) which forms the host rock for significant disseminated and vein gold mineralization. The date is interpreted as a cooling age following regional metamorphism, and is consistent with other ages in the area. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-93 Biotite
160.9 ± 2.5 Ma

Wt % K= 6.940
Rad. Ar= 4.539 x 10^{-5} cm^3/g
K-Ar 4257 % Atmos. Ar= 2.5

From a quartz monzonitic gneiss.
(115 O/10) From a block in dredge tailings beside the Sulphur Creek Road, 6.0 km northwest of the abandoned townsite of Dominion, Yukon; 63°41.7'N, 138°45.5'W; UTM zone 7, 610826E, 7064774N; sample MLB-89-355. Collected and interpreted by J.K. Mortensen.

The sample is an angular block dredged from underlying bedrock. It consists of moderately to strongly foliated quartz augen-bearing muscovite-biotite-feldspar-quartz gneiss of the mid-Permian Sulphur Creek orthogneiss (Mortensen, 1990). The date is interpreted as a cooling age following regional metamorphism, and is consistent with other ages in the area. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-94 Muscovite
148.8 ± 2.9 Ma

Wt % K= 9.112
Rad. Ar= 5.494 x 10^{-5} cm^3/g
K-Ar 4103 % Atmos. Ar= 61.9

From a quartz-muscovite schist.
(115 O/15) From a roadcut on the Hunker Creek Road 0.25 km east-southeast of the confluence of the right and left forks of Hunker Creek, Yukon; 63°54.7'N, 138°52.9'W; UTM zone 7, 603929E, 7088700N; sample MLB-88-114. Collected and interpreted by J.K. Mortensen.

The sample is from a band of pyritic muscovite (± quartz) schist that is interlayered with impure quartzites of the Klondike Schist (Mortensen, 1990). The sample has also yielded a nearly identical Rb-Sr muscovite age of 143 Ma. These dates are interpreted as cooling ages following regional metamorphism, and are consistent with other ages in the area. See GSC 91-95 for references and further discussion of these and related ages for the Klondike District.

GSC 91-95 Muscovite
165.1 ± 2.9 Ma

Wt % K= 8.964
Rad. Ar= 6.022 x 10^{-5} cm^3/g
K-Ar 4196 % Atmos. Ar= 2.3

From a quartz-muscovite schist.
(115 O/14) From a bulldozer trench on the crest of the ridge between Upper Bonanza and Eldorado creeks, 0.4 km west-southwest of the old workings of the Lone Star gold mine, Yukon; 63°53.2'N, 139°13.8'W; UTM zone 7, 586920E, 7085394N; sample MLB-89-270a. Collected and interpreted by J.K. Mortensen.

The sample is from quartz-muscovite schist of the Klondike Schist (Mortensen, 1990) that forms the host rock for disseminated and vein gold mineralization. The dates are interpreted as a cooling age following regional metamorphism.

This series of thirteen dates (GSC 91-83 to 95) was obtained as part of a detailed investigation of the cooling history of metamorphic rocks in the Klondike District and the place of gold-bearing mesothermal quartz veins (Mortensen et al., in press) in the overall thermal evolution of the area. Most ages fall in the Middle to Late Jurassic range of ages typical for much of the Yukon-Tanana Terrane (e.g. Mortensen, in press). There are subtle regional variations in the cooling ages in the Klondike District, however, suggesting slightly different cooling rates (perhaps related to differential uplift rates) across the area.

The Mount Burnham orthogneiss yields distinctly anomalous, mid-Cretaceous K-Ar cooling ages (GSC 91-84 and 91-85). These ages are only slightly younger than the U-Pb monazite age of 111 Ma determined for the same sample, and indicate that this body cooled from 700 to 300°C over a period of approximately 12 Ma. Schists adjacent to the Mt. Burnham orthogneiss also record mid-Cretaceous cooling ages (GSC 91-86 and 91-87). Several portions of the Yukon-Tanana Terrane, especially those which include Mississippian augen orthogneisses such as the Mt. Burnham orthogneiss, yield mid-Cretaceous cooling ages. Hansen et al. (1991) have used this as a criteria for distinguishing various subterranes of the Yukon-Tanana Terrane.

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GSC 91-96 Biotite
180.8 ± 3.8 Ma

Wt % K= 5.981
Rad. Ar= 4.422 x 10^{-5} cm^3/g
K-Ar 4249 % Atmos. Ar= 6.4

From a biotite quartz monzonite.
(115 O/13) From a ridge crest at the head of Jim Creek, 3.7 km north of Indian River, Yukon; 63°48.8'N, 139°35.9'W; UTM zone 7, 569014E, 7076774N; sample MLB-89-307. Collected and interpreted by J.K. Mortensen.

The sample is from a massive, unfoliated, medium grained, garnet-bearing biotite quartz monzonite informally termed the Jim Creek pluton. This body intrudes biotite-garnet grade carbonaceous Nasina Series metasediments, and is mineralogically unlike any other intrusions in the area. The early Middle Jurassic K-Ar igneous cooling age indicates that it is likely an outlier of the suite of dominantly Late Triassic-Early Jurassic intrusions that are widespread farther to the south and west (see Mortensen, in press, for review). The age is identical to cooling ages in metamorphic rocks in this area (see discussion following GSC 91-95), indicating that intrusion occurred during regional post-metamorphic cooling.

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GSC 91-97 Muscovite
140.1 ± 2.0 Ma

Wt % K= 8.496
Rad. Ar= 4.810 x 10^{-5} cm^3/g
K-Ar 4194 % Atmos. Ar= 29.4

From a quartz vein.
(115 O/15) From a bulldozer trench near the top of a small hill, 1.0 km north-northeast of King Solomon Dome, Yukon; 63°52.6'N, 138°56.5'W; UTM zone 7, 601112E, 7084705N; sample MLB-89-263. Collected and interpreted by J.K. Mortensen.

See GSC 91-98 for discussion.

GSC 91-98 Muscovite
134.5 ± 1.5 Ma

Wt % K= 7.441
Rad. Ar= 4.040 x 10^{-5} cm^3/g
K-Ar 4195 % Atmos. Ar= 1.5

From a quartz vein.
(115 O/15) From a bulldozer trench near the top of a small hill, 1.0 km north-northeast of King Solomon Dome, Yukon; 63°52.6'N, 138°56.5'W; UTM zone 7, 601112E, 7084705N; sample MLB-89-264. Collected and interpreted by J.K. Mortensen.

These two samples are from gold-bearing, low-sulphide, mesothermal quartz veins at the SHEBA occurrence (Mortensen et al., in press). Sample MLB-89-263 (GSC 91-97) is from muscovite disseminated throughout a 1-1.5 m wide coarse-grained quartz vein, and sample MLB-89-264 (GSC 91-98) is from a narrow (<1 cm) band of fine grained muscovite with minor galena in the center of an irregular, 6-15 cm wide vein. The two sample localities are separated by about 50 m. These two dates provide the first conclusive evidence for the age of the mesothermal veins that were the source of the bulk of gold recovered by placer mining in the Klondike District (Mortensen et al., in press). Metamorphic cooling ages in the King Solomon Dome area are mainly in the range of 140-150 Ma (see discussion following GSC 91-95). The latest Jurassic to earliest Cretaceous ages for the veins indicate that veining occurred during late post-metamorphic cooling in this area.

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GSC 91-99 Biotite
79.1 ± 2.2 Ma

Wt % K= 6.084
Rad. Ar= 1.911 x 10^{-5} cm^3/g
K-Ar 4141 % Atmos. Ar= 33.7

From a biotite-quartz-feldspar porphyry.
(116 B/2) From a low bluff between the mouths of Germaine and Goring creeks, 0.8 km west of the Klondike Highway, Yukon; 64°02.4'N, 138°54.6'W; UTM zone 7, 602071E, 7102949N; sample MLB-88-158. Collected and interpreted by J.K. Mortensen.

The sample is from a 40 cm diameter block of fresh biotite-bearing quartz-feldspar porphyry in a felsic volcanic breccia that is part of a sequence of interlayered immature fine to coarse clastic sediments and felsic volcaniclastic rocks that occurs on the southwest flank of the Tintina Trench in the northern Klondike District (Mortensen, 1990). The felsic rocks are lithologically identical to, and are thought to be the extrusive equivalents of Eocene quartz-feldspar porphyry dykes and plugs that are widespread in the region adjacent to the Tintina Fault in western Yukon (Mortensen, 1990). The Late Cretaceous age is therefore surprising, and likely reflects the presence of excess Ar in the analysis. The crystallization age of the sample is presently being tested by U-Pb zircon dating.

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GSC 91-100 Biotite
84.3 ± 1.7 Ma

Wt % K= 5.809
Rad. Ar= 1.948 x 10^{-5} cm^3/g
K-Ar 4247 % Atmos. Ar= 9.6

From a biotite-feldspar-quartz gneiss.
(115 O/2) From bedrock exposed in new placer workings on the west bank of Barker Creek 10 km above its mouth, Yukon; 63°05.7'N, 138°56.9'W; UTM zone 7, 603575E, 6997629N; sample MLB-89-269. Collected and interpreted by J.K. Mortensen.

The sample is from a strongly foliated biotite-feldspar-quartz gneiss (probably an orthogneiss) with well developed compositional layering on a scale of 2-20 cm. Foliaform quartz-feldspar veins and sweats to 2 cm thick are present locally. The date represents the last time that these rocks cooled through the closure temperature of biotite, but it is uncertain whether this reflects very prolonged cooling following regional deformation and metamorphism, or whether the area has been thermally reset by younger magmatism. This region is not well mapped, however Late Triassic and/or Early Jurassic felsic to ultramafic intrusions are widespread immediately south of the sample locality (H.S. Bostock, unpublished field notes, 1935-37), and mid-Cretaceous plutons are likely also present.

GSC 91-101 Muscovite
163.0 ± 2.2 Ma

Wt % K= 8.413
Rad. Ar= 5.580 x 10^{-5} cm^3/g
K-Ar 4184 % Atmos. Ar= 1.3

From a muscovite-quartz-feldspar schist.
(116 C/9) From the east shore of the Yukon River, 0.3 km below the mouth of Shell Creek, Yukon; 64°30.2'N, 140°25.5'W; UTM zone 7, 527618E, 7153034N; sample M-1050. Collected and interpreted by J.K. Mortensen.

The sample is from a muscovitic quartz-feldspar schist which is interlayered with grey, carbonaceous, quartz-feldspar-biotite-muscovite schist of the Nasina Series (Mortensen, 1988). The date is interpreted as a cooling age following regional metamorphism in this region (see discussion following GSC 91-80).

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GSC 91-102 Biotite
110.9 ± 1.7 Ma

Wt % K= 7.190
Rad. Ar= 3.195 x 10^{-5} cm^3/g
K-Ar 4121 % Atmos. Ar= 12.1

From a biotite granodiorite.
(116 C/8) From the north shore of the Yukon River, 6.75 km below the mouth of Cassiar Creek, Yukon; 64°21.4'N, 140°17.3'W; UTM zone 7, 534366E, 7136756N; sample M-1124. Collected and interpreted by J.K. Mortensen.

The sample is from a large body of massive, unfoliated, medium grained, equigranular biotite granodiorite informally termed the Mt. Carmacks pluton (Mortensen, 1990). This pluton is genetically associated with W-base metal skarn mineralization at the RAIL, ROAD, and TRACK occurrences. The date is older than expected, as most intermediate composition plutons in this area have yielded ages in the range of 60-70 Ma (Mortensen, 1990). U-Pb dating of zircon from this sample, however, also gives an age of 110 Ma. The Mt. Carmacks pluton (and possibly the large Fanning Creek pluton mapped by Mortensen (1988) farther to the northwest near the Alaska border) is therefore the same age as volcanic rocks of the Mount Nansen Group and much of the Dawson Range batholith in the Dawson Range northwest of Carmacks (Carlson, 1987).

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GSC 91-103 Hornblende
95.4 ± 1.7 Ma

Wt. % K= 0.822
Rad. Ar= 3.129×10^{-6} cm^3/g
K-Ar 4092 % Atmos. Ar= 10.4

From a medium grained biotite-quartz-feldspar crystal tuff.
(105 J/04) Outcrop is a roadcut on the North Canol Road, elevation 3300', about 25 km bearing east-northeast of the town of Ross River, Yukon; 62°3'54"N, 131°58'40"W, UTM zone 9, 344399E, 6884790N; sample GGA-85-21B1. Collected and interpreted by S. Gordey.

See GSC 91-104 for interpretation and references.

GSC 91-104 Hornblende
112.1 ± 3.3 Ma

Wt. % K= 0.680
Rad. Ar= 3.055×10^{-6} cm^3/g
K-Ar 4095 % Atmos. Ar= 34.8

From a medium grained biotite-hornblende-quartz-feldspar crystal tuff.
(105 K/08) Outcrop at elevation 5800', about 5 km bearing north-northwest of the west end of Tay Lake, Yukon; 62°24'41"N, 132°9'33"W, UTM zone 8, 646750E, 6923050N; sample GGAG-86-31C3. Collected by S. Gareau and interpreted by S. Gordey.

These two samples (GSC 91-103, 104) were collected from the mid-Cretaceous South Fork Volcanics (Wood and Armstrong, 1982; Gordey, 1988) at widely separated localities, from densely welded crystal tuffs. The ages are interpreted as eruption ages. It is uncertain from field relations whether the two samples represent the same or different eruptive events. Although the difference in ages suggests the latter, it is unclear whether this difference could also be an artifact of dissimilar welding and cooling histories, or slight differences in alteration. A biotite age of 100 ± 2 Ma for sample GGA-85-21B1 (GSC 89-102) was reported previously (Hunt and Roddick, 1990).

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GSC 91-105 Whole rock
51.8 ± 0.9 Ma

Wt % K= 1.035
Rad. Ar= 2.113×10^{-6} cm^3/g
K-Ar 4140 % Atmos. Ar= 40.1

From a basalt.
(105 E/5) From the west side of the Klondike Highway, 750 m north of Braeburn Lodge, Yukon; 61°29.2'N, 135°55.7'W; UTM zone 8, 450560E, 6817145N; sample MLB-89-364. Collected and interpreted by J.K. Mortensen.

The sample is from a disrupted outcrop of moderately vesicular, highly fractured, dark grey-brown, medium to dark chocolate brown weathering basalt. Vesicules are commonly lined with white zeolites. Some pieces appear to display poorly developed columnar jointing. This unit was mapped by Tempelman-Kluit (1984) as part of the mainly Quaternary age Selkirk Lavas. Although fresh basaltic rocks as old as 20 Ma have been documented in western Yukon (Mortensen and Roddick, 1990), the age determined for this sample is highly anomalous. Two explanations are possible; either this volcanic unit is part of a previously unrecognized Eocene volcanic episode in western Yukon, or more probably, the sample contained significant excess Ar (possibly related to the slight but pervasive alteration noted in thin section), and the age is unreliable.

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GSC 91-106 Hornblende
156.3 ± 5.0 Ma

Wt % K= 1.290
Rad. Ar= 8.213 x 10^{-6} cm^3/g
K-Ar 4255 % Atmos. Ar= 3.9

From a gneissic granodiorite.
(105 G/1) From a prominent knob on a ridge crest 9.2 km due east of the southern tip of Fire Lake, Yukon; 61°11.0'N, 130°21.4'W; UTM zone 9, 427048E, 6783761N; sample MLB-89-334. Collected and interpreted by J.K. Mortensen.

The sample is from a body of moderately foliated hornblende-biotite granodiorite of the Simpson Range plutonic suite (Mortensen and Jilson, 1985) that gives U-Pb zircon and titanite crystallization ages of about 350 Ma (Mortensen, in press). The age is interpreted as a cooling age following regional metamorphism. See GSC 91-110 for references and further discussion of this and other metamorphic cooling ages in this area.

GSC 91-107 Muscovite
104.6 ± 1.8 Ma

Wt % K= 7.941
Rad. Ar= 3.325 x 10^{-5} cm^3/g
K-Ar 4250 % Atmos. Ar= 5.7

From a gneissic granite.
(105 G/12) From the crest of a low knob 16.4 km S32°E from the mouth of the Hoole River; 61°37.4'N, 131°32.3'W; UTM zone 9, 365429E, 6834641N; sample MLB-89-320. Collected and interpreted by J.K. Mortensen.

See GSC 91-108 for discussion.

GSC 91-108 Muscovite
113.4 ± 1.8 Ma

Wt % K= 8.829
Rad. Ar= 4.015 x 10^{-5} cm^3/g
K-Ar 4198 % Atmos. Ar= 1.1

From a gneissic granite.
(105 G/11) From an exposure on the north side of the Robert Campbell Highway 5.2 km west of Mink Creek, Yukon; 61°42.4'N, 131°27.0'W; UTM zone 9, 370460E, 6843740N; sample MLB-89-328. Collected and interpreted by J.K. Mortensen.

The samples are from strongly foliated, coarse grained, feldspar augen orthogneiss of the Mink Creek orthogneiss (Mortensen and Jilson, 1985; Mortensen, in press). Samples from other portions of this unit have given Early Mississippian U-Pb zircon and Rb-Sr whole rock ages, and U-Pb monazite and Rb-Sr muscovite ages that range from 94-112 Ma. The ages are interpreted as cooling ages following regional metamorphism. See GSC 91-110 for references and further discussion of these and other metamorphic cooling ages in this area.

GSC 91-109 Hornblende
419.0 ± 12.0 Ma

Wt % K= 0.318
Rad. Ar= 5.826 x 10^{-6} cm^3/g
K-Ar 4256 % Atmos. Ar= 11.0

From a foliated quartz monzonite/ granodiorite.
(105 H/4) From a northwest-trending ridge crest, 13.8 km N65°W of Peak 6317", Yukon; 61°05.2'N, 129°53.1'W; UTM zone 9, 452264E, 6772559N; sample MLB-89-335. Collected and interpreted by J.K. Mortensen.

The sample is from massive to moderately foliated, medium grained quartz monzonite/granodiorite of the Tuchitua River pluton (Mortensen and Jilson, 1985). This body has given U-Pb zircon and titanite ages of 357 ± 1 Ma and 354 ± 1 Ma, respectively. The K-Ar hornblende age is therefore impossibly old, and must reflect the presence of a considerable amount of excess Ar. The age is considered meaningless. See GSC 91-110 for reference.

GSC 91-110 Muscovite
104.6 ± 2.0 Ma

Wt % K= 6.543
Rad. Ar= 2.739 x 10^{-5} cm^3/g
K-Ar 4254 % Atmos. Ar= 2.8

(105 G/8) From a gneissic quartz monzonite. From a south-facing slope 5.6 km due east of Peak 7721', Yukon; 61°22.0'N, 130°23.3'W; UTM zone 9, 425779E, 6804215N; sample MLB-89-326a. Collected and interpreted by J.K. Mortensen.

The sample is from a strongly foliated, locally mylonitic, medium grained orthogneiss of quartz monzonitic composition (Mortensen and Jilson, 1985; Mortensen, in press). A sample from a nearby locality has given a U-Pb zircon age of 361 Ma (Early Mississippian). The age is interpreted as a cooling age following regional metamorphism.

Metamorphic cooling ages in this portion of the Yukon-Tanana Terrane follow the same general pattern as that observed elsewhere in the terrane. Most of the terrane yields a range of mainly Jurassic cooling ages; however areas which include Early Mississippian peraluminous orthogneiss typically give Early Cretaceous cooling ages. This has been interpreted (e.g. Hansen et al., 1991) to indicate that different parts (subterranes) of the Yukon-Tanana Terrane cooled at very different rates, and that these subterranes have been juxtaposed by Early Cretaceous extensional faults.

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GSC 91-111 Hornblende
180.9 ± 2.6 Ma

Wt % K= 1.499
Rad. Ar= 1.108 x 10^{-5} cm^3/g
K-Ar 4253 % Atmos. Ar= 2.7

(105 G/10) From a hornblende granodiorite. From an east-trending spur ridge, 6.2 km west of Big Campbell Creek, Yukon; 61°32.5'N, 130°56.6'W; UTM zone 9, 396694E, 6824460N; sample MLB-89-322a. Collected and interpreted by J.K. Mortensen.

The sample is from a massive, unfoliated hornblende diorite with pink K-feldspar phenocrysts. This body gives U-Pb zircon and titanite ages of 184-189 Ma (Early Jurassic) (Mortensen and Jilson, 1985; Mortensen, in press). The K-Ar date indicates relatively rapid cooling following intrusion. See GSC 91-110 for references.

GSC 91-112 Biotite
110.1 ± 1.5 Ma

Wt % K= 7.302
Rad. Ar= 3.223 x 10^{-5} cm^3/g
K-Ar 4172 % Atmos. Ar= 2.3

See GSC 91-113 for interpretation and location.

GSC 91-113 Muscovite
103.0 ± 1.9 Ma

Wt % K= 8.595
Rad. Ar= 3.541 x 10^{-5} cm^3/g
K-Ar 4173 % Atmos. Ar= 2.8

(105 G/2) From a quartz monzonite. From a ridge crest 2.8 km due south of the south end of Fire Lake, Yukon; 61°09.5'N, 130°30.6'W; UTM zone 9, 418739E, 6781157N; sample FL-46. Collected and interpreted by J.K. Mortensen.

The sample is from a massive, unfoliated muscovite-biotite quartz monzonite of the Selwyn Plutonic Suite, which intrudes the Yukon-Tanana Terrane northeast of the Tintina Fault Zone. See GSC 91-117 for discussion of these and other ages for intrusions of this suite.

GSC 91-114 Biotite
103.4 ± 1.8 Ma

Wt % K= 7.688
Rad. Ar= 3.181 x 10^{-5} cm^3/g
K-Ar 4174 % Atmos. Ar= 2.4

See GSC 91-115 for interpretation and location.

GSC 91-115 Muscovite
103.8 ± 2.2 Ma

Wt % K= 8.719
Rad. Ar= 3.621 x 10^{-5} cm^3/g
K-Ar 4175 % Atmos. Ar= 3.4

From a quartz monzonite.

(105 G/2) From a ridge crest 3.0 km south-southeast of Peak 7721', Yukon; 61°20.4'N, 130°30.9'W; UTM zone 9, 418939E, 6801396E; sample KG-31. Collected and interpreted by J.K. Mortensen.

The sample is from a massive, unfoliated muscovite-biotite quartz monzonite of the Selwyn Plutonic Suite, which intrudes the Yukon-Tanana Terrane northeast of the Tintina Fault Zone. See GSC 91-117 for discussion of these and other ages for intrusions of this suite.

GSC 91-116 Biotite
100.0 ± 2.9 Ma

Wt % K= 7.087
Rad. Ar= 2.832 x 10^{-5} cm^3/g
K-Ar 4252 % Atmos. Ar= 1.9

From a megacrystic quartz monzonite.

(105 G/11) From a ridge crest 12.0 km due north of Peak 6959', Yukon; 61°32.1'N, 131°06.8'W; UTM zone 9, 387635E, 6823999N; sample MLB-89-321. Collected and interpreted by J.K. Mortensen.

See GSC 91-117 for interpretation and muscovite pair.

GSC 91-117 Muscovite
90.8 ± 1.4 Ma

Wt % K= 8.576
Rad. Ar= 3.104 x 10^{-5} cm^3/g
K-Ar 4251 % Atmos. Ar= 4.1

From a megacrystic quartz monzonite.

(105 G/11) From a ridge crest 12.0 km due north of Peak 6959, Yukon; 61°32.1'N, 131°06.8'W; UTM zone 9, 387635E, 6823999N; sample MLB-89-321. Collected and interpreted by J.K. Mortensen.

The sample is from a massive, unfoliated, K-feldspar megacrystic, muscovite-biotite quartz monzonite of the Selwyn Plutonic Suite, which intrudes the Yukon-Tanana Terrane northeast of the Tintina Fault Zone.

The plutons dated in this study, along with two other members of the Selwyn Plutonic Suite in this area, give very consistent U-Pb zircon and monazite crystallization ages of 112.4 Ma (Mortensen and Jilson, 1985; Mortensen, unpublished data). These ages coincide exactly with U-Pb monazite, Rb-Sr muscovite and K-Ar biotite and muscovite cooling ages for metamorphic minerals in the augen gneiss-bearing portion of the Yukon-Tanana Terrane, which forms wall rocks for two of the three plutons dated here. This indicates at least a temporal relationship between the plutons and the tectonism responsible for final uplift and cooling of this part of the terrane.

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1985: Evolution of the Yukon-Tanana terrane: evidence from southeastern Yukon Territory; Geology, v. 13, p. 806-810.

GSC 91-118 Botite
60.1 ± 0.8 Ma

Wt % K= 7.054
Rad. Ar= 1.675 x 10^{-5} cm^3/g
K-Ar 4097 % Atmos. Ar= 11.9

From a biotite quartz monzonite.

(105 L/3) From 3.2 km due north of the mouth of Bearfeed Creek, Yukon; 62°12.0'N, 135°04.9'W; UTM zone 8, 495800E, 6896300N; sample MLB-88-185. Collected and interpreted by J.K. Mortensen.

The sample is from a biotite quartz monzonite, shown by Campbell (1967) as Unit 21c. There is no evidence for thermal overprinting on this rock, and the date therefore indicates that this is an Eocene pluton. See GSC 91-119 for references and discussion of this and related dates.

GSC 91-119 Biotite
57.0 ± 0.8 Ma

Wt % K= 6.620
Rad. Ar= 1.490 x 10^{-5} cm^3/g
K-Ar 4182 % Atmos. Ar= 5.5

From a porphyry dyke.

(105 L/3) From 4.1 km due north of the mouth of Bearfeed Creek, Yukon; 62°12.4'N, 135°05.2'W; UTM zone 8, 495500E, 6897000N; sample MLB-88-182. Collected and interpreted by J.K. Mortensen.

The sample is from a N25°E trending, medium to dark green, biotite-plagioclase-hornblende porphyry dyke 1.5-2.0 m wide that cuts reddish brown weathering, amygdaloidal, olivine-pyroxene-phyric basalt. This age gives the time of emplacement for this body.

These two dates (GSC 91-118,119) indicate that Eocene igneous activity was likely widespread in this area, and included both plutonic rocks and hypabyssal intrusions.

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1967: Reconnaissance geology of Glenlyon map area, Yukon Territory (105L); Geological Survey of Canada, Memoir 352, 92p.

DISTRICT OF KEEWATIN (GSC 91-120 to GSC 91-121)

GSC 91-120 Whole rock
1395 ± 13 Ma

Wt % K= 7.959
Rad. Ar= 6.511 x 10^{-4} cm^3/g
K-Ar 4008 % Atmos. Ar= 0.1

From a hydrothermally altered metagreywacke.

(66 A/5) From a diamond drill hole END-3, N.W.T.; 64°15.5'N, 97°53'W; zone 14, 554114E 7126081N; sample END-3-295'. Collected and interpreted by A.R. Miller.

For interpretation and references see GSC 91-121.

GSC 91-121 Whole rock
1378 ± 15 Ma

Wt % K= 8.031
Rad. Ar= 6.454 x 10^{-4} cm^3/g
K-Ar 4009 % Atmos. Ar= 0.1

From a hydrothermally altered quartz+feldspar porphyry.

(66 A/5) From a diamond drill hole END-3, N.W.T.;64°15.5'N, 97°53'W; zone 14, 554114E 7126081N; sample END-3-472'. Collected and interpreted by A.R. Miller.

This sample and GSC 91-120 were collected from diamond drill hole END-3; this drill hole was drilled to define uranium mineralization in the END GRID area, Schultz Lake map area.

In the exploration area the metaturbidite sequence, considered part of the 2.8 Ga Woodburn Group, consists of fine- to medium-grained greywacke-pelite with minor banded magnetite iron-formation and has been intruded by quartz+feldspar porphyry. The latter is related to a ~1.9 Ga medium grained fluorite-bearing granite complex.

Variable hydrothermal alteration related to unconformity-type uranium mineralization has overprinted both rock units. The most intense alteration, samples END-3-295' (GSC 91-120) and END-3-472' (this sample), is represented by illite accompanied by varible but always assessory quantities of non-potassium-bearing phases. The metagreywacke at 295' consists of greater than 95% clay-sized illite with trace quantities of micron-sized hematite and zircon distributed through the illitic mat. The absence of detrital quartz grains in this sample is anomalous compared to less altered metagreywacke and metagreywacke outside the alteration envelope. The inverse relationship between detrital quartz grains and modal illite indicates that this sample lies within a zone of intense alteration, desilicification.

The quartz+feldspar porphyry at 472' has retained its excellent igneous texture inspite of intense illite replacement. The assemblage consists of illite+quartz+chlorite+hematite. Illite completely replaces feldspar phenocrysts and the matrix to phenocrysts which consisted of fine-grained feldspar and quartz. Relict anhedral quartz phenocysts remain in the illitic mat. Minor chlorite after biotite is sparsely distributed in the illitic mat. The loss of quartz at the expense of illite in the matrix to phenocrysts and the ragged anhedral form of quartz phenocrysts suggests desilicification.

The ages of 1395 ± 13 Ma and 1375 ± 15 Ma are interpreted to record the time of intense illitization of Archean metaturbidite and early Proterozoic granite. These ages are concordant with a K-Ar age on illite (Hunt and Roddick, 1988, see GSC 88-47) and a 1400 Ma U-Pb age on pitchblende from the Kiggavik uranium deposit (Fuchs and Hilger, 1989), the latter located approximately 16 km to the northeast. These ages from the END GRID area are also concordant with a K-Ar age on illite from the basal Thelon Formation (Hunt and Roddick, 1988, see GSC 88-50). The concordancy of illite ages from different lithologies having similar intense illitization, associated with uranium mineralization and their distribution over a considerable area especially in the northeastern sub-Thelon basement suggests that hydrothermal alteration and initial unconformity-type uranium mineralization may be no older than 1400 Ma.

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MANITOBA
(GSC 91-122 to GSC 91-158)

Cooling history of Lac du Bonnet Granitic Batholith, Manitoba

Samples WD3-1 to 11, (GSC 91-122 to 132) were collected from a deep borehole (about 1250m) located on the southeastern contact of the Lac du Bonnet granitic Batholith, Manitoba, NTS 62 I/7, 50°16'N, 95°51'W; UTM zone 15, 296889E 5571948N. The sample depths, ages, and K-Ar data are listed in Table 1.

All ages correspond to a late Kenoran event and show an overall younging with depth (Kamineni et al., 1991). Deviations from this overall trend suggest complex cooling of the batholith. Collected and interpreted by D.C. Kamineni.

Samples WB1-1 to 13, (GSC 91-133 to 145) were collected from a deep borehole (about 1250m) located at the centre of the Lac du Bonnet granitic Batholith, Manitoba, NTS 62 I/7, 50°16'N, 95°51'W; UTM zone 15, 296889E 5571948N. The sample depths, ages, and K-Ar data are listed in Table 2.

All ages correspond to a late Kenoran event and show an overall younging with depth (Kamineni et al., 1991). Deviations from the overall trend indicate complex cooling of the Lac du Bonnet Batholith. Collected and interpreted by D.C. Kamineni.

Samples WA1-1 to 13, (GSC 91-146 to 158) were collected from a deep borehole (about 1250m) located near the northern contact of the Lac du Bonnet Granitic Batholith, Manitoba, NTS 62 I/7, 50°16'N, 95°51'W; zone 15, 296889E 5571948N. The sample depths, ages, and K-Ar data are listed in Table 3.

All ages correspond to a late Kenoran event and, in contrast to WB1 and WD3 series samples, are older and show little or no younging with depth indicating that the northern part of the batholith cooled earlier and at a relatively uniform rate (Kamineni et al., 1991). Collected and interpreted by D.C. Kamineni.

REFERENCE

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1991: Cooling history of Lac du Bonnet Batholith, Manitoba: deduced from K-Ar biotite ages; Geological Association of Canada, Program with Abstracts, v. 16, p. A63.

Table 1. Biotite K-Ar ages from southeastern borehole

GSC #	Sample #	K-Ar #	Depth (metres)	Wt % K	Rad. Ar ($x10^{-3} cm^3/g$)	% Atmos. Ar	Age (Ma) ± 2 sigma
GSC 91-122	WD3-01	4027	12.5	7.406	1.309	0.2	2271 ± 22
GSC 91-123	WD3-02	4028	135.2	7.920	1.433	0.05	2301 ± 19
GSC 91-124	WD3-03	4029	245.7	7.679	1.351	0.3	2265 ± 20
GSC 91-125	WD3-04	4030	365.0	6.954	1.197	0.08	2237 ± 21
GSC 91-126	WD3-05	4031	485.6	7.394	1.258	0.02	2221 ± 19
GSC 91-127	WD3-06	4032	605.4	7.047	1.197	0.03	2219 ± 16
GSC 91-128	WD3-07	4033	725.0	7.443	1.213	0.1	2167 ± 17
GSC 91-129	WD3-08	4034	844.6	7.565	1.278	0.1	2213 ± 18
GSC 91-130	WD3-09	4035	964.7	7.685	1.304	0.1	2218 ± 19
GSC 91-131	WD3-10	4036	1085.0	7.337	1.183	0.1	2154 ± 16
GSC 91-132	WD3-11	4037	1119.6	7.454	1.229	0.1	2182 ± 18

Table 2. Biotite K-Ar ages from centre borehole

GSC #	Sample #	K-Ar #	Depth (metres)	Wt % K	Rad. Ar ($x10^{-3}$ cm^3/g)	% Atmos. Ar	Age (Ma) ± 2 sigma
GSC 91-133	WB1-01	4054	5.0	7.695	1.484	0.1	2384 ± 23
GSC 91-134	WB1-02	4232	100.8	7.793	1.477	0.01	2362 ± 17
GSC 91-135	WB1-03	4055	201.2	7.741	1.473	0.1	2367 ± 23
GSC 91-136	WB1-04	4233	299.0	7.417	1.362	0.03	2320 ± 20
GSC 91-137	WB1-05	4234	407.0	6.871	1.229	0.1	2286 ± 21
GSC 91-138	WB1-06	4235	500.0	6.904	1.188	0.1	2236 ± 19
GSC 91-139	WB1-07	4236	600.0	7.409	1.300	0.1	2261 ± 23
GSC 91-140	WB1-08	4237	696.2	7.182	1.230	0.4	2230 ± 16
GSC 91-141	WB1-09	4238	809.0	7.380	1.291	0.04	2258 ± 22
GSC 91-142	WB1-10	4239	900.0	7.634	1.369	0.1	2290 ± 19
GSC 91-143	WB1-11	4056	1000.8	7.538	1.326	0.1	2264 ± 24
GSC 91-144	WB1-12	4057	1103.0	7.481	1.299	0.1	2248 ± 26
GSC 91-145	WB1-13	4240	1203.2	7.676	1.323	0.1	2238 ± 22

Table 3. Biotite K-Ar ages from northern borehole

GSC #	Sample #	K-Ar #	Depth (metres)	Wt % K	Rad. Ar ($x10^{-3}$ cm^3/g)	% Atmos. Ar	Age (Ma) ± 2 sigma
GSC 91-146	WA1-01	4050	8.0	7.588	1.484	0.04	2403 ± 23
GSC 91-147	WA1-02	4223	99.1	7.704	1.480	0.1	2379 ± 17
GSC 91-148	WA1-03	4051	199.0	7.662	1.447	0.2	2357 ± 23
GSC 91-149	WA1-04	4224	310.0	7.781	1.516	0.04	2398 ± 17
GSC 91-150	WA1-05	4225	398.0	7.809	1.493	0.1	2373 ± 18
GSC 91-151	WA1-06	4226	500.5	7.751	1.478	0.02	2370 ± 17
GSC 91-152	WA1-07	4227	615.5	7.813	1.488	0.1	2368 ± 17
GSC 91-153	WA1-08	4228	706.0	7.837	1.472	0.03	2350 ± 17
GSC 91-154	WA1-09	4229	819.5	7.799	1.495	0.03	2377 ± 17
GSC 91-155	WA1-10	4230	912.4	7.783	1.497	0.03	2381 ± 17
GSC 91-156	WA1-11	4231	997.8	7.845	1.488	0.04	2363 ± 18
GSC 91-157	WA1-12	4052	1129.0	7.775	1.478	0.04	2366 ± 22
GSC 91-158	WA1-13	4053	1199.1	7.651	1.476	0.1	2385 ± 31

QUEBEC
(GSC 91-159)

GSC 91-159 Muscovite
1824 ± 17 Ma

Wt % K= 8.498
Rad. Ar= 1.042 x 10^{-3} cm^3/g
K-Ar 4099 % Atmos. Ar= 0.3

From a foliated muscovite-biotite tonalite pluton.

(35 G/16) Outcrop on west side of Lac Watts, Québec; 61°49'N, 74°15' W; UTM zone 18, 539520E 6853782N; sample SAB-87-P24. Collected and interpreted by R. Parrish, M. St-Onge and S. Lucas.

The sample was collected from a relatively large tonalite pluton which intruded layered mafic cumulates of the Watts Group prior to its collision with the Superior Province margin (St-Onge and Lucas, 1990a). The Watts Group has been dated at 1998 ± 2 Ma by U-Pb zircon geochronology, while the tonalite pluton records an approximate U-Pb age of 1870-1880 Ma (Parrish, 1989).

The pluton cross-cuts an early fault within the Watts Group but was itself deformed during collisional tectonism. The pluton is restricted to the hangingwall of a late thrust fault associated with post-thermal peak, greenschist facies metamorphism (ca. 450°C, 4-5 kb; Bégin, 1989). The K-Ar muscovite study was undertaken to provide an approximate (maximum) age for the post-thermal peak out-of-sequence thrusting in the Cape Smith Belt.

The 1824 ± 17 Ma age is older than expected for several reasons. First, K-Ar whole rock ages from Cape Smith Belt rocks (summarized in Taylor, 1982) range between 1450-1750 Ma, suggesting cooling below about 250°C during this period. Second, syn-thermal peak deformation within the Cape Smith Belt is known to be synchronous with collision of the Narsajuaq terrane (St-Onge and Lucas, 1990b; Lucas and St-Onge, 1991), dated to occur after 1825 Ma (Parrish, 1989). It is plausible that cooling of the pluton to the K-Ar closure temperature for muscovite occurred in the age range 1824 ± 17 Ma due to relatively rapid syn-faulting exhumation of the thrust stack. Another explanation of the age would be that wedging of the Narsajuaq terrane near the base of the thrust stack resulted in syn-thermal peak thrusting below the wedge and post-thermal peak faulting, possibly involving normal movement, above the wedge at retrograde conditions. In this scenario, the 1824 ± 17 Ma age may provide a minimum time constraint for the collision of Narsajuaq terrane with the Superior Province margin.

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NEW BRUNSWICK
(GSC 91-160 to GSC 91-166)

GSC 91-160 Biotite
1082 ± 13 Ma

Wt % K = 7.692
Rad. Ar = 4.434 x 10^{-4} cm^3/g
K-Ar 4138 % Atmos. Ar. = 1.2

From a granitoid rock.

(21 N/8) Sampled form a large gravel pit off west side of Hwy. No. 2 in the Madawaska River valley and just north of St. Jacques in northwestern New Brunswick; 47°25'N, 65°48'W; UTM zone 20, 288789E 5255051N; sample PC 2/87. Collected by V. Prest and M. Rappol.

This sample is a granulite facies granitoid rock consisting of quartz, fresh plagioclase and perthite, considerable ortho- and clinopyroxene, brown biotite, and minor olive green hornblende.

The sample was a glacial erratic boulder from valleyside gravels overlying till and a polished and striated Devonian bedrock surface. The abundance and variety of granitoid and other erratics strongly suggests a source area to the northwest, i.e. southeast-flowing Laurentide ice. The well preserved glacial striae, however, record a major flow to the northwest (310°) with a younger flow to the northeast (35°). The K-Ar date confirms the presence of southeast-flowing Laurentide ice. The sets of striae therefore reflect the outward flow of Appalachian ice during the full and retreatal phases of the last glaciation.

Rappol (pers. comm., 1990) reported southeasterly-trending striae at several nearby sites that pre-date the northwesterly striae.

GSC 91-161 Hornblende
1004 ± 12 Ma

Wt. % K= 1.235
Rad. Ar= 6.450 x 10^{-5} cm^3/g
K-Ar 4135 % Atmos. Ar= 0.34

From a granitoid rock.
(21 J/13) Sampled from a boulder pile on north side of road on south side of Thompkins Ridge, south of Bell Grove, New Brunswick; 46°51.3'N, 67°35.3'W; UTM zone 19, 607615E 5189800N; sample PC 3/87. Collected by V. Prest and M. Rappol.

This sample is a retrogressed granulite facies granitoid rock with practically no alteration of the feldspar; olive-brown to green hornblende with scant alteration to biotite along fractures. There are patches of what was probably early-formed pyroxene, now altered to biotite and epidote aggregates.

This sample was a small boulder of a typical granitoid rock from a pile of field stone believed to be within the area of a south-flowing lobe of Laurentide ice in the Saint John River valley south of Grand Falls. The Proterozoic age determination confirms the influence of Laurentide ice.

GSC 91-162 Biotite
950 ± 21 Ma

Wt % K= 5.318
Rad. Ar= 2.586 x 10^{-4} cm^3/g
K-Ar 4139 % Atmos. Ar= 3.5

From a granitoid rock.
(21 J/13) Sampled from a boulder-strewn field on west side of road, south of Bells Grove, New Brunswick; 46°55'N, 67°33.5'W; UTM zone 19, 609776E 5196694N; sample PC 5/87. Collected by Rappol and Prest.

A fresh-looking granitoid rock with clean microcline-perthite; quartz only 10 to 15 percent; scant brown biotite with some alteration to chlorite and some relict hornblende.

Numerous cobble to boulder size erratics are believed to lie within the area covered by a lobe of Laurentide ice in the Saint John River valley during the Late Wisconsinan. The K-Ar Proterozoic dating confirms the presence of Laurentide ice in this area.

GSC 91-163 Hornblende
424 ± 6 Ma

Wt. % K= 0.592
Rad. Ar.= 1.099 x 10^{-5} cm^3/g
K-Ar 4137 % Atmos. Ar= 3.4

From a granite cobble.
(21 J/4) Sampled from a borrow pit in terrace deposits on south side of Meduznekead River at bridge crossing to Belleville, about 11 km NW of Woodstock, New Brunswick; 46°10'N, 67°43'W; UTM zone 19, 599075E 5113148N; sample PC 7b/87. Collected by V. Prest.

This sample has large oscillatory zoned, lath-shaped, plagioclase crystals together with large stubby zoned feldspar; groundmass plagioclase is considerably altered (mica). There are large tabular and stubby hornblende crystals and good brown biotite.

This is a rare 'granite' cobble amongst the several basic and highly altered granitoid rocks and varied gneisses. Site is just 1.4 km north of the bouldery area (3.4 km north of Richmond Corner) that may be an end moraine. This deposit may relate to a lobe of Laurentide ice in the Saint John River valley but it is more likely from a lobe of southeast flowing ice in the Meduznekead River valley, i.e. from Maine. Silurian-age rocks occur both north and northwest of the sample site. This is from a definite magmatic rock and a Silurian age is quite acceptable.

GSC 91-164 Hornblende
702 ± 10 Ma

Wt. % K= 0.483
Rad. Ar= 1.610 x 10^{-5} cm^3/g
K-Ar 4134 % Atmos Ar = 18.7

From a granitoid glacial erratic cobble.
(21 I/3) Sampled near Dover on the east side of the Peticodiac River, New Brunswick. Specimen is representative of the few erratics in this area; 46°00'N, 64°41.5'W; UTM zone 20, 369007E 5095221N; sample PC 9/87. Collected by A. Seaman (#85-10), submitted by V. Prest.

This coarse textured granitoid rock has abundant, highly saussuritized (leucoxene, mica, epidote) feldspar and large clean, green hornblende in a quartz groundmass.

The Proterozoic date may well indicate that the cobble was derived by northward ice flow from the nearby Caledonia Highlands. Such northward flow has not been generally recognized, but has been suggested by Seaman (pers. comm., 1987).

GSC 91-165 Biotite
384 ± 6 Ma

Wt. % K= 7.490
Rad. Ar= 1.245×10^{-4} cm^3/g
K-Ar 4136 % Atmos. Ar= 1.7

From a granite boulder.
(21 I/4) Sampled on a road-side cut through low till knob 4 km west of Canaan Forks bridge on Hwy. 112. Same site as GSC 90-113 (sample PC 14/87, see Hunt and Roddick, 1991) and lying within the Pennsylvanian basin, New Brunswick; 46°01'N, 65°36'W; UTM zone 20, 298733E 5098968N; sample PC 16/87. Collected by V. Prest.

Sample has a quartz-rich groundmass, with small tabular crystals of plagioclase, clean, brown biotite interspersed throughout with only minor alteration to chlorite.

This fine grained, equigranular granite boulder from the clayey sand till knoll, is associated with numerous other large boulders and cobbles of granitoid and volcanic rocks including a rhyolite porphyry. No glacial erratics were noted for many kilometres east of the till knob but immediately to the west and over a distance of 8 km fieldstones are common. Farther west and northwest they are again rare to absent.

The K-Ar dating of both this sample and PC-14-87 (GSC 90-113, Hunt and Roddick, 1991) confirms the presence of Devonian-age erratics along Hwy. 112 over a width of about 8 km. The most logical source is the Mississippian-age coarse conglomerate some 15 to 20 km to the south and forming the north limb of the Caledonia Highlands. Alternatively the erratics could have been derived from a small stock at the eastern end of Cumberland Bay, Grand Lake, though this stock has been assigned to the Mississippian.

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GSC 91-166 Biotite
409 ± 13 Ma

Wt. % K= 5.417
Rad. Ar= 9.663×10^{-5} cm^3/g
K-Ar 4133 % Atmos. Ar= 2.0

From a granitoid boulder.
(21 I/13) Roadside boulders on Hwy. 420 at junction with logging road at about 10 km west of Sillikers, New Brunswick; 46°54'N, 65°58'W; UTM zone 20, 274037E 5198106N; sample PC 19/87. Collected by V. Prest.

Site is only 15 to 20 km south of Devonian outcrops forming part of the New Brunswick Hylands. The age dating confirms the transport of erratics from the New Brunswick Highlands by southeast flowing ice, which is in accord with striae evidence. It appears unlikely that Laurentide ice invaded this region.

LABRADOR
(GSC 91-167 to GSC 91-172)

GSC 91-167 Biotite
927 ± 18 Ma

Wt % K= 4.604
Rad. Ar= 2.169×10^{-4} cm^3/g
K-Ar 4084 % Atmos. Ar= 1.5

From a granite.
(13 A/3) Headwaters of St. Lewis river, southeast Labrador; 52°05'N, 57°21.7'W; UTM zone 21, 475217E 5770150N; sample CG86-697B. Collected and interpreted by C.F. Gower and W.D. Loveridge.

The K-Ar dates GSC 91-167 to 172 are from three plutons located within the Pinware terrane of eastern Labrador (Table 4). The results presented in Table 4 are excerpted from a larger study, of Grenvillian magmatism in the eastern Grenville province (Gower et al., in press).

The ages of intrusion of the three plutons, as measured by U-Pb on zircon, fall within a limited range 956 ± 1 to 966 ± 3 Ma (Table 4). The U-Pb data on zircon from the

Upper St. Lewis River (east) monzonite and the Upper St. Lewis River (west) granite are essentially concordant but U-Pb measurements on zircon from the two River Bujeault headwaters quartz syenite samples fall on a discordia line, indicating the presence of inherited zircon of probable age 1530 ± 30 Ma.

Rb-Sr dates on biotite record disturbances to the Rb-Sr systems up to 160 Ma after Grenvillian plutonism. The K-Ar ages on biotite from GSC 91-171 (CG86-698) and GSC 91-167 (CG86-697B) are interpreted (Gower et al., in press) as being anomalously high due to the presences of excess ^{40}Ar for the following reasons:
1) both K-Ar ages are considerably higher than Rb-Sr ages on the same biotite.
2) the K-Ar age on biotite GSC 91-171 is (nominally) higher than the K-Ar age on coexising hornblende, opposite to the usual sequence.
3) U-Pb results on zircon from both rocks indicate the presence of inherited zircon; incorporation of pre-existing rock in these melts provides a source for excess ^{40}Ar.

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Table 4. Comparison of U-Pb, K-Ar and Rb-Sr geochronological data (Ma)

Unit	U-Pb zircon	K-Ar hornblende	K-Ar biotite	Rb-Sr biotite
GSC 91-169, 168 (CG84-195) Upper St Lewis River (east) monzonite	966 ± 3	946 ± 13	914 ± 13	
GSC 91-167 (CG86-697B) Riviere Bujeault headwaters quartz syenite	964 ± 5		927 ± 18	811 ± 9 822 ± 9
GSC 91-170, 171 (CG86-698) Riviere Bujeault headwaters quartz syenite	964 ± 5	926 ± 16	953 ± 12	888 ± 12
GSC 91-172 (CG86-700) Upper St Lewis River (west) granite	956 ± 1		911 ± 10	

GSC 91-168 Biotite
914 ± 13 Ma

Wt % K= 7.296
Rad. Ar= 3.377 x 10^{-4} cm^3/g
K-Ar 3963 % Atmos. Ar= 0.4

From a massive monzonite.
(13 A/3) Tributary of Upper St. Lewis River, southeast Labrador; 52°13'N, 57°2.5'W; UTM zone 21, 497153E 5784920N; sample CG84-195. Collected and interpreted by C.F. Gower and W.D. Loveridge.

See GSC 91-169 for hornblende pair and GSC 91-167 for interpretation and references.

GSC 91-169 Hornblende
946 ± 13 Ma

Wt % K= 0.892
Rad. Ar= 4.313 x 10^{-5} cm^3/g
K-Ar 3964 % Atmos. Ar= 1.8

From a massive monzonite.
(13 A/3) Tributary of Upper St. Lewis River, southeast Labrador; 52°13'N, 57°2.5'W; UTM zone 21, 497153E 5784920N; sample CG84-195. Collected and interpreted by C.F. Gower and W.D. Loveridge.

See GSC 91-168 for biotite pair and GSC 91-167 for interpretation and references.

GSC 91-170 Hornblende
926 ± 16 Ma

Wt % K= 0.615
Rad. Ar= 2.892 x 10^{-5} cm^3/g
K-Ar 3973 % Atmos. Ar= 5.8

From a alkali-feldspar quartz syenite.
(13 A/3) Headwaters of St. Lewis river, southeast Labrador; 52°6.2'N, 57°2.5'W; UTM zone 21, 497146E 5772314N; sample CG86-698. Collected and interpreted by C.F. Gower and W.D. Loveridge.

See GSC 91-171 for biotite pair and GSC 91-167 for interpretation and references.

GSC 91-171 Biotite
953 ± 12 Ma

Wt % K= 7.066
Rad. Ar= 3.448 x 10^{-4} cm^3/g
K-Ar 3974 % Atmos. Ar= 2.3

From an alkali-feldspar quartz syenite.
(13 A/3) Headwaters of St. Lewis river, southeast Labrador; 52°6.2'N, 57°2.5'W; UTM zone 21, 497146E 5772314N; sample CG86-698. Collected and interpreted by C.F. Gower and W.D. Loveridge.

See GSC 91-170 for hornblende pair and GSC 91-167 for interpretation and references.

GSC 91-172 Biotite
911 ± 10 Ma

Wt % K= 7.576
Rad. Ar= 3.492 x 10^{-4} cm^3/g
K-Ar 3975 % Atmos. Ar= 0.5

From a massive, coarse grained granite.
(13 A/3) Headwaters of St. Lewis river, southeast Labrador; 52°14.5'N, 57°20.4'W; UTM zone 21, 476784E 5787754N; sample CG86-700. Collected and interpreted by C.F. Gower and W.D. Loveridge.

See GSC 91-167 for interpretation and references.

Outside Canada
People's Republic of China
(GSC 91-173 to GSC 91-187)

GSC 91-173 Biotite
122.8 ± 2.1 Ma

Wt % K= 6.870
Rad. Ar= 3.393 x 10^{-5} cm^3/g
K-Ar 4106 % Atmos. Ar= 1.9

From a biotite granite.
From 5 km west-southwest of the Linglong gold mine, Zhaoyuan area, eastern Shandong Province, People's Republic of China; sample MLB-89-133c. Collected and interpreted by J.K. Mortensen.

The sample is from a weakly foliated biotite granite phase of the Linglong Granite, which gives U-Pb zircon and monazite ages of 154 Ma (J.K. Mortensen, unpublished data; K.H. Poulsen et al., unpublished data). The date is interpreted as a thermal reset age. See GSC 91-177 for references and discussion of this and related samples.

GSC 91-174 Biotite
126.9 ± 4.6 Ma

Wt % K= 6.598
Rad. Ar= 3.371 x 10^{-5} cm^3/g
K-Ar 4107 % Atmos. Ar= 1.9

From a biotite-muscovite granite.
From 6 km south of the Linglong gold mine, Zhaoyuan area, eastern Shandong Province, People's Republic of China; sample MLB-89-134. Collected and interpreted by J.K. Mortensen.

The sample is from massive, unfoliated, medium grained biotite-muscovite granite of the Luanjiahe Granite, which gives U-Pb zircon and monazite ages of 154 Ma (Mortensen, unpublished data; K.H. Poulsen et al., unpublished data). The date is interpreted as a thermal reset age. See GSC 91-177 for references and discussion of this and related samples.

GSC 91-175 Biotite
123.7 ± 2.5 Ma

Wt % K= 7.144
Rad. Ar= 3.557×10^{-5} cm^3/g
K-Ar 4105 % Atmos. Ar= 8.6

See GSC 91-176 for location, interpretation, and hornblende pair.

GSC 91-176 Hornblende
158.0 ± 2.3 Ma

Wt % K= 1.011
Rad. Ar= 6.489×10^{-6} cm^3/g
K-Ar 4104 % Atmos. Ar= 18.6

From a hornblende-biotite granodiorite. From 12 km northeast of the Linglong gold mine, Zhaoyuan area, eastern Shandong Province, People's Republic of China; sample MLB-89-135a. Collected and interpreted by J.K. Mortensen.

The sample is from massive, unfoliated, medium grained hornblende-biotite granodiorite of the Guajialing Granite, which gives U-Pb zircon and titanite ages of 124 Ma (J.K. Mortensen, unpublished data; K.H. Poulsen et al., unpublished data). The biotite age (GSC 91-175) is interpreted to reflect cooling soon after emplacement. The hornblende age (this sample) is anomalously old, and likely reflects the presence of excess Ar in the sample. See GSC 91-177 for references and discussion of this and related samples.

GSC 91-177 Biotite
125.3 ± 1.8 Ma

Wt % K= 7.240
Rad. Ar= 3.654×10^{-5} cm^3/g
K-Ar 4111 % Atmos. Ar= 4.0

From a hornblende-biotite granodiorite. From 9 km northeast of the Jiaojia gold mine, Zhaoyuan area, eastern Shandong Province, People's Republic of China; sample MLB-89-149. Collected and interpreted by J.K. Mortensen.

The sample is from massive, unfoliated, medium grained hornblende-biotite granodiorite of the Guajialing Granite that contains abundant K-feldspar phenocrysts to 15 cm in length. This sample gives a U-Pb zircon age of 127 Ma (J.K. Mortensen, unpublished data; K.H. Poulsen et al., unpublished data). The age is interpreted to reflect cooling soon after emplacement.

This K-Ar data for granitic rocks in the Zhaoyuan area (GSC 91-173 to 177), together with U-Pb zircon, monazite and titanite ages for the same samples, resolve some long-standing problems concerning the age of intrusive rocks and contained gold mineralization in this part of Shandong Province (Poulsen et al., 1990). The Linglong Granite is the main host for disseminated and vein gold mineralization in the area, and had previously been considered to be Precambrian in age. These data clearly indicate that the Linglong Granite, and the closely related Luanjiahe Granite were emplaced at about 155 Ma (Late Jurassic), and were subsequently themselves intruded by the Guajialing Granite at about 125-127 Ma. The emplacement of these younger granites thermally reset the K-Ar biotite ages in the older phases.

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GSC 91-178 Hornblende
1995 ± 20 Ma

Wt % K= 1.046
Rad. Ar= 1.483×10^{-4} cm^3/g
K-Ar 4261 % Atmos. Ar= 0.2

From a mafic gniess. From 42 km southeast of Zhaoyuan, eastern Shandong Province, People's Republic of China; sample MLB-89-132b. Collected and interpreted by J.K. Mortensen.

The sample is from a biotite-hornblende-plagioclase gneiss of the lower Minshan Formation, Jiaodong Group. See GSC 91-181 for references and discussion of this and related samples.

GSC 91-179 Hornblende
1823 ± 18 Ma

Wt % K= 1.050
Rad. Ar= 1.286×10^{-4} cm^3/g
K-Ar 4262 % Atmos. Ar= 0.1

From a mafic gneiss.
From 42 km southeast of Zhaoyuan, eastern Shandong Province, People's Republic of China; sample MLB-89-132a. Collected and interpreted by J.K. Mortensen.

The sample is from a hornblende-plagioclase gneiss of the lower Minshan Formation, Jiaodong Group. See GSC 91-181 for references and discussion of this and related samples.

GSC 91-180 Biotite
988 ± 13 Ma

Wt % K= 7.250
Rad. Ar= 3.706 x 10^{-4} cm^3/g
K-Ar 4163 % Atmos. Ar= 0.4

From a pelitic schist.
From 40 km southeast of Zhaoyuan, eastern Shandong Province, People's Republic of China; sample MLB-89-131b. Collected and interpreted by J.K. Mortensen.

The sample is from a garnetiferous biotite-cordierite-sillimanite schist of the upper Pengkuan Formation, Jiaodong Group. See GSC 91-181 for references and discussion of this and related samples.

GSC 91-181 Hornblende
684 ± 9 Ma

Wt % K= 0.597
Rad. Ar= 1.930 x 10^{-5} cm^3/g
K-Ar 4263 % Atmos. Ar= 2.1

From a mafic gneiss.
From 10 km southeast of Zhaoyuan, eastern Shandong Province, People's Republic of China; sample MLB-89-130. Collected and interpreted by J.K. Mortensen.

The sample is from an amphibolite dyke intruding the Pengkuan Formation, Jiaodong Group.

This set of four K-Ar ages (GSC 91-178 to 181) from the Zhaoyuan area is part of an investigation of the geology and origin of gold mineralization in this region (Poulsen et al., 1990, and unpublished data). Most previous workers have concluded that the granitoid rocks in the area were generated during the upper amphibolite to granulite grade metamorphism that affected the Jiaodong Group. Data reported here demonstrate that peak metamorphic conditions occurred at about 1850 Ma, but most if not all of the granitic intrusions were emplaced in mid-Mesozoic time. There is little evidence of any regional thermal influence associated with the plutonism, as metamorphic ages from the Jiaodong Group have not been significantly reset.

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GSC 91-182 Biotite
123.9 ± 1.9 Ma

Wt % K= 7.128
Rad. Ar= 3.554 x 10^{-5} cm^3/g
K-Ar 4241 % Atmos. Ar= 3.7

From a rhyolite dyke.
From 1.5 km south of Jinchanggouliang gold mine, Inner Mongolia, People's Republic of China; sample TAA-89-140. Collected by B.E. Taylor and interpreted by J.K. Mortensen.

The sample is from a biotite-feldspar-phyric porphyry dyke that cuts gold vein mineralization in underground workings at the Jinchanggouliang gold mine. The date gives the age of emplacement of the dyke, and sets a minimum age for the gold veining (Poulsen et al., 1990, and unpublished data; Lin Baoqin et al., in press). See GSC 91-187) for references and discussion of this and related samples.

GSC 91-183 Biotite
127.8 ± 1.9 Ma

Wt % K= 7.316
Rad. Ar= 3.766 x 10^{-5} cm^3/g
K-Ar 4242 % Atmos. Ar= 1.9

From a granodiorite.
From 4 km southeast of the Jinchanggouliang gold mine, Inner Mongolia, People's Republic of China; sample MLB-89-78. Collected and interpreted by J.K. Mortensen.

The sample is from the hornblende-biotite granodiorite main phase of the Xiduimiangou pluton, which gives a U-Pb zircon age of 131 Ma (Poulsen et al, 1990, and unpublished data; Lin Baoqin et al., in press). The date gives the age of emplacement of the pluton. See GSC 91-187 for references and discussion of this and related samples.

GSC 91-184 Biotite
120.2 ± 1.9 Ma

Wt % K= 7.049
Rad. Ar= 3.407 x 10^{-5} cm^3/g
K-Ar 4243 % Atmos. Ar= 2.1

From a felsic tuff.
From 1.2 km northeast of the Jinchanggouliang gold mine, Inner Mongolia, People's Republic of China; sample MLB-89-99. Collected and interpreted by J.K. Mortensen.

The sample is from a biotite-phyric felsic tuff, which gives a maximum possible U-Pb zircon age of 129 Ma (Poulsen et al, 1990 and unpublished data; Lin Baoqin et al., in press). The K-Ar date gives the age of eruption of the tuff. See GSC 91-187 for references and discussion of this and related samples.

GSC 91-185 Biotite
149.1 ± 2.7 Ma

Wt % K= 6.569
Rad. Ar= 3.969 x 10^{-5} cm^3/g
K-Ar 4244 % Atmos. Ar= 3.2

See GSC 91-186 for location, interpretation and hornblende pair.

GSC 91-186 Hornblende
202.8 ± 2.8 Ma

Wt % K= 0.480
Rad. Ar= 4.005 x 10^{-6} cm^3/g
K-Ar 4245 % Atmos. Ar= 9.1

From a granodiorite.
From 0.5 km west of the Erdaogou gold mine, western Liaoning Province, People's Republic of China; sample MLB-89-126. Collected and interpreted by J.K. Mortensen.

The sample is from the Laoshang hornblende-biotite granodiorite, which forms a small plug intruding latest Jurassic felsic volcanic rocks dated by U-Pb zircon at 145 Ma (Poulsen et al, 1990 and unpublished data; Lin Baoqin et al., in press). A K-Ar hornblende age of 157 Ma had been obtained previously for this body by the No. 6 Geological Team of Liaoning Province (Shang Ling, personal communication, 1990). Field relationships indicate that the hornblende ages are geologically too old, and presumably reflect the presence of excess Ar in the samples. The K-Ar biotite date is within error of the U-Pb age for the volcanic wall rocks, suggesting that the granodiorite plug is likely a synvolcanic pluton. See GSC 91-187 for references and discussion of this and related samples.

GSC 91-187 Biotite
207.2 ± 3.1 Ma

Wt % K= 7.224
Rad. Ar= 6.165 x 10^{-5} cm^3/g
K-Ar 4246 % Atmos. Ar= 1.2

From a quartz monzonite.
From 2 km southwest of the Erdaogou gold mine, western Liaoning Province, People's Republic of China; sample MLB-89-72. Collected and interpreted by J.K. Mortensen.

The sample is from a weakly foliated, coarse grained, porphyritic biotite quartz monzonite that underlies a large area west of Erdaogou. This body dated by U-Pb zircon at 220 Ma (Poulsen et al, 1990 and unpublished data; Lin Baoqin et al., in press). The K-Ar biotite date either reflects relatively slow cooling following intrusion of the body, or thermal disturbance related to regional metamorphism and deformation and/or intrusion of younger plutons in the area.

This set of K-Ar ages (GSC 91-182 to 187) from the Erdaogou/Jinchanggouliang area is part of an investigation of the geology and origin of gold mineralization in this region (Poulsen et al., 1990 and unpublished data; Lin Baoqin et al., in press). These ages, together with U-Pb zircon and titanite data from the same samples (K.H. Poulsen et al., unpublished data; J.K. Mortensen, unpublished data) indicate that intermediate to felsic magmatism occurred over an extended period in this area, from at least 220-123 Ma. Field relationships demonstrate that gold veins formed at about 123-129 Ma, and are likely genetically related to the Xiduimiangou pluton.

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APPENDIX

The numbers listed below refer to the individual sample determination numbers, e.g. (GSC) 62-189, published in the Geological Survey of Canada age reports listed below:

	Determinations		Determinations
GSC Paper 60-17, Report 1	59-1 to 59-98	GSC Paper 74-2, Report 12	73-1 to 73-198
GSC Paper 61-17, Report 2	60-1 to 60-152	GSC Paper 77-2, Report 13	76-1 to 76-248
GSC Paper 62-17, Report 3	61-1 to 61-204	GSC Paper 79-2, Report 14	78-1 to 78-230
GSC Paper 63-17, Report 4	62-1 to 62-190	GSC Paper 81-2, Report 15	80-1 to 80-208
GSC Paper 64-17, Report 5	63-1 to 63-184	GSC Paper 82-2, Report 16	81-1 to 81-226
GSC Paper 65-17, Report 6	64-1 to 64-165	GSC Paper 87-2, Report 17	87-1 to 87-245
GSC Paper 66-17, Report 7	65-1 to 65-153	GSC Paper 88-2, Report 18	88-1 to 88-105
GSC Paper 67-2A, Report 8	66-1 to 66-176	GSC Paper 89-2, Report 19	89-1 to 89-135
GSC Paper 69-2A, Report 9	67-1 to 67-146	GSC Paper 90-2, Report 20	90-1 to 90-113
GSC Paper 71-2, Report 10	70-1 to 70-156	GSC Paper 91-2, Report 21	90-1 to 90-187
GSC Paper 73-2, Report 11	72-1 to 72-163		

GSC Age Determinations Listed by N.T.S. Co-ordinates

1-M	62-189, 190; 63-136, 137; 66-170, 171; 70-145, 146, 147, 152
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13-M/3	87-19
13-M/6	87-6

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