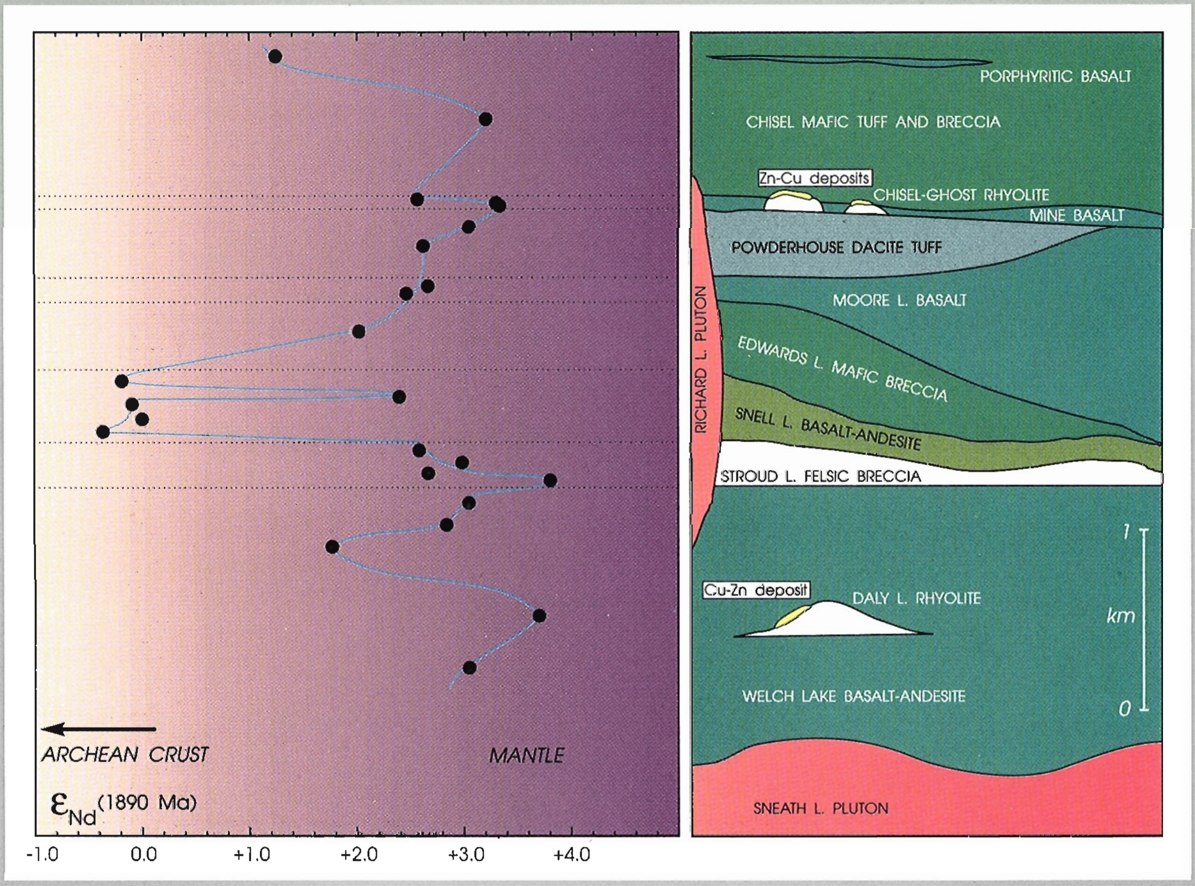
GEOLOGICAL SURVEY OF CANADA
PAPER 92-2

RADIOGENIC AGE AND ISOTOPIC STUDIES: REPORT 6



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REPORT 6

1992

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Cover description

Illustrated is a neodymium-isotopic profile of the volcanic rock stratigraphy through the 1.89 Ga Chisel Lake section of the Flin Flon belt, Manitoba. The isotopic variation suggests a complex origin for the magmas in this ancient volcanic arc, involving partial melting of the sub-oceanic mantle and varying extents of contamination by Archean (>2.5 Ga) crust (see discussion by Stern and others, this volume). Colour illustration courtesy of Richard Stern.

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CONTENTS

RADIOGENIC AGE AND ISOTOPIC STUDIES: REPORT 6

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 6 contains papers from regions across Canada. This year we have added two new additions to our report. Miscellaneous ^{40}Ar-^{39}Ar ages with procedures will be published annually similar to the compilation of K-Ar ages. In this paper, results of ^{40}Ar-^{39}Ar step heating analyses will be presented. These analyses may be minor components of larger studies or single analyses from specialized problems, and in order to make the data available in a timely fashion, the results are published here with a brief geological description and interpretation. Also in this report is a listing which includes a list of authors and contents pertaining to all our reports. We thank T. Barker for compiling this index.

The geochronology section over the years has been fortunate to have such high calibre students come to work in our laboratory. These include COSEP summer students and students from the University of Waterloo CO-OP program. We would like to thank all past and present students for their excellent work.

INTRODUCTION

La parution intitulée *Radiogenic Age and Isotopic Studies* fait partie d'une collection de rapports publiés annuellement, qui présentent des données provenant du Laboratoire de géochronologie (rattaché à la Division géoscientifique du continent de la Commission géologique du Canada). Le but principal de la collection est de fournir rapidement à la communauté géologique les données géochronologiques et les autres données sur les isotopes radiogéniques produites par le laboratoire. On trouve dans ces rapports une présentation complète des données, une description du lien qui existe entre ces données et leur contexte géologique, ainsi qu'une interprétation relativement courte des résultats. Le lecteur doit toutefois savoir que certaines données reproduites dans le présent document proviennent de travaux en cours et que des publications plus détaillées pourraient suivre. D'autres données géochronologiques et isotopiques de laboratoire, publiées dans diverses revues scientifiques ou faisant l'objet d'une publication séparée de la CGC, sont résumées à la fin du document.

Ce rapport, le sixième, contient des articles provenant de différentes régions du Canada et comprend plusieurs ajouts. D'abord, diverses descriptions d'analyses par la méthode ^{40}Ar-^{39}Ar, qui seront publiées annuellement comme c'est déjà le cas pour les datations K-Ar, mais aussi les résultats d'analyses d'échantillons soumis à des paliers de température par la méthode ^{40}Ar-^{39}Ar. Ces analyses peuvent faire partie d'études plus vastes ou être les seules de ce type effectuées dans le cadre de travaux visant des problèmes précis; pour que les données puissent être accessibles rapidement, les résultats ne sont accompagnés que d'une brève description de la géologie et d'une interprétation sommaire. Ce rapport contient en dernier lieu un index qui fait une liste des auteurs, des sujets et du contenu de tous les rapports de la collection publiés auparavant. Nous aimerions remercier T. Barker d'avoir compilé cet index et B. Blair de nous avoir donné accès à la liste de mots-clés de la base de données du centre GEOSCAN.

La Section de la géochronologie a eu, au cours des ans, la chance de recevoir dans son laboratoire des étudiants de haut calibre, notamment des étudiants participant au PEEAC et des étudiants inscrits au programme coopératif de l'université de Waterloo. Nous aimerions donc remercier tous les étudiants d'hier et d'aujourd'hui de leur excellent travail.

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 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 prominently 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.

Le Laboratoire de géochronologie dépend non seulement des ressources financières et des compétences scientifiques de la Division géoscientifique du continent, dont il fait partie, mais également d'autres groupes de la Commission géologique. La Section de la minéralogie de la Division des ressources minérales, en particulier, se charge de la séparation des minéraux et de la réduction en poudre des échantillons généralement volumineux (de 10 à 30 kg), un travail fastidieux qui demande une préparation soignée. À cette fin, nous aimerions remercier G. Gagnon, B. Machin, R. Christie et R. Delabio. La Section de minéralogie nous fournit également d'excellentes photomicrographies par balayage électronique de grains de minéraux, qui servent aux études morphologiques. Certaines d'entre elles ont été avantageusement reproduites sur la page couverture de rapports précédents de la collection. Nous tenons à remercier A. Roberts, D. Walker et L. Radburn de leur aide en minéralogie. Ne reste qu'à souligner l'apport de la Section de chimie analytique de la Division des ressources minérales, qui nous a permis d'utiliser un spectromètre d'absorption atomique pour effectuer des analyses du potassium en vue de datations par la méthode K-Ar. Nous remercions donc chaleureusement tous ces collaborateurs.

Patricia A. Hunt
Réginald J. Thériault

- 1 -

New U-Pb and ^{40}Ar-^{39}Ar age determinations from northern Ellesmere and Axel Heiberg islands, Northwest Territories and their tectonic significance

H.P. Trettin[1], R.R. Parrish[2], and J.C. Roddick[2]

Trettin, H.P., Parrish, R.R., and Roddick, J.C., 1992: New U-Pb and ^{40}Ar-^{39}Ar age determinations from northern Ellesmere and Axel Heiberg islands, Northwest Territories and their tectonic significance; in Radiogenic Age and Isotopic Studies: Report 6; Geolocial Survey of Canada, Paper 92-2, p. 3-30.

Abstract

New U-Pb and ^{40}Ar-^{39}Ar age determinations from northern Ellesmere and Axel Heiberg islands refine the history of granitoid plutonism and related tectonic events in the northern part of the Arctic Islands.

A granodiorite at Phillips Inlet yielded a zircon age of 965 ± 2 Ma, confirming the Grenville age of the Mitchell Point belt of the crystalline basement of the Pearya Terrane.

Dating of a syntectonic granodiorite at Ayles Fiord (monazite age 475 ± 1 Ma, ^{40}Ar-^{39}Ar biotite plateau age 467.8 ± 2.5 Ma), shifts the onset of the M'Clintock Orogeny of the Pearya Terrane from the early Middle Ordovician, as previously assumed, to the Early Ordovician.

Small, post-tectonic high-level plutons of variable compositions are common in northern Axel Heiberg and Ellesmere islands. Late Devonian ages have been obtained for a tonalite pluton (zircon age 368 ± 3 Ma, ^{40}Ar-^{39}Ar biotite age 360 ± 2 Ma) and a trachyandesite plug (zircon age 368 ± 3 Ma) both in northern Axel Heiberg Island, and also for a tonalite pluton in northwestern Ellesmere Island (^{40}Ar-^{39}Ar hornblende plateau age 364 ± 2 Ma). All three intrusions are probably slightly younger than the deformation of a foreland basin farther south in the Arctic Islands (Ellesmerian Orogeny, latest Devonian-Early Carboniferous).

A small monzodiorite pluton at Henson Bay, northwestern Ellesmere Island, was emplaced 334.5 ± 1.5 Ma ago (zircon and titanite), nearly coeval with deposition of the upper Visean Emma Fiord Formation, the oldest known unit of the Sverdrup Basin. This is the first evidence that thermal uplift or initial rifting of the Sverdrup Basin was accompanied by magmatism.

A quartz monzodiorite-granodiorite pluton on Marvin Peninsula, northernmost Ellesmere Island has yielded a ^{40}Ar-^{39}Ar hornblende plateau age of 94.2 ± 0.7 Ma, confirming approximate correlation with the large bimodal 92 ± 1 Ma Wootton intrusion of northwestern Ellesmere Island. Together with the Strand Fiord volcanic formation of western Axel Heiberg Island, these plutons demonstrate a major rift event in Late Cretaceous (Cenomanian) time.

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Résumé

Les nouvelles datations par les méthodes U-Pb et ^{40}Ar-^{39}Ar de roches provenant du nord de l'île d'Ellesmere et de l'île Axel Heiberg permettent d'affiner l'histoire du plutonisme granitoïde et des événements tectoniques associés dans la partie septentrionale de l'archipel arctique.

Dans la région de l'inlet Phillips, un granodiorite a donné un âge de 965 ± 2 Ma sur zircon, confirmant l'âge grenvillien attribué à la zone de la pointe Mitchell faisant partie du socle cristallin du terrane de la Pearya.

Dans la région du fjord Ayles, la datation d'une granodiorite syntectonique (âge de 475 ± 1 Ma sur monazite; âge ^{40}Ar-^{39}Ar de 467,8 ± 2,5 Ma, plateau de la biotite) a permis de déplacer l'amorce de l'orogenèse de M'Clintock du terrane de la Pearya, en la faisant passer du début de l'Ordovicien moyen, comme présumé, à l'Ordovicien précoce.

Les petits plutons post-tectoniques de niveau élevé et de composition variable sont nombreux dans le nord de l'île Axel Heiberg et de l'île d'Ellesmere. Des âges correspondant au Dévonien tardif ont été obtenus pour un pluton tonalitique (datation sur zircon à 368 ± 3 Ma, datation ^{40}Ar-^{39}Ar sur biotite à 360 ± 2 Ma) et pour un neck de trachyandésite (datation sur zircon à 368 ± 3 Ma), tous deux observés dans le nord de l'île Axel Heiberg, ainsi que pour un pluton tonalitique de la partie nord-ouest de l'île d'Ellesmere (âge ^{40}Ar-^{39}Ar de 364 ± 2 Ma, plateau de la hornblende). Il se peut que ces trois intrusions soient légèrement plus récentes que la déformation d'un bassin d'avant-pays situé plus au sud dans l'archipel arctique (orogenèse ellesmerienne, Dévonien terminal-Carbonifère précoce).

Dans la région de la baie Henson, dans le nord-ouest de l'île d'Ellesmere, un petit pluton de monzodiorite s'est formé il y a 334,5 ± 1,5 Ma (zircon et titanite), presque en même temps que la mise en place de la Formation d'Emma Fiord du Viséen supérieur, la plus ancienne unité connue du bassin de Sverdrup. C'est le premier indice qui permette de supposer que le soulèvement thermique ou le rifting initial du bassin de Sverdrup ait été accompagné de phénomènes magmatiques.

Dans la région de la péninsule Marvin, à la limite septentrionale de l'île d'Ellesmere, un pluton de monzodiorite-granodiorite quartzique a donné un âge ^{40}Ar-^{30}Ar de 94,2 ± 0,7 Ma (plateau de la hornblende), confirmant une corrélation approximative avec la vaste intrusion bimodale de Wootton, datée à 92 ± 1 Ma et observée dans le nord-ouest de l'île d'Ellesmere. Au même titre que la formation volcanique de Strand Fiord, dans l'ouest de l'île Axel Heiberg, ces plutons indiquent qu'il y a eu un important épisode de rifting au Crétacé tardif (Cénomanien).

GEOLOGICAL FRAMEWORK OF NORTHERN ELLESMERE AND AXEL HEIBERG ISLANDS

The Cambrian to Upper Devonian rocks of the Innuitian Orogen on Ellesmere and Axel Heiberg islands (Fig. 1) are assigned to the Franklinian mobile belt (Fig. 2). On Ellesmere Island, three major tectonic elements have previously been distinguished within the pre-Devonian succession of this belt (Trettin, 1987a): a southeastern, unstable shelf that is contiguous with the cratonic cover of North America (Franklinian Shelf), a northern microcontinent (Pearya), and an intervening deep-water basin. The deep-water basin in turn was divided into a southeastern subprovince consisting exclusively of deep-water sediments, and a northwestern subprovince containing volcanics and shallow marine sediments in addition to abundant deep-water sediments.

Four major successions were recognized in the Pearya Terrane: (1) a crystalline basement, late Middle to early Late Proterozoic in age; (2) metasedimentary and metavolcanic units of Late Proterozoic to earliest Ordovician age; (3) arc-type metavolcanic rocks and associated metasedimentary rocks of probable Early-Middle Ordovician age; and (4) fossiliferous sedimentary and volcanic successions of late Middle Ordovician to Late Silurian age that are weakly metamorphosed or unmetamorphosed.

It was recognized that Pearya is related to the Caledonides by the Grenville age of its crystalline basement, the presence of Early Ordovician ultramafic rocks, and the occurrence of an Ordovician deformation (M'Clintock Orogeny). In contrast, The Franklinian mobile belt has an Archean-Early Proterozoic crystalline basement and lacks an Ordovician orogenic belt or foreland basin. The two tectonic elements also differ in structural trends.

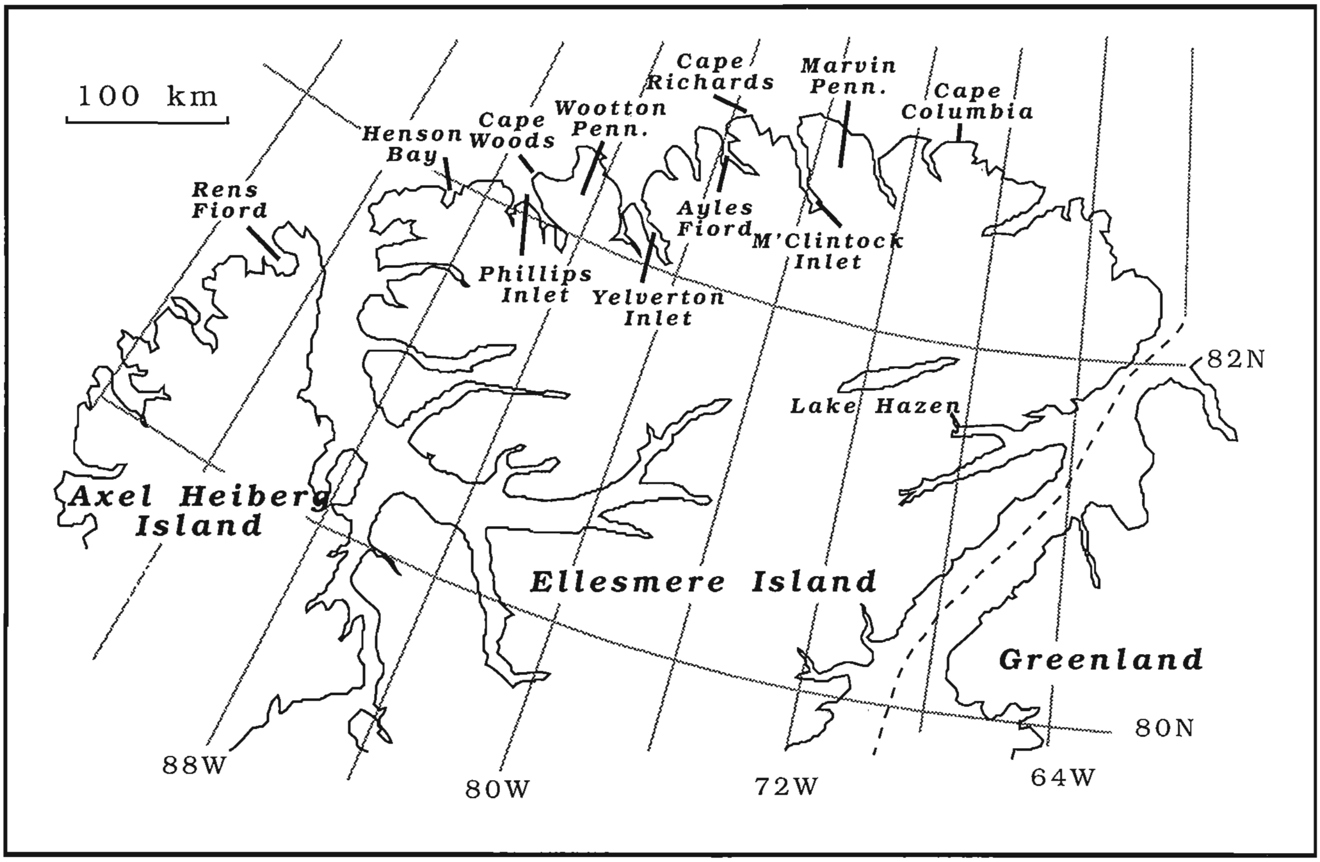


Figure 1. Map of northern Ellesmere Island showing geographic features referred to in text.

Time and mechanism of the accretion of Pearya still are problematic. It was previously inferred that Pearya was accreted to the deep-water basin by sinistral transpression during a major deformation in latest Silurian time (Trettin, 1987a). More recent studies (Bjornerud, 1991; Klaper, 1992) indicate that the accretion probably occurred prior to mid-Early Silurian (late Llandovery) time as a late Llandovery flysch of uniform composition overlaps parts of Pearya and of the deep-water basin. Moreover, the northwestern part of the deep-water basin includes a suspect terrane that is stratigraphically more closely related to Pearya than to the Franklinian mobile belt. Thus the main suture probably lies within the sedimentary-volcanic subprovince. However, mid-Early Silurian deformation was restricted to the northwestern part of the deep-water basin and was less pronounced than during the orogenic events extending from latest Silurian to Early Carboniferous. It appears that strike slip occurred repeatedly but the sites, ages, and magnitudes of these movements are still under study and will not be discussed in this paper.

In more southerly parts of the Arctic Islands, local movements occurred in Late Silurian to Early Devonian time (Boothia and Inglefield uplifts). A foreland basin developed in Middle and Late Devonian time that received clastic sediments from northern and eastern sources (Embry and Klovan, 1976; Embry, 1991a; McNicoll et al., 1992). The terminal Ellesmerian Orogeny commenced in the northern part of the mobile belt, advanced southeast and south, and affected the preserved part of the foreland basin in latest Devonian-Early Carboniferous time (Ellesmerian Orogeny; Thorsteinsson and Tozer, 1970; Harrison, 1991; Harrison et al., 1991).

Lower Carboniferous (Viséan) to Upper Cretaceous strata of the Sverdrup Basin (Thorsteinsson, 1974; Davies and Nassichuk, 1991; Mayr, in press; Embry, 1991b) unconformably overlie Devonian and older rocks of the Franklinian mobile belt, including Pearya. The Sverdrup Basin was deformed from latest Cretaceous to late Eocene or early

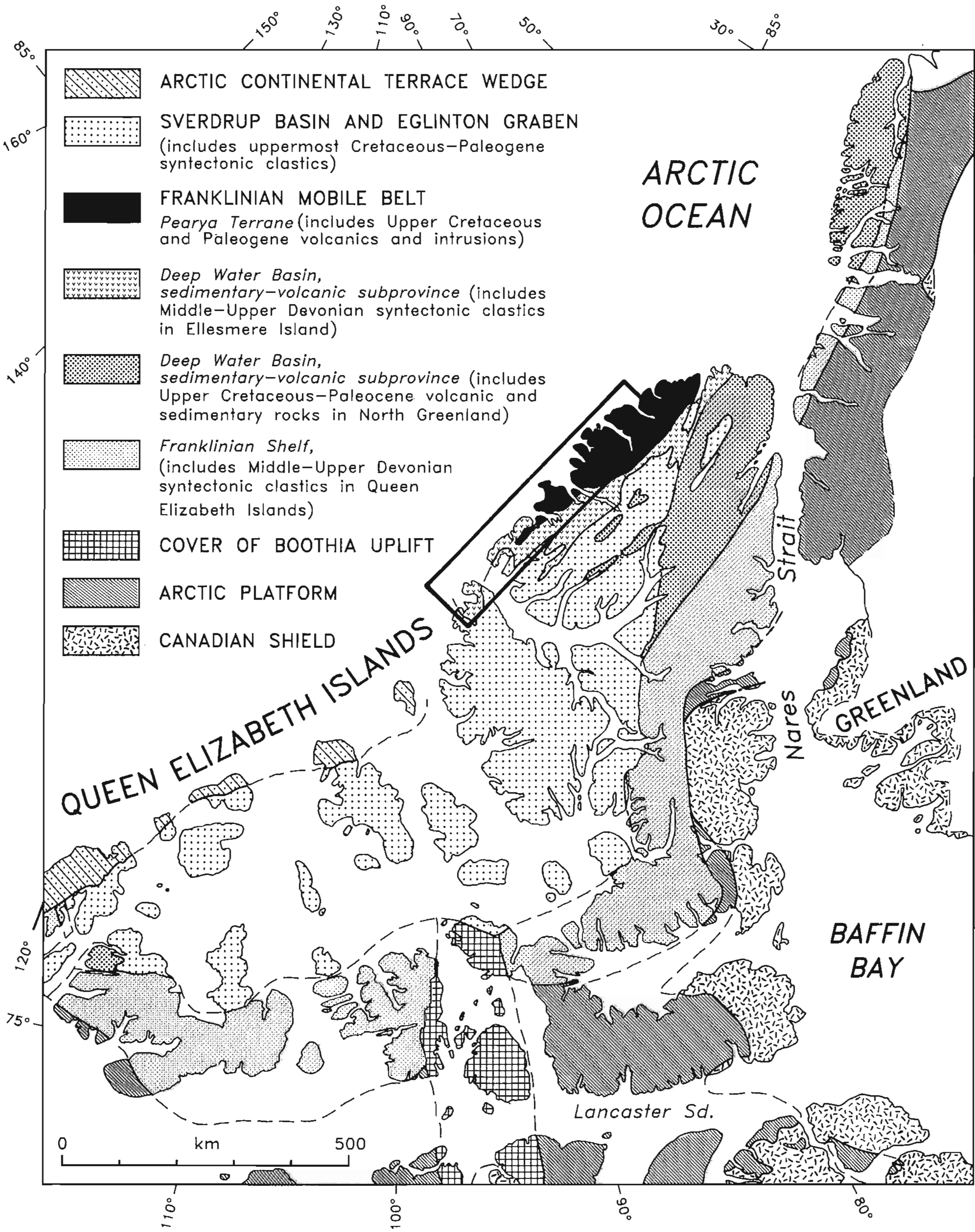


Figure 2. Geological provinces of the Arctic Islands and location of study area.

Table 1. U-Pb isotopic data

Sample[1] analysis	wt.[2] (mg)	U (ppm)	Pb[3] (ppm)	$^{206}Pb/^{204}Pb$ [4]	Pb_c [5] (pg)	$^{208}Pb/^{206}Pb$ [3]	$^{206}Pb/^{238}U$ [6]	$^{206}Pb/^{238}U$(Ma) [7]	$^{207}Pb/^{235}U$ [6]	$^{207}Pb/^{235}U$(Ma) [7]	$^{207}Pb/^{206}Pb$ [6]	corr. coef.	$^{207}Pb/^{206}Pb$(Ma) [7]
88-TM-136C													
A,n	0.0017	3376	276.0	3264	10	0.06	0.08529±0.10%	527.6±1.0	0.7057±0.11%	542.2±1.0	0.06001±0.05%	0.88	603.9±2.3
B,n	0.0353	1438	102.4	3292	7	0.11	0.07097±0.10%	442.0±0.8	0.5529±0.12%	446.9±0.9	0.05651±0.05%	0.91	472.5±2.3
C,eq	0.0408	1875	123.5	2308	139	0.11	0.06584±0.10%	411.1±0.8	0.5107±0.13%	419.0±0.9	0.05626±0.07%	0.87	462.6±3.0
MON-A,eu	0.0665	3414	1186	6852	161	4.18	0.07662±0.12%	475.9±1.1	0.5979±0.13%	475.9±1.0	0.05659±0.03%	0.97	475.7±1.5
MON-B,eu	0.0444	4527	1401	8081	120	3.63	0.07635±0.11%	474.3±1.0	0.5954±0.12%	474.3±0.9	0.05656±0.03%	0.97	474.6±1.5
88-TM-208B													
C-1,su	0.0937	233.9	42.22	15910	16	0.08	0.18127±0.09%	1073.9±1.7	1.9755±0.10%	1107.2±1.3	0.07905±0.03%	0.96	1173.2±1.2
C-2,su	0.0739	242.2	44.33	6953	30	0.07	0.18495±0.09%	1094.0±1.8	2.0429±0.10%	1129.9±1.4	0.08011±0.03%	0.95	1199.7±1.3
S-1,eu	0.0475	256.6	46.00	18270	8	0.06	0.18390±0.09%	1088.2±1.7	2.0016±0.10%	1116.1±1.3	0.07894±0.03%	0.95	1170.6±1.2
S-2,eu	0.0383	337.0	55.08	8164	17	0.04	0.16968±0.09%	1010.4±1.6	1.8933±0.10%	1078.8±1.3	0.08093±0.03%	0.95	1219.7±1.3
S-3,eu	0.0428	253.1	40.33	4638	24	0.08	0.16080±0.08%	961.2±1.5	1.5797±0.10%	962.3±1.2	0.07125±0.03%	0.94	964.7±1.4
88-TM-103C													
A,eq,eu	0.1625	339.9	22.21	6044	34	0.22	0.05948±0.09%	372.4±0.7	0.4663±0.11%	388.6±0.7	0.05686±0.04%	0.94	486.2±1.6
B,el,eu	0.2132	332.7	24.35	12090	24	0.23	0.06586±0.09%	411.2±0.7	0.5854±0.10%	467.9±0.7	0.06446±0.03%	0.96	756.9±1.3
C,n,eu	0.1185	402.6	26.68	10550	17	0.24	0.05952±0.09%	372.7±0.6	0.4637±0.10%	386.8±0.6	0.05650±0.03%	0.95	472.2±1.4
D,rd,an	0.1840	183.1	68.05	14670	46	0.21	0.32040±0.08%	1791.7±2.6	6.1745±0.10%	2000.8±1.7	0.13977±0.03%	0.96	2224.4±1.0
88-TM-107A													
A,eq,eu	0.0619	298.4	51.00	8614	20	0.19	0.15101±0.08%	906.6±1.4	2.6485±0.10%	1314.2±1.4	0.12720±0.03%	0.96	2059.7±1.0
B,el,eu	0.0740	541.8	32.40	10980	14	0.13	0.05854±0.09%	366.7±0.6	0.4383±0.10%	369.1±0.6	0.05430±0.03%	0.96	383.7±1.3
C,el,eu	0.0627	604.7	35.67	6181	23	0.12	0.05844±0.09%	366.1±0.6	0.4347±0.10%	366.5±0.6	0.05395±0.04%	0.94	369.0±1.6
88-TM-117D													
A,su,	0.1125	412.3	31.15	11210	14	0.60	0.05301±0.09%	332.9±0.6	0.3884±0.10%	333.2±0.6	0.05315±0.03%	0.95	335.1±1.4
B,su	0.1228	289.5	20.55	12370	11	0.36	0.05840±0.09%	365.9±0.6	0.4479±0.10%	375.8±0.6	0.05563±0.03%	0.95	437.8±1.4
C,su	0.1096	295.6	18.80	7676	14	0.34	0.05302±0.09%	333.0±0.6	0.3884±0.10%	333.2±0.6	0.05314±0.04%	0.94	334.6±1.6
S-1,TIT	0.1454	124.1	11.62	265	245	0.98	0.05323±0.16%	334.3±1.0	0.3907±0.47%	334.9±2.7	0.05324±0.38%	0.69	339.0±17.3
S-1,TIT	0.1446	126.4	11.52	329	197	0.94	0.05310±0.13%	333.5±0.8	0.3887±0.34%	333.4±1.9	0.05309±0.27%	0.68	332.4±12.3

[1]MON = monazite; TIT = titanite; eu = euhedral; su = subhedral; an = anhedral; el = elongate; n = needles; eq = equant; na = not abraded (all other fractions of zircon were abraded)
[2]Weighing error estimated to 0.001 mg.
[3]Radiogenic Pb.
[4]Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU
[5]Total common Pb in analysis corrected for fractionation and spike.
[6]Corrected for blank Pb and U, and common Pb; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean in when expressed in Ma.
[7]Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

Oligocene time (Eurekan Orogeny), and the syntectonic clastic sediments shed from related uplifts reside in the Eureka Sound Group (Miall, 1991).

PRESENT INVESTIGATIONS

Attempts to unravel the complex tectonic history of the northern part of the Arctic Islands by means of isotopic age determinations have been made from 1960 onward. However, the K-Ar method, used exclusively until the mid-seventies, yielded misleading results for rocks older than Late Devonian because of Devonian and younger thermal overprints. It was not until the advent of the Rb-Sr isochron and U-Pb concordia techniques that Precambrian and early Palaeozoic metamorphic-plutonic events could be deciphered (Sinha and Frisch, 1975, 1976; Trettin et al., 1982, 1987).

Here we continue this line of research by providing U-Pb and/or ^{40}Ar-^{39}Ar ages for seven granitoid plutons, ranging in age from early Late Proterozoic to Late Cretaceous. Where both methods have been applied to the same samples the results were compatible. Using the comprehensive time scale of Harland et al. (1989) with corrections for the Ordovician by Tucker et al. (1990) and for the Devonian-Carboniferous boundary by Claoué-Long et al. (1992), we relate these dates to the stratigraphic-structural record of the Arctic Islands and other regions. However, it is prudent to recall that all boundaries older than the Arenig-Llanvirn are uncertain. Further discussions of the genesis of the intrusions are deferred to forthcoming reports with chemical analyses.

We also attempted to date metamorphic events of Ordovician and Silurian or Devonian age by means of ^{40}Ar-^{39}Ar determinations on schists. The results are inconclusive but are reported fully to provide guidance for future research.

ANALYTICAL TECHNIQUES

U-Pb analytical methods for zircon and monazite are outlined in Parrish et al. (1987) and for titanite in Parrish et al. (1992). U and Pb blanks were approximately 1 and 15 picograms, respectively. The U-Pb data are presented in Table 1 and in Figures 4, 7, 14, 16 and 20. In the concordia plots, errors are shown at the 2σ level.

The ^{40}Ar-^{39}Ar incremental release spectra for muscovite, biotite, and hornblende were obtained by methods outlined in Roddick (1990). Data are given in Table 2. Cumulative per cent ^{39}Ar vs. age and Ca/K plots are presented in Figures 8, 9, 11, 12, 15, 17 and 22. Errors on step ages are shown at the 2σ level. The Ca/K ratio plots have been normalized to the average Ca/K ratio of the mineral separate in each sample, and a summary of relevant information, including total fusion and plateau ages, is listed in Table 2. Co-ordinates for all sample locations are listed in Table 3.

PHILLIPS INLET PLUTON, BASEMENT OF PEARYA (88-TM-208B)

The crystalline basement of the Pearya Terrane consists mainly of gneiss with lesser amounts of schist and amphibolite. It occurs in three major, fault-bounded belts that are similar lithologically but have different orientation and structural trends (Fig. 3). They have been named the Cape Columbia, Deuchars Glacier, and Mitchell Point belts. In addition, a small inlier occurs east of Henson Bay in a different structural belt.

The Proterozoic age of this basement was first established by Sinha and Frisch (1975) on the basis of Rb-Sr data from widely scattered localities in the Mitchell Point belt. These yielded isochrons of 726 ± 12 Ma and 802 ± 19 Ma (recalculated), interpreted as minimum ages. Rb-Sr data from the Cape Columbia belt produced a 1060 ± 18 Ma isochron age, and bulk zircons of two pegmatite samples had ^{207}Pb-^{206}Pb ages of 980 and 926 Ma, respectively, all within the range of ages from the Grenville and Sveconorwegian provinces of the North Atlantic region (Sinha and Frisch, 1976). This inference was supported by U-Pb zircon analyses on massive, plutonic granitoids of the Deuchars Glacier Belt, one of which had an upper intercept age of 1037 +25/-20 Ma (Trettin et al., 1987).

The lithological similarity of the three belts, and the proximity of the Deuchars Glacier and Mitchell Point belts suggested that the latter might also be ca. 1 Ga old rather than ca. 700-800 Ma as suggested by the first Rb-Sr ages. To verify this and to elucidate the intrusive history of the belt, we dated a small granitic intrusion in its western part, here referred to as the Phillips Inlet pluton (Fig. 3), which is in fault contact with Lower Silurian or older metasedimentary rocks on the south. The pluton is massive and has undergone considerable deformation, alteration and recrystallization. This involved brecciation and rotation of feldspar, growth of muscovite and chlorite between feldspar aggregates, producing an irregular, wavy foliation, and introduction and/or remobilization of quartz.

A representative thin section of the dated sample is medium grained and composed of albite 32%, K-feldspar 14%, quartz 23%, chlorite 19%, mica (mainly muscovite with minor green biotite) 10%, calcite 1%, and opaque minerals (trace).

Zircons from sample 88-TM-208B of the Phillips Inlet pluton were generally clear, pink in colour, mainly sharply faceted, and relatively equant in shape. Although of excellent clarity, many grains contained inclusion-bearing regions suggestive of inherited zircon cores inside an euhedral to subhedral magmatic rim. This visible evidence for inheritance is borne out in most of the U-Pb analyses (Fig. 4), each of which comprised 3 to 8 crystals. Four of five analyses are substantially discordant, but analysis S-3 is less than 0.3% discordant. It cannot be proven that this analysis is totally free of inheritance, but we suggest that its minor discordance is the result of slight Pb loss; consequently we infer that the tentative age of magmatic crystallization of the pluton is given by the ^{207}Pb-^{206}Pb age of this analysis, 965 Ma, and infer its error at ± 2 Ma.

The U-Pb age confirms that the Mitchell Point Belt is broadly coeval with the ca. 1 Ga Cape Columbia and Deuchars Glacier belts. It indicates a late phase of granitic plutonism within the basement of Pearya, perhaps analogous to a brief but extensive phase of plutonism in the eastern Grenville Province, which occurred between 966 and 956 Ma (Gower, 1990). The specific tectonic origin of the pluton remains uncertain because of its altered state, but the presence of inherited zircon confirms that it was derived, at least partly, from crustal sources, possibly older portions of the intruded crystalline terrane.

THE M'CLINTOCK OROGENY REVISITED

In the M'Clintock Inlet area of northernmost Ellesmere Island, a deformation of Ordovician age, known as the M'Clintock Orogeny, is apparent from two angular unconformities at the base of different parts of succession IV (succession IV: Egingwah and Harley Ridges Groups of Caradoc-Ashgill age; Trettin, 1987a) of the Pearya Terrane (Fig. 5, 6). 1) Boulder conglomerate of early Caradoc age (Cape Discovery Formation, member A; Egingwah Group) overlies metamorphosed arc-type volcanics (andesite and basalt) and associated sedimentary rocks (Maskell Inlet assemblage; succession II of Pearya Terrane); and carbonate strata of early Caradoc age (Cape Discovery Formation, member B) overlie an ultramafic and mafic intrusion (Bromley Island body of Thores Suite) structurally associated with this assemblage. 2) A low-angle regional unconformity exists within the Ashgill part of succession IV, which east of M'Clintock Inlet superposes the Taconite River Formation directly on the volcanic M'Clintock Formation, the intervening Ayles Formation having been eroded. This regional unconformity becomes a high-angle unconformity at the head of M'Clintock Inlet where boulder conglomerate of the Taconite River Formation overlies schist of succession II that probably is Cambrian or Late Proterozoic in age. The folding and metamorphism of the schist probably is pre-Caradoc in age but the Caradoc cover (Egingwah Group), if deposited, was removed during an early Ashgill uplift.

The most significant feature of the M'Clintock Orogen is a series of faults that juxtapose the Maskell Inlet assemblage, and/or associated fault slices of ultramafic-mafic rocks (Thores Suite), with succession II of the Pearya Terrane. The Ordovician age of these faults is apparent from the fact that they are unconformably overlain by succession IV. This

Table 2. ^{40}Ar-^{39}Ar data

TEMP. (°C)	$^{36}Ar_{tr}$	$^{37}Ar_{Ca}$ (X 10^{-9} cc STP)[a]	$^{38}Ar_{Cl}$	$^{39}Ar_K$	^{40}Ar	% Atmos. ^{40}Ar	Apparent age Ma ±2σ[b]	^{39}Ar (%)
88-TM-136C Biot. (197) (15.17 mg)								
600	0.036	0.047	0.033	0.316	16.09	65.6	207.0	8.9
700	0.010	0.030	0.019	2.296	98.12	3.1	456.1	1.2
750	0.007	0.017	0.036	4.901	212.34	0.9	470.7	1.0
800	0.009	0.012	0.073	9.405	404.57	0.6	469.0	0.9
850	0.011	0.023	0.048	5.633	242.25	1.3	466.0	1.8
(900) -LOST - estimated								7.4
935	0.001	0.016	0.016	2.003	86.68	0.4	472.4	2.5
970	0.003	0.030	0.020	2.592	112.39	0.7	471.9	1.4
1000	0.002	0.049	0.032	4.282	185.71	0.3	473.6	0.7
1050	0.004	0.095	0.047	6.193	266.81	0.4	470.4	0.5
1100	0.002	0.213	0.049	6.880	295.45	0.2	470.2	1.1
1150	0.000	0.329	0.056	7.621	327.26	0.0	470.7	0.6
1250	0.002	0.927	0.016	2.000	85.97	0.6	469.1	1.1
1450	0.005	0.174	0.004	0.518	22.14	6.1	443.9	5.5
Total[c]	0.10	2.00	0.49	58.99	2543.9	1.1	467.8	2.5
Conc.(/g)	6.5	131.6	32.09	3889.	167690.			
88-TM-206B Musc. (198) (15.44mg)								
600	0.026	0.092	0.082	0.531	20.16	38.6	270.8	6.9
700	0.009	0.174	0.044	1.049	38.91	7.1	387.2	2.9
750	0.007	0.144	0.023	0.922	38.15	5.1	435.1	2.4
800	0.007	0.247	0.021	1.360	56.60	3.8	442.7	1.9
850	0.008	0.236	0.018	2.306	95.41	2.6	445.4	1.6
900	0.005	0.218	0.026	3.451	142.71	1.0	451.5	0.7
935	0.003	0.187	0.030	3.892	160.94	0.6	453.0	1.3
970	0.002	0.142	0.028	2.652	109.57	0.4	453.3	1.5
1000	0.004	0.201	0.040	2.289	93.49	1.2	445.7	1.7
1050	0.003	0.304	0.061	1.841	74.30	1.0	441.7	2.8
1100	0.004	0.386	0.062	1.534	62.10	1.9	439.2	0.9
1150	0.003	0.537	0.078	1.617	69.87	1.5	467.4	4.9
1250	0.004	0.897	0.075	0.593	29.98	3.7	526.0	3.9
1450	0.008	0.840	0.071	0.185	12.05	19.9	559.0	31.2
Total[c]	0.09	4.60	0.66	24.22	1004.2	2.7	445.4	2.5
Conc.(/g)	6.0	298.3	42.7	1569.	65040.			
88-TM-OHT-40 Hbld. (204) (82.57mg)								
700	0.029	0.657	0.012	0.112	15.90	53.5	688.3	45.1
800	0.017	0.536	0.002	0.071	31.74	15.9	2331.2	35.2
900	0.009	0.715	0.001	0.087	28.58	9.7	2025.5	33.5
1000	0.007	4.755	0.033	0.486	21.37	9.2	444.8	3.3
1020	0.004	2.738	0.018	0.186	17.63	6.7	869.2	13.9
1040	0.003	3.415	0.025	0.209	14.30	6.8	667.1	8.2
1060	0.001	3.499	0.023	0.213	11.08	3.7	542.1	12.2
1080	0.008	16.760	0.108	0.868	40.04	5.8	479.4	5.6
1100	0.001	23.904	0.172	1.172	46.05	0.4	437.2	4.1
1110	0.001	1.642	0.013	0.091	3.10	12.6	342.6	18.8
1120	0.000	1.199	0.008	0.062	2.18	0.9	393.3	40.7
1130	0.000	1.460	0.011	0.070	2.76	0.8	437.6	17.8
1140	0.000	2.582	0.017	0.121	4.80	0.8	439.7	14.0
1150	0.000	4.047	0.027	0.184	7.42	1.6	441.8	16.2

Table 2. (cont'd.)

88-TM-OHT-40 Hbld. (204) (82.57mg) (cont'd.)								
1160	0.001	5.285	0.033	0.240	9.72	1.6	445.4	5.9
1170	0.001	8.399	0.055	0.386	15.42	2.7	434.7	8.4
1180	0.002	12.076	0.074	0.559	22.53	2.0	441.1	4.4
1190	0.002	8.911	0.061	0.409	16.53	3.0	438.2	9.0
1200	0.001	7.176	0.049	0.333	13.23	2.7	432.1	10.2
1220	0.002	11.748	0.086	0.546	21.55	2.4	431.0	6.2
1250	0.003	15.538	0.110	0.722	28.56	2.6	431.5	5.6
1300	0.003	24.385	0.167	1.170	45.08	1.9	423.9	1.8
1375	0.001	3.319	0.025	0.158	6.76	3.4	457.7	10.1
1550	0.004	1.157	0.026	0.161	4.21	25.5	230.7	76.5
Total[c]	0.10	165.90	1.16	8.62	430.5	6.8	509.6	3.7
Conc.(/g)	1.2	2009.	14.0	104.4	5214.			
88-TM-OHT-3 Musc. (200) (21.38mg)								
600	0.016	0.010	0.012	0.292	9.43	49.9	191.5	11.4
700	0.008	0.008	0.003	0.800	17.39	13.6	220.4	2.2
750	0.010	0.019	0.003	0.748	20.31	14.1	270.1	5.2
800	0.019	0.008	0.008	1.989	57.07	9.7	297.5	1.4
850	0.013	0.001	0.012	7.701	218.26	1.7	318.1	1.1
900	0.010	0.000	0.039	20.682	587.17	0.5	322.2	0.5
935	0.002	0.022	0.021	11.876	333.68	0.2	320.0	0.8
970	0.002	0.021	0.013	6.747	188.57	0.4	317.9	0.6
1000	0.003	0.005	0.012	5.605	157.92	0.6	319.7	0.4
1050	0.006	0.014	0.019	9.816	278.67	0.6	321.9	0.3
1100	0.004	0.003	0.015	7.634	217.47	0.6	322.9	0.5
1150	0.003	0.010	0.020	11.213	322.08	0.3	326.3	0.9
1250	0.000	0.004	0.004	2.926	82.54	0.2	321.2	1.2
1450	0.000	0.018	0.004	0.696	20.08	0.6	326.6	4.9
Total[c]	0.10	0.14	0.19	88.73	2510.7	1.1	319.3	1.8
Conc.(/g)	4.5	6.70	8.69	4150.	117430.			
88-TM-103C Biot. (199) (15.89mg)								
600	0.010	0.012	0.032	0.299	9.47	31.5	251.0	7.5
700	0.011	0.037	0.264	2.536	83.98	4.0	356.9	1.0
750	0.008	0.035	0.646	6.274	204.19	1.1	360.9	0.7
800	0.002	0.037	0.976	9.500	306.18	0.2	360.8	0.5
850	0.002	0.023	0.728	7.059	227.58	0.2	360.7	0.9
900	0.001	0.011	0.213	2.094	67.66	0.5	360.5	1.6
950	0.001	0.026	0.187	1.858	59.95	0.5	360.1	1.2
1000	0.002	0.044	0.322	3.254	105.52	0.6	361.2	0.6
1050	0.002	0.051	0.460	4.585	147.72	0.3	360.0	0.7
1100	0.002	0.129	0.946	9.306	299.52	0.2	360.3	0.5
1150	0.002	0.259	0.733	7.106	228.71	0.3	359.9	0.6
1450	0.002	0.989	0.558	5.174	166.51	0.3	359.8	0.8
Total[c]	0.04	1.65	6.06	59.05	1907.0	0.7	359.7	2.0
Conc.(/g)	2.8	104.0	381.6	3716.	120013.			
88-TM-208C Hbld. (205) (89.88mg)								
700	0.045	0.603	0.091	0.172	16.90	79.1	240.3	52.0
800	0.014	0.390	0.041	0.189	10.07	41.9	351.9	28.5
900	0.006	0.715	0.061	0.196	7.68	24.2	338.2	9.0
1000	0.008	12.394	1.530	1.050	36.33	6.5	366.2	4.8

Table 2. (cont'd.)

88-TM-208C Hbld. (205) (89.88mg) (cont'd.)								
1030	0.004	18.874	3.420	1.966	66.36	1.6	375.3	3.2
1060	0.009	36.913	8.054	4.626	151.75	1.7	365.3	0.7
1080	0.004	26.272	5.907	3.412	110.56	1.1	363.2	0.5
1100	0.003	21.656	4.899	2.853	92.27	1.1	362.7	1.0
1120	0.001	5.700	1.217	0.748	24.13	1.4	360.8	2.3
1140	0.001	4.561	0.950	0.581	18.83	1.1	363.2	2.9
1160	0.003	13.756	2.955	1.653	53.78	1.6	362.9	1.0
1180	0.002	11.966	2.541	1.354	44.48	1.6	365.9	2.3
1200	0.002	5.928	1.214	0.624	20.54	2.6	363.3	2.9
1220	0.002	4.822	0.964	0.492	16.43	3.7	364.5	2.8
1250	0.002	4.617	0.879	0.446	14.60	3.4	358.9	3.6
1300	0.002	9.505	1.638	0.797	26.55	1.8	370.2	8.7
1550	0.003	11.951	1.924	1.009	33.65	3.1	366.2	2.1
Total[c]	0.11	190.62	38.29	22.17	744.9	4.4	364.0	2.1
Conc.(/g)	1.2	2121.	426.0	246.7	8288.			
77-TM-320A Hbld. (206) (89.86mg)								
700	0.032	0.386	0.213	0.255	10.64	88.6	58.5	25.9
800	0.020	0.473	0.223	0.227	7.92	75.1	105.8	8.0
900	0.005	1.211	0.701	0.238	3.48	43.4	101.1	7.4
1000	0.015	30.430	15.475	3.699	34.16	12.9	98.3	1.0
1030	0.013	45.630	18.681	5.714	47.77	8.0	94.1	0.5
1060	0.006	33.512	13.833	4.422	36.05	4.9	94.8	0.5
1080	0.003	9.345	3.762	1.327	11.03	8.1	93.5	2.7
1100	0.003	3.998	1.547	0.595	5.24	14.4	92.3	4.0
1200	0.007	13.501	5.561	1.964	17.83	11.8	97.8	2.1
1550	0.004	8.589	3.001	1.072	9.96	13.0	98.7	1.3
Total[c]	0.11	147.07	63.07	19.51	184.1	17.3	95.4	0.8
Conc.(/g)	1.2	1637.	701.1	217.2	2048.			

[a]All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar and Ca, Cl and K denote Ar derived from these elements. ^{40}Ar denotes trapped plus radiogenic Ar. Atmos. ^{40}Ar assumes a trapped Ar component of atmospheric composition. [b]Errors from steps are analytical only and do not include the error in the irradiation parameter J. [c]Includes the integrated age. The uncertainty in J (0.6% 2σ) is included in the error.

Table 3. Location of dated samples

Field no.	NTS	Zone	UTM East	North
88-TM-208B	340 C	16X	504000	9102750
88-TM-136C	340 E	17X	522150	9185000
88-TM-OHT-40	340 F	17X	472750	9127200
88-TM-OHT-3	340 F	17X	550000	9130000
88-TM-103C	560 D	15X	479200	9026600
86-TM-107A	560 A	15X	491200	8989050
88-TM-208C	340 C	17X	465750	9077700
88-TM-117D	340 F	17X	494850	9157900

relationship suggests that the orogeny was caused by the collision of a volcanic arc (Maskell Inlet assemblage) with a miogeocline (succession II). The significance of the Thores Suite is uncertain: it could be interpreted either as an Alaskan-type subarc plutonic complex or as a dismembered ophiolite that lacks a sheeted dyke complex and pillow basalt. The orogeny was characterized by regional metamorphism up to amphibolite grade and was accompanied and followed by granitoid plutonism. The largest dated plutons are the Markham Fiord pluton, an arc-type intrusion, and the Cape Richards complex, a post-tectonic intrusion with alkalic characteristics.

For the dating of the deformation, the following information was available prior to the present study (Fig. 6):

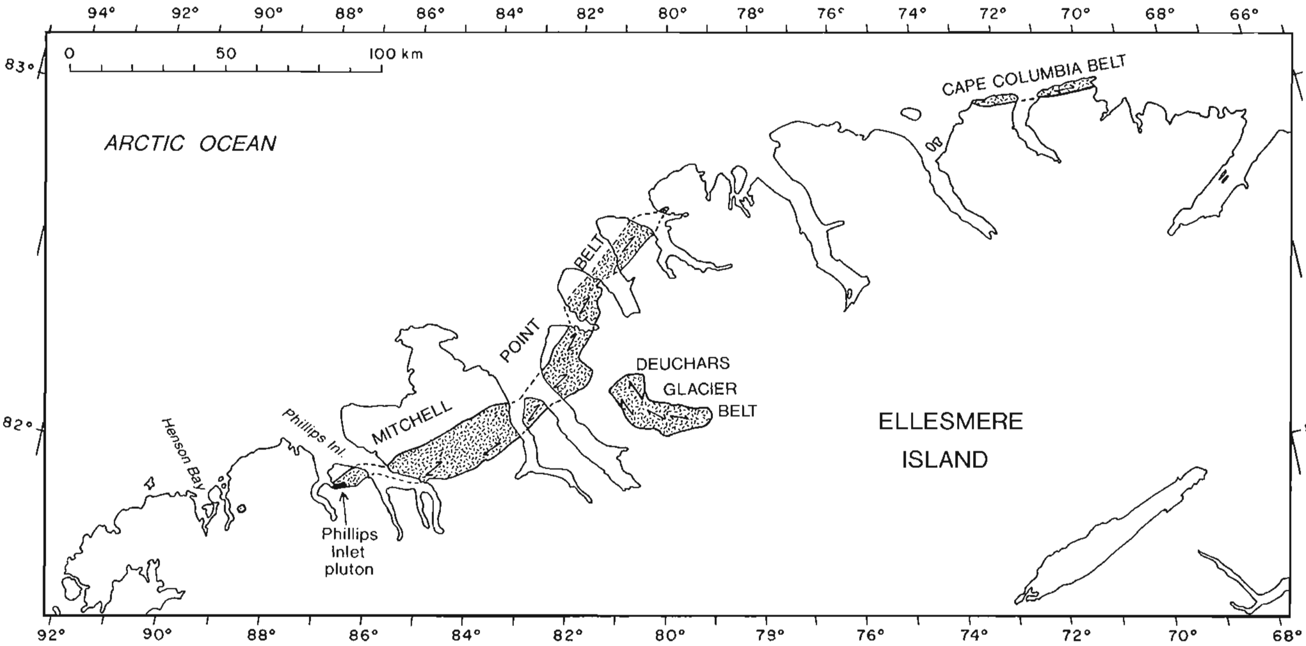


Figure 3. Crystalline basement of Pearya Terrane: generalized outcrop areas, gneissic trends, and location of dated pluton.

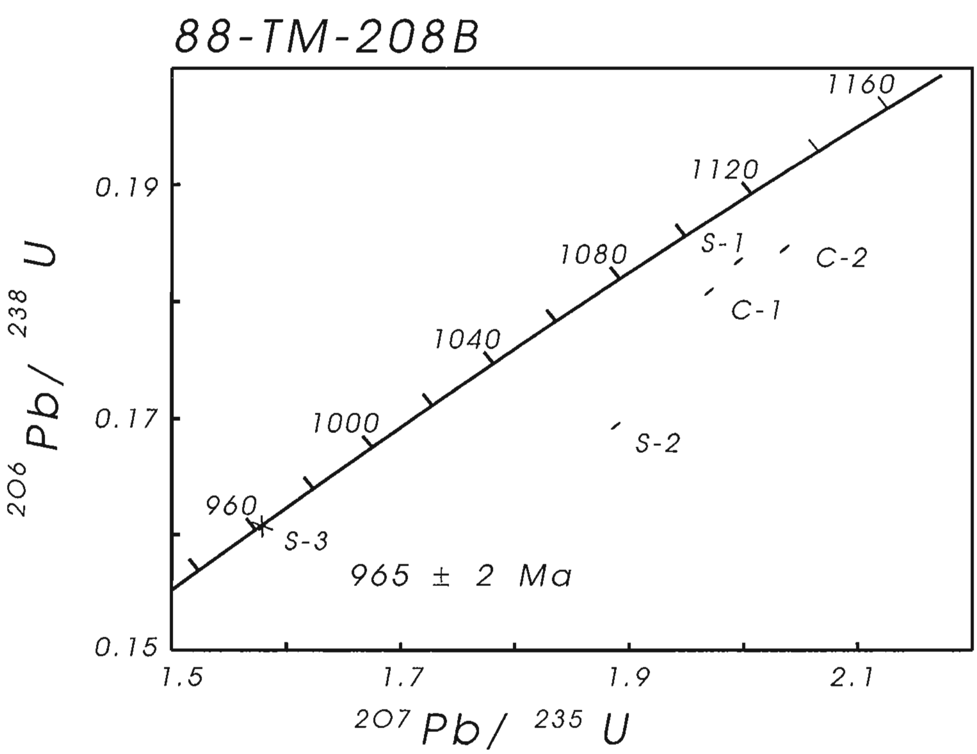


Figure 4.

Concordia diagram for Phillips Inlet pluton.

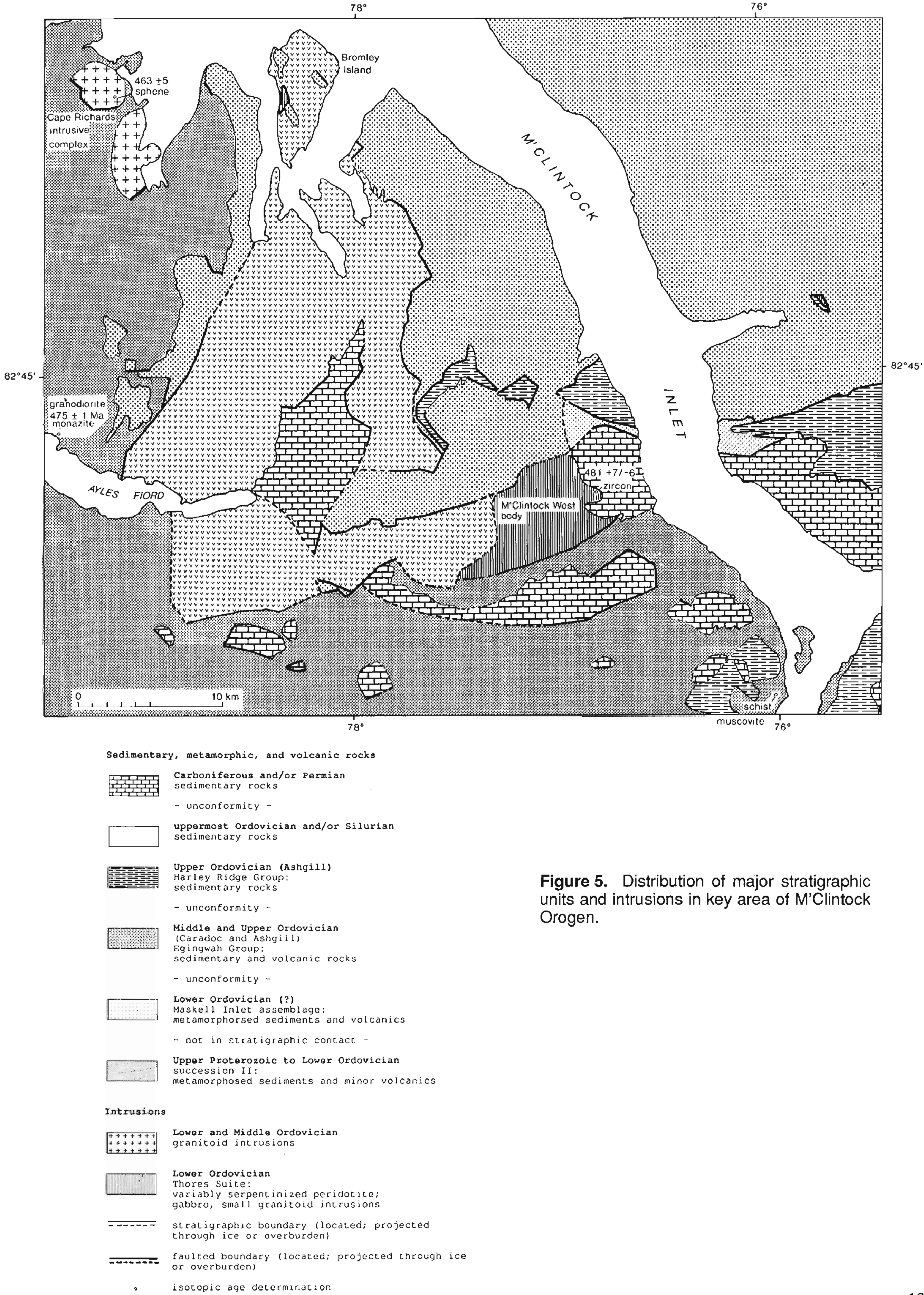


Figure 5. Distribution of major stratigraphic units and intrusions in key area of M'Clintock Orogen.

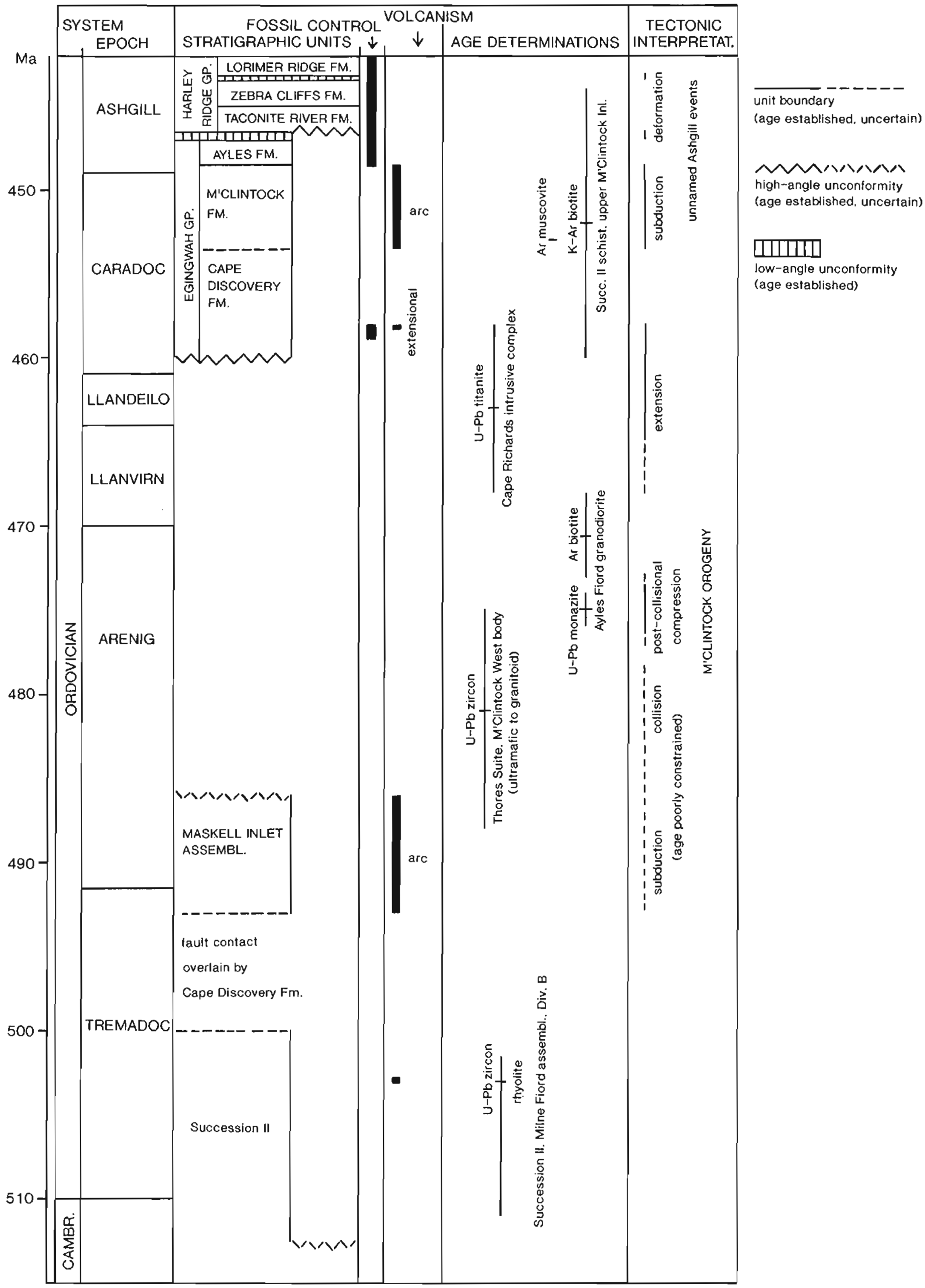


Figure 6. Correlation of Ordovician stratigraphic units and plutons, M'Clintock Orogen.

1. Succession II of the Pearya Terrane probably ranges in age from Late Proterozoic to Early Ordovician. Metavolcanic rocks high in the succession have yielded a U-Pb (zircon) age of 503.2 +7.8/-1.7 Ma (Trettin et al., 1987), probably Tremadoc.
2. A minor granitic sheet intruding the M'Clintock West body of the Thores Suite has given a U-Pb zircon age of 481 +7/-6 Ma (Trettin et al., 1982), late Arenig.
3. The Markham Fiord pluton (east of area shown in Fig. 6), interpreted to be of arc origin, has yielded a U-Pb (zircon) age of 462 ± 11 Ma (Trettin et al., 1982; not shown in Fig. 5 and 6), and the post-tectonic Cape Richards intrusive complex a U-Pb (titanite) age of 463 ± 5 Ma (Trettin et al., 1987), both Llandeilo.
4. Muscovite from a schist of succession II at upper M'Clintock Inlet, below the unconformity with the Taconite River Formation, produced a K-Ar age of 452 ± 8 Ma (Stevens et al., 1982), late Caradoc.
5. Fossils of early Caradoc age occur in the upper part of member A of the Cape Discovery Formation, and fossils of Ashgill age in the uppermost M'Clintock Formation and in the Ayles, Taconite River, Zebra Cliffs, and Lorimer Ridge formations (Fig. 6). These formations, as mentioned, are part of the unconformably overlying succession and post-date the M'Clintock Orogeny.

In summary, most previous data, combined with previous time scales, indicated that the main orogeny occurred in early Middle Ordovician (Llanvirn-Llandeilo) time. The somewhat younger K-Ar age on the mica schist at upper M'Clintock Inlet was regarded as a cooling age. In order to determine the age of the deformation more precisely, U-Pb and ^{40}Ar-^{39}Ar determinations were made on a small syntectonic granodiorite sheet at upper Ayles Fiord and a ^{40}Ar-^{39}Ar determination was made on the schist.

Syntectonic granodiorite at upper Ayles Fiord (88-TM-136C)

A concordant sheet of granodiorite intrudes the crest of a small anticline formed in schist of succession II of the Pearya Terrane north of upper Ayles Fiord (Fig. 5). In the region north of Ayles Fiord, succession II consists of sedimentary and volcanic rocks deformed mostly in the greenschist facies although staurolite occurs locally. This is the only granitic intrusion encountered during a few reconnaissance traverses but others may be present. The host rock is a pyritic schist, composed of quartz, muscovite, biotite, and minor garnet (almandine-pyrope). It is inferred that emplacement was broadly coeval with the folding and metamorphism of the host rocks.

The dated sample is composed of feldspar 52% (oligoclase 29%, K-feldspar 13%), quartz 31%, biotite 15%, chlorite 1%, and trace amounts of muscovite, chlorite, tourmaline, zircon, monazite, and apatite. Quartz and feldspar form myrmekitic intergrowths. Microscopic sigmoidal folds involving feldspar are apparent in some thin sections.

Most zircons from the intrusion form slender, elongate, euhedral crystals with high uranium concentrations, but some are small, equant, and euhedral to subhedral. Analyses of the equant crystals indicate inheritance, whereas analyses of two slender crystals indicate Pb loss (Fig. 7). Two analyses of

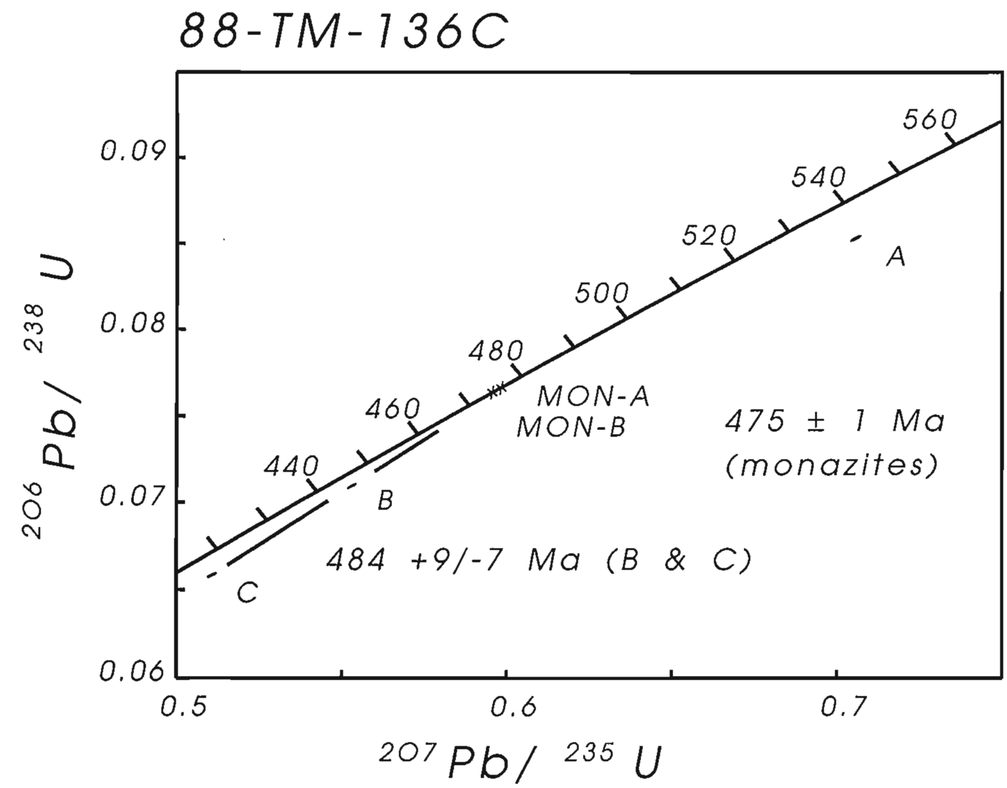


Figure 7.

Concordia diagram for Ayles Fiord granodiorite.

euhedral monazite with high uranium concentrations yield concordant ages at 474.3-475.9 Ma; accordingly, we interpret 475 ± 1 Ma as the age of crystallization of the granodiorite intrusion. The intercept age of the zircons with Pb loss is consistent with this age.

Biotite from the same sample yielded a plateau for the last 54% of the total of the gas released (Fig. 8). The total fusion and plateau ages, listed in Table 2, are 467.8 ± 2.5 and 471.0 ± 2.5 Ma, respectively, both slightly younger than the U-Pb age, due to cooling shortly after intrusion.

Schist at upper M'Clintock Inlet (88-TM-206B)

The schist at upper M'Clintock Inlet, previously dated by the K-Ar method, as mentioned, is unconformably overlain by the Taconite River Formation (Trettin, 1987a, Fig. 6). The new ^{40}Ar-^{39}Ar sample was taken from strata more than 500 m below the unconformity, and weathering effects are not apparent. The sample is composed mainly of quartz with smaller proportions of plagioclase, muscovite, biotite, chlorite, and garnet.

The prepared muscovite separate unfortunately was impure. Impurities were mainly a fine intergrowth of quartz and probably feldspar (Ca = 1.9% in the concentrate). The incremental release spectrum for muscovite is "hump-shaped", with a total fusion age of 445 ± 3 Ma; the top of the "hump" climbs to about 453 Ma for a gas fraction with about 27% ^{39}Ar (Fig. 9). As the age of each step varies, so does the Ca/K, indicating that muscovite with a low Ca/K is responsible for that part of the spectrum with the oldest age in the hump. It can be safely assumed that the minimum age of cooling of muscovite is 453 Ma.

Discussion

Deformation and metamorphism associated with the M'Clintock Orogeny is inferred to have been partly synchronous with the intrusion of the leucogranite at ca. 475 Ma. The youngest dated rock from the stratigraphic succession below the Ordovician unconformity is the 503 Ma rhyolite from succession II of the Pearya Terrane, which shifts the beginning of the M'Clintock Orogeny from early Middle Ordovician, as previously assumed, to the Early Ordovician (Tremadoc or Arenig). The origin and tectonic significance of the granitoid intrusion within the Thores Suite requires clarification, but the apparent age of 481 +7/-6 Ma may further restrict the collision to between 488 and 475 Ma.

The deformation probably preceded the ca. 468-471 cooling age of the biotite in the Ayles Fiord granodiorite and also the emplacement of the post-tectonic Cape Richards intrusive complex, dated at 463 ± 5 Ma on titanite.

The age determinations on the schist at upper M'Clintock Inlet are problematic. According to the current time scale (Fig. 6), the 453 Ma ^{40}Ar-^{39}Ar age, as well as the 452 Ma K-Ar age, are too young to be interpreted as cooling ages related to the M'Clintock Orogeny. At a depth of probably less than 1 km, the cooling should have been completed before the deposition of the nonmarine Cape Discovery Formation, at ca. 460 Ma or slightly later. On the other hand,

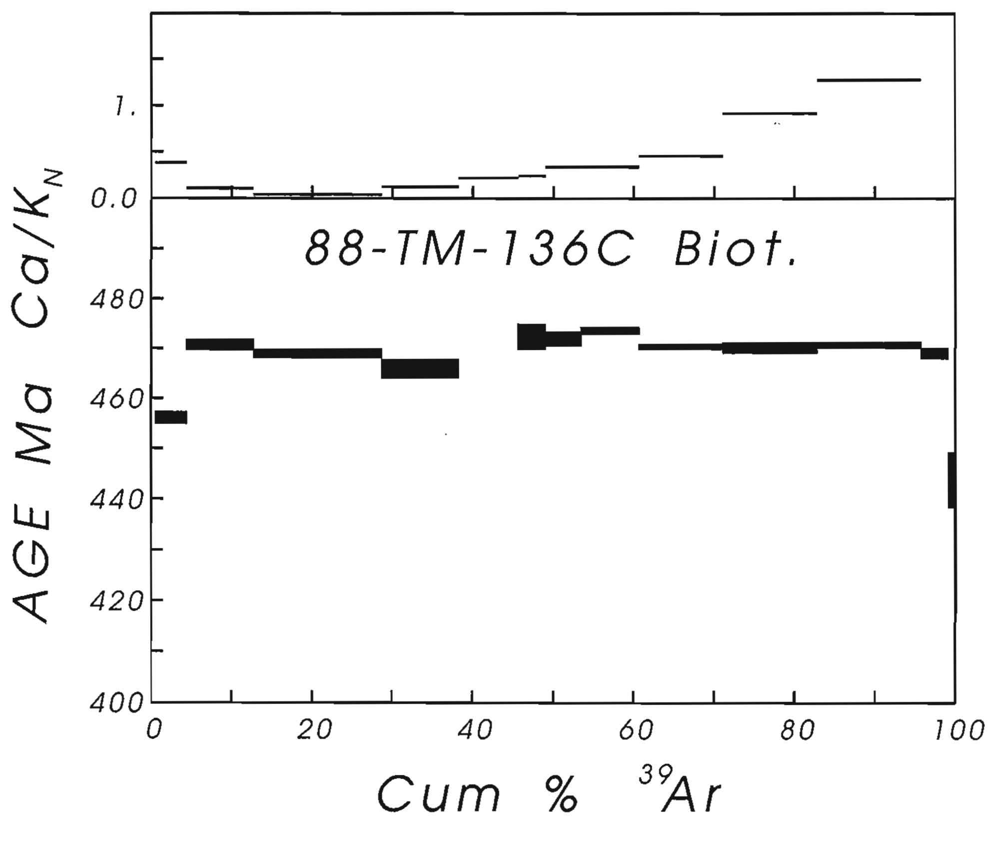


Figure 8.

^{40}Ar-^{39}Ar incremental-release spectrum of biotite from Ayles Fiord granodiorite. One gas release step was lost in the analysis during 900°C heating, and is noted as a gap in the age spectrum.

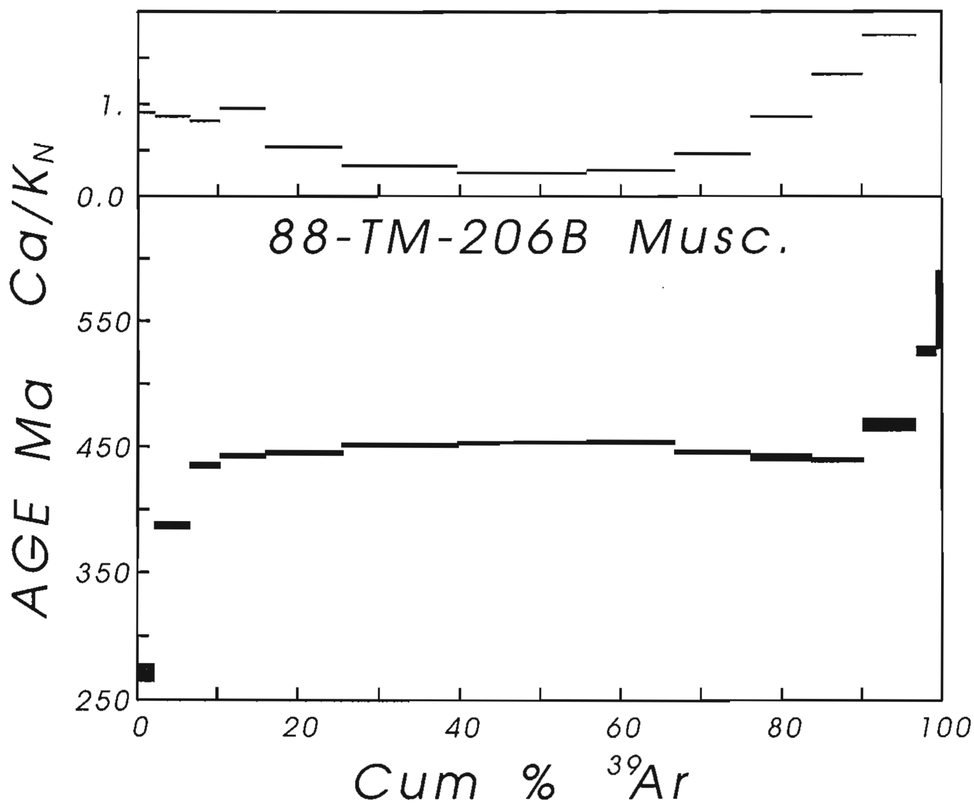


Figure 9.
^{40}Ar-^{39}Ar incremental-release spectrum of muscovite from schist at upper M'Clintock Inlet.

both ages coincide approximately with the age of the volcanic M'Clintock Formation. It is therefore suggested that both record a thermal overprint related to late Caradoc volcanism.

It was earlier inferred that the M'Clintock Orogeny had no counterpart in the Franklinian mobile belt, but that it had links to Ordovician events in the Caledonian-Appalachian mobile belt. How does the revised age assignment affect these two conclusions?

No significant unconformities are known to exist within the Lower Ordovician part of the Franklinian Shelf and deep water basin throughout the Arctic Islands and western and central North Greenland, nor are there syntectonic clastic sediments derived from northerly sources (Trettin et al., 1991; Higgins et al., 1991a,b).

Tectonic events of Early Ordovician and perhaps older age are known from various parts of the Caledonides. In contrast to the Franklinian mobile belt, an eastward-widening hiatus occurs in the shelf succession of northeastern Greenland at the base of the late Early Ordovician (early Arenig) and younger Wandel Valley Formation, which oversteps progressively older formations from west to east (Peel, 1985; Higgins et al., 1991a,b). The youngest strata beneath the unconformity (Tavsens Iskappe Group) are of earliest Ordovician (early Tremadoc) age, but Upper and Middle Cambrian units are truncated farther east. This hiatus indicates an uplift in northeastern Greenland, probably "related to the progressive closure of the Iapetus Ocean" (Higgins et al., 1991a) and perhaps also to the M'Clintock Orogeny (Surlyk, 1991).

A complex, polyorogenic history has been inferred for the central part of the Scandinavian Caledonides from ^{40}Ar-^{39}Ar determinations and geological data (Dallmeyer and Gee, 1986a,b). During the earliest phase, an accretionary prism, flanked by a volcanic arc in the west, formed over a westward dipping subduction zone. It cooled to temperatures below ca. 500°C in Arenig to Caradoc time, but eastward translation onto the Baltoscandian platform followed in the Silurian and Devonian.

In the Grampian Terrane of Scotland, compressional deformation and Barrovian metamorphism affected Upper Proterozoic-Lower Cambrian clastic sediments in earliest Ordovician (Tremadoc) time at ca. 500 Ma (Hutton, 1989 and references therein). This event, the Grampian Orogeny, is the earliest of a series of Palaeozoic terrane-accretion events in the British Isles.

In the Appalachians of central Newfoundland, emplacement of Lower Ordovician ophiolites of the Exploits Subzone of the Dunnage Zone onto the Gander Zone, the continental margin of Gondwana, occurred in late Arenig time (Coleman-Sadd et al., 1992). This event was closely followed by granitoid plutonism in the late Arenig (474 +6/-3 Ma) and by granitoid plutonism (465 ± 2 Ma and 464 ± 2 Ma), regional metamorphism, and emergence in Llandeilo-Llanvirn time. "The Penobscot Orogeny was roughly synchronous with the Taconian Orogeny on the opposite margin of the Iapetus, in which ophiolite of the Notre Dame Subzone was emplaced westward onto the continental margin of Laurentia" (op. cit., p. 348).

In summary, the present age assignment for the M'Clintock Orogeny supports the hypothesis that during this event the Pearya Terrane was part of the Caledonian-Appalachian belt and not yet part of the Franklinian mobile belt.

SILURIAN-DEVONIAN METAMORPHISM AND INTRUSION IN THE FOOTWALL OF THE PETERSEN BAY FAULT ZONE, NORTHWESTERN ELLESMERE ISLAND

The Petersen Bay Fault Zone juxtaposes Proterozoic gneiss of the Mitchell Point belt with Ordovician and Silurian units of the deep-water basin (Fig. 10). The metamorphic grade of the Clements Markham Fold Belt is mostly below the greenschist facies, but increases to amphibolite facies towards the fault zone where staurolite, kyanite, and garnet occur (Frisch, 1976; Trettin and Frisch, 1987; Klaper and Ohta, unpublished data). On the peninsula between Yelverton and Kulutingwak inlets, this metamorphism affects complexly folded strata of the Petersen Bay assemblage, a unit of sedimentary and volcanic rocks of probable Ordovician to Early Silurian age, and the overlying Imina Formation, a flysch of Early Silurian (late Llandovery to possibly Wenlock age). The age of the Petersen Bay assemblage is based on lithological correlation with the less metamorphosed Kulutingwak Fiord assemblage farther south (Bjornerud, 1991), which has yielded conodonts (G.S. Nowlan, pers. comm., 1992). The age of the Imina Formation is based on fossil identifications from the Imina Formation itself and from underlying and overlying units (Trettin and Frisch, 1987 and references therein). The Petersen Bay assemblage is intruded by a pegmatite discussed below and by diabase dykes of probable Late Cretaceous age.

Metamorphic isograds in the Petersen Bay assemblage and Imina Formation are approximately parallel with the fault zone but cut across the trends of fold axes formed in the Imina Formation, indicating that the metamorphism postdates an early phase of folding of the two units but was related to movements on the Petersen Bay Fault Zone (Klaper and Ohta, unpublished data). These movements and the related metamorphism clearly are younger than Early Silurian (Llandovery). Moreover, the absence of unconformities

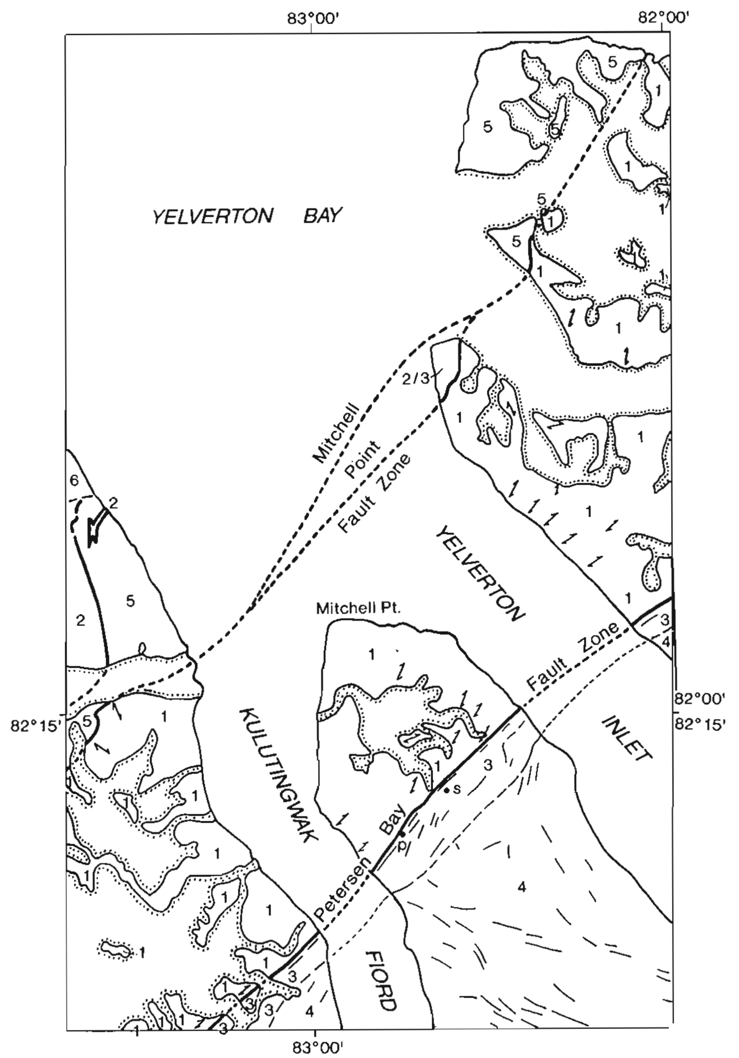
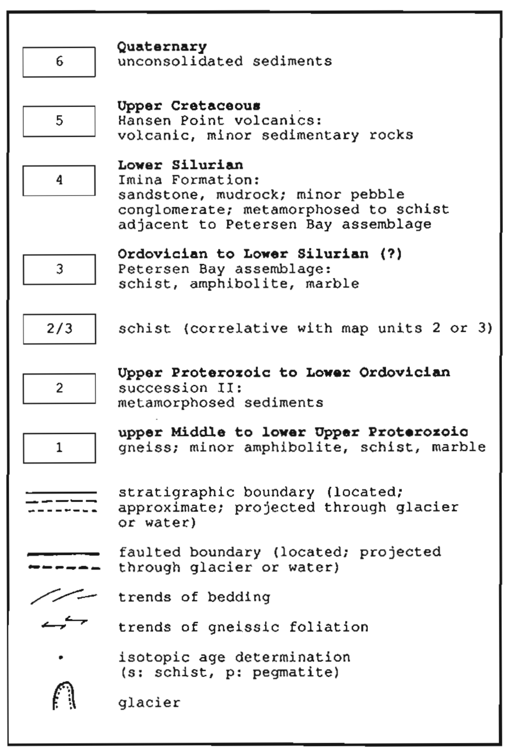


Figure 10.

Generalized geology in vicinity of Petersen Bay Fault Zone, northwestern Ellesmere Island.



within the Llandovery to middle Ludlow successions both of deep-water basin and Pearya Terrane suggests that the movements occurred later than middle Ludlow. An upper age limit is defined by the absence of regional metamorphism in the upper Lower Carboniferous (Viséan) and younger strata of the Sverdrup Basin. Thus, available geological evidence places the movements on the Petersen Bay Fault Zone and related metamorphism in the middle Ludlow to Early Carboniferous interval.

The Mitchell Point belt is terminated on the north by the Mitchell Point Fault zone, which affects rocks as young as Late Cretaceous. This fault zone probably also originated in Late Silurian to Early Carboniferous time, but was a site of normal faulting in the Late Cretaceous and of reverse faulting in the early Tertiary (Trettin and Frisch, 1987; Ohta and Klaper, unpublished data). Because of the absence of strata younger than the Imina Formation, the post-Llandovery history of the Petersen Bay Fault Zone is uncertain, but analogy with the Mitchell Point Fault Zone suggests recurrent movements.

To date the metamorphism, samples of schist and pegmatite were collected from the Petersen Bay assemblage by Y. Ohta (Norsk Polar Institutt, Oslo, Norway). It is assumed that the pegmatite was produced by the same metamorphic event as the schist, but at greater depth.

Schist (sample 88-TM-OHT-40)

The sample is composed mainly of quartz with lesser amounts of hornblende, small amounts of plagioclase and calcite, and trace amounts of biotite. Hornblende crystals are up to about 7 mm long and show sieve texture, containing numerous inclusions mainly of quartz ranging in size approximately from 15 to 250 μm.

In the ^{40}Ar-^{39}Ar analysis, the early gas release shows erratic ages as high as 2330 Ma, indicating the presence of excess argon (Table 2; Fig. 11). After release of this gas the ages are more consistent, with release from 42 to 96% ^{39}Ar indicating an age of 433 ± 3 Ma. These fractions may also contain some excess argon as the spectrum suggests a decrease to a minimum age of 424 Ma for the last significant gas release. The sample's age is therefore inferred to be at least 424 Ma, but possibly as old as 436 Ma. The integrated age of all gas is significantly older at 509 ± 4 Ma because of the presence of excess argon.

Pegmatite (sample 88-TM-0HT-3)

A representative thin section of the pegmatite consists of medium grained quartz 66%, albite 19%, and muscovite 15% with trace amounts of garnet.

The ^{40}Ar-^{39}Ar spectrum of muscovite (Fig. 12) is irregular with the initial steps in the first 5% gas release suggesting disturbance at an age of 200 Ma or less. The remaining gas fractions have ages ranging from 318 to 326 Ma with an age of 322 ± 2 Ma for 87% of the gas. The integrated age is similar at 319 ± 2 Ma. The sample shows disturbance in the high temperature release but the limited age range suggests that the disturbance is minor or that it has almost completely reset the muscovite from an age older than 326 Ma.

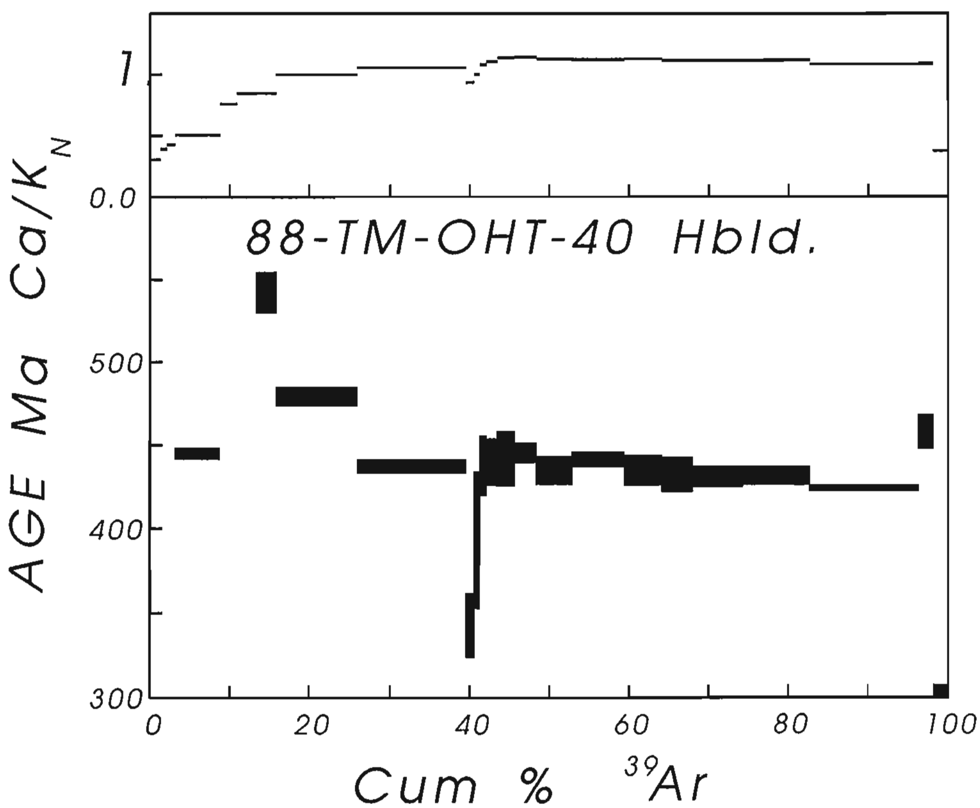


Figure 11.

^{40}Ar-^{39}Ar incremental-release spectrum of hornblende from schist south of Petersen Bay Fault Zone.

Discussion

As mentioned above, the metamorphism south of the Petersen Bay Fault Zone clearly is no older than the Llandovery-Wenlock boundary, ca. 430 Ma, and probably younger than middle Ludlow, ca. 415 Ma. The 433 Ma plateau age therefore appears to be too old. The plateau age of the pegmatite, on the other hand, is near the Early-Late Carboniferous boundary and considerably younger than expected. It should be noted that the Petersen Bay Fault Zone was a site of recurrent movements and hypabyssal intrusion that probably have affected the ^{40}Ar-^{39}Ar ages. The two age determinations cannot be rejected unequivocally but require confirmation by other samples and methods.

LATE DEVONIAN GRANITOID PLUTONISM IN NORTHERN AXEL HEIBERG AND NORTHWESTERN ELLESMERE ISLANDS

The distribution of Phanerozoic plutons and compositionally related dyke/plug swarms in northern Axel Heiberg and northwestern Ellesmere islands is shown in Figure 13. The relatively large Cape Woods pluton has a titanite age of 390 ± 10 Ma (Trettin et al., 1987), probably late Early Devonian (Pragian/Emsian). The remaining intrusions are small. Three of these, discussed here, have Late Devonian to earliest Carboniferous(?) isotopic ages, and one has a late Early Carboniferous age (Petersen Bay pluton, see below). All of these rocks were intruded after the main deformation in their host rocks.

Tonalite north of Rens Fiord, Axel Heiberg Island (88-TM-103C)

On the west coast of northernmost Axel Heiberg Island, south-southwest of Cape Thomas Hubbard, four small plutons, occurring in a 4.4 km long, south-southwest trending belt, intrude quartzite and slate of the Lower Cambrian Grant Land Formation of the Northern Heiberg Fold Belt (Fig. 3 of Trettin, 1987b, map-unit 7a). The dated sample is a medium grained tonalite, composed of plagioclase (zoned oligoclase-andesine) 48%, quartz 26%, biotite 26%, and trace amounts of chlorite, apatite and magnetite.

Four zircon analyses from this sample indicate that zircon inheritance is common; a precise age is precluded because all four analyses are very discordant. Nevertheless, the low uranium contents, large zircon grain size and equant shape, and strong abrasion of the zircons indicate that Pb loss probably was unimportant. The four analyses are aligned in an array (Fig. 14) with a lower intercept of 366 ± 6 Ma (MSWD = 236). A regression of fractions A, B, and C has an MSWD of 18 and an age of 360 ± 3 Ma. We interpret these data to indicate that the crystallization of the pluton occurred 360-370 Ma ago (Frasnian-Fammenian). There is a preponderance of Precambrian zircons in this sample with a mean age of about 2 Ga.

A ^{40}Ar-^{39}Ar determination on biotite (Fig. 15) from this sample has a plateau over 95% of gas release and defines an age of 360 ± 2 Ma (Tournaisian), indicating rapid cooling following intrusion.

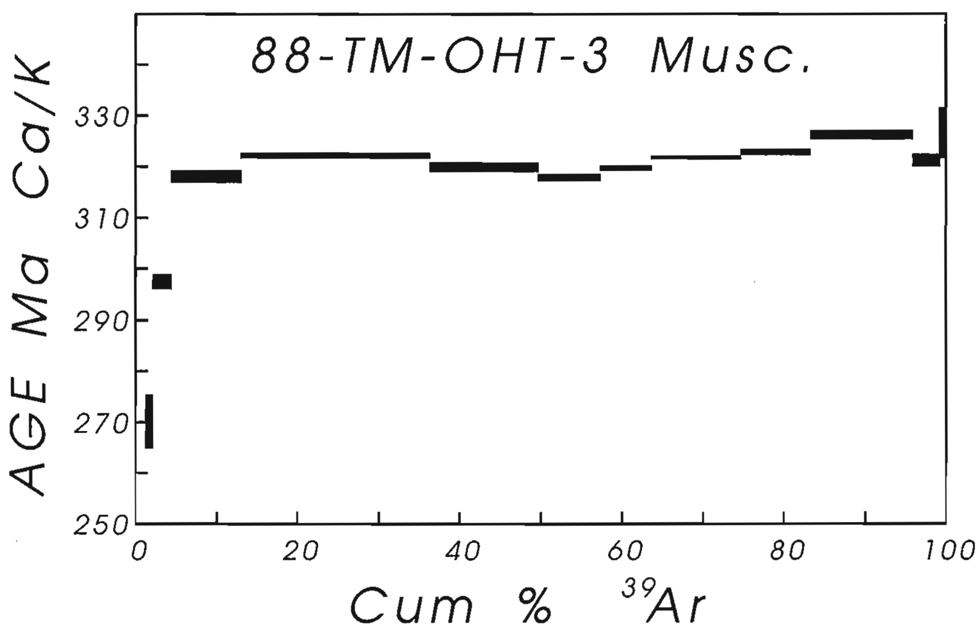


Figure 12.

^{40}Ar-^{39}Ar incremental-release spectrum of muscovite from pegmatite south of Petersen Bay Fault Zone.

The present determinations confirm and refine a previous K-Ar (biotite) age of 367 ± 25.5 Ma (recalculated from Lowden et al., 1963).

Trachyandesite south of Rens Fiord (86-TM-107A)

South of eastern Rens Fiord, 12 or more plugs and dykes of porphyritic felsite intrude quartzite and slate of the Lower Cambrian Grant Land Formation. They lie in a belt, 11 km wide and up to about 1 km wide, that is parallel to the predominant northwest-southeast structural trend of this area (Fig. 3 of Trettin, 1987b, map-unit 7c). The dated sample is from a dyke, about 700 m long. The felsic intrusions are similar in mineralogy and texture. Phenocrysts, mainly of plagioclase (zoned from sodic andesine to sodic labradorite or altered to albite) with small proportions of K-feldspar and of biotite, variably replaced by chlorite, are set in a groundmass of feldspar with variable amounts of quartz and minor amounts of biotite and chlorite. In addition, the rocks are partly replaced by calcite and dolomite. Two specimens showing little carbonate alteration were analyzed chemically. Major element classification according to Irving and Baragar (1971) places them in the fields of calc-alkaline rhyolite and dacite, respectively, whereas trace element ratios after Winchester and Floyd (1977) place both in the field of trachyandesite. Such discrepancies are common in analyses from lower and middle Palaeozoic volcanics of northern Ellesmere and Axel Heiberg islands. The trace element ratios always have given more consistent results and we therefore prefer the term trachyandesite, assuming that the matrix quartz is of replacement origin.

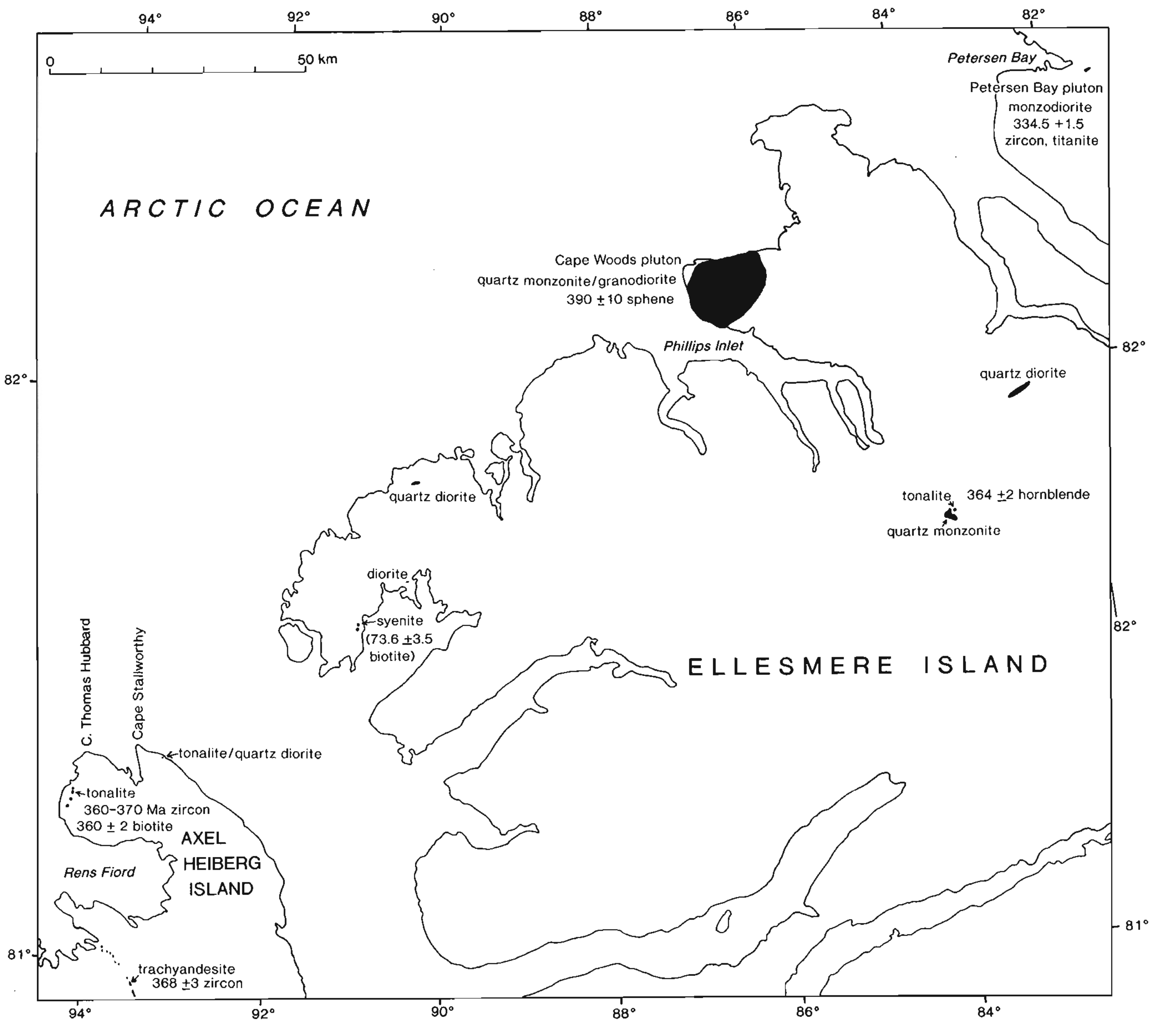


Figure 13. Distribution of Phanerozoic granitoid plutons and major dykes or dyke/plug swarms in northern Axel Heiberg Island and northwestern Ellesmere Island.

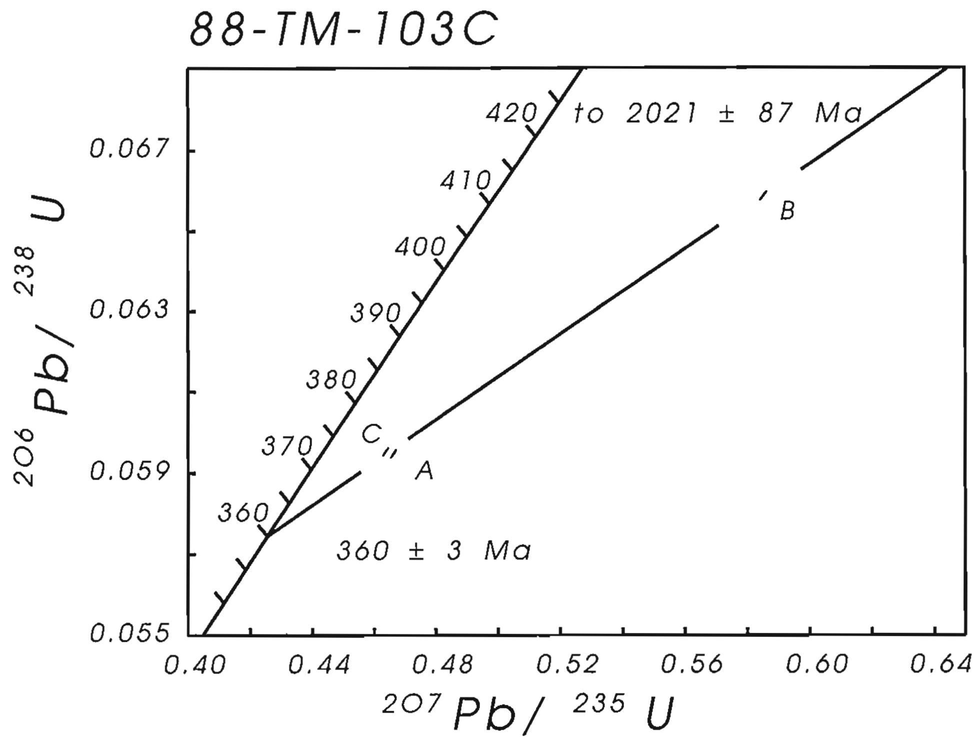


Figure 14.

Concordia diagram for tonalite north of Rens Fiord.

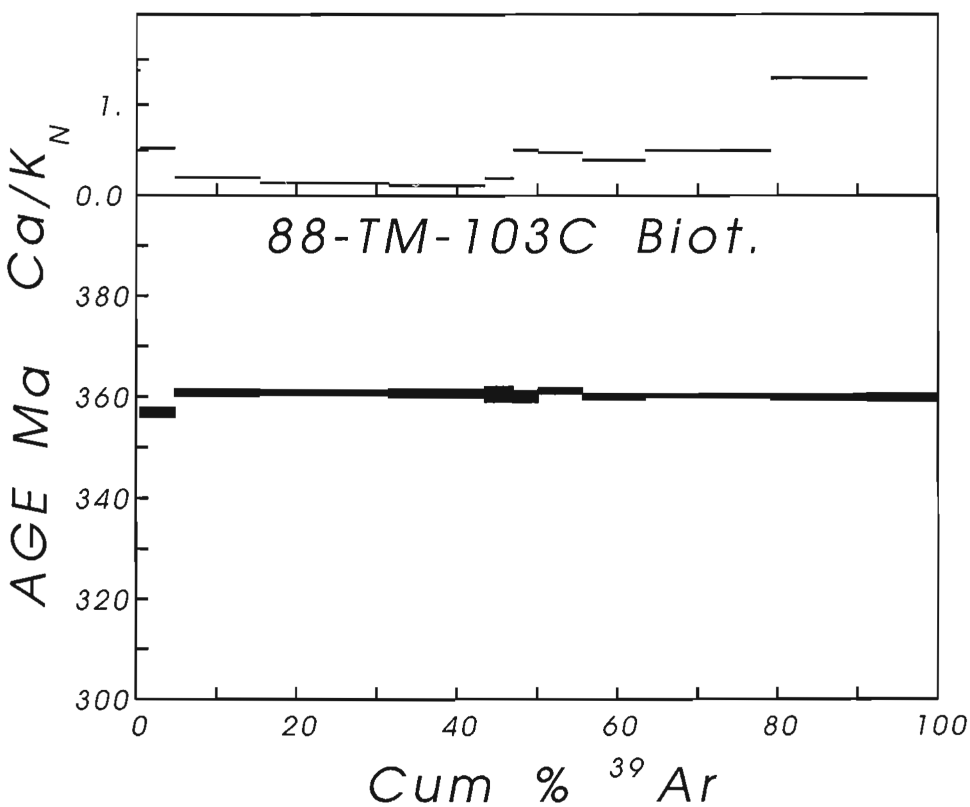


Figure 15.

^{40}Ar-^{39}Ar incremental-release spectrum of biotite from tonalite north of Rens Fiord.

The U-Pb systematics of this sample are similar to the previous sample 88-TM-103C in that zircon inheritance is present and the analyses are generally discordant. Zircons have moderate uranium concentrations (about 300-600 ppm), and are relatively elongate, with some inclusions and cloudy areas. The lower intercept of three abraded zircon fractions is 365.7 ± 0.5 Ma with an MSWD of 1.2 and upper intercept of 2436 Ma (Fig. 16). Analysis C is nearly concordant with a ^{207}Pb-^{206}Pb age of 369.0 ± 1.6 Ma. Using this as an older limit, we interpret the crystallization age to be between 365 and 371 Ma, and accept 368 ± 3 Ma as the preferred age. This is near the Frasnian-Famennian boundary in the Late Devonian.

Tonalite south-southeast of Phillips Inlet (88-TM-208C)

A small plug of tonalite intrudes turbidites of the undifferentiated Imina and Lands Lokk formations (upper Llandovery to lower Ludlow) in an area about 19 km south-southeast of the head of Phillips Inlet (Fig. 13) in the Clements Markham Fold Belt. A thin section of the dated sample is composed of feldspar 56% (mainly plagioclase), quartz 18%, hornblende 19%, biotite 3%, chlorite 3%, apatite 1%, and trace amounts of opaque minerals and epidote. Clinopyroxene is present in related samples of tonalite nearby.

These intrusions lie 0.7 to 5 km northwest of a major west-southwest-trending, linear fault that separates the Silurian Imina and Lands Lokk formations on the northwest from the Lower Cambrian Yelverton assemblage on the southeast.

A ^{40}Ar-^{39}Ar determination on hornblende gives a plateau age of 364.1 ± 2.0 Ma, representing 84% of the total gas (Fig. 17). Because this pluton is late to post-tectonic, having intruded relatively cool country rocks, we interpret 364 Ma (Famennian) as the approximate age of emplacement of this intrusion.

Discussion

Although the three dated intrusions are different in mineral composition and setting, they are discussed together because they are close in age (Fig. 18). The presence of inherited zircon in the two dated samples from northern Axel Heiberg Island shows that the granitoids in this area are at least partly of crustal derivation. The crustal sources either were crystalline basement rocks of Early Proterozoic age or sedimentary rocks derived from such basement. In contrast, upper intercept ages from volcanic and intrusive rocks of the Pearya Terrane are Middle Proterozoic in age (Fig. 19), in accord with the Grenville-age of the crystalline basement in that region.

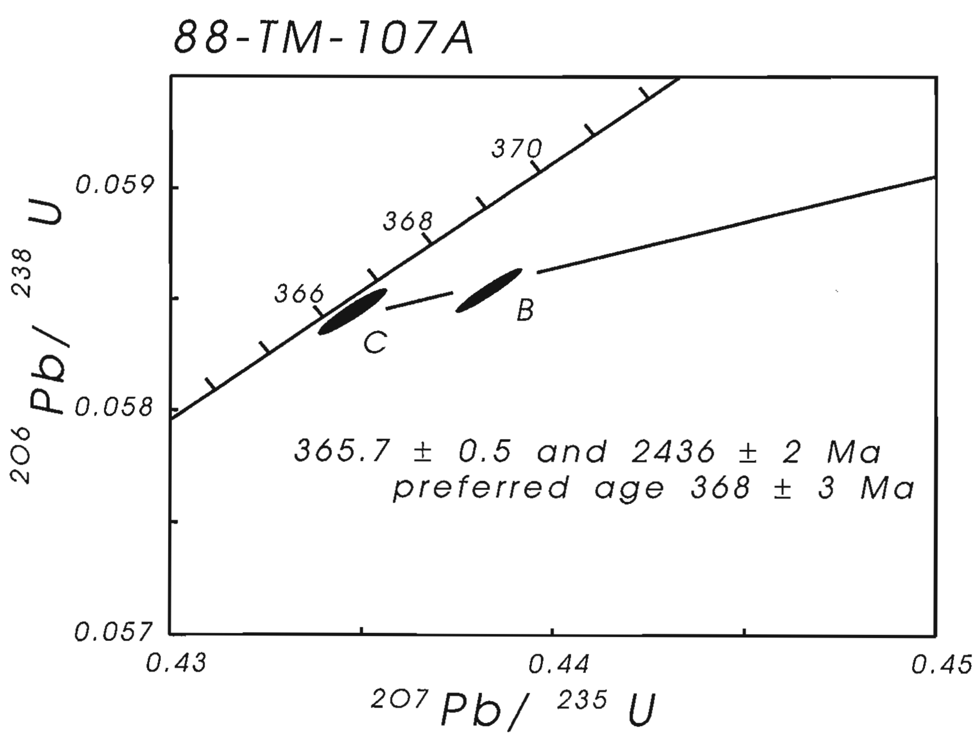


Figure 16.

Concordia diagram for trachyandesite south of Rens Fiord.

The geological evidence bearing on the tectonic significance of these intrusions is limited because strata of Late Devonian and earliest Carboniferous age are not preserved in the region. According to present time scales, all three fall in the Late Devonian (late Frasnian and Famennian), were coeval with the deposition of the youngest preserved formations in the Middle-Late Devonian foreland basin in the central and southern parts of the Arctic Islands (Okse Bay Group, Fig. 18), and predate the deformation of the foreland basin. Deformation apparently advanced south and southeast, affecting the foreland basin in latest Devonian-earliest Carboniferous time (cf. Thorsteinsson and Tozer, 1970; Embry and Klovan, 1976; Embry, 1991a; Harrison, 1991; Harrison et al., 1991). However, the correlation of the Upper Devonian sedimentary and plutonic rocks remains problematic because of uncertainty about the radiometric ages of the Devonian and Early Carboniferous Stage boundaries. Estimates by Harland et al. (1989) will have to be revised because of a new, precise age for the Devonian-Carboniferous boundary (Claoué-Long et al., 1992), which is 9 Ma younger.

According to the most widely accepted hypothesis for the opening of the Amerasian Basin of the Arctic Ocean Basin, the North Yukon and Arctic Alaska were part of the Franklinian mobile belt prior to Mesozoic counterclockwise rotation (Lawver and Scotese, 1990). It is interesting to note that both areas contain granitoid intrusions of Late Devonian age (cf. Mortensen and Bell, 1991 and references therein).

A rift-related Early Carboniferous intrusion in northwestern Ellesmere Island (88-TM-117D)

A small granitoid intrusion, about 1.3 km long and termed the Peterson Bay pluton, intrudes metasedimentary rocks of succession II of the Pearya Terrane (probably Late Proterozoic or Cambrian) in an area about 2.5 km east-southeast of Petersen Bay, northwestern Ellesmere Island (Fig. 13). The dated sample is a medium to predominantly fine grained monzodiorite composed of feldspar 88% (sodic albite 64%, K-feldspar 24%), biotite 6%, chlorite 2%, quartz 2%, epidote 1%, opaque minerals 1%, and trace amounts of hornblende and calcite. The pluton is truncated on its southeastern side by a major fault.

Good quality euhedral magmatic zircon and titanite were abraded and analyzed from the sample, and they yielded excellent concordant data. Only fraction B was discordant, with clear evidence of inherited zircons (Table 1). Two zircon analyses graze the concordia curve adjacent to 333 Ma, but with slightly older $^{207}Pb/^{206}Pb$ ages of 335 ± 1.5 Ma (Fig. 20). This slight discrepancy could result from either slight amounts of Pb loss or an initial deficiency in ^{230}Th-derived ^{206}Pb (see Coleman and Parrish, 1991, for further details on the latter explanation). An age range of 333-336 Ma is permissible for this intrusion, and we assign an age of 334.5 ± 1.5 to cover the uncertainties.

The age of the pluton corresponds to late Early Carboniferous, latest Viséan (Fig. 18). In the Arctic Islands the late Viséan is represented by predominantly nonmarine,

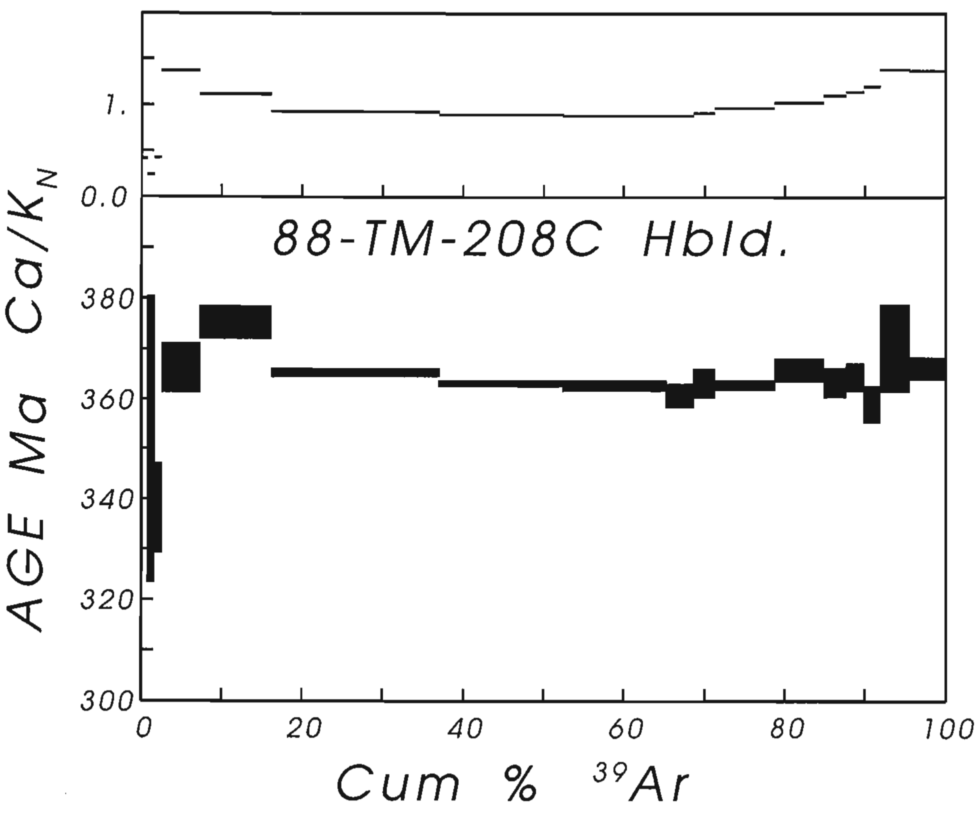


Figure 17.

^{40}Ar-^{39}Ar incremental-release spectrum of hornblende from granodiorite south-southeast of Phillips Inlet.

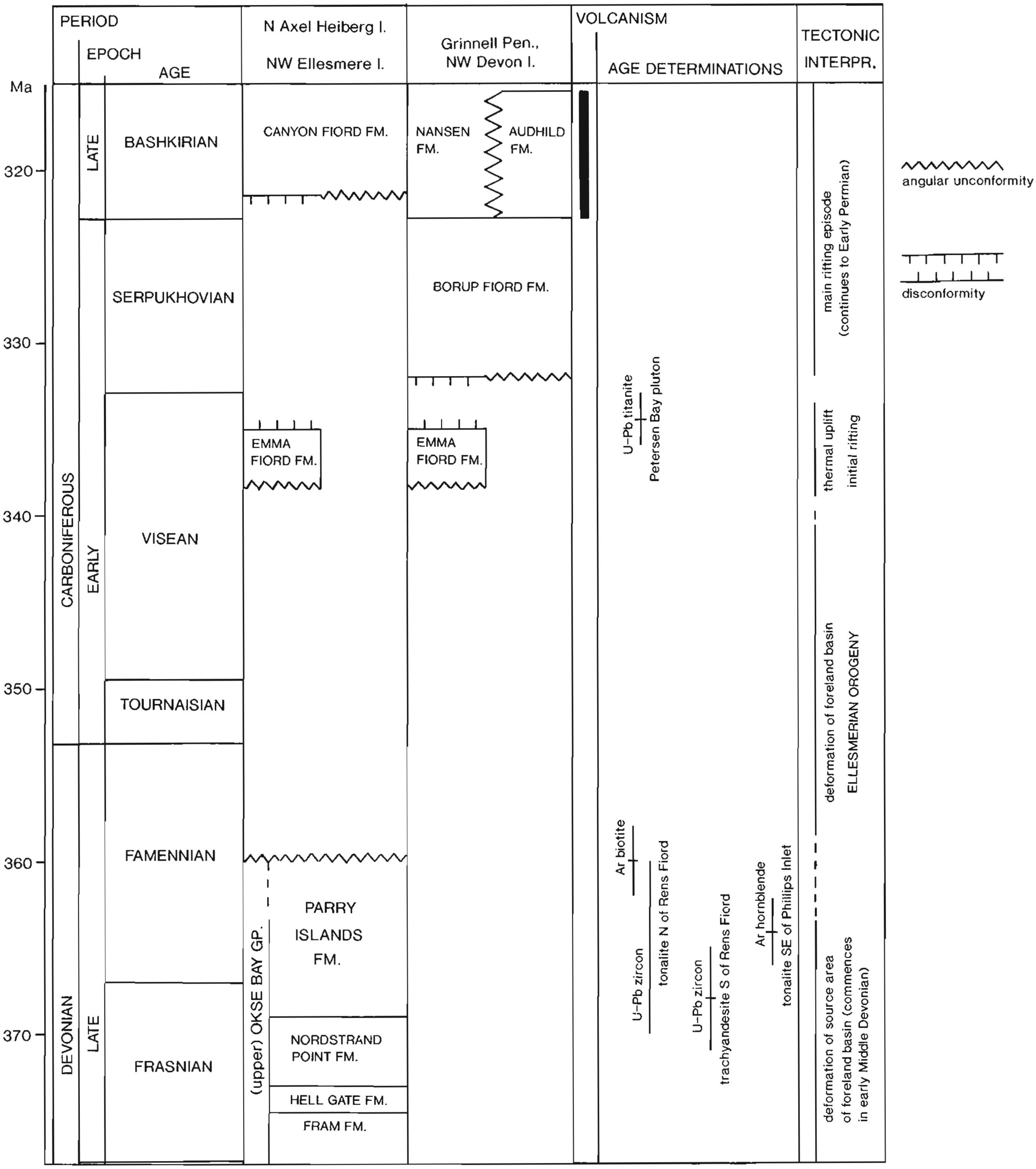


Figure 18. Correlation of Upper Devonian and Lower Carboniferous formations and intrusions; stratigraphy adapted from Embry, 1991a (Devonian) and Utting et al. (1989)(Carboniferous); time scale from Harland et al. (1989).

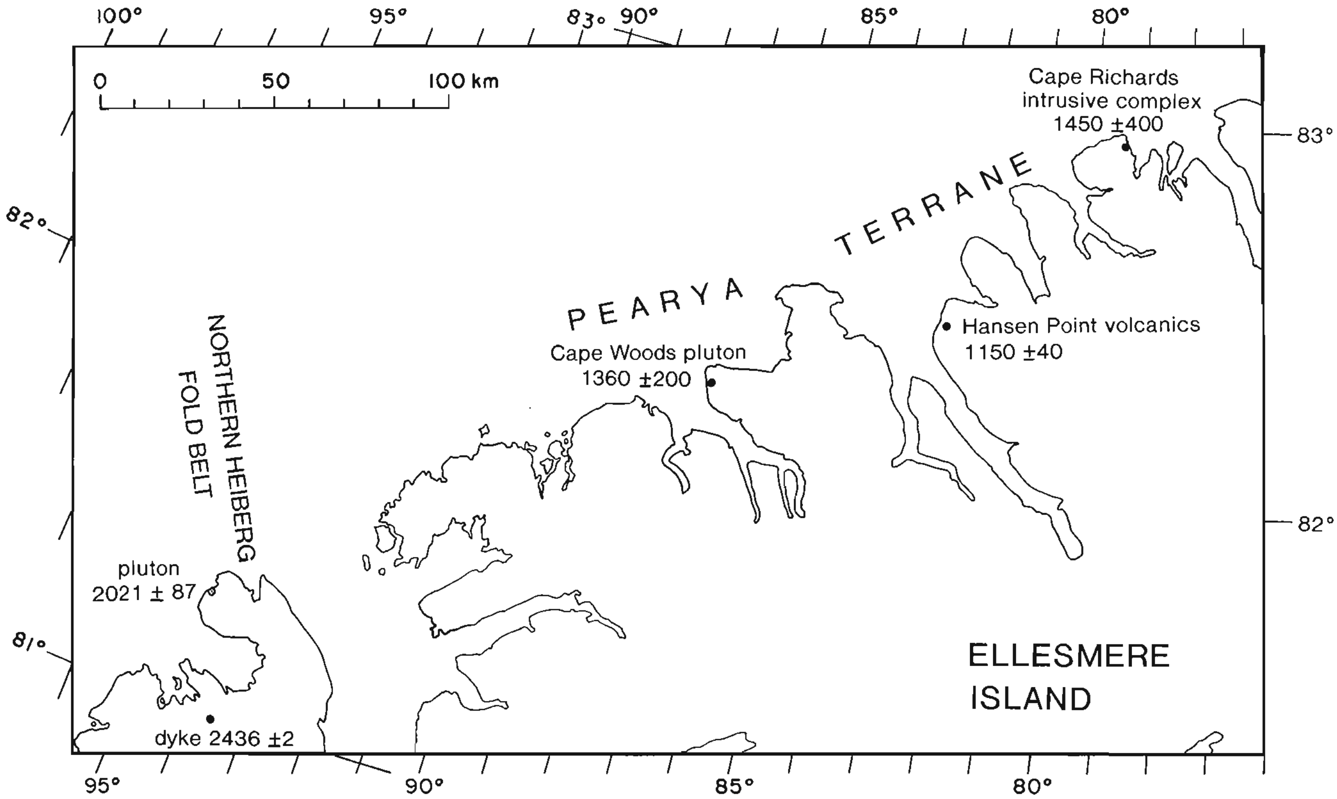


Figure 19. Upper intercept ages from zircon determinations on Ordovician to Cretaceous volcanic and intrusive rocks, Pearya Terrane and Northern Heiberg Fold Belt (this paper) and Pearya Terrane (Trettin et al., 1987; Trettin and Parrish, 1987).

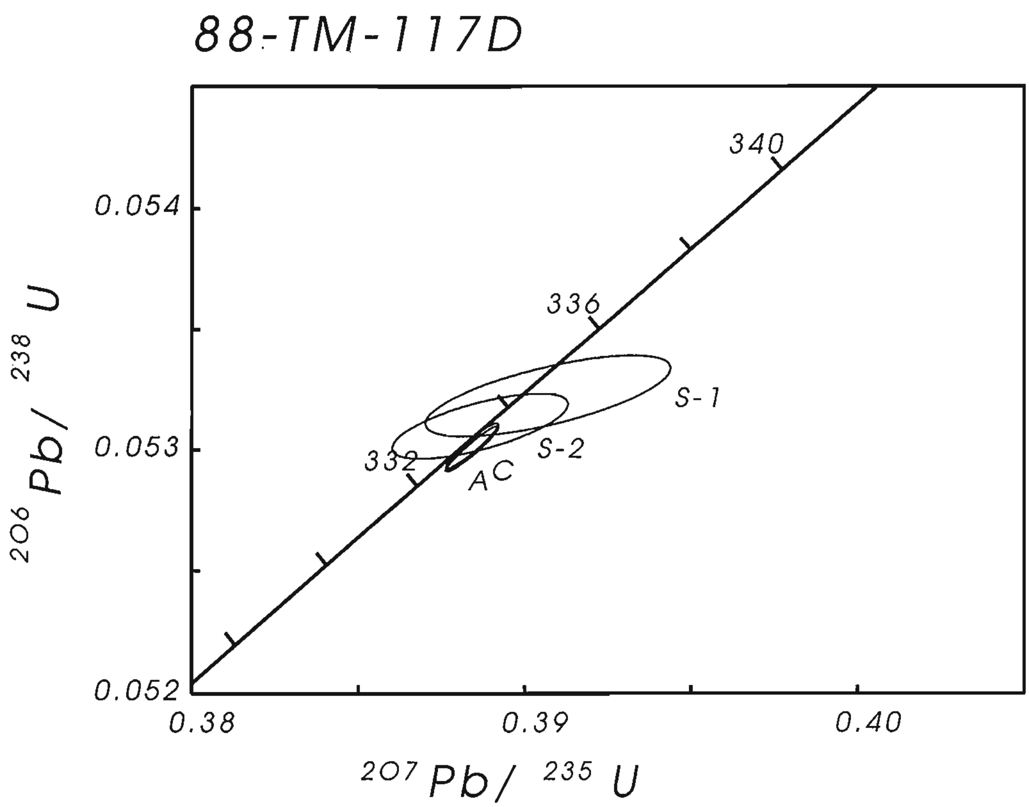


Figure 20.

Concordia diagram for Petersen Bay pluton.

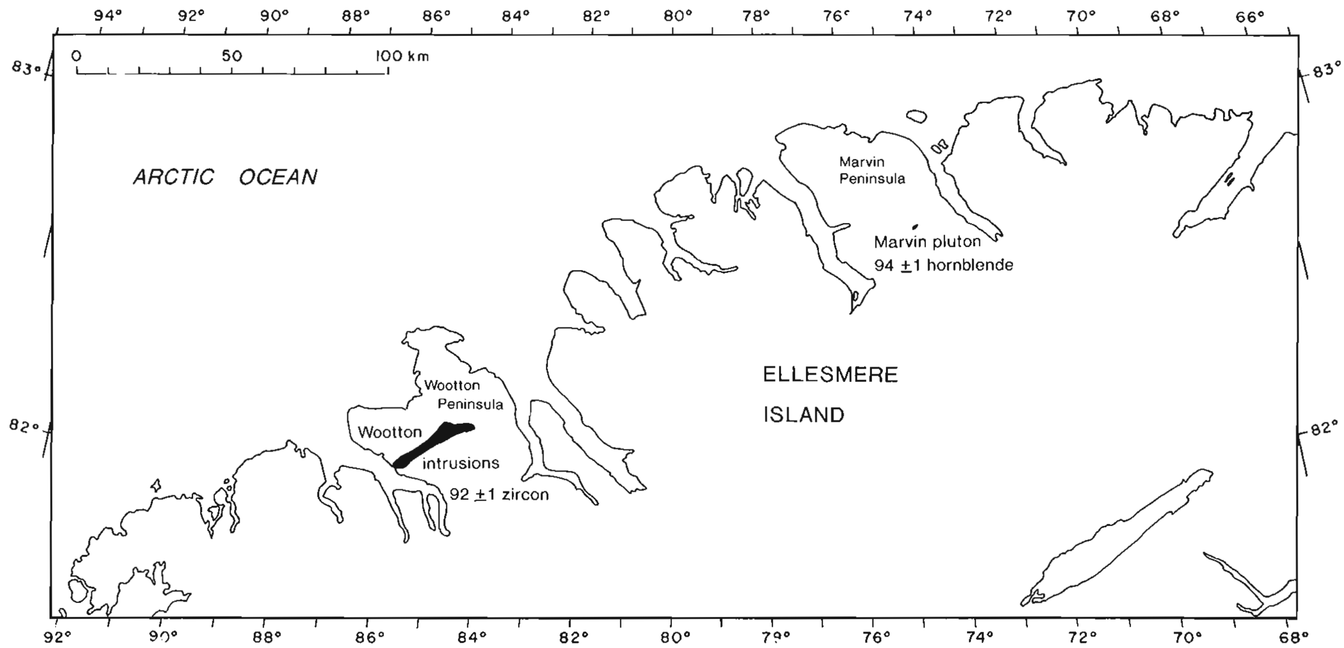


Figure 21. Dated Cenomanian intrusions, northern Ellesmere Island.

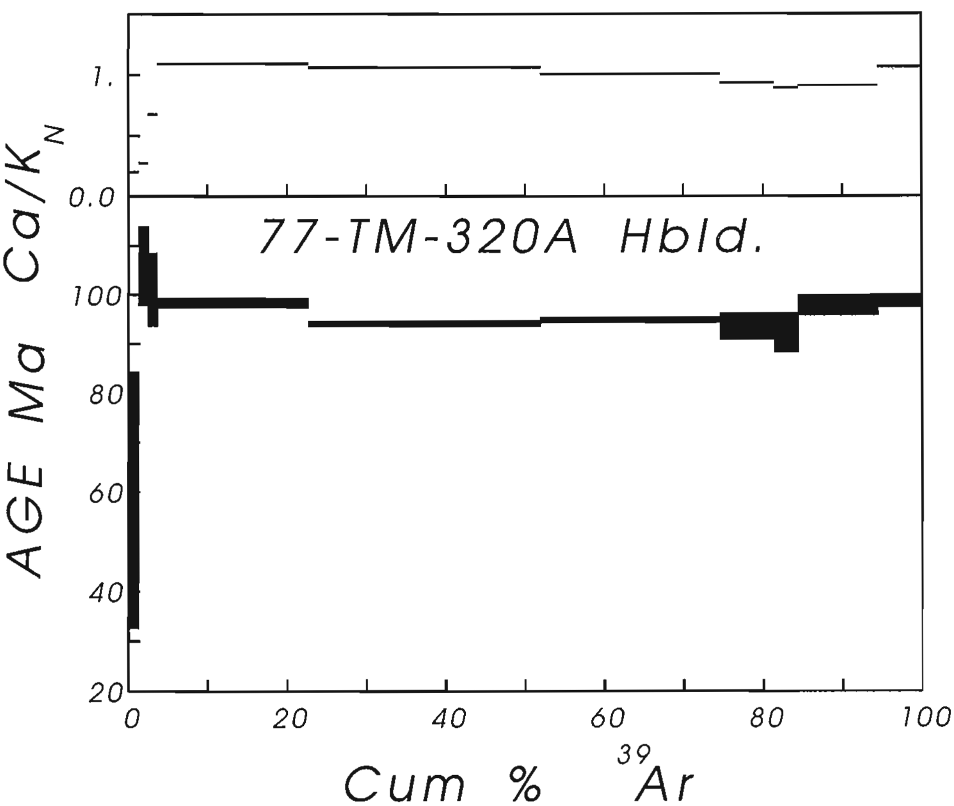


Figure 22.

^{40}Ar-^{39}Ar incremental-release spectrum of hornblende from Marvin pluton.

in part lacustrine, clastic sediments, preserved only in a few widely separated areas (Thorsteinsson, 1974; Davies and Nassichuk, 1991; Utting et al., 1989; Mayr, in press). These are the oldest known deposits of the Sverdrup Basin, perhaps produced by a thermal uplift that preceded the rifting which generated the Sverdrup Basin proper (Beauchamp et al., 1989). Applying the palynological data of Utting et al. (1989) to the isotopically calibrated time scale, the Petersen Bay pluton appears to be slightly younger than the Emma Fiord Formation, but the time scale is not well enough constrained to be certain. This is the first reported igneous rock of Viséan age in the Arctic Islands, and it shows that the initial thermal uplift or rifting of the Sverdrup Basin was accompanied by limited magmatism. Three phases of localized basaltic volcanism (Thorsteinsson, 1974; Trettin, 1988) were associated with the subsequent rifting, which extended in age from latest Early Carboniferous (Serpukhovian) to Early Permian.

Late Cretaceous Marvin pluton, Ellesmere Island (77-TM-320A)

A small quartz monzodiorite-quartz monzonite-granodiorite pluton intrudes Upper Ordovician sediments on Marvin Peninsula, northernmost Ellesmere Island (Fig. 21). Two hornblende separates yielded K-Ar dates of 94.2 ± 10 Ma and 91.6 ± 9.6 Ma (Stevens et al., 1982). These ages coincide with the 92.0 ± 1.0 Ma U-Pb (zircon) age of a large bimodal intrusion on Wootton Peninsula, northwestern Ellesmere Island (Trettin and Parrish, 1987). No minerals suitable for U-Pb dating were recovered from the Marvin pluton. To verify the earlier determinations with their broad confidence limits, the previous hornblende concentrate was analyzed by the ^{40}Ar-^{39}Ar method. The incremental-release spectrum (Fig. 22) provides a plateau age involving four steps with 62% of the gas released of 94.2 ± 0.7 Ma age, only slightly older than that of the Wootton intrusion. A similar apatite fission track date of 92.4 ± 6.4 Ma obtained by R.H. McCorkel (pers. comm., 1990) is in agreement with these data.

All three ages are Cenomanian and permit correlation with the upper part of the basaltic Strand Fiord Formation of western Axel Heiberg Island (Embry and Osadetz, 1988; Embry, 1991b). Together, these intrusive and volcanic rocks indicate a major magmatic event in the Arctic Islands, probably related to the tectonic events of the Amerasian Basin and Alpha Ridge of the Arctic Ocean.

CONCLUSIONS

New isotopic age determinations confirm the Grenville age of the crystalline basement of the Pearya Terrane and demonstrate that the M'Clintock Orogeny began in the Early Ordovician (prior to 475 Ma) rather than in the early Middle Ordovician as previously assumed. Both conclusions strengthen the hypothesis that the terrane was part of the Caledonian mobile belt prior to its accretion to the Franklinian mobile belt.

Combined U-Pb and ^{40}Ar-^{39}Ar determinations indicate that intrusion of widespread but small granitoid plutons of variable compositions occurred in northern Axel Heiberg and northwestern Ellesmere islands ca. 360-370 Ma ago (Late Devonian-early Tournaisian?) prior to the deformation of a foreland basin farther south.

Discovery of a monzodiorite pluton in northwestern Ellesmere Island with a crystallization age of ca. 335 Ma indicates that the initial rifting of the Sverdrup Basin in the late Early Carboniferous (Viséan) was accompanied by felsic magmatism.

Finally, further evidence in support of ca. 94-92 Ma (Late Cretaceous, Cenomanian) bimodal magmatism has been presented.

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U-Pb ages from the Archean Whitehills-Tehek lakes supracrustal belt, Churchill Province, District of Keewatin, Northwest Territories

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Roddick, J.C., Henderson, J.R., and Chapman, H.J., 1992: U-Pb ages from the Archean Whitehills-Tehek lakes supracrustal belt, Churchill Province, District of Keewatin, Northwest Territories; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 31-40.

Abstract

U-Pb dating of zircons from a metadacite at the structural base of the Woodburn Lake group establishes an age of 2796 to 2752 Ma for volcanism in the Whitehills-Tehek lakes area of the central Churchill Province. Most detrital zircons separated from immediately overlying psammitic metagreywacke yield an age of 2769 ± 11 Ma. The supracrustal rocks are polydeformed and intruded by late- to post-kinematic granitic batholiths that yield a combined U-Pb zircon age of 2612 ±4 Ma. The youngest granitic rocks in the region are undeformed dykes that give a U-Pb monazite age of 1835 ± 1 Ma. These results support the correlation of Woodburn Lake group with Prince Albert Group to the northeast, and establish a late Archean age for the last significant deformation in the region of Whitehills and Tehek lakes.

Résumé

La datation par la méthode U-Pb d'une métadacite provenant de la base structurale du groupe de Woodburn Lake a permis d'assigner un âge de 2 796 à 2 752 Ma au volcanisme de la région des lacs Whitehills et Tehek, au centre de la province de Churchill. La plupart des zircons détritiques provenant des métagrauwackes psammitiques immédiatement sus-jacents donnent un âge de 2 769 ± 11 Ma. Les roches supracrustales sont polydéformées; des batholites granitiques tardi-cinématiques à post-cinématiques les recoupent et donnent un âge U-Pb sur zircon combiné de 2 612 ±4 Ma. Les plus jeunes roches granitiques de la région sont des dykes non déformés dont l'âge U-Pb sur monazite est de 1 835 ± 1 Ma. Ces résultats corroborent la corrélation du groupe de Woodburn Lake avec le Groupe de Prince Albert au nord-est et permettent d'associer ladernière déformation importante dans la région des lacs Whitehills et Tehek à l'Archéen tardif.

INTRODUCTION

The central Churchill Province (Fig. 1), like the Slave Province, is characterized by large expanses of granitoid rocks separated by isolated remnants of supracrustal rocks. Both contain deformed Archean supracrustal rocks and late Archean (Kenoran) granites. The Churchill Province, however, also contains belts of deformed early Proterozoic supracrustal and late- to post-kinematic early Proterozoic (Hudsonian) granites. Distinction between these Archean and Proterozoic rocks and various orogenic events on the basis of field evidence has been difficult and therefore is critically dependent upon radiometric dating of appropriate units. Results of U-Pb zircon dating presented here, in conjunction with previously published results, provide essential data to interpret the tectonic evolution of part of this large, geologically complex region.

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GEOLOGICAL SUMMARY

The Whitehills-Tehek lakes area, about 50 km north of the village of Baker Lake, District of Keewatin (Fig. 1) is located within a region inferred to be composed mainly of deformed Archean rocks surrounded by regions more extensively affected by early Proterozoic deformation (Henderson et al., 1991). The extent of this Archean craton is currently constrained by several major geological features. The deformed and metamorphosed supracrustal rocks of the lower Proterozoic Amer Group indicate significant deformation to the northwest (Patterson, 1986), and to the south outliers of Hurwitz Group rocks (Fig. 1) containing early Proterozoic mafic sills were deformed after ≈2.16 Ga (Patterson and Heaman, 1991). Other limits are the 1.81 Ga Wager shear zone (Henderson and Roddick, 1990) on the northeast, the early Proterozoic Daly Bay complex (Gordon, 1988) on the Hudson Bay coast to the east, and to the west

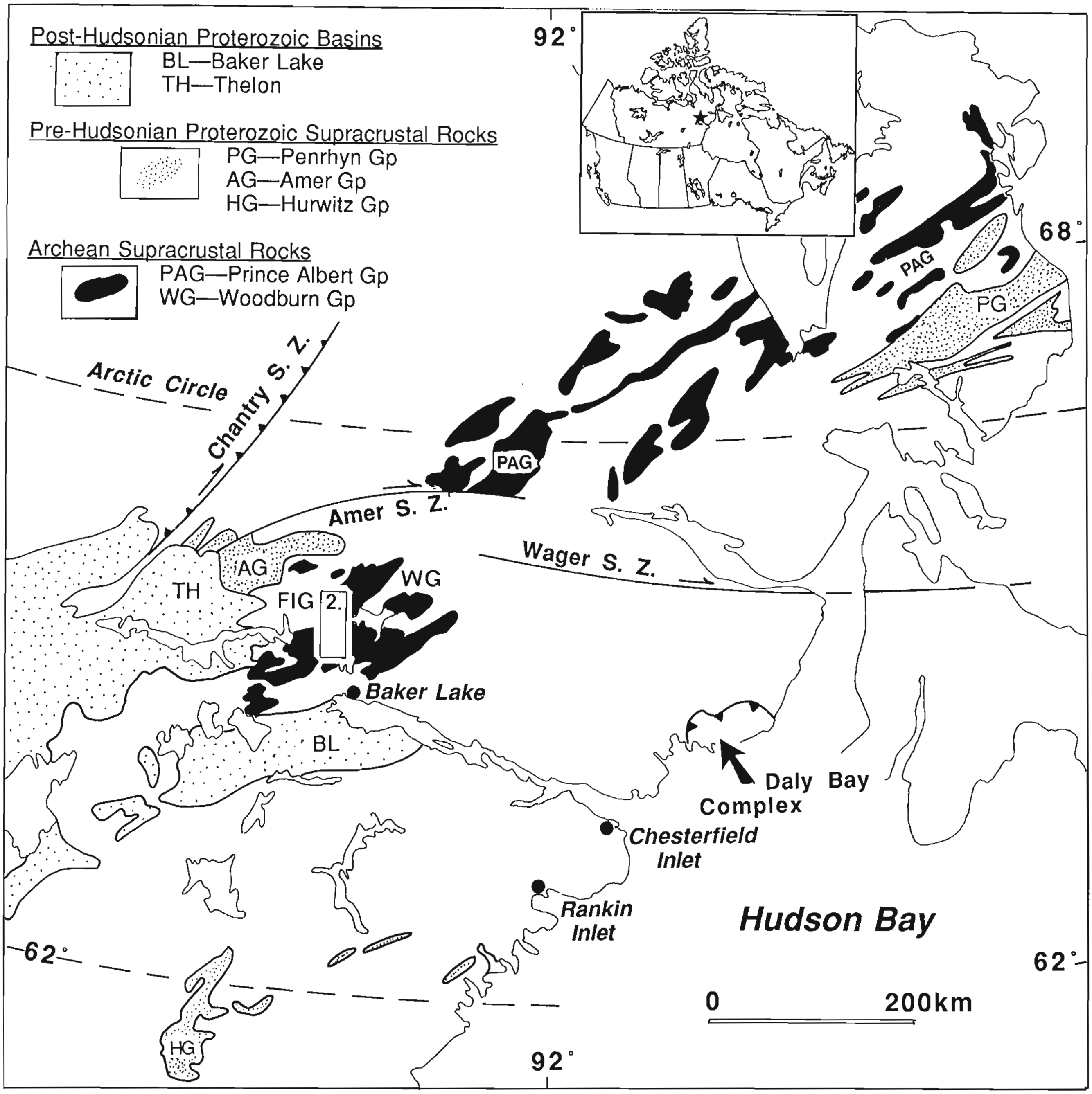

Figure 1. Geological map of the region northwest of Hudson Bay showing the distribution of Precambrian supracrustal rocks and major shear zones. The Whitehills-Tehek area is noted as Figure 2, about 50 km north-northwest of Baker Lake village.

the 1.7-1.8 Ga igneous and sedimentary rocks which fill the Thelon and Baker Lake basins (Fig. 1). These continental basins form an overlap sequence on Archean granitoid rocks, many of which record major felsic magmatism in the time interval 2.61 to 2.58 Ga (LeCheminant and Roddick, 1991).

Despite these suggested boundaries to this Archean terrane, interpretations remain controversial. Early Proterozoic deformation is likely to have occurred within parts of the region. For example, just west of the Amer shear zone near its intersection with the Thelon Basin sediments (Fig. 1), LeCheminant and Roddick (1991) obtained a U-Pb zircon age of 2610 +13/-12 Ma (originally reported as 2617 ± 20 Ma in Tella (1984)) for foliated granite unconformably overlain by deformed Amer Group basal quartzite. This is one basis for assigning the Amer Group an early Proterozoic age and concluding that deformation that affected it is Hudsonian in age, despite the fact that no geochronology is available on Amer Group rocks. In the Woodburn Lake map area north of Tehek Lake, Fraser (1988) showed that Woodburn Lake group rocks pass from greenschist to granulite grade, and obtained U-Pb discordia intercept ages for zircons from the granulite complex of 2662 +78/-53 and 1746 +86/-83 Ma. He interpreted the metamorphism and magmatism to be late Archean but choose to give no geological significance to the 1746 +86/-83 Ma lower intercept. This lower intercept may represent an early Proterozoic metamorphic event, although, on the basis of the age of an undeformed mafic syenite in the Amer Lake area, metamorphism must be >1.84 Ga (Tella et al., 1985b). Patterson (1986) inferred that extensive Proterozoic deformation occurred in the Whitehills-Tehek region. In a synthesis of the tectonics of Amer fold belt, she proposed that the Whitehills-Tehek belt (Fig. 1, WG) comprised the metamorphic and plutonic internides of the Amer belt. However, Ashton (1982, 1988) showed that Archean supracrustal rocks in the internide region northwest of Tehek Lake (Fig. 2) are at greenschist grade and are intruded by structurally isotropic to weakly foliated granites which give 2.6 Ga U-Pb zircon ages. With these differing interpretations of the extent of Proterozoic events in the region it is important to provide radiometric dates on rocks which more clearly constrain the geological evolution.

GEOLOGY OF THE WHITEHILLS-TEHEK LAKES AREA

In the region northwest of Hudson Bay (Fig. 1), the term Woodburn Lake group was first applied by Ashton (1982) to a succession of metavolcanic and metasedimentary rocks that occurs northwest of Tehek Lake (located 50 km NNW of Baker Lake). Subsequent mapping (Taylor, 1985; Fraser, 1988; Henderson et al.,1991) has extended the sequence over a larger area (Fig. 1) and linked it to other metasedimentary and metavolcanic sequences. Fraser (1988), mapping in the Woodburn Lake map area north of Tehek Lake, concluded that the Woodburn Lake group comprises an apparently conformable succession consisting of a 'lower sequence' of metavolcanic rocks and volcanogenic metasedimentary rocks and an 'upper sequence' of quartzite-dominated, platform type metasedimentary rocks. A similar stratigraphy was proposed by Ashton (1988) but with an unconformity between the lower and upper sequences. In the Whitehills-Tehek lakes area Henderson et al. (1991) could not establish an unconformity between the two sequences but showed that exposures include ultramafic, komatiitic (also described by Annesley, 1989), andesitic, and bimodal metavolcanic rocks, metagreywacke, iron-formation, and quartzite. On the basis of lithological similarities, Fraser (1988) correlated Woodburn Lake group with Prince Albert Group to the northeast. This correlation therefore implies that the Woodburn Lake group is Archean because U-Pb zircon dating of metarhyolite from Prince Albert Group in western Melville Peninsula by Frisch (1982) yielded a concordia upper-intercept age of 2879 Ma, and Schau (1982) concluded that the Prince Albert Group probably formed between 2.7 and 3.0 Ga.

Within the Whitehills-Tehek lakes area, bedding in Woodburn Lake group rocks is overprinted by two tectonic foliations (Fig. 2). S1 is the pervasive structural fabric throughout the area. It is locally overprinted by S2 crenulation cleavage which is axial-planar to a macroscale northwesterly-vergent recumbent F2 antiform (Fig. 3). Where S1 is not overprinted by S2, it dips uniformly southeast. Quartzite overlies most other lithological units in different places, suggesting that it was the strong member during a northwesterly-vergent, thrust-dominated D1 event. The youngest folds in the region are upright north-south trending, and east-west trending structures with no associated cleavage or mesoscopic folds. The north-south folds form a tight W-shaped synform between two lobes of structurally isotropic, late- to post-kinematic granite. Metamorphic grade is greenschist facies throughout most of the area but sillimanite has been noted at one locality in the most northern part of the area (S. Tella, pers. comm., 1991). The youngest granitic rocks identified in the region are undeformed granitic dykes up to 3 m wide and several tens of metres long.

ANALYTICAL PROCEDURES

Sample preparation, including air abrasion of zircon (Krogh, 1982) and U-Pb analytical methods are those outlined in Parrish et al. (1987) for zircon and monazite and in Parrish et al. (1992) for titanite mineral ages. U and Pb blanks were approximately 2 and 15 picograms, respectively. The analytical results for five samples are presented in Table 1. Linear regressions of data on concordia diagrams utilize a York (1969) error treatment with uncertainties as indicated in Table 1 for Pb/U ratios. Correlation coefficients, calculated using techniques detailed in Roddick (1987), range from 0.75 to 0.96 for the zircon fractions. Uncertainties in ages are stated at the 2σ level.

GEOCHRONOLOGY

In the Whitehills-Tehek lakes area, the structural base of the Woodburn Lake group consists of meta-andesite, dacite and interbedded iron formation (Henderson et al., 1991). To establish the maximum age of the exposed Woodburn Lake

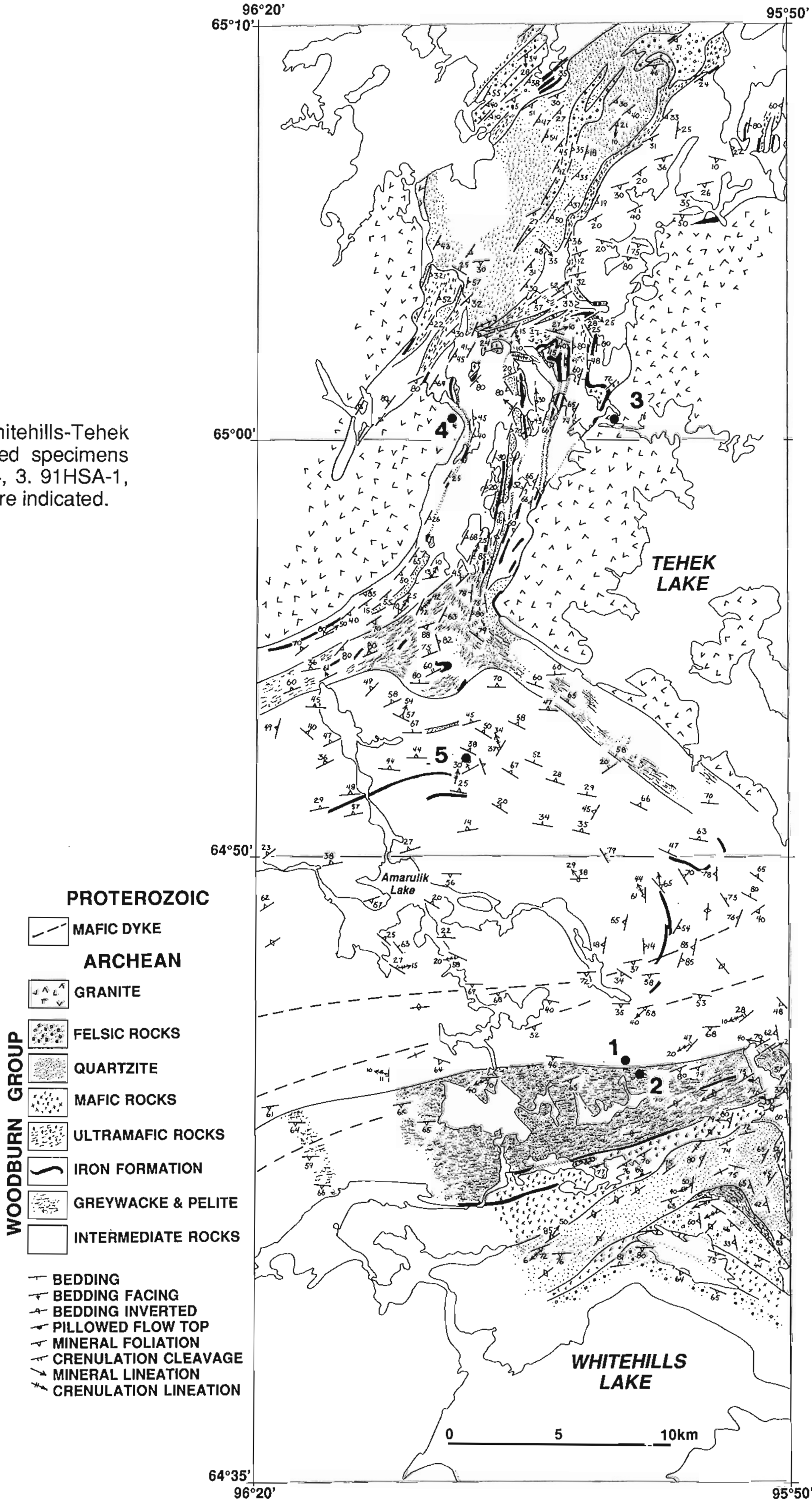

Figure 2.

Geological map of the Whitehills-Tehek area. Locations of five dated specimens (1. 89HSA-75, 2. 89HSA-74, 3. 91HSA-1, 4. 91HSA-2, 5. 89HSA-34) are indicated.

group a fine grained massive metadacite quartz porphyry (89HSA-75) was collected from near the contact with the overlying metagreywacke, about 10 km north of Whitehills Lake (Fig. 2). It is composed of plagioclase (An_{30}), quartz, chlorite, epidote, muscovite and carbonate. It is the major rock unit of the Woodburn Lake group and the protolith of the dated rock is believed to be a crystal tuff. Zircons in the sample are a uniform population of mainly translucent, pink to tan crystals 50 to 150 μm in size. Grains are subhedral to euhedral and mostly equidimensional but about 20% of the population consists of crystals with length to width (L/W) ratios of about 2:1. Crystal facets are sharp and well developed. In smaller clear grains not obscured by translucence no cores or zoning are visible. The sample also contains large (>150 μm) golden brown titanite fragments, some with facets and well developed igneous terminations.

Four multiple grain zircon fractions form a discordant array with some scatter about a regression line (MSWD=9.9) with an upper intercept age of 2787 ± 9 Ma (Fig 4). The lower intercept age is about 900 Ma. Omission of the most divergent analysis with the large error ellipse (A) does not significantly improve the fit (MSWD=9.4) nor change the intercept ages. The scatter suggests that there could be some inherited zircon in the analyzed grains. On this basis the minimum age is given by the ^{207}Pb-^{206}Pb age of fraction A at 2759 ± 7 Ma (Table 1). The igneous age of the metadacite is therefore constrained between the limits of 2796 and 2752 Ma. A discordant titanite analysis yields a younger ^{207}Pb-^{206}Pb age of 2739 ± 3 Ma. The titanite U-Pb system has a closure temperature of about 600°C (Tucker et al., 1987; Heaman and Parrish, 1991), well above temperatures implied by the regional greenschist metamorphism recorded in the Woodburn Lake group exposed in the area. It is possible, however, that partial resetting of the U-Pb system occurred during the metamorphism. The titanite then records an age between the zircon age and the metamorphic age, shown below to have occurred prior to 2616 Ma. On this basis the metamorphism and deformation of the Woodburn Lake group took place after 2736 Ma, the minimum ^{207}Pb-^{206}Pb age of the titanite.

A sample of psammitic medium grained tuffaceous metagreywacke (89HSA-74) was collected to examine the age of possible source rocks for the Woodburn Lake group. It was collected from the metasedimentary unit immediately overlying the metadacite (Fig. 2). The metagreywacke has a weak foliation, no visible bedding and is composed of

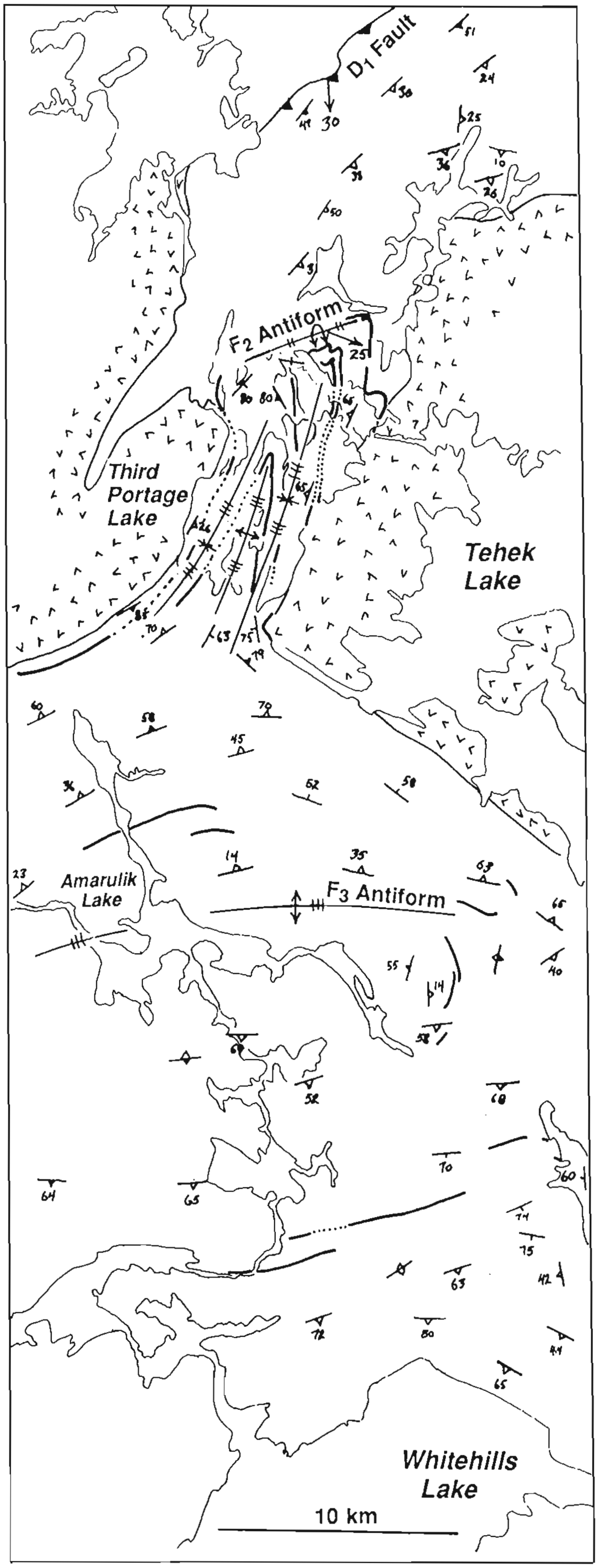

Figure 3. Structural map of the Whitehills-Tehek area showing F2 and F3 fold axial-surface traces, and an inferred D1 low-angle thrust fault. Dotted lines in Third Portage Lake link observed magnetic iron-formation beds that appear to be coincident with aeromagnetic anomalies. Note that both D1 and D2 are northwesterly-vergent, and D3 folds are upright with north-south and east-west trends.

Table 1. U-Pb analytical data

Fraction size µm[a]	wt.[2] (µg[b])	U (ppm)	Pb* (ppm)	$^{206}Pb/^{204}Pb$	Pb[c] (pg)	$^{208}Pb^*$ %	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	Apparent Age (Ma) $^{206}Pb/^{238}U$	Apparent Age (Ma) $^{207}Pb/^{206}Pb$
89HSA-75 Metadacite; Z1869; 64°45.1'N, 95°59.1'W											
Ti +105 †	54	54	57	566	154	50.0	0.5161±.13%	13.493±.17%	0.18961±.10%	2682.8	2738.8±3.3
A +105	6	92	58	488	38	15.4	0.5217±.36%	13.804±.35%	0.19190±.20%	2706.4	2758.5±6.5
B -105+74	6	139	78	1545	17	12.9	0.4770±.14%	12.403±.15%	0.18856±.06%	2514.4	2729.6±2.1
C -74+62	7	87	54	1402	15	12.8	0.5214±.25%	13.952±.24%	0.19406±.09%	2705.2	2776.9±2.8
D -125+74	4	110	71	1570	8	14.6	0.5330±.17%	14.284±.18%	0.19437±.05%	2754.0	2779.5±1.7
89HSA-74 Metagreywacke; Z1868; 64°44.8'N, 95°58.4'W											
A +105	7	152	92	1682	20	13.6	0.5085±.11%	13.222±.12%	0.18860±.05%	2650.0	2730.0±1.7
B +105	5	259	107	1495	20	10.3	0.3724±.11%	8.0620±.12%	0.15699±.05%	2040.9	2423.5±1.8
C ≈74	7	147	83	982	30	15.4	0.4688±.16%	11.761±.17%	0.18196±.07%	2478.2	2670.8±2.3
D ≈47	1	72	42	327	7	11.9	0.5038±.47%	12.793±.48%	0.18415±.24%	2630.3	2690.6±7.9
E -74	5	366	198	1482	34	14.2	0.4583±.13%	11.290±.15%	0.17868±.05%	2431.8	2640.7±1.7
91HSA-1 Granite - east; Z2593; 65°00.5'N, 95°59.8'W											
B ≈40	5	177	102	750	33	13.5	0.4915±.15%	11.876±.18%	0.17525±.09%	2577.0	2608.5±2.9
C -105+74	6	148	82	1850	14	12.4	0.4785±.14%	11.602±.15%	0.17585±.05%	2520.7	2614.1±1.7
D -10+74	10	128	74	976	37	13.4	0.4934±.13%	12.046±.15%	0.17707±.07%	2585.2	2625.7±2.3
91HSA-2 Granite - west; Z2594; 65°00.5'N, 96°08.8'W											
B -105+74	3	162	94	968	14	21.3	0.4518±.23%	10.803±.23%	0.17343±.07%	2403.1	2591.1±2.4
C -105+74	1	77	55	112	23	28.7	0.5033±1.3%	12.191±1.4%	0.17570±.57%	2627.8	2612.7±19.
D -105+74	1	51	34	131	13	24.8	0.4869±2.1%	11.727±2.1%	0.17468±.49%	2557.3	2603.0±17.
89HSA-34 Granitic Dyke; Z1866; 64°52.0'N, 96°08.0'W											
A Mz +250	12	5007	4887	27822	45	67.6	0.3292±.09%	5.091±.10%	0.11217±.03%	1834.3	1834.9±1.0
B Mz +250	8	2667	5230	11157	38	84.1	0.3246±.09%	5.016±.10%	0.11207±.03%	1812.2	1833.2±1.1

[a] Ti = titanite; Mz = monazite; † = not abraded
[b] Weight measured to ± 1 µg; this uncertainty to be included in U and Pb concentrations.
Pb* - radiogenic Pb
[c] - total common Pb in analysis
Errors are 1 std. error of mean in % except 207/206 age errors which are 2 std. errors in Ma.

plagioclase (An_{30}), quartz, epidote, titanite, muscovite and chlorite. Zircons from this sample do not show typical rounding characteristic of detrital zircons but rather are similar to the zircons from the underlying metadacite both in colour and morphology. They have sharp crystal facets with no rounding present. The only significant difference from the metadacite population is in the L/W of the grains. In this sample the grains range from equant to L/W of 3:1, with significantly more (60%) in the range 2:1 to 3:1. This similar morphology suggests that the metadacite could be the source for the zircons and perhaps minor transport selectively concentrated the more elongate grains.

Although individually discordant, four of five multi-grain zircon fractions from sample 89HSA-74 define a discordia line (Fig. 5) with some scatter from expected experimental errors (MSWD=11). The upper intercept age for this line is 2769 ± 11 Ma and the lower intercept is about 1350 Ma. The fifth analysis (D) could be younger since it has a ^{207}Pb-^{206}Pb age of 2691 ± 8 Ma, although if Pb loss took place at a similar time to the co-existing zircons (1350 Ma) the age could be as old as 2750 Ma. Most of the zircons appear to be from a single population with an age of about 2770 Ma or, considering the slight scatter in the data, more than one population with very similar ages. Within the error limits, the age is similar to that of the metadacite and it is possible that these detrital zircons were derived primarily from that unit or related igneous rocks and would therefore appear to be locally derived. Since zircon ages from this unit are the same or younger than the underlying metadacite, there is no evidence on the possible age of the basement to the Woodburn Lake group.

Samples for U-Pb dating were collected from the two granite bodies which intrude the W-shaped synform just west of Tehek Lake (Fig. 2). Both granites are coarse grained, granular textured in outcrop, and apparently unfoliated. Thin-section examination of sample 91HSA-1 (eastern body)

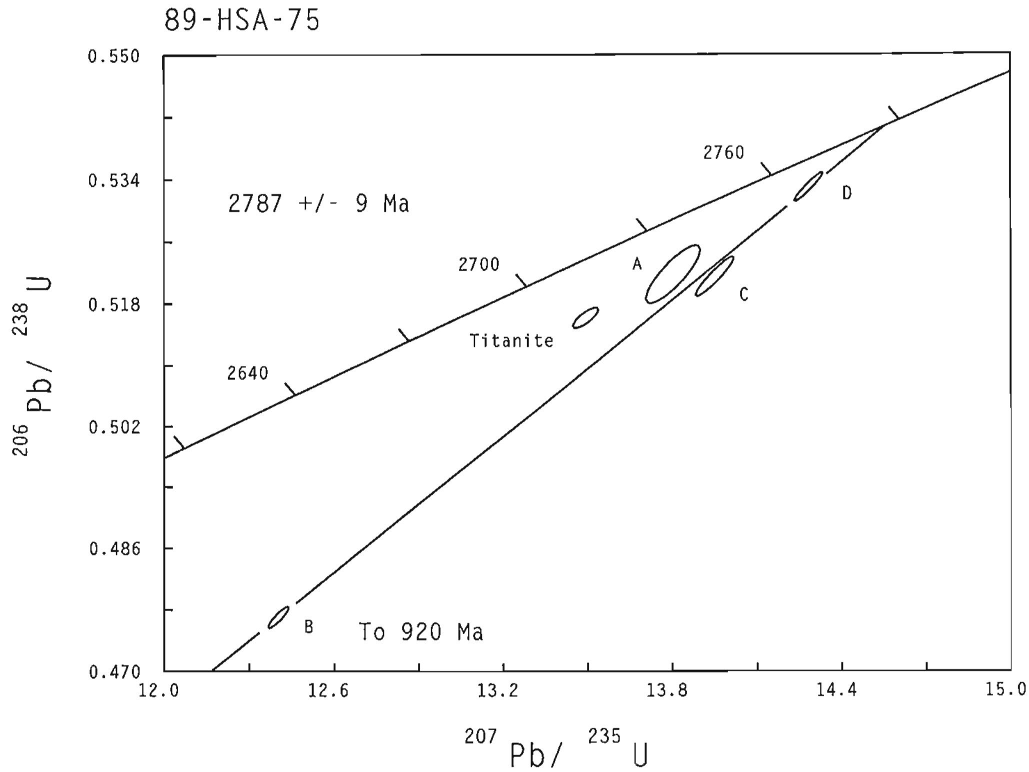

Figure 4.

U-Pb concordia diagram for zircons and a titanite from the metadacite quartz porphyry (89HSA-75) in the Woodburn Lake group.

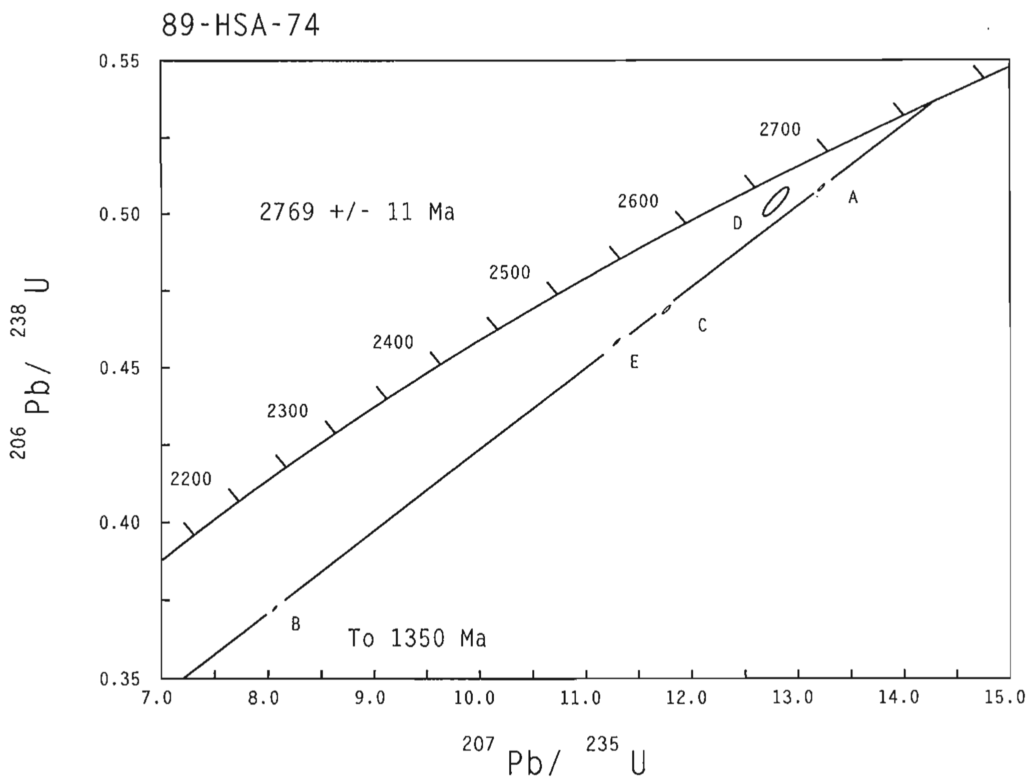

Figure 5.

U-Pb concordia diagram for zircons from a metagreywacke (89HSA-74) in the Woodburn Lake group.

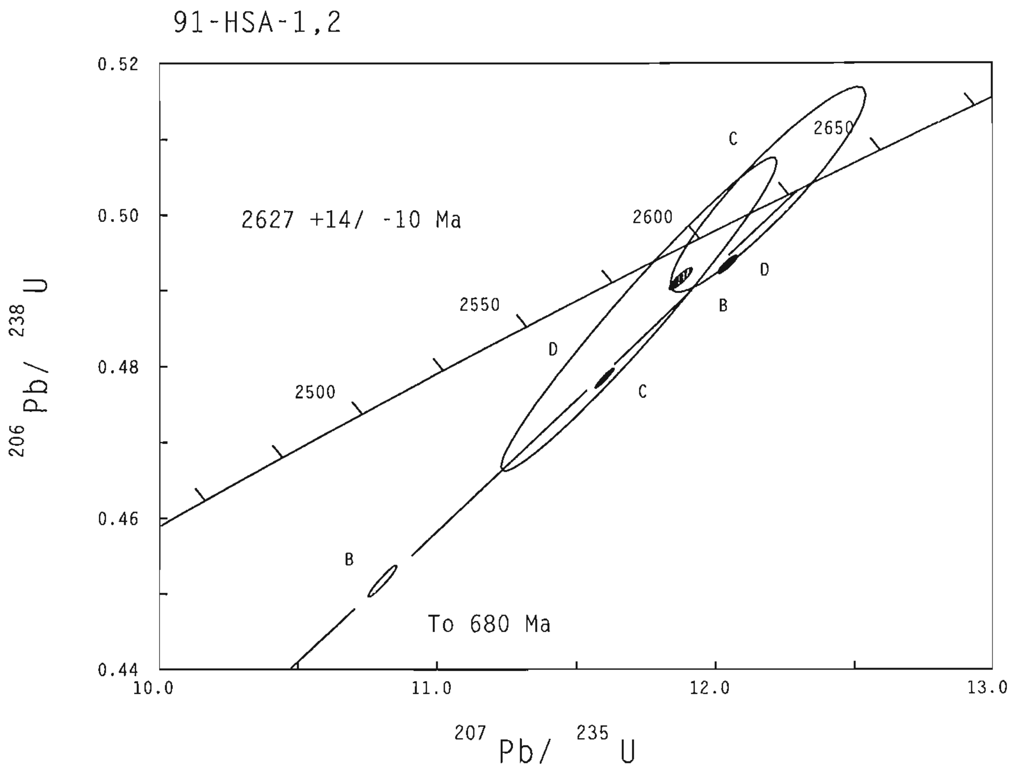

Figure 6.

U-Pb concordia diagram for two granite bodies intruding the W-shaped synform just west of Tehek Lake; filled ellipses for sample 91HSA-1 (east body) and open ellipses for sample 91HSA-2 (west body).

however, reveals mortar-textured, coarse feldspar grains and a moderate foliation defined by fine grained recrystallized quartz. Sample 91HSA-2 (western body) shows weak optical strain in quartz, and some quartz sub-grain development. Both granites contain perthitic microcline, oligoclase, quartz, and minor chloritized biotite. Sample 91HSA-1 contains minor titanite and plagioclase is strongly epidotized. These granites, which could be lobes of a common subsurface pluton, were thus intruded late in the deformation history and provide a minimum age for the last major deformation in the area.

The zircon populations of both samples are dominated by grains that are <125 µm in size and most in the range 50-100 µ m. Zircons are more abundant in sample 91HSA-1 with clear to translucent, colourless to pale pink-tan crystals. Grains are subhedral to euhedral with L/W ratios of 1:1 to 3:1 and are inclusion free. Crystal facets are sharp and well developed and no cores are seen. Zircons in sample 91HSA-2 are similar but are generally more translucent with the >100 µm grains being opaque white. The opacity appears to be caused by fracturing or fine pitting on the grain surfaces. The smaller grains are clear to pale pink-yellow in colour and have igneous growth zoning.

Figure 6 shows U-Pb results for granite samples 91HSA-1 and -2 plotted on a concordia diagram. Both zircon populations have low U contents (50-180 ppm) and the large error ellipses for sample 91HSA-2 analyses C and D result from the small two grain (1 µg) samples analyzed (Table 1).

The three analyses for sample 91HSA-1 show significant scatter beyond a linear trend, indicating inherited zircon is present. The best age estimate is from the youngest fraction (B) with a ^{207}Pb-^{206}Pb age of 2609 ± 3 Ma. This analysis overlaps those of sample 91HSA-2 and suggests that the two intrusions could be the same age. A regression of all sample 91HSA-2 fractions and fraction B of sample 91HSA-1 (MSWD<1) results in an age of 2612 ± 4 Ma and a lower intercept of about 450 Ma. This age for the granites, taken in conjunction with the titanite ^{207}Pb-^{206}Pb age of 2736 Ma from the metadacite, establishes that the major deformation in the region took place between about 2736 Ma and 2608 Ma (minimum granite age).

The youngest granitic rocks in the region are several undeformed dykes crosscutting the Woodburn Lake group. One sample (89HSA-34) was collected from an easterly trending 3m wide dyke located 10 km southwest of Tehek Lake (Fig. 2). It is coarse grained and composed of K-feldspar, plagioclase, quartz and muscovite. No zircon is present in the sample but minor amounts of clear yellow-green subhedral monazite were separated. Two analyses of single grains of monazite (Table 1), one of which is concordant, have similar ^{207}Pb-^{206}Pb ages at 1835 ± 1 Ma interpreted as the igneous age. This age is similar to the 1.82-1.83 Ga ages of abundant late- to post-kinematic Hudsonian calc-alkaline granitic plutons (LeCheminant et al., 1987) intruding deformed early Proterozoic Penrhyn Group north of the Wager shear zone, and is distinctly older than ca. 1.74-1.76 Ga rhyolites and rapikivi granites emplaced during late stages in the formation of the Thelon and Baker Lake basins (Dudás et al., 1991; LeCheminant 1992; Fig. 1).

DISCUSSION

The Whitehills-Tehek belt is a northwesterly-vergent late Archean fold-thrust belt in central Churchill Province. U-Pb dating of critical units in the area further constrains the geological evolution of the region. The oldest exposed unit of the Woodburn Lake group metavolcanics and metasediments has an age between 2796 and 2752 Ma. A minimum ^{207}Pb-^{206}Pb age of 2736 for titanite from this basal unit is interpreted to represent partial resetting of the U-Pb system in the mineral by regional metamorphism, indicating that the group was metamorphosed and deformed after this time. Overlying metagreywacke contains zircons with morphology and U-Pb ages similar to the dated basal metadacite unit. Last stages of deformation are represented by two exposures of a slightly deformed granite with a combined age of 2612 ± 4 Ma. Undeformed granitic dykes record minor Proterozoic magmatism at 1835 ± 1 Ma.

The age of the Woodburn metadacite is similar to a U-Pb zircon age of 2798 +24/-21 Ma, determined by Tella et al. (1985a), for a dacite porphyry from southeast of the Amer fold belt. They considered the dacite porphyry to be of tuffaceous origin and to be part of a metamorphosed mafic volcanic sequence which structurally underlies quartzite and phyllitic schist. The similar field relationships and age of metavolcanics to those in the Whitehills-Tehek lakes area suggest that Woodburn Lake group rocks can be further extended to the northwest.

Ashton (1988) obtained U-Pb zircon ages of 2621 ± 2 Ma for 'coarse grained granite' and 2599 ± 5 Ma for 'foliated granite' intruding Woodburn Lake group in the northeast corner of the Whitehills-Tehek area (Fig. 2). These ages, and the age reported here for the two granite bodies, are similar to ages of granitoids located west and south of Whitehills-Tehek area (LeCheminant and Roddick, 1991), and confirm the existence of widespread late Archean granitic plutonism within the central Churchill Province. The absence of penetrative early Proterozoic (Hudsonian) deformation in the Whitehills-Tehek area raises the possibility that the adjacent Amer belt (Patterson, 1986) to the northwest could be a late Archean rather than a Hudsonian fold-thrust belt.

Finally, the occurrence of Archean iron-formation-hosted gold mineralization in the Woodburn Lake group (Henderson et al., 1991) indicates that similar styles of mineralization should be sought along strike to the northeast in possibly correlative Prince Albert Group rocks.

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- 3 -

Age of rhyolite and provenance of detrital zircons in a granulestone from George Lake area, Slave Province, Northwest Territories

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van Breemen, O., Bursey, T., and Jefferson, C.W., 1992: Age of rhyolite and provenance of detrital zircons in a granulestone from George Lake area, Slave Province, Northwest Territories; in Radiogenic Age and Isotopic Studies: Report 6, Geological Survey of Canada, Paper 92-2, p. 41-47.

Abstract

The age of deposition of an iron-formation sequence in the George Lake area is established at 2683 ±2 Ma, based on U-Pb zircon ages from a discordant rhyolite sill and detrital zircon from a granulestone interbedded with turbidites. These data support correlation with the turbidite-hosted iron-formations overlying the main volcanic sequences in the Back River volcanic complex and the Hackett River volcanic belt of the eastern Slave Province.

Detrital zircons of various morphology in the granulestone do not provide evidence for a significantly older source terrane. Five concordant analyses of stubby crystals, interpreted to derive from nearby volcanic rocks, have ages in the range 2683 Ma to 2697 Ma, whereas one concordant analysis was slightly older at 2722 Ma, close to the oldest volcanic ages in the Slave Province.

Résumé

L'âge d'une séquence ferrifère dans la région du lac George a été établi à 2 683 ± 2 Ma, d'après des datations par la méthode U-Pb sur des zircons d'un filon-couche discordant de rhyolite et sur des zircons détritiques provenant d'une roche à granules (granulestone) interstratifiée avec des turbidites. Ces données appuient une corrélation faite avec les formations ferrifères logées dans des turbidites qui sont sus-jacentes aux principales séquences volcaniques du Complexe de Back River et de la Zone de Hackett River, dans l'est de la province des Esclaves.

Les zircons détritiques à morphologie diverse de la roche à granules n'indiquent pas que le terrane originel soit beaucoup plus ancien. Cinq analyses concordantes de cristaux trapus, interprétés comme provenant de roches volcaniques voisines, ont donné des âges variant entre 2 683 Ma et 2 697 Ma, tandis qu'une autre analyse concordante a livré un âge légèrement plus ancien de 2 722 Ma, valeur proche de celles des roches volcaniques les plus vieilles de la province des Esclaves.

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INTRODUCTION

The George Lake area, in the northeastern part of the Archean Slave Structural Province (Fig. 1), is underlain by a sequence of turbidite-hosted iron-formations which trends northwest for tens of kilometres in a linear belt (Olson, 1989,1990). Part of this belt is overlain by the Proterozoic Goulburn Group (Chandler and Holmberg, 1990).

Turbiditic rocks in the George Lake area contain two stratigraphically distinct auriferous iron-formations, and are cut by lenticular bodies of quartz-feldspar phyric rhyolite-dacite (Jefferson et al., 1992a). The purpose of this study was to establish the minimum age of the turbidites by dating zircons from the intrusions, and the maximum age by dating single detrital zircons from a granulestone interbedded with the turbidites. Ages of the detrital zircons also yield information on the sedimentary provenance and tectonic setting of deposition, although provenance ages generally introduce a bias towards felsic compositions (Davis et al., 1990). This note provides final geochronologic results which were reported in preliminary form by Bursey (1991) and van Breemen et al. (1992a).

GEOLOGY

Two iron-formations in the George Lake area, Northwest Territories, are being explored for gold by the George Lake Joint Venture (Olson, 1989, 1990; Chandler and Holmberg, 1990). These iron-formations are hosted by turbidites of the Beechey Lake Group of the Yellowknife Supergroup (Frith, 1987) which are intruded by rhyolite sills and contain minor tuffs. George Lake lies north of the Back River volcanic complex and east of Hackett River volcanic belt, which are part of an approximately 300 km volcanic belt trending north-northwest (Mortensen et al.,1988).

Jefferson et al. (1989, 1991, 1992b) noted that the iron-formations of George Lake are similar, both in lithology and general stratigraphic setting, to those of iron-formation C, as recognized in the Beechey Lake Group in the Back River and Hackett River areas (Table 1), and those in the Hood River belt, as summarized by Jefferson et al. (1992c); both are turbidite-hosted, characterized by magnetite-chert facies, and contain minor tuff and lapilli tuff beds. Jefferson et al. (1989, 1992b) and Lambert et al. (1990) have proposed a three-fold stratigraphic subdivision of the Back River volcanic

Table 1. Lithostratigraphic correlation and available radiometric dates after Bursey (1991), Jefferson et al. (1992a), van Breemen et al. (1992a). Locations are given in Figure 1.

Back River Volcanic Complex[1]	Hackett River Volcanic Belt[2]	George Lake[3]
turbidites of Beechey Lake Group (BLG)	turbidites of BLG	turbidites of BLG
Iron Formation C in BLG turbidites with felsic volcanic and epiclastic rocks (undated)	Iron Formation C in BLG turbidites with felsic volcanic and epiclastic rocks (undated)	Iron Formation in BLG turbidites with felsic epiclastic rocks and sill max. 2683 ± 2 Ma (detrital zircons) min. 2683 ± 2 Ma (zircons from sill)
Iron Formation B with carbonaceous and calcareous epiclastic rocks	Iron Formation B with carbonaceous and calcareous epiclastic rocks	base not exposed
2692 ± 2 Ma rhyolite dome in upper part of contiguous mafic to felsic volcanic sequence	2689 ± 2 Ma on Musk; 2687-2696 Ma on Yava; synvolcanic rhyolite domes in upper part of contiguous mafic to felsic volcanic sequence	
Iron Formation A with carbonaceous and calcareous epiclastic rocks above contiguous mafic to felsic volcanic sequence (undated)	base not exposed	
base not exposed		

[1] stratigraphy after Jefferson et al., 1992b; Lambert et al., 1990; U-Pb date from van Breemen et al. (1987).
[2] stratigraphy after Frith (1987); Jefferson et al.,1976; U-Pb dates from Mortensen et al. (1988)
[3] stratigraphy after Jefferson et al. (1992a); U-Pb dates from this report.

complex, based on mapping and facies associations of three different iron-rich sedimentary sequences. The lower two (A and B) mark hiatuses in volcanism and contain only minor tuffaceous material. The upper, turbidite-hosted iron-formation (C) appears to be a distal exhalative expression of volcanism which took place during turbiditic sedimentation. Rare tuffs and breccias intercalated with the iron-formation indicate minor volcanism coeval with the exhalative activity that formed the iron formation. This inference is tested at George Lake.

The detrital zircons dated in this study are from a thick bedded granulestone and quartz wacke turbidite unit (Bursey, 1991). This granulestone is most extensive in a lower tongue which overlies and is interbedded with a lower iron-

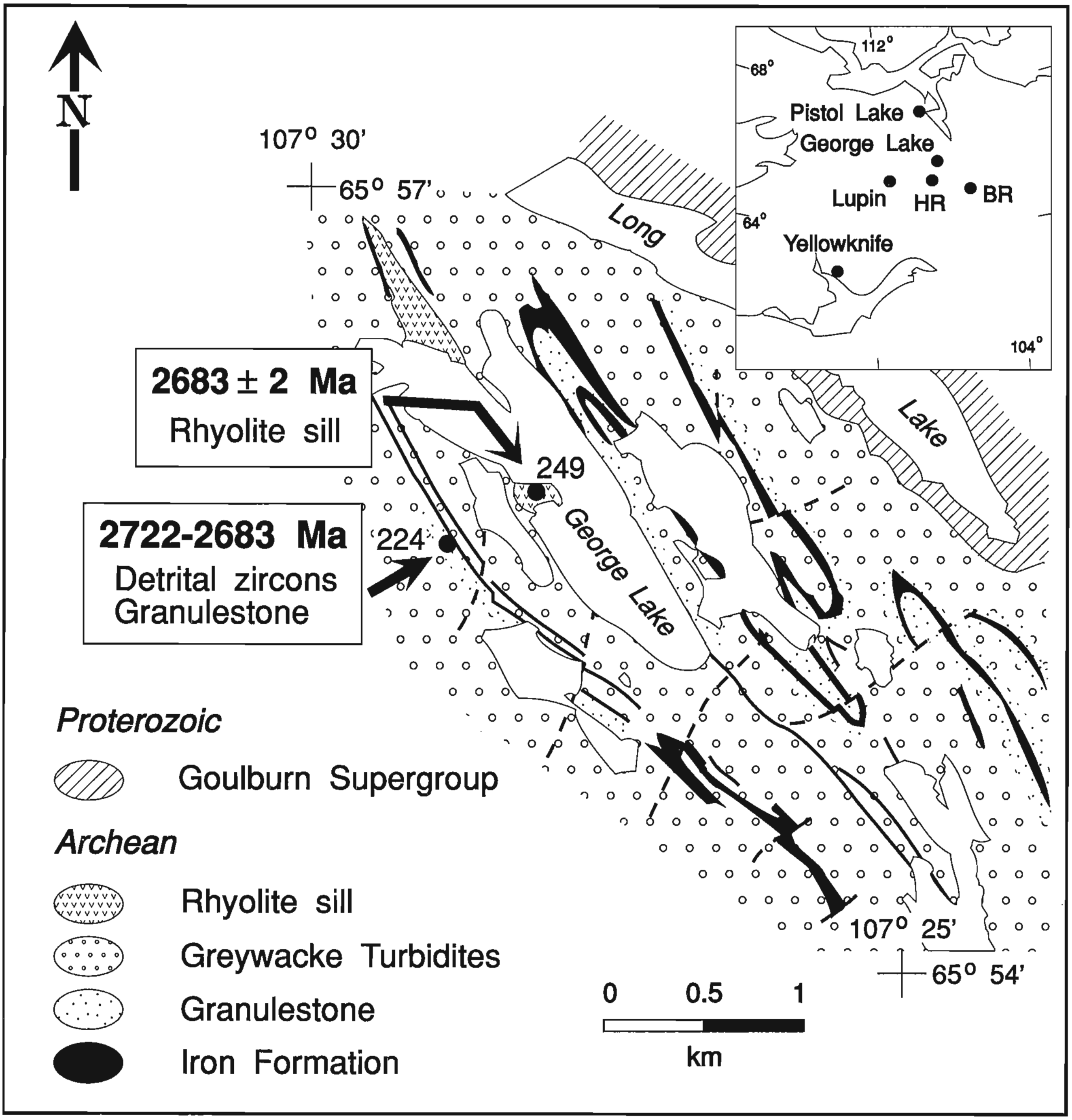

Figure 1. Geology of George Lake area showing localities of samples analyzed: #249, rhyolite; #224, granulestone. Inset: HR – Hackett River, BR – Back River. After Jefferson et al. (1992a).

formation. An upper tongue, restricted to the eastern area, is also intercalated with and overlies the upper iron-formation. In the western part of George Lake area, the unit is finer grained, but still distinctive from the adjacent turbidites. Eastward thickening of the granulestone suggests a proximal source to the east. About 25% of the source was volcanic as suggested by coarse grit-sized fragments of quartz and feldspar, pseudohexagonal quartz, volcanic rock fragments and resorption embayments in quartz grains (Bursey, 1991). The remainder of the source area appears to have been metamorphic, as is inferred from the high percentage of both strained and unstrained polycrystalline quartz with sutured boundaries, and granoblastic polygonal quartz. The grade of this metamorphism is, however, unconstrained. Archean sedimentary and volcanic rocks in the George Lake area experienced greenschist to lower amphibolite facies metamorphism (Chandler and Holmberg, 1990).

Detrital zircon sample 224 is from a lenticular bed of granulestone, part of the lower tongue of the quartz wacke to quartzose granulestone unit described above (Fig. 1). The unit is a moderately well sorted quartz greywacke with an average of 80% framework consisting of 30-40% strained polycrystalline quartz. A small percentage of feldspar, sub-rounding of grains and moderate degree of sorting suggest that the framework grains were transported over a moderate distance. A matrix content of about 20% may reflect mass deposition (e.g. turbiditic flow).

Sample 249 was collected from a porphyritic rhyolite sill at George Lake which is locally discordant to bedding (Fig. 1). The lack of any pyroclastic detritus, alignment of feldspars or flow-banding supports the field interpretation that this massive unit is neither tuff nor flow. The sill cuts greywacke and mudstone that stratigraphically overly the upper iron-formation, mudstone and the lower tongue of the granulestone, successively.

GEOCHRONOLOGY

U-Pb analyses are all of single abraded grains of zircon, following techniques detailed in Parrish et al. (1987). Treatment of analytical errors is described in Roddick (1987). Isotopic data are presented in Table 2 and displayed in Figure 2.

Rhyolite sill

Zircons recovered from sample 249 are clear and elongate to equidimensional (Fig. 3). Many edges are rounded and are interpreted in terms of igneous resorption (Sullivan and van Staal, 1990). Fractures are common. U concentrations are low (25-46 ppm; Table 2); data points are concordant with overlapping uncertainties, providing a precise crystallization age of 2683 ± 2 Ma (Fig. 2).

Granulestone

Two distinct zircon populations were apparent in the concentrate from sample 224: the majority consist of stubby crystals, a type commonly found in volcanics of the Slave Province (Fig. 3); others are more elongate prisms (length to width ratios of 2:1). Some grains manifest slight rounding; others have sharp corners. No cores were evident. In the

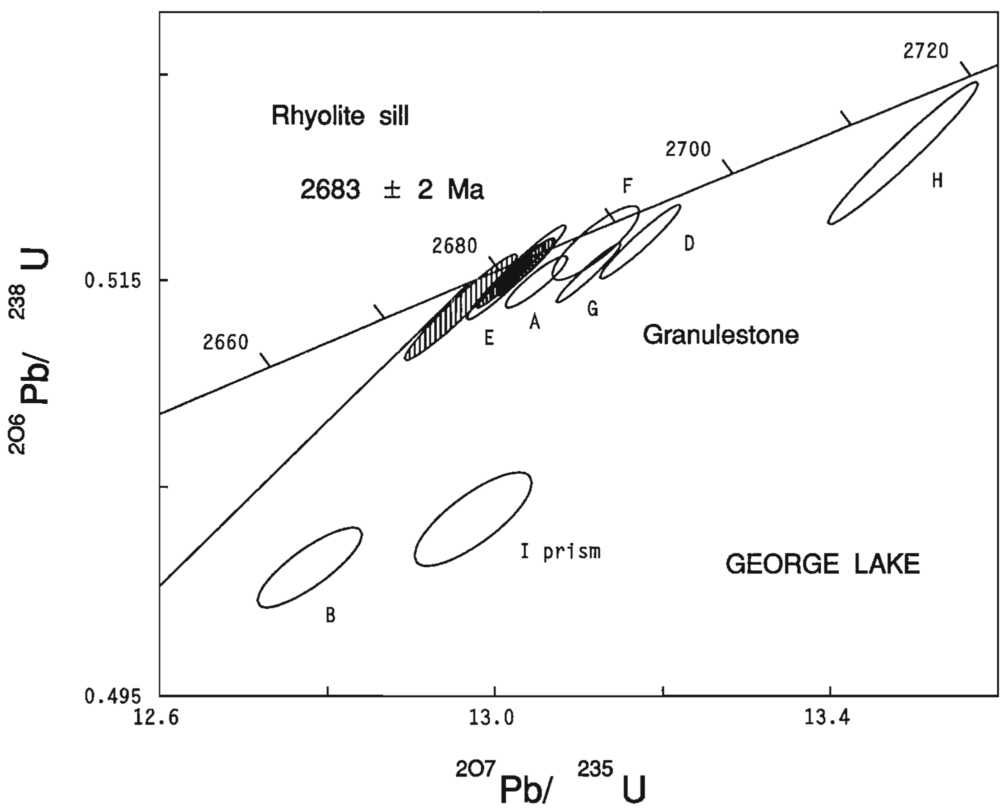

Figure 2.

Concordia diagram showing U-Pb data from rhyolite sill (hatched error envelopes) and granulestone (open error envelopes).

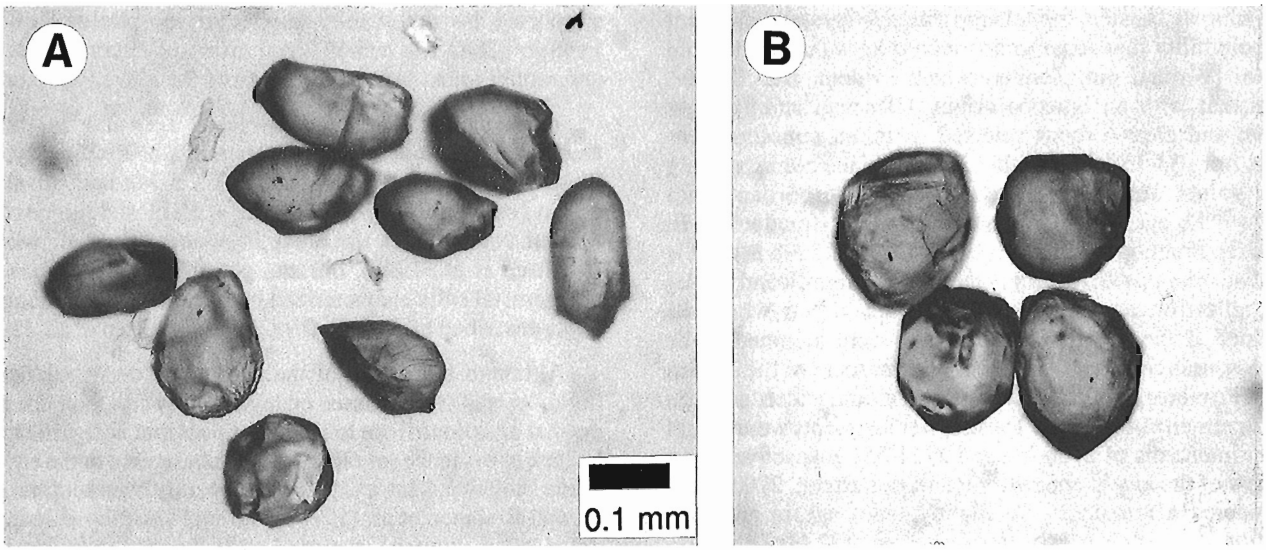

Figure 3. Photomicrographs of: **A**, zircons from rhyolite; and **B**, stubby zircons analyzed from granulestone.

Table 2. U-Pb zircon data

Fraction size µm[1]	wt. (µg[b])	U (ppm)	Pb[2] (ppm)	^{206}Pb[3] / ^{204}Pb	Pb[4] (pg)	^{208}Pb[2] %	^{206}Pb[5] / ^{238}U	^{207}Pb[5] / ^{235}U	^{207}Pb[5] / ^{206}Pb	Apparent Age (Ma)[6] ^{206}Pb/^{238}Pb	^{207}Pb/^{206}Pb
1. Rhyolite (89-JP-249)											
A N1 ≈200*120	0.007	46	26	988	10	7.3	0.5152(.15)	13.019(.16)	0.97	0.18326(.04)	2682.6(1.4)
B N1 ≈200*120	0.009	63	37	906	20	10.6	0.5156(.14)	13.033(.15)	0.94	0.18335(.05)	2683.4(1.7)
C N1 ≈200*120	0.009	25	14	545	13	6.9	0.5137(.25)	12.959(.26)	0.97	0.18335(.05)	2680.1(2.2)
2. Granulestone - George Lake (89-JP-224)											
A M2 stubby≈200	0.015	50	31	834	27	14.9	0.5149(.12)	13.049(.14)	0.91	0.18380(.06)	2687.5(1.9)
B M2 stubby≈180	0.009	59	37	406	38	17.2	0.5011(.19)	12.779(.24)	0.83	0.18494(.14)	2697.7(4.5)
D M2 stubby≈160	0.007	56	36	787	17	17.2	0.5169(.17)	13.173(.18)	0.96	0.18484(.05)	2696.8(1.6)
E M2 stubby≈160	0.007	48	30	642	17	15.8	0.5154(.22)	13.025(.23)	0.97	0.18330(.06)	2682.9(1.8)
F M2 stubby≈160	0.004	137	87	1091	15	16.4	0.5168(.17)	13.120(.20)	0.85	0.18411(.10)	2690.3(3.4)
G M2 stubby≈160	0.006	99	63	3969	5	17.3	0.5154(.14)	13.112(.15)	0.97	0.18452(.04)	2694.0(1.2)
H M2 stubby≈160	0.005	34	21	416	13	15.5	0.5212(.33)	13.487(.33)	0.98	0.18768(.08)	2722.0(2.4)
I M2 ≈220*80	0.003	145	82	314	46	8.8	0.5035(.22)	12.974(.27)	0.78	0.18690(.17)	2715.0(5.6)
K M2 ≈220*80	0.003	156	93	400	36	18.4	0.4801(.21)	12.092(.23)	0.83	0.18268(.13)	2677.4(4.3)
L M2 ≈220*80	0.005	130	79	91	271	12.0	0.5180(.27)	13.516(1.0)	0.65	0.18923(.85)	2735.5(28)

Notes: All analyses are on single zircons [1] sizes in microns 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; [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] ^{207}Pb/^{206}Pb model age; errors are two sigma in Ma.

following discussion, model ages intimate crystalization, but the possibility that some grains record ages that are too old cannot be ruled out. Zoning, where evident, was regular, consistent with an igneous origin. Of seven stubby, light brown and clear zircons selected, uranium concentrations range from 34-137 ppm (Table 2). Five of the corresponding data points are concordant to slightly discordant with ^{207}Pb-^{206}Pb ages from 2683 Ma to 2697 Ma (fractions A, E, D, F, G). Fraction B has a ^{207}Pb-^{206}Pb age of 2698 Ma but is 4% discordant (with respect to the origin). Significantly older crystallization ages for this discordant zircon would be obtained if the average, Pb loss age were assumed to be 500 Ma, as is commonly found in igneous rocks of the eastern Slave Province, or 1000 Ma which is seldom exceeded (e.g., van Breemen et al., 1987). These lower intercepts would yield upper intercepts of 2706 Ma and 2721 Ma, respectively, but in view of the low U concentration of this zircon, 2706 Ma is considered a maximum. A slightly older age is given by fraction H which is nearly concordant at 2722 Ma (Fig. 2).

The three more elongate grains selected were, of necessity, poorer in quality. Uranium concentrations are slightly higher, compared with the stubby grains, at 130-156 ppm (Table 2) and all three analyses are discordant (I 4%, K 7%, L 2%). Only data point I is plotted because the uncertainty for grain L is large and because K is too discordant. Assuming a minimum zircon age corresponding to the ^{207}Pb-^{206}Pb age and a maximum age based on an average Pb loss at 500 Ma, the time of crystallization of grains I and K can be bracketed as 2740-2710 Ma and 2723-2673 Ma, respectively.

DISCUSSION

An age of 2683 ± 2 Ma obtained from the discordant rhyolite sill defines the minimum age of the iron-formation sequence; the maximum possible age is provided by the youngest detrital age obtained from the granulestone, 2683 ± 2 Ma. Agreement of these ages supports the model of minor volcanism coeval with exhalative iron-formation proposed by Jefferson et al. (1992b, c).

Concerning the lithostratigraphic correlations suggested by Jefferson et al. (1991) and van Breemen et al. (1992a), (Table 1), the new data indicate an age difference of about 10 Ma between the turbidite-hosted iron-formation of George Lake and the top of the main volcanic sequences in the Back River volcanic complex (2692 ± 2 Ma, van Breemen et al., 1987) as well as the Hackett River volcanic belt (2687-2696 Ma on Yava, 2689.3 ± 2 Ma on Musk; Mortensen et al., 1988). The evidence is thus consistent with the proposed correlations but, as it falls short of proof, further geochronology is needed to determine the age of volcanic rocks immediately associated with the Hackett-Back River magnetite iron-formations (Jefferson et al., 1992b).

Both the similarity in age of the detrital zircons to that of the time of deposition and the absence of apparently metamorphic zircons support a hypothesis of minimal admixture from diverse source terranes and a proximal source, as inferred from the petrology. This inference is supported by the stubby subhedral morphology of the younger (2683 Ma to 2697 Ma) group of zircons of a type commonly found in volcanic rocks of the Slave Province.

The only sufficiently constrained ages that might be associated with a metamorphic source, albeit indirectly, are those from stubby grain H at 2722 ± 3 Ma and prismatic zircon I at 2740-2710 Ma. These ages are close to some older ages of volcanism in the Slave Province, near Yellowknife (Isachsen et al., 1990), but are, possibly, not as old as the supracrustal suite characterized by quartzites and ultramafic rocks described by Roscoe (1992) and (Isachsen et al., 1991).

Although our dating of the detrital zircon population is highly biased (we focussed on high quality crystals), the new detrital ages contribute to a body of data that is significant in its lack of evidence for older Archean basement in the eastern Slave Province. Such evidence has recently been summarized by van Breemen et al. (1992b) and includes: the absence of dated older gneisses, the marked difference between high ^{207}Pb-^{204}Pb leads in volcanic massive sulphide deposits in the western Slave versus low ^{207}Pb-^{204}Pb leads in such deposits in the eastern Slave (Thorpe et al., 1992), as well as Nd and Pb isotopic ratios for igneous rocks that are indicative of juvenile crust in the east (Davis and Gariépy, 1992; Davis and Hegner, in press). Throughout the Slave Province, detrital zircons from the turbidite-volcanic assemblage of the Yellowknife Supergroup have yielded, thus far, "syn-volcanic" ages with few older Archean ones (Schärer and Allègre, 1982; Mortensen et al., 1992).

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- 4 -

The timing of emplacement, and distribution of the Sparrow diabase dyke swarm, District of Mackenzie, Northwest Territories

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Bostock, H.H. and van Breemen, O., 1992: The timing of emplacement and distribution of the Sparrow diabase dyke swarm, District of Mackenzie, Northwest Territories; in *Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 49-55.*

Abstract

A U-Pb baddeleyite age of 1827 ± 4 Ma is presented for the time of crystallization of a 50 m-wide dyke of the northwest-trending Sparrow dyke swarm. The age is a minimum for deposition of the Nonacho Group, a continental assemblage of coarse clastics which is intruded by the dykes. This ca. 50 000 km^2 swarm is oriented approximately parallel to the northwesterly oriented branch of the conjugate Bathurst-McDonald fault system and was emplaced within the time constraints for movements along the McDonald fault.

Résumé

La cristallisation d'un dyke de 50 m de largeur, faisant partie de l'essaim de Sparrow à direction nord-ouest, a été établie par la méthode U-Pb sur baddeleyite à 1 827 ± 4 Ma. Cet âge constitue la limite minimale de sédimentation du Groupe de Nonacho, un assemblage continental de roches clastiques grossières recoupé par des dykes. Cet essaim, couvrant environ 50 000 km^2, est presque parallèle à l'embranchement de direction nord-ouest du réseau conjugué des failles de Bathurst-McDonald; sa mise en place s'insère dans la période de temps au cours de laquelle les déplacements le long de la faille McDonald ont eu lieu.

INTRODUCTION

The Sparrow dyke swarm (McGlynn et al., 1974) comprises northwesterly trending mafic dykes that intrude Early Proterozoic and older rocks across an extensive area southeast of Great Slave Lake. This area encompasses northwestern Taltson magmatic zone and a part of the Keewatin hinterland immediately to the east (Fig. 1). The age of emplacement of these dykes is important because they represent a period of crustal extension whose setting within the context of tectonic events in surrounding terranes could not previously be determined due to the large uncertainty inherent in available time constraints. In addition, these dykes intrude the Nonacho Group, a continental assemblage of coarse clastics (Aspler, 1985) whose age of deposition can be constrained only partially and indirectly from relations with north-south directed sinistral strike-slip faulting (pre-1906 Ma., Bostock and Loveridge, 1988) which affects these sediments. The emplacement age therefore provides a test for the minimum constraint age indicated by this faulting.

PREVIOUS WORK

Diabase dykes in the region about Taltson and Nonacho Lakes, although recorded by various earlier reconnaissance geologists (Henderson, 1936; Wilson, 1941; Mulligan and Taylor, 1954) were first examined in detail by McGlynn et al. (1974). The dykes were recognized as belonging to a northwesterly trending, steeply dipping swarm, the Sparrow

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dyke swarm. McGlynn et al. (1974) reported a well-defined paleomagnetic direction for the Sparrow dykes at 12°N, 069°W (dm=9°, dp=6°) based on data from 10 sampling sites. However, no reliable field test is available to establish if the remanence is primary or secondary (K. Buchan, pers. comm., 1992). The Nonacho Group sediments have a similar, but more poorly defined, paleopole at 13°N, 068°W (dm=20°, dp=15°, 12 sites), which is predominantly secondary on the basis of a negative fold test (McGlynn et al., 1974). The age of the least altered dyke was determined at about 1.70 Ga by the $^{39}Ar/^{40}Ar$ whole rock method (McGlynn et al., 1974).

GENERAL GEOLOGY

Extent of the dyke swarm

The majority of dykes in the Sparrow dyke swarm are concentrated within an area stretching from a few kilometres west of Deskenatlata Lake on the west, to Shark Lake and the eastern margin of central Nonacho basin on the east (Fig. 1). Likewise their greatest abundance is bounded on the north within a few kilometres of Great Slave Lake shear zone, and on the south, a few kilometres north of Pilot Lake. Only one dyke of Sparrow trend was found to cut mylonites of the Great Slave Lake shear zone (about 1 km from its southeast margin). West of this area no Sparrow dykes are known, but to the east and south a few dykes are recognized by Fahrig and West (1986) as belonging to the swarm. The southern and eastern limits of the swarm are therefore uncertain. Northeast of Nonacho basin, where the least mapping has been done, the limits of the swarm are most poorly known. Within this area dykes are only locally common and there appear to be substantial regions where few or none occur. The approximate extent of the swarm is shown in Figure 1.

Northwesterly trending Mackenzie dykes (1267 ± 2 Ma., LeCheminant and Heaman, 1989) have been interpreted within the eastern limits of the Sparrow swarm on the basis of northwest-linear aeromagnetic anomalies (Fahrig and West, 1986). Sparrow dykes in contrast typically show little aeromagnetic response, a feature likely due to slight but variable alteration which is not readily detectable in hand

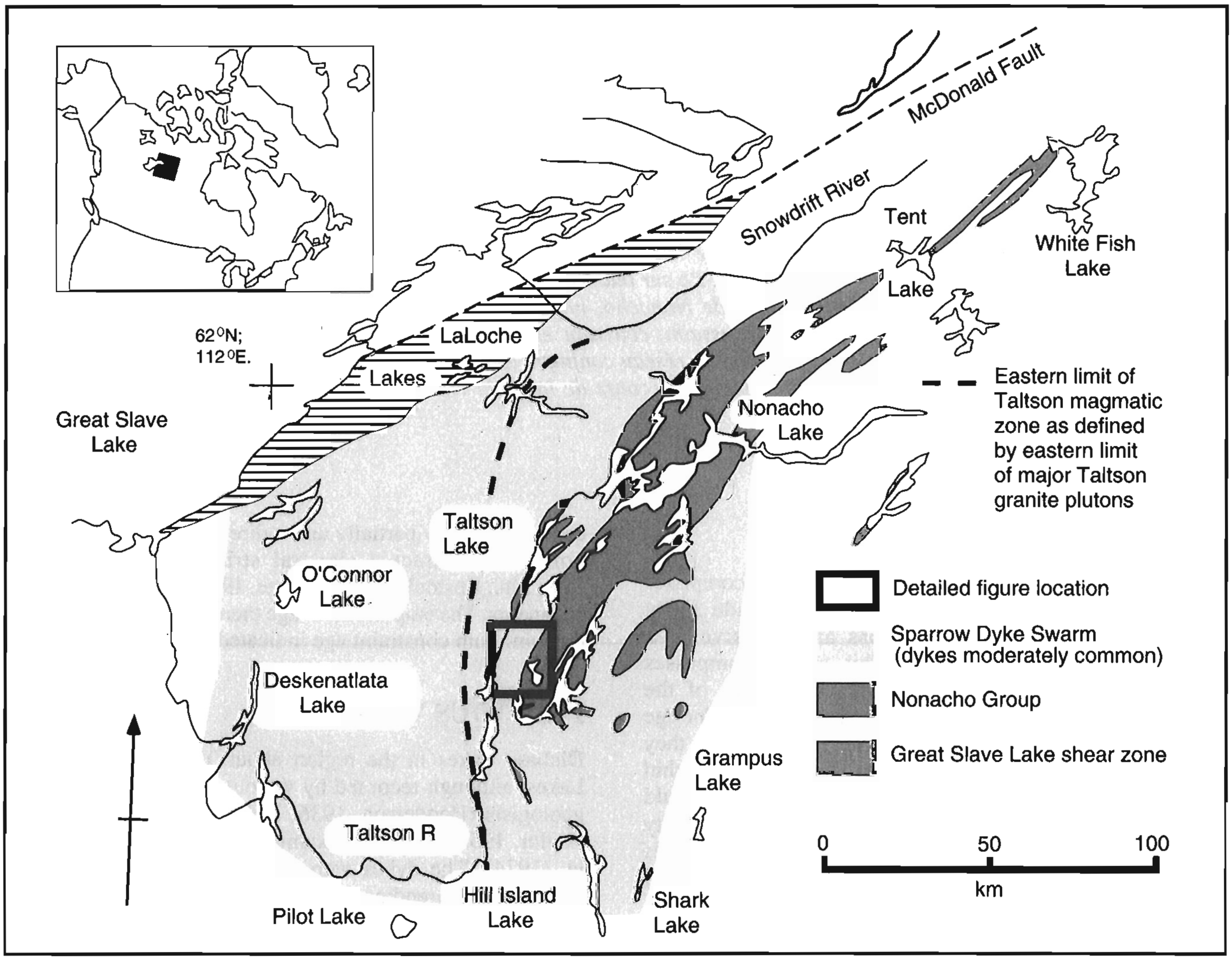

Figure 1. Regional extent of Sparrow dyke swarm.

specimen. Juxtaposition of the two dyke swarms enhances the difficulty of determination of the northeastern limits of the Sparrow swarm.

Rock description

The Sparrow dykes are dark grey to greenish in colour, and typically unfoliated although some are sheared along their margins. In hand specimen the smaller dykes are aphanitic to microlitic or porphyritic, whereas the larger ones are diabasic to gabbroic in texture. Amygdules up to several millimetres in diameter filled with quartz, chlorite, carbonate and locally epidote are present in some dykes, and a few contain up to several per cent by volume of wall-rock xenoliths up to about 20 cm in diameter. In thin section, the major minerals are intermediate to basic plagioclase and clinopyroxene. Pigeonitic pyroxene with low positive 2V is present in some dykes. Two thirds of the dykes examined contain small amounts of quartz. Most dykes show partial alteration of pyroxene to brown-green and/or blue-green hornblende. Sericite is a local product of plagioclase alteration. Chlorite is widely present, small amounts of red-brown or brown-green biotite appear locally and patches of epidote are less common. In most dykes opaque minerals consist of partially altered skeletal crystals of oxides. In some, oxide cores are surrounded by a fringe of titanite. The larger dykes contain interstitial segregations of quartz, sodic plagioclase and locally potash feldspar within which needles of apatite are concentrated. More rarely tiny bladed prisms of brownish to yellowish baddeleyite can be seen to project into these segregations from their margins. Xenocrystic zircon with stubby rounded prismatic form, cracks, and internal zoning, typical of zircon in the wall rocks, is rare.

Structural relations

Most Sparrow dykes trend about 320 degrees and dip steeply. In different parts of the swarm the mean trend varies by roughly ± 15 degrees. Limiting trends vary roughly from 300 degrees to 360 degrees. Dykes along the northerly trending section of Taltson River valley northeast of Pilot Lake commonly have most northerly trends and tend to be more fractured and altered. A very few northeasterly trending dykes, otherwise similar to the northwesterly trending ones, have been observed in the northern part of the swarm, but these have not been traced for more than a few tens of metres. In view of the irregular structure of many northwesterly trending Sparrow dykes it is possible that these apparently aberrant dykes represent magma that locally followed cross fractures. The Sparrow dykes vary from a few millimetres to 50 m but most are less than 2 m in width. The contacts are typically irregularly cross-jointed with minor displacement of the margins. Most display chilled margins. Many of the smaller dykes vary substantially in width over short distances along strike. Anastomosing dykes are present locally. In a few cases dykes with chilled margins have been intruded across earlier dykes at a small angle, but there is no other evidence that the later dykes are significantly younger than those which they transect. Where most abundant, Sparrow dykes commonly occur in isolated clusters of several dykes, but toward the borders of the swarm single dykes are typical.

Most of the Sparrow dykes have little topographic expression, but there is a prominent set of lineaments that trend northwest to north-northwestward across northern Taltson magmatic zone which are approximately parallel to the dykes. These lineaments are at least locally occupied by diabase dykes. Quartz and quartz-carbonate veins are present along some of them. South of O'Connor Lake, where some regional contacts between granite and gneiss are intersected by dykes and lineaments at roughly 45 degrees, there is evidence of sinistral offset of these contacts by up to several kilometres. Similar sinistral offsets are evident crossing the west contact of the main Nonacho basin (Fig. 2, Aspler, 1985) although not all lineaments show significant offset. One particularly well exposed dyke in the O'Connor Lake region has been intruded by quartz veins along its northern margin and the contact zone has subsequently undergone sinistral shear. The relations demonstrate that significant sinistral shear took place after emplacement of some of the dykes but the time interval between dyke emplacement and shearing could have been short, or the dykes could have been emplaced during a period of spasmodic sinistral shear within which later shearing was concentrated along specific fractures or fracture zones.

GEOCHRONOLOGY

The Jerome Lake sample

The central part of the Jerome Lake dyke is well exposed on an island about 20 m across in the central part of the extreme northwest bay of Jerome Lake (Fig. 2). The margins of the dyke are not exposed and the width of the bay allows that this could be one of the largest dykes of the swarm possibly exceeding 50 m in thickness at this point. The island is composed of massive, uniform, dark grey-green, very tough, medium grained gabbro. Thin sections show that plagioclase is delicately internally zoned with sodic rims and patches of chloritic alteration in cores. Brown-green hornblende with local blue-green fringes rims clinopyroxene. About 2% of skeletal opaque oxides are present. Interstitial segregations of graphically intergrown quartz-andesine containing apatite needles form about 15% of the rock. Baddeleyite crystals up to about 30 μm length locally project into these segregations.

Baddeleyite data

Analytical techniques for baddeleyite analysis are identical to those of zircon (Parrish et al., 1987), although the baddeleyite is not abraded. Treatment of analytical errors follows Roddick (1987) and regression analysis York (1969). The isotopic data are presented in Table 1 and displayed in Figure 3.

Baddeleyite fractions have variable U concentrations ranging from 213 ppm to 2075 ppm (Table 1). There is no correspondence between U content and discordance of isotope ratios, which vary only slightly, from 1.2% to 2.0%

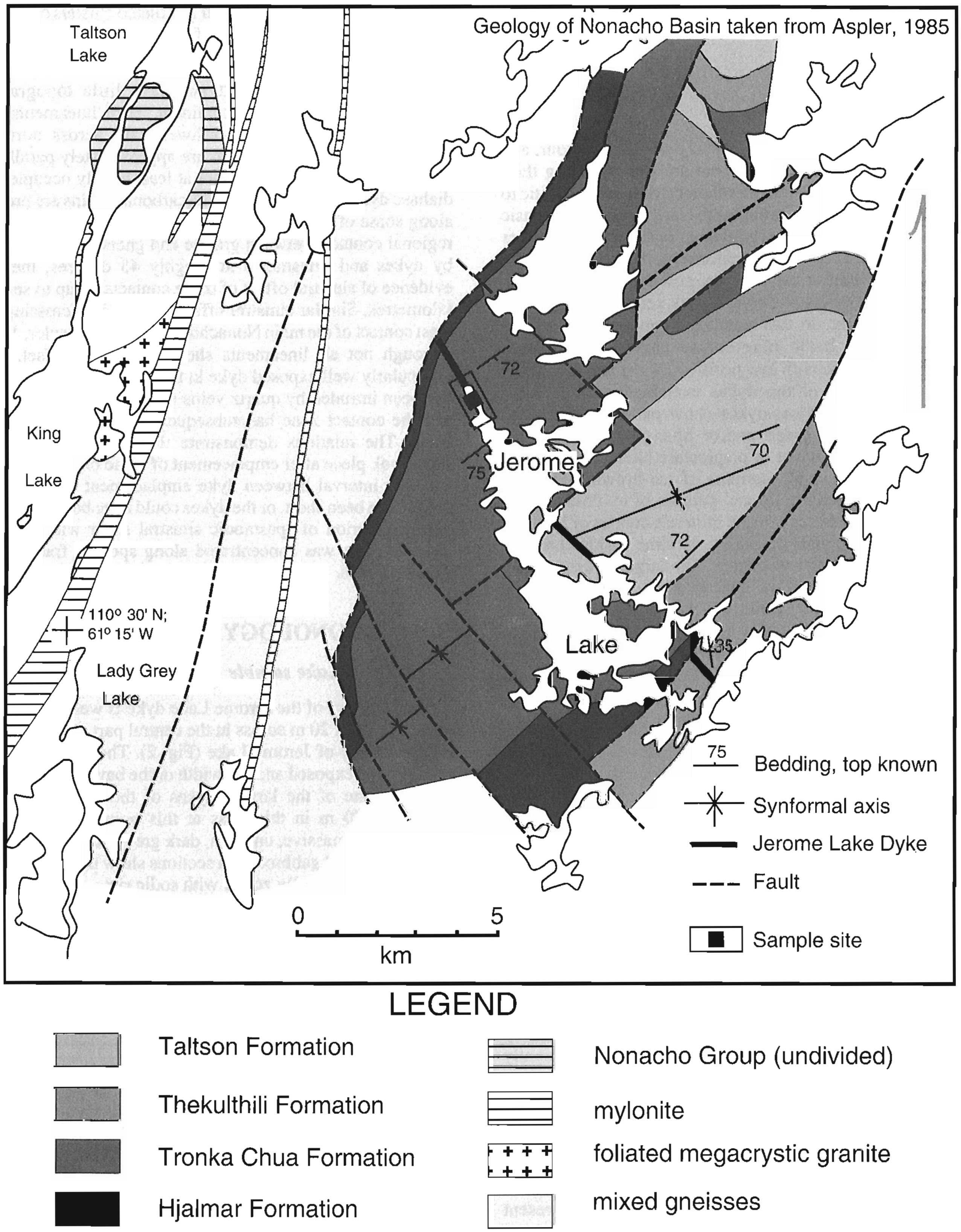

Figure 2. Location of Jerome dyke sampling locality.

(Fig. 3). The five data points fit a line within error (mean square of deviates = 0.15) and a regression analysis provides upper and lower intercept ages of 1830 +38/-8 Ma and 455 Ma, respectively. If fractions C and F with large analytical uncertainties are excluded, a regression line is obtained corresponding to upper and lower intercepts of 1827 +56/-8 Ma and 277 Ma. These intercept ages are preferred as other studies show that lead loss in baddeleyite is mostly recent, at least where there is no evidence for subsequent metamorphic overprinting. Even though pyroxenes have thin amphibolitic rims, the dyke was emplaced at the end of orogenic activity and baddeleyite crystals are pristine. In addition, the age can be more closely constrained with a younger limit of 1823 Ma, provided by the ^{207}Pb-^{206}Pb age for fraction B, and an upper limit of 1831 Ma based on an assumed maximum Pb loss age of 500 Ma for fractions A, B and E. Thus, based on these three fractions, 1827 ± 4 Ma is the age and uncertainty assigned to igneous crystallization.

Table 1. U-Pb isotopic data

Fraction,[1] size	Weight (mg)	U (ppm)	Pb^2 (ppm)	$^{206}Pb^3/^{204}Pb$	Pb_c^4 (pg)	$^{208}Pb^2$ %	$^{206}Pb^5/^{238}U$	$^{207}Pb^5/^{235}U$	R	$^{207}Pb^5/^{206}Pb$	Age(Ma)[6]
Jerome Lake dyke (88-BK-173A; 61° 8' 16"; 110° 18' 35")											
A,≈40, baddeleyite	0.002	518	162	616	33	1.5	0.3233±.11	4.975±.19	0.69	0.11160±.14	1825.7±5.0
B,≈30, baddeleyite	0.001	2075	649	2361	17	1.7	0.3217±.09	4.946±.11	0.88	0.11152±.05	1824.3±1.8
C,≈30, baddeleyite	0.001	416	131	382	22	2.3	0.3220±.17	4.954±.28	0.62	0.11159±.22	1825.5±8.1
E,≈40, baddeleyite	0.005	352	110	1077	33	2.0	0.3206±.10	4.929±.13	0.78	0.11151±.08	1824.2±3.0
F, ≈40, baddeleyite	0.004	213	67	411	44	1.5	0.3238±.14	4.989±.25	0.61	0.11175±.20	1828.1±7.2

Notes: [1] width in microns; [2]radiogenic Pb; [3]measured ratio, corrected for spike and fractionation; [4]total common Pb_c 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] $^{207}Pb/^{206}Pb$ model age; errors are two sigma in Ma.

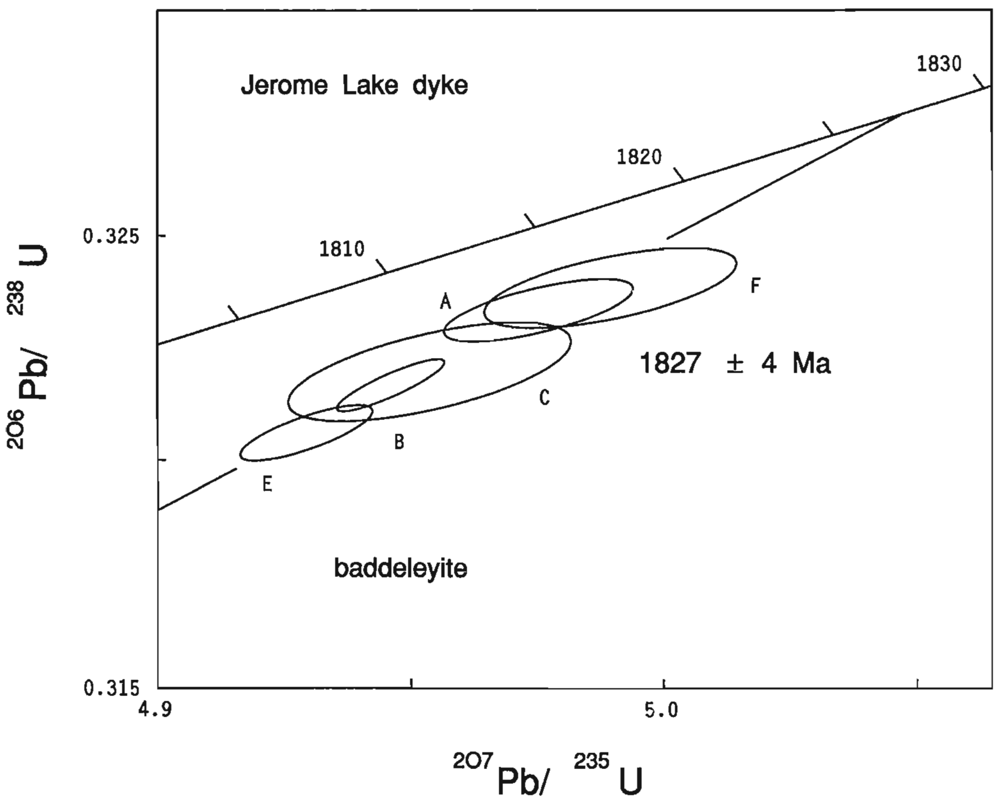

Figure 3.

U-Pb concordia plot of baddeleyite fractions from Jerome dyke. Regression line is for fractions A, B and E, and the lower intercept is 277 Ma.

DISCUSSION

Emplacement of the Sparrow dyke swarm occurred toward the end of a period of deformation involving continental collision between the Slave Province and western Churchill Province ca. 1.97 Ga ago (Bowring and Grotzinger, 1989; Tirrul and Grotzinger, 1990). Ductile interaction between these two continental blocks took place along the Great Slave Lake shear zone during the period 2.0-1.9 (Hanmer et al., 1992), an interval which also included all of the known magmatism within Taltson magmatic zone (1.99-1.88 Ma; Bostock et al., 1987; Bostock and Loveridge, 1988; Bostock et al., 1991). According to Aspler and Donaldson (1986), deposition of Nonacho Group of continental clastic sediments was linked to north-south brittle-ductile, sinistral strike slip faulting associated with the above event (Hanmer et al., 1992). This hypothesis is consistent with the inference from the age of pre-1.91 Ga shearing which has affected the group (Bostock and Loveridge, 1988) and with the 1827 ± 4 Ma age of the cross-cutting Sparrow dykes.

Available evidence suggests that the Sparrow dykes were emplaced during a subsequent tectonic reactivation. A remote collision far to the west is thought to have caused brittle intracratonic indentation of the Slave Province into the Churchill Province along the northwest-trending Bathurst and northeast-trending Mcdonald faults that are sinistral and dextral respectively (Gibb and Thomas, 1981; Hildebrand et al., 1987; Henderson et al., 1990). Related northeast- and northwest-trending conjugate faults occur within and west of Wopmay orogen where they postdate 1.84 Ga granitoid intrusions (Hoffman, 1980; Bowring, 1984; Hoffman et al., 1988; Hildebrand et al., 1987). A minimum age of 1.74 Ga for this compressional event is provided by Rb-Sr biotite uplift ages between Bathurst and McDonald faults (Henderson et al., 1990). Farther east, conjugate faulting of similar age is found in the Dubawnt Lake area (Peterson, 1992), where 1850 +30/-10 Ma minette dykes and syenite stocks (Tella et al., 1985) were linked to the brittle indentation by Hoffman (1980). These intrusions are associated with the volcano-plutonic assemblage of the Christopher Island Formation (Rainbird and Peterson, 1990) predating the Thelon sedimentary basin; formation of this basin at the apex of the Bathurst and McDonald faults is inferred to be synchronous with the indentation event (Henderson et al., 1990).

The Sparrow dykes were emplaced approximately parallel to the northwesterly trend of one of the branches of this Bathurst-McDonald conjugate system. Aspler and Donaldson (1986) have drawn attention to similarly oriented conjugate faults which cut Nonacho Group southeast of McDonald fault (Fig. 2). The Sparrow dykes were emplaced along some but not all of these faults and were at least locally succeeded by sinistral strike slip. They thus appear to be emplaced not only within the time constraints for movements along the Bathurst-McDonald fault system but also in conformation with its geometry. Synchronous intrusion cannot be proven, however, nor can a hypothesis be offered relating dyke orientation to the stress regime of faulting.

Also of interest is the absence of Sparrow dykes north of McDonald Fault. Although this distribution might be explained by regional stresses, or contrasting properties between adjacent Slave and Churchill crusts, it is also possible that pre-existing structural trends have exerted control on magma emplacement; namely, the planar McDonald fault system is at right angles to the dyke swarm and could have impeded lateral magma propagation. Much later, the same fault system appears to have functioned in a similar manner as a barrier to propagation of the 1.27 Ga Mackenzie dykes, which are more abundant northwest of the fault.

Some of the major middle to late Precambrian dyke swarms of the Canadian Shield have been thought to result from continent wide rifting associated with mantle plume activity (Fahrig, 1987; LeCheminant and Heaman, 1989; Heaman et al., 1992). In contrast to these swarms the extent of the Sparrow swarm appears to be restricted, as are the width and length of individual dykes. Furthermore, individual dykes commonly have more irregular contacts, more widely dispersed trends, and typically show low grade metamorphism. These characteristics likely reflect a combination of more local factors, including stress field and magma availability. A less energetic mantle plume might fit these criteria, but it is also possible that other sources in the mantle were tapped in response to cratonic reactivation.

Farther afield, in terranes not separated from the Sparrow dykes by the McDonald Fault, there has been little tectonic activity yet recognized that could be closely associated with the emplacement of this dyke swarm. In the Trans Hudson Orogen most of the plutonism including emplacement of the Wathaman-Chipewyan batholith was over by 1.83 Ga but was succeeded by subaerial volcanism in the Flin Flon belt and by emplacement of felsic sills within the Kisseynew gneisses at this time (Gordon et al., 1990). The Sparrow dykes are geochronologically coeval with these later rocks. In the Wager Bay region to the northeast, possibly related events may be associated with emplacement of a major belt of "calc-alkaline" plutons (ages 1823 ± 3 and 1826 +4/-3 Ma), which intruded Penrhyn Group metasedimentary rocks (LeCheminant et al., 1987).

ACKNOWLEDGMENTS

Widespread use of baddeleyite (ZrO_2) as a reliable mineral medium for U-Pb geochronology of mafic dykes suggested this attempt to date emplacement of the Sparrow swarm directly by this method. We acknowledge the contribution of Larry Aspler who, on the basis of his mapping of the Nonacho Group, suggested the Jerome Lake dyke as a suitable member of the swarm for this purpose (Fig. 2). We are also indebted to Ken Buchan who reviewed earlier paleomagnetic data (McGlynn et al., 1974) for inclusion in this report. The manuscript has been critically read by T.D. Peterson, R. Ernst and K.L. Buchan. U-Pb systematics of baddeleyite has been discussed with J.K. Mortensen.

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U-Pb age determinations from the Glennie Lake Domain, Trans-Hudson Orogen, Saskatchewan

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Abstract

U-Pb zircon and titanite age determinations are presented for five samples from the Brownell Lake area in the southern part of the Glennie Domain and one sample from the Laonil Lake area in the north-central part of the domain. These data are important for refining the understanding of the timing of key events in the evolution of the domain including plutonism, volcanism, deformation, gold mineralization and molasse-type sedimentation. Late granodiorite of the Carroll Lake Block in the Brownell Lake area has been dated at 1834 ±3 Ma and provides a maximum age for phase two deformation. The Brownell Lake and the Maynard Creek Plutons, intrusions within the Brownell Lake Greenstone Belt, yield ages of 1831 ± 9 Ma and 1832 +9/-3 Ma, respectively, and provide minimum age constraints on the timing of D2 deformation and maximum constraints on the timing of the third phase of deformation. Single detrital zircons from a meta-arkose of the Wapawekka Lake Formation within the Brownell Lake Greenstone Belt are 1848 to 1888 Ma, which corresponds to ages of plutonism and volcanism within the Glennie Domain. Plagioclase porphyry that intrudes the Wapawekka Lake Formation is 1834 ± 5 Ma in age, constraining deposition of these sediments to the period between 1848 ±1 to 1834 ±5 Ma. A rhyolite from the upper part of the Pine Lake Metavolcanics in the Laonil Lake map area has been dated at 1838 ±2 Ma, providing further evidence for the timing of volcanism in the Glennie Domain.

Résumé

L'âge U-Pb sur zircon et titanite de cinq échantillons provenant de la région du lac Brownell, dans la partie sud du Domaine de Glennie, et d'un échantillon provenant de la région du lac Laonil, dans le centre nord du domaine, a été déterminé. Ces datations sont importantes pour mieux comprendre la chronologie des événements-clés de l'évolution du domaine (plutonisme, volcanisme, déformation, minéralisation aurifère et sédimentation molassique). Dans la région du lac Brownell, la granodiorite tardive du Bloc de Carroll Lake a été datée à 1 834 ±3 Ma; ce résultat permet d'assigner un âge maximal à la deuxième phase de la déformation. Les plutons du Brownell Lake et du Maynard Creek, qui recoupent par intrusion la Zone de roches vertes de Brownell Lake, ont été datés à 1 831 ± 9 Ma et 1 832 +9/-3 Ma, respectivement, ce qui permet d'associer l'âge minimal à la déformation D2 et l'âge maximal à la troisième phase de la déformation. L'âge de zircons détritiques individuels d'une arkose métamorphisée de la Formation de Wapawekka Lake, faisant partie de la Zone de roches vertes de Brownell Lake, varie entre 1 848 et 1 888 Ma, ce qui correspond aux âges du plutonisme et du volcanisme au sein du Domaine de Glennie. Le porphyre à plagioclases qui recoupe par intrusion la Formation de Wapawekka Lake a été daté à 1 834 ±5 Ma, ce qui limite l'accumulation de ces sédiments à l'intervalle de temps compris entre 1 848 ±1 Ma et 1 834 ±5 Ma. Dans la région du lac Laonil, une rhyolite de la partie supérieure des Métavolcanites de Pine Lake a été datée à 1 838 ±2 Ma, fournissant ainsi des indices chronologiques supplémentaires sur le volcanisme dans le Domaine de Glennie.

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INTRODUCTION

The Glennie Domain is one of a number of distinct lithostructural subdivisions of the internal "Reindeer Zone" of the Early Proterozoic Trans-Hudson Orogen (Lewry and Sibblad, 1977; Hoffman, 1981, 1990; Stauffer, 1984; Lewry and Collerson, 1990). Much of the domain is characterized by arcuate, spatially discontinuous belts of Lower Proterozoic supracrustal rocks, including island arc volcanogenic successions, within variably reworked granitoids and granitoid gneisses (Fig. 1). High strain zones define major parts of the eastern and western boundaries of the Glennie Domain and are also common internally at contacts between supracrustal assemblages and large granitoids (Lewry and Macdonald, 1988; Delaney, 1988, 1989). The north-south trending Tabbernor Fault Zone, at or near the eastern boundary of the Glennie Domain, is characterized by tight north-south folds, locally high metamorphic gradients, and early mylonitic and late brittle deformation zones (Sibbald, 1978; Macdonald, 1975; Wilcox, 1990). The western boundary of the domain is defined by the Stanley Shear Zone, a one to four kilometre wide north-northeasterly trending straight zone subjected to early heterogeneous ductile strain and subsequent late brittle deformation (Lewry et al., 1980; Lewry, 1981). In the western part of the domain, in the Iskwatikan and Hunter Bay domes, an assemblage of Lower Proterozoic rocks containing shallow-dipping zones of high strain overlies Archean basement gneisses (Lewry and Slimmon, 1985; Chiarenzelli, 1989; Chiarenzelli et al., 1987; Lewry et al., 1990).

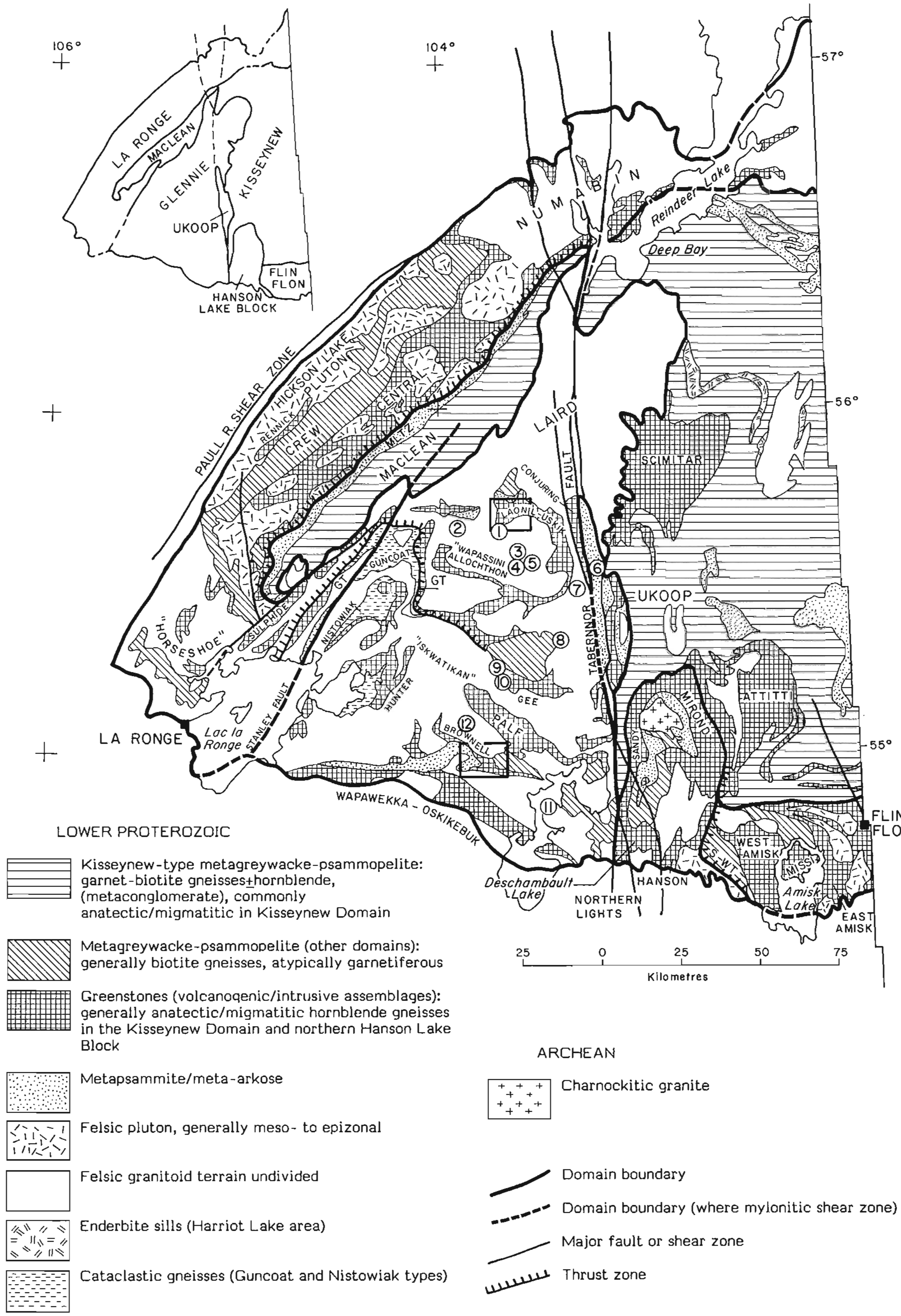
Supracrustal rocks of the Glennie Domain can be subdivided into three distinct assemblages: volcanogenic, arkosic, and pelitic/psammopelitic. Volcanogenic assemblages are composed of volcanics, syn-volcanic intrusives, volcaniclastics and associated sediments. The volcanics in most of these assemblages, including the Pine Lake Metavolcanics, Brownell Lake Group and Gee Lake Metavolcanics, are predominantly mafic to intermediate flows and fragmentals; felsic varieties such as the sodic rhyolites and dacites of the Pine Lake Metavolcanics in the area north of Laonil Lake (Fig. 1) are rare. Preliminary geochemical investigations suggest that most of the volcanics are tholeiitic in character and were emplaced either in a primitive island arc setting or mid ocean ridge spreading centre (Delaney, unpublished data; Slimmon, pers. comm., 1991).

The second variety of supracrustal assemblage consists of arkose, conglomeratic arkose, grit and conglomerate and includes the Porky Lake siliciclastics (Lewry, 1977; Delaney, 1986), Ourom Lake meta-arkoses (Budding and Kirkland, 1956; Delaney, 1987; Delaney et al., 1988) and Wapawekka Lake Formation (Padgham, 1966, 1967, 1968; Delaney, 1988, 1989). Although arkosic successions throughout the Trans-Hudson Orogen have been broadly correlated (Macdonald and Broughton, 1980; Lewry, 1983; Delaney et al., 1988; Stauffer, 1990), relationships between these distinctive molasse-like assemblages are unclear even within the Glennie Domain. For example in the Pine Lake Greenstone Belt, the Ourom Lake Formation meta-arkose is intercalated with the Pine Lake Metavolcanics, whereas in the Brownell-Wapawekka Lakes Greenstone Belt, arkosic rocks of the Wapawekka Lake Formation appear to conformably overlie volcanics and volcaniclastics of the Wapawekka Narrows and Churchman Island Formation (Padgham, 1966, 1967, 1968; Delaney, 1988, 1989).

Pelitic and psammopelitic assemblages occur east of Sadler Lake (Chakrabarti, 1968, 1969; Slimmon, 1989), around Glennie Lake (Lewry, 1977) and north of Porky Lake in the core of the Ray Lake Synform (Johnston, 1968; Lewry, 1977; Delaney, 1986). In some areas, such as north of Porky Lake, these pelitic rocks are the facies equivalent of the arkosic rocks; relationships elsewhere are obscure.

This paper presents U-Pb zircon and titanite age determinations for five samples from the Brownell Lake area in the southern part of the Glennie Domain and one sample from the Laonil Lake area in the north-central part of the domain (Fig. 1). These data are important for refining the understanding of the timing of key events in the evolution of the domain including plutonism, volcanism, deformation, gold mineralization and molasse-type sedimentation.

Figure 1. Geological elements of the major part of the Reindeer Zone, Trans Hudson Orogen in Saskatchewan (modified from the Saskatchewan Geological Survey, 1987). Boxes show location of the Laonil Lake (northern box) and Brownell Lake (southern box) areas discussed in this paper. The numbers are locations of some of the U-Pb and Sm-Nd age determinations referred to in the text: 1. Laonil Lake Intrusive Complex; 2. unnamed "gneiss"; 3. Eyahpaise Lake Pluton; 4. Wykes Lake Pluton; 5. Late intrusive that cuts Eyahpaise Lake Pluton; 6. Ourom Lake meta-arkose; 7. Wood Lake Batholith; 8. Daniel - Tri Lake mafic-ultramafic complex; 9. Gee Lake rhyolite; 10. aplite; 11. Deschambault Narrows tonalite; 12. Carroll Lake Gneiss. The following domains/subdomains are located in the figure: Kisseynew = Kisseynew Domain; Maclean = Maclean Lake Belt; Glennie = Glennie Domain; Ukoop = Ukoop Lake Segment; La Ronge = La Ronge Domain; Crew = Crew Lake Belt; Numabin = Numabin Complex; Horseshoe = La Ronge Horseshoe terrane; Iskwatikan = Iskwatikan Subdomain; Laird = Laird Lake Complex; Flin Flon = Flin Flon Domain. Greenstone belts: Central = Central Metavolcanic Belt; Rennick = Rennick Lake; Sulphide = Sulphide-Hebden-MacKay Lakes; Hunter = Hunter Bay; Wapawekka-Oskikebuk; Brownell = Brownell Lake; Palf = Palf Lake; Gee = Gee Lake; Laonil Uskik = Pine Lake; Conjuring = Conjuring segment of Pine Lake. Flin Flon greenstone belts: West Amisk; Missi (Island); East Amisk; Hanson = Hanson Lake Volcanics; Northern Lights = Northern Lights Volcanics. Hornblende gneiss complexes (volcanogenic): Scimitar = Scimitar Lake Complex; Attitti = Attitti Lake Complex; Sandy = Sandy Narrows; Mirond = Mirond Lake. Shear zones or thrusts: MLTZ = McLennan Lake Tectonic Zone; S-WT = Sturgeon-Weir Thrust; GT = Guncoat Thrust.


106°
104°
57°
56°
55°
LA RONGE
MACLEAN
GLENNIE
KISSEYNEW
UKOOP
FLIN FLON
HANSON LAKE BLOCK
N U M A B I N
Reindeer Lake
Deep Bay
PAULL R. SHEAR ZONE
HICKSON LAKE PLUTON
RENNICK
CREW
CENTRAL
METZ
MACLEAN
LAIRD
FAULT
CONJURING
LAONIL-USKIK
SCIMITAR
"WAPASSINI ALLOCHTHON"
GUNCOAT
GT
UKOOP
TABERNOR
"HORSESHOE"
SULPHIDE
NISTOWIAK
"SKWATIKAN"
HUNTER
GEE
PALF
BROWNELL
MIROND
SANDY
ATTITTI
LA RONGE
Lac la Ronge
STANLEY FAULT
WAPAWEKKA - OSKIKEBUK
Deschambault Lake
NORTHERN LIGHTS
HANSON
S-WT
WEST AMISK
MISSI
Amisk Lake
EAST AMISK
FLIN FLON
25 0 25 50 75
Kilometres
LOWER PROTEROZOIC
Kisseynew-type metagreywacke-psammopelite: garnet-biotite gneisses±hornblende, (metaconglomerate), commonly anatectic/migmatitic in Kisseynew Domain
Metagreywacke-psammopelite (other domains): generally biotite gneisses, atypically garnetiferous
Greenstones (volcanogenic/intrusive assemblages): generally anatectic/migmatitic hornblende gneisses in the Kisseynew Domain and northern Hanson Lake Block
Metapsammite/meta-arkose
Felsic pluton, generally meso- to epizonal
Felsic granitoid terrain undivided
Enderbite sills (Harriot Lake area)
Cataclastic gneisses (Guncoat and Nistowiak types)
ARCHEAN
Charnockitic granite
Domain boundary
Domain boundary (where mylonitic shear zone)
Major fault or shear zone
Thrust zone

PREVIOUS GEOCHRONOLOGY OF THE GLENNIE DOMAIN

Detailed U-Pb zircon, monazite, apatite and titanite geochronology investigations have been completed by Chiarenzelli (1989) in the western part of the Glennie Domain. The geochronology database is more fragmentary, however, for most of the domain, including the areas discussed in this paper (see Delaney et al., 1988). Chiarenzelli (1989) recognized Archean rocks which are at least 2800 Ma old interlayered with or intruded by younger Archean granitoids which are approximately 2500 Ma old in nappe sheets of the Nistowiak and Iskawatkin Lakes area. Elsewhere in the domain the oldest known rocks are the 1893 ±35 Ma Caroll Lake gneiss, from the south-central part of the domain (Fig. 1, 2; Van Schmus et al., 1987) and a quartz diorite phase from the synvolcanic Laonil Lake Intrusive Complex which has been dated at 1889 ±9 Ma (Chiarenzelli, 1989). Two ages have been determined for volcanic sequences within the Glennie Domain: 1881 ±7 Ma for the Palf Lake rhyolite (Chiarenzelli, 1989) and ca. 1.87 Ga for a highly strained rhyolite from a felsic volcanic horizon in the Gee Lake Metavolcanics (Fig. 2; Slimmon, 1988; Delaney et al., 1990). In addition, an age of 1837 ±7 Ma was obtained for a dioritic gneiss, interpreted to be an andesitic flow or shallow intrusive, that was collected in the Nistowiak Lake area (Fig. 2; Chiarenzelli, 1989). A major phase of tonalitic to granodioritic plutonism during the period 1846-1859 Ma

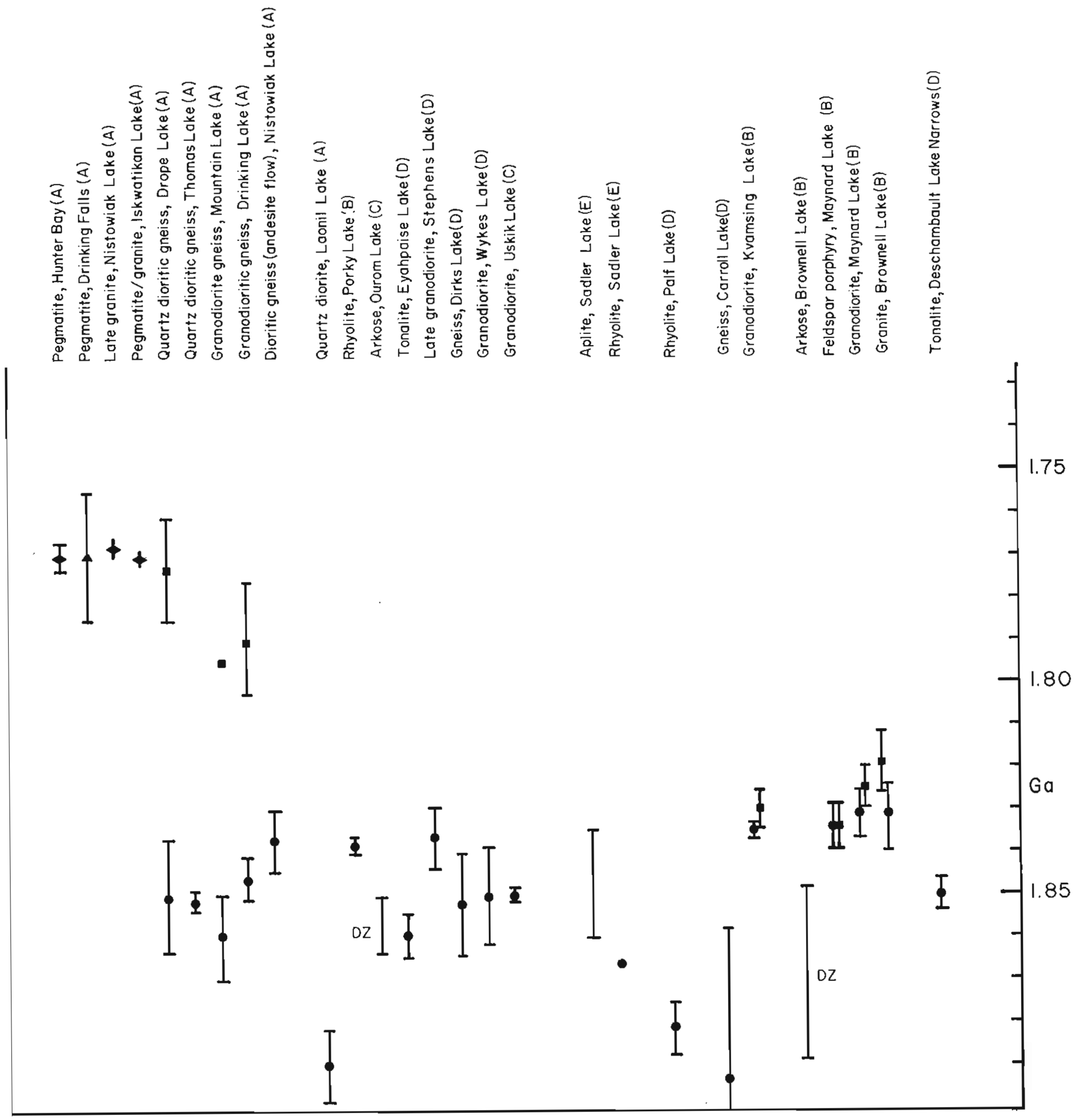


Figure 2. Summary chart of Proterozoic U-Pb age determinations for the Glennie Domain. Diamond = monazite; square = titanite; triangle = allanite; circle = zircon; DZ = detrital zircons. Bars on either side of symbol shows errors. Sources: A = Chiarenzelli, 1989; B = this paper; C = Delaney et al., 1988; D = Van Schmus et al., 1987.

(Fig. 2; Van Schmus et al., 1987; Delaney et al., 1988; Chiarenzelli, 1989) formed bodies such as the Eyahpaise Pluton, Wykes Lake Granodiorite, Wood Lake Batholith, Deschambault Narrows tonalite and intrusions in the Nistowiak-Iskwatkin Lakes area. Dating of detrital zircons from the Ourom Lake Formation meta-arkose revealed that this unit was derived from rocks 1850-1863 Ma old (Delaney et al., 1988). A second, though apparently more restricted, phase of plutonism at 1837 Ma formed the small intrusion in the core of the Eyahpaise Pluton (Van Schmus et al., 1987). Four monazite fractions from syntectonic to post-tectonic granites and pegmatites in the Nistowiak – Iskwatkin Lakes area yielded dates of 1762-1770 Ma, interpreted to represent ages of crystallization (Chiarenzelli, 1989). U-Pb titanite ages, also from rocks in the western part of the domain, range from 1773 to 1790 Ma and were interpreted to represent the timing of titanite closure following peak metamorphic conditions (Chiarenzelli, 1989). A mafic-ultramafic complex in the Daniel-Tri Lakes area, west of Wood Lake, has a Sm-Nd crystallization age of 1.71 ±0.05 Ga (Hegner et al., 1989).

GEOLOGY OF THE BROWNELL LAKE AREA

The Brownell Lake area, located in the southern part of the Glennie Lake Domain, is composed of four distinct lithotectonic elements: the Oskikebuk Block, the Brownell Lake Greenstone Belt, the Carroll Lake Block and the Kvamsing Cataclastic Belt (see inset of Fig. 3; Delaney, 1988). The southern boundary of the map area is defined by the northern margin of the dioritic Oskikebuk Block, which is separated from the Brownell Lake Greenstone Belt by the east-west trending Hartley Shear Zone.

The Brownell Lake Greenstone Belt is composed of supracrustal and granitoid rocks that have undergone amphibolite facies metamorphism (Delaney, 1988). Supracrustal rocks are divided into two main assemblages: the Brownell Lake Group, in the eastern part of the area, and the Wapawekka Lake Formation of the Wapawekka Group on the west side of Brownell Lake (Padgham, 1966, 1968; Delaney, 1988, 1989). The Brownell Lake Group is composed of both volcanic and sedimentary rocks; volcanic rocks include mafic to intermediate flows, associated intrusives and volcaniclastic rocks which predominate over felsic volcaniclastics and volcanics. The Wapawekka Lake Formation consists predominantly of arkose (sample 8822-376) intruded by locally abundant plagioclase porphyry sills (sample 9022-94) and minor intermediate volcanics and volcaniclastics.

There are several distinct granitoid bodies in the Brownell Lake Greenstone Belt. An early diorite and quartz monzonite body lies south of the eastern end of Brownell Lake; other small intrusions are composed of tonalite to granodiorite or porphyritic granodiorite. The greenstone belt is cut by two later composite plutons (see Fig. 3) – the Brownell Lake Pluton (sample 8822-377) which is composed of porphyritic granite and diorite phases, and the Maynard Creek Pluton (sample 8822-378) which is composed of leucocratic granodiorite and quartz diorite phases.

The Carroll Lake Block to the north is separated from the Brownell Lake Greenstone Belt by the curvilinear Lake Shear Zone, the trace of which passes through Brownell Lake. Only the southeast part of this block has been investigated in any detail (Scott, 1981; Delaney, 1988) and in that area it consists mainly of a late-stage leucocratic granodiorite (sample 8822-375). Near the southern margin of this intrusive is a discontinuous belt of older granitoid, supracrustal and cataclastic rocks. The Rustad Lake Quartz Monzodiorite forms much of the southern margin of the Carroll Lake Block.

The Kvamsing Cataclastic Belt comprises moderately to generally strongly deformed rocks consisting mainly of amphibolite; deformed pelitic and calc silicate-bearing rocks are subordinate. Locally there are undeformed remnants of amygdaloidal mafic volcanics and basic intrusives.

Three episodes of deformation are recognized in the Brownell Lake area (Delaney, 1988). The oldest phase of deformation, D1, formed mylonites and strongly deformed rocks only recognized in the Kvamsing Cataclastic Belt and the Carroll Lake Block. The second phase of deformation, D2, resulted in significant faulting and folding. The major shear zones which bound the Brownell Lake Greenstone Belt probably formed during this episode, as did the east-trending overturned folds and the strong L-S or locally L-fabric within the Brownell Lake Greenstone Belt and Carroll Lake Block. Phase three deformation resulted in a series of north-northeast-trending folds, including the Brownell Lake Anticline, Maynard Creek Anticline and Brownell Lake Syncline, which deform all of the lithotectonic elements. A second episode of movement along high strain zones flanking the Brownell Lake Greenstone Belt also occurred during this period.

Several gold showings have been identified in the Brownell Lake Greenstone Belt. These occur in quartz veins at or near the margins of the Brownell Lake Pluton and within or adjacent to high strain zones in volcanics of the Brownell Group and dioritic phases of the Maynard Creek Pluton (Delaney, 1990a, b, in press).

U-Pb AGE DETERMINATIONS FROM THE BROWNELL LAKE AREA

This paper reports on five U-Pb age determinations from the Brownell Lake map area (Fig. 3) and one U-Pb age determination from the Laonil Lake area (described below). Preliminary data for four of the Brownell Lake rocks have been summarized by Delaney et al. (1990). The data presented in this paper supplant the previous summaries.

Analyses of the Carroll Lake granodiorite (8822-375), Wapawekka Lake meta-arkose (8822-376), Brownell Lake granite (8822-377), and Maynard Creek granodiorite (8822-378) were performed in the Geochronology Laboratory at the Geological Survey of Canada. Analyses of the Maynard Creek porphyry (9022-94) and the Pine Lake meta-rhyolite (9022-92), which is from the Laonil Lake area, were performed at the Jack Satterly Laboratory, Royal Ontario Museum (ROM).

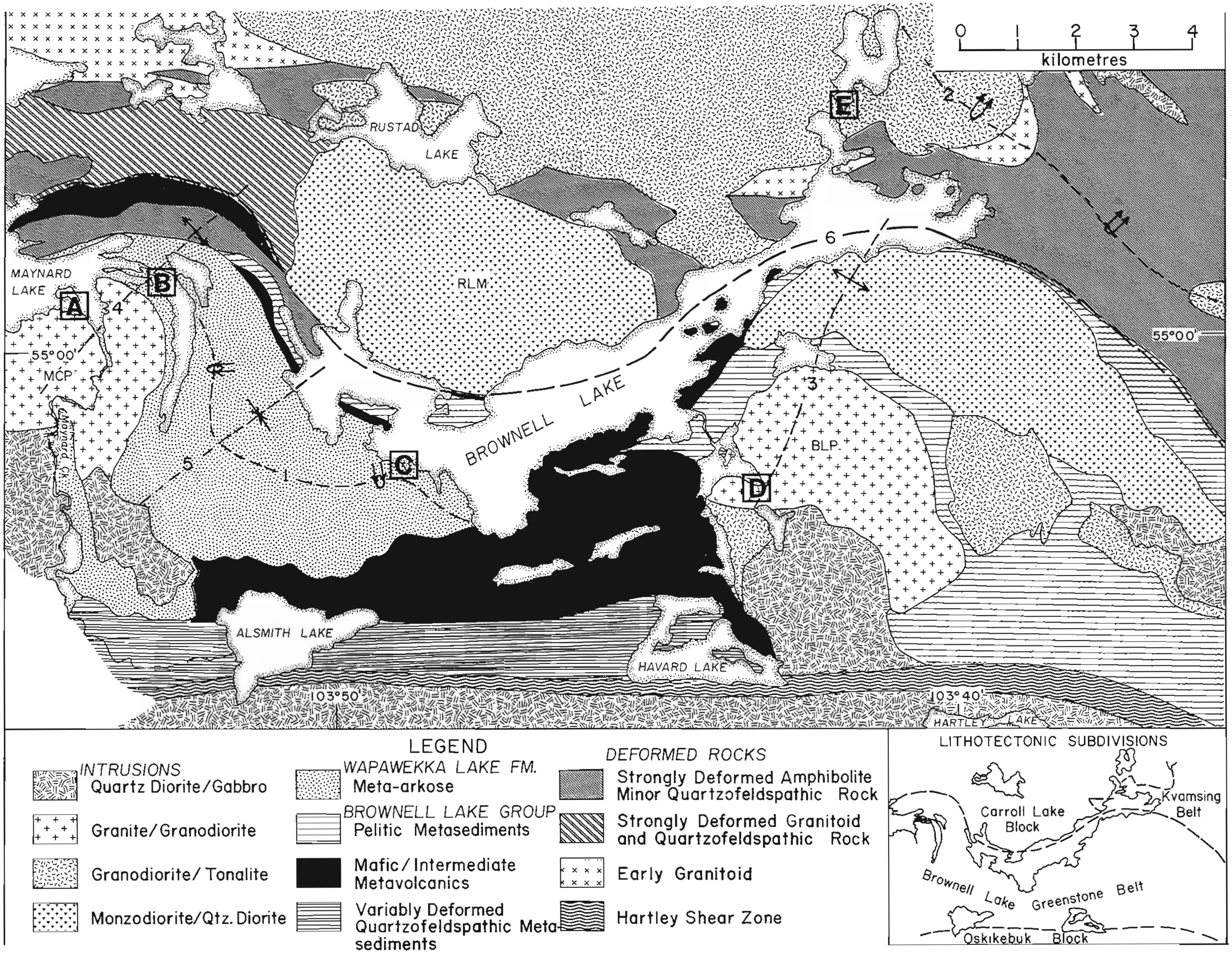


Figure 3. Geological sketch map of the Brownell Lake area. BLP = Brownell Lake Pluton; MCP = Maynard Creek Pluton; RLM = Rustad Lake Quartz Monzodiorite. The following features are numbered on the map: 1. Brownell Creek Syncline; 2. unnamed overturned antiform in Carroll Lake Block; 3. Brownell Lake Anticline; 4. Maynard Creek Anticline; 5. Brownell Lake Syncline; 6. Sole of the Lake Shear Zone. Letters in boxes show locations of geochronology samples: A = 8822-378; B = 9022-94; C = 8822-376; D = 8822-377; E = 8822-375.

Isotopic data are presented in Table 1 and displayed in the concordia plots (Fig. 4-8). U-Pb analytical methods are those outlined in Parrish et al. (1987) for samples analysed at the GSC, and Krogh (1973) for those analyzed at the ROM. Techniques included air abrasion (Krogh, 1982) for all zircon fractions. The locations of samples from the Brownell Lake area are plotted in Figure 3; exact localities are described in the appendix.

Carroll Lake granodiorite (sample 8822-375)

The buff to pinkish buff, medium-grained Carroll Lake granodiorite is generally moderately foliated and strongly lineated, with the lineation defined by elongate quartz grains up to 10 mm in length. The granodiorite body is deformed by a southeast-plunging overturned D2 antiform near the margin of the Carroll Lake Block (E, Fig. 3). This granodiorite intrudes poorly understood highly deformed granitoid rocks which comprise much of the western part of the Carroll Lake Block. A sample of gneiss from this package of highly deformed rocks has been dated at 1893 ±35 Ma (U-Pb zircon; Fig. 1, 2, Van Schmus et al., 1987).

Two fractions of elongate, cloudy, euhedral igneous zircons were analyzed from the Carroll Lake granodiorite. The zircons, which were only of fair quality, are discordant, having experienced some Pb loss. A minimum age for magmatic crystallization is best interpreted to be about 1834 ±3 Ma, based on the ^{207}Pb-^{206}Pb ages of the two zircon fractions (Fig. 4). A two point regression of the zircon data yields an upper intercept of 1833 ±8 Ma, consistent with this estimate. Two fractions of clear, golden brown titanite were also analyzed from this sample. The cooling age of titanite (closure temperature of titanite is approximately 600°C (Tucker et al., 1987; Heaman and Parrish, 1991)) based on the ^{207}Pb-^{206}Pb ages of nearly concordant analyses, is interpreted to be approximately 1830 ± 10 Ma.

Wapawekka Lake Formation meta-arkose (sample 8822-376)

Light grey to buff meta-arkose is the predominant lithology of the Wapawekka Lake Formation in the western part of the Brownell Lake Greenstone Belt. It is composed mostly of feldspar and quartz, with minor amounts of biotite, muscovite and magnetite. Southeast of Maynard Lake, the locality of sample 8822-376 (C, Fig. 3), the unit is weakly deformed, typically thin- to medium-bedded, and contains trough crossbedding. In this area, the meta-arkose lies in the core of an overturned, east-trending D2 syncline, the Brownell Creek Syncline.

Ten high quality, single zircon grains of various morphologies were selected for analysis from the meta-arkose. These included subfacetted to subrounded equant grains and elongate, well-facetted euhedral crystals which were all subsequently strongly abraded (Krogh, 1982). Individual zircons were analyzed from the meta-arkose to ensure that each analysis represented a unique age and not a mixture of several ages. The grains produced concordant to moderately discordant analyses. The ages of the most concordant analyses ranges from 1848 to 1888 Ma, with most grains in the 1850-1870 Ma time span (Fig. 5). The scatter in data points is interpreted to reflect a range in source ages,

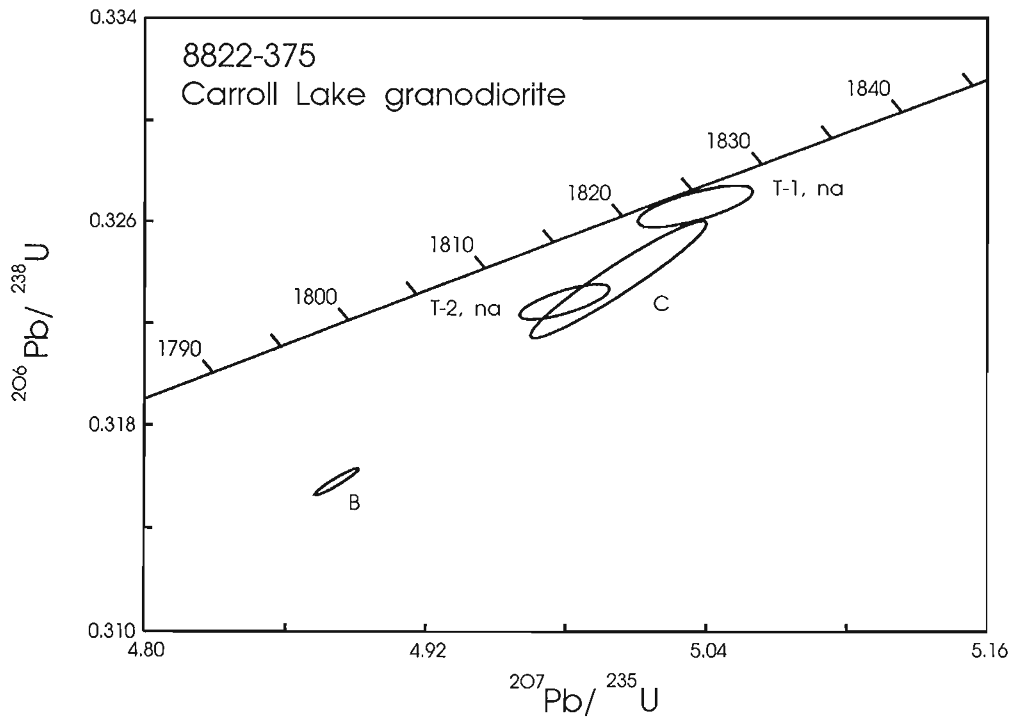


Figure 4.

U-Pb concordia diagram of zircon and titanite from sample 8822-375, Carroll Lake granodiorite. Error ellipses are 2σ.

coupled with differing amounts of Pb loss. The detrital zircon data indicate that the age of the Wapawekka Lake Formation meta-arkose is younger than 1847.6 ± 1.4 Ma, the age of concordant fraction G.

Maynard Creek feldspar porphyry (sample 9022-94)

Along the southern margin of the Brownell Creek Syncline there are several sills of plagioclase porphyry intruding meta-arkose of the Wapawekka Lake Formation. The greatest concentration of sills is between the southeast arm of Maynard Lake and Alsmith Lake (B, Fig. 3; Delaney, 1988). The plagioclase porphyry is characterized by medium- to rarely coarse-grained phenocrysts of plagioclase in a fine grained groundmass composed of quartz, feldspar and biotite. Individual sills range in thickness up to a few metres and are typically foliated as a result of D2 deformation.

The sample contained, for the most part, a single zircon population characterized by tiny turbid crystals with fractures and resorbed surfaces. A single, colourless, uniquely transparent crystal (fraction 4 in Table 1, Fig. 6) is interpreted to be a xenocryst, with a ^{207}Pb-^{206}Pb age of 1864 Ma. Three multi-grain zircon fractions (fractions 1,2,5) and a single titanite fraction (labelled T in Fig. 6) define a discordia line with an upper intercept age of 1834 ± 5 Ma and a lower intercept age of ca. 100 Ma. The 1834 ±5 Ma upper intercept age is interpreted as the best estimate for the emplacement age of this porphyry.

Brownell Lake granite (sample 8822-377)

This sample, a pinkish-buff, medium grained granite containing euhedral to subhedral phenocrysts of microcline, was collected from the northern part of the Brownell Lake Pluton (D, Fig. 3). Although locally massive, more typically this granite is foliated and lineated, structures attributed to the second and third phases of deformation. The pluton is deformed by a north-northeast-trending D3 antiform called the Brownell Lake Anticline (Padgham, 1968; Delaney, 1988).

Zircons from the Brownell Lake Pluton are elongate, euhedral, light brown frosted crystals with many needle-like inclusions. A modified York linear regression (see Parrish et al., 1987) of four zircon fractions result in a chord with an upper intercept of 1831 ±9 Ma, which is interpreted as the age of crystallization of the pluton (Fig. 7). Discordance of the zircon analyses is probably a result of Pb loss enhanced by the high U contents (>1500 ppm) of the poor-quality zircons. Two fractions of clear, golden brown, irregular-shaped titanite were also analyzed from this sample. The age of titanite is interpreted to be 1819 ± 7 Ma, based on the ^{207}Pb-^{206}Pb ages of the titanite fractions. Because it is slightly younger than the interpreted zircon age, 1819 ±7 Ma may represent the cooling age of the pluton.

Maynard Creek granodiorite (sample 8822-378)

This sample was collected from the northern part of the Maynard Creek Pluton, a buff to pinkish-buff, medium grained, leucocratic granodiorite which is typically foliated and lineated (A, Fig. 3). As in the case of the Brownell Lake

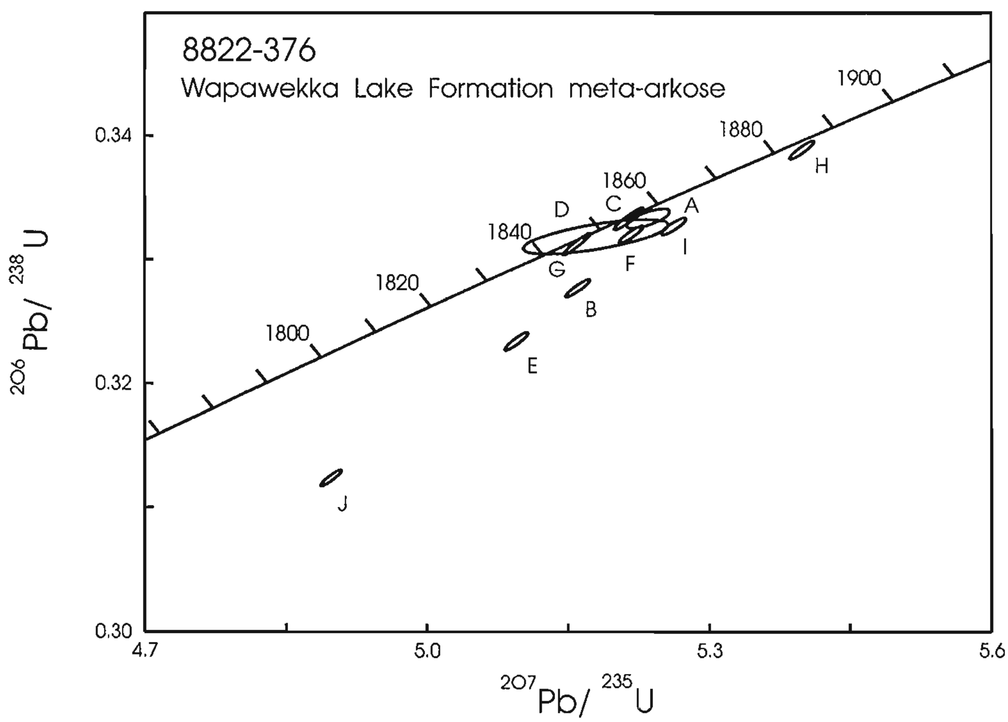


Figure 5.

U-Pb concordia diagram of detrital zircons from sample 8822-376, Wapawekka Lake Formation meta-arkose. Error ellipses are 2σ.

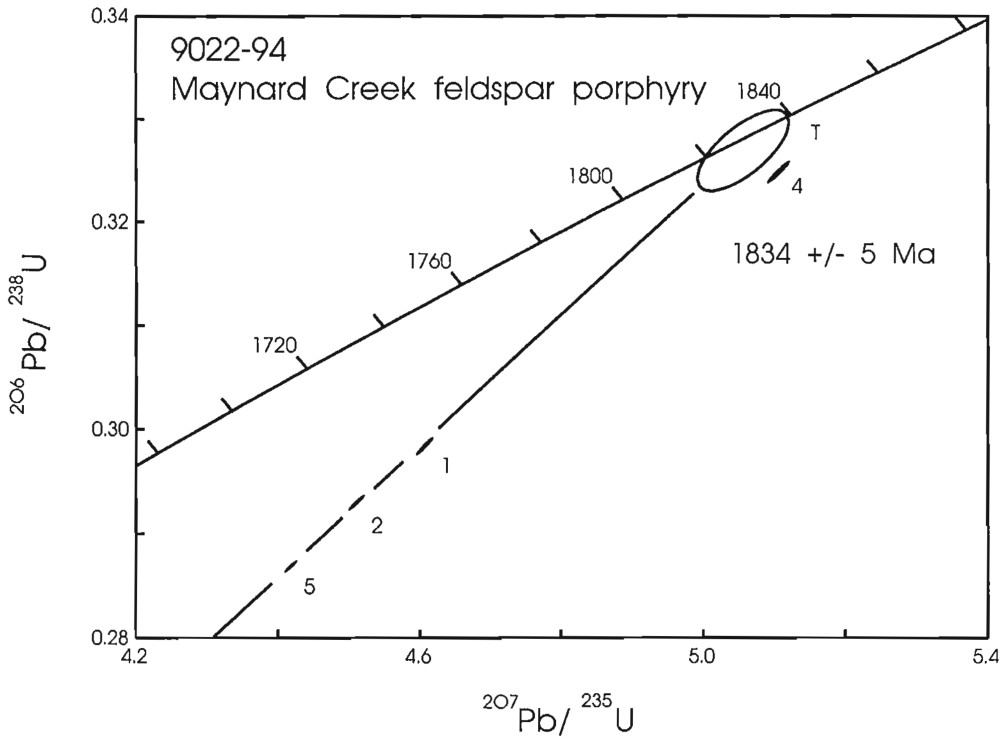


Figure 6.

U-Pb concordia diagram of zircon and titanite from sample 9022-94, Maynard Lake plagioclase porphyry. Error ellipses are 2σ.

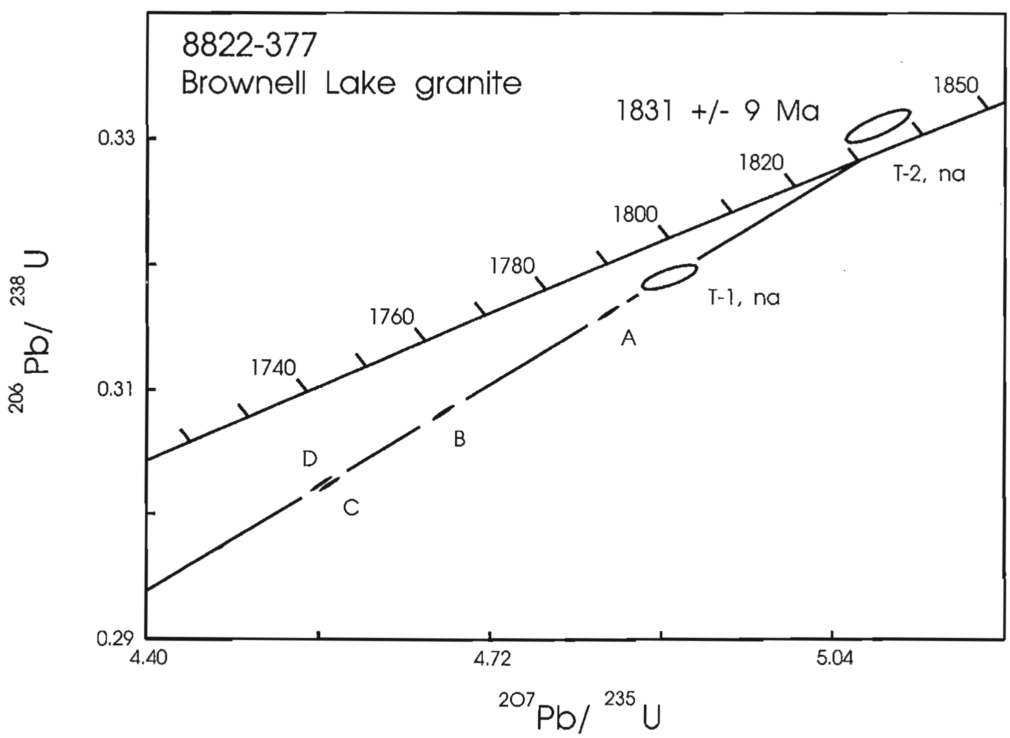


Figure 7.

U-Pb concordia diagram of zircon and titanite from sample 8822-377, porphyritic granite from the Brownell Lake Pluton. Error ellipses are 2σ.

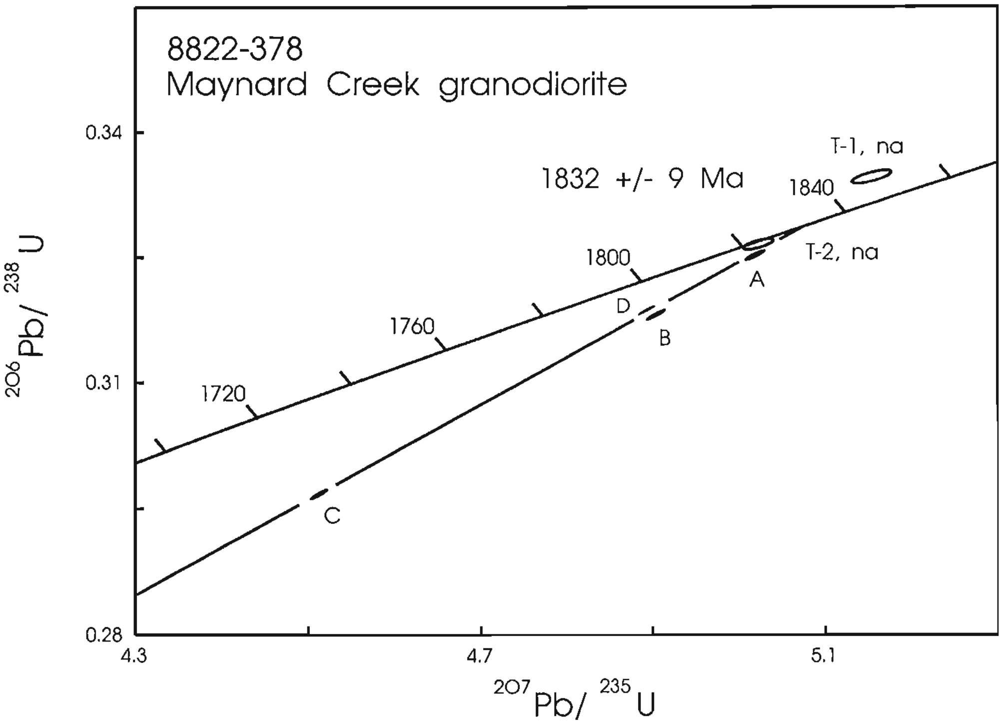


Figure 8.

U-Pb concordia diagram of zircon and titanite from sample 8822-378, granodiorite from the Maynard Creek Pluton. Error ellipses are 2σ.

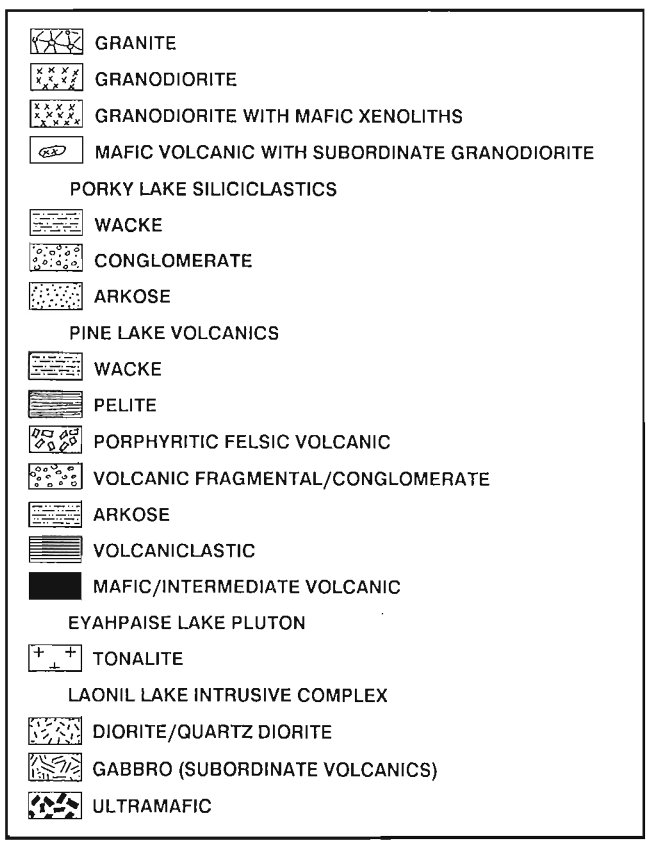
Pluton, the Maynard Creek Pluton is deformed by a northeast-trending third phase antiform, the Maynard Creek Anticline (Padgham, 1968; Delaney, 1988).

Very clear, euhedral, elongate, igneous zircons with minor inclusions were analyzed from the Maynard Creek Pluton. A linear regression of 4 zircon points has an upper intercept of 1832 ±9 Ma (Fig. 8). The least discordant point A is quite close to concordia, lending confidence to this interpretation. In addition the ^{207}Pb-^{206}Pb age of fraction A, 1830.8 ± 2.2 Ma, is a minimum age of the pluton; it therefore is older than 1828.6 Ma. We infer that a realistic estimate of the age of the zircon is 1832 +9/-3 Ma, which is interpreted to be the crystallization age of the pluton. Analyses of two fractions of clear, golden brown, irregular-shaped titanite result in interpreted cooling ages of 1825 ± 5 Ma, based on their ^{207}Pb-^{206}Pb ages. This cooling age is within the analytical error of the formation age of the pluton.

GEOLOGY AND U-Pb AGE DETERMINATION FROM THE LAONIL LAKE AREA

The Laonil Lake map area comprises Lower Proterozoic supracrustal and granitoid rocks that record a complex history of magmatism, deformation and metamorphism (Fig. 9). The oldest rocks are from the Laonil Lake Intrusive Complex which consists mostly of gabbro and older multiphased diorite to quartz diorite. Within this admixture of basic intrusives are layers, lenses, sills and dykes of melanocratic gabbro to ultramafic rock as well as remnants and xenoliths of mafic volcanics, volcaniclastics and sediments. A quartz

Figure 9 Legend



diorite phase of the Laonil Lake Intrusive Complex has yielded a U-Pb zircon date of 1889 ±9 Ma (Chiarenzelli, 1989). On its southern side, the Laonil Lake Shear Zone separates the complex from the Eyahpaise Lake Pluton, a medium-grained leucocratic tonalite which has been dated at 1859 ± 5 Ma (U-Pb zircon age). On the north side of the complex, the Pine Lake Metavolcanics comprise a basal sequence of mafic to intermediate flows and volcaniclastics that are succeeded by felsic volcanics, fragmentals and coarse to fine sediments. Although a strong contact foliation masks relationships between the Laonil Lake Intrusive Complex and the supracrustal rocks, Delaney (1986) has identified a number of factors which support an unconformable contact.

Southeast of Laonil Lake, at Ourom Lake on the Churchill River, the Ourom Lake meta-arkoses are intercalated within the Pine Lake Metavolcanics. Single detrital zircon grains from these arkoses have yielded U-Pb ages between 1850-1863 Ma (Delaney et al., 1988). Separated by a zone of high strain on the northwest side of the Pine Lake Metavolcanics are the Porky Lake siliclastics, a thick succession of arkose, conglomerate and wacke exposed in the core of a synform. North of Pine Lake, the Pine Lake Metavolcanics have been intruded by granodiorite. Although at least three episodes of deformation can be distinguished in the metavolcanics, expressions of these events are poorly preserved in the Laonil Lake Intrusive Complex.

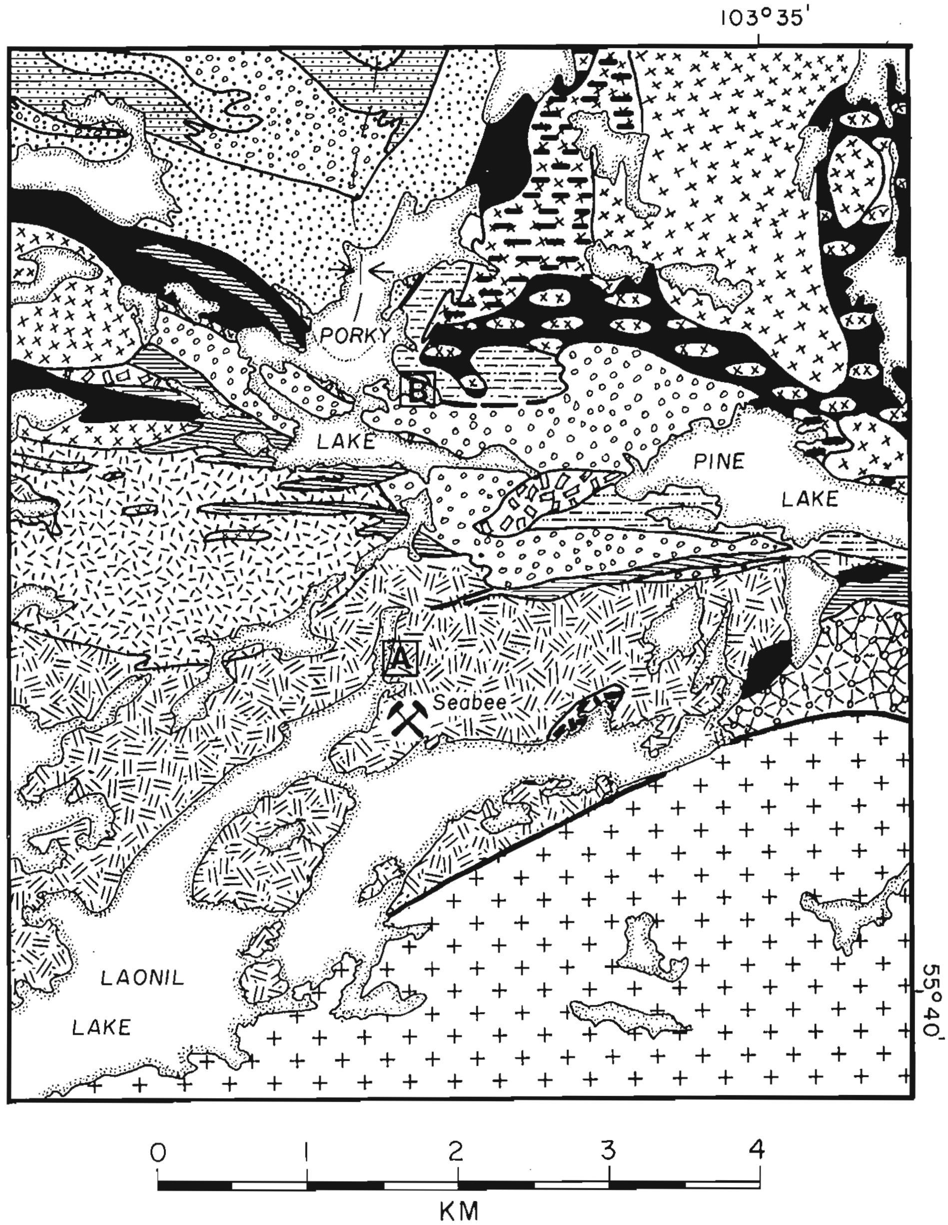


Figure 9. Geological sketch map of the Laonil Lakes area. Letters in boxes show locations of geochronology samples: A = Laonil Lake Intrusive Complex; B = Pine Lake Meta-rhyolite.

Pine Lake meta-rhyolite (sample 9022-92)

This sample was collected from an outcrop of buff rubbly soda rhyolite along the shore of Porky Lake (B, Fig. 9); a photo of this outcrop is included as Plate 4 in Lewry (1977). The zircon population in this sample consisted of tiny (<80 µm) colourless to light pink prisms. All four zircon fractions analyzed from this sample are colinear and less than 1% discordant with ^{207}Pb-^{206}Pb ages that fall between the narrow range 1837-1839 Ma (Table 1, Fig. 10). The best-fit regression line constructed to pass through all four analyses defines an upper intercept of 1838 ± 2 Ma and a lower intercept age near 0 Ma. The more discordant analysis (fraction 4) consisted of a more magnetic fraction that contained some fractures. The 1838 Ma age is interpreted as the time of felsic volcanism in the Pine Lake Greenstone Belt.

Table 1. U-Pb Analytical Data for the Brownell Lake and Laonil Lake map areas, Saskatchewan

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)
8822-375, Carroll Lake granodiorite													
T-1,cl,na	0.216	44.73	19.12	471	438	0.407	0.3266 ± 0.13	1821.9 ± 4.0	5.035 ± 0.24	1825.2 ± 4.1	0.1118 ± 0.18	0.72	1829.1 ± 6.4
T-2,cl,na	0.264	45.08	19.09	599	416	0.411	0.3228 ± 0.11	1803.6 ± 3.4	4.979 ± 0.20	1815.8 ± 3.3	0.1119 ± 0.14	0.74	1829.9 ± 4.9
B,125,c,eu	0.006	411.2	134.7	4615	11	0.094	0.3158 ± 0.08	1769.2 ± 2.6	4.882 ± 0.10	1799.2 ± 1.6	0.1121 ± 0.03	0.95	1834.1 ± 1.1
C,125,c,eu	0.001	103.3	33.01	262	10	0.036	0.3237 ± 0.36	1807.9 ± 11.2	5.002 ± 0.38	1833.3 ± 4.0	0.1121 ± 0.11	0.96	1833.3 ± 4.0
8822-376, Wapawekka Lake Formation meta-arkose													
A,e,cl,sf	0.007	124.9	43.21	515	38	0.092	0.3333 ± 0.12	1854.4 ± 3.8	5.235 ± 0.22	1858.3 ± 3.8	0.1139 ± 0.16	0.72	1862.7 ± 5.8
B,e,cl,sr	0.009	79.80	26.43	1123	14	0.061	0.3277 ± 0.11	1827.1 ± 3.5	5.160 ± 0.13	1846.1 ± 2.2	0.1142 ± 0.05	0.93	1867.6 ± 1.8
C,e,cl,sr	0.008	66.33	22.84	984	11	0.088	0.3333 ± 0.14	1854.3 ± 4.3	5.215 ± 0.15	1855.0 ± 2.6	0.1135 ± 0.05	0.94	1855.9 ± 1.9
D,e,cl,sf	0.007	161.1	55.53	168	152	0.094	0.3318 ± 0.21	1847.2 ± 6.7	5.179 ± 0.75	1849.2 ± 12.7	0.1132 ± 0.63	0.68	1851.5 ± 22.8
E,e,cl,sf	0.007	117.6	38.44	1733	10	0.061	0.3234 ± 0.11	1806.4 ± 3.5	5.095 ± 0.12	1835.3 ± 2.1	0.1143 ± 0.05	0.92	1868.3 ± 1.8
F,e,cl,sr	0.007	126.8	43.20	1942	9	0.079	0.3320 ± 0.10	1848.1 ± 3.3	5.217 ± 0.12	1855.4 ± 2.1	0.1140 ± 0.05	0.91	1863.6 ± 1.9
G,e,cl,sr	0.006	190.1	66.21	3558	6	0.109	0.3313 ± 0.14	1844.5 ± 4.3	5.159 ± 0.14	1846.0 ± 2.4	0.1130 ± 0.04	0.96	1847.6 ± 1.4
H,e,cl,sf	0.010	98.80	33.08	1607	13	0.034	0.3389 ± 0.11	1881.2 ± 3.5	5.399 ± 0.12	1884.8 ± 2.1	0.1156 ± 0.05	0.92	1888.7 ± 1.7
I,el,cl,eu	0.004	356.0	119.7	1300	22	0.061	0.3327 ± 0.10	1851.3 ± 3.3	5.262 ± 0.13	1862.8 ± 2.2	0.1147 ± 0.06	0.88	1875.6 ± 2.2
J,el,cl,eu	0.008	165.4	52.74	2150	12	0.073	0.3124 ± 0.10	1752.5 ± 3.2	4.898 ± 0.12	1801.9 ± 1.9	0.1137 ± 0.04	0.93	1859.5 ± 1.5
9022-94, Maynard Creek feldspar porphyry													
1,c,fr	0.009	544	158	6912	15	0.021	0.2984 ± 0.15	1683.4 ± 5.1	4.607 ± 0.14	1750.6 ± 4.9	0.1120 ± 0.07	0.95	1831.8 ± 2.6
2,c,fr	0.024	488	139	4906	46	0.017	0.2930 ± 0.13	1656.3 ± 4.2	4.511 ± 0.13	1733.0 ± 4.5	0.1120 ± 0.03	0.95	1826.9 ± 1.1
T	0.872	40	14	662	1110	0.131	0.3269 ± 0.60	1823.4 ± 21.9	5.056 ± 0.64	1828.7 ± 23.4	0.1122 ± 0.20	0.70	1834.7 ± 7.3
4,cl	0.006	71	23	4076	4	0.063	0.3248 ± 0.17	1813.3 ± 6.2	5.106 ± 0.16	1837.0 ± 5.9	0.1140 ± 0.06	0.95	1864.0 ± 2.2
5,c,fr	0.062	617	172	1686	413	0.016	0.2869 ± 0.10	1625.8 ± 3.3	4.418 ± 0.10	1715.7 ± 3.4	0.1117 ± 0.03	0.73	1827.3 ± 1.1
8822-377, Brownell Lake Pluton porphyritic granite													
A,el,c,eu	0.033	1508	474.3	24980	39	0.047	0.3162 ± 0.09	1770.9 ± 2.7	4.831 ± 0.10	1790.2 ± 1.7	0.1108 ± 0.03	0.96	1812.8 ± 1.0
B,el,c,eu	0.026	1876	575.9	7526	124	0.048	0.3082 ± 0.10	1731.8 ± 2.9	4.676 ± 0.11	1763.0 ± 1.8	0.1100 ± 0.03	0.96	1800.2 ± 1.1
C,el,c,eu	0.018	2293	690.6	13200	61	0.049	0.3024 ± 0.09	1703.2 ± 2.7	4.570 ± 0.10	1743.8 ± 1.7	0.1096 ± 0.03	0.96	1792.8 ± 1.0
D,el,c,eu	0.044	1710	514.9	26330	54	0.049	0.3024 ± 0.10	1703.4 ± 2.9	4.562 ± 0.11	1742.4 ± 1.8	0.1094 ± 0.03	0.97	1789.5 ± 1.0
T-1,cl,na	0.060	88.87	30.89	390	287	0.155	0.3190 ± 0.15	1784.9 ± 4.7	4.887 ± 0.26	1800.1 ± 4.4	0.1111 ± 0.18	0.73	1817.7 ± 6.7
T-2,cl,na	0.057	82.71	29.40	397	259	0.137	0.3310 ± 0.20	1843.4 ± 6.3	5.082 ± 0.29	1833.0 ± 5.0	0.1113 ± 0.18	0.81	1821.3 ± 6.5
8822-378, Maynard Creek Pluton granodiorite													
A,el,cl,eu	0.007	168.9	59.38	1260	18	0.144	0.3252 ± 0.10	1815.3 ± 3.2	5.019 ± 0.12	1822.5 ± 2.1	0.1119 ± 0.06	0.87	1830.8 ± 2.2
B,el,cl,eu	0.006	160.2	53.20	1479	13	0.102	0.3182 ± 0.11	1780.7 ± 3.3	4.904 ± 0.12	1802.9 ± 2.0	0.1118 ± 0.05	0.91	1828.7 ± 1.8
C,el,cl,eu	0.009	95.13	28.32	1974	8	0.057	0.2967 ± 0.11	1674.9 ± 3.2	4.512 ± 0.12	1733.2 ± 1.9	0.1103 ± 0.05	0.90	1804.3 ± 1.9
D,el,cl,eu	0.007	420.7	142.1	6510	9	0.121	0.3187 ± 0.09	1783.5 ± 2.8	4.893 ± 0.10	1801.1 ± 1.7	0.1113 ± 0.03	0.95	1821.5 ± 1.1
T-1,cl,na	0.146	124.6	53.37	582	677	0.374	0.3346 ± 0.12	1860.6 ± 3.8	5.154 ± 0.22	1845.1 ± 3.8	0.1117 ± 0.16	0.74	1827.7 ± 5.7
T-2,cl,na	0.096	145.4	52.40	688	429	0.171	0.3266 ± 0.10	1821.7 ± 3.1	5.023 ± 0.18	1823.2 ± 3.0	0.1116 ± 0.12	0.73	1824.9 ± 4.5
9022-92, Pine Lake meta-rhyolite													
1,<80,eu	0.024	128	45	40560	6	0.139	0.3282 ± 0.14	1829.4 ± 5.1	5.081 ± 0.13	1832.6 ± 4.8	0.1123 ± 0.06	0.95	1837.0 ± 2.2
2,<80,eu	0.007	574	198	549179	5	0.110	0.3284 ± 0.14	1830.5 ± 5.1	5.086 ± 0.12	1833.7 ± 4.4	0.1123 ± 0.07	0.95	1837.7 ± 2.7
3,<80,eu	0.011	242	85	156130	3	0.128	0.3275 ± 0.52	1826.2 ± 19.0	5.075 ± 0.51	1831.9 ± 18.7	0.1124 ± 0.06	0.95	1838.5 ± 2.1
4,<80,eu	0.027	167	58	45167	7	0.128	0.3259 ± 0.16	1818.7 ± 5.8	5.054 ± 0.16	1828.4 ± 5.9	0.1125 ± 0.04	0.95	1839.4 ± 1.4

** 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, fr = fractures, eu = euhedral, sf = subfaceted, sr = subrounded

Weighing error = 0.001 mg.

\+ Radiogenic Pb.

* Measured ratio, corrected for spike, and Pb fractionation of 0.09% +/- 0.03%/AMU; for samples 9022-92 and 9022-94 also corrected for blank Pb

\# 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 in when expressed in Ma.

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

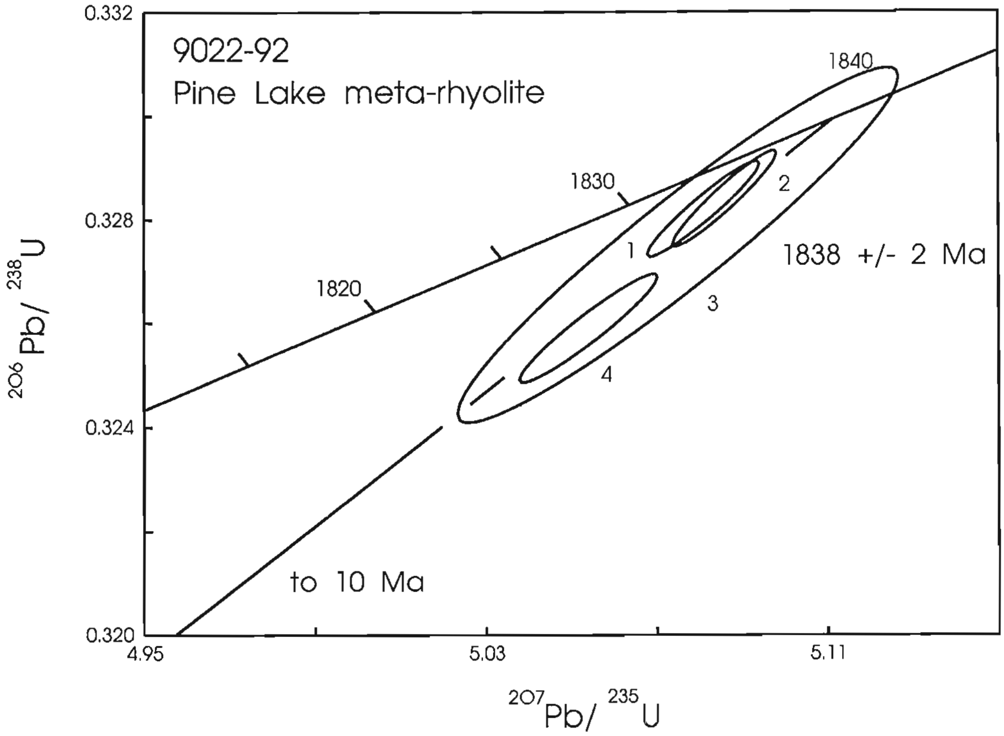


Figure 10.

U-Pb concordia diagram of zircons from sample 9022-92, Pine Lake meta-rhyolite at Porky Lake. Error ellipses are 2σ.

DISCUSSION

Ages of emplacement of granitoid bodies constrain the timing of deformation in the Brownell Lake area. The Carroll Lake granodiorite, interpreted to be 1834±3 Ma in age, is deformed by major D2 folding into an east-southeast- trending overturned antiform, thereby constraining the second phase of deformation to be younger than 1834 ±3 Ma.

The Brownell Lake Pluton and the Maynard Creek Pluton have been dated at 1831 ± 9 Ma and 1832 +9/-3 Ma respectively. Both of these plutons contain structures formed during D3 deformation and may have been intruded during this deformation event (Padgham, 1968). Thus the timing of the third phase of deformation is constrained to be approximately 1831 Ma or younger. The Brownell Lake and Maynard Creek plutons cross-cut D2 fabrics within the Brownell Lake Greenstone Belt. The second phase of deformation is therefore likely to have occurred between 1837 and 1829 Ma (the value of 1837 Ma is obtained from the maximum age of the late granodiorite of the Carroll Lake Block and 1829 Ma is obtained from the minimum age of the Maynard Creek Pluton).

Mesothermal gold mineralization in the Brownell Lake area occurs in quartz veins in the margins of and within country rocks adjacent to the Brownell Lake Pluton; in shear zones in the dioritic phase of the Maynard Creek Pluton; and in high strain zones within volcanics of the Brownell Lake Group that were reactivated during D3 deformation. On the basis of age determinations reported in this paper, gold mineralization is younger than approximately 1831 Ma, the age of the Brownell Lake Pluton. This relationship is similar to one in the Flin Flon Domain where mesothermal gold mineralization post-dates 1834 Ma (Ansdell and Kyser, 1991).

The Wapawekka Lake Formation meta-arkose is younger than about 1.85 Ga, as zircons of this age occur as detrital grains within the metasediment. The zircons have a fairly closely defined age span, with most of the grains in the 1.85-1.87 Ga range. These dates correspond to ages obtained for episodes of plutonism and volcanism elsewhere in the Glennie Domain (Fig. 2). Feldspar porphyry that intrudes the Wapawekka Lake Formation is dated at 1834 Ma. Therefore deposition of these sediments is tightly constrained between 1848 and 1834 Ma.

Elsewhere in the Glennie Domain, meta-arkose of the Ourom Lake Formation, which is intercalated with the Pine Lake Metavolcanics, contains detrital zircons that range in age from 1863-1850 Ma (Delaney et al., 1988). Deposition of the arkoses was interpreted to have occurred at 1850 Ma on the basis of an intrusive relationship between the 1850 Ma Wood Lake Batholith and rocks of the Pine Lake Metavolcanics. Subsequent re-examination of an apparent intrusive contact revealed that this relationship is equivocal. Felsic volcanics from the upper part of the Pine Lake Metavolcanics in the Porky Lake area have been dated at 1838 ±2 Ma (this paper), and may represent a more realistic estimate for the minimum age of deposition for arkoses of the Ourom Lake Formation.

Detrital zircons from arkoses of the Missi Group (Stauffer, 1990) within the Flin Flon Domain of the Trans Hudson Orogen were dated using the single zircon Pb-evaporation technique and had ^{207}Pb-^{206}Pb ages between 2092 ±38 Ma and 1869 ±9 Ma (Ansdell et al., 1991). The time span of Missi sedimentation may have varied: in the Flin Flon area, intrusive relationships suggest that sedimentation was completed by 1840 Ma (Ansdell and Kyser, 1991); whereas, farther to the east a rhyolitic flow intercalated with rocks correlated with the Missi Formation has been dated at 1832 ± 2 Ma (Gordon et al., 1990). Although the provenance of the sediments comprising the Missi Formation was probably different, sedimentation occurred during a similar period for arkosic sediments of the Wapawekka Lake Formation and the Ourom Lake Formation arkoses which may have implications for the concurrence of Lower Proterozoic molasse sedimentation throughout the Trans-Hudson Orogen (Delaney et al., 1988).

The age of 1838 Ma for the rhyolite of the Pine Lake Metavolcanics at Porky Lake is similar to an age of 1837 ± 7 Ma that Chiarenzelli (1989) obtained from dioritic gneiss, interpreted to be an andesitic flow or shallow intrusive, collected in the Nistowiak Lake area. The date provides additional evidence of at least three episodes of volcanism in the Glennie Domain which occurred during the periods 1889-1881 Ma, ca. 1866 Ma, and 1837-1838 Ma (Fig. 2).

There were two main episodes of Proterozoic plutonism in the Glennie Domain. A major phase of tonalitic to granodioritic plutonism during the period 1846-1859 Ma formed many of the major intrusions. A second episode of granodioritic to granitic plutonism, during the period 1837 to 1831 Ma, formed small intrusions that are apparently more common in the southern part of the domain. Plutonism in the Glennie Domain occurred during a similar time span to that of the Flin Flon Domain (Ansdell and Kyser, 1991).

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APPENDIX 1

SAMPLE LOCATIONS AND DESCRIPTIONS

8822-375: granodiorite, Carroll Lake Block; southern shore of unnamed lake located north of eastern arm of Brownell Lake, Saskatchewan (63M/4); 103°41.49'W-55°02.17'N; UTM 13u, 583575E-6099625N.

8822-376: meta-arkose, Wapawekka Lake Formation; southern shore of Brownell Lake, Saskatchewan (63L/13); 103°50.07'W-54°59.34'N; UTM 13u, 574275E-6093825N.

9022-94: plagioclase porphyry, southern shore of most eastern arm of Maynard Lake, Brownell Lake area, Saskatchewan (63M/4); 103°54'W-55°00.40'N; UTM 13u, 570320E-6095735N.

8822-377: granite, Brownell Lake Pluton; southern shore of Upper Lake "29" (Delaney, 1988), Brownell Lake area, Saskatchewan (63L/13); 103°43.24'W-54°58.55'N; UTM 13u, 581825E-6092900N.

8822-378: granodiorite, Maynard Creek Pluton; southern shore of Maynard Lake at mouth of Maynard Creek, Brownell Lake area, Saskatchewan (63M/4); 103°54.05'W-55°00.40'N; UTM 13u, 570325E-6095735N.

9022-92: rhyolite, Pine Lake metavolcanics; eastern shore of Porky Lake, Laonil Lake area, Saskatchewan (63M/12); 103°35'W-55°40'N; UTM 13u, 589115E-6169680N.

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Nd-isotopic stratigraphy of Early Proterozoic Amisk Group metavolcanic rocks from the Flin Flon belt, Manitoba and Saskatchewan[1]

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Stern, R.A., Syme, E.C., Bailes, A.H., Galley, A.G., Thomas, D.J., and Lucas, S.B., 1992: Nd-isotopic stratigraphy of Early Proterozoic Amisk Group metavolcanic rocks from the Flin Flon belt, Manitoba and Saskatchewan; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 73-84.

Abstract

Initial ε_{Nd} values of Amisk Group metavolcanic rocks from the Flin Flon and Snow Lake areas are much more variable than has previously been recognized, ranging from -0.4 to +4.8. The low ε_{Nd} values for certain volcanic horizons, such as the Vick Lake andesite tuff in the Flin Flon area and the Snell Lake basalt/andesite flows in the Snow Lake area, indicate the early involvement of older, probably Archean, light REE-enriched lithosphere. Mystic Lake granodiorite and a tonalite from Athapapuskow Lake have ε_{Nd} (1906 Ma) values of -6.1 and -3.9, respectively, indicating their derivation from largely Archean crustal sources. The arc volcanic rocks from the Snow Lake area have lower ε_{Nd} values than those in the Flin Flon area, suggesting that the Snow Lake arc segment interacted to a greater degree with the older crust. Volcanogenic massive sulphide ores were deposited during episodes of isotopically primitive silicic magmatism. This suggests a coupling between ore deposition and the input of mantle-derived heat or fluids.

Résumé

Les valeurs initiales de ε_{Nd} caractérisant les métavolcanites du Groupe d'Amisk observées dans les régions de Flin Flon et de Snow Lake sont beaucoup plus variables qu'on ne l'avait établi; elles s'échelonnent de - 0,4 à + 4,8. Les faibles valeurs de ε_{Nd} obtenus dans le cas de certains horizons volcaniques, comme le tuf andésitique du lac Vick dans la région de Flin Flon et les coulées de basalte-andésite du lac Snell dans la région de Snow Lake, indiquent un apport précoce de la lithosphère plus ancienne (probablement archéenne) riche en ETR légers. La granodiorite du lac Mystic et une tonalite du lac Athapapuskow donnent des valeurs de ε_{Nd} (1 906 Ma) de - 6,1 et de - 3,9, respectivement, indiquant qu'elles dérivent de sources crustales en grande partie archéennes. Les roches d'arc volcanique de la région de Snow Lake présentent des valeurs de ε_{Nd} plus faibles que celle de la région de Flin Flon, laissant supposer que l'intercation entre le segment d'arc de Snow Lake et la croûte plus ancienne a été plus importante. Les sulfures massifs d'origine volcanique se sont mis en place durant des épisodes d'intrusion de magma sursaturé isotopiquement primitif, ce qui indique un lien entre la minéralisation et l'apport de chaleur ou de fluides provenant du manteau.

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INTRODUCTION

The following article presents the interim results of Nd isotopic studies carried out in support of the NATMAP Shield Margin Project (Lucas et al., 1992). The overall goal of the isotopic study is to place new constraints on the geological evolution and tectonic setting of Lower Proterozoic rocks within the NATMAP project area. The initial emphasis of the study has been on the characterization of Amisk Group metavolcanic rocks within the Flin Flon belt. Subsequent studies will be extended into the Kisseynew gneiss belt and the Paleozoic subsurface. To aid in the discussion that follows, the terms Flin Flon segment (west) and Snow Lake segment (east) are used to refer to the rocks within specific regions of the larger Flin Flon belt (Fig. 1).

PREVIOUS WORK

Prior to this study, there were 14 published Nd isotopic determinations for Amisk Group metavolcanic rocks from the Flin Flon belt (Chauvel et al., 1987; Thom et al., 1990). These previous studies reported relatively juvenile isotopic signatures for the analyzed rocks, with initial ε_{Nd} values of +3.2 to +4.8. The isotopic data, in combination with the geochemical data which showed a high proportion of tholeiitic vs. calc-alkaline rocks (Parslow and Gaskarth, 1984; Gaskarth and Parslow, 1987; Syme, 1990; Thom et al., 1990; Watters, 1991), were consistent with an intraoceanic arc/back-arc setting for the metavolcanic rocks (Gaskarth and Parslow, 1987; Syme, 1990; Thom et al., 1990).

Compared to the Flin Flon belt, the metavolcanic rocks from the La Ronge-Lynn Lake Domain, about 200 km to the northwest and north, have a higher proportion of calc-alkaline rocks, and are interpreted as being evolved island arc or continental arc assemblages (Watters and Pearce, 1987; Syme, 1990; Thom et al. 1990). Indeed, the available Nd isotopic data for those belts (Chauvel et al., 1987; Hegner et al., 1989; Thom et al., 1990), showed initial ε_{Nd} values extending to lower numbers (-0.6 to +4.9) than the Flin Flon rocks, implying greater proximity to, and involvement of, Archean crust. Note that Archean upper crust at 1.9 Ga would have strongly negative ε_{Nd} values (e.g. -5 to -10).

GEOLOGICAL AND GEOCHRONOLOGICAL CONTEXT

Rock powders used in this study were obtained from areas where detailed geological mapping has recently been completed: the Flin Flon-White Lake (Bailes and Syme, 1989), Athapapuskow Lake (Syme, 1987, 1988), Bootleg-Birch Lakes (Thomas, 1991), and Chisel-Morgan-Anderson Lakes (Bailes, 1990b) map sheets.

The Flin Flon belt is composed of (Fig. 1): 1. sub-greenschist- to amphibolite-grade subaqueous and lesser subaerial volcanic rocks and associated sediments (Amisk Group), 2. an unconformably overlying sequence of terrestrial sedimentary and minor metamorphosed volcanic rocks (Missi Group), and 3. a complex suite of mafic to felsic plutons ranging from syn-Amisk to post-Missi. Numerous post-metamorphic faults divide the Flin Flon segment of the belt into discrete fault-bounded blocks (Fig. 2), each containing a distinct stratigraphic sequence, structural style, metamorphic grade, and suite of intrusions (Bailes and Syme, 1989).

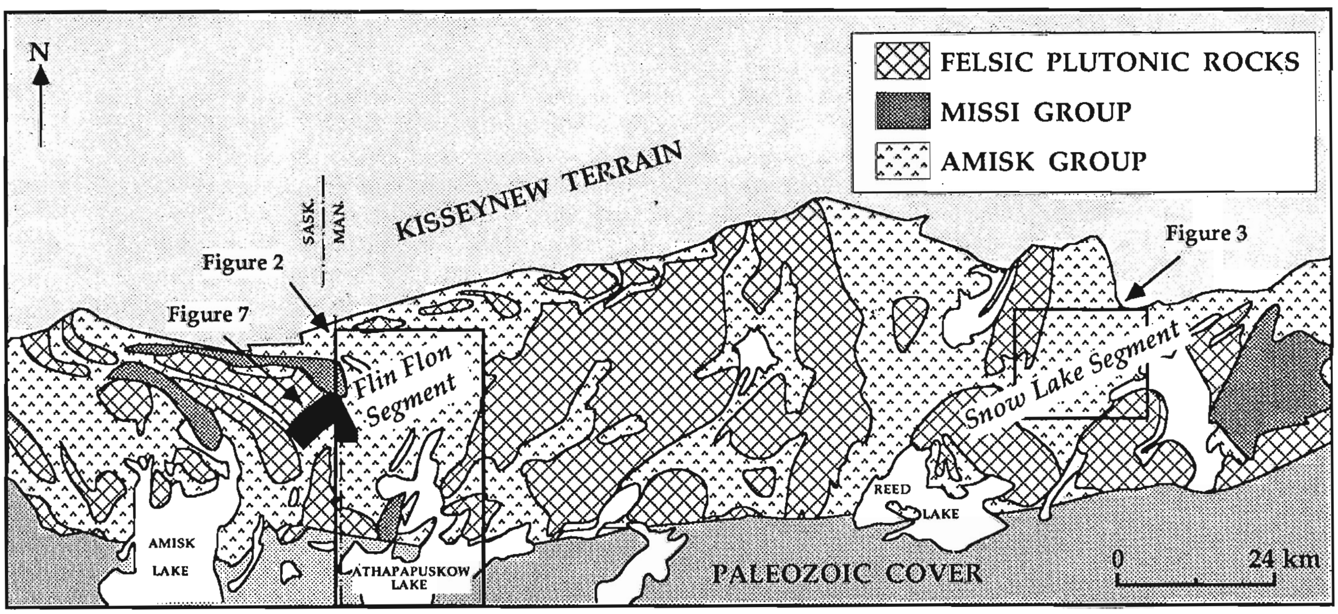

Figure 1. Location map showing the Flin Flon and Snow Lake segments of the Flin Flon metavolcanic belt. Open boxes show location of Figures 2 and 3. Filled region shows the general location of samples used to construct the Nd-isotopic profile of Fig. 7.

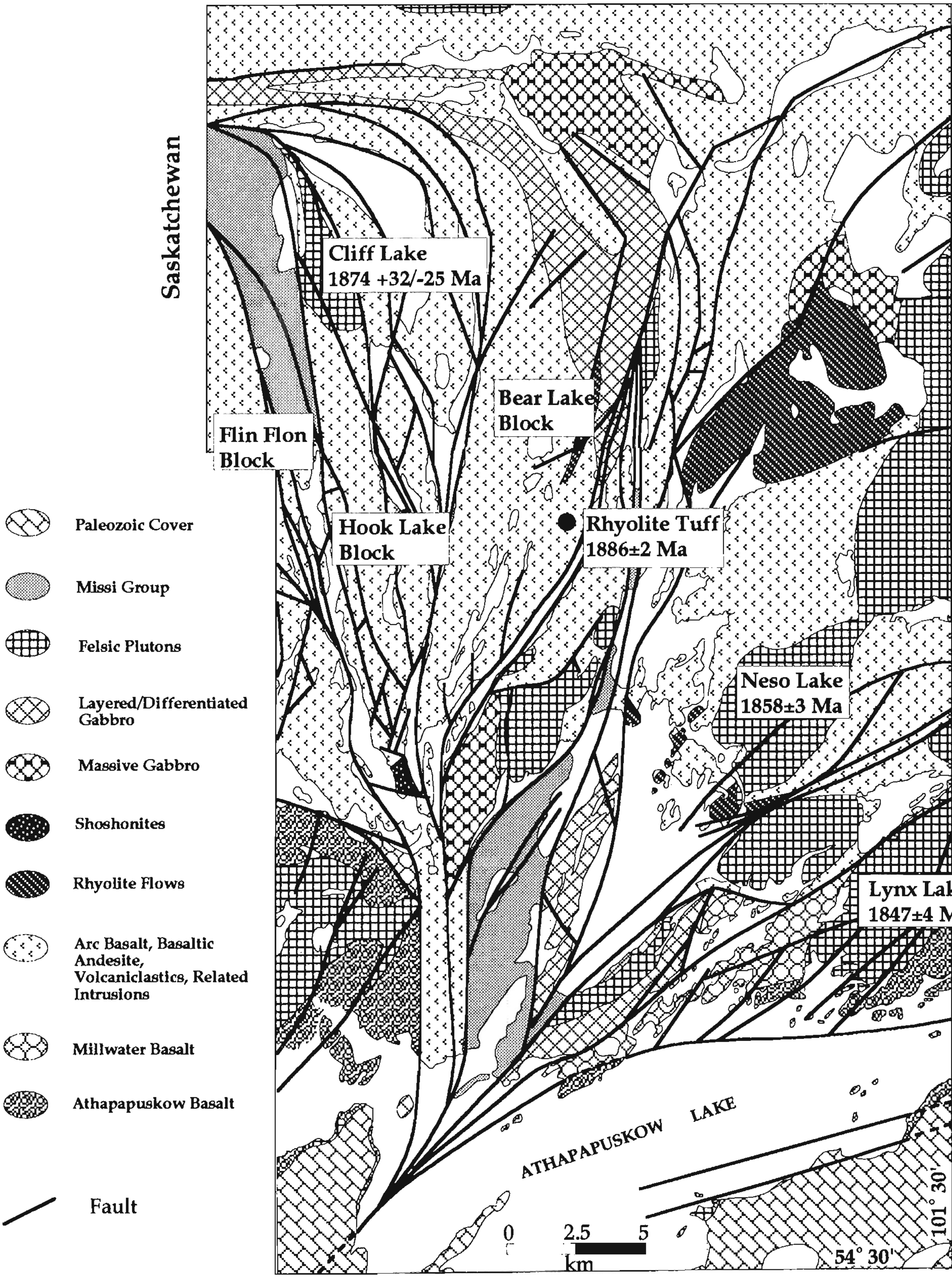

Figure 2. Geology of part of the Flin Flon metavolcanic belt in Manitoba, showing some of the fault-bounded blocks (after Bailes and Syme, 1989; Syme, 1988). References to age determinations are as follows: Lynx Lake pluton, Cliff Lake pluton, rhyolite tuff (Gordon et al., 1990); Neso Lake pluton (Syme et al., 1991).

The Amisk Group rocks within the Flin Flon segment can be subdivided into three distinct geochemical assemblages: arc-type, MORB-type, and shoshonitic (Gaskarth and Parslow, 1987; Syme, 1990). The arc-type assemblage is the most voluminous, and consists of tholeiitic and lesser calc-alkaline basalt to basaltic andesite flows, ferrobasalt flows, andesitic to dacitic tuffs, rhyolite flows, and numerous synvolcanic gabbroic intrusions. The rocks are interpreted as arc-type due to their depletion of high field strength (HFS) elements (P, Hf, Zr, Ti, Y) and enrichment in large ion lithophile (LIL) elements (Rb, Ba, K, Th, Sr) relative to typical mid-ocean ridge basalt (MORB) (Syme, 1990). The Bear Lake, Hook Lake, Flin Flon, and Bakers Narrows fault-bounded blocks (Fig. 2) contain arc assemblages.

The MORB-type assemblage, in fault contact with the arc assemblage, is represented by basaltic flows in the Athapapuskow Lake area ('Athapapuskow basalt'; Fig 2; Syme, 1987), which show little or no high field strength element depletion and minor enrichment in large ion lithophile elements relative to MORB (Syme, 1988, 1990). The overall trace element patterns resemble those of modern back-arc basalts. The Athapapuskow basalts are themselves in fault contact with basaltic flows having chemical characteristics transitional between the Athapapuskow basalt and the arc-type basalt, termed 'Millwater basalt' (Fig. 2; Syme, 1987). For the purposes of discussion, the transitional Millwater basalts have been included within the MORB-type assemblage.

The shoshonitic assemblage, a relatively minor component of the Flin Flon segment, is represented by high-K, intermediate-silica clasts from a volcanic conglomerate of the Hook Lake block (Fig. 2; Syme, 1988), and by the Neso Lake pyroxenite through tonalite pluton (Syme, 1987).

In the Snow Lake segment, where there are fewer late faults, a relatively intact volcanic stratigraphy measuring about 3 km thick has been established near Chisel Lake (Fig. 3), referred to as the Chisel Lake section (Bailes, 1988, 1990a; Bailes and Galley, 1989). Geochemical characterization suggests that the rocks of the Chisel Lake section represent an arc-type assemblage (Bailes, 1988).

The best age constraints on arc-related volcanism in the Flin Flon belt are 1886 ± 2 Ma obtained on a rhyolite tuff from the Bear Lake block of the Flin Flon segment (Gordon et al., 1990), and 1892 ± 4 Ma for a felsic breccia from the Chisel Lake section in the Snow Lake segment (Machado and David, 1992). An age of 1858 ± 3 Ma was obtained on the Neso Lake pluton (Syme et al., 1991), which constrains the age of shoshonitic magmatism. The age of the MORB-type assemblage is presently unknown. For the calculation of ε_{Nd} values, 1890 Ma was adopted as the nominal crystallization age for the arc assemblage rocks, and 1900 Ma for the MORB assemblage. As will be discussed below, the lack of a precise age of the MORB-type assemblage will not change the interpretations that follow. The ages of other rocks and plutons are referenced in the text.

ANALYTICAL ASPECTS

Analytical procedures are similar to those described by Thériault (1990). Following standard treatment of Nd isotopic data, the initial $^{143}Nd/^{144}Nd$ ratios were calculated from the measured $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios and assumed crystallization ages. The initial $^{143}Nd/^{144}Nd$ ratios are expressed in ε_{Nd} units, which compare the $^{143}Nd/^{144}Nd$ ratio of the sample with the $^{143}Nd/^{144}Nd$ ratio of a chondritic uniform reservoir (CHUR) at the same age. Present day values for CHUR used for these calculations are $^{143}Nd/^{144}Nd=0.512638$ and $^{147}Sm/^{144}Nd= 0.1966$ (Jacobsen and Wasserburg, 1984).

A realistic assessment of the analytical uncertainty of the Nd data is the ability to replicate analyses of the same rock samples, which for this study is better than ±0.4 ε_{Nd} units when calculated at the crystallization age. This level of error takes into account possible uncertainties in the ages of crystallization, which are generally within ±10 Ma for the rocks being considered. A change in the absolute age of 10 Ma will increase or decrease the ε_{Nd} value by 0.1 unit or less.

Only the least epidotized and least silicified samples were analyzed. It is highly unlikely that any of the isotopic variations reported here are due to secondary alteration processes, such as epidotization or silicification. Studies of hydrothermally-altered and mylonitized rocks show that Nd and Sm are generally immobile (Farmer and DePaolo, 1987; Barovich and Patchett, 1992), and this is confirmed by experimental data which show the rare earth elements to be sparingly soluble in H_2O- or chloride-bearing fluids (Cullers et al., 1973; Flynn and Burnham, 1978; Wendtlandt and Harrison, 1979). Furthermore, the direct correlation of ε_{Nd} values with rock lithology (see below) demonstrates that the isotopic variations reflect those of the magmas.

ISOTOPIC RESULTS

Flin Flon Segment Overview

Figure 4 shows ε_{Nd} values vs. crystallization ages of all analyzed rocks from the Flin Flon segment in Manitoba and Saskatchewan. The rocks have been grouped according to the name of their fault-bounded blocks. The range of initial ε_{Nd} values of all analyzed rocks from five fault-bounded blocks is from +2.1 to +4.8, which is greater than the isotopic range previously reported in the literature (see above). The arc assemblage shows the greatest isotopic variability, with ε_{Nd} values from +2.1 to +4.8 (Fig. 4), whereas, the MORB assemblage (Athapapuskow and Millwater basalts) has a limited range of initial ε_{Nd} values of +3.1 to +3.9. The shoshonite assemblage also shows a relatively narrow range of initial ε_{Nd} values, from +3.7 to +4.6 (Fig. 4).

Within the arc assemblage there is no general relationship between initial ε_{Nd} and bulk compositional parameters such as SiO_2 content. For example, rhyolite flows from the Flin Flon block (Mine rhyolite, Myo Lake member) and Bear Lake block (Solodiuk rhyolite) have initial ε_{Nd} values from +4.1 to +4.6. Low ε_{Nd} values (down to +3.1) were measured for basaltic to basaltic andesite flows of the Hidden Lake-Burley Lake sequence and for plagioclase-phyric basalts of the Hook Lake block. However, the lowest values came from an andesitic pyroclastic unit called the Vick Lake tuff of the Bear Lake block. The initial ε_{Nd} value of this unit is $+2.3 \pm 0.2$, significantly lower than any other Amisk Group supracrustal

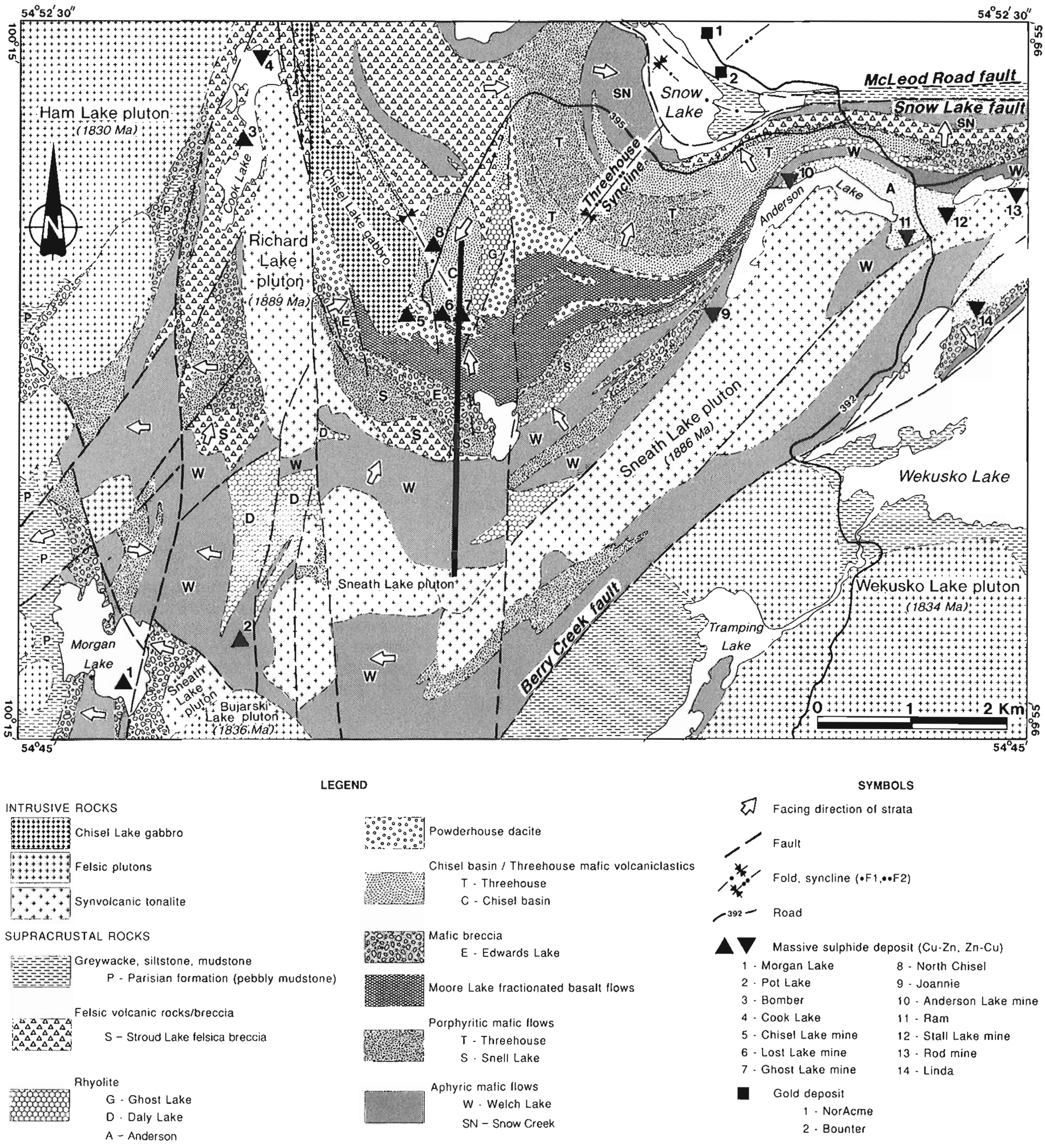

Figure 3. General geology of the Snow Lake area (from Bailes and Galley, 1991). Heavy line marks approximate location of profile shown in Figure 8.

rock from the Flin Flon segment (Fig. 4). The isotopic data require these rocks to have contributions from much older light REE-enriched lithosphere, such as Proterozoic crust >1900 Ma or Archean crust. This isotopic constraint combined with the alkaline affinities of the Vick Lake tuff (Bailes and Syme, 1989) indicate that parts of the Flin Flon arc may have interacted with Archean lithosphere at a relatively early stage. Three analyses of the Mikanagan Lake sill, which intrudes the Vick Lake tuff, yielded initial ε_{Nd} values of +2.5 to +3.6 (Fig. 4). That these values tend to be rather low and have a considerable range suggests that the parental magmas may have also interacted with older lithosphere or the Vick Lake tuff itself.

Figure 5 shows a histogram of initial ε_{Nd} values comparing the arc assemblage with the MORB assemblage. As the two assemblages have partly overlapping ε_{Nd} values, it would not be easy to distinguish them on the basis of the Nd isotopes alone. Nevertheless, a distinction of the arc assemblage is that it includes rocks with higher initial ε_{Nd} values. This suggests that the mantle sources to the arc and MORB-type assemblages may have been isolated from one another. The mantle source to the arc assemblage was probably oceanic lithosphere within a subduction zone mantle wedge, modified by the addition of fluids and melts. This supra-subduction zone mantle probably did not have a unique Nd isotopic composition, due to any original isotopic heterogeneities and isotopic variations caused by the addition of subduction-related fluids and melts. On the basis of Figure 5 it can be seen that the arc assemblage, and thus the supra-subduction mantle, was dominated by ε_{Nd} values between +3.75 and +4.25 at 1890 Ma. The mantle source to the MORB-type assemblage, on the other hand, had ε_{Nd} values between +3 and +4. Syme (1990) suggested that the MORB-type Athapapuskow basalts were similar to modern back-arc magmas, and therefore their source regions may

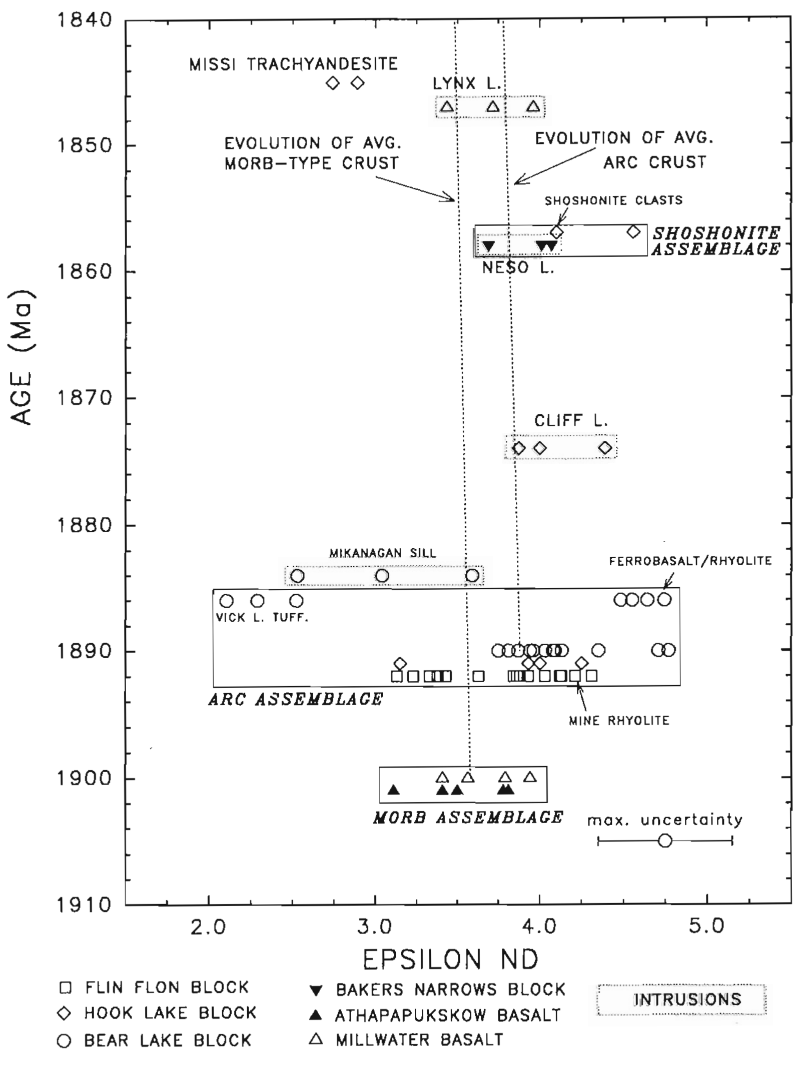

Figure 4.

Compilation of epsilon Nd values vs. age for metavolcanic and plutonic rocks from the Flin Flon area. Age of Missi trachyandesite intrusion based on data by Ansdell et al. (1991). Other references to age determinations in caption to Fig. 2 and in text.

have been the asthenosphere rather than supra-subduction zone lithosphere. In modern arcs there are also clear isotopic as well as trace element differences between arc and back-arc magmas (Gill, 1981). The slight large ion lithophile element enrichment in the Athapapuskow basalts and the presence of the 'transitional' Millwater basalts suggests, however, that this asthenospheric source region may also have been influenced by a subduction component.

Representatives of the shoshonite assemblage, the Neso Lake tonalite to pyroxenite pluton and shoshonite clasts from a volcanic conglomerate, have initial ε_{Nd} values of +3.7 to +4.6 (Fig. 4). These primitive values suggest negligible contributions from older crustal sources. These magmas may have been generated at 1858 Ma from large ion lithophile element-enriched source regions in the mantle that had been formed during 1890 Ma arc magmatism.

The other felsic plutons examined (Cliff Lake tonalite, Lynx Lake granodiorite) have initial ε_{Nd} values of +3.5 to +4.3, also showing that they were largely juvenile additions to the Flin Flon arc. In Figure 4, the plutons plot close to a Nd-isotope evolutionary line for the Flin Flon arc assemblage. Therefore, one permissible origin for the felsic plutons is partial melting of lower parts of the Flin Flon arc crust.

The youngest unit examined in the Flin Flon segment is a trachyandesite intrusion occurring within Missi metasedimentary rocks in the Athapapuskow Lake area. Its initial ε_{Nd} value of +2.7 indicates a small Archean lithospheric component (Fig. 4).

Detail of Bear Lake block, Flin Flon segment

As one of the largest fault-bounded blocks, and containing a well-preserved internal stratigraphy, the Bear Lake block has been studied in some detail. The results from this block also demonstrate well the coherence of the Nd isotopes with the observed geological stratigraphy.

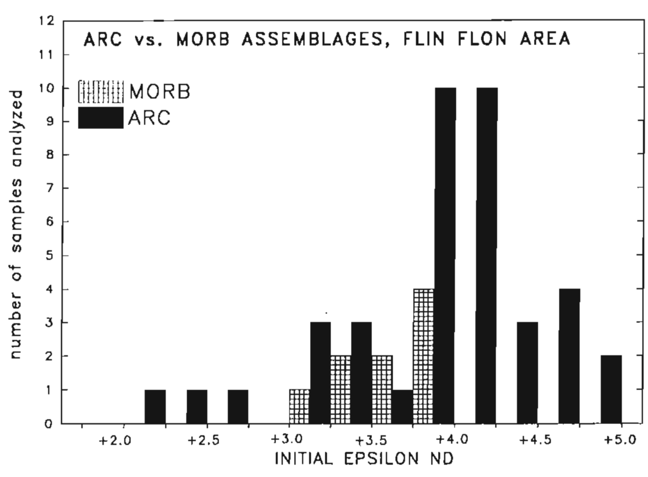

Figure 5. Histogram of initial ε_{Nd} values, comparing the MORB-type assemblage and the arc-type assemblage from the Flin Flon segment.

Figure 6 shows the Nd-isotopic data for the Bear Lake block, plotted against a simplified geological section. Two samples of subaqueous basaltic andesite (unit 1) from the lowest part of the section have ε_{Nd} values of about +4.8, whereas seven others have ε_{Nd} of +4.0 ± 0.2. The former have the highest Ni contents, which might indicate that they were the most primitive magmas among a series of magmas which assimilated older crust in amounts proportional to their extent of fractionation. Nevertheless, the possibility that the isotopic variation reflects original source heterogeneity cannot be excluded.

Solodiuk Lake rhyolite flows (unit 13) and dacitic to andesitic tuffs (units 26 and 24) have ε_{Nd} values of about +4, which are identical to the values for most of the underlying basaltic andesites (Fig. 6). Thus, the basaltic andesite flows through dacitic tuffs of the lower part of the Bear Lake block, except for two samples of basaltic andesite, may have been on the same liquid line of descent, possibly derived from the same volcanic centre.

The eruption of Two Portage ferrobasalt (unit 5) marks a slight, but distinct change to more positive ε_{Nd} values of ≈ +4.6 ± 0.1. A sample of rhyolite crystal tuff (unit 27) interbedded with ferrobasalt gives an identical value (Fig. 5), consistent with previous suggestions that the rhyolites may be differentiation products of ferrobasalt (Bailes and Syme, 1989).

Eruption of the pyroclastic Vick Lake andesite tuff (unit 25) marks an abrupt change in the isotopic composition of the erupted magmas (Fig. 6). The ε_{Nd} value of this unit is +2.3, significantly lower than earlier units. A dacitic flow (unit 18a) which overlies the Vick Lake tuff has the same value, +2.3. These rocks, with their high Na_2O, K_2O, Sr, and Ba abundances and relatively low ε_{Nd} values, must have been derived in part from older, light REE-enriched lithosphere.

Detail of Flin Flon block, Flin Flon segment

The Flin Flon fault-bounded block was studied in detail because it contains the Flin Flon mine, which is the largest Cu-Zn deposit in the Flin Flon belt. Figure 7 shows a simplified stratigraphy of the block (after Thomas, 1989; Bailes and Syme, 1989) and corresponding ε_{Nd} (1890 Ma) values for samples obtained from Saskatchewan and Manitoba. The stratigraphically lowest unit, called the Creighton member in Saskatchewan and South Main basalt in Manitoba, is dominated by mafic, pillowed flows and breccias. This unit is overlain by rhyolite flows and fragmental rocks of the Myo Lake member (equivalent to Mine rhyolite and South Main rhyolite of Manitoba) and contemporaneous massive sulphide deposits. The rhyolites are successively overlain by mafic flows of the Bomber Lake, Newcor, Hapnot Lake, and Louis Lake members, their probable equivalents in Manitoba being the Hidden Lake basalt and Burley Lake basaltic andesite sequence (Bailes and Syme, 1989). The stratigraphically youngest unit, called the Douglas Lake assemblage, consists of mafic to intermediate volcaniclastic rocks and minor mafic flows.

The ε_{Nd} values show limited variation throughout the section, ranging from +3.2 to +4.3. The highest measured values are for the rhyolites associated with the ore deposit. These values suggest little or no involvement of older lithosphere in the generation of the rhyolites. Stratigraphically higher units show a slight trend to progressively lower ε_{Nd} values, suggesting an increasing involvement of older lithosphere in younger magmas. It is important to note here that the isotopic profile for the Flin Flon block is quite distinct from the profile reported above for the Bear Lake block (compare Fig. 6 and 7). The implication is that the volcanic sequences within these two blocks are not correlative, a conclusion consistent with the geological interpretations (Bailes and Syme, 1989).

Chisel Lake section, Snow Lake segment

The Nd-isotope stratigraphy and the geological stratigraphy of the Chisel Lake section are presented in Figure 8. Three samples of Welch Lake aphyric basaltic andesite to andesite have identical ε_{Nd} values of +3.0, and one Welch Lake basalt sample has a distinctly lower value of +1.8. The cause of the isotopic variation of the Welch Lake units is uncertain, but it may be related to interaction with Archean crust (see below). A sample of Daly Lake rhyolite flows, approximately coeval with Welch basalt, yields an even higher ε_{Nd} value of +3.7, showing that the rhyolite is not directly related to Welch basalt via fractionation.

A sample from the lowermost part of the Stroud Lake felsic breccia has an ε_{Nd} value of +3.8, identical to the Daly Lake rhyolite. Three stratigraphically higher samples of the Stoud Lake breccia, including one which was collected only a few metres from its upper contact, yielded identical values of +2.8 ± 0.2. It has recently been reported that a sample of the Stroud Lake breccia collected from the same location as one of the latter three samples contains inherited zircons with ages of 2650 Ma, 2716 Ma, and 2824 Ma (Machado and David, 1992). Thus, a reasonable explanation for the lower and variable ε_{Nd} values in the Stroud Lake horizon is the differential incorporation of Archean crust with strongly negative ε_{Nd} values into the parental magmas to the rhyolite. By implication, other rocks with low ε_{Nd} values in the Chisel section, such as the underlying Welch basalts, may also have interacted with Archean crust.

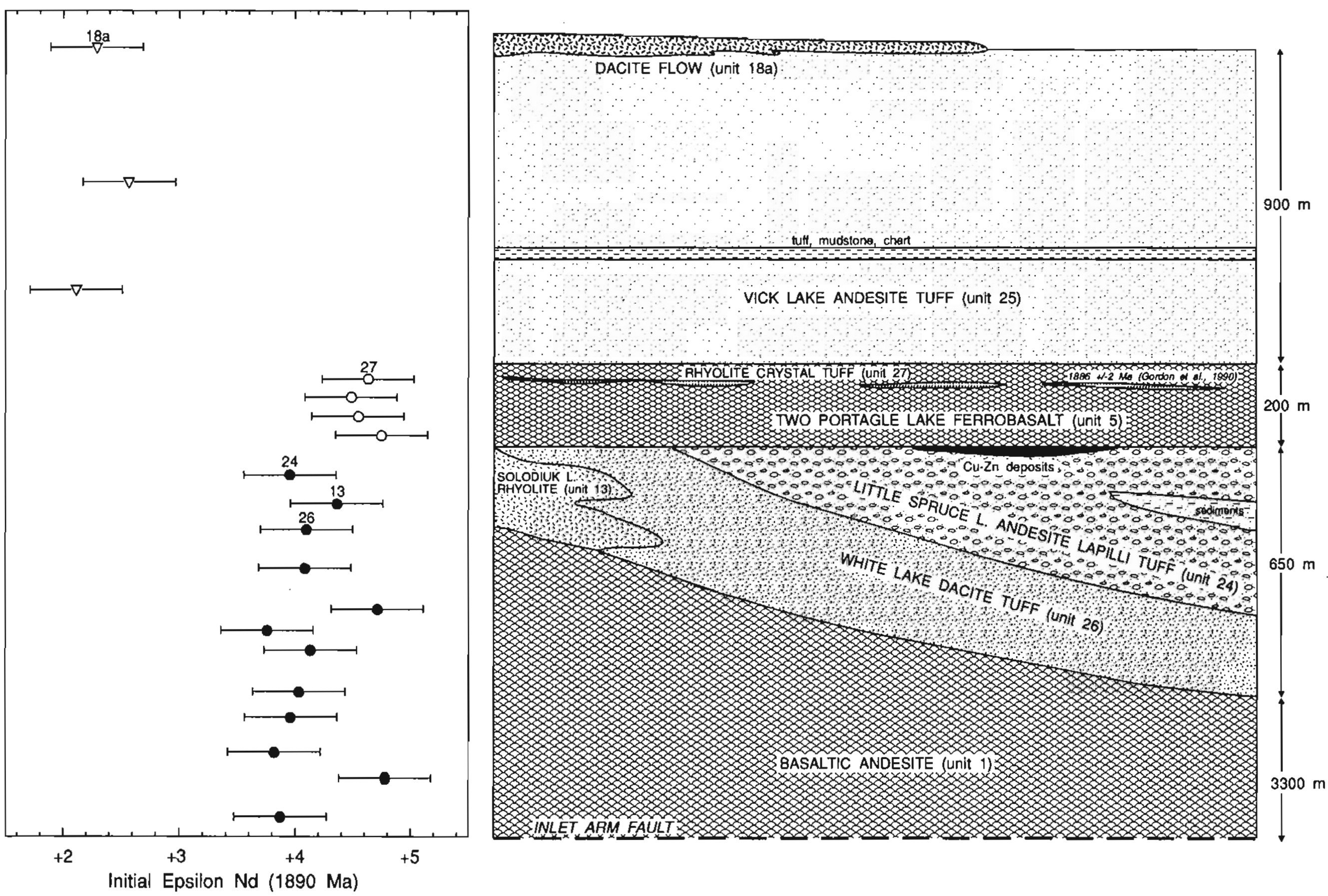

Figure 6. Detailed Nd-isotopic and geological stratigraphy of the Bear Lake block, Flin Flon segment (see Fig. 2). Geology summarized from Bailes and Syme (1989).

The overlying Snell Lake plagioclase-phyric basalt to andesite flows yielded conclusive isotopic evidence of the presence of older crust at depth. A sample of Snell basalt collected a few metres above the contact with the Stroud Lake breccia yielded an ε_{Nd} value of -0.4. Three other basaltic andesite to andesite samples from higher in the unit have an average ε_{Nd} value of -0.1 ± 0.1 (Fig. 8). These are the lowest epsilon Nd values measured to date for the Amisk Group metavolcanic rocks from the Flin Flon belt. They clearly show that these magmas interacted with much older light REE-enriched lithosphere. On the basis of the zircon data for the Stroud Lake horizon, the magmas have probably been contaminated by Archean crust. The effects of contamination on the Snell magmas were variable, as one sample of basalt gave an ε_{Nd} value of +2.4.

The units overlying the Snell horizon show a progressive return to isotopically primitive ε_{Nd} values, beginning with the Edwards Lake mafic volcanic wacke at +2.0, followed by the Moore Lake basalt flows at +2.5, Powderhouse dacite tuff at +2.8, and ending with the Ghost Lake rhyolite flow at +3.3 (Fig. 8). The progressive increase in ε_{Nd} values may reflect a gradual diminishing of the effects of crustal contamination. The Mine basalt and Chisel basin mafic tuff which immediately overlie the sulphide deposit have values of +2.6

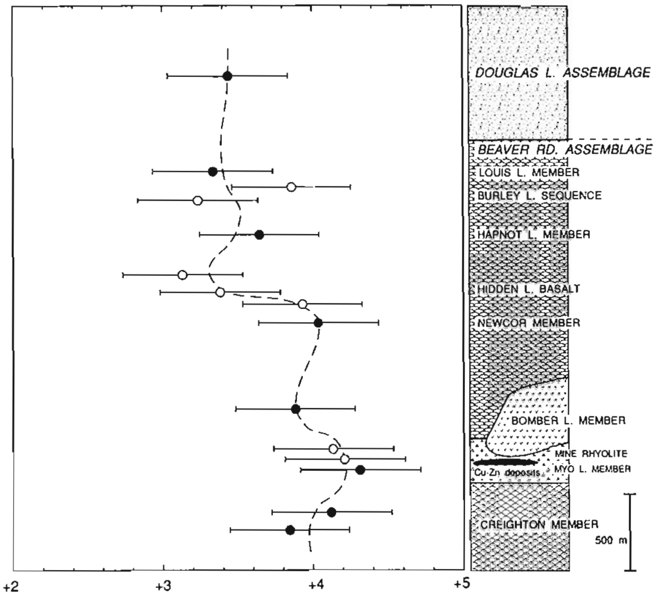

Figure 7. Isotopic and geological stratigraphy of the Flin Flon block. Geology and nomenclature summarized from Thomas (1989) and Bailes and Syme (1989).

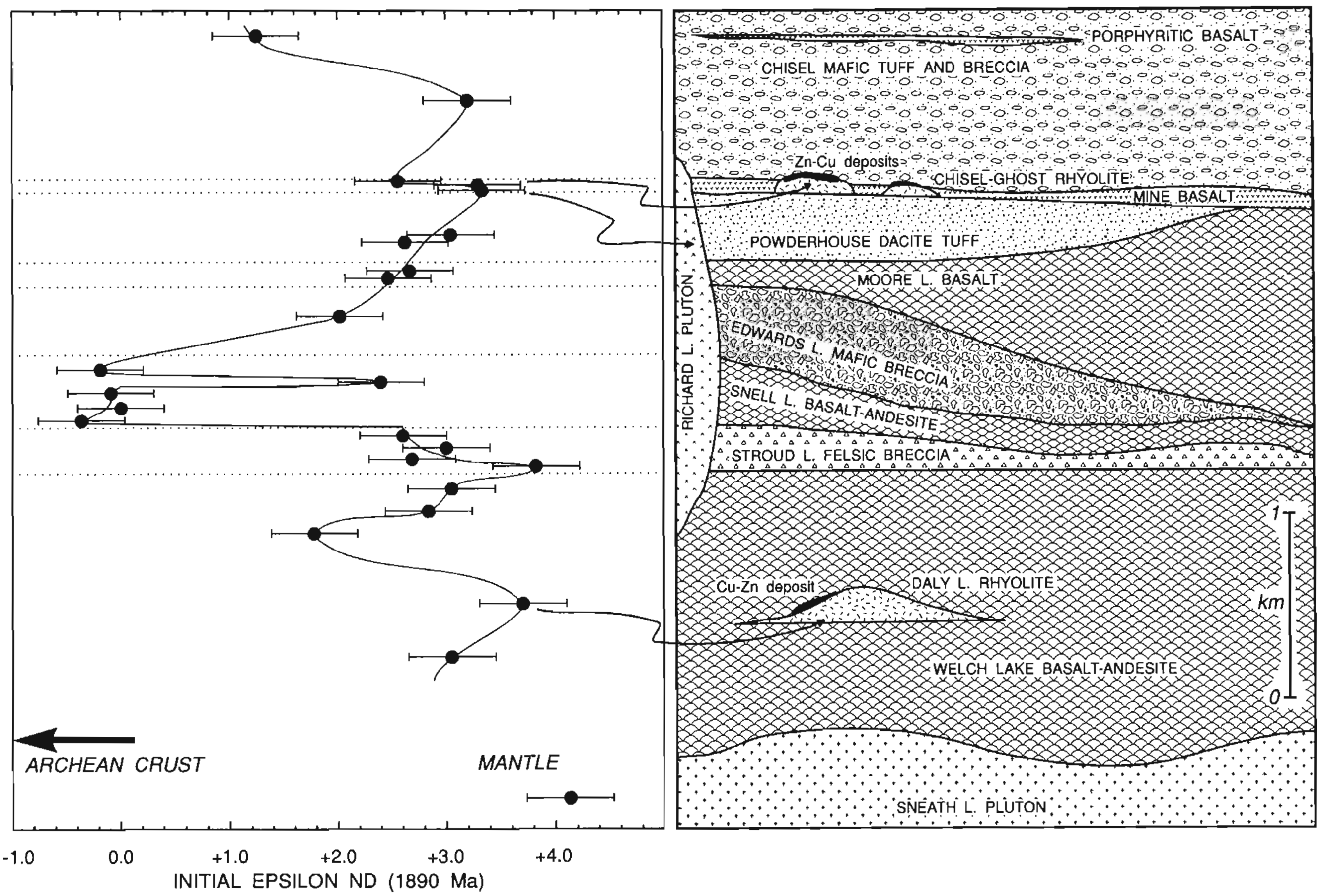

Figure 8. Isotopic and geological stratigraphy of the Chisel Lake section, Snow Lake segment. Geology summarized from Bailes (1990a) and Bailes and Galley (unpub. data). See Figure 3 for location.

and +3.2, respectively. The stratigraphically highest sample analyzed, a Chisel basin basalt flow, again shows evidence of interaction with Archean crust, having an ε_{Nd} value of +1.2.

METALLOGENESIS

The massive sulphide deposits in the Flin Flon and Snow Lake segments are stratigraphically associated with felsic volcanic rocks, some of which have been analyzed in this study. In the Flin Flon segment, the felsic volcanic rocks associated with ore deposits include the Little Spruce Lake andesite lapilli tuff (Bear Lake fault-bounded block; Fig. 6) and the South Main and Mine rhyolites (Flin Flon fault-bounded block; Fig. 7). These rocks all have similar initial ε_{Nd} values of $+4.0 \pm 0.3$. In the Snow Lake segment, the ore-related felsic rocks are the Chisel-Ghost Lakes rhyolite and the Daly Lake rhyolite flows (Fig. 8) which have initial ε_{Nd} values of +3.3 and +3.7, respectively. Thus, the rhyolitic magmas associated with the ores are among the most isotopically juvenile within their stratigraphic sequences, and were most likely on the liquid line of descent of mantle-derived magmas. Sulphide deposition occurred during or shortly after periods of mantle-derived felsic magmatism.

The isotopic data also provide further constraints on the plutonic activity that may have provided heat or fluids necessary for the formation of the sulphide deposits (Bailes and Galley, 1989). In the Snow Lake segment, the Richard Lake pluton may be a sub-volcanic equivalent to the Ghost Lake rhyolite on the basis of their identical initial ε_{Nd} values (both +3.3) and $^{147}Sm/^{144}Nd$ ratios (0.126 and 0.129, respectively), and also their similar whole-rock geochemistry (Bailes, 1988). Similar comparisons indicate that the Sneath Lake pluton and the Daly Lake rhyolite (Fig. 7) may be equivalents.

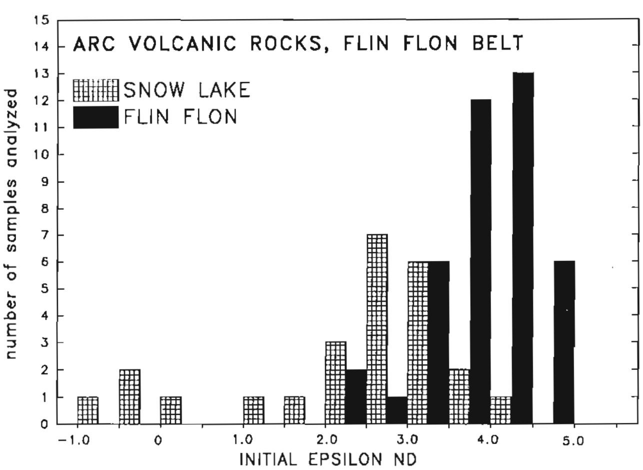

Figure 9. Histogram comparing the number of analyzed arc-type metavolcanic rocks from the Flin Flon segment with those from the Snow Lake segment. Note the higher values in the Flin Flon segment.

CONCLUSIONS AND IMPLICATIONS

The initial $^{143}Nd/^{144}Nd$ ratios (expressed as ε_{Nd} units) of Amisk Group metavolcanic rocks from the Flin Flon belt provide important constraints on their petrogenesis, tectonic setting, and metallogenesis. The important conclusions are as follows:

1. Amisk Group rocks have much more isotopic variability than previously recognized.

The range of initial ε_{Nd} values is considerable, from -0.4 to +4.8, which is similar to the range of values found in the La Ronge-Lynn Lake domain to the northwest. Such a large range of ε_{Nd} values is not consistent with an intraoceanic arc well-removed from the influences of older crust. Isotopic evidence for the existence of older crust near in the Flin Flon area is found within an assemblage of variably-strained volcanic and intrusive rocks near Mystic Lake and the West Arm of Lake Athapapuskow, in fault contact with the Flin Flon arc rocks (Thomas, 1991; Syme, 1988). Foliated Mystic Lake granodiorite and foliated quartz-megacrystic tonalite from these respective areas have ε_{Nd} (1906 Ma) values of -6.1 and -3.9, and late Archean depleted mantle extraction ages. The Mystic Lake granodiorite, dated at 1906 ± 2 Ma (Heaman et al., 1991, in press), also contains inherited zircons with a minimum age of 1951 Ma (Heaman et al., in press). If these rocks were magmas at 1906 Ma, then their source regions must have consisted largely of earliest Proterozoic or Archean crust. The presence of late Archean zircons within 1892 Ma rhyolite of the Snow Lake segment (Machado and David, 1992) provides compelling evidence for the involvement of Archean crust during the earliest stages of arc magmatism at $\approx$ 1890 Ma.

2. Arc metavolcanic rocks from the Flin Flon and Snow Lake segments have different isotopic signatures.

Figure 9 shows a histogram of initial ε_{Nd} values for ca. 1890 Ma metavolcanic rocks, comparing the Flin Flon and Snow Lake segments. It is apparent that the Flin Flon segment has higher overall initial ε_{Nd} values, and thus the two segments probably evolved independently. The Snow Lake segment must have interacted with older crust to a greater degree than the Flin Flon segment.

3. Within the Flin Flon segment, the arc-type assemblage may have been derived from a different mantle source than the MORB-type assemblage.

Arc assemblage rocks were probably derived largely from sub-arc mantle with ε_{Nd} values between +3.8 and +4.8, whereas MORB-types were derived from back-arc (asthensopheric?) sources with ε_{Nd} values of +3 to +4 (Fig. 5). The Nd isotopic results are consistent with the geochemical distinctions between these assemblages (Syme, 1990).

4. The Nd-isotopic profiles are useful for stratigraphic correlation of the volcanic rocks.

The three detailed sections presented here (Fig. 6, 7, 8) demonstrate the direct correlation of the Nd isotopic values with the mapped stratigraphy. The presence of isotopically distinct horizons such as the Snell Lake basalt (Fig. 8) will greatly aid future attempts at correlating units in the Snow Lake area. It is evident by comparing the three sections that each is unique and represents a separate volcanic centre.

5. Massive sulphide deposits are associated with periods of isotopically-primitive rhyolitic magmatism.

The rhyolitic units associated with the massive sulphide deposits are among the most isotopically primitive units within a given sequence of rocks. The Nd isotopes may therefore be useful for assessing the economic potential of volcanic sequences. Synvolcanic intrusions, such as the Richard Lake and Sneath Lake plutons are also isotopically primitive. This suggests a coupling between ore deposition and input of mantle-derived heat or fluids.

ACKNOWLEDGMENTS

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Geochronology of a tonalite from the Sherridon area, Manitoba

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Hunt, P.A., and Froese, E., 1992: Geochronology of a tonalite from the Sherridon area, Manitoba; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 85-88.

Abstract

Deformed intrusive grey tonalite gneiss forms an important lithological component of felsic gneisses correlative with ca. 1.9 Ga Amisk Group in the Sherridon area of Manitoba. One of these tonalite gniess bodies yielded a U-Pb zircon concordia upper intercept age of 1816 ± 2 Ma. The age of this synmetamorphic intrusion is regarded as representing the time of peak metamorphism within the southern flank of the Kisseynew gneiss belt.

Résumé

Dans la région de Sherridon au Manitoba, les gneiss tonalitiques gris, intrusifs et déformés, constituent une composante lithologique importante des gneiss felsiques correspondant au Groupe d'Amisk (1,9 Ga env.). L'un de ces massifs de gneiss tonalitique a donné un âge U-Pb sur zircon de 1 816 ± 2 Ma (intercept supérieur sur diagramme Concordia). L'âge de cette intrusion synmétamorphique est considéré comme coïncidant avec la période de métamorphisme maximal qui a marquée le flanc méridional de la zone gneissique de Kisseynew.

INTRODUCTION

The Sherridon area lies within the southern flank of the Kisseynew gneiss belt (Fig. 1). The rocks of the southern flank are highly metamorphosed equivalents of units forming the Flin Flon volcanic belt, i.e. the older Amisk Group (1886 ± 2 Ma) and the younger Missi Group (1832 ± 2 Ma); both groups were metamorphosed at 1814 +17/-11 Ma (Gordon et al., 1990). The stratigraphic units established for the Flin Flon volcanic belt were employed in recent mapping of the Kisseynew gneisses (Ashton, 1992; Zwanzig and Schledewitz, 1992). Rocks in the vicinity of Sherridon, originally mapped as the Sherridon Group (Bateman and Harrison, 1946), pose some difficulty in an attempt to place them into a stratigraphic framework. Although Ashton and Froese (1988) suggested that they be assigned to the Amisk Group, supporting geochronological evidence is lacking so far (Ashton et al., 1992), and Zwanzig and Schledewitz (1992) retained the Sherridon Group as a separate unit. A previous attempt to date the Sherridon Group (Ashton et al., 1992), however, was not successful as two felsic gneisses failed to contain zircon. To place a minimum age on these felsic gneisses, a tonalite gneiss, that is intruded into the Sherridon Group, was collected with the hope of obtaining an Amisk age.

SHERRIDON GROUP GNEISSES

The most common rocks of the Sherridon Group are medium grained quartz-rich gneisses (Froese and Goetz, 1981). Two previous samples gave U-Pb titanite ages of 1804 ± 3 Ma and 1808 ± 2 Ma (Ashton et al., 1992) indicating that the Kisseynew gneisses at Sherridon cooled through the closure

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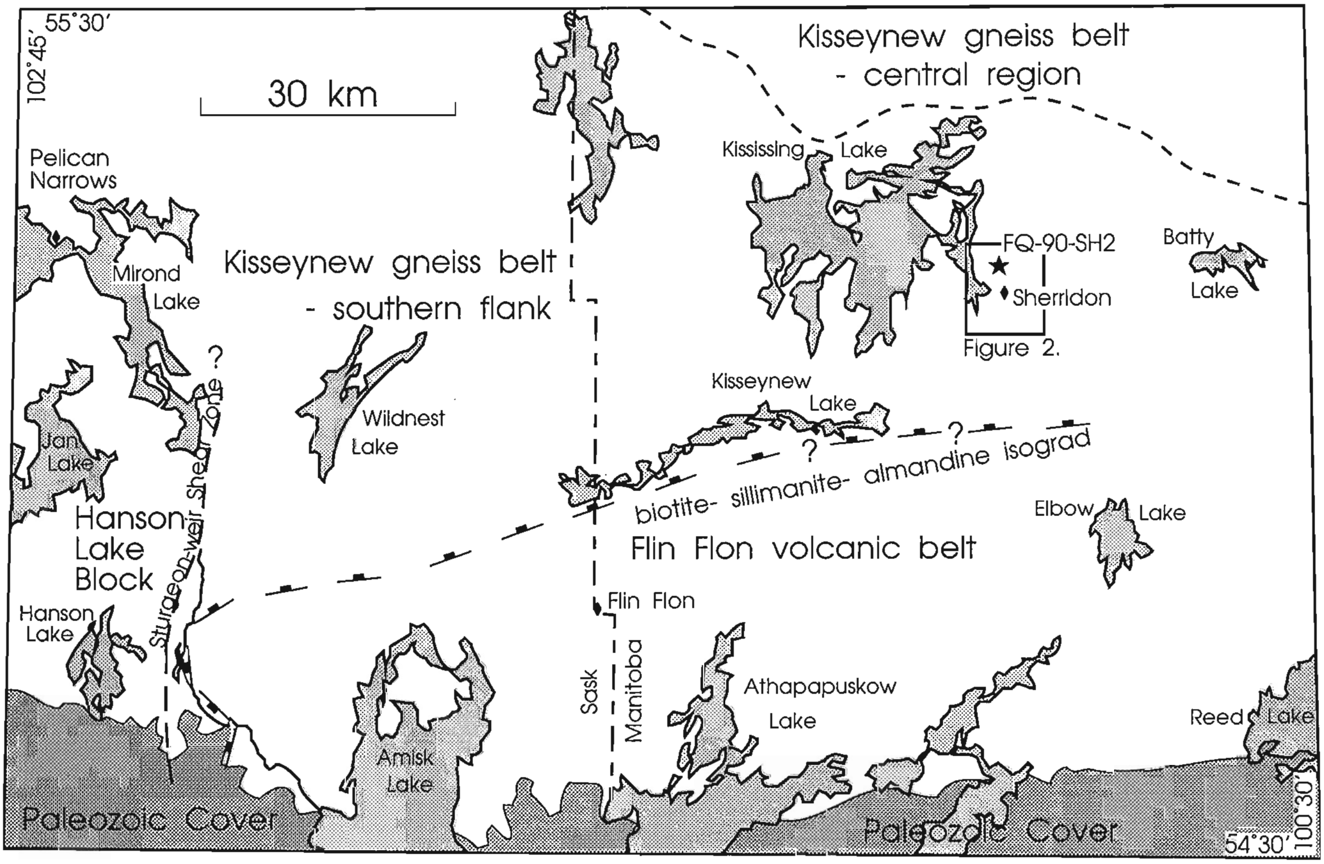

Figure 1. Sketch map of the western Flin Flon volcanic belt and southern flank of the Kisseynew gneiss belt (from Ashton et al., 1992).

Table 1. U-Pb zircon data

Fraction	Wt. mg[a]	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)
Tonalite Gneiss											
L	0.006	599	185	3418	21	0.1	0.3228 ± 0.09%	4.938 ± 0.11%	0.91	0.11096 ± 0.04%	1815.3 ± 1.6
MB	0.008	206	64	1432	23	0.3	0.3228 ± 0.10%	4.947 ± 0.13%	0.81	0.11114 ± 0.08%	1818.1 ± 2.8
M	0.007	283	87	1097	36	0.2	0.3236 ± 0.09%	4.958 ± 0.13%	0.79	0.11113 ± 0.08%	1818.0 ± 2.9
S	0.008	453	140	4373	16	0.2	0.3229 ± 0.08%	4.939 ± 0.10%	0.93	0.11092 ± 0.04%	1814.6 ± 1.3

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
[a] Sample weight error of ± 0.001 mg in concentration uncertainty
[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 $^{207}Pb/^{235}U$.

temperature of titanite at about 1806 Ma. The quartz-rich gneisses contain highly foliated concordant lenses of tonalite; contacts are transposed into the foliation and the nature of the original contacts has been obscured, although these tonalite bodies are inferred to be intrusive in origin. The age of the tonalites could range in age from Amisk to synmetamorphic. A sample was collected for geochronology from a 50 m thick gneissic tonalite body near Cold Lake Manitoba (Fig. 2).

ANALYTICAL TECHNIQUES

U-Pb analytical methods follow those outlined by Parrish et al. (1987). Other techniques included a strong air abrasion of all zircon fractions (Krogh, 1982). U and Pb blanks were approximately 1 and 20 picograms, respectively. The U-Pb data are presented in Table 1 and Figure 3. In the concordia plot, errors are shown at the 2σ level.

RESULTS

Sample FQ-90-SH2 was collected 2 km north-northwest of Sherridon, Manitoba, from a bend in the power line, (55°8'31.5"N, 101°5'57.0"W, Fig. 2). It is a dark grey gneiss with light pink feldspar. Grain size is about 1mm, with quartz aggregates up to 1 cm. The rock has a granoblastic metamorphic texture and consists of 20% quartz, 50% plagioclase, 20% biotite, 10% hornblende. It occurs within quartz-rich paragneisses of the Sherridon Group.

Zircons from the tonalite consisted of very small (<62μm) stubby prismatic grains with a length:width ratio of 2:1. Crystals were generally clear and colourless with no internal fractures or inclusions. It is difficult to determine from the morphology of the zircons whether they are igneous or metamorphic, though the fine grain size may suggest a metamorphic origin. Internal older cores were not observed in any of the grains.

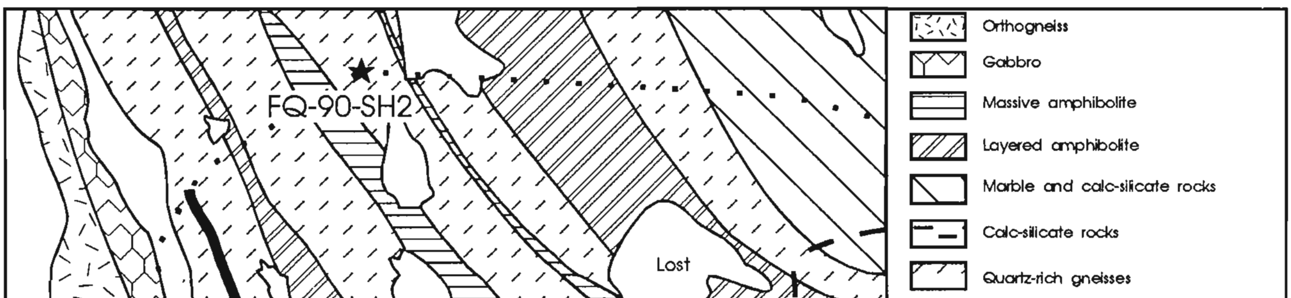

Figure 2. Simplified geological map of the Sherridon area (from Froese and Goetz, 1981).

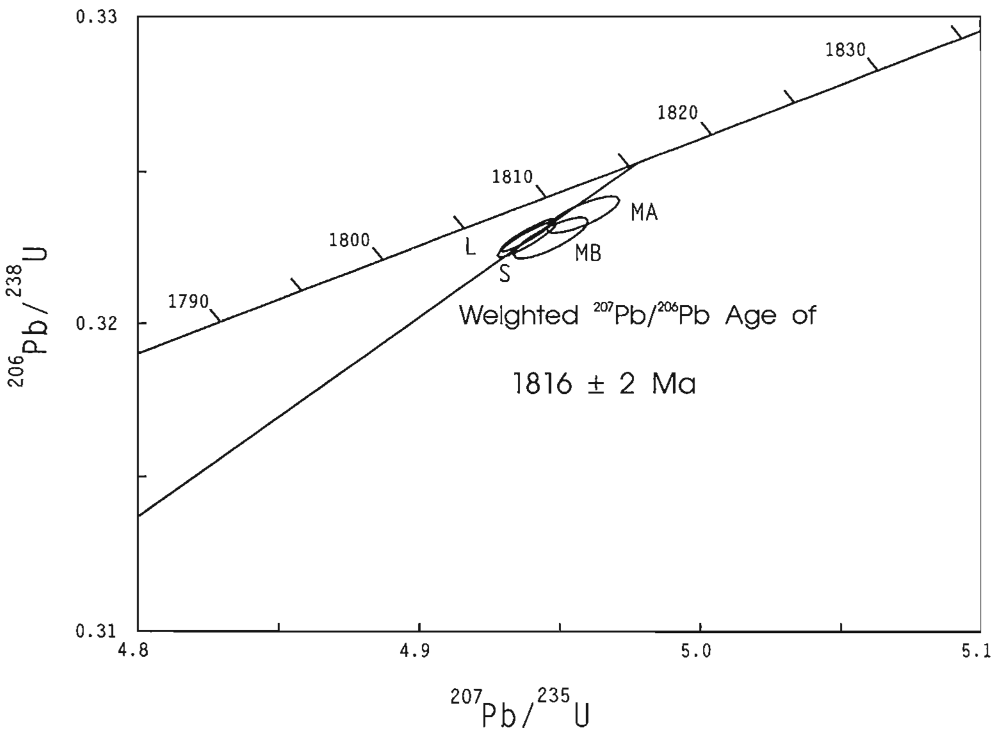

Figure 3.

Concordia diagram showing results of U-Pb zircon analyses from a grey gneiss FQ-90-SH2, Sherridon, Manitoba.

Four zircon fractions were analyzed (Table 1). The four fractions plot slightly below concordia, with error ellipses overlapping (Fig. 3). This pattern of similar degrees of discordance for all analyzed zircons could be attributed to the fact that all fractions are approximately the same size and may have been affected by similar lead loss. A weighted average of the ^{207}Pb-^{206}Pb ages of the four fractions gives a concordia upper intercept age of 1816 ± 2 Ma, which we interpret as the age of zircon crystallization. This is a mininum age of zircon crystallization since the Pb loss may have occured hundreds of millions of years ago, instead of recently as the 1816 Ma age implies.

The interpreted upper intercept age of 1816 Ma can be interpreted in several ways. The zircons could be igneous, dating the time of crystallization, or they could be recrystallized or formed during metamorphism, dating the time of metamorphism. Titanite ages from rocks in the Sherridon area are 1808 Ma and 1804 Ma (Ashton et al. 1992), suggesting that peak metamorphism occurred prior to ca. 1810 Ma. Either igneous and metamorphic origin for the zircon is consistent with this data; in any case it is apparent that metamorphism of the tonalite occurred in the 1.82-1.81 Ga interval. This is also consistent with the 1814 +17/-11 Ma age of metamorphism inferred by Gordon et al. (1990).

ACKNOWLEDGMENTS

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Age constraints on the Puskuta Lake shear zone from U-Pb dating of granitoid rocks, Kapuskasing uplift, northern Ontario[1]

Robert W. Sullivan[2] and Alain D. Leclair[2]

Sullivan, R.W. and Leclair, A.D., 1992: Age constraints on the Puskuta Lake shear zone from U-Pb dating of granitoid rocks, Kapuskasing uplift, northern Ontario; in *Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 89-96.*

Abstract

U-Pb zircon and titanite ages of 2682 ± 3 Ma and 2644 ±2 Ma, respectively, for the undeformed granodiorite of the Dishnish batholith, combined with previously published ages from the granodioritic mylonite of the Puskuta Lake shear zone indicate that the latest shear deformation was post-2682 Ma. If the deformation occurred above the titanite closure temperature of ca. 600°C, then the shearing must have occurred prior to 2665 Ma. However, if the deformation occurred below the titanite closure temperature, the shearing was post-2665 Ma and possibly post-2644 Ma. These results provide some constraints for the timing of dextral movement along major auriferous shear zones in the central Superior Province.

Résumé

Des âges U-Pb sur zircon et titanite de 2 682 ± 3 Ma et de 2 644 ± 2 Ma ont été obtenus pour la granodiorite non déformée du batholite de Dishnish; combinées aux âges assignées dans une publication antérieure à la mylonite granodioritique de la zone cisaillée du lac Puskuta, ces données indiquent que la déformation par cisaillement la plus récente a eu lieu après 2 682 Ma. Si les roches ont été déformées au-dessus de la température de fermeture de la titanite (autour de 600 °C), le cisaillement doit avoir eu lieu avant 2 665 Ma. Cependant, si elles l'ont été au-dessous de cette température, le cisaillement doit s'être produit après 2 665 Ma et peut-être après 2 644 Ma. Ces résultats permettent de restreindre un peu l'intervalle de temps au cours duquel il y a eu des mouvements dextres le long des principales zones de cisaillement à minéralisation aurifère du centre de la province du lac Supérieur.

INTRODUCTION

The Kapuskasing uplift is a northeast-trending zone characterized by low- to high-grade gneisses, granitoid intrusions and geophysical anomalies that transects the Abitibi and Wawa volcano-plutonic and Quetico and Opatica metasedimentary-plutonic subprovinces. The geology of this region is described by Bennett et al. (1967), Thurston et al. (1977), Percival (1981, 1990) and Leclair (1992 and references therein). In the Kapuskasing-Groundhog-Missinaibi River area (Leclair, 1992; Fig. 1), the Kapuskasing uplift comprises three geological-geophysical entities: the Groundhog River, Val Rita and northern Chapleau blocks (Percival and McGrath, 1986). These tectonic blocks are for the most part bounded by major cataclastic zones across which there has been up to 15 km of differential block movement. They are interpreted to collectively reveal the structure of the Archean crust of the central Superior Province from paleodepths of 10-15 km down to 30-35 km (Leclair, 1990).

The westernmost part of the Kapuskasing-Groundhog-Missinaibi River area roughly coincides with the transition between volcano-plutonic terranes of the Wawa subprovince

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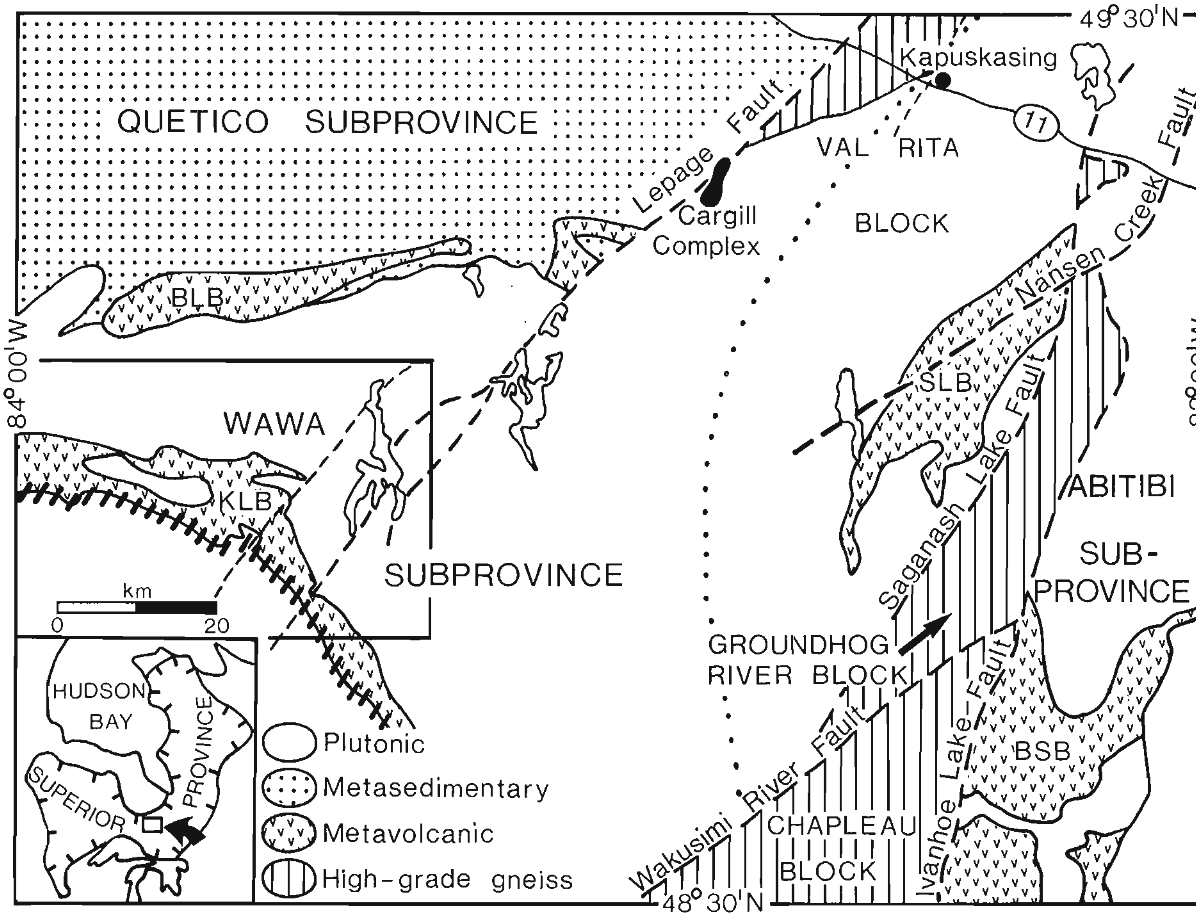

Figure 1. Regional geology and major structural elements of the Kapuskasing-Groundhog-Missinaibi River area encompassing the central segment of the Kapuskasing uplift and adjacent parts of the Abitibi, Wawa and Quetico subprovinces in the central Superior Province (modified from Percival and McGrath, 1986). 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-Strachan belt; SLB, Saganash Lake belt; KLB, Kabinakagami Lake belt; BLB, Buchanan Lake belt. The geology of the western part of the area (outlined) and sample locations are presented in Figure 2.

and generally higher-grade tonalitic gneisses of the Val Rita block (Fig. 1). The transition is marked by the continuation of the Kabinakagami Lake metavolcanic belt, Puskuta Lake shear zone and Dishnish batholith (Fig. 2) across the southern projection of the Lepage fault which, farther north, defines an abrupt contact between Quetico and Kapuskasing rocks.

This paper is the final instalment of a regional geochronology study which was undertaken to date major geological events in lower and upper crustal rocks of the central Kaspuskasing uplift, such as tonalitic magmatism, regional deformation, high-grade metamorphism, late to post-tectonic plutonism and discrete ductile shearing (see Leclair and Sullivan, 1991; Sullivan and Leclair, 1992). Here, we discuss the significance of age determinations obtained from granodiorite of the Dishnish batholith and mylonite of the Puskuta Lake shear zone. The U-Pb zircon and titanite ages are used to place constraints on the latter part of the deformational history of the Puskuta Lake shear zone.

PREVIOUS GEOCHRONOLOGY

One of the main objectives of this study was to define the age of the Puskuta Lake shear zone. To this end, a sample of granodioritic mylonite from the southern margin of the shear zone, inferred to be derived from the adjacent Dishnish batholith, was collected and dated. However, the significance of the zircon age at 2708 ±4 Ma and the titanite age at 2665 ± 4 Ma obtained from this sample was equivocal (see Leclair and Sullivan, 1991, sample 6). One preliminary interpretation was that the zircon age possibly reflected an inherited component in the granodiorite protolith and the titanite age gave the age of the latest shearing event along the Puskuta Lake shear zone. This interpretation was based on the assumption that titanite formed and/or was reset during high-T ductile deformation. An alternative explanation was offered, wherein the titanite age was interpreted as the igneous

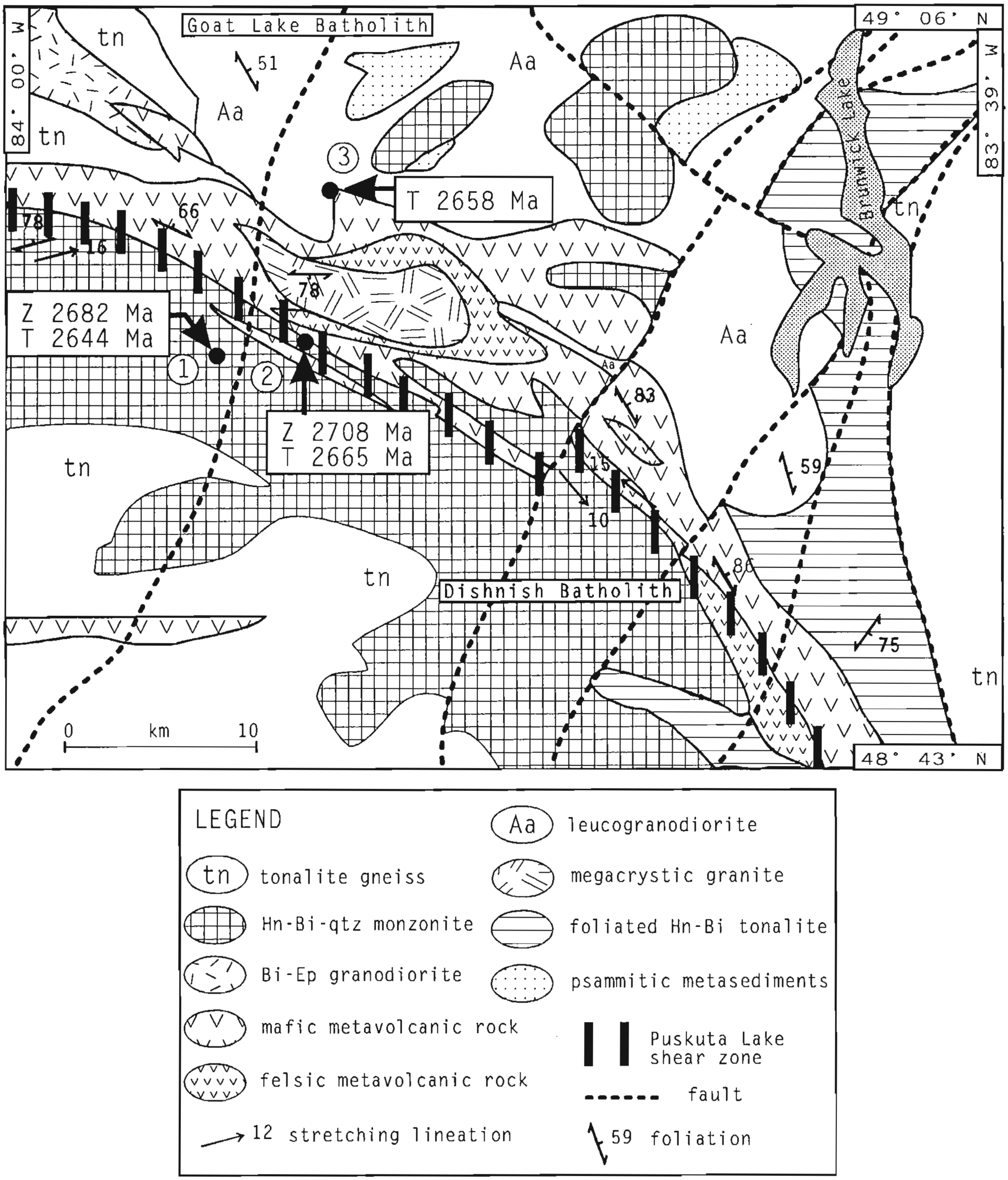

Figure 2. Simplified geological map for the western part of the Kapuskasing-Groundhog-Missinaibi River area (modified from Bennett et al., 1967; Thurston et al., 1977; Leclair, 1992). Geochronology samples are: 1 - granodiorite of the Dishnish batholith (this paper), 2 - granodioritic mylonite of the Puskuta Lake shear zone (this paper; Leclair and Sullivan, 1991) and 3 - leucogranodiorite of the Goat Lake batholith (Leclair and Sullivan, 1991). In the legend Hn = hornblende, Bi = biotite, Ep = epidote and Qtz = quartz.

cooling age of the protolith. In this paper we report additional U-Pb results for the same tectonite sample and, in light of new U-Pb zircon and titanite ages from flanking isotropic granodiorite of the Dishnish batholith, propose further interpretations for the age of the Puskuta Lake shear zone. For previous geochronology on the Goat Lake batholith referred to in this paper, see Leclair and Sullivan (1991, sample 5).

GENERAL GEOLOGY

The general geology of the western part of the Kapuskasing-Groundhog-Missinaibi area (Fig. 2) shows that amphibolite-facies metavolcanic rocks of the Kabinakagami Lake belt are in sheared contact with massive granitoid rocks of the Dishnish batholith along the Puskuta Lake shear zone, and are intruded in the north by leucogranodiorite of the Goat Lake batholith (Leclair, 1990, 1992).

Dishnish batholith

The Dishnish batholith forms a large, irregular-shaped, dominantly massive body of quartz monzonite, granodiorite and monzogranite which is continuous with Percival's (1981) unit 10c in the adjacent map area to the south. Field relations and map pattern (Fig. 2) indicate that these rocks are intrusive into biotite-epidote tonalitic gneisses to the south. Along the northern margin of the batholith, the granitoid rocks display a progressively more intense shear fabric, grading into protomylonite and mylonite of the Puskuta Lake shear zone. This highly sheared contact with metavolcanic rocks of the Kabinakagami Lake belt is inferred to have been previously intrusive based on the occurrences of lenses of layered amphibolite enclosed in deformed quartz monzonite, and the presence of granitoid rocks of Dishnish batholith affinities north of the shear zone.

Table 1. U-Pb zircon and titanite analytical data

Fraction description[f]	Wt. µg	U ppm	Pb^a ppm	$^{206}Pb^b/^{204}Pb$	Pb^c pg	$^{208}Pb^a$ %	$^{206}Pb^d/^{238}U$	$^{207}Pb^d/^{235}U$	Corr[e] Coeff	$^{207}Pb^d/^{206}Pb$	$^{207}Pb/^{206}Pb$ Age (Ma)
1. Dishnish batholith: L863-1-89 (48° 55′ 13" N 83° 51′ 10" W)											
A, +105, cl, euh, S	5	268	147	559	71	5.0	0.5120 ±.11	12.878 ±.16	0.75	0.18242 ±.11	2675.0 ±3.6
B, +105, cl, euh	8	150	86	691	52	7.8	0.5160 ±.11	13.042 ±.15	0.82	0.18333 ±.08	2683.3 ±2.8
C, -74, =>4:1,cl, euh	15	186	106	5182	16	10.5	0.5020 ±.09	12.583 ±.10	0.95	0.18181 ±.03	2669.5 ±1.1
XM, +105, "mantle", S	1	70	36	85	30	3.9	0.4895 ±1.3	12.246 ±1.3	0.85	0.18145 ±.71	2666 ±23
XMC, +105 "mant/core"	1	34	20	63	22	14.7	0.4830 ±2.7	12.472 ±2.5	0.89	0.18730 ±1.2	2719 ±39
EE, Tit, darkest, NA	298	142	160	1033	1165	53.7	0.5135 ±.08	12.686 ±.12	0.79	0.17917 ±.08	2645.2 ±2.6
FF, Tit, dark, NA	325	150	130	861	1583	41.0	0.5040 ±.08	12.438 ±.14	0.77	0.17898 ±.09	2643.5 ±3.0
FF2, Tit, dark, NA	363	131	147	1153	1142	54.4	0.5021 ±.10	12.397 ±.13	0.86	0.17909 ±.07	2644.4 ±2.3
GG, Tit, lighter, NA	283	208	204	705	2364	47.1	0.5085 ±.10	12.536 ±.17	0.76	0.17880 ±.11	2641.7 ±3.7
GG2, Tit, lighter, NA	334	136	154	669	1929	54.1	0.5099 ±.10	12.581 ±.17	0.76	0.17897 ±.12	2643.4 ±3.9
2. Puskuta Lake shear zone mylonite: L868-5-89 (48° 56′ 03"N 83° 47′ 08" W)											
A, -105+74, cl, euh	9	12	7	236	16	7.6	0.5237 ±.54	13.442 ±.54	0.98	0.18614 ±.11	2708.4 ±3.7
B, -74+62, cl, euh	2	152	86	973	7	8.2	0.5121 ±.26	13.015 ±.27	0.97	0.18432 ±.06	2692.1 ±2.0
C, -74+62, cl,euh	12	39	22	475	29	7.3	0.5108 ±.16	12.960 ±.21	0.84	0.18403 ±.11	2689.5 ±3.7
D, +105, cl, euh, S	4	151	85	2285	8	6.1	0.5182 ±.13	13.239 ±.13	0.96	0.18531 ±.04	2701.0 ±1.3
AA, Tit, cl, darker	132	133	72	1425	345	5.3	0.5073 ±.09	12.670 ±.12	0.88	0.18113 ±.06	2663.2 ±1.9
BB, Tit, cl, darker	128	130	71	1322	355	5.2	0.5061 ±.09	12.663 ±.12	0.88	0.18149 ±.06	2666.5 ±2.0
CC,Tit,cl,lighter, NA	115	129	70	942	448	4.9	0.5121 ±.11	12.743 ±.15	0.84	0.18048 ±.08	2657.3 ±2.7

[f] size before abrasion in microns; cl = clear, euh = euhedral, S = single grain; 4:1 = L:B ratio; Tit = titanite, rest are zircon; fractions abraded unless indicated NA = not abraded

[a] = radiogenic Pb

[b] = measured ratio, corrected for fractionation and spike

[c] = total common Pb in analysis corrected for fractionation and spike

[d] = corrected for blank Pb and U, common Pb.

[e] = correlation coefficient of errors in $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$.

Errors are 1 standard error of the mean in % except $^{207}Pb/^{206}Pb$ age errors which are 2 standard errors in Ma.
Initial common Pb compositions from Cumming and Richards (1975).

Puskuta Lake shear zone

The Puskuta Lake shear zone (Fig. 1, 2) has an arcuate, concave to the southwest map pattern. It extends for at least 75 km, from the eastern Wawa subprovince into the Kapuskasing uplift (Leclair, 1990; Fig. 2). The roughly 2-km-wide zone of mylonite contains a steep shear foliation (dipping 60-80°N to NE) and a shallow-plunging stretching lineation (10-25°) which are interpreted to be the result of mainly dextral transcurrent displacement (Leclair, 1990). The mylonitic fabric is transected by diabase dykes of the Matachewan/Hearst swarm, dated at 2454 Ma (Heaman, 1988) and pegmatite dykes which are cut by north- to northeast-trending brittle faults. In a regional tectonic context, the Puskuta Lake shear zone is regarded as a segment of a crustal-scale structure that continues through high-grade rocks of the Kapuskasing uplift and joins gold-rich shear zones in the southern Abitibi greenstone belt (Leclair et al., unpublished data).

ANALYTICAL METHODS

Zircon and titanite fractions were extracted from crushed rock samples by conventional Wilfley table, heavy liquid and Frantz magnetic separation techniques. Zircons were strongly air abraded (Krogh, 1982). Titanite was not abraded. Carefully selected single and multi-grain fractions were dissolved and analyzed using procedures outlined in Parrish et al. (1987, 1992). Analytical blanks for zircon were typically 0 and 15 pg for U and Pb respectively, and 1 and 25 pg for U and Pb for titanite. Analytical data are presented in Table 1 and displayed on concordia diagrams in Figures 3 and 4. All quoted age errors are at the 2σ(95% confidence interval).

U-Pb RESULTS

1. Dishnish granodiorite (L863-1-89)

A sample of homogeneous, undeformed granodiorite of the Dishnish batholith, located about 5 km from the southernmost recognizable shearing effects of the Puskuta Lake shear zone (Fig. 2), was selected for U-Pb zircon and titanite analyses. The rock is light grey, medium grained and equigranular, with 5 to 25% combined hornblende, biotite, epidote and titanite. It shows porphyritic texture with K-feldspar phenocrysts up to 1 cm long.

A clear, slightly coloured euhedral zircon population that is considered to be igneous in origin, predominates. Some grains had rod and bubble inclusions but no cores were seen. Analyses of three fractions (A, B and C) of this zircon type form a linear trend which we interpret as resulting from Pb loss (Fig. 3). Fraction B is concordant at 2682 ±3 Ma and is considered to be the crystallization age of the batholith. There are also darker zircons with obvious cores. In an attempt to determine the ages of the mantle and core phases from this type, a single large zircon (X) was abraded and then broken. Carefully selected fragments were then analyzed (Fig. 2) to represent the mantle, (fraction XM, 1 grain) and the mantle-core (fraction XMC, 2 grains). These are discordant but the mantle result (^{207}Pb-^{206}Pb age of XM = 2666 Ma) is consistent with the 2682 Ma igneous age, whereas fraction

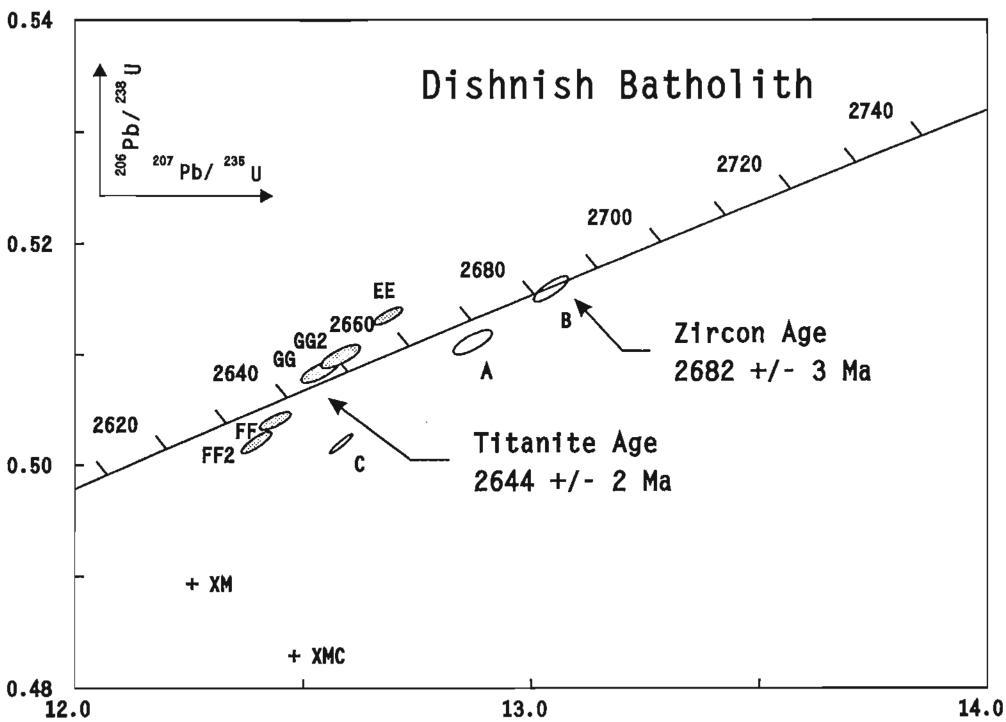

Figure 3.

U-Pb concordia diagram of zircon (open ellipses) and titanite (shaded ellipses) data from sample 1, the Dishnish batholith. Analyses XM and XMC are imprecise and plotted without error ellipses.

XMC, which is mixture of mantle and core, indicates an older core with a minimum age, based on the ^{207}Pb-^{206}Pb result, of ca. 2719 Ma.

The sample has abundant brown titanite, generally forming blocky fragments but with some euhedral crystals. Clarity ranges from nearly opaque to clear. Five fractions EE, FF and a duplicate FF2, GG and a duplicate GG2 were analyzed. Three fractions (EE, GG and GG2) plotted above concordia (Fig. 3). The reason for this is not clear. The duplicates were from the same titanite concentrate but were analyzed in separate chemistry batches. The fact that the duplicate fractions gave coherent results indicates the problem originates in the sample and not the analytical method. It is possible that the fractions which plotted above the curve contained small inclusions that did not dissolve (e.g., zircon) and which retained parent U but released the unsupported radiogenic Pb, and thus perturbed the U-Pb ratios. Alternatively, the reverse discordance of some of the analysis may indicate that these fractions have either gained radiogenic Pb or lost U, but it is not clear why it affected only some of the grains. We consider the best estimate of the age to be a weighted ^{207}Pb-^{206}Pb age, at 2644 ± 2 Ma. This is considered to be the igneous titanite cooling age of the Dishnish batholith.

2. Puskuta Lake granodioritic mylonite (L868-5-89)

The mylonite, and the analyzed fractions were previously described in Leclair and Sullivan (1991) but are described again for continuity. The rock is a light grey, fine-grained, strongly sheared granodiorite tectonite displaying quartz ribbons and K-feldspar augen. The mylonite sample comes from the southern margin of the Puskuta Lake shear zone within a band of highly deformed granitoid rocks flanked on both sides by metavolcanic rocks of the Kabinakagami Lake belt (Fig. 2). These sheared granitoid rocks were previously inferred to have been derived from a protolith of homogeneous, massive granodiorite which was part of the Dishnish batholith (Leclair and Sullivan, 1991). However, as discussed later, it now appears more likely that granitoid rocks of another source were incorporated in the deformation zone.

Several populations of zircon occur in the sample but a clear, colourless to light brown, euhedral type predominates. Some grains of this type are internally featureless whereas others have fine zoning and clear bubble and rod inclusions. Another population is more translucent, dark brown, generally anhedral, and contains probable cores. The titanite from the sample varies from small, equant, clear, light to dark brown grains, some of which have crystal faces. Some euhedral titanite grains were seen in thin section.

The zircons were picked from the clear, colourless, euhedral type and grains with visible cores or other internal features were excluded. The three fractions of zircon (A, C, D) and the titanite fractions (AA, BB, CC,) were previously reported by Leclair and Sullivan (1991). The reported titanite age is 2665 ±4 Ma. Zircon fraction B is a new analysis, not previously reported. The analyzed zircons are now considered to be magmatic and not xenocrystic which was an alternative interpretation (Leclair and Sullivan, 1991). The four zircon results define a linear array (MSWD = 0.03), which is interpreted to result from surface-correlated Pb loss (Fig. 4). A regression of these data gives an upper intercept age of

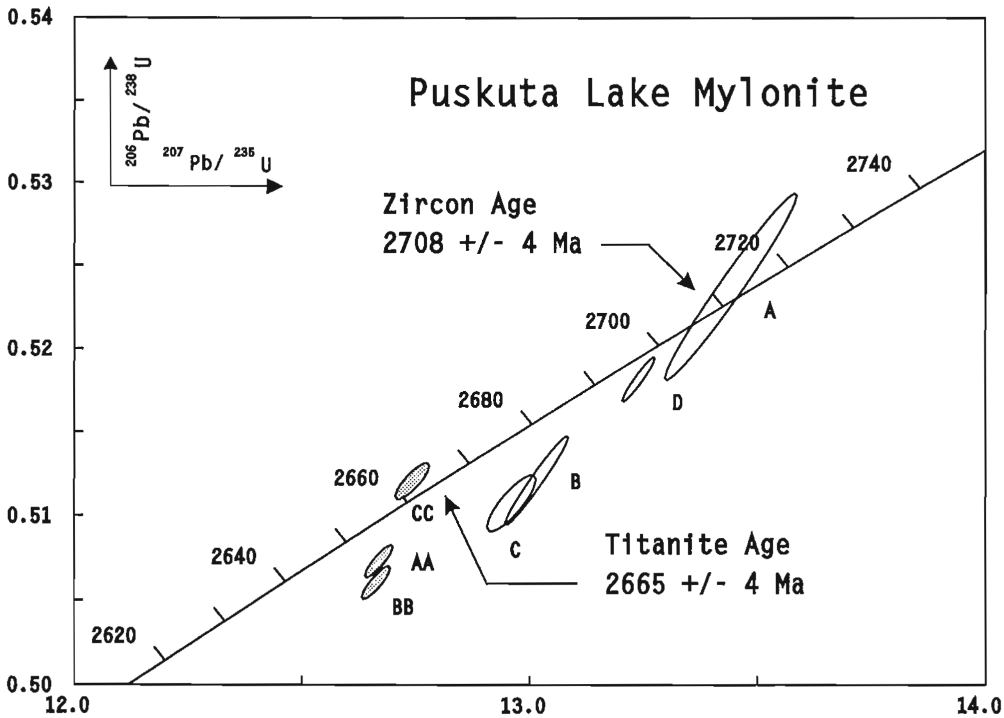

Figure 4.

U-Pb concordia diagram of zircon (open ellipses) and titanite (shaded ellipses) data from sample 2, the Puskuta Lake mylonite.

2706 Ma, which agrees within error with the previously reported concordant age of 2708 ±4 Ma given by fraction A (Leclair and Sullivan, 1991).

The 2708 Ma zircon age is interpreted as the crystallization age of the granodiorite protolith of the mylonite sample. The 2665 Ma titanite age is simply interpreted as the U-Pb closure age of the titanite in the tectonite. As discussed below, the significance of this 2665 Ma age in constraining the age of the Puskuta Lake shearing event depends on whether the deformation occurred at temperatures above or below the closure temperature of titanite, estimated at ca. 600°C.

DISCUSSION

The zircon and titanite ages of the Dishnish batholith are 2682 ±3 and 2644 ±2 Ma respectively. The protolith of the tectonite, however, appears to have formed at 2708 ±4 Ma, predating the formation of the Dishnish batholith by ca. 25 million years. These new U-Pb results indicate that the granodiorite of the Dishnish batholith is not the protolith of the dated granodioritic mylonite from the Puskuta Lake shear zone, as previously assumed. Instead, the two analyzed samples most likely represent disparate granitoid rocks of different age and provenance that were juxtaposed as a result of transcurrent movement along the shear zone.

Based on field relations alone, the age of the latest ductile deformation along the Puskuta Lake shear zone is bracketed between the emplacement of the Dishnish batholith at 2682 ± 3 Ma and cross-cutting diabase dykes of the Matachewan/ Hearst swarm, at 2454 Ma (Heaman, 1988). An additional lower age constraint may include a set of late pegmatite dykes that cut cleanly across the mylonitic fabric. These dykes have been inferred to be associated with the Goat Lake leucogranodiorite, which is interpreted to have a minimum emplacement age of 2658 +10/-8 Ma, given by U-Pb dating of titanite (Leclair and Sullivan, 1991). However, the dykes themselves have not been dated.

The 2665 ± 4 Ma age obtained from titanite in the granodioritic mylonite records the age of the last cooling through its closure temperature of ca. 600°C. The deformation of the Puskuta Lake shear zone occurred under amphibolite-facies conditions, at temperature ranging from about 500 to 700°C. This range encompasses the closure temperature of titanite. The exact temperature at which deformation occurred is not known but two alternatives are presented: a) if the deformation occurred at temperatures above ca. 600°C then the 2665 Ma titanite age must be interpreted as a cooling age after the shearing event and therefore, the shearing would be pre-2665 Ma; b) if the temperature during deformation was below 600°C then the 2665 Ma titanite age records the igneous cooling age of the protolith of the tectonite, and the shearing event was post-2665 Ma.

Portions of the Dishnish batholith were effected by the shearing, therefore its U-Pb zircon age constrains the latest shear deformation to post-2682 Ma. If alternative b) above is correct, then the titanite age of 2644 ±2 Ma for the batholith, must represent a maximum age for the latest ductile movement along Puskuta Lake shear zone.

The Puskuta Lake shear zone represents a favourable structural site for gold mineralization, owing to its brittle-ductile nature and proximity to a multilayered metavolcanic sequence (cf. Colvine et al., 1988; Wong et al., 1991). It has been correlated with other major auriferous shear zones in the central Superior Province (Leclair et al., unpublished data). These regional east-trending deformation zones may have been active since ca. 2690 Ma (Corfu et al., 1989), through the deposition of the Timiskaming Group, spanning the interval from 2685 to 2675 Ma (Corfu et al., 1991). The U-Pb results presented in this paper place additional constraints on the deformational history of these shear zones.

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U-Pb age constraints on Early Proterozoic mafic magmatism from the southern Superior and western Grenville provinces, Ontario

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Prevec, S.A., 1992: U-Pb age constraints on Early Proterozoic mafic magmatism from the southern Superior and western Grenville provinces, Ontario; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 97-106.

Abstract

U-Pb zircon ages from metamorphosed gabbro-anorthosite intrusions in the Grenville and Superior Provinces are presented. A gneissic anorthosite from the Grenville Province to the west of Lake Nipissing in Mercer Township is dated at 1222 ± 2 Ma. The Wanapitei intrusion, a metamorphosed gabbro-norite located within the Grenville Province and adjacent to the Grenville Front, is dated at 1746 +6/-5 Ma, with a Grenvillian lower intercept. A metagabbro from the Superior-Southern Province boundary to the east of Sudbury shows Archean inheritance with a Proterozoic intrusion age.

Résumé

Des âges U-Pb sur zircon associés à des intrusions de gabbro et d'anorthosite métamorphisés des provinces de Grenville et du lac Supérieur sont présentées dans cet article. Une anorthosite gneissique de la province de Grenville, à l'ouest du lac Nipissing (canton de Mercer), a été datée à 1 222 ± 2 Ma. L'intrusion de Wanapitei, composée de gabbro et de norite métamorphisés, est observée dans la province de Grenville, près du front; elle a été datée à 1 746 +6/-5 Ma (intercept inférieur au Grenvillien). Un métagabbro échantillonné à la limite entre les provinces du lac Supérieur et du Sud, à l'est de Sudbury, présente un héritage archéen; son intrusion remonte au Protérozoïque.

INTRODUCTION

Three mafic igneous intrusions have been dated as part of a study to determine the extent and genesis of Early Proterozoic mafic magmatism in the Sudbury area and the northwestern Grenville Province (Fig. 1). To the west of Sudbury 2480 to 2450 Ma mafic magmatism occurs as intrusive lopoliths and sills (e.g. East Bull Lake, Shakespeare-Dunlop intrusions; James et al., 1983; James and Harris, 1977, respectively; Krogh et al., 1984), volcanics (basal Huronian tholeiites; e.g. Jolly, 1987) and as dyke swarms (Matachewan; Heaman, 1989). The presence of the ca. 2450 Ma River Valley gabbro-anorthosite pluton (Ashwal and Wooden, 1989; Heaman, pers. comm., 1990) immediately to the south of the Grenville Front Boundary Fault suggests that this magmatic episode was not restricted to the area north of the Southern Province.

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Most of the 2450 Ma mafic magmatism, both intrusive and extrusive, appears to be geographically related to early Proterozoic rifting along the Murray Fault system and its subsidiaries, or perhaps along a continuation thereof in the case of the River Valley pluton. Previous workers have speculated that undated gabbro-anorthosites and anorthosites farther south in the Grenville Province may also be related (e.g. Ashwal and Wooden, 1989; Rousell, pers. comm., 1988). Two of these intrusions, the Wanapitei Complex and the Mercer anorthosite are included in this study. A third body located between these two, the St. Charles anorthosite, has a ca. 1210 Ma age (e.g. Prevec and Baadsgaard, 1991a).

The ages of these intrusions are important in establishing and constraining the extent of early Proterozoic magmatism across the Grenville Front, related to rifting and formation of the Huronian depositional basin.

A small, highly-deformed intrusion immediately to the east of the Sudbury Igneous Complex in Falconbridge Township was sampled also. Based on petrographic and tectonic evidence (Prevec and Baadsgaard, 1991b), an affinity with the ca. 2480 Ma rocks to the west is suggested.

ANALYTICAL TECHNIQUES

U-Pb analytical methods used follow those described by Parrish et al. (1987). All zircon fractions were air-abraded using the technique of Krogh (1982), dissolved in microcapsules (Parrish, 1987), and spiked with a mixed ^{205}Pb-^{233}U-^{235}U isotopic tracer (Parrish and Krogh, 1987). Isotopic analyses were conducted as described by Roddick et al. (1987) and the errors assessed by numerical propagation (Roddick, 1987). Data regressions use the method of York (1969). Total Pb blanks associated with the samples in this study were less than 10 pg. Isotopic data are presented in Table 1 and displayed in concordia diagrams in Figures 3, 5, and 7.

MERCER ANORTHOSITE (SPA-89-22)

Geology and geochronology

The Mercer Anorthosite lies in Falconer Township, south of the western arm of Lake Nipissing. It is hosted by migmatitic to gneissic quartz dioritic to monzonitic rocks which are interlayered with biotite gneiss similar to that which is host

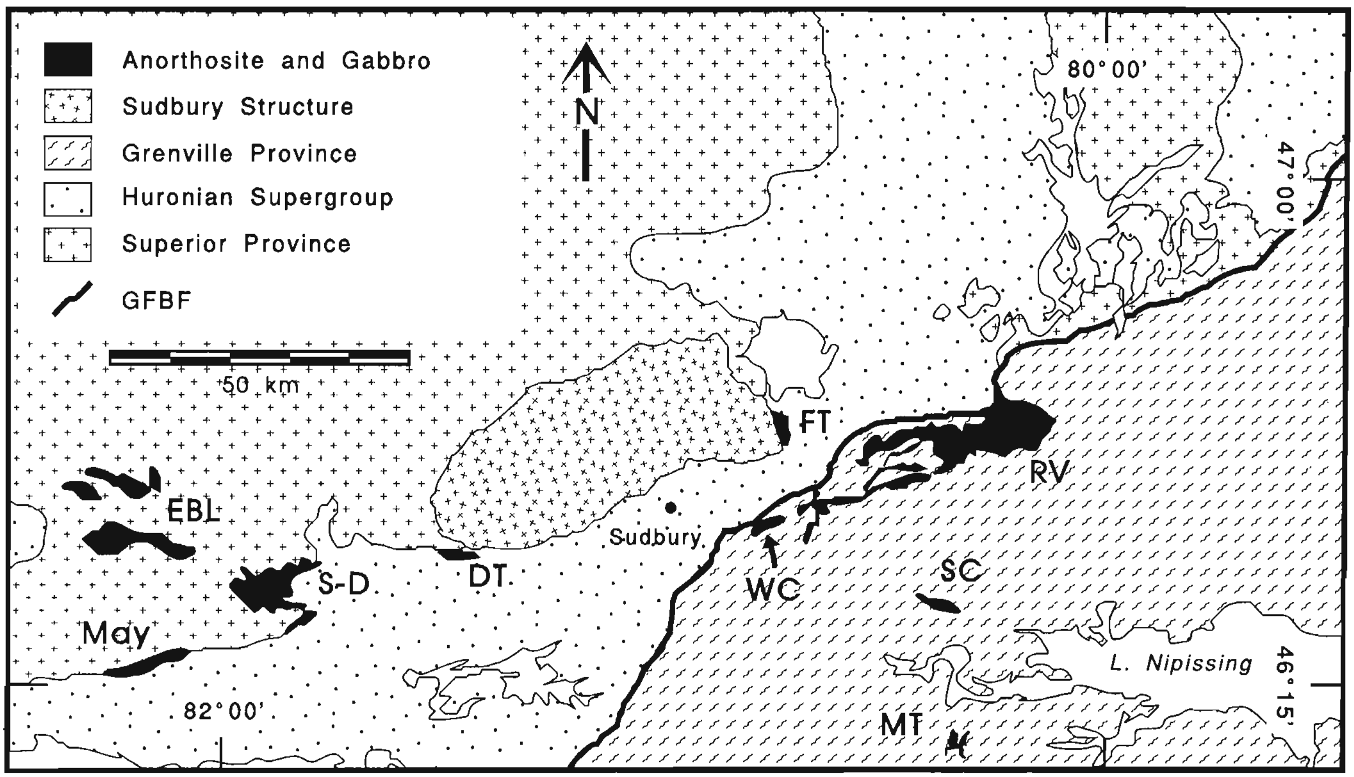

Figure 1. Map of Sudbury area, Ontario, showing gabbro-anorthosite locations (from Ashwal and Wooden, 1989). Abbrev. are GFBF = Grenville Front Boundary Fault; EBL = East Bull Lake; May = May Twp.; S-D = Shakespeare-Dunlop; DT = Drury Twp.; FT = Falconbridge Twp.; WC = Wanapitei Complex; RV = River Valley pluton; SC = St. Charles Anorthosite; MT = Mercer Anorthosite.

Table 1. U-Pb zircon analytical data

Fraction, Size[a]	Wt. µg	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)
Mercer Anorthosite, SPA-89-22											
BA,+105,NM1,fr	31	232	55	3768	24	18.4	0.2078 ±.08%	2.323 ±.10%	0.92	0.08107 ±.04%	1223.2 ±1.6
BB,+105,NM1,cd	38	88	20	4833	9	13.5	0.2073 ±.09%	2.315 ±.10%	0.91	0.08099 ±.04%	1221.2 ±1.7
Wanapitei Intrusion, SPA-89-31											
A,+105,NM1,cd,fr	57	288	94	10298	28	15.1	0.2915 ±.08%	4.187 ±.10%	0.96	0.10418 ±.03%	1699.9 ±1.1
BA,+105,NM1,cl,fr	14	240	80	3378	18	13.0	0.3038 ±.08%	4.430 ±.10%	0.92	0.10576 ±.04%	1727.6 ±1.5
BB,+105,NM1,cd,fr	38	311	107	5618	38	16.1	0.3031 ±.08%	4.420 ±.10%	0.95	0.10579 ±.03%	1728.0 ±1.2
CA,+149,NM1,cl,s	13	279	96	7030	10	14.9	0.3078 ±.08%	4.518 ±.10%	0.94	0.10648 ±.03%	1740.1 ±1.2
CC,+149,NM1,cd,s	16	1289	441	33261	11	17.6	0.2961 ±.10%	4.330 ±.11%	0.96	0.10607 ±.03%	1732.9 ±1.2
Falconbridge Twp. Gabbro, SPA-89-08A											
A,+105,NM5,o,s	11	566	274	308	455	23.2	0.3800 ±.12%	7.283 ±.37%	0.66	0.13902 ±.30%	2215.1 ±10.4
BA,-105,NM5,cd,s	2	3693	2154	15770	12	23.4	0.4472 ±.09%	9.811 ±.10%	0.96	0.15913 ±.03%	2446.4 ±1.0
BB,-105,NM5,o	9	772	419	6018	29	17.5	0.4463 ±.09%	10.248 ±.10%	0.96	0.16654 ±.03%	2523.2 ±1.0
CA,-105,NM5,o	3	962	597	2810	27	24.8	0.4608 ±.08%	11.167 ±.10%	0.94	0.17577 ±.04%	2613.4 ±1.2
CB,-105,NM5,cd,fr	5	134	72	1876	10	11.8	0.4696 ±.10%	11.289 ±.11%	0.92	0.17435 ±.04%	2599.9 ±1.5

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
[a] Numbers (i.e. +105, -149) indicate zircon size larger or smaller than given diameter, in microns
[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.
NMn = Non-magnetic at n degrees of tilt
cl = Clear grains; cd = Clouded grains; fr = Fractured grains; o = Opaque; s = Single grain analysis

to the Wanapitei intrusion farther north (Lumbers, 1975). Apart from Lumbers' regional mapping, a very few samples from this unit were included in a geochemical study of anorthosite genesis (Simmons and Hanson, 1978). The anorthosite and all of the surrounding rocks are typically gneissic and are strongly deformed, as is reflected in the outcrop pattern (Fig. 2). Sample SPA-89-22 was obtained from one of the northern extensions of the intrusion (UTM zone 17, 556850E, 5113100N), and is a relatively unaltered, albeit recrystallized rock consisting largely of andesine, less than 20 modal per cent amphibole, and accessory biotite, garnet and apatite.

Two distinct zircon morphologies were identified, consisting of extremely clear, subhedral grains with rounded tips and coarser, often fragmented anhedral, darker and more fractured grains (fractions BA and BB). Four zircon fractions were analyzed after moderate abrasion. The clear fractions had extremely low U concentrations (≤6 ppm U) and produced very low-precision data, but the data from the "poorer-quality" fractions mutually overlap just below concordia and indicate an average ^{207}Pb-^{206}Pb age of 1222.2 ± 2.0 Ma (Fig. 3).

Discussion

The 1222 Ma age is interpreted to be an igneous crystallization age for the Mercer anorthosite. The observation that the two high precision analyses are slightly discordant allows for some potential flexibility in the age if an episodic lead-loss event is suggested. Using a Grenville-aged (ca. 1000 Ma, arbitrarily) lower intercept, an upper intercept of ca. 1240 Ma can be obtained.

The age of 1222 ± 2 Ma for the Mercer Anorthosite is coincident with a relatively imprecise U-Pb zircon age of 1206 ± 18 Ma for the St. Charles Anorthosite (from additional work subsequent to Prevec and Baadsgaard, 1991a) which lies about 20 km to the north of Mercer Township. These two ca. 1220 Ma anorthosite ages are not coincident with any other mafic (or felsic) magmatism in the western Grenville Province. These ages slightly postdate ca. 1240 Ma intrusive ages for the Sudbury dyke swarm farther to the north (Krogh et al., 1987), and the Mulock Granite (Lumbers et al., 1991). Possible (as yet undated) local correlatives to the southwest include the Eau Claire Anorthosite, the Rutter Granite and the Bigwood Nepheline Syenite (Easton, pers. comm. 1992).

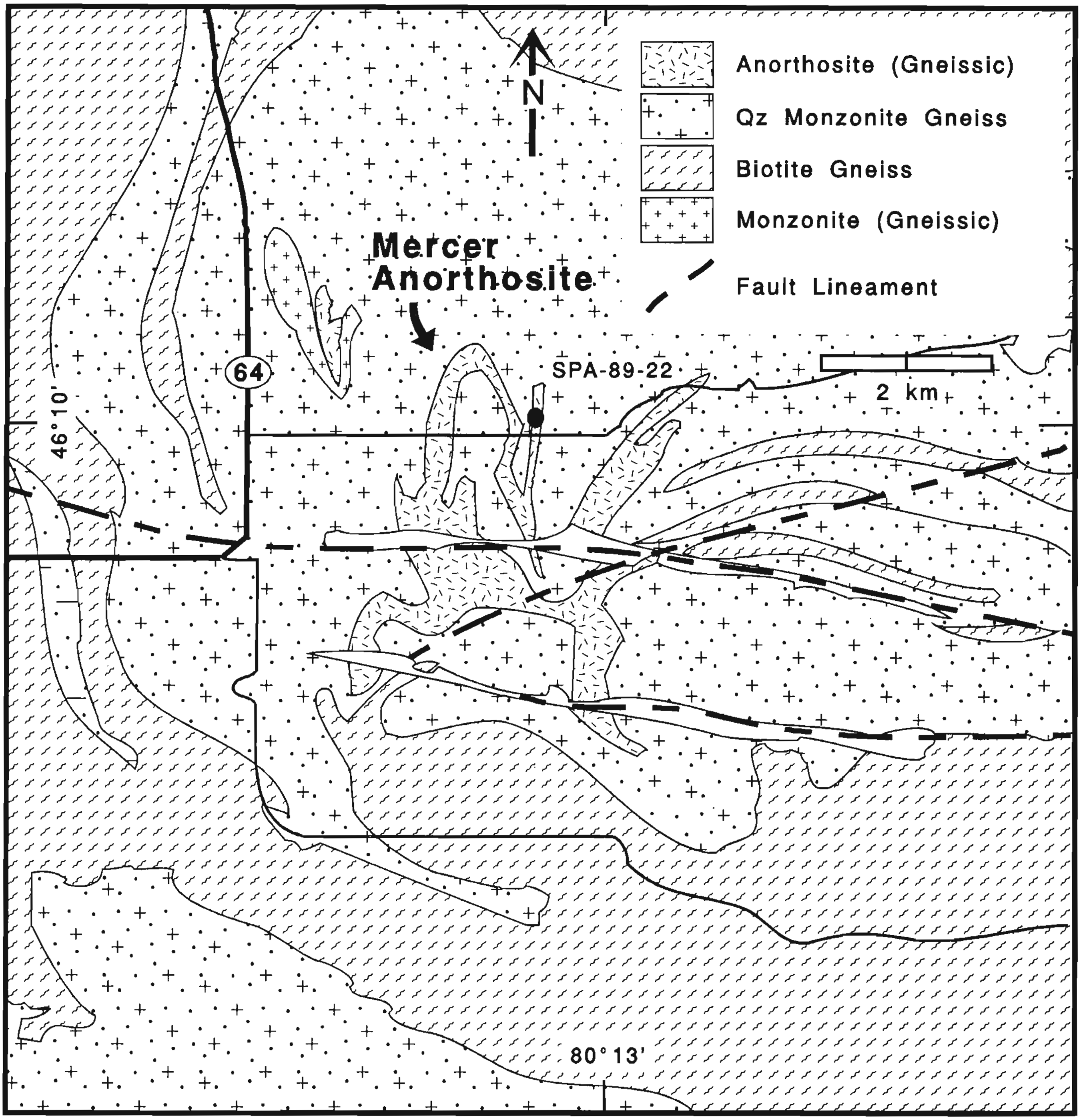

Figure 2. Location map for the Mercer Anorthosite sample (from Lumbers, 1975), showing Highway 64 and other access roads. Lakes are shown beneath fault traces, where present.

WANAPITEI INTRUSION (SPA-89-31)

Geology and geochronology

The Wanapitei intrusion lies immediately to the south of the Grenville Front Tectonic Zone and southwest of the River Valley pluton (Fig. 1). The intrusion was previously mapped by Lumbers (1975) and was the subject of a petrological-geochemical studies by Rousell and Trevisiol (1988). It is hosted by migmatitic biotite (para-) gneiss of the Grenville Province, and is cut by anorthosite on its southeast margin (Lumbers, 1975) (Fig. 4). This latter unit may be related to subsidiary gabbro-anorthositic intrusions mapped as River Valley material (e.g. Lumbers, 1975) which extend to the area immediately east of the Wanapitei intrusion in the form of the Red Deer anorthosite. The Wanapitei intrusion consists largely of granular, recrystallized two-pyroxene-plagioclase norite, or its amphibolitized equivalent, and is locally gneissic along the margins. Olivine cumulates showing coronitic textures are preserved. Sample SPA-89-31 is a plagioclase poikiloblastic hornblende rock from the hornblende norite zone of Rousell and Trevisiol (1988), from the south-central part of the complex (UTM zone 17, 515250E, 5144765N).

Zircons were typically subhedral to anhedral, stubby prismatic grains with aspect ratios varying from 2:1 to 5:1. Grains were variably fractured grains, commonly containing fine grained opaque inclusions. One single-grain fraction (CC) was euhedral. The zircons were strongly abraded and five fractions varying from as many as five grains (fraction A) to two of single grains were analyzed (fractions CA, CB). A four-point regression through the resultant discordia, shown in Figure 5, produces an upper intercept age of 1747 +6/-5 Ma, with a Grenvillian (ca. 996 Ma) lower intercept. This is interpreted as the crystallization age of the norite. One single-grain fraction which falls below the discordia (fraction CC) has a ^{207}Pb-^{206}Pb age of about 1733 Ma. This is consistent with a 1747 Ma crystallization age but with less Grenville-aged lead-loss than the other four fractions.

Discussion

The age of 1747 +6/-5 Ma for the Wanapitei intrusion correlates with the age of 1742.1 ± 1.4 Ma (van Breemen and Davidson, 1988) for the granitoid of the Killarney Complex some 60 km to the west. Both the Wanapitei and Killarney intrusive complexes are immediately adjacent to or cut by the Grenville Front Boundary Fault and/or mylonite (Davidson, 1986). These ages are comparable to those of ca. 1730 Ma granitoid orthogneiss from the northwestern Grenville Province (Krogh et al., 1971;), ca. 1700 to 1750 Ma granite plutonism in the Southern Province (Davidson et al., 1992), a ca. 1760 Ma granite-rhyolite terrane in Wisconsin, ca. 1700 to 1780 Ma orogenic material underlying southwestern and mid-continental North America (van Schmus and Bickford, 1981; van Schmus et al., 1987) and with ca. 1750 to 1720 Ma arc activity in southeastern Laurentica-Baltica (cf. Gower et al., 1991). The major lithological difference between the Wanapitei intrusion and other possible correlatives is that the latter suites consist entirely of felsic material (intrusive and extrusive), with the exception of rocks of the Colorado Province which include tholeiite-dominated bimodal volcanics (van Schmus and Bickford, 1981). If the Wanapitei

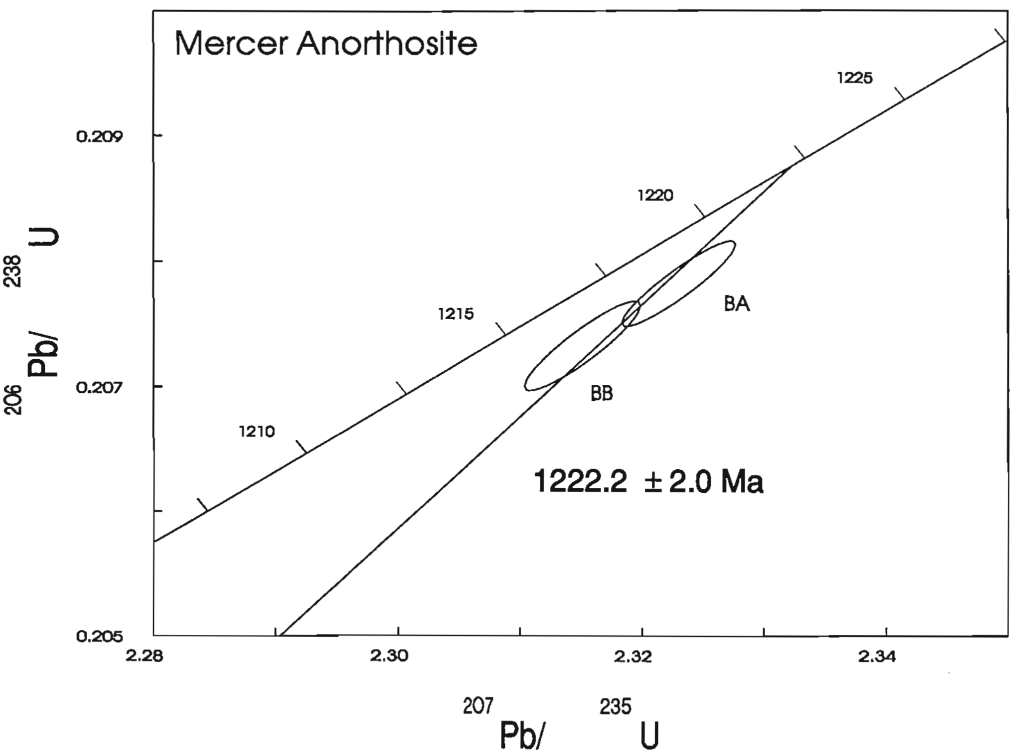

Figure 3.

U-Pb concordia plot for the Mercer Anorthosite, sample SPA-89-22. Error ellipses reflect the 2σ uncertainty.

intrusion is related to orogenic activity it represents the only manifestation of mafic magmatism associated with otherwise felsic plutonism and volcanism yet identified.

FALCONBRIDGE TOWNSHIP INTRUSION (SPA-89-08A)

Geology and geochronology

A metagabbroic intrusion in Falconbridge Township, just northeast of the Falconbridge mine was originally mapped as hornblende syenite (Thomson, 1957), and lies between norite of the Sudbury Intrusive Complex, Huronian mafic volcanics, and Superior Province granitoids, as shown in Figure 6. The Falconbridge Township intrusion consists of amphibolitized and saussuritized metagabbro, showing intercumulus clinopyroxene textures between plagioclase preserved as pseudomorphs. Late greenschist-facies metamorphism and quartz-filled fracturing postdates amphibolitization. This unit is petrologically identical to an intrusion in Drury Township near the southwestern corner of the Sudbury Structure (Fig. 1), which has a zircon age of 1854 ± 14 Ma (Prevec and Baadsgaard, 1991a). The intrusion is cut by late olivine diabase dykes (Thomson, 1957) which are members of the Sudbury dyke swarm and provide a minimum age for the Falconbridge gabbro of 1238 ± 4 Ma (Krogh et al., 1987). Both the Drury and the Falconbridge Township intrusions are cut by Early Proterozoic dykes (i.e. Card, 1965; Thomson, 1957, respectively) which may be Matachewan and/or possibly Nipissing Diabase (Thomson, 1957; Card, pers. comm. 1992) (ca. 2450; Heaman, 1989; ca. 2219 Ma; Corfu and Andrews, 1986, respectively). The Drury and Falconbridge bodies have experienced comparable metamorphic histories, based on petrographic evidence (Prevec and Baadsgaard, 1991b), and are much more extensively altered than the 1850 Ma (Krogh et al., 1984), relatively pristine Sudbury Norite (e.g. Naldrett and Hewins, 1984). Sample SPA-89-08A is a coarse grained amphibole-plagioclase rock with cumulate texture pseudomorphs well-preserved, from the north end of the intrusion (UTM zone 17, 517730E, 5162135N).

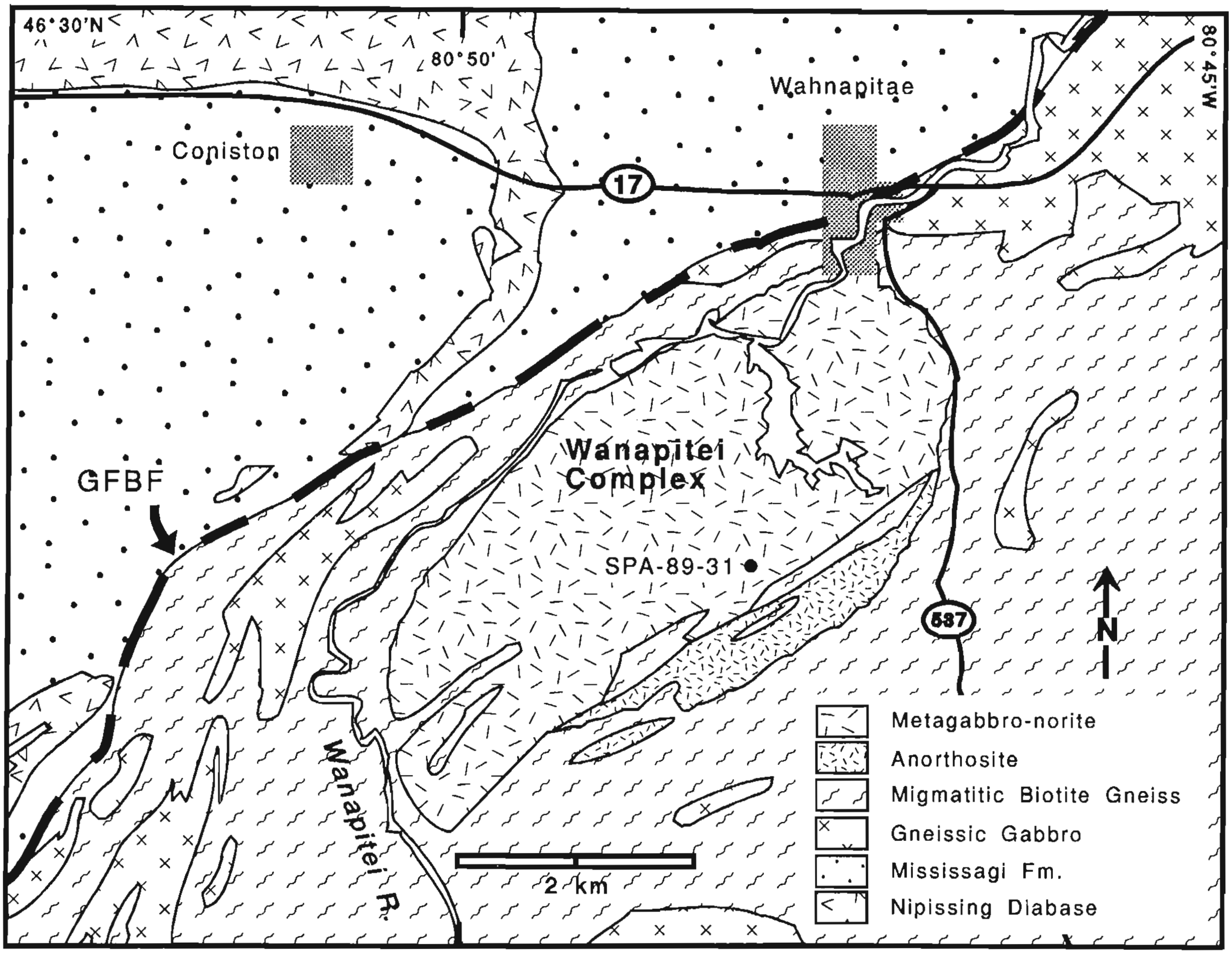

Figure 4. Location map for the Wanapitei Complex sample (from Dressler, 1984), showing the towns of Coniston and Wahnapitae and Highways 17 and 537 for reference. GFBF=Grenville Front Boundary Fault.

Zircons were without exception of extremely poor quality. Although euhedral grain shape was not uncommon, grains were red-brown and usually opaque or very weakly transluscent. Fraction BA, consisting of a single small grain, was relatively transluscent and devoid of obvious internal structure such as the zoning observed in grains in fraction BB. After light abrasion on most fractions (except BA which was heavily abraded), five fractions were analyzed. All fractions were highly discordant, (Fig. 7) with ^{207}Pb-^{206}Pb ages ranging from 2215 Ma to 2613 Ma. A regression through three points, used as a reference line, indicates an Archean upper intercept and a Sudbury-aged (ca. 1850 Ma) lower intercept. Fraction BA lies above this reference discordia and has a ^{207}Pb-^{206}Pb age of 2446 Ma. Fraction CA lies below the line, with a ^{207}Pb-^{206}Pb age of 2613 Ma.

Discussion

A large amount of Archean, probably ca. 2700 Ma inheritance is indicated by the zircon data. The data for fraction CA are consistent with an Archean provenance which has undergone relatively recent lead-loss. The proximity of this intrusion to the Sudbury Structure such that it is actually in contact with the Sudbury Norite on the west side (Thomson, 1957), supports ca. 1850 Ma lead-loss related to the intrusion of the Sudbury Igneous Complex. However, the data for fraction BA preclude straightforward interpretation in terms of Archean inheritance in a Sudbury-aged intrusion, and suggest a ca. 2450 Ma component. A line from 1850 Ma through fraction BA produces a geologically meaningless upper intercept at ca. 2550 Ma (indicated by dashed line in Figure 7). Unfortunately lack of exposure of contact relationships and the absence of up-to-date geological mapping preclude definitive interpretation of these zircon data based on field relationships for the Falconbridge Township. intrusion. The available information suggests that a ca. 2450 Ma intrusion age is possible, with prominent Archean inheritance and episodic Pb-loss at 1850 Ma as a result of the emplacement of the Sudbury Igneous Complex.

SUMMARY AND CONCLUSIONS

The age of the Mercer Anorthosite combined with contemporaneous intrusion of the nearby St. Charles Anorthosite suggests minor magmatic activity at ca. 1220 Ma in the western Grenville. The fact that possibly correlative units nearby include another anorthosite and a nepheline syenite suggests that this magmatism was anorogenic. The presence of euhedral, very low-U zircons may indicate later metamorphic effects, possibly of Grenvillian age.

Age data for the Wanapitei Complex suggest an affinity, temporally and perhaps tectonically, with the Killarney Complex and the potentially related granite-rhyolite terranes to the southwest. Although the Wanapitei Complex is a relatively minor intrusion, it represents a departure from the felsic magmatism which characterizes the other contemporaneous suites, in that it is gabbro-noritic without an associated felsic component.

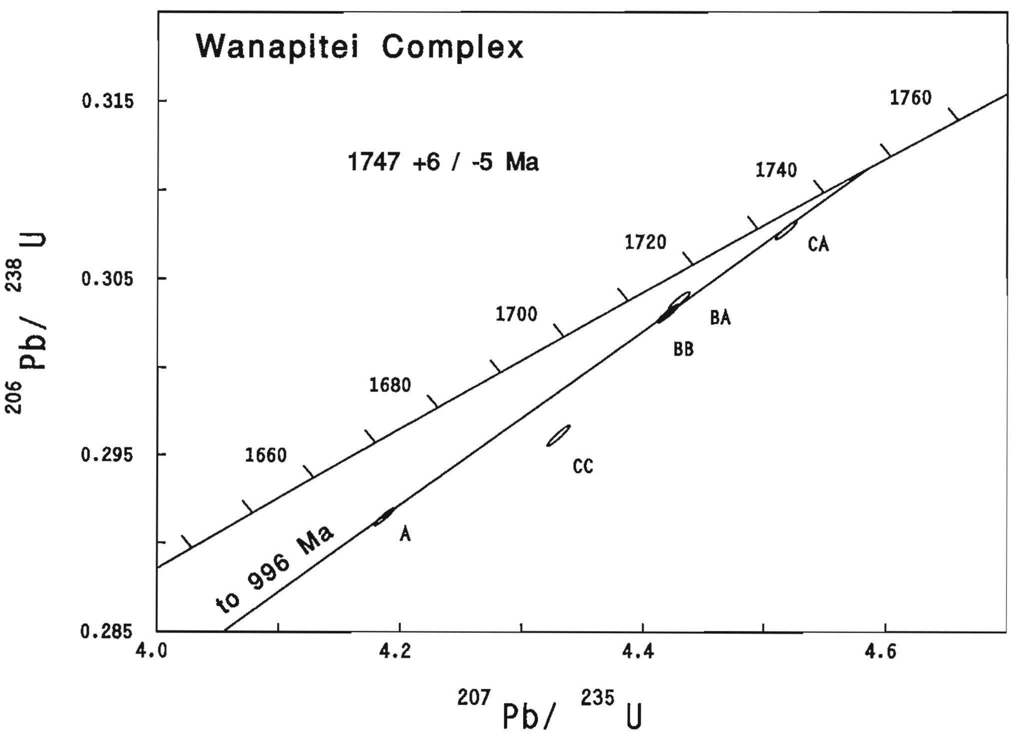

Figure 5.

U-Pb concordia plot for the Wanapitei Complex, sample SPA-89-31. Error ellipses reflect the 2σ uncertainty. The 4-point regression excluding fraction CC has an MSWD of 4.8.

The isotopic characterisitics of zircons from the Falconbridge Township intrusion infer an Early Proterozoic emplacement age with significant late Archean component, possible sources of which include Abitibi granitoids and detrital grains from Huronian sediments, or higher-grade Levack gneiss-type material (representing potential lower crust in the Sudbury area). The observation that the inheritance clearly predates 2450 Ma material favours an Archean age (as opposed to a younger Creighton-Murray Granite age) for the granitoid which hosts the intrusion on its east flank, although the inheritance may reflect deeper-level contamination. Further analyses of single, relatively clear grains is required to establish the nature of the Early Proterozoic contributions (both at 2450 and 1850 Ma) to these systematics.

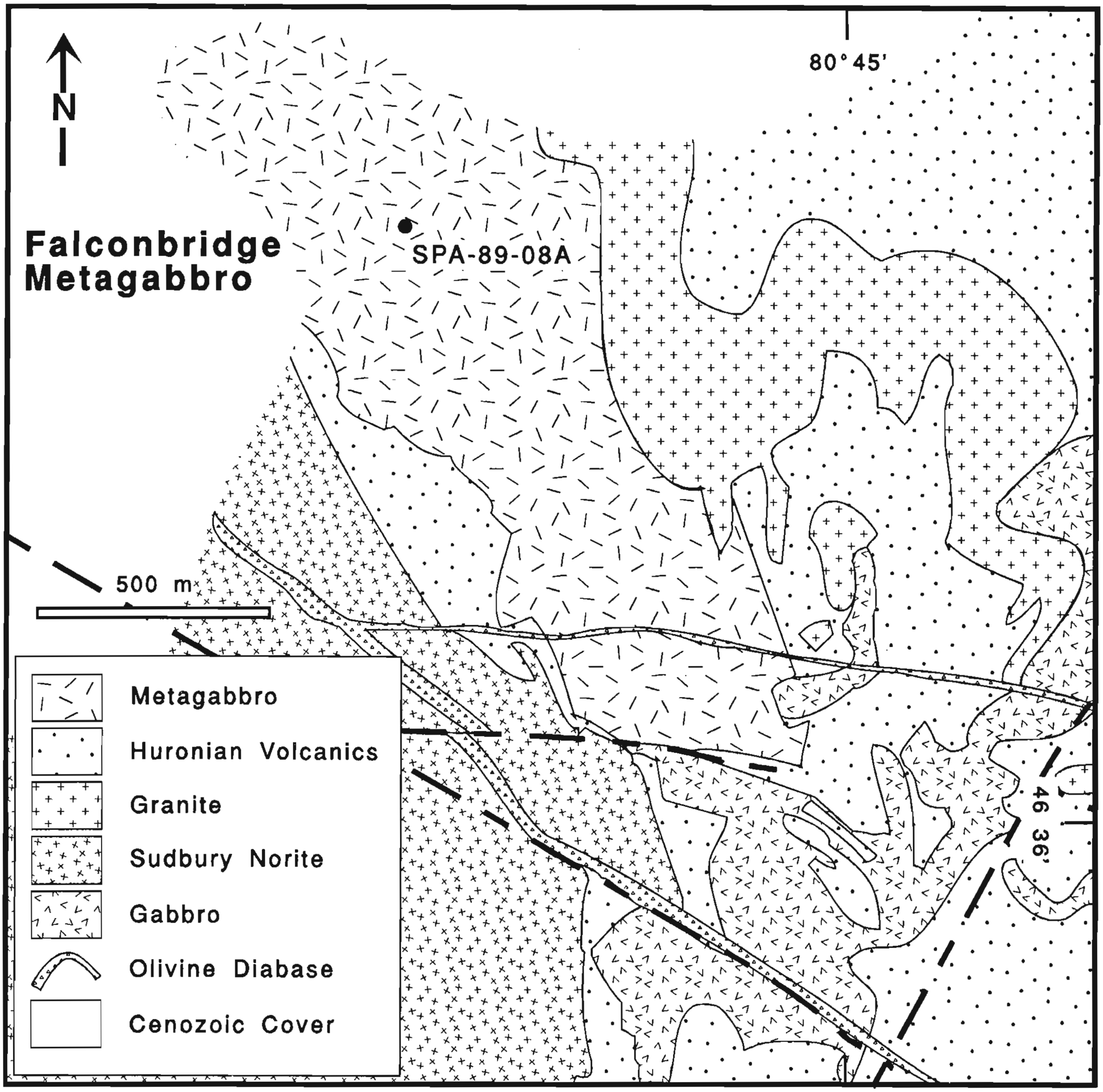

Figure 6. Location map for the Falconbridge Twp. metagabbro sample SPA-89-08A (based on Thomson, 1957). "Olivine diabase" and "Gabbro" are Sudbury dykes and Nipissing Diabase, respectively.

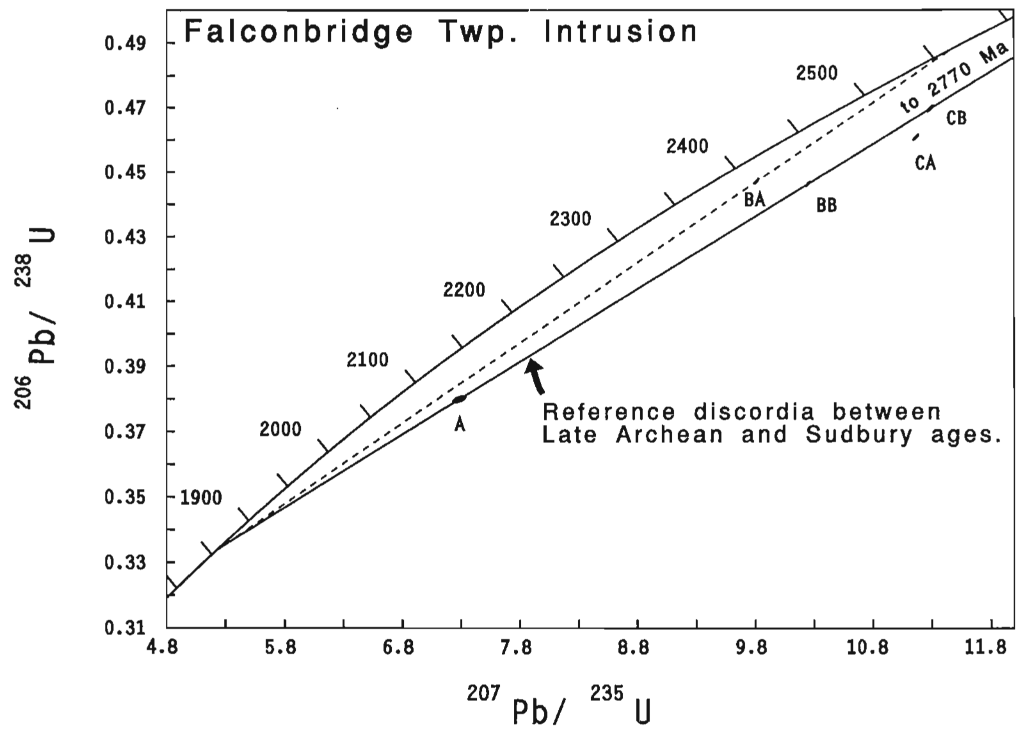

Figure 7.

U-Pb concordia plot for the Falconbridge Twp. intrusion, sample SPA-89-08A. Error ellipses reflect the 2σ uncertainty. The dashed line indicates a maximum age for fraction BA assuming 1850 Ma lead-loss (see text).

ACKNOWLEDGMENTS

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- 10 -

Circa 1.75 Ga ages for plutonic rocks from the Southern Province and adjacent Grenville Province: what is the expression of the Penokean orogeny?

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Davidson, A., van Breemen, O., and Sullivan, R.W., 1992: Circa 1.75 Ga ages for plutonic rocks from the Southern Province and adjacent Grenville Province: what is the expression of the Penokean orogeny?; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 107-118.

Abstract

It has been argued that folding and metamorphism of the Huron Supergroup in the Southern Province of Ontario was due to Penokean orogeny. Existing geochronology, however, does not record events in the range 1.89-1.83 Ga now assigned to this orogeny in its type area south of Lake Superior. The five new U-Pb ages of plutonic rocks in the Southern Province and in the adjacent Grenville Province, reported here, also fail to identify it. These ages lie in the 1.75-1.70 Ga range, comparable to those of previously dated plutons in the region and to the Yavapai and Mazatzal orogenies farther west in the mid-continent.

Résumé

Il a été établi que le Supergroupe de Huron, observé dans la province du Sud, en Ontario, avait été plissé et déformé au cours de l'orogenèse pénokéenne. La géochronologie actuelle, cependant, ne fait pas état d'événements datant de l'intervalle 1,89 à 1,82 Ga, maintenant attribué à cette orogenèse dans sa région type, au sud du lac Supérieur. Les cinq nouvelles datations par la méthode U-Pb qui sont présentées ici, effectuées sur des roches plutoniques de la province du Sud et de la province de Grenville adjacente, ne se situent pas dans l'intervalle susmentionné. Les âges sont compris entre 1,75 et 1,70 Ga et sont comparables à ceux obtenus dans le cas des plutons antérieurement datés dans la région et dans le cas des orogenèses de Yavapai et de Mazatzal, plus à l'ouest, dans le milieu du continent.

INTRODUCTION

Deformation and metamorphism of Paleoproterozoic supracrustal rocks in Michigan, Minnesota and northern Wisconsin are ascribed to Penokean orogeny (e.g. Cannon, 1973; Van Schmus, 1976; Medaris, 1983). Earlier assigned an age range of 1.9-1.7 Ga (Goldich et al., 1961) and equated with Hudsonian orogeny (in the sense of Stockwell, 1982), the Penokean orogeny is now generally considered to have ceased by about 1.83 Ga (Hoffman, 1989).

Although it has been recognized (Card et al., 1972) that the Huron Supergroup of the Southern Province in Ontario was folded before intrusion of Nipissing diabase at 2.2 Ga (Corfu and Andrews, 1986), an event named Blezardian

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orogeny by Stockwell (1982), much of the folding and metamorphism in these rocks has been assumed to have been caused by Penokean orogeny, and the region has been commonly referred to as part of the Penokean fold belt (e.g. Card et al., 1972; Card, 1978; Zolnai et al., 1984; Hoffman, 1989). This interpretation was made, in part, on the basis of 2.2 - 1.8 Ga Rb-Sr whole rock ages (Fairbairn et al., 1969). Correlation between Ontario and Michigan across a 200 km interval covered by mid-continental rift deposits (≈1100 Ma) and the northern margin of the Paleozoic Michigan basin, however, may not be justified for the reasons that follow: 1) The supracrustal rocks in the two regions, despite earlier suggestions of stratigraphic correlation (see discussion in Morey, 1973), are now known to be different in age (Huron Supergroup <2.47 Ga >2.2 Ga, Marquette Range Supergroup < 2.1 Ga >1.85 Ga); 2) Orogenic igneous rocks of Penokean age (sensu stricto) have not been identified in Ontario. The only intrusive unit whose primary age is in the Penokean range is the Sudbury irruptive (1850 ± 1 Ma; Krogh et al., 1984); as this unit is now widely accepted to have been generated by extraterrestrial impact, it can have no relationship to orogeny; 3) K-Ar and Rb-Sr ages obtained from the Huron Supergroup and other rocks in the Ontario Southern Province do not cluster in the Penokean age range; most are younger, and as many are older as lie within the Penokean range (Stockwell, 1982).

To what event, or events, then, can be assigned the deformation, plutonism and metamorphism that characterize the Southern Province fold belt in Ontario? The 1.85 Ga Sudbury event is a local time marker that allows recognition that the adjacent Huron Supergroup was previously deformed. A possible clue to pre-Nipissing (Blezardian) orogeny may be given by the Creighton, Murray and Skead granite plutons, which lie adjacent to the south range of the Sudbury irruptive (Dressler, 1984). A U-Pb zircon age of 2333 +33/-22 Ma was obtained for the Creighton granite by Frarey et al. (1982); and 2388 +20/-13 Ma for the Murray granite by Krogh et al. (1984). These granites were deformed before intrusion of the Sudbury irruptive, but the relative age of this deformation with respect to intrusion of Nipissing diabase has not been reported.

All of the other dated plutons that intrude Huronian rocks of the Southern Province fold belt have given ages younger than 1.75 Ga. These are (Fig. 1): Cutler, ≤1.73 Ga (Rb/Sr, Wetherill et al., 1960); Mongowin, 1.73 Ga (Rb-Sr, Van Schmus, 1971); Croker Island, 1.47 Ga (Rb-Sr, Van Schmus, 1965); Chief Lake, 1.75 Ga (Rb-Sr, Davis et al., 1966) and 1.72 Ga (recalculated U-Pb, Krogh and Davis, 1968); Bell Lake, 1471 ± 3 Ma (U-Pb, van Breemen and Davidson, 1988); Killarney, 1742 ± 1.4 Ma (granite) and 1733 +7/-6 Ma (porphyry) (U-Pb, van Breemen and Davidson, 1988); a Rb-Sr isochron age of 1623 ± 74 Ma for the Killarney granite was reported previously by Wanless and Loveridge (1972). Van Schmus et al. (1975) obtained a U-Pb zircon age of 1500 ± 20 Ma for granite intersected in drill core beneath Paleozoic rocks on Manitoulin Island. Attempts to obtain a more satisfactory age for the Cutler granite and an age for the previously undated Eden Lake pluton are reported herein. Of the only other two small granite plutons in the Southern Province (Fig. 1), the Balsam Lake pluton has a K-Ar muscovite age of 1215 ± 45 Ma (Frarey, 1985), and the Daisy Lake pluton has not been dated. Despite the wide range in reliability of the above ages, it is clear that there were three distinct episodes of granite plutonism in the Ontario Southern Province: namely at ≈2.35 Ga, 1.75-1.70 Ga and 1.50-1.45 Ga; granites of Penokean age have yet to be identified. Most granitoids near the Grenville Front yield wholly or partly reset K-Ar ages (see Stockwell, 1982; Easton, 1986); the K-Ar age of the Balsam Lake stock is probably no exception.

Mylonitization along southeast-dipping zones that mark the Grenville Front, the northwest margin of the late Mesoproterozoic Grenvillian orogen, has affected the ≈1.75 and ≈1.45 Ga plutons that lie along it. It has proven difficult to correlate Huron Supergroup formations of the Southern Province with metasedimentary gneiss units in the immediately adjacent Grenville Province (Frarey, 1985), but it has been known for some time that granitoid orthogneiss in the same zone has an age of ≈1.73 Ga (Krogh et al., 1971). A 1746 +6/-5 Ma age for the mafic Wanapitei Complex just southeast of the Grenville Front is presented by Prevec (1992). Farther into the interior of the Grenville Province metaplutonic rocks are identified with ages in the 1.50-1.45 Ga range as well as thoroughly reworked granitoid rocks with ages of ≈1.7 Ga (Lumbers, 1975; van Breemen et al., 1986; Corrigan, 1990). New isotopic age data for three metaplutonic units of the older group in the Grenville Front Tectonic Zone are presented in this paper.

Analytical techniques of zircon and titanite analysis are described in Parrish et al. (1987, 1992). In order to remove thin metamorphic overgrowths and improve concordance of the isotopic systems, zircons have been abraded in the manner described by Krogh (1982). Treatment of analytical errors follows Roddick (1987) and regression analysis York (1969). Isotopic data are presented in Table 1.

SOUTHERN PROVINCE

Samples were collected from the Cutler and Eden Lake plutons for dating by the U-Pb method. The rationale was to confirm the earlier Rb-Sr age obtained for the Cutler batholith and to compare this with the age of the similar trondhjemitic phase of the previously undated Eden Lake pluton. In summary, a U-Pb titanite age was obtained for the Cutler batholith (Fig. 2a), in good agreement with the earlier Rb-Sr age. Inheritance was found in zircons from the Cutler batholith and only a few fractions were analyzed. The zircons from the Eden Lake pluton proved to be varied, indicating a multi-event history and a large component of Archean inheritance. While a tentative U-Pb zircon age was obtained for the pluton (Fig. 2b), titanite is secondary and contained too little uranium (ca. 1 ppm) for a reliable age.

Cutler batholith

The Cutler batholith is exposed along the north shore of North Channel (Lake Huron) between the villages of Cutler and Algoma Mills. It is situated in the core of a major anticline in the Huron Supergroup, where it intrudes the McKim Formation (Giblin et al., 1979), formerly a thin-bedded shale which has been converted to garnet-staurolite mica schist in the vicinity of the batholith (Cannon, 1973). The plutonic rocks range from grey biotite tonalite to pink 2-mica granite; both are relatively fine grained and are commonly associated with pegmatite. Contacts with the McKim schists are complicated by veining and included screens, particularly along the crest of the anticline where the exposure of the batholith narrows to the east.

The sample was collected from a low, blasted outcrop on the south side of the gravel access road to Serpent River Indian Reserve, 1 km from its intersection with Highway 17, west of Cutler village. This outcrop is free of inclusions of metasediment and of pegmatite. The rock is fresh, grey, equigranular biotite tonalite with a very weakly developed foliation. Thin section examination reveals that quartz is recrystallized but not notably strained, and that a small amount of secondary muscovite and chlorite is associated with biotite. Accessory minerals are pyrite, allanite, apatite,

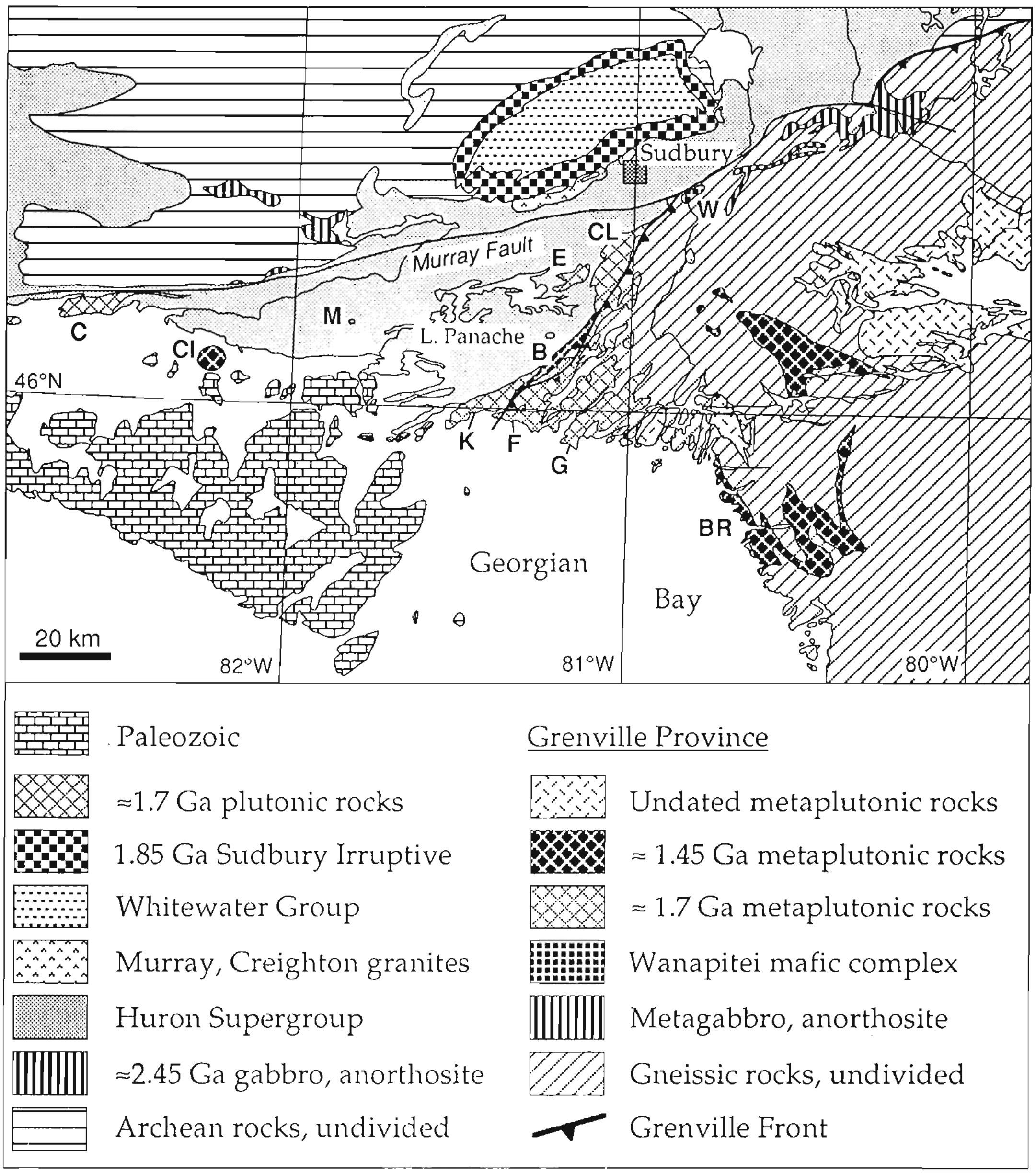

Figure 1. Generalized geology of the Sudbury region. Dated plutonic rock units mentioned in the text are identified as follows: **C** - Cutler, **Cl** - Croker Islands, **M** - Mongowin, **K** - Killarney, **B** - Bell Lake, **E** - Eden Lake, **CL** - Chief Lake, **W** - Wanapitie Complex, **F** - Fox Islands, **G** - Grondine, **BR** - Britt.

titanite and minor zircon. Zircon has the form of elongate needles (length to breadth ratio ≈5:1) and some appear markedly metamict.

The zircon population contains several types but the bulk consists of a tan, anhedral variety with rounded terminations, coarse euhedral zoning, and minor bubble and rod inclusions. These are considered to be primary igneous zircons (fraction C), however, very obvious, rounded, clear cores, some with bubble and rod inclusions were noted in these tan zircons, indicating an inherited component. In addition, minor clear overgrowths and rounded buds on the ends of some of the zircons indicate a secondary zircon growth event. Fraction C has a high uranium content (1636 ppm) (Table 1) and is very discordant with a ^{207}Pb-^{206}Pb age of 1701 +/- 16 Ma (Fig. 2a). The latter is taken as a minimum age for the intrusion of the batholith. A clear, colourless, generally rounded type of zircon with rod and bubble inclusions (eg. fraction F, ^{207}Pb-^{206}Pb age of 2321 Ma) probably represents the cores seen in the bulk zircon. These and a dark brown

Table 1. Isotopic data

Fraction,[1] size	Weight (mg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$ [3]	Pb[4] (pg)	^{208}Pb[2] %	$^{206}Pb/^{238}U$ [5]	$^{207}Pb/^{235}U$ [5]	R	$^{207}Pb/^{206}Pb$ [5]	Age(Ma)[6]
Cutler batholith (88-DM-14a; 46° 12' N, 82° 29' W)											
C,-74+62,5:1	0.014	1636	419	274	1110	17.1	0.2232(.14)	3.209(.50)	0.67	0.10426(.42)	1701.3(16)
F,+105,cl,euh,S	0.004	365	155	2609	14	11.2	0.3822(.13)	7.790(.14)	0.97	0.14783(.04)	2320.9(1.2)
H,-105+74,5:1,S	0.003	475	206	2660	12	4.9	0.4113(.15)	9.484(.15)	0.98	0.16725(.03)	2530.3(1.1)
A,Tit,darkest,NA	0.107	160	52	571	551	11.6	0.3021(.10)	4.403(.23)	0.67	0.10569(.18)	1726.3(6.4)
X,Tit,dark,NA	0.150	124	41	886	385	11.6	0.3073(.10)	4.499(.18)	0.76	0.10618(.12)	1734.8(4.4)
Y,Tit,dark	0.085	133	44	506	415	11.5	0.3080(.11)	4.518(.26)	0.68	0.10639(.21)	1738.5(7.5)
Z,Tit,light,flat,NA	0.142	127	42	877	372	11.3	0.3049(.09)	4.460(.17)	0.74	0.10610(.12)	1733.4(4.4)
B,Tit,lightest,NA	0.160	52	15	485	305	3.8	0.2883(.13)	4.029(.27)	0.70	0.10137(.20)	1649.3(7.3)
Eden Lake pluton (87-DM-359; 46° 18' 24" N, 81° 06' 44" W)											
AA,-149, cl,S	0.002	51	35	244	13	27.1	0.4891(.72)	12.460(.74)	0.98	0.18477(.15)	2696.1(5.1)
CC,-105+74,cl,S	0.002	205	111	449	28	9.0	0.4877(.39)	11.920(.41)	0.96	0.17726(.11)	2627.4(3.6)
EE,-105+74,tl,S	0.002	408	213	2230	12	7.1	0.4704(.20)	12.742(.21)	0.99	0.19644(.03)	2796.9(1.1)
GG,-105+74,cl,S	0.004	190	121	1203	19	8.6	0.5577(.25)	16.206(.26)	0.99	0.21076(.04)	2911.4(1.2)
LL,-62,cl,euh	0.005	9	5	145	11	7.9	0.4670(.65)	11.014(.70)	0.80	0.17106(.43)	2568.0(14)
KK,-37,>5:1,NA	0.014	636	173	5128	26	9.9	0.2571(.08)	3.794(.10)	0.95	0.10702(.03)	1749.3(1.2)
ST,-37,<=3:1,NA	0.004	512	162	66	689	9.3	0.2984(.53)	4.970(2.2)	0.70	0.12082(1.9)	1968.3(66)
QR,-37,4:1,NA	0.004	770	217	1322	38	9.6	0.2667(.09)	4.050(.12)	0.82	0.11012(.07)	1801.4(2.5)
AB,-37,4:1	0.003	559	165	890	33	8.4	0.2814(.10)	4.570(.14)	0.77	0.11779(.09)	1923.0(3.3)
HH,-37,4:1	0.010	808	258	4712	32	7.8	0.3066(.09)	4.899(.10)	0.93	0.11590(.04)	1893.9(1.3)
JJ,-37,4:1	0.008	648	208	3373	29	8.0	0.3074(.08)	4.919(.10)	0.93	0.11605(.04)	1896.2(1.4)
Fox Islands granite (87-DM-339; 45° 56' 51" N; 99° 21' 50" W)											
A,-105+74	0.008	416	124	3811	15	9.6	0.2839(.09)	4.052(.10)	0.95	0.10353(.03)	1688.3(1.2)
B,-105+74	0.008	588	170	7642	15	8.3	0.2797(.10)	3.949(.11)	0.95	0.10242(.03)	1668.4(1.3)
C,-74+62	0.009	569	168	6473	14	9.3	0.2825(.08)	4.045(.10)	0.95	0.10386(.03)	1694.2(1.2)
E,-62	0.009	733	224	9864	12	9.5	0.2912(.09)	4.153(.10)	0.94	0.10344(.03)	1686.8(1.2)
Granite orthogneiss, "Grondine complex" (87-DM-266; 45° 55' 20" N; 99° 9' 13" W)											
O,-105+74	0.002	708	212	3429	9	8.1	0.2896(.12)	4.131(.13)	0.95	0.10346(.04)	1687.1(1.6)
P,-105+74,stubby	0.004	446	137	4133	7	8.7	0.2953(.09)	4.237(.11)	0.96	0.10407(.03)	1697.9(1.1)
Q,-74+62,equant	0.007	403	125	5673	9	9.2	0.2961(.09)	4.256(.10)	0.95	0.10424(.03)	1700.9(1.1)
R,-74+62	0.008	385	122	1768	32	9.6	0.3019(.10)	4.357(.12)	0.88	0.10469(.06)	1708.9(2.1)
Quartz dioritic orthogneiss, "Grondine complex" (87-DM-273; 45° 55' 46" N; 99° 10' 35" W)											
W,+105	0.015	285	96	762	101	13.3	0.3074(.55)	4.437(.65)	0.90	0.10469(.28)	1709.0(11)
X,-105+74	0.023	231	70	4428	21	10.9	0.2851(.09)	4.037(.10)	0.94	0.10270(.04)	1673.6(1.3)
Y,-74+62,equant	0.010	219	73	2049	20	15.1	0.2970(.10)	4.267(.11)	0.94	0.10421(.04)	1700.3(1.5)
Z,-74+62	0.019	354	103	6183	18	11.5	0.2720(.09)	3.803(.10)	0.95	0.10141(.03)	1650.1(1.2)

Notes: [1] mineral phase is zircon unless marked Tit = titanite; crystals are abraded unless marked NA = not abraded; sizes in microns before abrasion, e.g. -74+62; less than and greater than 74 microns and 62 microns, respectively; S indicates analysis of a single crystal; L:B = length to breath ratio, e.g. 5:1; euh = euhedral; cl = clear; tl = translucent; [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] ^{207}Pb-^{206}Pb model age; errors are two sigma in Ma.

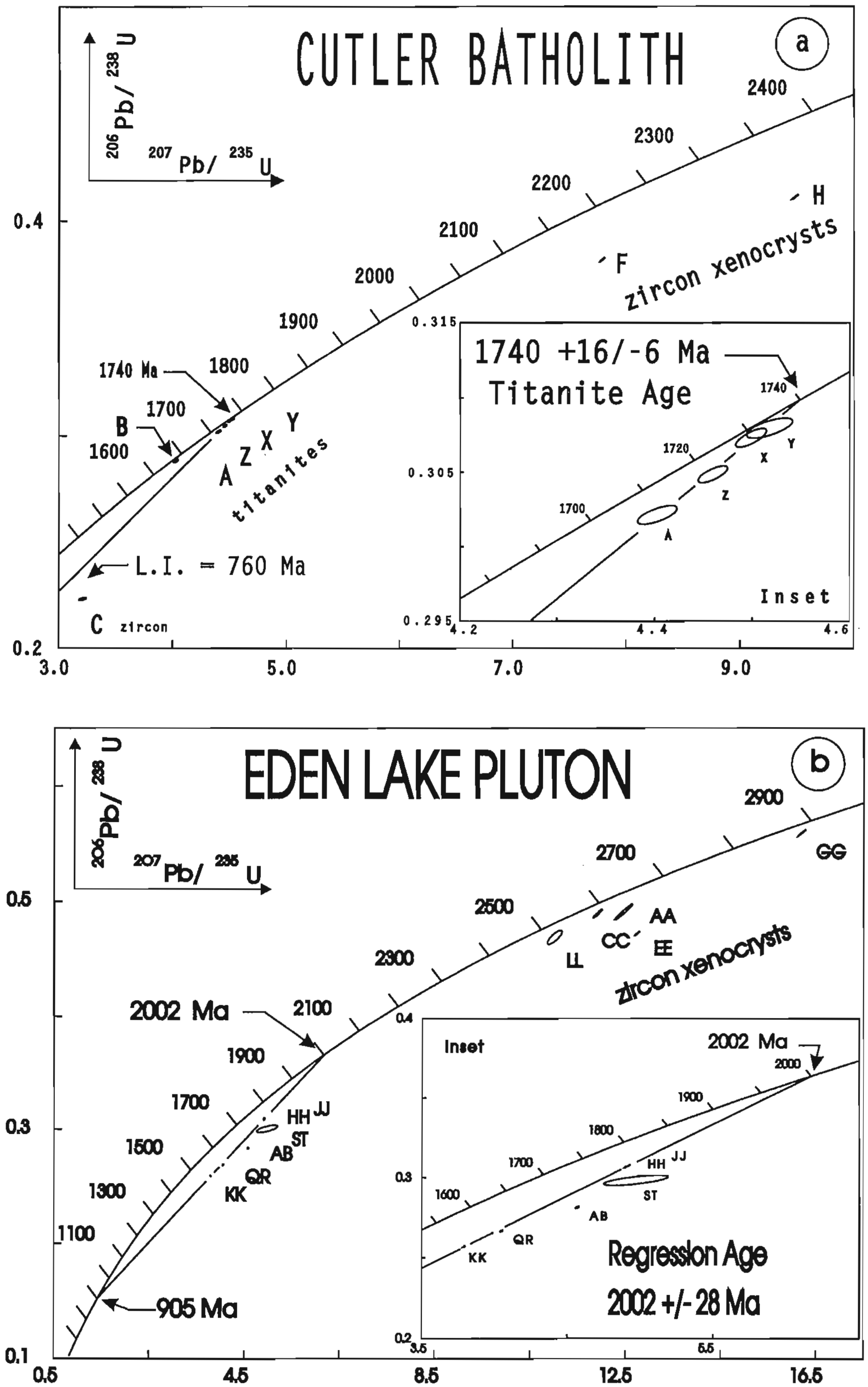

Figure 2. Concordia diagrams for plutons in the Southern Province: **a**, the Cutler batholith; **b**, Eden Lake pluton.

variety (possibly fraction H, ^{207}Pb-^{206}Pb age of 2530 Ma) are considered to be zircon xenocrysts, probably of different ages, assimilated from the country rock. A few round, equant pitted grains provide evidence that at least some of the zircons are of detrital origin. Only three zircon fractions were analyzed in favour of the abundant titanite in the sample.

The titanite consists mostly of blocky, clear, often lustrous fragments, but crystal facets were also observed. The colour ranged from dark to very light brown. A few fragments had both dark and light phases with a sharp boundary between. Fraction B consisted of clear, very light brown fragments and gave younger apparent ages (^{207}Pb-^{206}Pb age of 1649 Ma) than the darker titanite, suggesting new titanite growth.

Four fractions (A, X, Y and Z) of the darker titanite were analyzed and yielded a somewhat discordant but linear data set (mean square of weighted deviates, MSWD = 0.55) that is attributed to recent Pb loss. Regression gave a lower intercept age 760 Ma and upper intercept age of 1740 +16/-6 Ma which is interpreted to be the cooling age of the titanite (Fig. 2a). This age may be slightly younger than the crystallization age of the Culter batholith.

Eden Lake pluton

The Eden Lake complex lies to the west of the Chief Lake batholith and extends as far as Lake Panache (Fig. 1). It was mapped by Card et al. (1975), who described it as "a number of mafic to felsic plutons, sills, and dikes which intrude the Huronian metasediments and Nipissing-type gabbroic rocks over an area of about 42 km^2 ..." (ibid., p. 21). According to the geological map of Card et al. (1975) the predominant rock types are trondhjemite and quartz monzonite, with gabbro, diorite, granodiorite and aplite occurring locally as small bodies. Like the Cutler batholith, these plutonic rocks are intimately involved with their country rocks.

The sample was collected from a large outcrop of uniform, grey biotite trondhjemite exposed at the shore of a bay in the northwest part of Wavy Lake, at the northeastern-most extremity of the complex. Included rafts of Mississagi sandstone of the Huronian Supergroup are present nearby, and the tonalite intrudes Nipissing gabbro to the south. In thin section the quartz and biotite are fairly strongly strained and the albitic plagioclase has irregular, patchy extinction. Allanite, apatite, titanite and zircon are accessory phases. Epidote, chlorite and secondary titanite are associated mainly with biotite. Fractured garnet grains occur sparingly. Zircon occurs as stubby prisms (≈2:1), larger, cored grains being enclosed by biotite and smaller clear prisms by quartz. The rock has clearly been affected by post-crystallization strain; whether this is an effect of intrusion of the neighbouring Chief Lake batholith or is a more distal effect of younger Grenvillian tectonics is not known.

The zircons consist of such a variety of habits, colours, internal and external features that classification into meaningful groups is difficult. There is, furthermore, abundant morphological evidence for inheritance in terms of obvious core-overgrowth relationships, particularly in the larger grains, all of which indicates multi-event history(s). Some zircons consist of a colourless core with bubble inclusions, surrounded by tan zircon with euhedral zoning, and an outer clear, unzoned, rounded mantle. Other grains have clear colourless cores and overgrowths. Zircons with euhedral twin terminations were noted as were anhedral buds. Rare parallel twins were also seen. Evidence for a detrital origin of some grains is provided by round, pitted zircons.

On a concordia diagram (Fig. 2b), the data plot in two separate groups with six close to a chord between 2.0 Ga and 0.9 Ga, and five with Archean ^{207}Pb-^{206}Pb ages ranging from 2.56 Ga to 2.91 Ga. Four of these Archean ages are for single grains of which EE and GG represent a dark pink type, whereas grains AA and CC characterize an abundant population of large clear euhedral zircons. The two zircons in fraction LL are fine, clear, euhedral and of "gem" quality. Although there are no visible cores their euhedral habit and multiply faceted terminations result from an overgrowth of undetermined age. It is assumed that abrasion has removed minor overgrowths on the Archean zircons analyzed.

The younger group (fractions AB, HH, JJ, KK, ST and QR) are interpreted to have formed largely during magmatic crystallization and emplacement of the Eden Lake pluton. These zircons are finer, light purple-tan, generally acicular with anhedral longitudinal zoning and perpendicular fractures. This population also contains a few grains that have been broken and subsequently healed with growth of new zircon. Also noted in the finer acicular population were bent and cracked crystals with elbow form that indicate post-crystallization strain. As no cores were visible in the analyzed fractions, the linear trend generated by these younger zircons is best interpreted in terms of an upper concordia intercept indicating the age of emplacement and the lower intercept indicating a time of Pb loss. Lead loss was, apparently, surface correlated as fraction KK contained the most acicular zircons (L:B >= 5:1, many of which were thin and platy) which were not abraded, whereas fractions HH and JJ were strongly abraded.

The essentially duplicate results of fractions HH and JJ are taken as evidence for a lack of inheritance and the following age estimates are based on this assumption. An estimate is taken from a regression of the four fractions HH, JJ, QR and KK yielding upper and lower inter- cepts of 2002 ± 28 Ma and 905 ± 65 Ma, respectively. This gives a MSWD of 71 indicating significant geological scatter, due primarily to the discordance of points KK and QR. A three point regression of fractions HH, JJ and KK yields identical intercept ages, whereas a three point regression of fractions QR, HH and JJ gives upper and lower intercepts of 1980 Ma and 813 Ma respectively. Thus the age of the Eden Lake pluton is tentatively assigned at 2002 with a realistic error of ±28 Ma. This age is consistent with the maximum age for the complex provided by the 2219 ± 4 Ma age for the Nipissing diabase (Corfu and Andrews, 1986). The analytical results of fractions AB and ST of more equant zircons, which plot to the right of the above regression line, are attributed to a minor component of cryptic Archean cores.

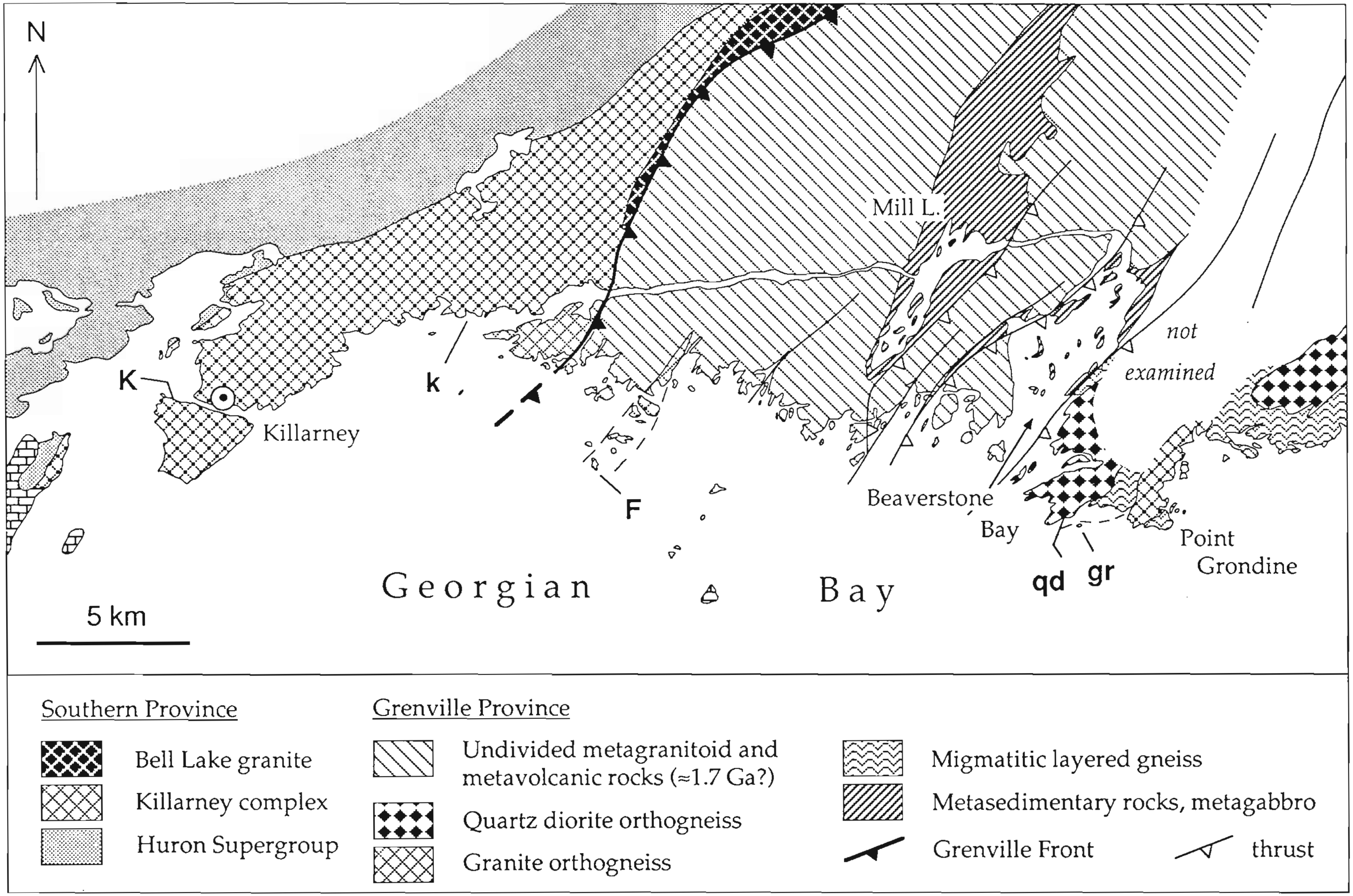

Figure 3. Simplified geology in the vicinity of the Grenville Front, north shore of Georgian Bay, showing sample sites for the Fox Islands granite (**F**) and the Grondine complex (**gr** - Hen Island granite, **qd** - quartz diorite orthogneiss). Also shown are sample sites in the Killarney complex (**K** - granite, **k** - porphyry) for which ages were reported in van Breemen and Davidson (1988).

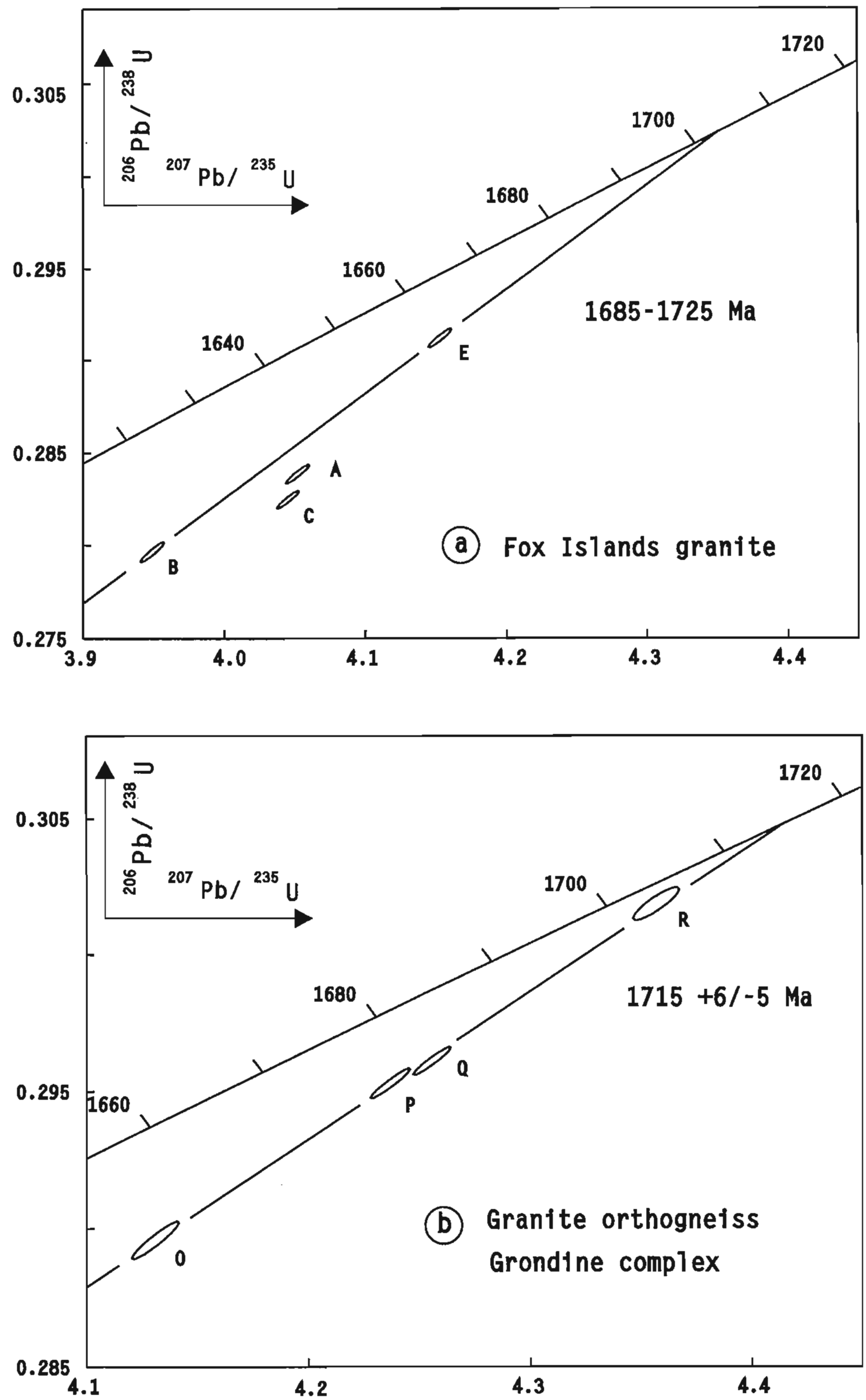

Figure 4. Concordia diagrams for plutons from the Grenville Front Tectonic Zone: **a**, Fox Islands granite **b**, granite orthogneiss, Grondine complex; **c**, quartz dioritic orthogneiss, Grondine complex.

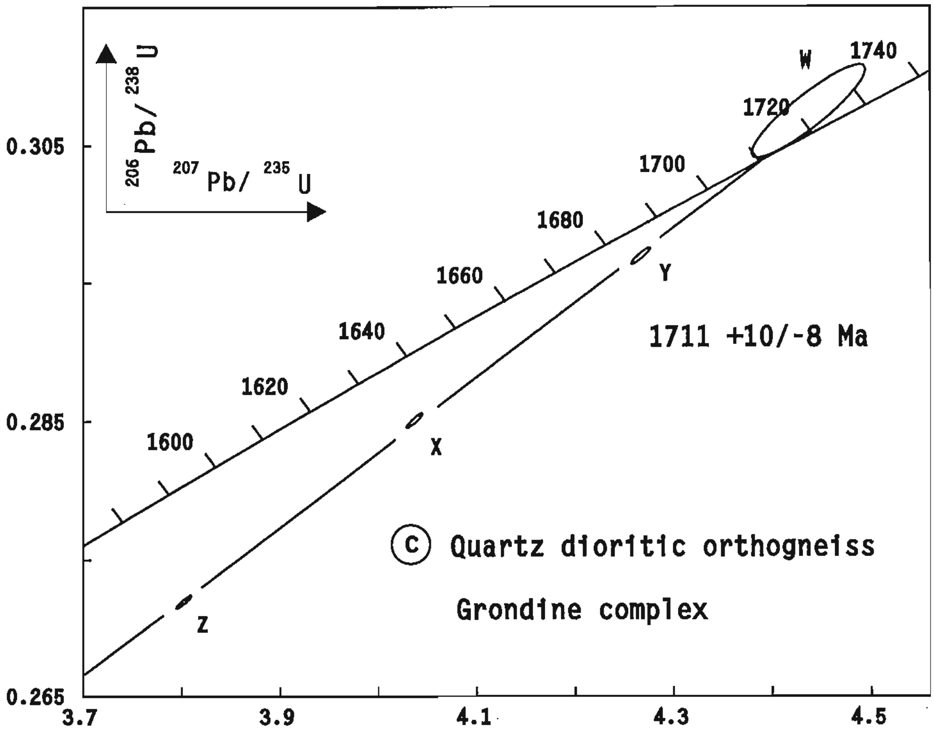

Figure 4. (cont'd.)

An alternative interpretation is that all of the younger fractions contain an inherited component and that the 2.0 Ga to 0.9 Ga linear trend is fortuitous. If this were the case, the age of emplacement could be younger than the youngest ^{207}Pb-^{206}Pb age of 1749 Ma, that of fraction KK. Further work is required.[1]

GRENVILLE FRONT TECTONIC ZONE

Most of the rocks in the Grenville Front Tectonic Zone exposed along the north shore of Georgian Bay east of Killarney are of plutonic origin. The following summary is taken from Davidson and Bethune (1988), and illustrated in Figure 3. Foliated to gneissic granite bodies within a few kilometres of the Grenville Front lie within fine grained sugary to foliated feldspathic gneisses, some of which exhibit shadowy lenticles with slightly different colour indices that may represent volcanic clasts. This assemblage is reminiscent of the association of rhyolite, tuff, hypabyssal porphyry and granite that characterizes the Killarney complex immediately west of the front, which has been dated at ≈1740 Ma (van Breemen and Davidson, 1988). Metasedimentary rocks, including quartzite and pelitic schist, are exposed in and around Mill Lake and Beaverstone Bay, but do not reach the outer coast (Fig. 3). A thrust sheet of protomylonitic granodiorite lies structurally above the metasedimentary rocks in Mill Lake and below those in the northern part of Beaverstone Bay. To the south, another thrust sheet intervenes; it consists of small-folded red granite enclosing a small metagabbro to monzodiorite pluton. Exposed east of the mouth of Beaverstone Bay are more thoroughly recrystallized granitoid rocks of various kinds, interleaved with highly migmatitic rocks of uncertain affinity; this association is referred to here as the Grondine complex.

Fox Islands granite

A sample of pink, foliated granite was collected from West Fox Island, 3 km southeast of the pronounced mylonite zone that marks the Grenville Front at the south shore of Philip Edward Island (Fig. 3). The Fox Islands granite has the form of a sheet, no more than a kilometre thick and tapering north-eastward, within pink, sugary leucogneiss that was probably derived from felsic volcanic rocks or hypabyssal intrusions. In most places it has a moderately well developed foliation inclined about 60° southeast and carrying a dip-parallel lineation imparted by streaks of recrystallized biotite. Once medium grained, the granite's quartz and feldspar are now recrystallized to fine lenticles and sugary aggregates respectively, but at the sample site small augen of original feldspar remain. This is reflected by the seriate texture seen in thin section. Elongate quartz grains are composed of low-strain polygonal crystals. The rock is a leuco-monzogranite, with the content of microcline slightly exceeding that of oligoclase. Apatite, fluorite, calcite and zircon are accessory. The zircon occurs throughout as isolated grains; crystals are weakly rounded and the larger ones are cored.

[1] Since completion of this report a whole grain analysis of a single, clear, euhedral monazite crystal from a newly collected sample of the Eden Lake pluton gave a ^{207}Pb-^{206}Pb age of 1740 ± 15 Ma. This new result indicates that the 2.0 Ga to 0.9 Ga linear trend (Fig. 2b) is fortuitous and that the age of the Eden Lake pluton is ca. 1740 Ma.

Separated zircons are abundant, mostly small and fractured, and contain many inclusions. They are prismatic, euhedral to subhedral, with length to breadth ratios varying from 1:1 to 3:1. They are zoned and have thin metamorphic overgrowths. Uranium concentrations are high, ranging from 410 to 730 ppm (Table 1). A scatter of data points (Fig. 4a) is interpreted in terms of an inherited component. A line through points B and E, corresponding to fractions with lowest ^{207}Pb-^{206}Pb ages has upper and lower intercepts of 1703 and 667 Ma (Fig. 4a). Given the likelihood that abrasion removed the metamorphic rims, the upper age is considered to approximate primary granite crystallization. This igneous age is constrained within the limits 1725 and 1685 Ma, where the latter approximates the ^{207}Pb-^{206}Pb age of the most concordant point E and the former the upper intercept generated by a line through E and a lower intercept of 1.0 Ga. This maximum lower intercept is based on the assumption that the zircons were affected both by recent Pb loss and by Grenville metamorphism, dated at 1.0 Ga by Krogh (1989), and this hypothesis is consistent with the results from the nearby Grondine complex (see below).

Grondine complex

Samples of two different kinds of orthogneiss were collected from the metaplutonic complex immediately east of the mouth of Beaverstone Bay, approximately 15 km southeast of the Grenville Front.

Granite orthogneiss

Pink granite orthogneiss, somewhat similar to the Fox Islands granite but more thoroughly recrystallized and containing hornblende as well as biotite, was sampled on Hen Island. The rock is now fine grained, but biotite and feldspar aggregates attest to its former coarser grain size. Feldspar is composed of about equal parts of oligoclase and weakly perthitic microcline forming a polygonal mosaic between elongate plates of annealed quartz. Hornblende and biotite have the form of dispersed aggregates which, along with quartz, define the foliation. Magnetite and apatite are the common accessory minerals. The thin section contains at least 40 zircon grains, mainly associated with the mafic mineral aggregates.

The extracted zircons have length to breadth ratios between 1:1 and 3:1, internal, sharply defined zonation, and overgrowths with rounded edges. Most grains contain fractures and some inclusions. Uranium concentrations are high, from 385 to 710 ppm (Table 1). On a concordia plot (Fig. 4b), data points fit a regression line (MSWD = 2.5) with upper and lower intercepts of 1715 +6/-5 Ma and 794 ± 126 Ma, respectively. The lower intercept may reflect both Grenvillian and a more recent lead loss event. The upper intercept age is interpreted as the time of igneous crystallization which accords with the most concordant point R consisting of strongly abraded prismatic crystals.

Quartz dioritic orthogneiss

The second sample from the Grondine complex is grey, biotite-hornblende orthogneiss of quartz diorite composition from an unnamed island 700 m north of Hen Island. An irregular, deformed body of this rock is separated from the Hen Island granite by a narrow unit of variable migmatitic gneiss (Fig. 3), of undetermined but probably supracrustal origin. The orthogneiss has a streaky aspect imparted by aggregates of fine grained mafic minerals (hornblende> biotite) and of plagioclase. Minor quartz forms drop-like grains within a subequigranular mosaic of the other minerals. Apatite, magnetite and zircon are accessory; the thin section contains at least 20 zircon grains; these are rounded and have zoned cores.

Extracted zircons are equant to subhedral prismatic with length to breadth ratios between 1:1 and 3:1. Both impurities and fractures are common. On a concordia plot (Fig. 4c) the four data points are aligned but show some scatter (MSWD = 10). Upper and lower intercept ages are 1711 +10/-8 Ma and 783 ± 82 Ma, respectively. The upper intercept age, indistinguishable from the age of the granite orthogneiss from Hen Island, is interpreted as the time of igneous crystallization because all fractions were strongly abraded, and all but fraction Y are prismatic.

DISCUSSION

It is clear from the isotopic data reported above that none of the sampled granitoid rocks records a Penokean age. The age obtained from titanite in the Cutler tonalite is in good accord with the Rb-Sr age determined earlier by Wetherill et al. (1960), and since neither this granite nor its country rocks appear to have been metamorphosed subsequent to intrusion, is considered to approximate closely the age of emplacement. The varied zircon population obtained from the Eden Lake trondhjemite is interpreted as recording a multi-event history. The zircons contain an abundant inherited component, either from the Huronian rocks which it intrudes or from the Archean substrate assumed to underlie them.

Although some of the ages obtained for granites in the Southern Province are of poor quality, the fact that none lie in the Penokean range raises the possibility that all of the granitoid intrusions, except for the Murray and Creighton granites and possibly the undated Skead granite, are related to either of two continent-wide episodes of igneous activity that occurred in post-Penoken but pre-Grenvillian time: 1) Yavapai/Mazatzal orogenies, prevalent in the southwest United States (Hoffman, 1989), identified in the Central Plain orogen south of the Penokean fold belt in the mid-continent (Sims and Peterman, 1986), and in the same age range as the Ketilidian and Labradorian orogenies, identified in and adjacent to the Grenville orogen along the Atlantic coast; 2) so-called anorogenic magmatism in the mid-continent, south of but overlapping the Central Plains and Penokean orogens, now known to be prevalent among the younger orthogneiss units so far dated in the Central Gneiss Belt of the southwestern Grenville Province. The fact that the present work has failed to identify plutonism in the range 1.89-1.83

Ga confirms the earlier suggestion of Card (1978) that the effects of the Penokean orogeny in the Southern Province of Ontario would appear to be represented only by deformation and low-grade metamorphism unrelated to plutonism; even this tectonism may be pre-Penokean.

The three zircon ages obtained for the granitoid rocks in the Grenville Front Tectonic Zone east of Killarney all lie in the range 1725-1710 Ma, a little younger than the Killarney and Wanapitei complexes, and a little older than the 1685 Ma age obtained by Corrigan (1990) for orthogneiss in northwest Britt domain, some 35 km east of Beaverstone Bay and 45 km southeast of the Grenville Front. These ages imply that igneous activity coeval with the Yavapai and Mazatzal orogenies played a major crust-forming role in this region of the Grenville Province, long before inception of the protracted Grenvillian orogeny. Future dating of other granitoid units in the region between the Bell Lake granite (1.47 Ga) and the Britt and Mann Island plutons (1.46 Ga) may identify more granitoid units of this mid-Proterozoic age.

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Circa 1.7 Ga Rb-Sr re-setting in two Huronian paleosols, Elliot Lake, Ontario and Ville Marie, Quebec

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Abstract

Isotopic compositions of Sr and isotope dilution abundance determinations of Rb and Sr on whole rock samples are reported for two paleosol profiles developed beneath strata of the Huronian Supergroup. One of these is beneath the Matinenda Formation near Elliot Lake; the other, beneath the Lorrain Formation at Ville Marie, Quebec. Large variations in Rb/Sr ratios in the uppermost parts of the paleosols make them sensitive indicators of times of weathering and/or later modifications. A poorly fitting regression line (MSWD = 54.4) through five data points from the Elliot Lake profile corresponds to an age of 1688 ±71 Ma. Ten samples from the Ville Marie weathering profile yield an errorchron (MSWD = 3.5) corresponding to an age of 1728 ± 17 Ma. These ages post-date, not only deposition of the Huronian strata but also later minor metamorphism. They may reflect post-Penokean uplift and cooling.

Résumé

Dans le présent article sont présentés la composition isotopique du Sr et des déterminations de contenu en Rb et Sr par dilution isotopique dans des échantillons de roche entière extraits de deux paléosols formés au-dessous des couches du Supergroupe de l'Huronien. L'un des paléosols se trouve au-dessous de la Formation de Matinenda près d'Elliot Lake; l'autre, au-dessous de la Formation de Lorrain dans la région de Ville-Marie, au Québec. Les grandes variations dans les rapports Rb/Sr obtenues dans la partie supérieure des paléosols en font de bons indicateurs de l'intervalle de temps associé à l'altération ou aux modifications ultérieures. Une droite de régression (mean square weighted deviate (MSWD) = 54,4) rejoignant les cinq points de données mal distribués du profil d'Elliot Lake indique un âge de 1 688 ± 71 Ma. Les dix échantillons du profil d'altération de Ville-Marie donnent une isochrone (MSWD = 3,5) correspondant à un âge de 1 728 ± 17 Ma. Ces âges sont postérieurs non seulement à la sédimentation des couches huroniennes, mais également au faible métamorphisme qui a suivi. Ils peuvent refléter un soulèvement et un refroidissement post-pénokéens.

INTRODUCTION

Precambrian paleosols have received considerable attention for their inferred fingerprinting of early atmospheric oxidation states (e.g. Pienaar, 1958; Robertson, 1964; Roscoe, 1969; Gay and Grandstaff, 1980; Retallack et al., 1984; Holland et al., 1989; Zbinden et al., 1988; Roscoe and Prasad, 1991). The differences in Fe^{3+}/Fe^{2+} of Late Archean weathering profiles in comparison to Early Proterozoic profiles is considered to record the transition from a reducing to an oxidizing paleoatmosphere.

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The change in atmospheric oxydation state, termed oxyatmoversion by Roscoe (1973), evidently occurred during deposition of the Huronian Supergroup. Formations in the lower part of the Supergroup, notably the Matinenda Formation near Elliot Lake, contain pyritic paleoplacers and authigenic pyrite, whereas hematitic paleoplacers and redbeds are present in the Cobalt Group which forms the uppermost part of the succession.

A paleosol profile beneath the quartz arenite of the Lorrain Formation in the Cobalt Group near Ville Marie, Quebec, studied by Rainbird et al. (1990), shows Fe retention, in contrast to Fe loss in sub-Matinenda paleosols, including one from Quirke Mine near Elliot Lake (Prasad and Roscoe, 1991) that was selected for this study. The locations of the two paleosols are shown in Figure 1. Rb-Sr isotopic investigations of the two profiles were undertaken to test the possibility that the paleosols could be dated isotopically, thereby establishing the time of oxyatmoversion and of deposition of uppermost Huronian formations within relatively narrow ranges. The dates obtained, however, are far younger than possible dates for deposition of Huronian rocks or for geologically reasonable dates of possible important metasomatic modifications of these rocks. We examine herein some possible causes of disruption of the Rb-Sr isotopic system.

GEOCHRONOLOGICAL CONSTRAINTS

The youngest known Archean rock near Huronian rocks is the 2665 +2/-1 Ma East Bull Lake syenite (Krogh et al., 1984, U-Pb zircon). The Matinenda Formation and the Dollyberry basalt flows, which unconformably overlie Archean rocks, are at about the same stratigraphic level in the lower part of the Huronian Supergroup as rhyolite of the Coppercliff Formation which has been dated at 2450 +25/-10 Ma (Krogh et al., 1984, U-Pb zircon). The supergroup was deformed during and after it was deposited. Later, at 2219 ± 4 Ma, it was intruded by numerous, extensive Nipissing gabbro sheets (Andrews et al., 1986). Southern parts of the Huronian belt were folded and metamorphosed prior to 1.75 Ga, most likely during the Penokean collisional orogeny at about 1.85 Ga (Card et al., 1972; Card, 1978; Hoffman, 1989). Lower greenschist minerals are present in Dollyberry basalt in the Quirke drillcore section, but there is little reason to suppose that paleosol beneath gently-dipping Lorrain quartzite at Ville Marie would have been much affected by the Penokean orogeny.

Post-Penokean events in the region, indicated by U-Pb zircon dates, include minor felsic magmatism southwest of Sudbury at about 1.75 Ga (van Breemen and Davidson, 1988; Davidson et al., 1992). There are several 1.5 Ga plutons in the Lake Huron-Manitoulin Island area. Pegmatite dykes associated with the Cutler granite intrusion 35 km south of Quirke Mine have yielded Rb-Sr dates around 1.7 Ga (Wetherill et al., 1960). Nearly all Rb-Sr and K-Ar mineral and whole rock dates from igneous and metamorphic rocks in the Penokean orogen are at least 100 to 200 Ma younger than would be expected from U-Pb zircon data. Similar discrepancies are found in comparable data for rocks throughout the Canadian Shield. Some of these may have resulted from mild effects of later orogenic events, but they have also been considered to reflect delayed closures of the isotopic systems accompanying prolonged post-orogenic uplifts and resultant cooling (e.g. Stockwell, 1980).

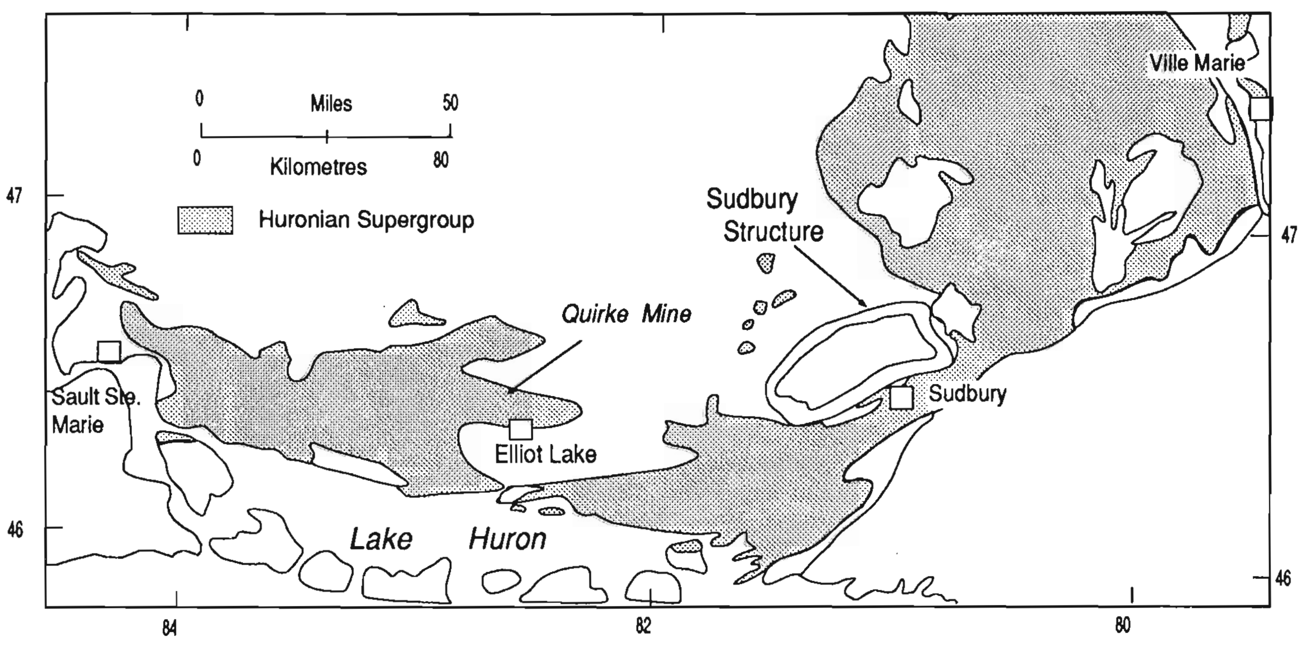

Figure 1. Location of the Elliot Lake and Ville Marie paleosols.

SUB-MATINENDA PALEOWEATHERING PROFILE AT QUIRKE MINE, ELLIOT LAKE AREA

A paleosol profile intersected in Quirke Mine underground drillhole 268 is the least complex of several studied by Prasad and Roscoe (1991). It is developed in faintly foliated basalt. The least altered basalt, 10 m beneath the base of the Matinenda Formation, contains greenschist minerals, calcite veinlets and minor sulphides. Progressive upward alteration becomes conspicuous at 5.5 m below the top of the section. Dominantly chloritic alteration changes upward to chlorite-sericite and to dominantly sericite alteration. The volcanic origin of the rock becomes unrecognizable at 1.5 m and the uppermost 90 cm consists almost entirely of sericite and resembles argillite. The contact between the paleosol argillite cap and overlying sorted clastic sediment of the Matinenda Formation is sharp. In some other profiles, clasts of residual argillite and pebbles of variably argillitized basalt and quartz are present at the base of the Matinenda Formation.

Mineralogical changes in the 268 profile are reflected by progressive upward changes in chemical composition. The compositon of the profile is relatively uniform up to 1.5 m below the top, but there are losses in Mg, Fe^{3+}, K and Rb relative to Al and gains in H_2O, S, CO_2 and Ca. Much greater changes in the uppermost 1.5 m include extreme losses in Na, Ca, Sr, CO_2, Mg, Cu and Zn, losses in Fe^{2+}, Fe^{3+}, Pb, Si, Mn, La, and Y, and gains in K, Rb, Ba, Be, S and U. In overlying Matinenda subarkose, ratios of elements to Al reflect the composition of matrix sericite plus grains of detrital potassium feldspar. These ratios are close to those in the argillaceous paleosol cap which was the source of the sericite precursor. It may be significant that Sr/Ca and Sr/Rb ratios are higher in basal subarkose than in the underlying residual argillite. The following changes in ratios between Rb, Sr, and associated K and Ca occur upward through the profile: K:Ca – 10x, Rb:K – 0.45x, Sr:Ca – 5.8x, Rb:Sr – 7.9x.

SUB-LORRAIN PROFILE, VILLE MARIE, QUEBEC

Paleosol developed on massive, red, coarse grained Archean Ville Marie granite 3 km northwest of Ville Marie, Quebec, on the east shore of Lake Temiskaming has been studied by Rainbird et al. (1990). It is a rare, possibly unique, occurrence of a well-exposed paleosol profile beneath Huronian strata younger than the Matinenda Formation. The granite is increasingly altered upward through a stratigraphic thickness of 10-20 m. In the lower part of the profile, plagioclase is altered to sericite, biotite is altered to iron-rich chlorite, and interstitial calcite is developed in the rock. This zone is transitional to an overlying homogeneous sericite-rich saprolith zone. Corestones are absent and "thin wisps of hematite" are present at the top of the zone according to Rainbird et al. (1990). The saprolith is overlain by 3-4 m of hematitic breccia that the latter workers considered a local, basal member of the Lorrain Formation. The breccia is overlain by "pebbly current-bedded pink feldspathic wacke" transitional to subarkose. Clasts are variably sericitized, and hematized metavolcanic rocks and granite are set in a matrix detritus.

Progressive upward changes in elemental concentrations relative to Al include: loss of virtually all Na; loss of most Ca and Sr; losses of Si, Cu, Pb, and Zn; gains of Rb, Ba, Mg and Fe^{3+}. The K:Al ratio in sericite is similar to that in the parent granite. The most notable difference in the Ville Marie profile compared to the Elliot Lake profile is the increase in Fe^{3+}. The following changes in ratios between Rb, Sr, and associated K and Ca occur upward through the profile: K:Ca – 100x, Rb:K – 1.3x, Sr:Ca – 2.0x, Rb:Sr – 6.7x.

ANALYTICAL METHODS

Whole rock powders were spiked with a mixed ^{87}Rb-^{84}Sr tracer solution and dissolved in an HF-HNO_3 mixture on a hot plate at 150°C overnight. Separation of Rb and Sr followed standard cation exchange chromatography. Total procedure blanks were approximately 130 pg for Sr and 60 pg for Rb. Mass analysis was carried out on a MAT 261 solid source mass spectrometer in static multicollection mode for Sr and in single collector peak jumping mode for Rb. Sr isotopic compositions were normalized to $^{88}Sr/^{86}Sr$ = 8.37521. Repeated measurements of NBS 987 standard during the period of analysis yielded $^{87}Sr/^{86}Sr$ = 0.710281 (mean of 10 analyses, 2σ error = 0.000013). $^{87}Sr/^{86}Sr$ were corrected to NBS 987 $^{87}Sr/^{86}Sr$ = 0.710250. External precision for $^{87}Rb/^{86}Sr$ ratios is estimated to be better than 1% (2σ error). Further analytical details are given by Thériault (1990).

Table 1. Rb and Sr isotope dilution abundance data and Sr isotopic compositions for paleosols from Elliot Lake, Ontario and Ville Marie, Quebec

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ (2σ)
Elliot Lake paleosol				
89RF-40	9.25	201.9	0.133	0.737925 (37)
89RF-41	19.60	200.7	0.2835	0.742022 (25)
89RF-42	78.03	169.4	9.259	0.956303 (35)
89RF-43	142.6	45.64	9.239	0.956303 (35)
89RF-44	217.6	23.87	28.16	1.39799 (4)
Ville Marie paleosol				
TR87-2	218.6	155.4	4.112	0.814560 (37)
TR87-3	341.8	69.01	14.86	1.08501 (7)
TR87-4	470.9	51.17	28.46	1.41123 (6)
TR87-6	470.6	44.42	33.16	1.54132 (4)
TR87-9	502.8	45.73	34.47	1.56255 (3)
TR87-11	613.1	21.84	99.89	3.05686 (10)
TR88-2	229.7	157.8	4.257	0.818693 (40)
TR88-3	325.9	100.2	9.633	0.953545 (27)
TR88-8	206.7	173.4	3.480	0.800863 (38)
TR88-9	329.4	107.5	9.069	0.939765 (29)

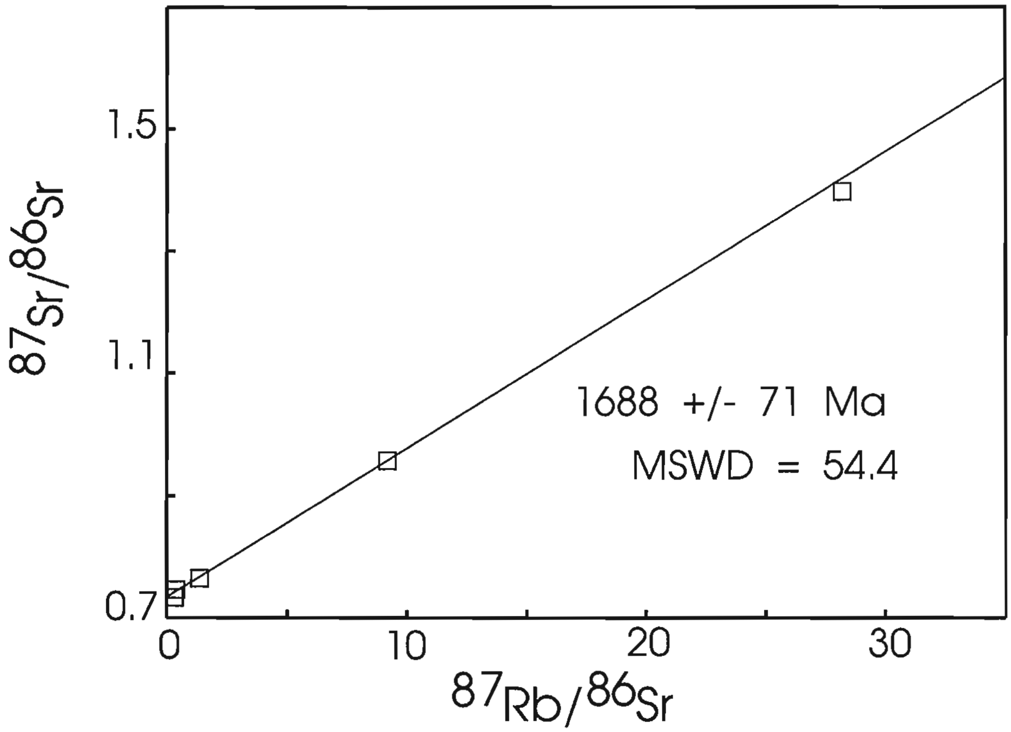

Figure 2.

$^{87}Rb/^{86}Sr$ vs. $^{87}Sr/^{86}Sr$ errorchron plot for 5 Elliot Lake paleosol whole rock samples. Samples are from the top argillitic 200 cm of the weathering profile. See text for analytical details.

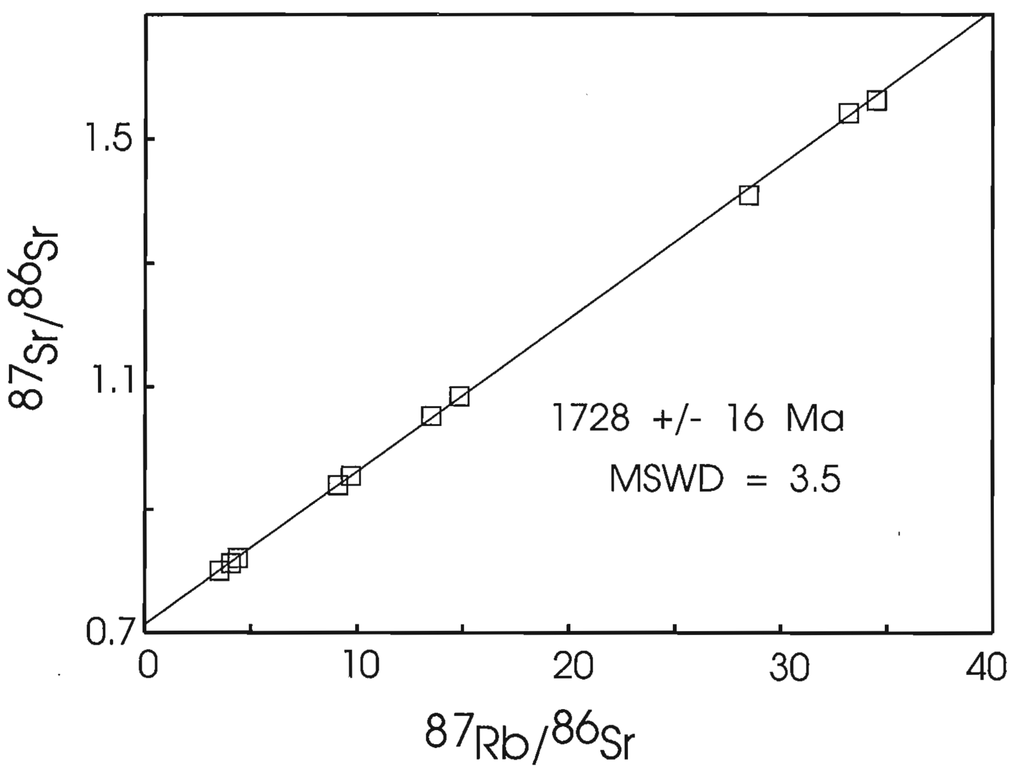

Figure 3.

$^{87}Rb/^{86}Sr$ vs. $^{87}Sr/^{86}Sr$ plot for 10 Ville Marie paleosol whole rock samples. Samples are from the transitional and saprolith zones of the weathering profile. See text for analytical details.

RESULTS

Data for ten samples from the Ville Marie paleosol and five from the Elliot Lake paleosol are presented in Table 1. The samples from the Elliot Lake paleosol were taken from the uppermost 200 cm of the weathering profile where chemical changes leading to formation of argillite as the end product of weathering of basalt are most pronounced. $^{87}Rb/^{86}Sr$ ratios vary from 0.1330 to 28.16, $^{87}Sr/^{86}Sr$ ratios from 0.737925 to 1.39799. The data define a poorly fitting line (MSWD = 54.4) corresponding to an age of 1688 ± 71 Ma (Fig. 2). Incomplete isotopic homogenization of Sr upon closure of the Rb-Sr system and/or later disturbance of the system may be responsible for the poorly defined linear array of data and elevated MSWD.

The Ville Marie samples are from the transitional and saprolith zones of the profile, where the most pronounced changes in geochemistry relative to the protolith are developed. The strong geochemical variations are reflected by $^{87}Rb/^{86}Sr$ which range from 4.11 to 34.5. Corresponding $^{87}Sr/^{86}Sr$ ratios vary from 0.800893 to 1.56255. The data yield an errorchron age of 1728 ± 17 Ma (MSWD = 3.5) (Fig. 3). Isotopic homogeneity of Sr upon closure of the Rb-Sr system and subsequent co-linear isotopic evolution of Sr is consistent with the relatively good fit of the data to a linear regression.

DISCUSSION

This attempt, and others (e.g. McFarlane and Holland, 1991), at dating Precambrian weathering events have been motivated by the very high Rb/Sr ratios developed in these profiles. The upward enrichment of Rb relative to Sr is directly related to enrichment of K relative to the loss of Ca which has been removed along with Na, Mg and, in some cases, Fe during the weathering process. Late post-orogenic disturbances of Rb-Sr isotopic systems are common if not normal, as indicated by discrepancies between U-Pb zircon dates and Rb-Sr dates on rocks in metamorphic and plutonic terranes. We had hoped, however, that the great Rb enrichment and Sr depletion in the sub-Huronian profiles coupled, in the case of the Ville Marie profile, with minimal metamorphism would have been especially conducive to the preservation of a record of Rb-Sr isotopic closure corresponding to the time of weathering.

The study has not yielded the age of formation of the paleosols. The Rb-Sr isotopic systems were apparently open or were reopened after the weathering, perhaps during deep burial, and were closed 500 to 700 Ma after deposition of the Huronian Supergroup and no less than 100 Ma after the Penokean collisional orogeny. Alternatively, the data could be interpreted as suggesting that the wide ranges of Rb/Sr ratios in the paleosols were produced largely during a 1.7 Ga metasomatic event, rather than entirely during weathering and diagenesis. There is no geological evidence, however, that pervasive regional metasomatism accompanied the limited 1.7 Ga magmatism known to have occurred in the southern part of the Huronian belt.

Addition of K (and associated Rb) is a notable feature of ancient weathering profiles, in contrast to K depletion in younger and modern profiles. This has led to doubts that the addition or apparent retention of alkalis actually occurred during the subareal weathering process. Rainbird et al. (1990) postulated that K was initially removed from the weathered Ville Marie granite, that kaolin was the final weathering product, and that K was subsequently restored during diagenesis in amounts adequate to convert the kaolin to illite, the precursor of sericite. They considered that sericite in the matrix of overlying Lorrain sandstone was originally detrital kaolinitic clay derived from soil, and that illitization of the interstitial clay as well as of the sub-Lorrain residual clay was controlled by permeability of the sandstone. Sericite is ubiquitous throughout the Huronian succession, however, and it occurs as the main component of thick beds in pelitic formations as well as in matrices of arenite and conglomerates, so such permeability control for its formation is not obvious. Moreover, unaltered as well as sericitic rock clasts and potassium feldspar clasts are common. If only the geological relationships are considered, it would seem possible that residual clay which was K-rich prior to burial was deposited in Huronian formations, that diagenetic and metamorphic modifications of buried residual clay and that transported clay minerals were essentially isochemical. In any case, the scenario proposed by Rainbird et al. (1990) would not result in isotopic dates significantly later than soil formation.

McFarlane and Holland (1991) found that Rb-Sr isotopic data on samples from three paleosols of different ages gave dates 300 to 650 Ma younger than permissible times of weathering. Tops of each of the paleosol profiles are enriched in K. The oldest, developed on 2.76 Ga Mount Roe basalt, Fortesque Group, Australia, shows Fe depletion, but paleosols developed on the ca 2.2 Ga Hekpoort basalt and the 1.92 Ga Ongeluk basalt in South Africa show Fe enrichment. McFarlane and Holland (1991) attributed the younger Rb-Sr isochron dates to resetting caused by alkali metasomatism, related to regional metamorphism in the case of the Mount Roe paleosol and to effects of nearby intrusions in the cases of the Hekpoort and Ongeluk paleosols. No such explanation can be postulated for K (and Rb) enrichment in sericite in Huronian rocks and sub-Huronian paleosols. There is no evidence for post-Huronian, post-Nipissing gabbro, post-Penokean orogeny pervasive alkali metasomatism.

Perhaps post-Huronian isotopic equilibration of Rb-Sr systems in Huronian paleosols resulted from minor migrations and exchanges of elements without important changes in bulk rock compositions. The late dates then would reflect cessation of isotopic equilibration as a result of cooling attendant on regional uplift and erosion.

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U-Pb and ^{40}Ar-^{39}Ar geochronology of granodioritic orthogneiss in the western Pelly Mountains, Yukon Territory

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Abstract

Granodioritic orthogneiss is an important component of a medium to high grade metamorphic assemblage which underlies a large area in the western Pelly Mountains. This unit was originally considered to be of Proterozoic to possibly Early Cambrian age. A sample of the orthogneiss gives U-Pb zircon and monazite ages of 110.8 ± 0.4 Ma, which we interpret as igneous crystallization ages. An ^{40}Ar-^{39}Ar age of 100.0 ± 1.0 Ma for biotite from the same sample indicates that the orthogneiss and associated metasedimentary schist and gneiss were ductilely deformed at mid-crustal levels, metamorphosed at amphibolite facies, and cooled through ~300°C, all within 11 million years following intrusion. The orthogneiss is intruded, and ductile fabrics within it are locally annealed, by the relatively high-level, undeformed, 85-95 Ma Nisutlin batholith.

Résumé

L'orthogneiss granodioritique est une importante composante d'un assemblage qui a subi un métamorphisme moyen à élevé et qui couvre une vaste zone dans la partie ouest des monts Pelly. L'âge de cette unité a été initialement établi entre le Protérozoïque et probablement le Cambrien précoce. Un échantillon de l'orthogneiss donne des âges U-Pb sur zircon et monazite de 110,8 ± 0,4 Ma, lesquels sont associés à la cristallisation ignée. Un âge ^{40}Ar-^{39}Ar sur biotite de 100,0 ± 1,0 Ma obtenu à partir du même échantillon indique que l'orthogneiss ainsi que les roches métasédimentaires associées (schiste et gneiss) ont subi une déformation ductile au niveau de la croûte intermédiaire, ont été métamorphisées au faciès des amphibolites et ont commencé à durcir à env. 300 °C, moins de 11 millions d'années après l'intrusion. L'orthogneiss est recoupé par des intrusions et ses fabriques ductiles internes sont localement recuites par le batholite de Nisutlin (85-95 Ma), une masse non déformée de profondeur relativement faible.

INTRODUCTION

Orthogneiss and metasedimentary schist and gneiss underlie a large area in west-central Quiet Lake map-area in the western Pelly Mountains of southeastern Yukon Territory (Fig. 1). The study area lies between the Tintina Fault Zone on the northeast and the Teslin Suture Zone on the southwest (Fig. 1). Tempelman-Kluit (1977) included these metamorphic rocks in his unit Pn+, and considered them to be mainly Proterozoic to possibly Early Cambrian in age. In this study, we have dated a sample of granodioritic orthogneiss from within the Pn+ package. The data provide both a crystallization age for the protolith of the orthogneiss unit, and a post-tectonic cooling age, and therefore have important implications for the timing of metamorphism and deformation of this gneissic package and associated rocks.

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REGIONAL SETTING AND SAMPLE DESCRIPTION

The Pn+ unit in west-central Quiet Lake map-area consists of orthogneiss, migmatite, and schist. Unit Pn+ was originally mapped as Proterozoic, and the metasedimentary package was mapped as Proterozoic to early Cambrian in age.

Orthogneiss and metasedimentary rocks of unit Pn+ display a concordant, gently-dipping foliation which contains a southeast-trending elongation lineation. Asymmetric microstructures such as S-C fabrics (Berthé et al., 1979), mica fish, shear bands, and asymmetric microfolds record top-to-the-southeast (or dextral) displacement under ductile flow conditions (Hansen, unpublished data). These paragneiss and orthogneiss units are intruded by the Nisutlin batholith which lacks any obvious tectonic fabric. K-Ar biotite ages reported for the Nisutlin batholith by Tempelman-Kluit (1977) range from 85-95 Ma. Deformation fabrics of the gneiss and schist are annealed along the contact with the Nisutlin batholith; this relation, and the presence of andalusite along the batholith contact indicate that the Nisutlin batholith intruded after ductile shearing ceased, and at a relatively shallow crustal level (Spicuzza and Hansen, 1989).

The sample collected for dating (Fig. 1) is from a biotite-feldspar augen orthogneiss unit of granodioritic bulk composition, which is part of the Pn+ unit. The orthogneiss is in apparently gradational contact with muscovite-biotite orthogneiss and metasedimentary gneiss and schist.

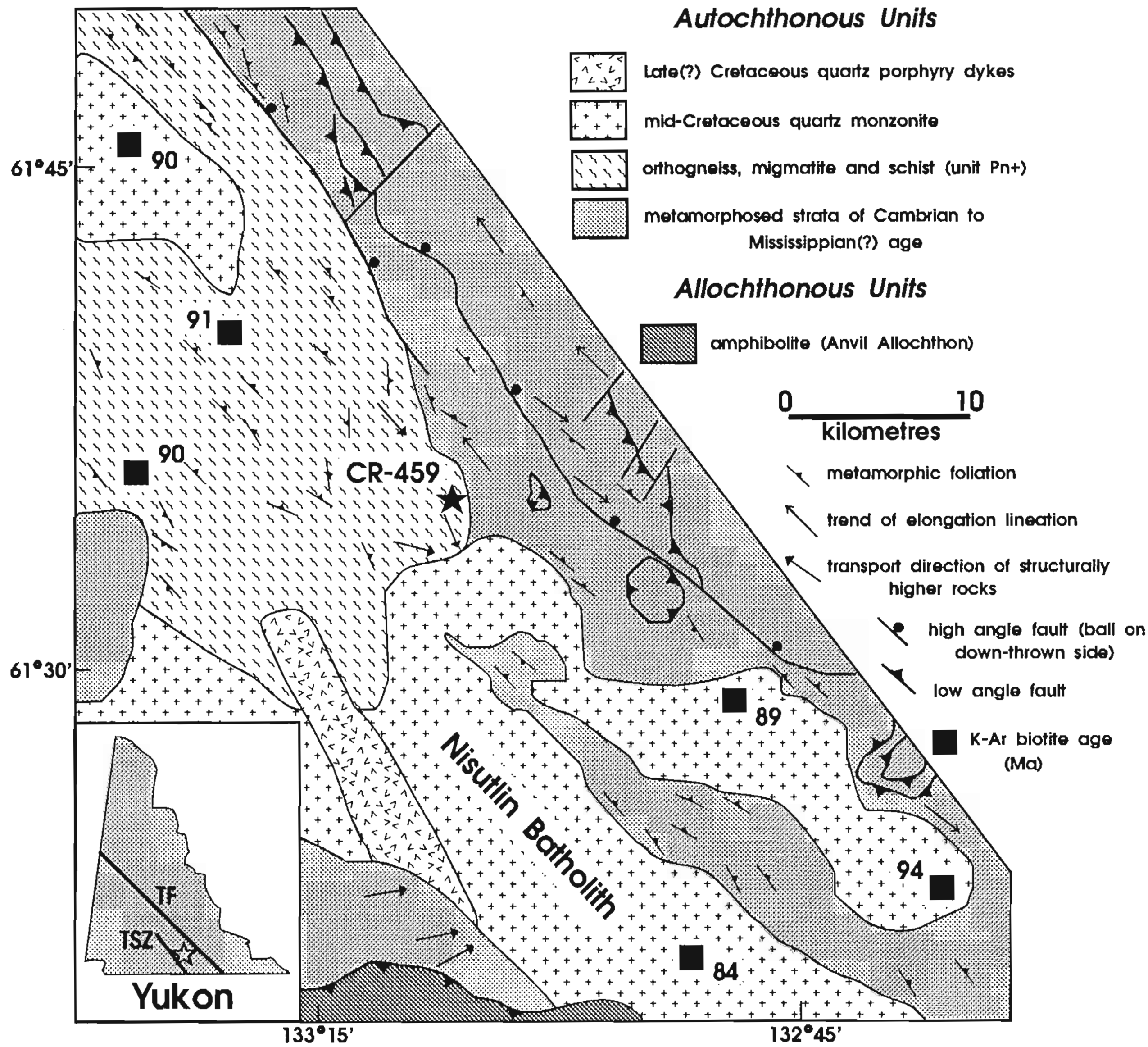

Figure 1. Simplified geology of the west-central part of Quiet Lake map area (modified from Tempelman-Kluit, 1977). Main structures shown on inset map: TF, Tintina Fault Zone; TSZ, Teslin Suture Zone. Star on inset map shows location of the study area.

ANALYTICAL METHODS

Zircon, monazite and biotite were separated from a 25 kg sample using conventional Wilfley table, heavy liquids, and magnetic separation techniques. Techniques for U-Pb dating of zircon and monazite are modified slightly from that described by Parrish et al. (1987). Techniques used for K-Ar and ^{40}Ar-^{39}Ar dating are described in this volume (Hunt and Roddick, 1992).

ANALYTICAL RESULTS

U-Pb analytical data are given in Table 1. U-Pb data are also shown in a conventional U-Pb concordia plot in Figure 2. Zircon recovered from the sample range from euhedral, equant, multifaceted grains to very elongate, euhedral prisms with multifaceted terminations. All grains are clear and colourless. Zoning is visible in the more prismatic grains, and slightly cloudy, rounded cores are present in some grains.

Table 1. U-Pb analytical data

Sample Description[1]	Wt (µg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$ (meas.)[3]	total common Pb (pg)	% ^{208}Pb[2]	$^{206}Pb/^{238}U$[4] (± % 1σ)	$^{207}Pb/^{235}U$[4] (± % 1σ)	$^{207}Pb/^{206}Pb$[4] (± % 1σ)	$^{207}Pb/^{206}Pb$ age (Ma; ± % 2σ)
Sample CR-459 (61°34.1'N, 133°04.7'W)										
A: N1,+105,p,a	16	886	36.0	1479	25	7.8	0.04021(0.11)0.0	0.4600(0.12)	0.08298(0.05)	1268.6(1.9)
B: N1,+105,e,a	82	577	33.4	7087	23	10.6	5518(0.10)	0.6753(0.11)	0.08876(0.03)	1399.0(1.1)
C: N1,+74,lp,a	19	1287	22.4	1247	22	9.8	0.01738(0.11)	0.1157(0.14)	0.04829(0.08)	113.4(3.8)
AA: monazite	18	5404	290	3021	35	70.4	0.01750(0.10)	0.1153(0.16)	0.04781(0.11)	89.8(5.0)
BB: monazite	30	3548	305	2974	38	81.5	0.01754(0.10)	0.1152(0.15)	0.04763(0.10)	80.8(4.7)

[1] N1, non-magnetic at 1 degree side tilt on Frantz isodynamic separator; s, stubby prisms; e, elongate prisms; m, multifaceted equant grains; a, abraded
[2] radiogenic Pb; corrected for blank, initial common Pb, and spike
[3] corrected for spike and fractionation
[4] corrected for blank Pb and U, and common Pb

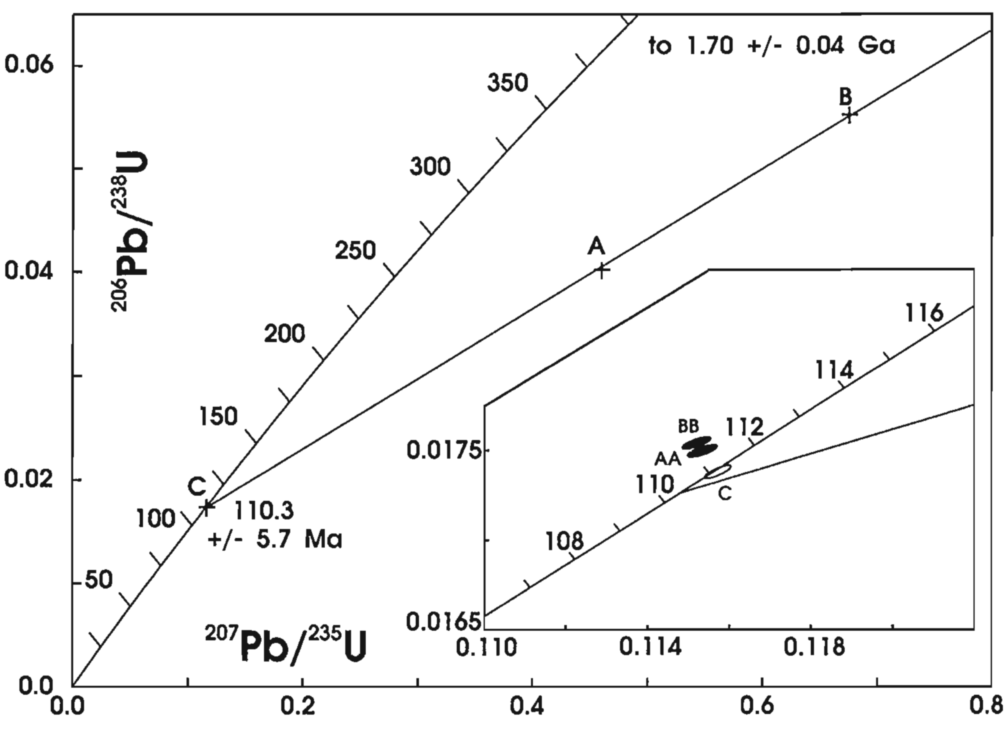

Figure 2.

U-Pb concordia plot for zircons (open) and monazites (solid) from the biotite orthogneiss unit.

Clear rod-, tube- and bubble-shaped inclusions are rare to abundant in the zircons. Monazite in the sample forms clear, pale yellow, subhedral prisms and tablets with abundant clear and opaque inclusions. Three fractions of zircon were selected for analysis, taking care to avoid any grains with visible cores. All fractions were strongly abraded. One fraction, which consisted of very elongate, inclusion-free prisms, is nearly concordant at about 111 Ma (Fig. 2). Together the three zircon fractions define a linear array which we interpret to indicate the presence of a significant inherited zircon component in two of the three fractions. A calculated regression line through the three analyses (MSWD = 481) gives upper and lower intercept ages of 1.70 ± 0.04 Ga and 110.3 ± 5.7 Ma. The upper intercept age is an average age for the inherited zircon component, and the lower intercept age corresponds to the time of intrusion of the protolith for the orthogneiss. A better estimate for the age of intrusion is given by the U-Pb monazite analyses. Both fractions of monazite plot above concordia, reflecting the presence of excess ^{206}Pb related to ^{230}Th disequilibrium effects. We consider the ^{207}Pb-^{235}U age of the reversely discordant monazite analyses to be the best estimate for the actual age of closure of the U-Pb system in this mineral. The two analyses give identical ^{207}Pb-^{235}U ages of 110.8 ± 0.4 Ma.

Fresh, reddish-brown biotite from the orthogneiss sample was also dated using a 3-step ^{40}Ar-^{39}Ar method. The two final steps, comprising 98% of the gas released, are in agreement, and correspond to an age of 100.0 ± 1.0 Ma. This is interpreted as the age at which the orthogneiss unit last cooled though the closure temperature for Ar diffusion in biotite (~320°C; Harrison et al., 1985).

DISCUSSION

The 111 Ma U-Pb crystallization ages reported here for the biotite feldspar augen orthogneiss unit indicate that the top-to-the-southeast (or dextral) displacement at mid-crustal levels occurred during or after 110 Ma. Shearing could have started prior to emplacement of the orthogneiss magma and continued after crystallization of the magma; therefore these dates do not necessarily constrain the initiation of ductile deformation. Ductile deformation which affected the orthogneiss must have ceased by 100 Ma, when the unit last cooled through the closure temperature of the Ar system in biotite. Ductile deformation must certainly have been complete by 85-95 Ma, the K-Ar biotite cooling ages from the crosscutting Nisutlin batholith. We conclude that syn-kinematic metamorphism of the metasedimentary host rocks within the Pn+ unit, locally up to sillimanite grade of the amphibolite facies (M.J. Spicuzza, unpublished data), occurred during, and possibly as a result of, mid-Cretaceous emplacement of the orthogneiss parent magma.

Our data indicate that the biotite feldspar augen orthogneiss of unit Pn+ of Tempelman-Kluit (1977) is mid-Cretaceous rather than Proterozoic in age. It is therefore considered to represent an early, deep-level phase of the Cassiar Suite of intrusions (Woodsworth et al., 1991). Metamorphism and local migmatization of associated metasedimentary strata, which include rocks of Cambrian to possibly as young as Mississippian age, likely occurred simultaneously with, and possibly as a result of, mid-Cretaceous emplacement of the granodioritic magma as a thick sill. Top-to-the-southeast (or dextral) crustal shear accompanied or closely followed sill intrusion.

Recognition of mid-Cretaceous protolith ages for orthogneiss bodies within unit Pn+ in the western Pelly Mountains indicates that the volume of mid-Cretaceous plutonic rocks in the western Pelly Mountains may be much greater than previously estimated.

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Age and Pb isotopic studies of Ag-Sn-base metal epigenetic mineralization in the Mount Mye area, east-central Yukon Territory

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Abstract

Epigenetic veins and mineralized breccia systems developed within the eastern part of the mid-Cretaceous Anvil Batholith in the Mount Mye area of east-central Yukon have recently been explored for their precious metal content. Muscovites in alteration zones adjacent to the mineralization give average ^{40}Ar-^{39}Ar ages of 100.6 ±1.1 Ma, indicating that the mineralization is related to early, highly peraluminous phases of the batholith. Pb isotopic compositions for galenas from veins and breccia zones in this area are highly radiogenic, and relatively uniform. The Pb may have been derived either from the host granitoids, or from sedimentary strata in the area. The mineralization resembles Ag-Sn-bearing veins in southern Bolivia.

Résumé

Les filons épigénétiques et les réseaux de brèches minéralisées mis en place dans la partie orientale du Batholite d'Anvil, une masse du Crétacé moyen observée dans la région du mont Mye, dans le centre est du Yukon, ont récemment fait l'objet d'une exploration minérale pour déterminer leur teneur en métaux précieux. Les muscovites présentes dans les zones d'altération adjacentes à la minéralisation donnent des âges ^{40}Ar-^{39}Ar moyens de 100,6 ± 1,1 Ma, indiquant que la minéralisation est liée aux phases hyperalumineuses précoces du batholite. La composition isotopique du Pb dans la galène des filons et des zones bréchifiées est, dans cette zone, relativement uniforme et indique que cet élément est radiogène. Le Pb pourrait provenir des granitoïdes encaissants ou des couches sédimentaires de la région. La minéralisation est semblable à celle des filons à Ag et Sn dans le sud de la Bolivie.

INTRODUCTION

We have carried out mineralogical, dating, and Pb isotopic studies of Ag, Sn and base metal-bearing veins and breccias in the Mount Mye area of east-central Yukon (Fig. 1). These occurrences have recently been explored extensively for their precious metal contents. Here we report ^{40}Ar-^{39}Ar ages for alteration zones associated with the occurrences, as well as Pb isotopic compositions of galenas from several samples. The results have implications for both the nature and genesis of the mineralization itself, and for the regional potential for similar mineralization elsewhere in the northern Cordillera.

REGIONAL SETTING

Veins in the Mount Mye area occur within the eastern part of the Anvil Batholith (Fig. 1), a composite intrusion of mid-Cretaceous age that was emplaced into metamorphosed middle Paleozoic and older sedimentary strata of the Selwyn Basin (e.g., Gordey and Irwin, 1987). The batholith forms part of the Selwyn Plutonic Suite (Anderson, 1988; Woodsworth et al., 1991), which includes batholiths and smaller bodies throughout much of east-central and southeastern Yukon (Fig. 1). The Selwyn Plutonic Suite comprises two distinct phases. Strongly peraluminous biotite

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and biotite-muscovite quartz monzonite and granite is the most abundant rock type. This phase is typically equigranular, but locally includes megacrystic and pegmatitic varieties. It is associated with tungsten skarn mineralization in several areas of southeastern Yukon (e.g., Anderson, 1988). Regional U-Pb and K-Ar dating studies (Mortensen and Gordey, unpublished data), indicate that this phase of plutonism began at about 112 Ma, and continued until ~99 Ma. A slightly younger suite of plutons is also widespread, particularly in east-central Yukon. It consists of mildly peraluminous, commonly hornblende-bearing, biotite granodiorite and quartz monzonite. These rocks are locally porphyritic, suggesting a relatively high level of emplacement. They are closely associated and appear to be comagmatic with local thick accumulations of welded dacitic tuff of the South Fork Volcanics. These volcanic rocks are mainly confined to several large caldera structures (e.g., Gordey and Irwin, 1987). The ring faults of these calderas cut the two-mica granitoids, and are locally intruded by the hornblende-bearing plutons. The hornblende-bearing intrusions and the South Fork Volcanics yield consistent U-Pb zircon ages of about 97 Ma (Mortensen and Gordey, unpublished data).

Two-mica-bearing phases of the Anvil Batholith are termed the Mount Mye phase. Hornblende-bearing and porphyritic plutonic bodies within the batholith are termed the Orchay phase in the western part of the batholith and the Marjorie phase in the southeastern part (e.g., Pigage and Anderson, 1985; Anderson, 1988; Gordey, 1990a,b). The margins of Mount Mye phase intrusions are commonly moderately to strongly foliated. This ductile shear fabric is cross-cut by plutons of the Orchay and Marjorie suites.

VEINS IN THE CODY RIDGE AREA

Silver-bearing mineralized zones yielding assay values in excess of 3000 g/tonne were reported by Doron Exploration (Inc.) on the east-facing steep slope of Cody Ridge in the eastern part of the Anvil Range (Fig. 2) (R. Robertson and G. Wallis, pers. comm., 1988). This prospect is situated due north of Mount Mye, at and above 5500 feet (1680 m) elevation. Polymetallic-precious metal-bearing mineralization occurs in veins and fracture-filled breccia bodies. The veins themselves are also brecciated. Examination of samples collected in 1986 revealed the presence of significant Sn values in the mineralization. The results reported here are from a detailed examination of 22 samples collected during a property visit in 1988 (Ballantyne et al., 1989; Ballantyne and Harris, 1990).

The granite-hosted mineralization in the Cody Ridge area occurs in zones of quartz (± pyrite, sericite) veinlets and stockwork zones associated with strong, pervasive sericitic alteration. Mineralizing events are related to multiple and repeated brecciation of these early quartz veins. Continued fluid introduction produced zones of micro-brecciated ore and gangue minerals contained in banded-colloidal-crystalline quartz and/or chalcedony and carbonate (rhodochrosite). The veins and/or filled fractures pinch and

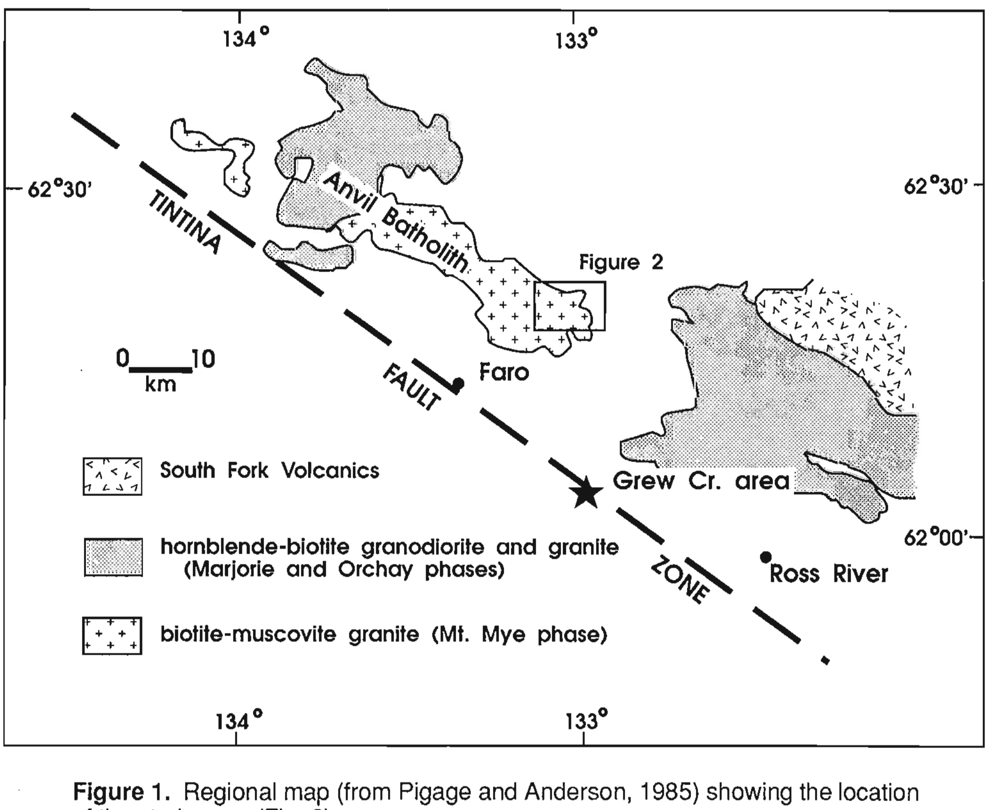

Figure 1. Regional map (from Pigage and Anderson, 1985) showing the location of the study area (Fig. 2).

swell to a maximum of a metre in width. Ore and gangue minerals appear to have been deposited intermittently in the fractures and around breccia fragments. The mineralized zones are inferred to have formed at relatively shallow levels.

Deformed metavolcanic and metasedimentary rocks that form a roof pendant on Cody Ridge (Fig. 2) are cut by quartz-feldspar-porphyry dykes. The relationship of the mineralization to these dykes, or to intermediate composition dykes (Unit Ks2 of Gordey, 1990a) is unknown.

Electron microprobe and SEM analysis were both necessary to identify and document the complex ore and gangue mineral assemblage. Complex ore fluids introduced Ag, Sn, Zn, Pb, Sb, Fe, As, Cu, Mn, Ca and Si during brecciation and vein and/or veinlet development. This complex elemental assemblage and resultant mineral assemblage is similar to Southern Bolivian Ag-Sn deposits (Ballantyne and Harris, 1990).

Studies thus far have found pyrite, non-silver bearing galena, Fe-rich sphalerite, arsenopyrite, stannite, canfieldite, acanthite, Ag-bearing tetrahedrite, native silver, semsegite, covellite, diaphorite, pyragyrite, miargyrite, and high temperature, needle-shaped cassiterite. These minerals are found as brecciated fragments, banded or micro-brecciated, encased in silica and pink rhodochrosite which may display open spaces and vugs.

The surfaces of mineralized samples are commonly stained black with manganese oxides and hydroxides. However, sulphides are generally fresh and unaltered in

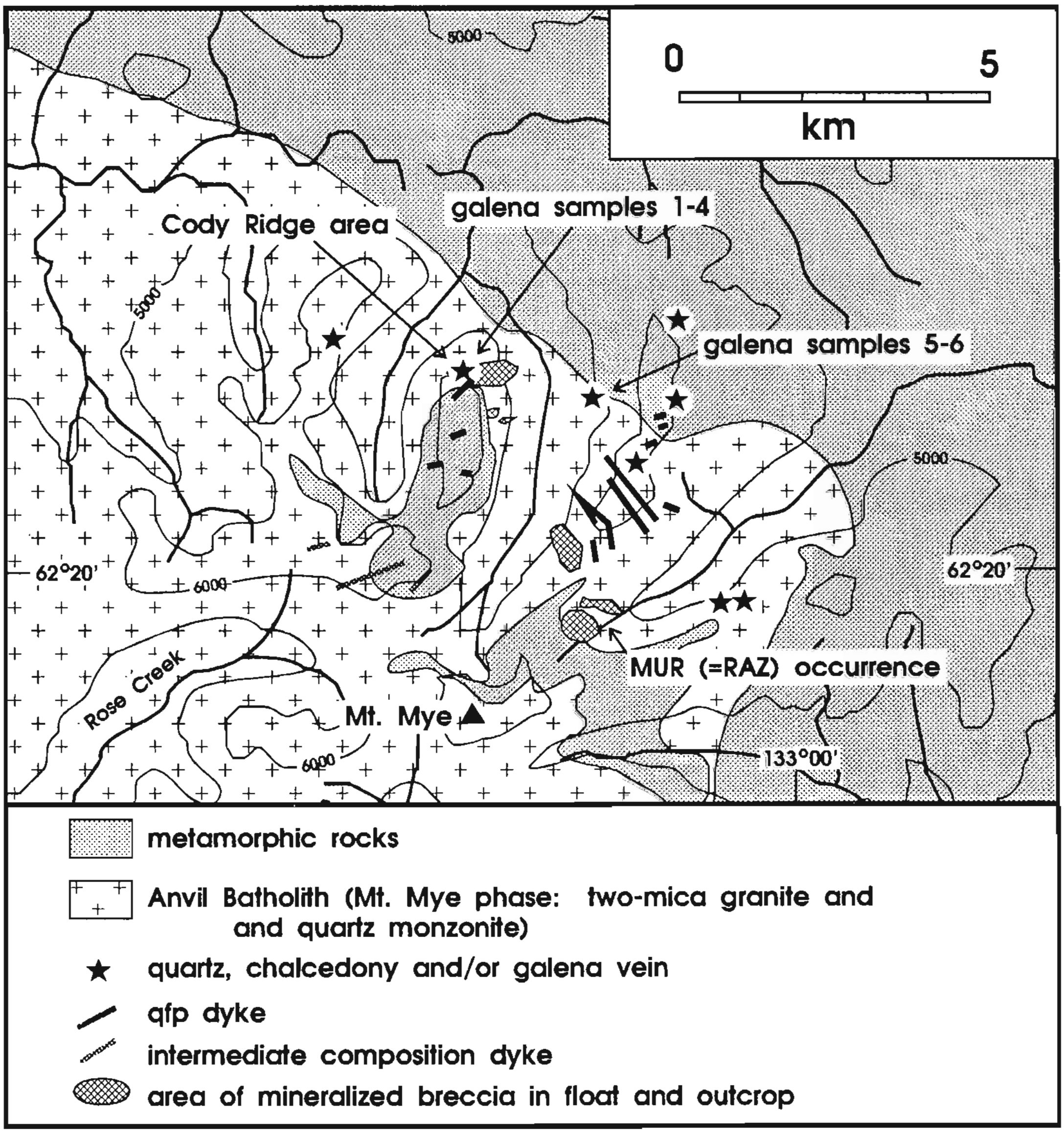

Figure 2. Simplified geology map of the eastern Anvil Range (modified from Gordey, 1990a,b; R. Robertson and G. Wallis, pers. comm., 1988).

surface samples. Pale yellow-green sericite is the predominant alteration product of the granite adjacent to the veinlets, veins or fractures. It appears throughout the mineralized zone with greatest abundance immediately adjacent to the mineralization.

Other similar occurrences have been discovered to the southeast of the Cody Ridge locality. These include the MUR and NA occurrences (Fig. 2) (Mortensen, 1977; Godwin et al., 1988; MINFILE, 1992).

^{40}AR-^{39}AR DATING RESULTS

We have obtained three-step ^{40}Ar-^{39}Ar step-heating age spectra for two samples of muscovite from alteration envelopes adjacent to mineralized veins. Analytical techniques are described elsewhere in this volume (Hunt and Roddick, 1992), and analytical data are given in Hunt and Roddick (1992) and summarized in Table 1. Sample 889004 is from altered granite adjacent to a chaotic stockwork of quartz veinlets containing pyrite and arsenopyrite. These veinlets also crosscut the Fe- and Mn-rich veins. Sample 889006 is from altered granite adjacent to a strongly manganese stained, brecciated veinlet. The two samples give consistent results. The first step (<2% of gas released) give spurious ages. The two final steps, comprising a total of 98% of the gas released from each sample, correspond to ages of 100.4 ± 1.1 Ma and 100.8 ± 1.1 Ma. The mineralization is thought to have formed at shallow crustal levels, and at temperatures that are unlikely to have exceeded the closure temperature of the ^{40}Ar-^{39}Ar system in muscovite (~350°C). Therefore we interpret the ages for the alteration muscovite samples to date formation of the mineralized vein systems, rather than as post-veining cooling ages.

Table 1. ^{40}Ar-^{39}Ar dating results for muscovite from alteration zones

Sample No.	#889004	#889006
Mineral	muscovite	muscovite
Location	62°21.8'N; 133°5.8'W	62°21.8'N; 133°5.8'W
Step 2 (age, % gas)	100.0 ± 0.8 Ma (47%)	100.3 ± 0.3 Ma (45%)
Step 3 (age, % gas)	100.8 ± 0.4 Ma (51%)	101.2 ± 0.7 Ma (53%)
Integrated ^{40}Ar-^{39}Ar age	100.6 ± 1.3 Ma	101.1 ± 1.1 Ma

PB ISOTOPIC STUDIES

Pb isotopic compositions of galenas (data in Table 2) were measured using Faraday collectors in static mode on a Finnegan MAT 261 solid source mass spectrometer. Sample loads of about 75 ng were analyzed at a filament temperature of 1200°C. Analyses of unknowns were corrected for instrumental mass fractionation of 0.135%/amu, based on more than 50 replicate analyses of Broken Hill galena standard T-1003, and the composition recommended by Richards et al. (1981).

Table 2. Pb isotopic compositions of galenas from epigenetic mineralization in the Mount Mye area

Sample No.	Description	^{208}Pb/^{204}Pb (%1σ)	^{207}Pb/^{204}Pb (%1σ)	^{206}Pb/^{204}Pb (%1σ)	Source
1: 879001	high-Ag vein and polymetallic breccia	39.285(0.05)	15.700(0.04)	19.200(0.04)	this study
2: 889022	high-Ag vein breccia with vugs and bands of quartz and rhodochrosite	39.282(0.05)	15.700(0.05)	19.207(0.04)	this study
3: 889008	high-Ag vein with vugs filled with quartz and rhodochrosite	39.317(0.05)	15.711(0.04)	19.211(0.04)	this study
4: 889003	low-Ag vugs with rhodochrosite veinlets	39.277(0.05)	15.700(0.04)	19.207(0.04)	this study
5: 879003	massive, low-Ag galena vein	39.293(0.05)	15.703(0.04)	19.210(0.04)	this study
6: 879002	massive, low-Ag galena vein	39.244(0.05)	15.682(0.04)	19.185(0.04)	this study
7: MUR	vein	39.325(?)	15.725(?)	19.230(?)	Godwin et al., 1988
8: RAZ (=MUR)	vein	39.293(?)	15.724(?)	19.217(?)	Godwin et al., 1988
9: NA	vein	39.245(0.16)	15.696(0.16)	19.188(0.08)	Godwin et al., 1998

Analyses of four samples of galena from veins in the main area of mineralization on Cody Ridge, as well as two samples from a massive galena vein across the valley to the east (Fig. 2), are shown in Table 2 and plotted on conventional $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ plots in Figure 3. Also included are analyses of galena from other similar veins in the same area (MUR occurrence [number 105K:53, MINFILE, 1992] and NA occurrence, [exact location unknown]), taken from a compilation by Godwin et al. (1988). Two average crustal growth curves are shown for reference; these are the "shale curve", constructed by Godwin and Sinclair (1982) for the Canadian Cordilleran miogeocline, and the "average upper crust" curve from Zartman and Doe (1981). The shale curve appears to reflect most closely Pb isotopic evolution within the Selwyn Basin.

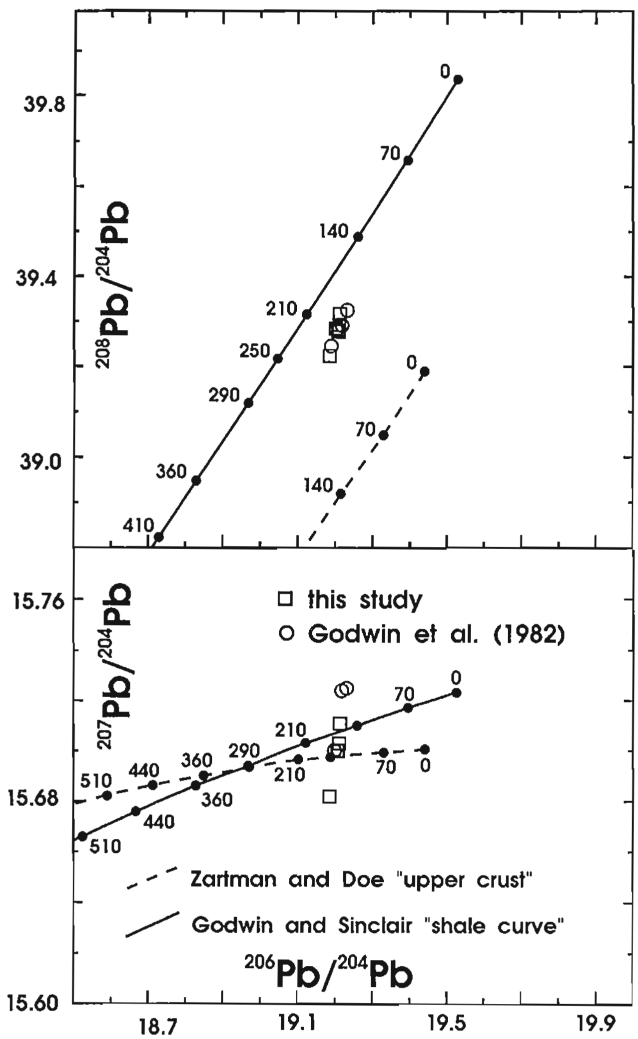

Figure 3. Pb isotopic compositions for galenas from epigenetic mineralization in the northastern Anvil Range. Crustal growth curves (with ages shown in Ma) are from Zartman and Doe (1981) and Godwin and Sinclair (1982).

Pb isotopic values from the Cody Ridge occurrences and from the other veins in the area cluster just below the shale curve on both the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ plots. This indicates that the Pb in these veins was derived largely from the upper crust, and was generated in an environment with a μ-value slightly lower than the average for typical Selwyn Basin sedimentary rocks. There is a small spread in the data that may reflect mixing of Pb from two or more isotopically distinct sources; however the data are insufficient to test this.

DISCUSSION

The ages we have obtained for alteration related to veining in the Cody Ridge area indicates that the mineralization is much older than the mid-Eocene, epithermal Au-Ag mineralization that occurs at Grew Creek in the Tintina fault zone southeast of Faro (Fig. 1) (e.g., Christie et al., 1992). The average age of 100.6 ± 1.1 Ma for veins on Cody Ridge is similar to the youngest U-Pb ages obtained thus far for plutons of the Mount Mye phase of the Anvil Batholith, but is distinctly older than ages for the younger phases of the batholith. The veins therefore apparently predate most of the young porphyritic dykes in this area, which are thought to be related to the 97 Ma Orchay phase and associated South Fork Volcanics. We conclude that the veins formed from hydrothermal systems established during or immediately following emplacement of the Mount Mye suite of intrusions. No initial common Pb compositions are presently available for the Mount Mye phase granitoids; however, the strongly peraluminous nature of these units suggests that these compositions are likely to be very radiogenic, possibly comparable to shale curve values. Pb isotopic compositions for late magmatic fluids should therefore not be expected to differ significantly from those of Pb derived from surrounding sedimentary country rocks by intrusion-driven geothermal systems. The Pb isotopic compositions of galena from the veins indicate that the Pb was derived from upper crustal sources, but cannot be used to further constrain the nature and origin of the mineralizing fluids.

The nature of the Cody Ridge mineralization is important because most of the exploration in this area was directed towards Faro-style, stratiform Pb-Zn-Ag deposits. Geochemical stream sediment sampling in the area was targeted for these elements, and resulting anomalies were interpreted within the context of a massive sulphide deposit model. The district has recently been surveyed as part of the National Geochemical Reconnaissance (NGR) stream sediment and water program (GSC Open File 2174, 1990). In the NGR data, multi-element stream sediment anomalies associated with Cody Ridge are coincident for the elements Zn, Pb, Ag, As, Sb, Sn, Mn, Fe, Cd and therefore directly reflect its complex mineral assemblage. The NGR data and "historic" anomalies should be reassessed in view of the possibility of more extensive mineralized systems of the Cody Ridge type (Ballantyne, 1991).

The mineralization exploration potential of peraluminous phases of the Selwyn Plutonic Suite has previously been thought to be mainly limited to W (± Sn) skarns (e.g., Anderson, 1988; Lynch, 1989; Emond, 1992; Emond and Lynch, 1992). However other Ag-Sn vein mineralization such as the ZETA occurrence (number 115P:47 in MINFILE, 1992) in the northern McQueston map-area (e.g., Abercrombie, 1990; Emond and Lynch, 1992; MINFILE, 1992) resemble that on Cody Ridge. It is possible that mineralized systems such as this have gone unrecognized elsewhere in east-central Yukon.

The Cody Ridge mineralization shows marked similarities to southern Bolivian ores. The Andean tin province ores are generally subvolcanic, and related to dacitic-rhyodacitic stocks. They typically have a complex paragenesis (high silver sulphosalts), zonation, telescoping and intense hydrothermal alteration (e.g., Cerro Rico, Potosi) (Villalpando, 1988).

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U-Pb, ^{40}Ar-^{39}Ar, and K-Ar ages for metamorphism of the Kluane and Aishihik assemblages in southwestern Yukon Territory

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Mortensen, J.K. and Erdmer, P., 1992: U-Pb, ^{40}Ar-^{39}Ar, and K-Ar ages for metamorphism of the Kluane and Aishihik assemblages in southwestern Yukon Territory; in Radiogenic Age and Isotopic Studies, Report 6; Geological Survey of Canada, Paper 92-2, p. 135-140.

Abstract

Medium to high grade metamorphic rocks of the Kluane and Aishihik assemblages adjacent to the Denali fault zone in southwestern Yukon give very consistent, Early Tertiary U-Pb monazite and xenotime, and K-Ar and ^{40}Ar-^{39}Ar muscovite and biotite ages. These data indicate that metamorphism and deformation in this area coincided with emplacement of the Paleocene-Eocene Ruby Range Batholith, which forms a very thick conformable sill. The metamorphic rocks record high cooling (and uplift) rates. Structural, petrological and geochronological studies in this area indicate that metamorphism occurred during a major period of west-vergent deformation and syntectonic plutonism.

Résumé

Les méthodes U-Pb sur monazite et xénotime (Tertiaire précoce) ainsi que K-Ar et ^{40}Ar-^{39}Ar sur muscovite et biotite ont permis d'obtenir des âges très cohérents dans le cas des roches de métamorphisme moyen à élevé des assemblages de Kluane et d'Aishihik, adjacentes à la zone de failles de Denali dans le sud-ouest du Yukon. Ces données indiquent que le métamorphisme et la déformation dans cette région ont eu lieu en même temps que la mise en place du batholite de Ruby Range du Paléocène-Éocène, un filon-couche concordant très épais. Les roches métamorphiques révèlent des taux de refroidissement (et de soulèvement) élevés. Des études structurales, pétrologiques et géochronologiques des roches de cette région permettent d'associer le métamorphisme à une importante période de déformation à vergence ouest et à un plutonisme syntectonique.

INTRODUCTION

A large area on the northeast side of the Denali Fault Zone in southwestern Yukon Territory is underlain by meta- morphic rocks (Fig. 1). Previous workers have correlated the metamorphic rocks in the study area in part with the Nisling Assemblage, a continental margin assemblage thought to be of early Paleozoic age or older (Wheeler and McFeely, 1991; Mortensen, in press), and in part with the Kluane Schist. The metamorphic assemblage is shown as "metamorphic rocks undivided" in the most recent geological compilation by Wheeler and McFeely (1991). Geological mapping and petrological studies led Erdmer (1989, 1990, 1991) to introduce the term "Aishihik assemblage" for metamorphic units previously included in the Nisling assemblage, and "Kluane assemblage" to include both the Kluane Schist *sensu stricto* as well as related rocks of somewhat higher metamorphic grade in the study area that were not originally considered part of the Kluane Schist. The Aishihik and Kluane assemblages are predominantly sedimentary in origin; however the protolith age(s) and tectonic evolution of the metamorphic terrane as a whole are poorly understood.

Several K-Ar cooling ages exist for metamorphic micas from both the Kluane and Aishihik assemblages in the study area. Samples from the Kluane assemblage generally yielded Eocene cooling ages, whereas Jurassic to Early Cretaceous cooling ages have been reported from the Aishihik

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assemblage. In this study, we have obtained U-Pb ages for metamorphic monazite and xenotime from both assemblages, as well as K-Ar and ^{40}Ar-^{39}Ar cooling ages for metamorphic biotite and muscovite. Our data allow us to constrain both the age of peak metamorphism and the post-metamorphic cooling history of the region. The results shed light on the nature of Early Tertiary tectonism and regional uplift in southwestern Yukon.

REGIONAL GEOLOGY

The geology of the study area is shown in simplified form in Figure 1. Supracrustal rocks southwest of the Ruby Range Batholith have been subdivided into three main units: the Aishihik assemblage, the Kluane assemblage, and the Dezadeash Formation. The Aishihik assemblage consists of biotite muscovite psammite and pelite that are commonly migmatitic and garnet and (or) sillimanite-bearing, micaceous quartzite, amphibolite, marble, calc-silicate, and minor granitic orthogneiss. It is lithologically equivalent to the biotite schist and marble units in Aishihik map-area described by Tempelman-Kluit (1974) and the Nisling assemblage of Wheeler and McFeeley (1991). The Kluane assemblage comprises mainly graphitic biotite schist with abundant andesine porphyroblasts, as well as minor muscovite-chlorite schist. The biotite schist units are locally garnet- and staurolite-bearing. The Kluane assemblage includes the hornfelsed schist or Kluane Schist unit of Tempelman-Kluit (1974), as well as rocks included in the Yukon Group by Kindle (1952) and Muller (1967). The Dezadeash Formation consists of weakly metamorphosed argillite, greywacke, conglomerate and minor volcanic rocks, and contains macrofossils of Late Jurassic and Early Cretaceous age.

The Aishihik and Kluane assemblages are lithologically distinct, and are in sharp contact along a regionally northeast-dipping surface that is interpreted as a pre-metamorphic fault. The Kluane assemblage appears to record a single period of deformation and metamorphism, whereas the Aishihik assemblage is polydeformed and polymetamorphic. Petrological studies by Erdmer (1991) suggest that the prograde metamorphism recorded by the Kluane assemblage may correspond to a relatively young retrograde event in the Aishihik assemblage. Peak metamorphic temperatures recorded by the Kluane and Aishihik assemblages are similar (Erdmer, 1991); garnet-biotite geothermometry indicates a range of temperatures of 525-665°C. Metamorphic temperatures generally appear to

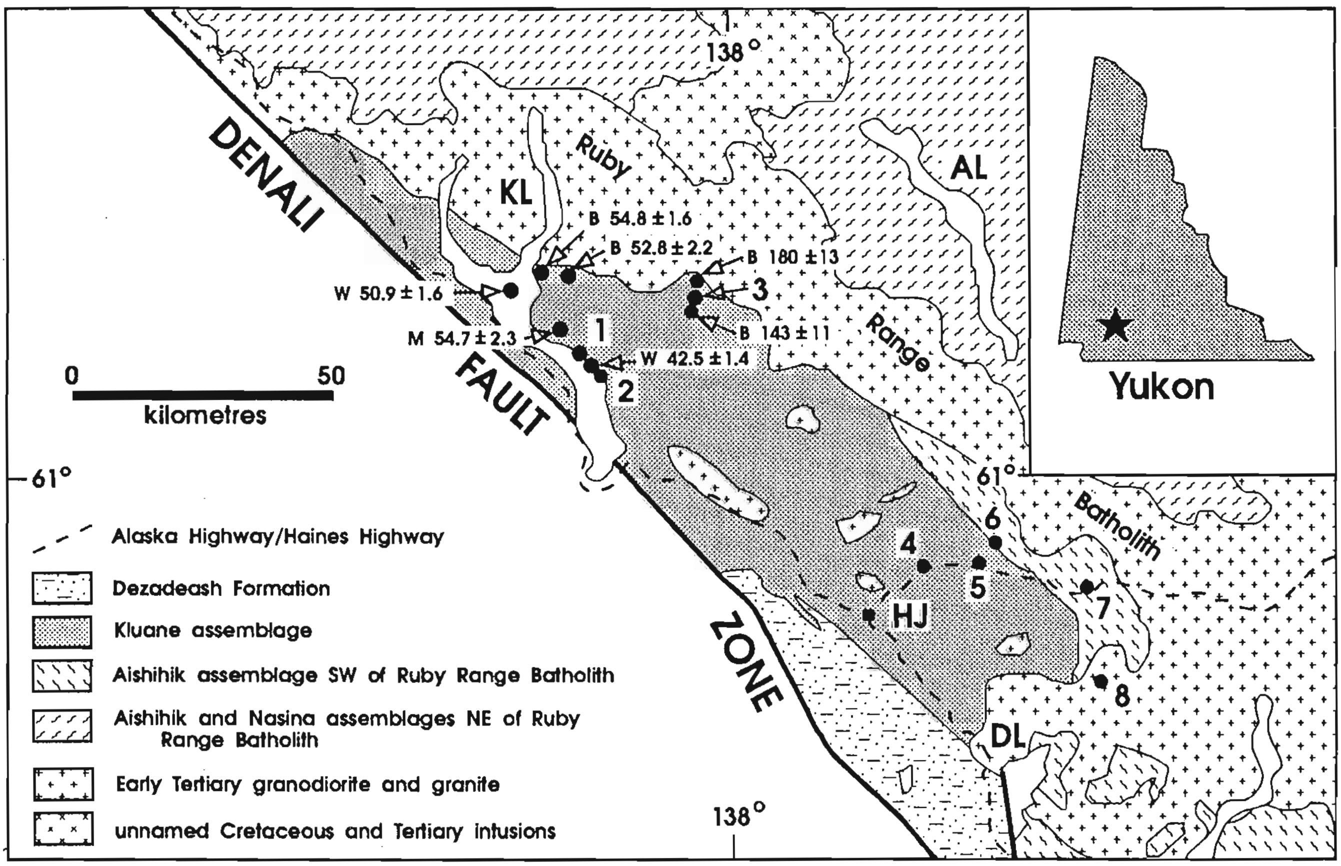

Figure 1. Simplified geological map of the study area (modified from Wheeler and McFeeley, 1991). Sample localities are shown by the numbered dots. Existing K-Ar ages for metamorphic rocks are also shown (B, biotite; M, muscovite; W, whole rock; data from Lowdon, 1960; Stevens et al., 1982; Farrar et al., 1988). HJ, Haines Junction; KL, Kluane Lake; DL, Dezadeash Lake; AL, Aishihik Lake.

increase slightly to the northeast. Geobarometric studies give conflicting results, and are not considered reliable. Layering and schistosity in the metamorphic rocks generally dip northeast.

The Ruby Range Batholith (Woodsworth et al., 1991) is an elongate body of mainly biotite-hornblende granodiorite that intrudes both the Kluane and Aishihik assemblages. The batholith appears to have intruded in a sill-like fashion, roughly parallel to compositional layering and schistosity in the enclosing schists and gneisses (S. Johnston, pers. comm., 1991).

Eisbacher (1976) suggested that the Kluane Schist may be a more highly metamorphosed equivalent of the Jura-Cretaceous Dezadeash Formation exposed immediately to the southwest (Fig. 1). Erdmer (1990) noted that in addition to differences in metamorphic grade, the Kluane Schist also differs lithologically from the Dezadeash Formation. The contact between the Deszadeash Formation and the Kluane assemblage is inferred to be faulted, and its original nature is unknown.

PREVIOUS DATING STUDIES

Available age data from metamorphic rocks southwest between the Ruby Range Batholith and the Denali fault zone are shown in Figure 1 (data from Lowdon, 1960; Stevens et al., 1982; and Farrar et al., 1988). K-Ar ages for muscovite, biotite and whole rock samples of mica-rich schist yielded ages in the range of 42.5-54.7 Ma (early to mid-Eocene) in the Kluane Lake area. Two K-Ar biotite ages from the Aishihik assemblage farther east gave ages of 143 and 180 Ma (Early Jurassic to Early Cretaceous). However, measured K-contents for these samples are very low, and the ages are considered somewhat suspect. More than 20 isotopic

Table 1. U-Pb analytical data

Sample Description[1]	Wt (μg)	U (ppm)	Pb[2] (ppm)	$^{206}Pb/^{204}Pb$ (meas.)[3]	total common Pb (pg)	% ^{208}Pb[2]	$^{206}Pb/^{238}U$[4] (± % 1σ)	$^{207}Pb/^{235}U$[4] (± % 1σ)	$^{207}Pb/^{206}Pb$[4] (± % 1σ)	$^{207}Pb/^{235}U$ age (Ma; ± % 2σ)
Sample 3 (PE-89-65; 61°18.1'N, 138°03.9'W)										
A: monazite, 4	40	5076	76.0	1821	62	46.3	0.00889(0.13)	0.0578(0.17)	0.04714(0.11)	57.0(0.2)
B: xenotime, 5	18	4789	39.9	791	61	5.2	0.00877(0.11)	0.0570(0.28)	0.04715(0.23)	56.3(0.3)
Sample 4 (PE-89-113; 60°51.6'N, 137°03.9'W)										
A: monazite, 1	8	5368	158	702	36	71.6	0.00880(0.11)	0.0571(0.42)	0.04703(0.36)	56.3(0.5)
B: monazite, 4	13	6670	134	1489	32	60.7	0.00875(0.10)	0.0566(0.23)	0.04691(0.19)	55.9(0.3)
Sample 5 (PE-89-111; 60°40.2'N, 136°46.3'W)										
A: monazite, 3	31	3833	101	1517	44	69.2	0.00897(0.10)	0.0578(0.18)	0.04675(0.13)	57.1(0.02)
B: monazite, 15	46	4401	121	1515	75	70.3	0.00898(0.12)	0.0579(0.18)	0.04675(0.12)	57.1(0.02)
Sample 6 (PE-89-117; 61°54.3'N, 137°01.5'W)										
A: monazite,10	24	8118	172	1885	58	61.6	0.00899(0.11)	0.0582(0.16)	0.04695(0.10)	57.4(0.02)
B: monazite, 8	45	8356	149	2941	72	54.3	0.00904(0.14)	0.0587(0.16)	0.04711(0.06)	57.9(0.02)
Sample 7 (PE-89-114; 60°49.6'N, 136°48.1'W)										
A: monazite, 5	31	11877	177	3420	60	45.8	0.00893(0.12)	0.0576(0.14)	0.04680(0.05)	56.9(0.02)
B: monazite, 9	26	9019	164	3823	34	55.5	0.00893(0.11)	0.0576(0.13)	0.04679(0.06)	56.8(0.01)

[1] number of grains analysed given by number
[2] radiogenic Pb; corrected for blank, initial common Pb, and spike
[3] corrected for spike and fractionation
[4] corrected for blank Pb and U, and common Pb.

ages have been determined for the Ruby Range Batholith; these include K-Ar biotite and hornblende ages, as well as Rb-Sr whole rock and U-Pb zircon ages. Most of the ages range from Paleocene to early Eocene (50-57 Ma); however some portions of the batholith have given Early and Late Cretaceous crystallization ages. Thus it appears to be a composite body, emplaced mainly in Early Tertiary time. A small felsic pluton that intrudes the Kluane assemblage has also given an Early Tertiary age (Farrar et al., 1988).

ANALYTICAL METHODS

Minerals for dating were separated from 3-5 kg samples using conventional Wilfley table, heavy liquids, and magnetic separation techniques. Techniques for U-Pb dating of monazite and xenotime are modified slightly from that described by Parrish et al. (1987). Techniques used for K-Ar and ^{40}Ar-^{39}Ar dating are described elsewhere in this volume (Hunt and Roddick, 1992). ^{40}Ar-^{39}Ar age determinations were made using three heating steps, and ages reported here represent the average of the final two heating steps, comprising >98% of the gas released.

RESULTS

K-Ar and ^{40}Ar-^{39}Ar analytical data are given in detail in Hunt and Roddick (1992). U-Pb analytical data are given in Table 1 and shown in a conventional U-Pb concordia plot in Figure 2. All of the age data are summarized in Table 2.

Abundant monazite was recovered from five sillimanite-bearing, biotite-rich schist and gneiss samples (3-7 in Fig. 1). Three of the samples are from the Kluane assemblage and two are from the Aishihik assemblage. The monazite consists of pale yellow to yellowish green, rounded subhedral discs, locally with abundant fine opaque inclusions. Xenotime was also recovered from sample 3. It occurs as pale yellow, stubby to elongate simple prismatic grains. Both the monazite and xenotime are interpreted as metamorphic in origin. Peak metamorphic temperatures estimated for the Kluane and Aishihik assemblages are significantly lower than the closure temperature of the U-Pb system in monazite (~700°C; Parrish, 1990); hence the U-Pb monazite ages are thought to date peak metamorphism in this area. Duplicate monazite analyses were done on four samples, and one fraction of monazite and one of xenotime were analyzed from the fifth sample. Most of the monazite analyses plot above concordia, reflecting the presence of excess ^{206}Pb related to ^{230}Th disequilibrium effects. These disequilibrium effects are negligible in xenotime, which has a much lower Th/U ratio than monazite. We consider the ^{207}Pb-^{235}U age of the reversely discordant monazite analyses to be the best estimate of the actual age of closure of the U-Pb system; hence this age is reported in Tables 2 and 3. The U-Pb monazite ages are very consistent at 55.9-57.9 Ma (latest Paleocene-early Eocene). The xenotime fraction from sample 3 is concordant, and gives a slightly younger age than monazite from the same sample (56.3 ± 0.3 Ma vs. 57.0 ± 0.2 Ma). It is uncertain whether this indicates a slightly lower closure temperature for the U-Pb system in xenotime than in monazite.

Table 2. Summary of U-Pb, K-Ar and 40Ar-39Ar age data

Sample Number	U-Pb monazite(m) or xenotime (x) age (Ma)	K-Ar biotite (b) age (Ma)	Ar-Ar muscovite (m) or biotite (b) age (Ma)
1 (PE-89-83)			m 43.4 (1.9)
2 (PE-89-85)			b 39.4 (0.9)
3 (PE-89-65)	m 57.0 (0.2) x 56.3 (0.3)	b 54.7 (1.8)	
4 (PE-89-113)	m 56.3 (0.5) m 55.9 (0.3)	b 52.0 (0.8)	
5 (PE-89-111)	m 57.1 (0.2) m 57.1 (0.2)	b 45.3 (0.9)	
6 (PE-89-117)	m 57.4 (0.2) m 58.0 (0.2)	b 52.0 (2.8)	
7 (PE-89-114)	m 56.9 (0.2) m 56.8 (0.1)	b 50.2 (0.8)	
8 (PE-89-101)			b 55.3 (0.7)

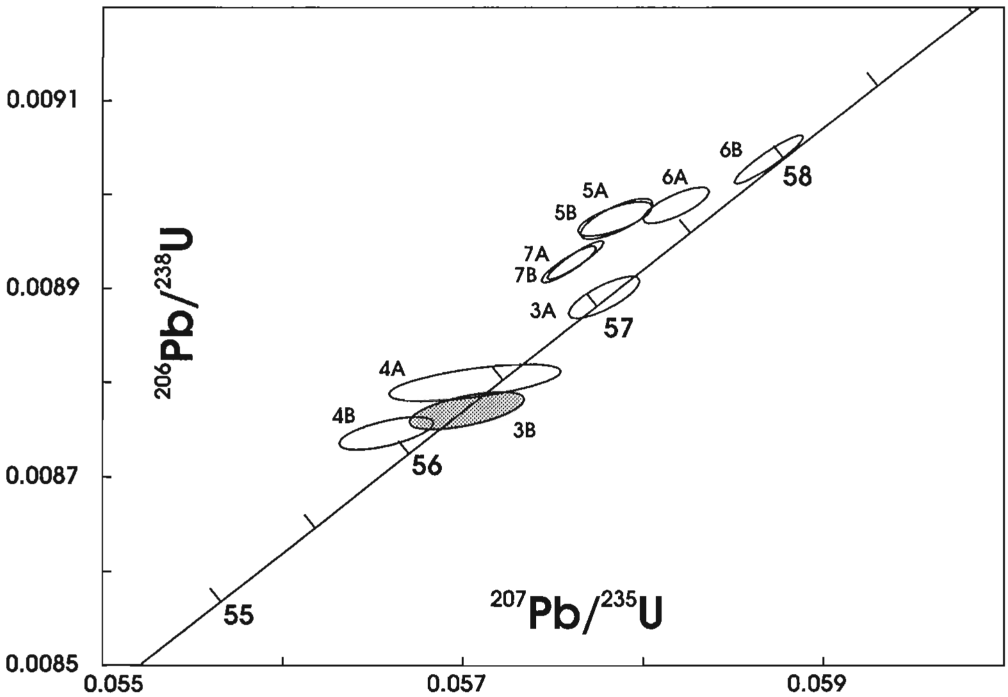

Figure 2.

U-Pb concordia plot for metamorphic monazite and xenotime (shaded) from samples of Kluane and Aishihik assemblages.

Table 3. Calculated cooling rates for metamorphic rock samples

Sample No.	Peak T (°C)	U-Pb monazite age (Ma)	K-Ar biotite age (Ma)	Cooling rate (°C/Ma)
3	560-660	57.0	54.7	122-165
4	~550-600	56.1	52.0	66-78
5	~550-600	57.1	45.3	39-46
6	556	57.7	52.0	48
7	~550-600	56.9	50.2	40-48

K-Ar biotite ages were determined for four of the five samples from which we obtained U-Pb monazite ages. In addition, ^{40}Ar-^{39}Ar ages for muscovite and (or) biotite were determined for three other samples. K-Ar and ^{40}Ar-^{39}Ar ages are generally consistent with ages reported previously for metamorphic rocks from this area. In particular, the ^{40}Ar-^{39}Ar ages of 43.4 and 39.4 Ma for samples 1 and 2 along the shore of Kluane Lake confirm the 42.5 Ma age that was reported by Farrar et al. (1988) from a nearby locality. Although sample 8 is shown in Figure 1 as being within the Ruby Range Batholith, this area is underlain at least in part by metamorphic rocks (Erdmer, 1990). The K-Ar biotite age of 54.7 Ma obtained for sample 3, together with the Paleocene/Eocene U-Pb monazite and xenotime ages for the same sample, indicate that the previously reported K-Ar biotite ages of 143 and 180 Ma for the same locality (Lowdon, 1960) are suspect, and should be disregarded.

DISCUSSION

Our data show that the terminal high grade metamorphic event in both the Aishihik and Kluane assemblages between the Ruby Range Batholith and the Denali fault zone occurred in latest Paleocene to early Eocene time. The close coincidence between the peak metamorphic ages (55.9-58.0 Ma) and the crystallization age of the Ruby Range Batholith (50-57 Ma) suggests that the batholith was emplaced as a syn-metamorphic sill. Structural data from the metamorphic rocks indicate that deformation associated with the metamorphism was west- or southwest-vergent. Although the Early Tertiary tectonism appears to be the only event recorded in the Kluane assemblage, there is evidence of overprinting of earlier fabrics and metamorphic assemblages in the Aishihik assemblage. Our data do not further constrain the possible protolith ages for the two metamorphic packages.

The combined K-Ar biotite and U-Pb monazite ages for single samples, together with peak metamorphic temperatures determined by the garnet-biotite geothermometer, permit post-peak metamorphic cooling rates to be estimated for several of the samples. As discussed above, the estimated closure temperature of 700°C for the U-Pb system in monazite is well above the peak metamorphic temperatures reached during metamorphism of the samples, and we infer that the cooling history of the samples began at the measured age of the monazites. Closure temperatures of 350°C and 280°C (as summarized in Heaman and Parrish, 1991) were used for the ^{40}Ar-^{39}Ar system in muscovite and the K-Ar system in biotite, respectively. The muscovite and biotite ages are interpreted as cooling ages. Calculated rates are given in Table 3. The data indicate extremely rapid cooling rates in the highest grade rocks adjacent to the Ruby Range Batholith. A cooling rate of 18°C/Ma is inferred for the 350-280°C interval for the area along the eastern shore of Kluane Lake, based on ^{40}Ar-^{39}Ar muscovite and biotite ages of samples 1 and 2. This area experienced peak metamorphic temperatures comparable to the rest of the Kluane assemblage (~540°C; Erdmer, 1991); however we have no direct constraint of the age of peak metamorphism in this area. The high cooling rates likely reflect rapid uplift rates in this region.

We interpret that the Ruby Range Batholith intruded as a synmetamorphic and syntectonic sill at about 56-58 Ma, during west-vergent deformation. The previously deformed and metamorphosed Aishihik assemblage may have been juxtaposed against the Kluane assemblage early in this interval along a west-vergent thrust fault. Rapid syn- to post-tectonic uplift is implied by the high cooling rates and early cooling through 280°C in the metamorphic rocks nearest to the batholith, whereas rocks farther southwest cooled more slowly. The entire structural stack was subsequently tilted to the northeast, possibly during offset along the Denali fault system. Juxtaposition of the Kluane assemblage and Dezadeash Formation may have occurred at this time.

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Age and provenance of felsic clasts in Bowser Basin, northern British Columbia

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Abstract

A dacite boulder, found in Bowser Lake Group strata of Early Callovian age, has a U-Pb zircon age of 160.7 ±0.7 Ma. Erosion of the volcanic source, or unroofing of a high-level intrusion must have been rapid. The most likely source rocks are presently located in both the hanging wall and footwall of the King Salmon fault. Older Hazelton Group volcanic rocks in Spatsizi and Iskut map areas are considered unlikely sources, as are felsic units farther south in Quesnellia.

Resumé

Un bloc de dacite, trouvé dans les couches du Groupe de Bowser Lake du Callovien précoce, a été daté par la méthode U-Pb sur zircon à 160,7 ±0,7 Ma. L'érosion de la source volcanique ou le décapage d'une intrusion de faible profondeur a dû être rapide. Les roches mères les plus probables sont actuellement situées dans le toit et le mur de la faille de King Salmon. Les roches volcaniques plus vieilles du Groupe de Hazelton qui sont observées dans les régions cartographiques de Spatsizi et d'Iskut ne sont pas considérées comme des sources probables, ce qui est également le cas des unités felsiques plus au sud dans la Quesnellie.

INTRODUCTION

Sediment in the northern sector of Bowser Basin (Spatsizi and adjacent map areas) was derived primarily from the oceanic Cache Creek Terrane, presently exposed in the hanging wall of King Salmon fault (Fig. 1). In particular, spectacular conglomerate units in the Bowser Lake Group (Middle to Upper Jurassic) consist almost entirely of clasts of radiolarian chert (Currie, 1984; Cordey et al. 1987); these clasts were derived from the Cache Creek Terrane when, beginning in the Aalenian, it was uplifted above sea level (Ricketts et al., unpublished data). A ubiquitous but minor component of the conglomerates that clearly is not derived from the Cache Creek source, includes clasts of felsic volcanic and\or intrusive rock; these generally constitute less than 1-2% of the framework. We attempt to identify this source by dating one of these clasts.

STRATIGRAPHY AND PETROGRAPHY

Distinctive, white-weathering clasts of dacite, some having porphyritic texture, occur in Callovian and Oxfordian conglomerate of the Bowser Lake Group. Most are less than 5-10 cm maximum width, are well rounded and exhibit some alteration. One (oversized) boulder 30 cm across, was extracted from a transgressive facies in a succession of coarsening-upwards shelf cycles on Tsatia Mountain. (Fig. 1 – section RAK 14-89, Ricketts, 1990; NTS 104H, 57°33.57'N; 129°6.7'W; UTM zone 9, 443456N 6379851E). The sample was located 260 m above a lenticular unit of conglomerate (known locally as the "waterfall conglomerate"), the top of which coincides with a shelf-slope break (Ricketts and Evenchick, 1991).

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Biostratigraphic constraints for the sample are good because several beds in the underlying 200 m-interval contain abundant Early Callovian *Cadoceras* and related ammonite faunas (Poulton et al., 1991). Overlying shelf deposits contain Middle Callovian and younger ammonite genera.

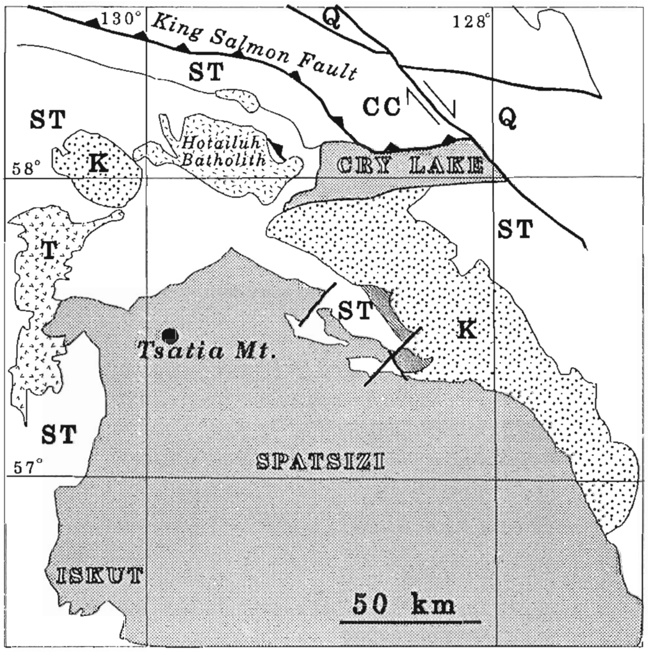

Figure 1. Location map showing sample site, Bowser Basin deposits (shaded), major tectonic structures and terrane boundaries, and map areas. ST = Stikinia; CC = Cache Creek Terrane; Q = Quesnellia; K = Sustut Basin deposits; T = Tertiary volcanic rocks.

The dated sample consists of moderately altered (calcite) plagioclase phenocrysts (10%) in a groundmass of altered plagioclase laths. There is no visible quartz, although up to 15% quartz phenocrysts are seen in other dacite samples. There are no penetrative structural fabrics in the dacite clasts.

U-Pb ANALYTICAL METHODS AND RESULTS

A small amount of zircon was obtained from the single clast of dacite using conventional mineral separation techniques involving crushing, grinding, and heavy liquid separation. Zircons were of euhedral igneous habit, with relatively equant and elongate habits, and lacked any visible evidence of older inherited cores. Minor inclusions in most crystals were quite common. Zircons were chosen for analysis which lacked cracks and alteration, and these were air-abraded using methods similar to Krogh (1982). U-Pb analytical procedures are outlined in Parrish et al. (1987). U and Pb blanks were 0.5 and 8-10 picograms, respectively. Analytical data are given in Table 1 and illustrated in Figure 2. Error ellipses in Figure 2 are shown at the 2σ level of uncertainty.

Due to the small amount of zircon recovered from this pebble, the amounts of zircon and radiogenic Pb in each analysis was small, the total Pb in each being less than 100 picograms. Nevertheless, despite the relatively large errors resulting from uncertainties in the composition of the blank and common Pb corrections, the three analyses are concordant and have overlapping errors indicating an age of 160.7 ±0.7 Ma. This age is interpreted as the crystallization age of the dacite in the source area, it corresponds to Early Callovian on the time scale of Harland et al. (1990). Since this conglomerate overlies Lower Callovian beds and underlies Middle Callovian rocks, rapid erosion, transport, and deposition of the dacite clasts are required. Also, the Early

Table 1. U-Pb analytical data for dacite boulder, sample RAK-8-14-89

Zircon[#] fraction	wt.[a] (mg)	U (ppm)	Pb[b] (ppm)	$^{206}Pb^{c}/^{204}Pb$	Pb_c^{d} (pg)	$^{208}Pb^{e}$ (%)	$^{206}Pb^{f}/^{238}U$	$^{207}Pb^{7}/^{235}U$(Ma)	corr. coef.	$^{207}Pb^{f}/^{206}Pb$	^{207}Pb age[g] ^{206}Pb(Ma)
A,e	0.010	237.5	5.941	181	23	0.10	0.02525±0.24%	0.1715±1.03%	0.68	0.04926±0.89%	160±41
B,p	0.016	163.6	4.041	492	9	0.09	0.02525±0.11%	0.1717±0.43%	0.54	0.04933±0.38%	164±18
C,p	0.009	137.2	3.403	89	27	0.09	0.02525±0.54%	0.1709±2.86%	0.65	0.04908±2.55%	152±119

[#]e, equant; p, prismatic
[a]Weighing error = 0.001 mg
[b]Radiogenic Pb
[c]Measured ratio, corrected for spike, and Pb fractionation of 0.09% ± 0.03%/AMU
[d]Total common Pb in analysis corrected for fractionation and spike
[e]Radiogenic ^{208}Pb, as expressed as the percent of total radiogenic Pb.
[f]Corrected for blank U, and total common Pb; common Pb composition was determined from isochron regression analysis and then used to calculate the radiogenic Pb composition; errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean in when expressed in Ma.
[g]Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

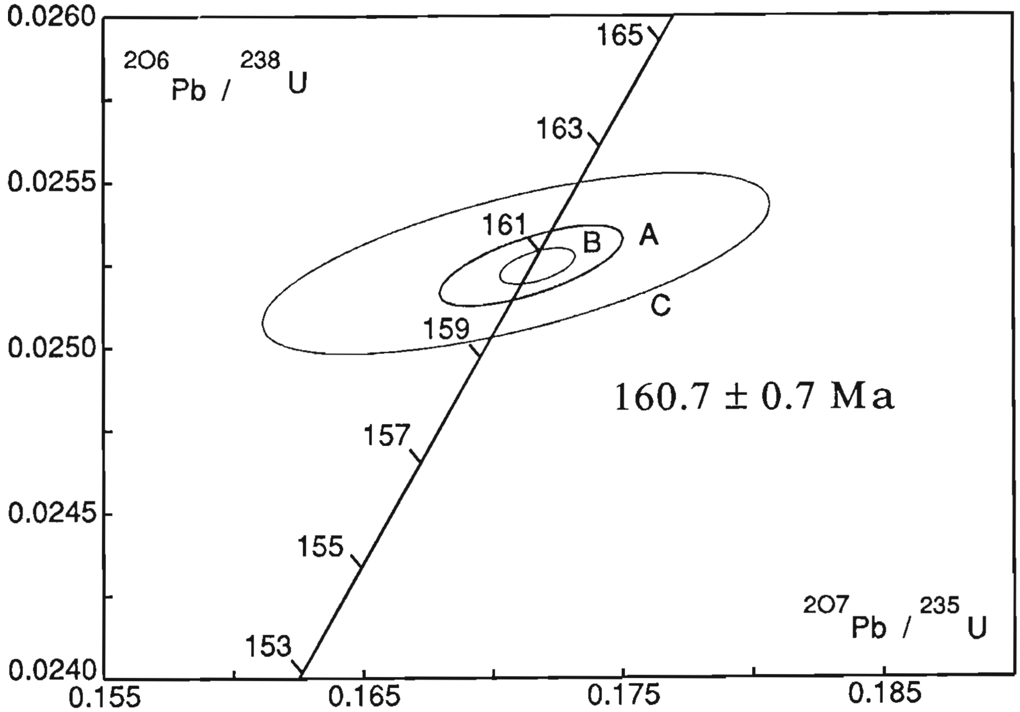

Figure 2.

U-Pb concordia plot for dacite boulder in Bowser Basin sediments. See Table 1 for analytical data.

Callovian period must extend to 161 Ma or slightly younger, and as such, comprises a constraint on the geological time scale in the Jurassic.

PROVENANCE CRITERIA

Based on textural and regional stratigraphic criteria, the following conditions relating to the provenance of the dacite clasts are inferred:

(1) The dacite and other felsic clasts were the products of first cycle erosion and deposition.

(2) Except for the dacite clasts, no other recognizable extraneous rock types entered the Cache Creek-derived sedimentary package in northern Bowser Basin.

(3) The U-Pb age and biostratigraphic age indicate that erosion of the volcanic source, or alternatively, unroofing of a high-level intrusion must have been rapid.

(4) Regional sediment transport in the northern coarse grained belt of Bowser Basin was towards the southwest.

POTENTIAL SOURCE AREAS

Unlikely sources

Felsic volcanic rocks of the Hazelton Group underlying Bowser Basin in Spatsizi and Cry Lake map areas (south of the King Salmon Fault), are mostly Early Jurassic in age and are not a likely source. Hazelton Group rocks in the Iskut map area to the west and southwest are pre-Bathonian (Anderson and Thorkelson, 1990) and are probably too far south and west of the area of dacite pebble accumulation.

Several potential source units bracketing the Early Callovian age exist in the central and southern parts of Quesnellia. If consideration is given to the hypothesis that Stikinia has been displaced northward with respect to Quesnellia (e.g. Gabrielse, 1985), then felsic source rocks could have shed debris into Bowser Basin from a more southerly location. However, it might be expected that additional clast lithologies derived from older sedimentary or metamorphic successions would also be present in the Jurassic Bowser Basin fill. In fact, the first indication of additional source rocks east of the basin is not seen until the Early Cretaceous when mica from the Omineca Belt was introduced into the Skeena Group (Tipper and Richards, 1976). Therefore, a Quesnellian source is considered unlikely for the felsic clasts.

Probable sources

A number of possible source rocks occur in the Cry Lake area. Felsic volcanic rocks are interbedded with Bowser-like conglomerate in the footwall of King Salmon fault, Mount Blair area (Gabrielse, 1991). Marine deposits associated with the conglomerate and volcanic units have early Bajocian faunas (H.W. Tipper, pers. comm., 1992), however the upper age limit of these deposits is not known.

The average of U-Pb ages for the Hotailuh Batholith is 170 ±9 Ma (Anderson, 1983) which at its upper age limit could have acted as a source of Bowser Basin sediment. North of Stikine River granodiorite plutons giving K-Ar ages ranging from 157.8 ±2.4 Ma to 160.8 ±2.5 Ma also intrude Triassic and Lower Jurassic volcanic and volcaniclastic assemblages (Gabrielse, 1979; Hunt and Roddick, 1987, p. 203), and cut younger structures possibly associated with the King Salmon Fault system (Gabrielse, 1991). Intruding the Cache Creek Terrane itself in the Dease Lake area are granodiorite stocks with K-Ar ages of 173 ±4 Ma and 160 ±20 Ma (Stevens et al. 1982, p. 16). Despite large errors in the age estimates these relatively shallow level intrusions provide viable alternatives as sources of the dacite clasts.

SUMMARY

U-Pb zircon dating of a dacite boulder from the Bowser Lake Group in Spatsizi map area indicates a dacite crystallization age of 160.7 ±0.7 Ma. Corresponding biostratigraphic control based on ammonite faunas indicates the host deposits are Early Callovian. Therefore, eruption and erosion of a volcanic source, or emplacement and unroofing of a high level intrusion and subsequent deposition of felsic debris must have been rapid.

The most likely source rocks for the felsic clasts were probably locally distributed. In the footwall of King Salmon Fault these include the latest intrusive phases of Hotailuh Batholith and associated plutons, some of which may have fed Bajocian and possibly younger felsic volcanic rocks in proximal Bowser Basin strata. Granodiorite stocks with ages that bracket the Early Callovian also occur in the hanging wall of King Salmon Fault and intrude the Cache Creek succession. It is not possible at present to distinguish between these alternative source types.

Based on petrographic criteria, a more southerly source area in Quesnellia seems unlikely. However, this interpretation in itself neither confirms or denies the hypothesis that Stikinia was displaced laterally in post-Callovian time relative to the western margin of composite North America.

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New Late Triassic and Early Jurassic U-Pb zircon ages from the Hotailuh Batholith, Cry Lake map area, north-central British Columbia

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Abstract

New Late Triassic (221 ± 3 Ma; Cake Hill Pluton) and Early Jurassic (184 ± 8 Ma; McBride River Pluton) U-Pb zircon ages from the Hotailuh Batholith agree with geological relationships and previously determined K-Ar ages. The new isotopic age for the McBride River pluton supersedes a previously published, poorly-constrained U-Pb (zircon) age of 166 ± 8 Ma. The new dates and a previously published U-Pb zircon date for the Three Sisters pluton of 170 ± 1 Ma indicate that three distinct plutonic episodes are represented in the Hotailuh Batholith. Late Triassic, Early Jurassic and Middle Jurassic plutons in the Hotailuh Batholith are parts of more widespread suites in northern Stikinia.

Résumé

Les nouveaux âges U-Pb sur zircon associés au Batholite de Hotailuh le font remonter au Trias tardif (221 ± 3 Ma; Pluton de Cake Hill) et au Jurassique précoce (184 ± 8 Ma; Pluton de McBride River); ils corroborent les liens géologiques et les datations par la méthode K-Ar établies antérieurement. Le nouvel âge isotopique du pluton de McBride annule un âge U-Pb sur zircon mal délimité de 166 ± 8 Ma déjà publié. Les nouvelles données et un âge U-Pb sur zircon antérieurement publié de 170 ± 1 Ma pour le pluton de Three Sisters indiquent que trois épisodes distincts caractérisent le Batholite de Hotailuh. Les plutons du Trias tardif, du Jurassique précoce et du Jurassique moyen du Batholite de Hotailuh font partie de suites plus vastes observées dans le nord de la Stikinie.

INTRODUCTION

Previous mapping and K-Ar and U-Pb geochronometry established composite Late Triassic and Early to Middle Jurassic plutonic suites in the Hotailuh Batholith (Anderson, 1983; Anderson et al., 1982; R.G. Anderson in Stevens et al., 1982b). Abundant intrusive and geological relationships and isotopic ages helped constrain the age of the Stikine and Three Sisters plutonic suites to circa 218-230 Ma and 166-188 Ma, respectively.

The purpose of this paper is to present new U-Pb results for the Late Triassic Cake Hill Pluton of the Stikine suite and the Early Jurassic McBride River Pluton of the Three Sisters suite in the batholith (Fig. 1). Advances in the analytical precision and accuracy of U-Pb analyses at the Geological Survey of Canada since 1982 (e.g., Parrish et al., 1987) permit resolution of the emplacement age for the Cake Hill Pluton, the most extensive and oldest of the Late Triassic plutons, previously estimated from K-Ar dates and intrusive relationships. Re-analysis of zircon from the same sample of the McBride River Pluton dated by Anderson et al. (1982)

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helps to resolve the apparent discordance between published K-Ar (hornblende) and U-Pb (zircon) ages for the pluton and to define a new Early Jurassic intrusive episode in the Triassic-Jurassic Hotailuh Batholith.

GEOLOGICAL SETTING

The Hotailuh (1100 km^2) and Stikine (300 km^2) batholiths typify the large, composite Late Triassic to Middle Jurassic ultramafic and granitic intrusions of northern and western Stikinia where they occur north and west of the Bowser Basin (Anderson, 1983, 1984, 1988; Woodsworth et al., 1991). The Hotailuh Batholith is a composite intrusion comprising plutonic suites of Late Triassic (Stikine Plutonic Suite) and Early and Middle Jurassic (Three Sisters Plutonic Suite) age. Both batholiths postdate the Tahltanian orogeny (Souther and Armstrong, 1966; Souther, 1971) and are mainly post-tectonic. They intrude polydeformed Mississippian and Permian metasedimentary rocks and metavolcanic rocks (Thorstad, 1980), undeformed Middle Triassic argillite, volcaniclastic rocks, andesite and chert ("Tsaybahe Group"; Read, 1983, 1984) and massive Middle and Upper Triassic volcaniclastic rocks, greywacke, and pillowed, clinopyroxene- and plagioclase-phyric basalt (Stuhini Group; Anderson, 1983). Contact metamorphosed volcanic and plutonic pendants are common and parts of the Jurassic plutonic suite are barely unroofed. Although rarely miarolitic, the suites are probably epizonal based on intrusive relations and preserved stratigraphy.

TRIASSIC STIKINE PLUTONIC SUITE

Four plutons represent the Triassic Stikine Plutonic Suite in Hotailuh Batholith (Fig. 2a) and intruded in the following order: 1) Stikine Pluton (intensely foliated biotite-hornblende diorite); 2) Cake Hill Pluton (massive to moderately foliated or lineated hornblende quartz monzodiorite, granodiorite, monzodiorite and quartz monzonite); 3) Gnat Lakes

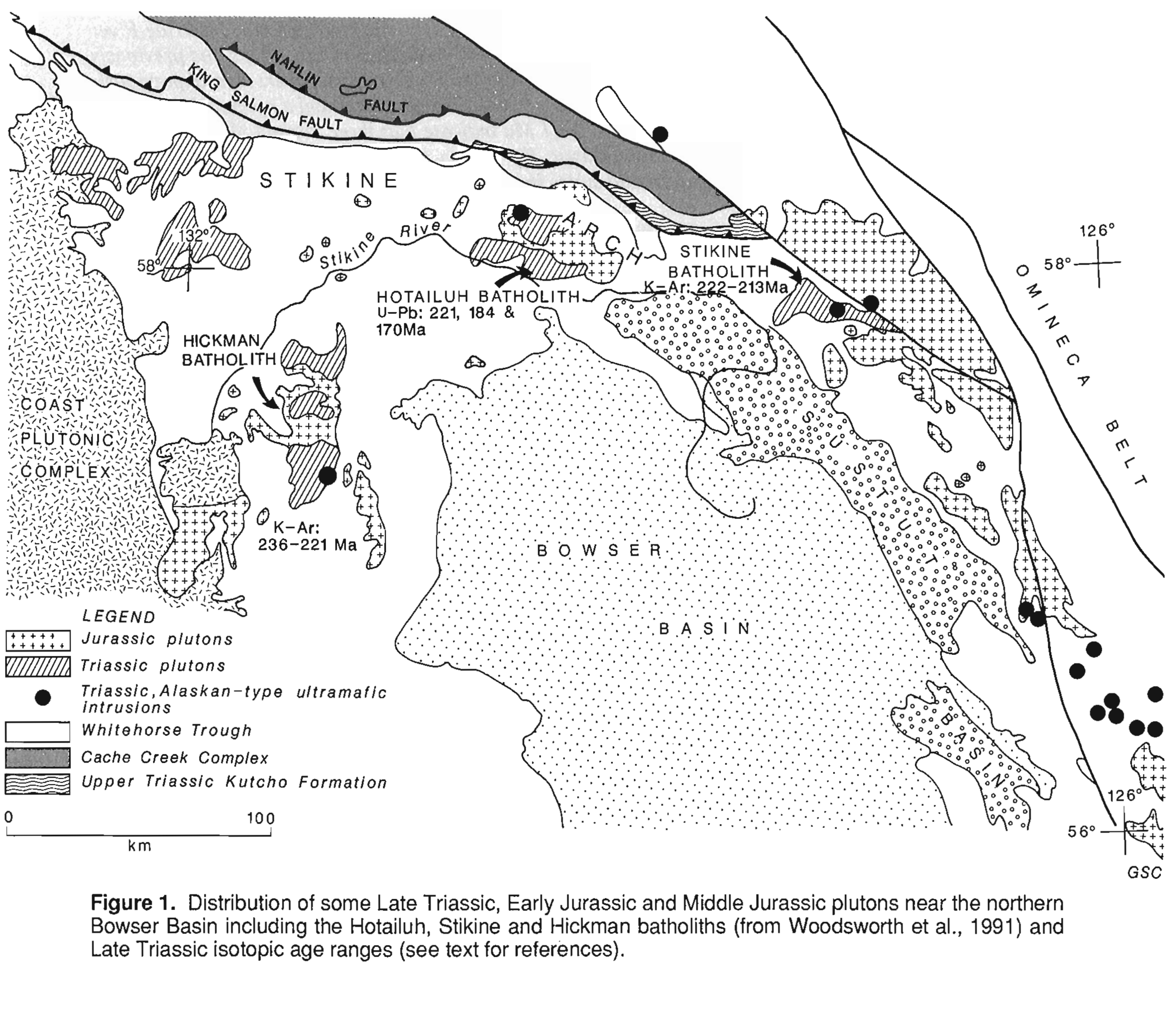

Figure 1. Distribution of some Late Triassic, Early Jurassic and Middle Jurassic plutons near the northern Bowser Basin including the Hotailuh, Stikine and Hickman batholiths (from Woodsworth et al., 1991) and Late Triassic isotopic age ranges (see text for references).

Ultramafite (hornblende clinopyroxenite and hornblendite); and 4) Beggerlay Creek Pluton (highly altered hornblende gabbro and diorite). The Cake Hill Pluton is the most extensive; the other plutons occur around its margins. Diagnostic interplutonic intrusive relations are inclusions of older plutons in younger plutons, intrusion of apophyses of younger plutons in older plutons and deformation of older plutons along intrusive contacts of younger plutons. Steep, east- to northeast-trending mineral foliation in the Stikine Pluton and steep northwest-trending mineral foliation in the Cake Hill Pluton distinguish them from the massive to locally compositionally-layered Beggerlay Creek and Gnat Lakes plutons.

JURASSIC THREE SISTERS PLUTONIC SUITE

The Jurassic Three Sisters Plutonic Suite consists of batholithic bodies (Three Sisters and McBride River plutons) and satellitic intrusions (Snowdrift Creek, Pallen Creek and Tanzilla plutons) (Fig. 2b,c). The new U-Pb zircon age for the McBride River pluton is older than a previously determined U-Pb zircon age for the potassic marginal phase of the Three Sisters Pluton (Anderson et al., 1982).

The McBride River Pluton comprises siliceous, mafic-poor hornblende-biotite granodiorite and quartz monzodiorite. It is characteristically homogeneous, massive, and contains rare mafic inclusions. In contrast, the Three Sisters Pluton is heterogeneous and composite, contains common clinopyroxene, hornblende, biotite, and mafic inclusions, is rarely foliated, and contains an expanded compositional suite. Five sequentially intruded, mafic to felsic phases are recognized (Anderson, 1979); they range in composition from diabase of the fine grained phase, hornblende gabbro or diorite of the mafic phase, quartz monzodiorite or granodiorite of the central phase, and quartz monzonite or quartz syenite of the potassic marginal phase, to leucocratic quartz diorite and granodiorite of the leucocratic phase.

STRATIGRAPHIC AGE CONSTRAINTS

Stratigraphic constraints on Triassic plutons

Relationships between nearly coeval intrusive and overlying rocks constrain the emplacement age of most of the Late Triassic plutonic suite (including the Cake Hill Pluton) to pre-Carnian or Norian (Anderson, 1979). Non-granitoid country rocks for the Triassic plutonic suite were not seen in contact with the Cake Hill Pluton although small outcrops of Permian carbonate and marble occur southwest and west of the batholith. The exception is the Beggerlay Creek Pluton, the intrusion of which metamorphosed rocks of the Middle and Upper Triassic Stuhini Group. The Triassic plutonic suite was the plutonic root for porphyritic flows and volcaniclastic rocks that nonconformably overlie it; the pluton is also nonconformably overlain by sedimentary rocks of the Middle and Upper Triassic Stuhini Group. Toarcian sedimentary rocks also nonconformably overlie the Cake Hill Pluton (Anderson, 1980; Henderson and Perry, 1981).

Stratigraphic constraints on Jurassic plutons

Intrusive and metamorphic relationships provide several maximum age constraints for emplacement of the Jurassic plutonic suite. Three Sisters Pluton is post-Toarcian; Snowdrift Creek Pluton is post-Pliensbachian; McBride River Pluton is post-Sinemurian or Toarcian (H. Gabrielse, unpublished data); and Pallen Creek and Tanzilla plutons are post-Carnian or Norian (Anderson, 1983; Stevens et al., 1982b). The Triassic plutonic suite formed part of the country rock for the emplacement of the Jurassic plutonic suite and locally was remobilized and metamorphosed as a result. The Jurassic plutonic suite intruded and metamorphosed the Mesozoic volcanic and sedimentary rocks and locally contains inclusions of them. Middle and Upper Triassic Stuhini Group rocks are intruded by all plutons in the suite; the Middle Triassic siliceous argillite occurs as inclusions in the potassic marginal phase of the Three Sisters Pluton. Sinemurian or Toarcian limestone- and greenstone-bearing conglomerate is hornfelsed adjacent the northeast margin of the McBride River Pluton (H. Gabrielse, unpublished data, 1992). Structures in Pliensbachian sedimentary rocks, related to early movements on the King Salmon Fault, are crosscut by the Snowdrift Pluton. Triassic-Jurassic volcanic rocks along the northeast margin of the batholith are crosscut and Toarcian sedimentary rocks were metamorphosed during intrusion of the Three Sisters Pluton.

Intrusive relations between the Three Sisters and McBride plutons were not established. Intrapluton intrusive relations in the Three Sisters Pluton indicate a general mafic to felsic, radially east to southeastward emplacement of four of the five phases (Anderson, 1979). The exception is the fine grained phase which occurs between the leucocratic and potassic marginal phases. The leucocratic phase, east of the fine-grained phase, is compositionally similar to the McBride River pluton, is anomalous within the petrological trend represented in the Three Sisters Pluton, and has petrographic textures indicating widespread mineral overgrowths (Anderson, 1983). The leucocratic phase has a K-Ar age (185 ± 28 Ma) that is concordant with the McBride Pluton (186 ± 13 Ma) but that is anomalously old compared with other phases of Three Sisters Pluton (R.G. Anderson in Stevens et al., 1982b). The leucocratic phase may represent the western margin of the McBride River Pluton which was metamorphosed during intrusion of the younger Three Sisters Pluton.

PREVIOUS K-AR AND U-PB ISOTOPIC DATING

Late Triassic Stikine plutonic site

Meaningful K-Ar ages for the Triassic suite (ranging from 218-230 Ma) are consistent with the stratigraphic constraint on the youngest age of emplacement of the Triassic plutonic suite but do not resolve the sequence of intrusion (Fig. 2a; Anderson, 1980; R.G. Anderson in Stevens et al., 1982b). For the Cake Hill Pluton, dated in this study, K-Ar age determinations include replicate analyses of hornblende from the interior of the pluton (218 ± 11 and 220 ± 11 Ma) and a

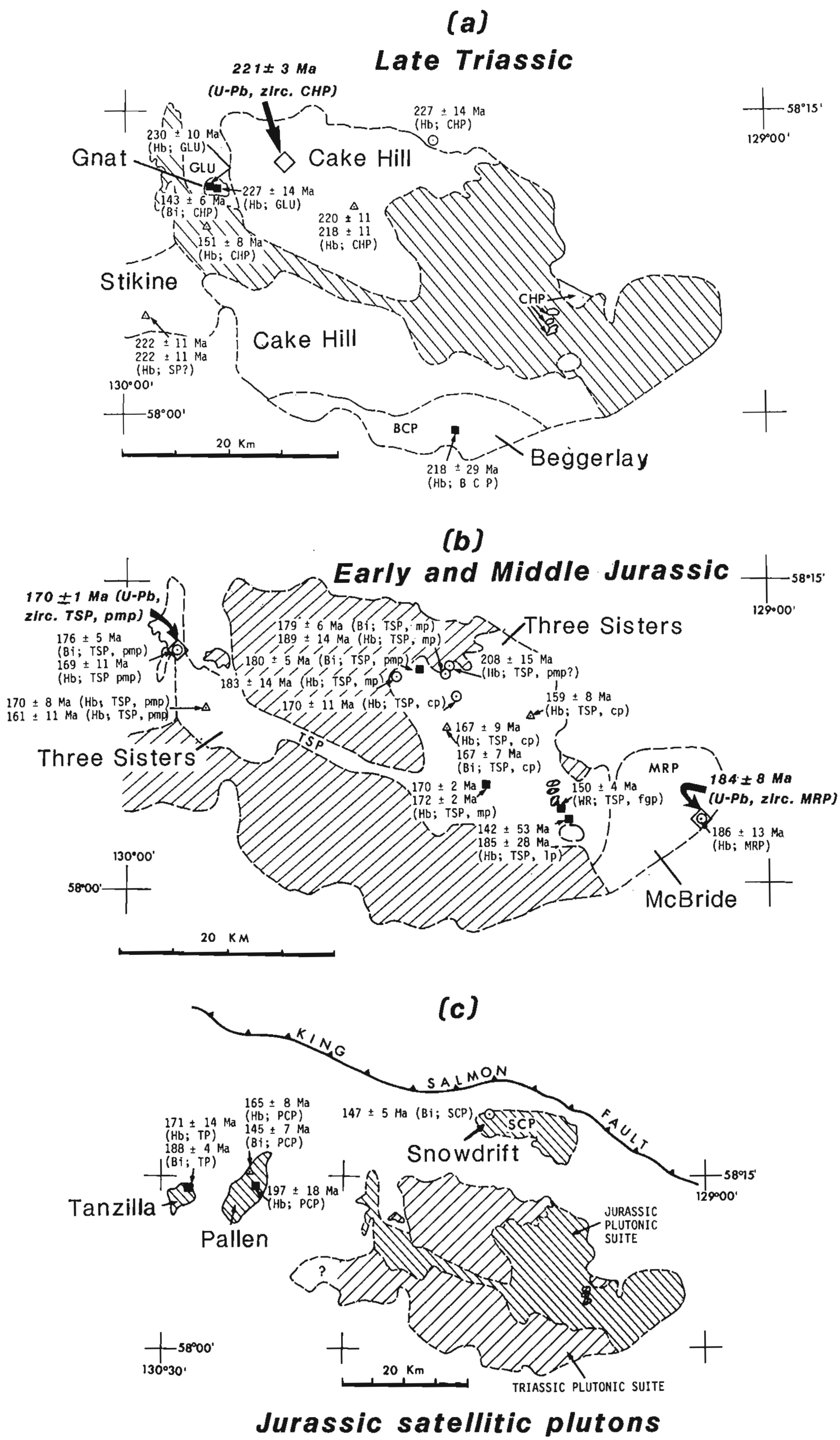
(a)
Late Triassic
221± 3 Ma
(U-Pb, zirc. CHP)
227 ± 14 Ma
(Hb; CHP)
58°15'
129°00'
230 ± 10 Ma
(Hb; GLU)
Gnat
GLU
Cake Hill
227 ± 14 Ma
(Hb; GLU)
143 ± 6 Ma
(Bi; CHP)
220 ± 11
218 ± 11
(Hb; CHP)
151 ± 8 Ma
(Hb; CHP)
Stikine
CHP
222 ± 11 Ma
222 ± 11 Ma
(Hb; SP?)
Cake Hill
130°00'
58°00'
BCP
20 Km
218 ± 29 Ma
(Hb; B C P)
Beggerlay
(b)
Early and Middle Jurassic
58°15'
129°00'
170 ±1 Ma (U-Pb,
zirc. TSP, pmp)
179 ± 6 Ma (Bi; TSP, mp)
189 ± 14 Ma (Hb; TSP, mp)
Three Sisters
176 ± 5 Ma
(Bi; TSP, pmp)
169 ± 11 Ma
(Hb; TSP pmp)
180 ± 5 Ma (Bi; TSP, pmp)
208 ± 15 Ma
(Hb; TSP, pmp?)
183 ± 14 Ma (Hb; TSP, mp)
170 ± 8 Ma (Hb; TSP, pmp)
161 ± 11 Ma (Hb; TSP, pmp)
170 ± 11 Ma (Hb; TSP, cp)
159 ± 8 Ma
(Hb; TSP, cp)
167 ± 9 Ma
(Hb; TSP, cp)
167 ± 7 Ma
(Bi; TSP, cp)
Three Sisters
TSP
MRP
184 ± 8 Ma
(U-Pb, zirc. MRP)
170 ± 2 Ma
172 ± 2 Ma
(Hb; TSP, mp)
150 ± 4 Ma
(WR; TSP, fgp)
142 ± 53 Ma
185 ± 28 Ma
(Hb; TSP, lp)
186 ± 13 Ma
(Hb; MRP)
130°00'
58°00'
McBride
20 KM
(c)
KING SALMON FAULT
165 ± 8 Ma
(Hb; PCP)
145 ± 7 Ma
(Bi; PCP)
147 ± 5 Ma (Bi; SCP)
SCP
171 ± 14 Ma
(Hb; TP)
188 ± 4 Ma
(Bi; TP)
Snowdrift
58°15'
129°00'
197 ± 18 Ma
(Hb; PCP)
Tanzilla
Pallen
JURASSIC
PLUTONIC
SUITE
?
58°00'
130°30'
20 Km
TRIASSIC PLUTONIC SUITE
Jurassic satellitic plutons

Table 1. U-Pb analytical data

Zircon fraction	Weight (mg)	U (ppm)	Pb* (ppm)	Measured $^{206}Pb/^{204}Pb$	$^{208}Pb^*$ (%)	Isotopic ratios $^{206}Pb^*/^{238}U$	$^{207}Pb^*/^{235}U$	$^{207}Pb^*/^{206}Pb^*$ age (Ma)
AN-78-569-3 Cake Hill quartz monzonite; 58°12'34"N, 129°45'00"W								
A. N +105	0.0718	166	5.8	1737	11.2	0.03455	0.2408	220.6 ± 7.3
B. N +105	0.0942	194	6.7	2601	10.2	0.03445	0.2401	219.9 ± 4.7
C. N +105	0.0707	152	5.4	2017	11.9	0.03448	0.2404	221.7 ± 5.8
AN-78-467 McBride River granodiorite; 58°03'12"N, 129°08'00"W								
A. N +105	0.0399	367	8.5	1401	12.4	0.02236	0.1533	181.6 ± 8.6
B. W +149	0.0276	571	19.1	3114	10.2	0.03318	0.2348	255.5 ± 4.4
C. W +105-149	0.0337	535	12.0	1347	12.8	0.02160	0.1485	188.3 ± 6.9
D. N +105	0.0295	242	6.4	901	12.7	0.02562	0.1747	169.8 ± 16.1

*radiogenic Pb, blank corrected.
Note: Numbers such as "+105" refer to size in microns. N= non-magnetic at <1° side tilt, 1.7 amps on Frantz magnetic separator; W= non-magnetic at <2°, >1°, 1.7 amps. Total procedural blanks for Pb in zircons ranged from 9 to 13 pg, and U blanks averaged <1 pg. Errors were propogated numerically (Roddick, 1987) and are quoted at the 2σ level, with $^{207}Pb/^{235}U$ and $^{206}Pb/^{238}U$ errors averaging 0.42% and 0.20%, respectively. All age determinations were calculated using the decay constants of Steiger and Jäger (1977).

hornblende date (227 ± 14 Ma) from a Cake Hill Pluton fragment in the overlying Middle and Upper Triassic Stuhini Group volcaniclastic rocks (Fig. 2a; R.G. Anderson in Stevens et al., 1982b). Intrusive relations suggest that the Cake Hill Pluton must have intruded before the Gnat Lakes Ultramafite that is dated at 230 ± 10 and 228 ± 14 Ma (both K-Ar, hornblende; Fig. 2a) (R.G. Anderson in Stevens et al., 1982b).

Jurassic Three Sisters plutonic site

Potassium-argon isotopic ages for the Jurassic plutonic suite range from 142 to 208 Ma (Fig. 2b,c; R.G. Anderson in Stevens et al., 1982b). A single K-Ar hornblende age of 186 ± 13 Ma (GSC 80-11; Stevens et al., 1982a) was attributed to excess argon because the same McBride River pluton sample yielded U-Pb zircon data interpreted to indicate an age of 166 ± 8 Ma (Anderson et al., 1982).

Most of the Three Sisters pluton in the northeastern part of the batholith (Fig. 2b) is thought to have been intruded at about 180-187 Ma based on the oldest K-Ar isotopic ages and stratigraphic constraints on maximum emplacement age using the revised Jurassic time scale of Harland et al. (1990). In the western part of the pluton (Fig. 2b), K-Ar isotopic ages range from 161 to 176 Ma and are concordant with the 170 ± 1 Ma U-Pb age for zircon from the potassic marginal phase of the Three Sisters Pluton (Anderson et al., 1982).

Figure 2. K-Ar and U-Pb geochronometry of batholithic Late Triassic (Fig. 2a) and Jurassic (Fig. 2b) plutons and plutons satellitic (Fig. 2b) to the Hotailuh Batholith; note change in scale for Fig. 2c (modified from Stevens et al., 1982b). Hatchured areas indicate distribution of Late Triassic plutons (hatchures with positive slope) and Jurassic plutons (hatchures with negative slope). New U-Pb dates for Late Triassic Cake Hill Pluton (CHP) and Early Jurassic McBride River Pluton (MRP) as well as previously published date (Anderson et al., 1982) for the potassic marginal phase (pmp) of the Three Sisters Pluton (TSP) are shown by diamond symbols. Other symbols denote previously published dates in Wanless et al., (1972; triangles), Stevens et al. (1982a; circles), and Stevens et al. (1982b; filled squares). Other abbreviations for Triassic plutons are: BCP = Beggerlay Creek Pluton; GLU = Gnat Lakes Ultramafite; SP = Stikine Pluton. For Jurassic plutons, other abbreviations are: PCP = Pallen Creek Pluton; SCP = Snowdrift Creek Pluton; TP = Tanzilla Pluton; and for Three Sisters Pluton, cp = central phase; fgp = fine-grained phase; lp = leucocratic phase; mp = mafic phase. Material dated is: Bi = biotite, Hb = hornblende,.WR = whole rock, and zirc. = zircon.

U-PB ISOTOPIC DATA

Sampling and analytical techniques

Zircon was concentrated from 25 kg of each rock sample, and individual zircon populations were separated initially on the basis of grain size and magnetic susceptibility. Grains selected for analysis were hand picked using a binocular microscope to isolate the most euhedral, optically homogeneous, fracture-free and inclusion-free grains. All zircon fractions were abraded to remove potentially altered outer parts of the grains (Krogh, 1982). Procedures for dissolution, separation of Pb and U, purification techniques employing a ^{205}Pb-$^{233-235}U$ spike, mass spectrometry, and data reduction were the same as those detailed in Parrish et al. (1987).

U-PB RESULTS AND INTERPRETATION

U-Pb analytical results for both samples are given in Table 1. Zircons from the Cake Hill Pluton quartz monzonite are pale yellow prisms with dipyramids characterized by

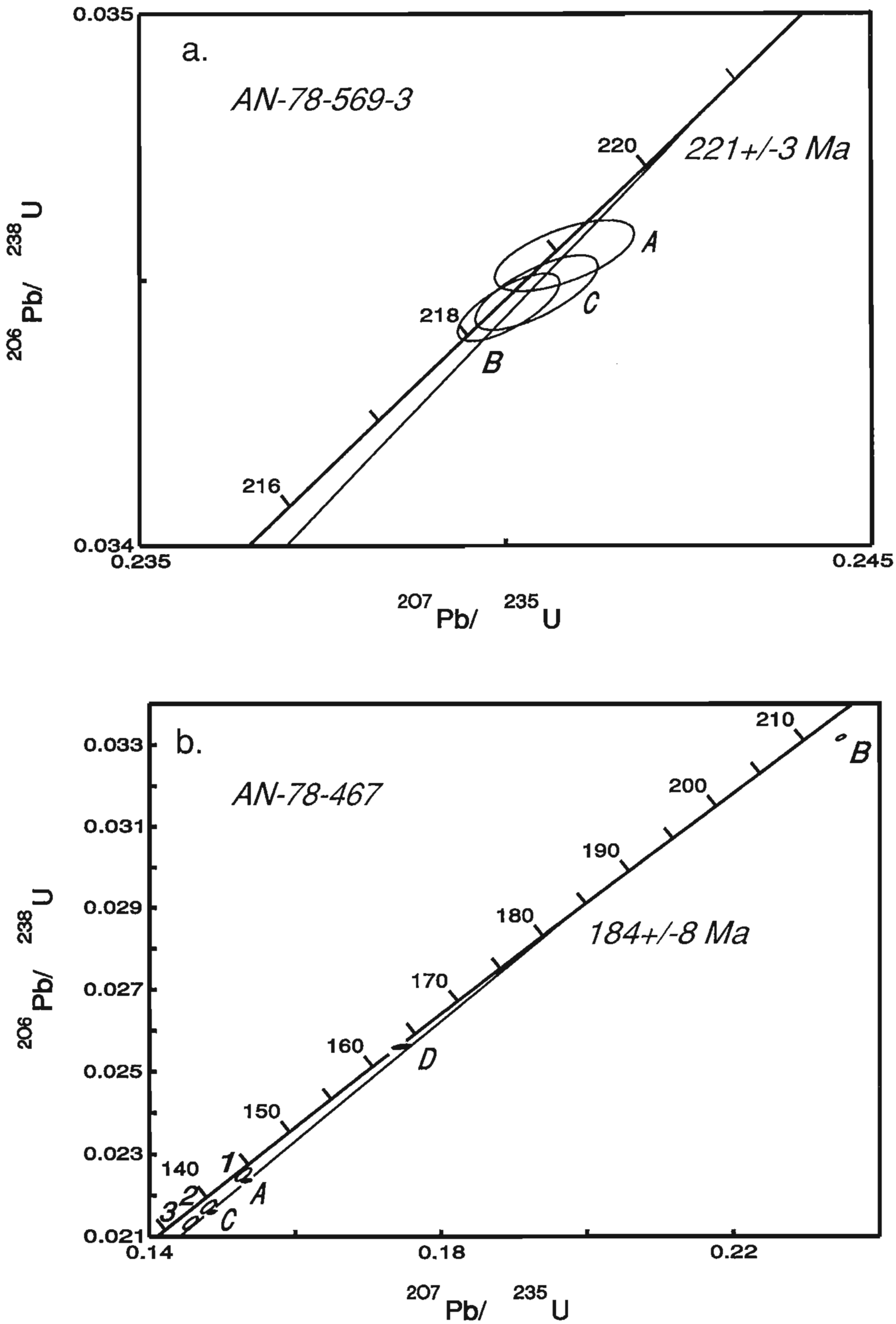

Figure 3. **(a)** U-Pb concordia plot for Cake Hill Pluton quartz monzonite. **(b)** U-Pb concordia plot for McBride River granodiorite. Fractions 1, 2, and 3 are from Anderson et al. (1982).

length:breadth ratios of 2:1 to 3.5:1. The grains have excellent clarity and rare inclusions of colourless bubbles or rods. Three analyses partially overlap each other and are 0.7-1.5% discordant (Fig. 3a). A weighted ^{207}Pb-^{206}Pb age for the sample is 221 ± 3 Ma and is the best estimate of the crystallization age of the sample.

Zircons from the McBride River Pluton granodiorite are pale yellow, euhedral, multifaceted grains with length: breadth of 1.5:1 to 2.5:1. The grains exhibit good to excellent clarity, rare inclusions of colourless tubes and bubbles, and no visible zoning. Four analyses display a pattern on a concordia plot that indicates both Pb loss and inheritance (Fig. 3b). Analyses A, C, and D have similar radiogenic $^{208}Pb(\%)$ (and hence Th/U) values that are unlike that for analysis B. It is unlikely that all of the zircons in B crystallized out of the same magma as zircons in the other fractions. The ^{207}Pb-^{206}Pb age of fraction B indicates an inherited component with a minimum age of 255 ± 4 Ma. A weighted ^{207}Pb-^{206}Pb age for fractions A, C, and D is 184 ± 8 Ma; we consider this to be the best estimate of the crystallization age of the sample.

The previously determined U-Pb data of Anderson et al. (1982) are also plotted in Figure 3b. These analyses show a systematic shift to lower ^{207}Pb-^{235}U, and ^{206}Pb-^{204}Pb values, and contain higher common Pb than our analyses, characteristics that indicate a systematic error in the analyses due to inaccurate estimation of the common Pb composition.

Discussion

The U-Pb crystallization date for zircon from the Late Triassic Cake Hill Pluton (221 ± 3 Ma) agrees closely with both previously determined K-Ar (hornblende) dates (220 ± 11 Ma and 218 ± 11 Ma) from a sample in the northern part of the Pluton near the zircon sample location (Fig. 2a).

The U-Pb crystallization age for the Cake Hill Pluton corroborates the minimum age derived from the stratigraphic constraints on its age and is consistent with the 223 ± 10 Ma Carnian-Norian stage boundary of Harland et al. (1990). It suggests that an earlier minimum estimate for intrusion of 227-230 Ma (based on a K-Ar age for hornblende in Cake Hill Pluton fragments in Stuhini Group rocks and in the crosscutting Gnat Lakes Ultramafite; R.G. Anderson in Stevens et al., 1982b) should be reduced. The older K-Ar ages in the Gnat Lakes Ultramafite are attributed to excess argon derived from clinopyroxene commonly enclosed in hornblende of the pluton's hornblendite and hornblende clinopyroxenite phases. As an estimate for intrusion of the Stikine Plutonic Suite in the Hotailuh Batholith, the 221 Ma date for the Cake Hill Pluton is not only concordant with available K-Ar dates from the batholith but is similar to K-Ar dates from the Stikine Batholith to the east (average hornblende K-Ar date of 217 ± 3 Ma; R.G. Anderson in Hunt and Roddick, 1987); and plutons in the Hickman batholith (236-221 Ma; Holbek, 1988) to the southwest; and U-Pb isotopic ages (226-221 Ma; Bevier and Anderson, 1991) for Late Triassic plutons in the Iskut map area to the south (Fig. 1).

The U-Pb crystallization age of 184 ± 8 Ma for the McBride River pluton is the same, within error, as a K-Ar (hornblende) age for the same sample. The date corroborates the Toarcian or older maximum age based on stratigraphic constraints and is consistent with the best estimates for the base and top of the Toarcian (187 ± 15 and 178 ± 11 Ma, respectively) by Harland et al. (1990). The date establishes a new, compositionally distinct, magmatic episode as an important part of the complex evolution of the Hotailuh Batholith. Toarcian magmatism is becoming recognized increasingly along the northeastern (e.g., Gabrielse et al., 1980; Woodsworth et al., 1991), northern (this study), and western (Brown et al., 1992; Macdonald et al., 1992) flanks of the Bowser Basin.

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U-Pb zircon ages for the Hazelton Group and Cone Mountain and Limpoke plutons, Telegraph Creek map area, northwestern British Columbia: age constraints on volcanism and deformation

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Abstract

U-Pb ages of 185 +7/-1 Ma and 175 +4/-1 Ma are reported for two volcanic rocks in an outlier of Hazelton Group within Telegraph Creek map area. This Toarcian to Aalenian succession unconformably overlies late Norian Stuhini Group strata, and correlates with Hazelton Group rocks to the east and south. Collectively, these Jurassic volcanic rocks record a significant and widespread phase of calc-alkaline arc volcanism in western and northern Stikine Terrane.

The pre-kinematic Cone Mountain pluton yielded a U-Pb zircon date of 184.7 ± 0.6 Ma and thus represents coeval Toarcian plutonism. The eastern margin of this pluton was mylonitized during a southwest-directed contractional event, perhaps synchronous with development of the regionally important King Salmon Fault. The Limpoke pluton, located 35 km northwest of the Hazelton Group outlier, intrudes Stuhini Group rocks, and has a U-Pb age of 194 ± 2 Ma.

Résumé

Des âges U-Pb de 185 +7/-1 Ma et de 175 +4/-1 Ma ont été obtenues pour deux échantillons de roches volcaniques provenant d'une butte-témoin du Groupe de Hazelton dans la région cartographique du ruisseau Telegraph. Cette succession du Toarcien à l'Aalénien repose en discordance sur des couches du Groupe de Stuhini du Norien tardif, en plus de correspondre aux roches du Groupe de Hazelton à l'est et au sud. Globalement, ces roches volcaniques jurassiques témoignent d'une phase significative et étendue de volcanisme d'arc calco-alcalin dans l'ouest et le nord du terrane de Stikine.

Le pluton pré-cinématique de Cone Mountain a donné un âge U-Pb sur zircon de 184,7 ± 0,6 Ma et, par conséquent, représente un épisode de plutonisme toarcien contemporain. La bordure orientale de ce pluton a été mylonitisée durant un événement de contraction à direction sud-ouest, peut-être synchrone avec la formation de la faille de King Salmon d'importance régionale. Le pluton de Limpoke, situé à 35 km au nord-ouest de la butte-témoin du Groupe de Hazelton, recoupe par intrusion des roches du Groupe de Stuhini; il a été daté par la méthode U-Pb à 194 ± 2 Ma.

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INTRODUCTION

We report U-Pb ages for four samples of volcanic and plutonic rocks that help constrain the distribution of early Mesozoic volcanism and timing of deformational events in Telegraph Creek map area, which was mapped at a 1:50 000 scale by British Columbia Geological Survey Branch personnel (Brown and Gunning, 1989; Brown and Greig, 1990; Brown et al., 1990, 1992). The study area, located approximately 40 km southwest of Telegraph Creek (Fig. 1), is underlain predominantly by Upper Triassic Stuhini Group and Lower to Middle Jurassic Hazelton Group rocks. These Mesozoic island-arc volcanic rocks typify Stikine Terrane (Fig. 1). Jurassic volcanic rocks in a 100 km^2 outlier west of Yehiniko Lake represent the northwesternmost extent of the Hazelton Group recognized thus far (Figs. 1 and 2).

LOCAL GEOLOGY

The study area is underlain by an outlier of Jurassic volcanic and sedimentary rocks of the Hazelton Group. The northwest portion of the study area (Fig. 2) was first mapped by Kerr (1948). He identified a thin horizon of Jurassic sedimentary rocks and a larger area of what he believed to be Upper Jurassic and/or Lower Cretaceous strata (now includes part of the Hazelton outlier). Souther (1972) combined Kerr's two units and assigned them an Early Jurassic age. However, Souther correlated the southern quarter of the outlier (from Crocus Mountain to Site A in Fig. 2) with the older Stuhini Group of Late Triassic age. Our more detailed work suggests that much of this outlier is also of Early to Middle Jurassic age.

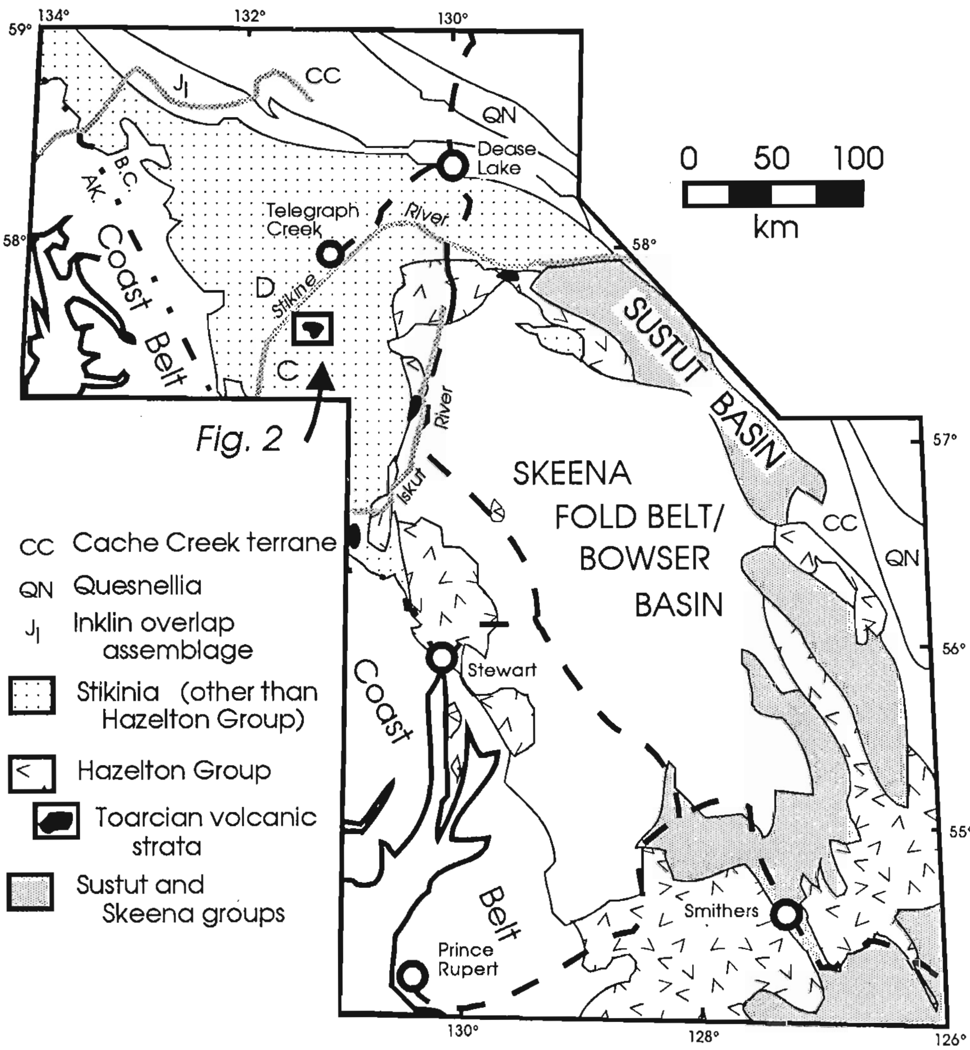

Figure 1. Regional setting and distribution of Hazelton Group rocks and the location of the study area (Fig. 2). Geology simplified from Wheeler and McFeely (1991). Also shown are U-Pb sample localities C and D.

The Hazelton Group section in the study area is characterized by exposures of moderately to gently dipping, maroon to mauve andesite flows and block-tuffs. These strata unconformably overlie deformed Upper Triassic rocks. Locally, the base of the Hazelton Group succession, southwest of Crocus Mountain (Site B in Fig. 2), comprises a tan weathering, aphanitic felsic flow. Stratigraphically above this, a discontinuous intravolcanic limy wacke horizon contains ammonite and *Weyla* fragments, belemnites, abundant Terebratulid brachiopods and scarce bivalves, that suggest a Toarcian age (locality F1 in Fig. 2) (Tipper, unpublished data, 1989). Based on rock type and age these sedimentary rocks correlate with the lower member of the Salmon River Formation in the upper part of the Hazelton Group (Anderson and Thorkelson, 1990). Overlying the wacke horizon are dark purple to maroon pyroxene and

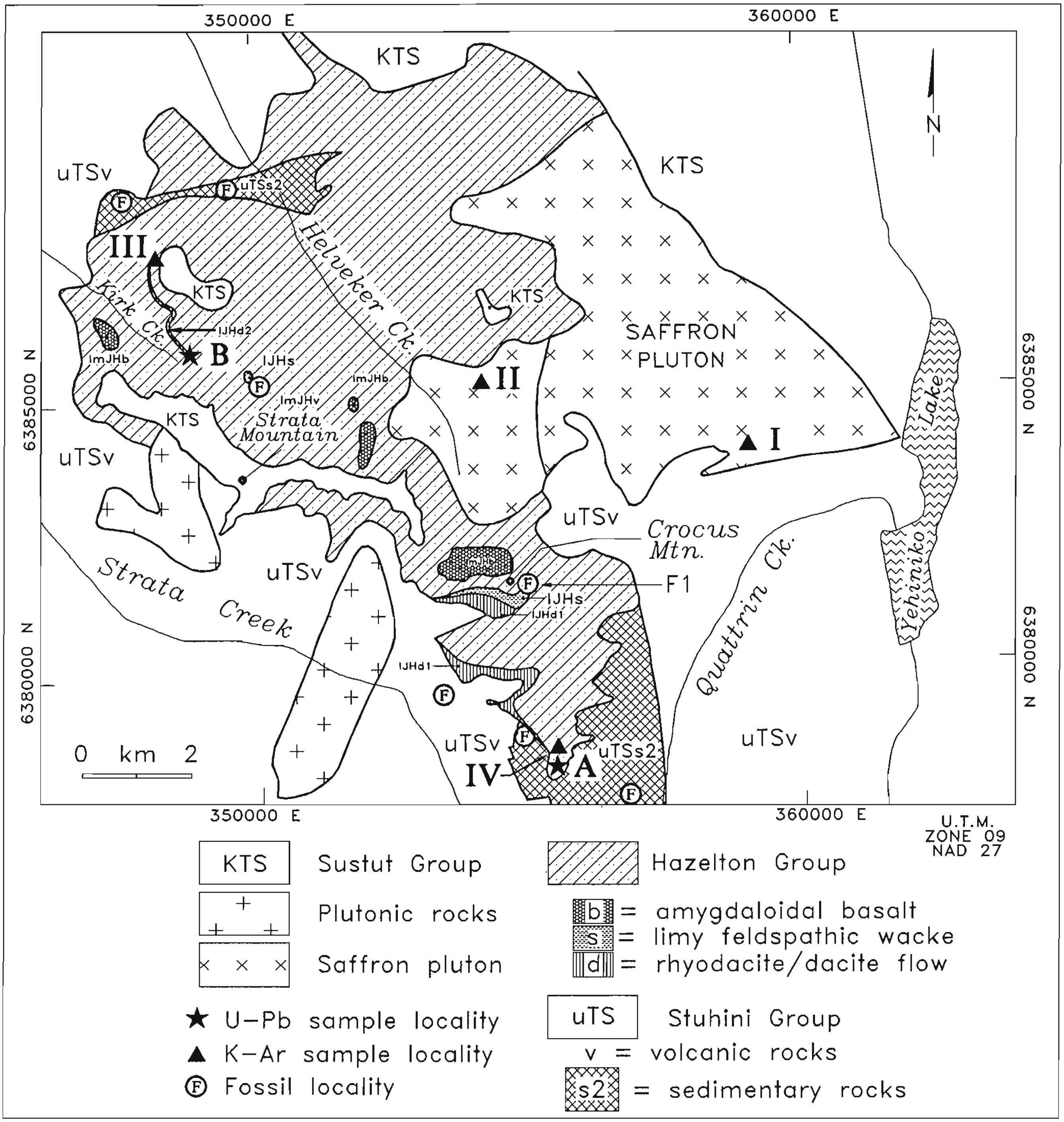

Figure 2. Simplified geology of the study area. U-Pb and K-Ar dating localities are numbered A and B and I to IV, respectively. Geology from Brown et al. (1990). Two distinct felsic units are labelled lJHd1 and lJHd2.

plagioclase-phyric andesite flows. The flows may be age-equivalent to the Snippaker Mountain volcanic facies, which is the upper member of the Salmon River Formation (ibid.).

Immediately to the south, a similar series of andesite tuffs, breccias and flows lies above an angular unconformity with fine grained Norian clastic rocks (Tozer, unpublished data, 1988; Orchard, unpublished data, 1990). Basalt pillow breccia and amygdaloidal flows are the highest stratigraphic unit and they may correlate with thick accumulations of Bajocian basalt in eastern Telegraph Creek map area. K-Ar whole rock ages from samples of Hazelton Group volcanic rocks within the outlier are 166 ± 6 Ma for pyroxene-plagioclase porphyritic andesite and 148 ± 5 Ma for greenschist grade, crystal lithic andesite lapilli tuff (localities III and IV respectively in Fig. 2; Brown et al., 1990).

The Saffron pluton and associated dykes, that were formerly included as part of the Yehiniko pluton of Brown and Greig (1990), intrude and locally hornfels the volcanic succession (Fig. 2). The Saffron pluton consists of distinctive pale brown to pink weathering, medium grained hornblende biotite granite to quartz monzonite that grades into a subordinate pale grey quartz monzodiorite phase. Biotite from granite and hornblende from quartz monzodiorite of the Saffron pluton yield concordant K-Ar ages of 162 ± 7 Ma (localities I and II, Fig. 2; Brown et al., unpublished data). These ages are somewhat younger than the ca. 172 Ma expected based on correlations with the Yehiniko Pluton (Holbek, 1988) and Three Sisters suite of the Hotailuh batholith (Anderson, 1983; Bevier and Anderson, 1991). The K-Ar ages do, however, provide a minimum age for the volcanic rocks in this area.

U-Pb GEOCHRONOLOGY

Sample selection

Two volcanic units were sampled for U-Pb dating in an attempt to constrain the age of the succession and to place a younger limit on the age of deformation that affected the underlying Upper Triassic rocks. Sample A (DBR-90-162) was collected directly above the basal unconformity exposed between Strata and Quattrin creeks (Fig. 2, 3). The sample is from an andesitic tuff unit within a sequence of tuffs, breccias

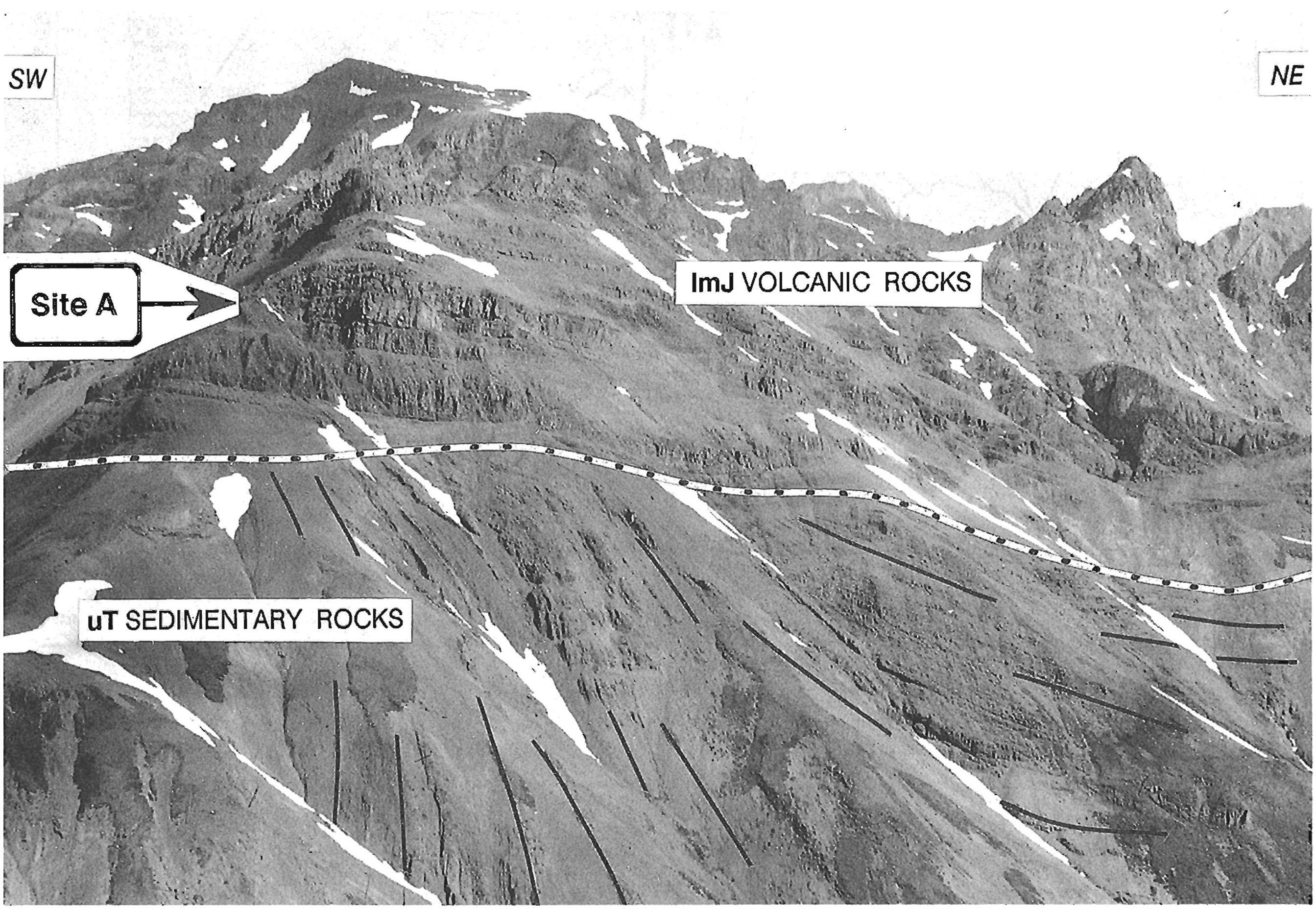

Figure 3. View northwest to gently dipping Lower to Middle Jurassic volcaniclastic rocks and flows unconformably overlying folded and faulted Norian siltstone and sandstone, showing site of sample A, on ridge between Strata and Quattrin creeks. Dotted line corresponds to the unconformity.

and flows that unconformably overlie Norian clastic rocks. Farther north near Kirk Creek, a series of andesite flows includes a distinct pink-weathering, hematitic, flow-banded, aphanitic rhyolite (Sample B; DBR-90-717) (Fig. 2, 4).

Sample C (DBR-90-40) is from a massive, medium grained biotite hornblende granodiorite of the Cone Mountain pluton in the footwall of the Cone Mountain fault, a mylonitic zone 500 m wide that comprises the eastern margin of the pluton (Fig. 1). The mylonite zone formed during a southwest-directed contractional event that occurred after emplacement of the pluton. Sample D (DBR-90-723) is from medium grained, biotite hornblende quartz monzodiorite of the Limpoke pluton (Fig. 1).

Analytical techniques

Zircon was concentrated from 25 kg of each rock sample, and individual zircon populations were separated initially on the basis of grain size and magnetic susceptibility. Grains selected for analysis were hand picked using a binocular microscope to isolate the most euhedral, optically homogeneous, fracture-free and inclusion-free grains. Most zircon fractions were abraded to remove potentially altered outer parts of the grains (Krogh, 1982). Procedures employed at the Geological Survey of Canada (GSC) for dissolution, separation of Pb and U, purification techniques employing a ^{205}Pb-$^{233-235}$U spike, mass spectrometry, and data reduction follow those detailed in Parrish et al. (1987). Error analysis was done using the method of Roddick (1987). Analytical methods used at the University of California at Santa Barbara (UCSB) are modified slightly from those of Parrish et al. (1987), and uncertainties were calculated using the error analysis method of Mattinson (1987). Pb and U blanks were 10-16 pg and <1 pg, respectively, for GSC analyses, and 10-50 pg and 2 pg, respectively, for UCSB analyses.

Analytical results

U-Pb analytical results are given in Table 1. Zircons from Sample A (Hazelton Group andesitic tuff) are pale yellow, euhedral grains with simple to slightly multifaceted, prismatic forms and excellent clarity. Length:breadth (l:b) is 2:1 to 5:1. The sample yielded only 0.04 mg zircon that was divided into 3 fractions for U-Pb analysis. Two analyses (B and C) overlap concordia, and the more precise of these (C) has a ^{206}Pb-^{238}U age of 184.8 ± 0.4 Ma and a ^{207}Pb-^{206}Pb age of 184.5 ± 7.5 Ma (Fig. 5). The sample cannot be younger than the minimum possible ^{206}Pb-^{238}U age of analysis C, and therefore the crystallization age of the sample is bracketed at 185+7/-1 Ma. Fraction A yields a ^{207}Pb-^{206}Pb age of 207 ± 11 Ma that indicates an Early Jurassic component of inherited Pb.

Sample B (Hazelton Group rhyolite) yielded colourless to pink, euhedral to subhedral, strongly fractured zircons (l:b = 1:1 to 3:1) with colourless to cloudy white inclusions. Unabraded fractions are discordant but a strongly abraded fraction (E) is concordant with a ^{206}Pb-^{238}U age of 175 ± 1 Ma (Fig. 6). The zircon systematics suggest that the discordance is due to Pb loss. Two discordant analyses (B, C)

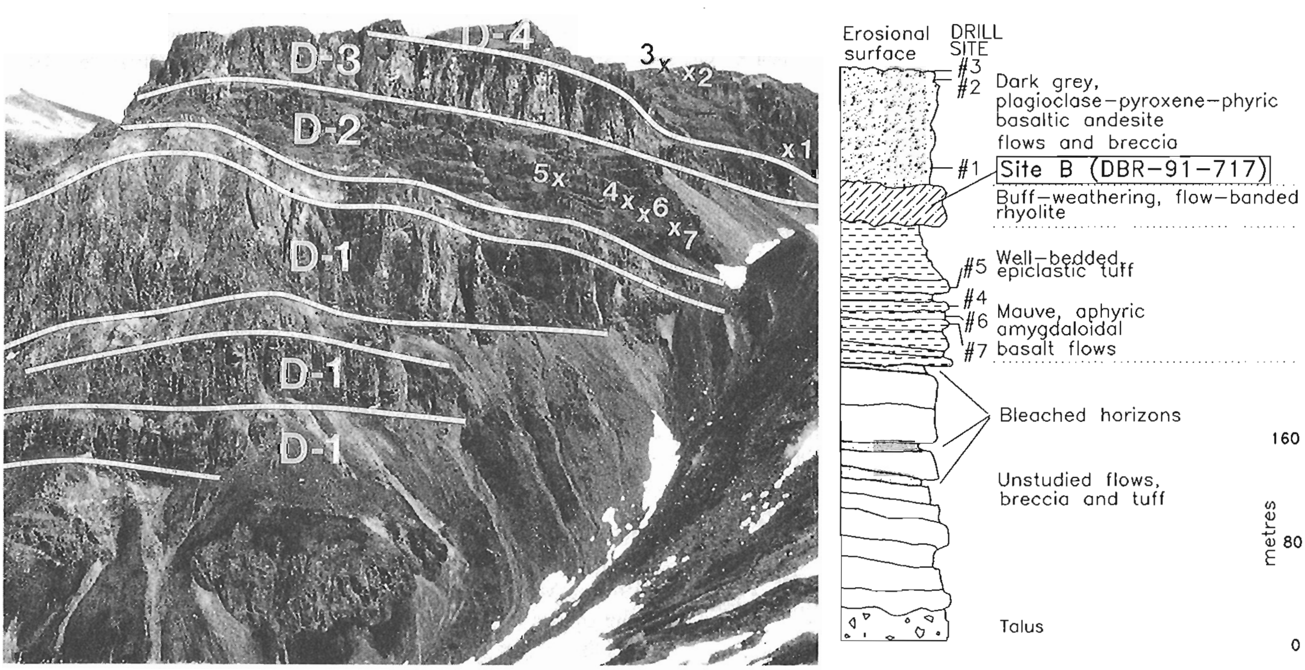

Figure 4. View of rhyolite flow from which sample B was collected, northeast of Kirk Creek, showing the stratigraphic position of this flow relative to paleomagnetic drill sites of Vandall at al. (1992)(small number with x's on photograph). The schematic column for the Kirk Creek cirque area illustrates the local stratigraphic position of the rhyolite (D-3 on photograph) relative to underlying basalt flows and epiclastic beds (D-2) and the overlying pyroxene-plagioclase phyric andesite flows (D-4). D-1 was not examined.

Table 1. U-Pb analytical data

Zircon Fraction	Wt. (mg)	U (ppm)	Pb[a] (ppm)	$^{206}Pb^{b}/^{204}Pb$	Pb[c] (pg)	$^{208}Pb^{a}$ (%)	$^{206}Pb^{d}/^{238}U$	$^{207}Pb^{d}/^{235}U$	Corr. Coeff.[e]	$^{207}Pb^{d}/^{206}Pb$	$^{207}Pb/^{206}Pb$ age (Ma)
Sample A (DBR-90-162) Hazelton Group andesitic tuff											
A. N1 +74-105 A	0.023	425	13	1298	14	8.8	0.03032±0.10%	0.2101±0.28%	0.56	0.05027±0.23%	207.3±10.8
B. N1 +62-74 A	0.013	527	15	784	16	10.7	0.02887±0.14%	0.1982±0.51%	0.60	0.04980±0.44%	185.8±20.5
C. N1 +149 A	0.010	1189	35	2198	10	11.1	0.02908±0.11%	0.1995±0.21%	0.65	0.04978±0.16%	184.5±7.5
Sample B[f] (DBR-90-717) Hazelton Group flow-banded rhyolite flow											
A. M2 +45-63	0.2	532	13	675	247	13.2	0.02389±0.15%	0.1632±0.21%	0.70	0.04954±0.15%	173±7
B. N2 +63-100	0.5	107	3	651	141	12.1	0.02617±0.15%	0.1806±0.21%	0.70	0.05005±0.15%	197±7
C. M2 +80-100	0.4	325	9	996	226	12.0	0.02663±0.15%	0.1828±0.18%	0.83	0.04978±0.10%	185±5
D. N2 +100-350	1.3	491	13	334	1120	12.0	0.02611±0.15%	0.1787±1.35%	0.43	0.04964±1.32%	178±15
E . N2+100-350 A	0.2	386	11	1266	108	11.7	0.02747±0.15%	0.1878±0.17%	0.89	0.04957±0.08%	175±4
Sample C (DBR-90-40) Cone Mountain granodiorite											
A. N1 +105 A	0.044	537	16	1576	28	12.6	0.02905±0.10%	0.1993±0.17%	0.70	0.04976±0.13%	183.9±5.9
B. N2 +105 A	0.089	429	13	1657	42	11.4	0.02906±0.09%	0.1993±0.15%	0.77	0.04973±0.10%	182.6±4.5
C. M2 +149 A	0.062	402	12	2460	18	13.5	0.02908±0.10%	0.1995±0.16%	0.70	0.04976±0.11%	183.9±5.3
D. N2+105-149 A	0.102	606	17	2597	42	11.8	0.02813±0.09%	0.1937±0.13%	0.82	0.04994±0.07%	192.1±3.5
Sample D[f] (DBR-91-723) Limpoke quartz monzodiorite											
A. N2 +30-45	1.1	759	20	5108	269	13.8	0.02568±0.15%	0.1767±0.15%	0.99	0.04991±0.02%	191±1
B. N2 +45-80	1.7	466	12	4297	100	12.7	0.02525±0.15%	0.1736±0.16%	0.96	0.04986±0.04%	188±2
C. N2 +80-100	1.8	422	11	3793	112	12.3	0.02615±0.15%	0.1799±0.22%	0.67	0.04991±0.17%	191±8
D. N2 +100-125	1.9	410	11	3910	111	12.2	0.02606±0.15%	0.1791±0.22%	0.68	0.04985±0.16%	188±8
E. N2 +125-350	0.8	304	8	3377	119	11.9	0.02569±0.15%	0.1767±0.15%	0.98	0.04989±0.03%	190±2
F. N2 +145-350	0.4	398	11	6104	47	12.8	0.02794±0.15%	0.1924±0.15%	1.00	0.04993±0.01%	192±1

N1 = non-magnetic at <1°, 1.7 amps on Frantz magnetic separator; N2 = non-magnetic at <2°, 1.7 amps; M2 = magnetic at >2°, 1.7 amps; A = abraded. Numbers such as +105 refer to size in microns. $^{207}Pb/^{206}Pb$ age errors are 2 std. errors in Ma. Blank compositions for GSC analyses are $^{206}Pb/^{204}Pb$=17.92; $^{207}Pb/^{204}Pb$=15.37; $^{208}Pb/^{204}Pb$=37.4. Blank compositions for UCSB analyses are $^{206}Pb/^{204}Pb$=18.6; $^{207}Pb/^{204}Pb$=15.5; $^{208}Pb/^{204}Pb$=38.0.

[a] Radiogenic Pb, corrected for blank and spike

[b] Corrected for fractionation and spike Pb

[c] Total common Pb in analysis in picograms

[d] Corrected for blank Pb and U and common Pb; errors are 1 std. error of mean in %

[e] Correlation coefficient of errors in $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$

[f] Analyses performed at UCSB by McClelland; all other analyses done at GSC by Bevier

with ^{207}Pb-^{206}Pb ages older than that observed for fraction E may contain (1) a slight component of inherited Pb or (2) surficial Pb along fractures with a composition different than that estimated for initial Pb. The latter interpretation is consistent with observations that a large percentage of the zircon population is strongly fractured and that the rhyolite contains alteration veins suggesting post-depositional hydrothermal alteration. Removal of the outer zircon surfaces by abrasion apparently removes the discordance. Allowing for slight Pb loss in analysis E, our best estimate of the crystallization age of this sample is 175 +4/-1 Ma.

Zircons recovered from sample C (Cone Mountain granodiorite) are pale brown, euhedral zircons (l:b = 1.5:1 to 4:1) with good to excellent clarity, inclusions of colorless tubes, bubbles, and black specks (ilmenite?), and multi-faceted crystal forms. Three analyses overlap concordia with indistinguishable ^{206}Pb-^{238}U ages that average 184.7 ± 0.6 Ma, and we consider this to be the best estimate of the crystallization age of the pluton (Fig. 7). Fraction D is highly discordant due to considerable Pb loss; however the ^{207}Pb-^{206}Pb age of 192 ± 4 Ma also indicates the presence of a minor inherited component in this fraction.

Sample D is a medium grained biotite hornblende quartz monzodiorite from the Limpoke pluton. Six discordant analyses define a chord with upper and lower intercepts of 194 ± 2 Ma and 19 Ma, respectively (MSWD = 1.2) (Fig. 8). The upper intercept is interpreted as the crystallization age and the zircon systematics suggest that the discordance is due to Pb loss, perhaps resulting from low temperature hydrothermal alteration. This interpretation is consistent with a K-Ar hornblende age of 182 ± 5 Ma obtained from the same sample site (Brown et al., unpublished data). The significance of the lower intercept age is uncertain.

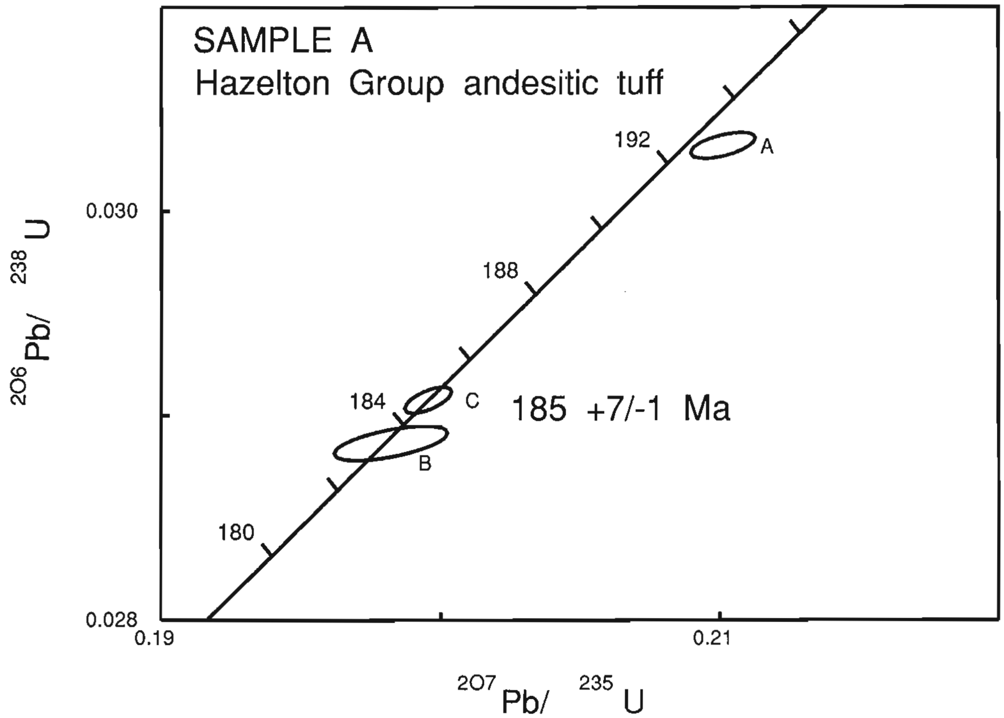

Figure 5.

U-Pb concordia plot for sample A, showing 2σ error ellipses.

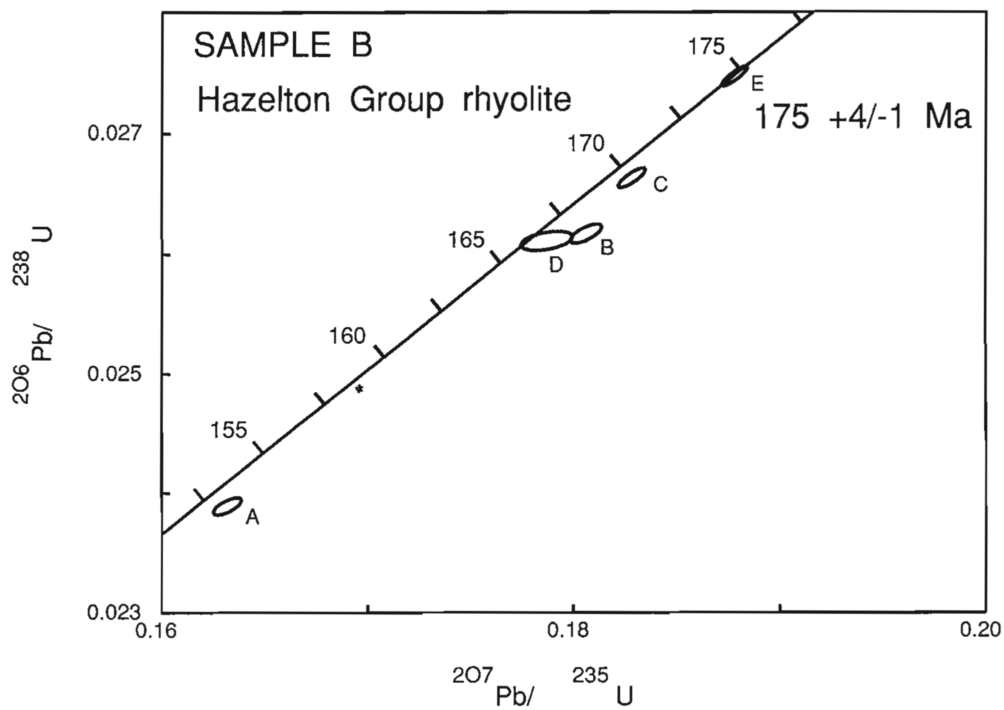

Figure 6.

U-Pb concordia plot for sample B, showing 2σ error ellipses.

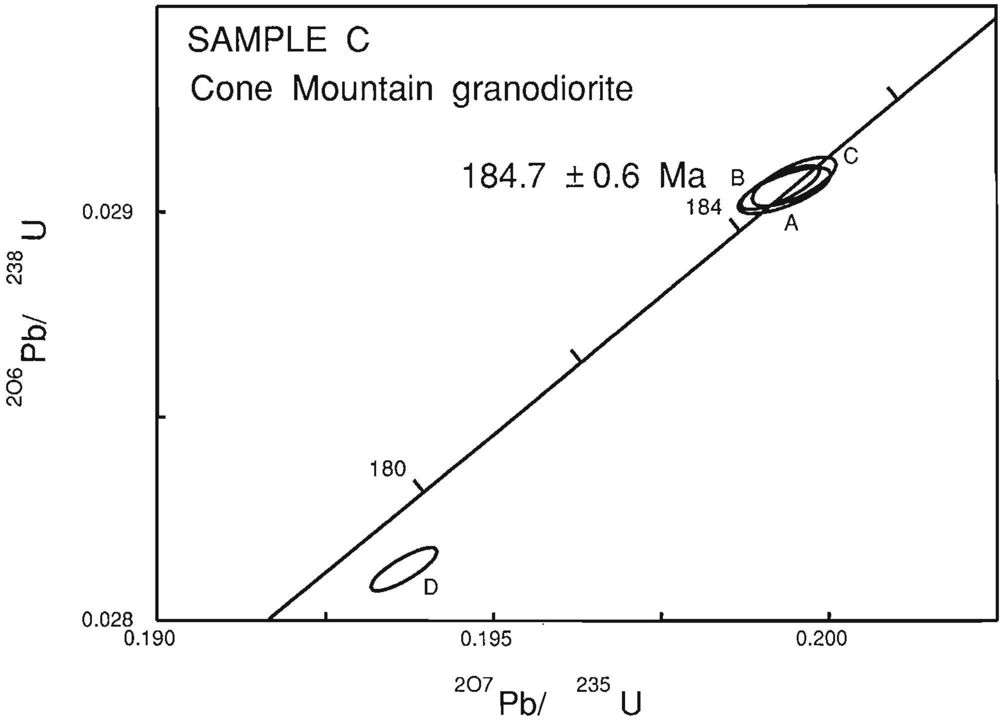

Figure 7.

U-Pb concordia plot for sample C, showing 2σ error ellipses. Similar plutons exposed 6 km to the east yield K-Ar dates of 177 ± 7 Ma, 182 ± 7 Ma and 163 ± 6 Ma, for biotite, hornblende and biotite respectively (Brown et al., 1990).

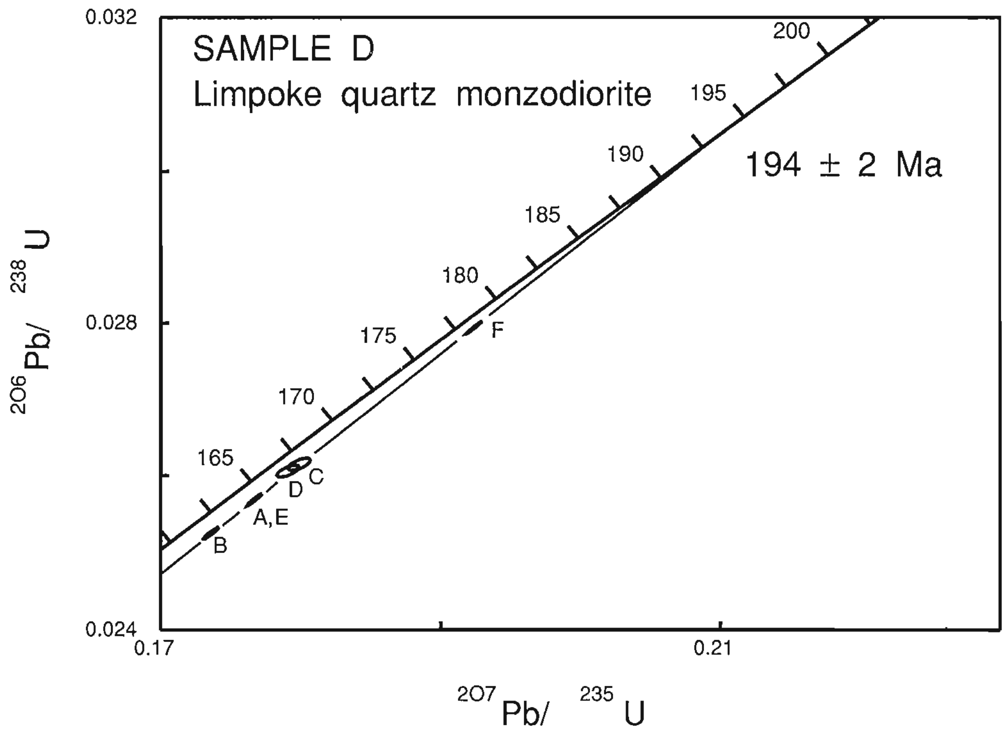

Figure 8.

U-Pb concordia plot for sample D, showing 2σ error ellipses. U-Pb discordia line was generated using the program of Ludwig (1992).

DISCUSSION

The Toarcian to Aalenian calc-alkaline volcanic and sedimentary rocks described and dated above span at least 10 million years of geological time and are similar to coeval volcanic rocks throughout northwestern Stikinia. These rocks include the Snippaker Mountain facies of the Salmon River Formation in the Iskut River area, about 100 km to the south-southeast (Fig. 1; Anderson and Thorkelson, 1990), strata east and northeast of Hankin Peak (70 km to the southeast; Evenchick, 1991) and the Mount Brock volcanic succession, 140 km to the east-northeast (Read, 1984; Read and Psutka, 1990; Thorkelson, 1992). The Limpoke pluton is coeval with the Texas Creek pluton near Stewart (Fig. 1), and the Cold Fish volcanic rocks of the Hazelton Group that lie 110 km to the east-northeast in Spatsizi River map area. Collectively, the Hazelton Group strata and nearby plutons, including the Cone Mountain and Saffron plutons, represent remnants of Toarcian to Aalenian arc magmatism exposed around the north and northwest margins of the Bowser Basin. The absence of Sinemurian to Pleinsbachian Hazelton Group strata in the study area suggests that the area was part of the Stikine Arch, a topographic high at that time.

The U-Pb ages presented above constrain the age of an episode of contractional deformation that affected the Stuhini Group to pre-Toarcian and post-Norian. Younger, southwest-directed contractional deformation affected the eastern margin of the Cone Mountain pluton and adjacent Paleozoic rocks of Stikine assemblage. This event may correlate with a contractional event along the King Salmon Fault (Thorstad and Gabrielse, 1986).

Comparison of U-Pb zircon and K-Ar whole rock ages for the Hazelton Group and the Limpoke pluton demonstrate the difficulty in interpreting K-Ar whole rock ages from this area. The K-Ar ages are invariably younger than the U-Pb ages, and therefore they have probably been partially reset by younger thermal or hydrothermal events.

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U-Pb geochronology of Cretaceous and Tertiary plutonic rocks of the Tagish Lake area, northeastern Coast Mountains, British Columbia

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Currie, L., 1992: U-Pb geochronology of Cretaceous and Tertiary plutonic rocks of the Tagish Lake area, northeastern Coast Mountains, British Columbia; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 163-170.

Abstract

U-Pb age determinations for Cretaceous and Tertiary plutons that intrude metamorphic rocks of Nisling terrane and Stikinia along the eastern margin of the Coast Plutonic Complex in northwestern British Columbia provide age constraints for the structural history of these terranes. The Stikinia-Nisling terrane boundary was broadly folded and offset by steep north-trending faults after 185 Ma and before the intrusion of a 127.2 ± 0.6 Ma granite. Deformation fabrics in a 101.6 ± 0.6 Ma granite are interpreted as the result of mid-Cretaceous or younger motion on the Llewellyn fault, which juxtaposes the metamorphic rocks of Stikinia and Nisling terrane against unmetamorphosed rocks of Stikinia to the east. To the northwest a hornblende granodiorite, thought to be part of the same plutonic complex, is 103 ± 0.7 Ma. A 55.9 ± 0.2 Ma age for an undeformed diorite tentatively constrains the age of the Mt. Switzer volcanic suite to Eocene or older. Coeval with the diorite is a quartz monzonite (55.7 ± 0.2 Ma) that is exposed 7 km to the east.

Résumé

Des datations U-Pb de plutons crétacés et tertiaires, qui recoupent par intrusion des roches métamorphiques du terrane de Nisling et de la Stikinie, le long de la bordure orientale du Complexe plutonique côtier dans le nord-ouest de la Colombie-Britannique, permettent de délimiter l'évolution structurale de ces terranes. La limite entre la Stikinie et le terrane de Nisling a été largement plissée et déplacée par des failles abruptes de direction nord après 185 Ma et avant l'intrusion d'un granite de 127,2 ± 0,6 Ma. Les fabriques de déformation dans un granite de 101,6 ± 0,6 Ma sont interprétées comme le résultat d'un déplacement remontant au Crétacé moyen, ou à une époque plus récente, le long de la faille de Llewellyn, qui juxtapose les roches métamorphiques de la Stikinie et du terrane de Nisling contre les roches non métamorphisées de la Stikinie à l'est. Au nord-ouest, une granodiorite à hornblende, qu'on croyait faire partie du même complexe plutonique, a été datée à 103 ± 0,7 Ma. Un âge de 55,9 ± 0,2 Ma associé à une diorite non déformée permet de limiter celui de la suite volcanique de Mt. Switzer, pour la situer à l'Éocène ou à une époque antérieure. Une monzonite quartzique (55,7 ± 0,2 Ma) qui affleure à 7 km à l'est est contemporaine de la diorite.

GEOLOGICAL FRAMEWORK

Cretaceous and Tertiary intrusive rocks in the Tagish Lake area lie on the eastern margin of the Coast Plutonic Complex of northeastern British Columbia, where they intrude metamorphic rocks of Nisling terrane and Stikinia (Currie, 1991, 1992; Currie and Parrish, in press; Fig. 1). The Nisling terrane is a narrow terrane interpreted by Wheeler et al. (1991) as representing a continental margin setting. It is separated from continental margin rocks of ancestral North America by oceanic rocks of Slide Mountain and Cache Creek terranes, and by primitive arc rocks of Stikinia and Quesnellia south of latitude 61°N (Fig. 1). How continental margin rocks of Nisling terrane came to be located outboard of oceanic and arc-like terranes remains a mystery. However, U-Pb age determinations for plutons from the Tagish Lake area

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constrain the structural history of Nisling terrane, making it possible to begin to compare the complex deforma- tional history of this terrane with tectonic events that affected other parts of the Cordillera.

In the Tagish Lake area metamorphic rocks from both Stikinia and Nisling terrane are polydeformed. Deformation fabrics that formed prior to the juxtaposition of the two terranes were transposed during the amalgamation of the two terranes along a sinistral transpressive ductile shear zone between 185 Ma and 170 Ma. This occurred during or before the accretion of Stikinia to the ancestral margin of North America (Currie and Parrish, in press). The resulting terrane boundary, referred to as the Wann River shear zone, and the associated penetrative ductile fabrics have been folded into broad open folds that are reflected in the map pattern (Fig. 1). Later faulting offset the terrane boundary, and was followed by intrusion of Cretaceous to Tertiary plutons. By some time in the Early Tertiary the metamorphic rocks were exposed and unconformably overlain by Mt. Switzer and Sloko Group volcanic rocks (Christie, 1957; Mihalynuk et al., 1989, 1990; Currie, 1990, 1991, 1992; Fig. 1).

The metamorphic rocks of Stikinia and Nisling terrane are presently juxtaposed against unmetamorphosed volcanic and sedimentary rocks of Stikinia along the Llewellyn Fault (Fig. 1), for which the magnitude, age and sense of movement are poorly constrained. Some west-side-down motion occurred on the fault after the deposition of Early Tertiary Sloko Group volcanic rocks; they are preserved at elevations of 1200 m west of the fault, but are not preserved east of the fault where peaks up to 1981 m are underlain by Triassic volcanic rocks (Fig. 1). However, the magnitude of offset may be less than the difference between these elevations (780 m), since the volcanics may have been deposited in an area of uneven topography (see Souther, 1971). In contrast, based on

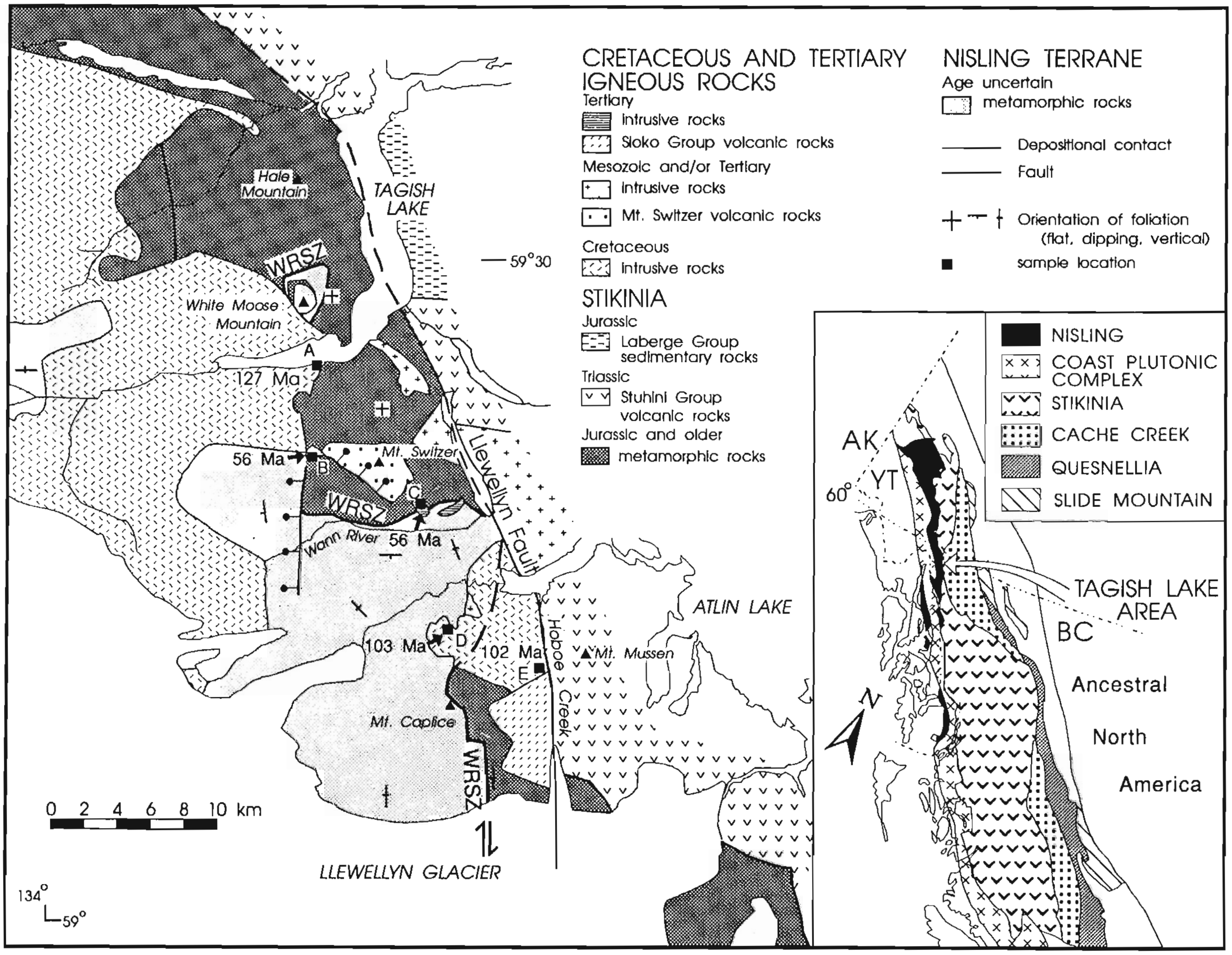

Figure 1. Geological map of the Tagish Lake area showing locations of samples discussed in the text: A, LC-89-35c; B, LC-89-75; C, LC-89-175c; D, LC-90-359; and E, LC-90-261c). Geology is based on Mihalynuk et al., 1989, 1990; Currie, 1990, 1991, 1992). The Wann River shear zone (WRSZ) forms the boundary between Nisling terrane and Stikinia. Inset shows the location of Nisling terrane and adjacent terranes (modified from Wheeler et al. 1991). AK - Alaska, BC - British Columbia, YT - Yukon Territory.

Table 1. U-Pb zircon analytical data for Cretaceous and Tertiary plutonic rocks in the Tagish Lake area

Fraction size[1]	Wt. µg	U ppm	Pb[2] ppm	$^{206}Pb^3/^{204}Pb$	Pbc[4] pg	^{208}Pb %	$^{206}Pb^5/^{238}U$	$^{207}Pb^5/^{235}U$	Corr. Coeff.	$^{207}Pb^6/^{206}Pb$	$^{207}Pb^7/^{206}Pb$	$^{207}Pb/^{206}Pb$ age (Ma)[6]	$^{207}Pb/^{206}Pb^2$ age (Ma)[7]
(A) LC-89-31 Undeformed, pink, medium-grained muscovite-biotite (± garnet) granite; locally contains orthoclase phenocrys up to 2 cm in diameter. UTM 8v, 538130E 6586250N.													
A +149	137	2216	42.3	4558	84	6.1	0.01988±0.21%	0.1333±0.22%	0.98	0.04864±0.04%	0.04860	130.4±2.1	128.9
B +149	89	2313	43.8	2118	124	3.0	0.01993±0.12%	0.1338±0.15%	0.88	0.04868±0.07%	0.04864	132.7±3.4	131.0
C +149	151	2389	42.8	7948	59	3.2	0.02059±0.16%	0.1432±0.17%	0.98	0.05044±0.04%	0.05040	215.2±1.7	213.6
E +149	22	991	20.6	859	38	6.3	0.02163±0.10%	0.1466±0.23%	0.67	0.04916±0.18%	0.04914	155.8±8.5	154.5
(B) LC-89-175c Undeformed fine- to medium-grained pink quartz monzonite from southeast of Mt.Switzer. UTM 8v, 534260E 6575550N													
A +149	28	302	2.66	664	7	11.2	0.008675±0.14%	0.05636±0.50%	0.58	0.04711±0.43%	0.04707	55.0±20	53.1
B +149	35	374	3.59	374	20	11.7	0.008686±0.15%	0.05715±0.69%	0.64	0.04772±0.61%	0.04767	85.4±27	83.3
C +149	76	324	3.70	508	33	11.4	0.010376±0.15%	0.06827±0.40%	0.59	0.04772±0.34%	0.04765	85.4±16	82.0
D +149	161	345	3.10	725	44	11.8	0.008792±0.11%	0.05740±0.27%	0.67	0.04734±0.21%	0.04730	66.6±10	64.6
E +149	165	332	3.08	1214	26	12.1	0.009076±0.17%	0.05945±0.27%	0.62	0.04751±0.21%	0.04734	74.7±10	70.9
F +149	110	337	2.99	523	40	11.6	0.008687±0.14%	0.05688±0.38%	0.63	0.04749±0.31%	0.04745	73.8±15	71.8
(C) LC-89-75c Undeformed medium-grained hornblende diorite. UTM 8v, 537250E 657890N.													
A +149	406	311	2.91	511	142	16.3	0.008707±0.16%	0.05659±0.36%	0.70	0.04714±0.27%	0.04712	56.2±13	55.5
B +149	352	291	2.72	656	89	16.3	0.008697±0.13%	0.05665±0.28%	0.71	0.04724±0.21%	0.04723	61.5±10	60.8
(D) LC-90-359 Undeformed medium-grained biotite-hornblende granodiorite from Mt. Caplice. UTM 8v, 544950E 6564725N													
A +149	52	310	5.40	1006	16	16.3	0.016150±0.14%	0.10741±0.32%	0.57	0.04823±0.26%	0.04820	110.7±12	108.9
B +149	148	301	5.23	1584	29	16.1	0.016187±0.10%	0.10767±0.19%	0.68	0.04824±0.14%	0.04820	111.1±6.8	109.2
C +149	162	279	4.92	1186	40	16.6	0.016324±0.10%	0.10867±0.19%	0.72	0.04828±0.13%	0.04824	113.1±6.3	111.3
(E) LC-90-261C Deformed medium- to coarse-grained granite from adjacent to the Llewellyn fault. UTM 8v, 551075E 6561325N													
A 105-149	161	562	9.76	2340	39	17.2	0.015923±0.10%	0.10596±0.14%	0.82	0.04826±0.08%	0.04822	112.2±3.9	110.3
B 105-149	123	427	7.41	1242	43	16.9	0.015964±0.10%	0.10620±0.20%	0.67	0.04825±0.15%	0.04821	111.6±7.2	109.6
C 105-149	182	353	6.41	948	73	16.6	0.016761±0.11%	0.11614±0.21%	0.69	0.05025±0.16%	0.05021	206.7±7.3	204.0

Notes: [1]sizes (i.e. +105) refer to length aspect of crystals in microns; [2]radiogenic Pb; [3]measured ratio, corrected for spike and fractionation; [4]total common Pb in analysis corrected for spike and fractionation; [5]corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; [6]corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jager (1977); for analytical details see Parrish et al. (1987); [7]corrected for initial Th exclusion, assuming whole rock Th/U=3/1 (see Schärer, 1984).

the juxtaposition of amphibolite facies metamorphic rocks west of the fault against unmetamorphosed rocks on the east, the net offset on the Llewellyn Fault has a component of west-side-up movement. The contrasting senses of offset suggest that movement on the Llewellyn Fault has been variable and long-lived. It is likely that the Llewellyn fault was also the locus of some strike-slip motion, but there are presently no reliable data to constrain either the magnitude or sense of this offset.

Five U-Pb age determinations for plutons from the Tagish Lake area are presented. Three undeformed plutons (labeled A to C in Fig. 1) were collected to obtain a youngest possible age for deformation that affected the metamorphosed rocks, and samples from undeformed and deformed parts of a fourth plutonic complex were collected to determine the oldest age limit for at least some of the deformation associated with movement on the Llewellyn Fault (labeled D and E, respectively in Fig. 1).

ANALYTICAL METHODS

U-Pb analyses of zircon fractions were performed at the Geological Survey of Canada in Ottawa using analytical procedures summarized by Parrish et al. (1987). Blanks for U are 1 picogram or less, and blanks for Pb range from 7-14 pico-grams. The clearest crystals were analyzed and all zircon fractions were strongly abraded prior to dissolution (Krogh, 1982). Analytical results are presented in Table 1.

Some of the analyses give discordant results. Three possible causes of discordance are: Pb-loss; ^{206}Pb deficiency; and inheritance of a component of older zircon. It is unlikely that Pb-loss is significant in these samples because they were all strongly abraded, and the U concentration of all samples except LC-89-35c is low to moderate (Table 1). The ^{206}Pb deficiency is caused by the exclusion of ^{230}Th from zircon during crystallization (Mattinson, 1973). Since ^{230}Th is an intermediate daughter product of the ^{238}U-^{206}Pb decay chain (half life = 7.52 x 10^4 Ma), the exclusion of ^{230}Th will result in a deficit of ^{206}Pb, and the ^{206}Pb-^{238}U age will be slightly less than the ^{207}Pb-^{235}U age. ^{206}Pb deficiency has been corrected in all analyses presented here using the method outlined in Schärer (1984); this method has been successfully applied to zircon analyses (Coleman and Parrish, 1991). Inheritance of an older component of zircon has affected some of the analyses, and is discussed below where appropriate.

RESULTS AND INTERPRETATIONS

Granite intruding fabrics in metamorphic rocks of Stikinia

The oldest dated pluton that is apparently unaffected by deformation that folded the Stikinia-Nisling terrane boundary (Wann River shear zone) is a pink granite with chilled margins (labelled A in Fig. 1); it also intrudes the metamorphic rocks of Nisling terrane and Stikinia (Currie, 1992), steep north-trending faults, and the shear fabric associated with the Wann River shear zone. Although this granite is shown in Figure 1 as a continuous body, it may contain phases of other lithologies and/or different ages (eg. see Mihalynuk et al., 1990).

U-Pb isotopic data for two zircon fractions (A and B) from the granite are concordant after correction for ^{206}Pb deficiency (Table 1, Fig. 2; Schärer, 1984), and overlap concordia between 126.4 and 127.6 Ma. It is therefore inferred that 127 ± 0.6 Ma is the crystallization age of the granite. Two additional fractions are discordant, primarily due to a component of inherited zircon. Some of the inherited zircon likely comes from the Early Jurassic Hale Mountain granodiorite (Currie, 1991) which is intruded by the granite (included in Stikinia in Fig. 1); the error ellipse for fraction E plots near a mixing line with a lower intercept of 127 Ma (the age of the granite) and an upper intercept of 185 Ma (the age of the Hale Mountain granodiorite). However, some of the inherited zircon must be older than Jurassic since both fractions C and E plot to the right of this mixing line.

The age of this sample is younger than a K-Ar age of 131 ± 3 Ma for hornblende from a sample inferred to belong to the same pluton, but collected to the north (Bultman, 1979; recalculated as 133 ± 3 Ma, Mihalynuk et al., 1989). A possible explanation for the anomalously old K-Ar age is the presence of excess Ar in the hornblende. The Early Cretaceous age for the granite provides a youngest possible age for the steep north-trending faults and the deformation that folded the Stikinia-Nisling terrane boundary and control the map pattern in the Tagish Lake area.

Hornblende diorite on the north side of Mt. Switzer

The hornblende diorite body on the north side of Mt. Switzer (labelled B in Fig. 1) intrudes previously deformed Jurassic plutonic rocks of Stikinia and the 127 ± 0.6 Ma granite (Fig. 1). Although the contact relationship between the diorite and the volcanic rocks was not observed, it is inferred that the diorite also intruded the Mt. Switzer volcanic suite because adjacent to the diorite body, conglomerates in the Mt. Switzer volcanic suite contain clasts of amphibolite, biotite schist, quartzite, granite and volcanic rocks, but lack diorite clasts. All of the lithologies represented in the conglomerate are exposed near the conglomerate; the granite clasts are likely derived from the 127 ± 0.6 Ma granite that outcrops to the west, the source of the metamorphic rocks is thought to be the Nisling Terrane which also lies to the west, and the volcanic rocks are assumed to be derived from the Mt. Switzer volcanic suite. The lack of diorite clasts suggests that the diorite was not present at the time the conglomerate was deposited, and is therefore younger than the Mt. Switzer volcanic suite.

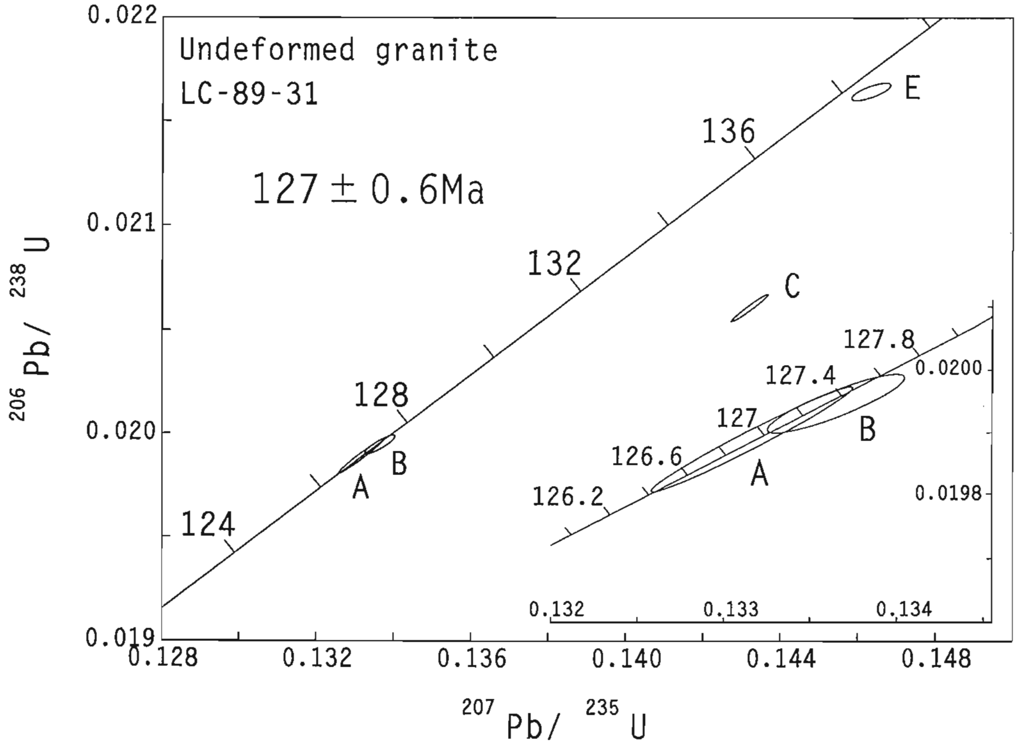

Figure 2.

U-Pb concordia diagram for undeformed granite (Sample LC-89-35c) from the south end of Tagish Lake. Data have been corrected for ^{206}Pb deficiency (Table 1).

U-Pb analyses for two fractions of large, clear, euhedral zircon with minor inclusions from the diorite body are concordant, overlapping concordia at 55.9 ± 0.2 Ma (Table 1; Fig. 3). The correction for ^{206}Pb deficiency (Table 1, Fig. 2; Schärer, 1984) increases the ^{206}Pb-^{238}U age of these analyses by about 0.1 Ma. The interpreted age of crystallization for the diorite is 55.9 ±0.2 Ma.

The Eocene age for the diorite provides a probable minimum age for the Mt. Switzer volcanic rocks.

Pink biotite quartz monzonite

A small pink biotite quartz monzonite body (labelled C in Fig. 1) intrudes the Wann River shear zone on the north side of the upper Wann River, where Permian gneiss of Stikinia (Currie, 1992) and metasedimentary rocks of Nisling terrane (Fig. 1) are juxtaposed. This shear zone is thought to have been active between 185 and 170 Ma (Currie and Parrish, in press).

Zircons from the quartz monzonite are clear, euhedral and contain inclusions. U-Pb isotopic data for three zircon fractions (A, B, and F) are nearly concordant to concordant and have ^{206}Pb-^{238}U ages of 55.7 ±0.3 Ma, before correction for ^{206}Pb deficiency and from 55.7 to 55.8 Ma after the correction is made (Table 1; Schärer, 1984; Fig. 4). Fraction (A) also has a ^{207}Pb-^{235}U age of 55.7 ±0.2 Ma and its error ellipse overlaps concordia at 55.7 ±0.2 Ma. The consistency of the results is evidence for a crystallization age of 55.7± 0.2 Ma. Three additional fractions plot to the right of concordia, primarily due to a component of older inherited zircon. Error ellipses for these fractions overlap mixing lines with lower intercepts of 55.7 Ma (the age of the quartz monzonite) and upper intercepts of 129 Ma (the age of a pluton that crops out west of the quartz monzonite) and 185 Ma (the age of the Hale Mountain granodiorite which outcrops north of the quartz monzonite and is included in Stikinia in Fig. 1). The inherited zircon component is therefore probably Jurassic to Cretaceous in age.

The age of the quartz monzonite is within error of that of the 55.9 ±0.2 Ma diorite exposed 5 km to the west, suggesting that these plutons may be genetically related.

Undeformed granodiorite from north of Mt. Caplice

An undeformed biotite-hornblende granodiorite (labelled D in Fig. 1) from north of Mt. Caplice and 6 km west of the Llewellyn Fault intrudes metamorphic rocks of the Nisling Terrane, Stikinia, and the tectonic boundary between them (Wann River shear zone; Fig. 1).

Pale yellow, euhedral zircons with inclusions have concordant U-Pb analyses that overlap concordia between 103.1 Ma and 103.7 Ma, after correction for ^{206}Pb deficiency (Table 1; Fig. 5). The error ellipse for a third fraction (C) is discordant, likely due to a component of inherited zircon, of probable Cretaceous age. If fractions A and B also contain a component of inherited zircon, the crystallization age of the granodiorite might be slightly younger than 103.1 Ma, and therefore the age of the granodiorite is interpreted as 103 ± 0.7 Ma. This age is supported by K-Ar cooling ages for hornblende and biotite (97.5 ± 3.4 Ma and 96.0 ± 3.4 Ma, respectively) for a sample inferred to belong to the same pluton

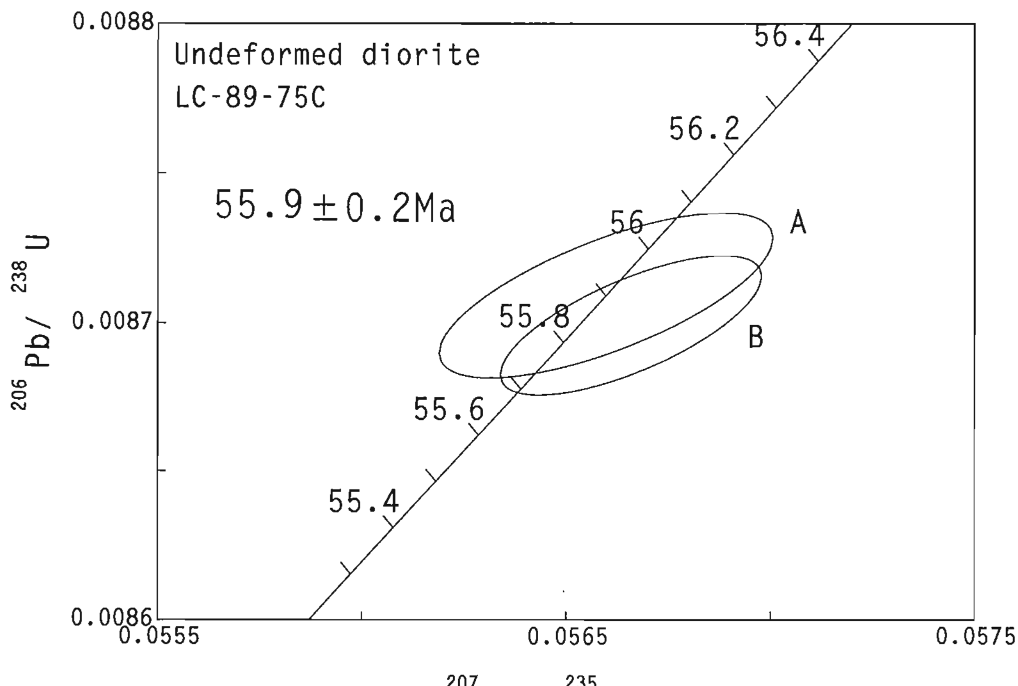

Figure 3.

U-Pb concordia diagram for diorite (Sample LC-89-75c) from southwest of Mt. Switzer. Data have been corrected for ^{206}Pb deficiency (Table 1).

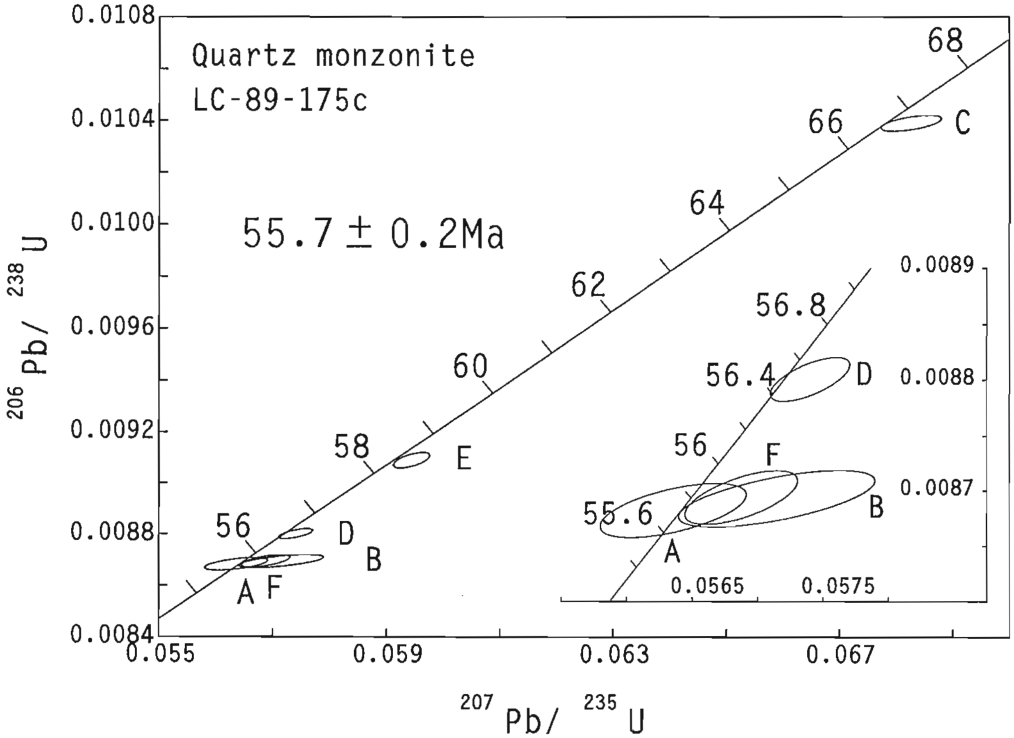

Figure 4.

U-Pb concordia diagram for quartz monzonite from north of the upper Wann River (Sample LC-89-175c). Data have been corrected for ^{206}Pb deficiency (Table 1).

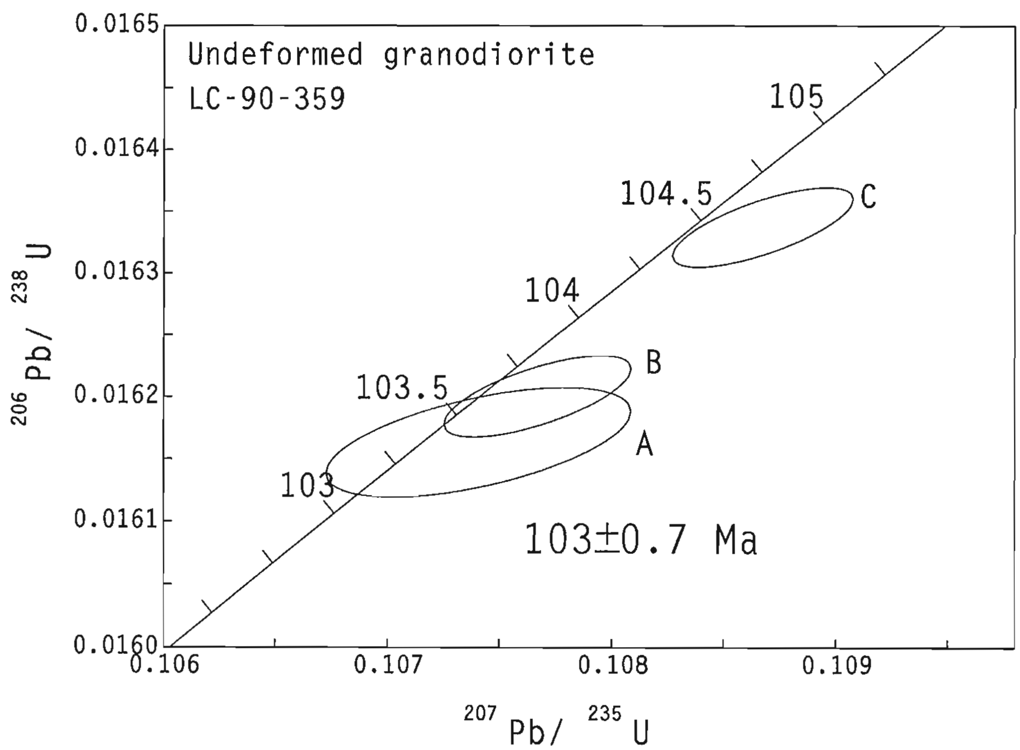

Figure 5.

U-Pb concordia diagram for undeformed granodiorite from north of Mt. Caplice (Sample LC-90-359). Data have been corrected for ^{206}Pb deficiency (Table 1).

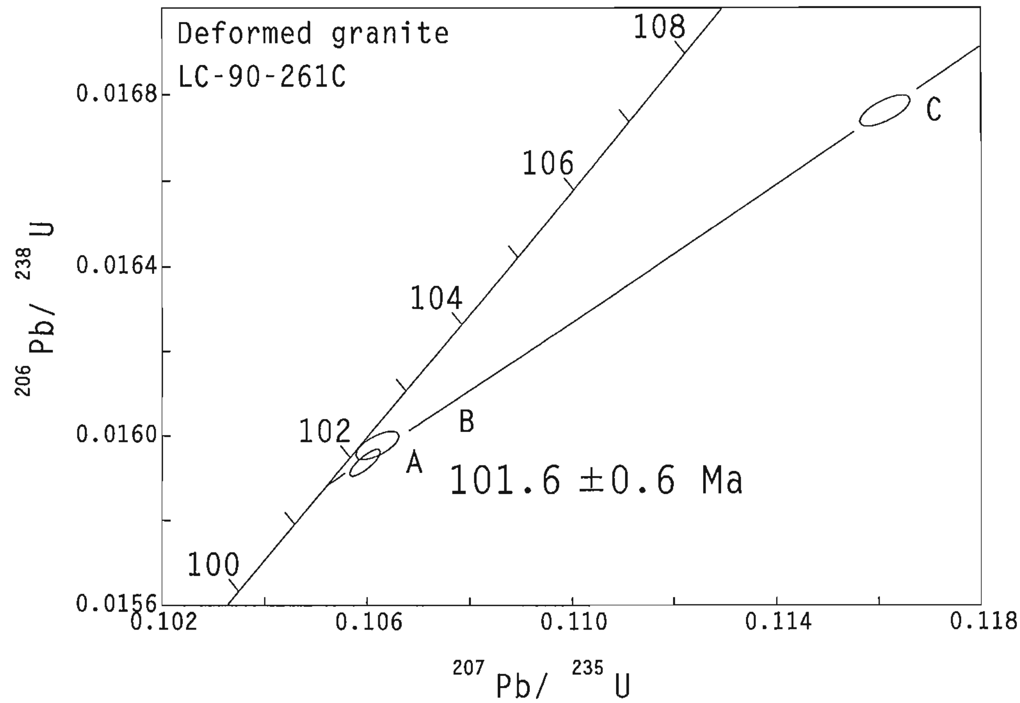

Figure 6.

U-Pb concordia diagram for deformed granite from adjacent to the Llewellyn Fault (Sample LC-90-261c). Data have been corrected for ^{206}Pb deficiency (Table 1).

that was collected to the north (UTM 8v 546333E 6564809N; unpublished data collected at the University of British Columbia ca. 1978, R.L. Armstrong pers. comm. 1991).

Deformed granite from the west side of the Llewellyn Fault

A moderately altered granite adjacent to and west of the Llewellyn Fault (labelled E in Fig. 1) is poorly indurated with brittle fabrics that dip steeply toward the west. Unlike the undeformed plutons, in which quartz grains have straight extinction and lack subgrains, quartz grains from this granite sample have well developed serrated subgrain boundaries. It is inferred that these deformation fabrics post-date crystallization of the granite and are due to movement on Llewellyn Fault.

Two of three analyses for zircon fractions for this sample are nearly concordant (A and B; Fig. 6), and after correction for ^{206}Pb deficiency one fraction (B) touches concordia at 102.2 Ma. The third fraction is highly discordant, primarily due to a component of inheritance, some of which may be older than 1.1 Ga. The data for these three zircon fractions yield a regression line with a lower intercept of 101.6 ±0.3 Ma (MSWD = 1.63), younger than 102.2 Ma, the age at which fraction B touches concordia. It is not possible to determine whether the younger age for the lower intercept is due to Pb-loss or whether fraction B contains a component of inherited zircon and the contact between its error ellipse and concordia is fortuitous. Consequently, the interpreted age for the granite includes both ages, and is 101.6 ±0.6 Ma.

The age of the granite (101.6 ±0.6 Ma) is close to the age of the granodiorite to the west (103 ±0.7 Ma), suggesting that these rocks may be phases of the same pluton. The Early Cretaceous age for this deformed granite indicates that some of the movement on Llewellyn Fault occurred after 101.6 ± 0.6 Ma.

CONCLUSIONS

The main conclusions, based on the U-Pb age determinations presented here are:

1. Broad folds which control the pattern of map units and fold the Stikinia-Nisling terrane boundary in the Tagish Lake area formed after 185 Ma and before 128 Ma.
2. At least some of the deformation associated with movement on the Llewellyn fault occurred after 102 Ma.
3. The age of the Mt. Switzer volcanic suite is tentatively inferred as 56 Ma or older.

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Miscellaneous ^{40}Ar-^{39}Ar ages and analytical procedures

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Roddick, J.C., Quigg, F.B. and Hunt, P.A., 1992: Miscellaneous ^{40}Ar-^{39}Ar ages and analytical procedures; in Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 171-177.

Abstract

^{40}Ar-^{39}Ar analytical procedures and step heating analyses are presented for a hornblende from the Boothia Uplift of Somerset Island, District of Franklin, with an interpreted age of 1918 ± 12 Ma; a biotite from a diabase dyke in the Eastern Townships of Quebec, with an interpreted age of 195.5 ± 1.4 Ma; and a biotite from a tuff in the Cordillera of Yukon Territory, with an interpreted age of 94.6 ± 1.0 Ma.

Résumé

Le présent arcticle fait état de la procédure rattachée aux datations par la méthode ^{40}Ar-^{39}Ar et des résultats d'analyses par paliers de température selon cette méthode, effectuées sur divers minéraux, notamment sur une hornblende provenant du Soulèvement de Boothia (île Somerset dans le district de Franklin), dont l'âge interprété est de 1 918 ± 12 Ma; sur une biotite extraite d'un dyke de diabase dans les Cantons de l'Est, au Québec, dont l'âge interprété est de 195,5 ± 1,4 Ma; et sur une biotite contenue dans un tuf de la Cordillère, au Yukon, dont l'âge interprété est de 94,6 ± 1,0 Ma.

INTRODUCTION

The Geochronology Laboratory of the Geological Survey of Canada performs step-heating ^{40}Ar-^{39}Ar geochronology on samples in support of geological studies. Some analyses are minor components of larger studies or single analyses for specialized problems which do not merit separate publication. To make the data available in a timely fashion the results are published here with a brief description of the geological context and an interpretation of the age spectra. This is the first report of this type and includes results for three samples: a hornblende separate from Somerset Island, N.W.T. and two biotite separates from Quebec and Yukon Territory. Details of current analytical techniques are also given.

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ANALYTICAL TECHNIQUES

Mineral separates for dating were obtained by standard magnetic and heavy-liquid techniques followed by ultrasonic washing in water and hand-picking to obtain better than 99.5% purity. The samples were weighed and wrapped in aluminium foil packets for neutron irradiation. They were arranged along with other samples in aluminium cans 40 mm x 19 mm diameter along with optical grade CaF_2 and flux monitors. The samples were irradiated in three separate irradiations in position 5C of the enriched uranium research reactor at McMaster University, Hamilton Ontario. Table 1 records the fast neutron dose received, as the J factor, for each sample. This reactor has an approximate fast neutron fluence of 3×10^{16} neutrons/cm^2, which results in a J/hour (Dalrymple et al., 1981) of about 2.5×10^{-4}. Flux variation over the volume of the irradiation containers ranges from 2 to 3%, mainly along the axis of the cans, and is measured with 10 to 12 monitors placed throughout a can. The errors quoted on the integrated ages take into account the measurement of this flux variation, though the individual step ages do not. Flux

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monitors used were MMHb-1 hornblende (Samson and Alexander, 1987) with an assumed age of 518.9 Ma (Roddick, 1983) and FCT-3 biotite, a new ^{40}Ar-^{39}Ar flux monitor prepared by the USGS, Reston from the Fish Canyon Tuff. Its age is 27.68 ± 0.03 Ma (M.J. Kunk, pers. comm. 1988).

For two of the three irradiations the McMaster reactor was on full power for about 15 hours per day. To determine the appropriate amount of ^{37}Ar and ^{39}Ar generated the formula of Wijbrans and McDougall (1987) for pulsed irradiations was used. The sum of these incremental irradiations for ^{37}Ar production can be equated to a single irradiation of the same time as the sum of irradiation times but with an adjusted time of termination of irradiation. This adjusted time for ^{37}Ar decay is sufficiently similar to that for ^{39}Ar decay to make insignificant differences to corrections for ^{39}Ar decay.

Argon extractions were performed in a double-vacuum resistance-heated tantalum furnace, similar in design to that described by Staudacher et al. (1978). The evolved gas was purified in three stages of gettering. Corrections are made for extraction and purification blanks of ≈ 3 X 10^{-10} cc (STP) of ^{40}Ar (atmospheric composition) below 1200°C and 2 X 10^{-9} cc at 1550°C and have an uncertainty of ±50%. There is an additional blank (not corrected for) from the Al foil packet material of about 8 X 10^{-9} ^{40}Ar which is evolved by 660°C, the melting point of Al.

A modified MS 10 mass spectrometer with a 0.4 Tesla permanent magnet and solid state electronics was used to analyze the argon. Data collection involved computer controlled voltage peak switching with twelve scans of the Ar isotope ion beams recorded by a Keithley™ 642 electrometer with a 10^{11} ohm resistor, feeding a Solartron™ DVM with output to a HP 300 computer for data reduction. The peak switching interval of 5 seconds delay and 4 seconds integration resulted in a detection limit of 1 X 10^{-12} cm^3 STP of argon. Intensities at masses 28, 43 and 80 were also monitored.

Data Table 1 lists the Ar isotope quantities after corrections for minor products of neutron reactions on K, Ca, and Cl. Mass spectrometer discrimination and sensitivity were monitored with each sample by analyzing atmospheric argon from a pipette system. Close monitoring of the sensitivity and careful gas clean-up procedures allow the MS 10 to be used as a precise manometer (Baksi, 1973). Using monitors with known amounts of K, Ca and Cl the concentrations of these elements in the samples are determined. The precision of these concentrations is estimated to be about 1% (1σ). In addition to the usual age spectrum plots the associated variations of Ca/K and Cl/K for each step are also displayed for some of the samples and used in the interpretation of the age spectra. In the plots, the element ratios are normalized to the integrated value of the ratio for all heating steps. Additional details of procedures for ^{40}Ar-^{39}Ar analyses are given in Roddick (1990). Errors in quoted ages and plotted age spectra are at the 2σ level.

RESULTS

AA92-275, Hornblende
1918 ± 12 Ma interpreted ^{40}Ar-^{39}Ar
Wt % K= 0.323, Wt % Ca= 8.15, ppm Cl= 524
From a metabasite.

Between Macgregor Laird Lake and Bellot Strait, about 8 km east-northeast of Leask Point, Somerset Island, District of Franklin; NTS 58B; 72°00.5'N, 94°57'W; UTM zone 15 432781E 7990782N; sample 76-DV-279b. Collected by C.D.S. de Vries and R.D. Stevens.

The sample is from a 70 m thick metabasite sheet, metamorphosed at granulite grade. This sample and two others were previously dated by conventional K-Ar dating and gave ages of 1942 to 2126 Ma with relatively high uncertainties of about ± 130 Ma (samples GSC 78-114 to 116; Wanless et al., 1979). ^{40}Ar-^{39}Ar dating was undertaken to improve on the uncertainties and to confirm the age, as excess argon is common in high grade terranes (Harrison and McDougall, 1981). Recent U-Pb zircon, monazite and sphene dating of crystalline rocks from the Boothia Uplift indicates a granulite metamorphism and magmatic event at ca. 1.9 Ga (unpublished data). A K-Ar hornblende age from a granodiorite gneiss gives an age of 1831 ± 20 Ma (Hunt and Roddick, 1992; GSC 92-51) and biotite K-Ar ages from quartzofeldspathic rocks in the area range from 1635 to 1742 Ma (GSC 63-17, 18, 92, Wanless et al., 1965; GSC 78-112, 113, Wanless et al., 1979). The conventional K-Ar age of this sample represents the oldest dated hornblende found in the map area.

The age spectrum for this sample (Fig. 1) is irregular with ages as high as 5.7 Ga showing the presence of significant excess argon in the initial gas release. This excess argon is correlated with a phase higher in Ca/K than the bulk of the hornblende (see Ca/K plot, Fig. 1) and is evolved by about 30% ^{39}Ar release at which a minimum age of 1854 Ma for the spectrum is reached. There is an increase in age to a maximum of 1948 Ma and then a decrease to a relatively constant age of 1918 ± 12 Ma for three steps in the last 29% gas release. This age is taken as the best estimate of the age of the sample and represents cooling through about 500°C, the Ar closure temperature for hornblende. The age is about 100 Ma younger than the integrated age which also includes the excess argon present in the initial release.

Table 1. Argon analytical data

TEMP. (°C)	$^{36}Ar_{tr}$	$^{37}Ar_{Ca}$	$^{38}Ar_{Cl}$	$^{39}Ar_K$	^{40}Ar	%Atmos. ^{40}Ar	APPARENT AGE Ma	$\pm 2\sigma$[b]	^{39}Ar (%)
		($\times 10^{-9}$ cm^3 STP)[a]							
AA92-275 Hornblende (76-DV-279-b; 40.98 mg; J=0.006663 ± 0.5% 1σ)									
700	0.046	0.663	0.039	0.020	82.04	16.7	5722	156.8	0.3
800	0.032	0.290	0.020	0.015	62.00	15.1	5768	98.2	0.2
850	0.016	0.143	0.010	0.015	20.42	23.0	3784	307.8	0.2
900	0.007	0.132	0.007	0.013	11.16	17.5	3098	447.4	0.2
950	0.006	0.283	0.019	0.014	12.32	14.4	3213	403.5	0.2
1000	0.007	0.790	0.054	0.051	21.90	9.1	2323	84.1	0.8
1050	0.016	2.712	0.200	0.202	69.79	6.8	2069	14.0	3.3
1080	0.008	4.427	0.327	0.338	103.31	2.3	1975	8.6	5.5
1100	0.012	10.012	0.741	0.774	222.45	1.7	1911	4.8	12.5
1110	0.007	5.690	0.423	0.442	122.43	1.6	1868	4.6	7.2
1120	0.007	5.640	0.417	0.436	119.37	1.8	1854	6.3	7.0
1130	0.004	4.121	0.304	0.318	88.23	1.5	1872	13.5	5.1
1150	0.007	5.469	0.406	0.425	122.29	1.6	1912	7.2	6.9
1175	0.003	6.168	0.455	0.473	138.49	0.6	1944	6.8	7.7
1225	0.004	11.563	0.834	0.871	255.35	0.5	1948	4.1	14.1
1275	0.003	5.117	0.365	0.380	109.28	0.8	1921	5.4	6.2
1350	0.002	6.822	0.502	0.524	150.23	0.4	1923	5.2	8.5
1550	0.017	11.141	0.825	0.868	251.34	2.0	1915	4.0	14.0
Total[c]	0.20	81.18	5.95	6.18	1962	3.1	2013	16	
Conc.(/g)	5.0	1981.07	145.14	150.77	47887				
AA92-237 Biotite (GGA-87-14A3 ; 17.08 mg; J=0.001810 ± 0.5% 1σ)									
600	0.048	0.219	0.319	1.097	16.23	88.1	5.8	0.61	11.4
800	0.059	0.105	0.269	1.239	49.75	35.2	83.0	0.7	12.8
900	0.010	0.053	0.348	1.623	57.04	5.3	105.5	0.5	16.8
950	0.004	0.047	0.185	0.866	30.09	4.4	105.4	0.8	9.0
1000	0.006	0.109	0.188	0.739	30.98	6.2	124.0	1.2	7.7
1050	0.006	0.161	0.339	1.361	55.43	3.0	124.6	0.9	14.1
1100	0.007	0.125	0.289	1.387	47.91	4.0	105.1	0.4	14.4
1450	0.007	0.250	0.291	1.333	43.05	4.6	97.9	0.7	13.8
Total[c]	0.15	1.07	2.23	9.64	330.5	13.2	94.6	1.0	
Conc. (/g)	8.7	62.64	130.4	564.6	19348				
AA92-90 Biotite (SD-870102; 3.17 mg; J=0.01239 ± 0.25% 1σ)									
700	0.012	0.034	0.099	0.317	5.52	62.9	138.9	28.0	1.9
800	0.015	0.089	0.194	0.721	10.99	41.0	190.7	3.9	4.3
900	0.005	0.111	0.377	1.927	19.73	7.6	199.9	1.1	11.4
1000	0.002	0.002	0.513	3.122	30.49	1.6	203.1	3.6	18.5
1050	0.003	0.112	0.396	2.589	24.66	4.2	193.3	3.3	15.4
1100	0.001	0.003	0.401	2.684	25.14	1.2	195.8	0.8	15.9
1150	0.001	0.024	0.511	3.580	33.78	1.3	197.0	1.4	21.2
1200	0.000	0.071	0.144	1.740	16.19	0.9	195.2	3.1	10.3
1250	0.001	0.047	0.003	0.076	0.82	29.4	161.7	73.1	0.5
1300	0.001	0.019	0.004	0.043	0.60	32.8	195.0	54.1	0.3
1500	0.003	0.111	0.008	0.057	1.27	67.3	156.0	159.8	0.3
Total[c]	0.04	0.62	2.65	16.86	169.2	7.8	195.9	1.6	
Conc.(/g)	14.1	196.25	836	5318	53371				

[a]All gas quantities have been corrected for decay, isotopes derived from minor interfering neutron reactions, and blanks. tr denotes trapped Ar and Ca, Cl and K denote Ar derived from these elements. ^{40}Ar denotes trapped plus radiogenic Ar. Atmos. ^{40}Ar assumes a trapped Ar component of atmospheric composition.
[b]Errors from steps are analytical only and do not include the error in the irradiation parameter J.
[c]Includes the integrated age. The uncertainty in J is included in the error.

The hornblende age of 1918 ± 12 Ma agrees with U-Pb dating in the area and is interpreted as a cooling age related to a major early Proterozoic thermal event at about 1.9 Ga which had cooled to the biotite K-Ar closure temperature of about 300°C by 1650-1700 Ma.

AA92-90, Biotite
195.5 ± 1.4 Ma interpreted ^{40}Ar-^{39}Ar
Wt % K= 6.12, Wt % Ca= 0.43, ppm Cl= 1623
From a diabase.

Outcrop is on a roadcut on highway 263, 2.4 km ESE of Lambton, eastern townships, Quebec; NTS 21 E/14; 45°49.7'N, 71°3.9'W; UTM zone 19 339605E 5076830N; sample SD-870102. Collected by E. Schwarz.

The sample is from a small outcrop of a diabase dyke which is the only exposed expression of a prominent N-S magnetic anomaly identified on aeromagnetic maps C21066G, C21067G, C41066G, and C41067G (Geological Survey of Canada, 1985a, b, c, d). The magnetic expression of the anomaly is reflected as a vertical magnetic component 0.4 gamma above the regional background over a width of 170 m. Along the N-S strike the magnetic anomaly extends more than 5 km north (aeromagnetic map limit) of the diabase outcrop and more than 37 km to the south of it (aeromagnetic map limit). It appears to be the only prominent magnetic feature in the region with this orientation. Dating was carried out to determine if this dyke is related to other igneous intrusions in the region.

The dyke intrudes folded siltstone and sandstone strata of the Lower Devonian St Francis Group (St-Julien, 1965). In thin section the texture of the diabase varies from coarse (0.5-1.0 mm) equigranular to a fine groundmass with phenocrysts of slightly altered pyroxene (some twinned) and fresh plagioclase laths. The groundmass is an intergrowth of slightly altered plagioclase, altered pyroxene, opaques and minor brown to green biotite. Alteration is variable in the equigranular textured zones with higher concentrations of biotite in areas where plagioclase is extensively altered. It is believed that the development of biotite is associated with deuteric alteration. A pure separate, of uniform brown biotite flakes was obtained from the dyke sample, despite the presence of the green variant. The K content at 6.1% is just within the expected range in K for biotite (7.1 ± 0.5% S.D.; Mitchell et al., 1988) and reflects a good quality concentrate.

The step heating ^{40}Ar-^{39}Ar data for this sample is typical of a biotite in an undisturbed geological environment. The spectrum is essentially flat at an age of about 196 Ma with a slight hump to an age just over 200 Ma for 20% of the ^{39}Ar (Fig. 2). There is no apparent correlation of this hump with the Cl/K as this ratio decreases in a monotonic fashion over the total gas release. The Ca/K variation is not plotted as most Ca derived ^{37}Ar had decayed before analysis. The pattern of small irregularities on step ages is common in quickly cooled biotites (Tetley and McDougall, 1978) but does not detract from obtaining meaningful ages from the plateau segment of the spectrum or, if the hump is significant ($\geq$4%, Tetley and McDougall, 1978), from the integrated age. In this case, the

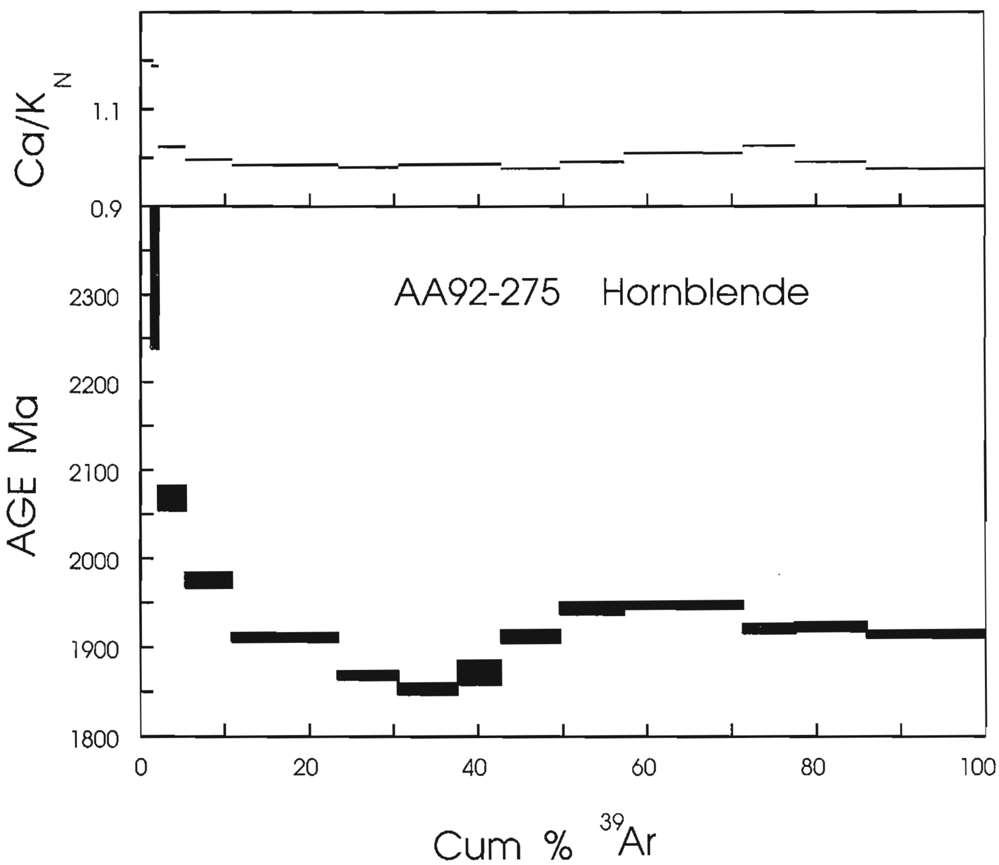

Figure 1.

^{40}Ar-^{39}Ar age spectrum for hornblende from dyke sample 76-DV-279-b. The variation of Ca/K for the steps (normalized to the integrated value for all steps) is also plotted.

biotite yields a plateau age of 195.5 ± 1.4 Ma for 63% of the gas release in excellent agreement with the integrated age of 195.9 ± 1.6 Ma.

The dyke, with an age of 196 Ma, does not appear to be related to any other igneous events known in the region. It is not related to the extensive alkaline igneous activity in southern Quebec. Doig and Barton (1968) distinguished four periods of alkaline magmatism associated with the St. Lawrence rift system. These occurred at 1000 to 820 Ma, 565 Ma, 450 Ma at 110-130 Ma. Subsequent work by Foland and workers (Foland et al., 1986, 1989) on the youngest event, the Monteregian Hills activity, has shown that nine of ten intrusions all have ages of 124 ± 1 Ma (see Foland et al., 1986, 1989). The tenth intrusion, at Oka, west of Montreal is about 10 Ma younger.

The orientation of the dyke is similar to the north-northeast-trending faults associated with the St Lawrence graben to the west and north. One of these faults, which passes through Mount Royal in Montreal, is known to have been active for at least 975 Ma (Philpotts, 1964). The dyke could be an easterly expression of these faults extending into the Appalachians.

On a more regional scale the age of this dyke is similar to ages of basalts of the Newark basin, New Jersey and nearby upper-Triassic-lower Jurassic basins, which have K-Ar ages of about 190 Ma (Seidemann et al, 1984). The formation of these basins is associated with the initial rifting and opening of the Atlantic Ocean. Recently, Dunning and Hodych (1990) have provided precise U-Pb ages which indicate sills along the eastern margin of North America associated with this rifting are 201 ± 1 Ma old. The eastern townships dyke may be a more remote expression of this major rifting event.

AA92-237, Biotite
94.6 ± 1.0 Ma interpreted ^{40}Ar-^{39}Ar
Wt % K= 4.5, Wt % Ca= 0.95, ppm Cl= 1732
From a quartz feldspar biotite crystal lithic tuff.

From 15.8 km northeast of the northeast end of Orchie Lake, Yukon; NTS 105 J/5; 62°16.24'N, 131°39.67'W; UTM zone 9, 361880E, 6906970N; sample GGA-87-14A3. Collected by S.P. Gordey.

The sample was collected to determine a K-Ar age of a tuff associated with the South Fork volcanics. Dating of other samples in the region suggests that the volcanics are about 95 Ma old (see GSC 92-32, Hunt and Roddick, 1992). Routine K analyses determined that the K content was anomalously low (4.43 wt.%) despite the fact that the mineral separate appeared to be of excellent quality. It is a golden brown biotite made up of subhedral laths with no apparent alteration or colour variation which might suggest chlorite is present. XRD analysis showed a possible serpentine alteration.

Step heating ^{40}Ar-^{39}Ar dating was undertaken on this sample to determine if a plateau age could be determined. The age spectrum is very irregular rising from an unrealistically low age of 6 Ma for the first 11% ^{39}Ar release to a peak age of 124 Ma for two steps at 60% release (Fig. 3). With additional heating the age decreases to 98 Ma in the final 14% gas release. There is some correlation of the highest

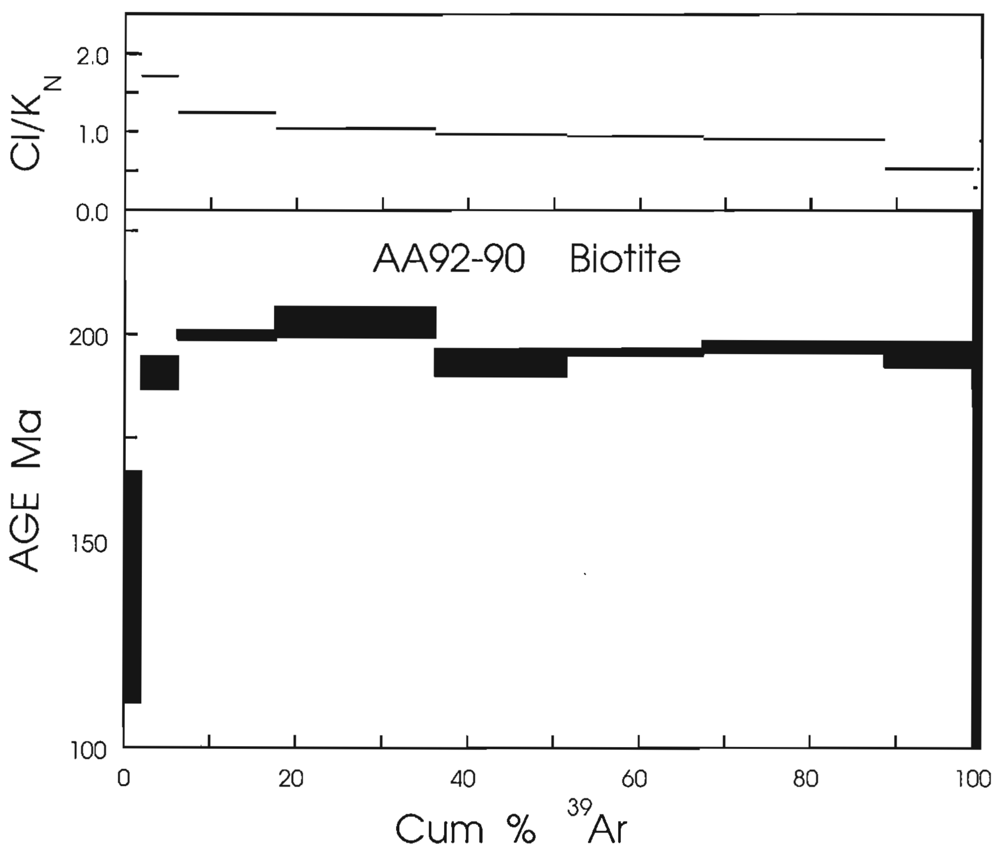

Figure 2.

^{40}Ar-^{39}Ar age spectrum for biotite from diabase dyke sample SD-870102. The variation of Cl/K for the steps (normalized to the integrated value for all steps) is also plotted.

ages with higher Cl/K but the correlation with Ca/K is irregular suggesting that the gas is released from sites with variable Ca/K. The Ca content is about 1%, a value typical of igneous biotite as represented by the chemical analyses of Deer et al. (1965). From their tabulation of 14 igneous biotites the mean Ca content is 0.59 wt.% with a standard deviation of ± 0.44 wt.%. In contrast the mean Ca content of 16 metamorphic biotites is lower at 0.16 ± 0.16 wt.%. The integrated age for the sample is 94.6 ± 1.0 Ma and is consistent with other ages of South Fork volcanics (see GSC 92-32, Hunt and Roddick, 1992). The young ages in the low temperature release could be interpreted to represent a recent re-heating event but there is no regional or local evidence for post-crystallization tectonism, thermal overprinting or metamorphism in the area. Consequently, the irregular spectra is believed to be caused by low-K alteration phases in the biotite.

It has been recognized for some time that incorrect conventional K-Ar ages are reflected by low K contents in biotite but there is no consistent pattern. Obradovich and Cobban (1975) showed that, for altered biotites with K contents below 5%, conventional K-Ar ages are usually too young but some may be too old. Clearly radiogenic Ar is lost but in some cases similar or greater amounts of K are lost to maintain or increase the K-Ar age.

Recently Hess and Lippolt (1986) have shown that ^{39}Ar, produced during neutron irradiation, and used to determine the K in a sample, may be lost from biotite during irradiation in the nuclear reactor. Using a suite of optically fresh and apparently unaltered biotites with a range of K contents, they showed ^{39}Ar loss is inversely correlated with K content, with 1% or more loss in biotite containing $\leq$5% K. They convincingly argued that the loss of ^{39}Ar cannot occur as direct recoil out of the crystal lattice, but rather is the result of a two stage process. In this process recoiled ^{39}Ar is enhanced in low-K alteration phases in the biotite and under the increased temperatures occurring during neutron irradiation (<190°C) this argon diffuses out of these alteration phases and is lost from the mineral grains. In a subsequent paper Hess et al. (1987) discussed age spectra produced by biotites with low K contents and showed that most biotites with K <8 wt.% have spectra with low ages in the initial release and which rise to plateau ages that may be greater than the known age of the rock. In some cases, however, the integrated ages are consistent with the accepted age. Recoil of ^{39}Ar out of biotite sites into low-K inclusions or alteration phases, which degas at low temperatures, can account for the observed spectra. In the case of the samples with correct integrated ages, it is probable that no recoiled ^{39}Ar is lost from the low-K phases during irradiation. Alternatively, but less likely, the amount of ^{39}Ar diffused out of these phases is proportional to the radiogenic Ar lost from the biotite since its formation, thus maintaining the original age of the sample.

In comparing the present South Forks biotite spectrum with these studies of biotite ^{40}Ar-^{39}Ar dating, it appears that the inclusions or alteration in the biotite may have retained the neutron induced ^{39}Ar because the integrated age, at 94.6 ± 1.0 Ma, is consistent with the age of associated volcanics. The age of the last step at 97.9 ± 1.3 Ma (error includes J uncertainty) is also in agreement with the expected age and may indicate that biotite is the only phase still retaining any Ar in the highest temperature release. Given the uncertainties associated with the interpretation of this single sample, the apparent age must be treated with caution.

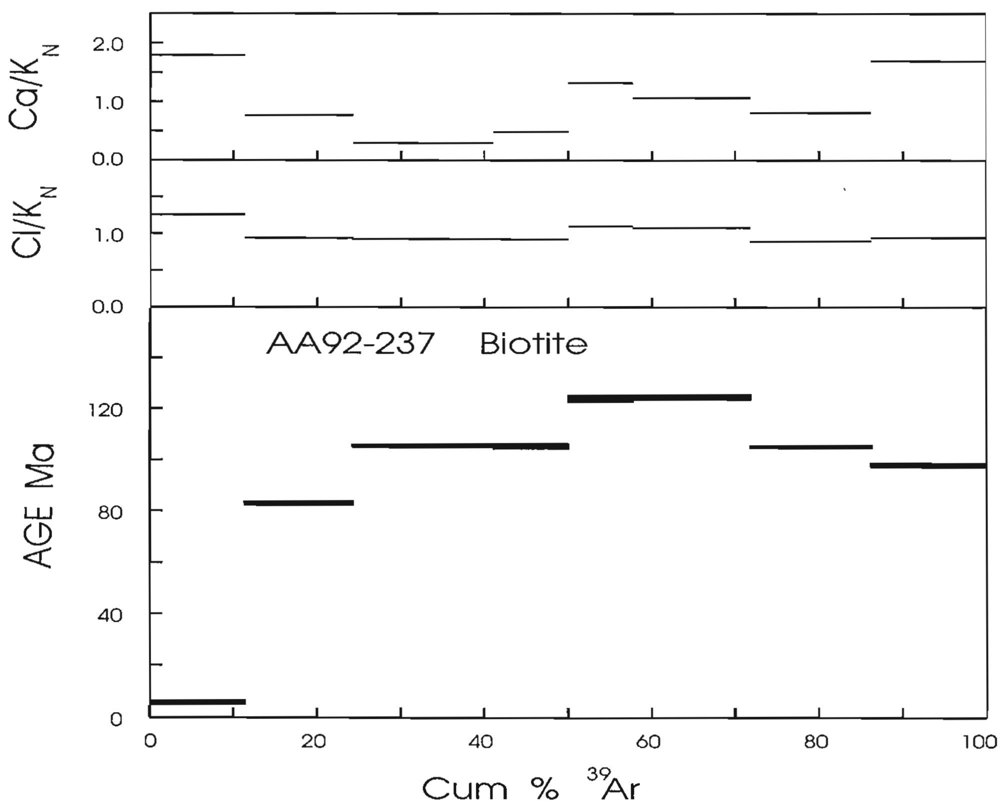

Figure 3.

^{40}Ar-^{39}Ar age spectrum for biotite from tuff GGA-87-14A3. The variation of Cl/K and Ca/K for the steps (normalized to the integrated value for all steps) are also plotted.

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A compilation of K-Ar and ^{40}Ar-^{39}Ar ages: Report 22

P.A. Hunt[1] and J.C. Roddick[1]

Hunt, P.A. and Roddick, J.C., 1992: A compilation of K-Ar and ^{40}Ar-^{39}Ar ages: Report 22; in *Radiogenic Age and Isotopic Studies: Report 6; Geological Survey of Canada, Paper 92-2, p. 179-226.*

Abstract

One hundred potassium-argon age determinations (including nineteen ^{40}Ar-^{39}Ar analyses) 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 les résultats de 100 datations (comprenant 19 analyses ^{40}Ar-^{39}Ar) par la méthode potassium-argon de la Commission géologique du Canada. Chaque datation est accompagnée d'une description de la roche ou du concentré minéral utilisé, ainsi que d'une brève interptétation expliquant la signification géologique du résultat lorsque possible. Les méthodes expérimentales sont aussi résumées. De plus, un index de toutes les datations par la méthode potassium-argon de la Commission géologique du Canada qui ont été publiées selon ce format de présentation a été préparé sur la base des numéros du SNRC.

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 1992 (Hunt and Roddick, 1992). In this new contribution 100 determinations are reported. For the first time a number of analyses (nineteen) using the ^{40}Ar-^{39}Ar technique are reported. An explanation of ^{40}Ar-^{39}Ar procedures used and general interpretation of the data are given below. 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

Conventional K-Ar

The data compiled here represent analyses carried out in 1990 and 1992, as well as a number from the District of Mackenzie analyzed in 1978. Potassium was analyzed by atomic absorption spectrometry on duplicate dissolutions of the samples. Conventional Argon extractions were carried out using a radio frequency 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 permanent 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

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(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 1978 samples are given in Wanless et al. (1973).

^{40}Ar-^{39}Ar Analyses

The Geochronology Laboratory has used the ^{40}Ar-^{39}Ar step heating technique for several years (Roddick, 1990). In 1991 the laboratory started using this technique in place of conventional K-Ar dating for certain minerals. In this technique a sample is irradiated in a nuclear reactor to convert some K atoms to ^{39}Ar. The ^{39}Ar is used as a measure of the K in the sample and a sample's age is determined by the measurement of the ^{40}Ar-^{39}Ar isotopic ratio (corrected for interfering isotopes and atmospheric Ar). By step-wise heating of a sample in a vacuum furnace ages can be calculated for Ar fractions released at incrementally higher temperatures. In general, ages determined from the higher temperature steps represent Ar released from more retentive sites in a mineral. For further analytical details see Roddick (1990) and for an explanation of the principles of the technique see McDougall and Harrison (1988) or Hanes (1991).

The analyses reported here consist of three heating steps, with the temperature of the first step selected to liberate most of the atmospheric argon but a minimum of the radiogenic argon from the sample. This step contains very little radiogenic Ar and usually is not reported. The next temperature step is selected to release about 50% of the radiogenic argon from the sample. A final fusion step releases any remaining Ar. The analyses are therefore essentially two age measurements and permit a comparative test of the consistency of the ages of argon released from a sample. If the ages of the two fractions are in agreement, it is assumed that a reliable age can be assigned to the sample. Should the ages differ then it is likely that there has been a disturbance to the K-Ar system in the sample.

The results are presented in a format similar to the previous K-Ar reports but with an additional section detailing the ages of the steps and the preferred age of the sample. The first age given represents the mean age of all three gas fractions weighted and summed according to the amounts of ^{39}Ar in each fraction, and is indicated as an integrated age with 2σ uncertainty. The error limit includes uncertainty in irradiation calibration of the amount of K converted to ^{39}Ar (J factor, typically $\pm$ 0.5-1.0% 2σ) which must be considered when comparing different sample ages. This age is equivalent to a conventional K-Ar age and, in samples that are not subject to recoil Ar loss (see McDougall and Harrison, 1988), is the age which would be determined by that technique. The percent atmospheric argon in the sample is given for this integrated age. The ages of the last two steps are given separately, along with their 2σ uncertainties and percentages of ^{39}Ar in the fractions. The uncertainties of these ages do not include irradiation calibration error since it does not contribute to uncertainties between heating steps on a single sample. If these two ages agree within their error limits the preferred age is a mean of these fractions weighted by the amounts of ^{39}Ar in the fractions. The error estimate for this age does include uncertainty in the irradiation calibration. This is termed the plateau age. If these two ages do not agree then one of the steps may be designated as the preferred age. In many cases of known complex geological history the age of highest temperature gas fraction may be the best estimate of the age as the lower temperature release may record a partial response to the most recent geological reheating of the sample. In some cases excess argon is present in the initial Ar released and in this case the highest temperature step is also the best estimate of the age of a sample. Some explanation of the reason for the preferred age is given in the geological discussion of a sample.

The potassium concentration of the sample is also given. This is determined from the calibration of the mass spectrometer as a precise manometer, the conversion factor for ^{39}Ar production, and the weights of the samples. The precision of this K concentration is one to four percent and is limited by errors associated with weighing of the small (4 to 30 mg) samples used for analyses.

The complete series of reports including the present one is given in the Appendix.

ACKNOWLEDGMENTS

We would like to thank Fred Quigg and Rejean Seguin for the argon extractions and Diane Bellerive for the potassium analyses.

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BRITISH COLUMBIA
(GSC 92-1 to 92-29)

GSC 92-1 Hornblende
164.1 ± 3.7 Ma

Wt % K= 0.385
Rad. Ar= 2.571 x 10^{-6} cm^3/g
K-AR 4340 % Atmos. Ar= 33.9

From a medium grained unfoliated hornblende-rich diorite.

(92 J/2) Collected from a west facing road cut on the east side of B.C. Highway 99, south of Pemberton, B.C.; 50°14′N, 122°52′W; UTM zone 10, 509500E, 5564300N; sample PCD-4-8; Collected by T.A. Vandall and J. Baker. Hornblende separation by R.M. Friedman, T.A. Vandall and J. Baker. Interpretation by T.A. Vandall and R.M. Friedman.

This sample and GSC 92-2 were collected from the Pemberton Diorite Complex which is situated within the western domain of the southern Coast Belt (Journeay, 1990). The geochronology of these samples was undertaken to establish minimum age constraints for thermal magnetizations identified by a paleomagnetic study of these rocks.

The complex is a heterogeneous assemblage of hornblende-rich diorite, quartz diorite and amphibolite with lesser granodiorite. Previous geochronology from the complex include a U-Pb zircon date of 113 ± 2 Ma (Friedman 1990) and a K-Ar hornblende date of 53 ± 15 Ma (Wanless et al., 1979). The 164.1 ± 3.7 Ma date is interpreted as a cooling age following pre-Late Jurassic intrusion of this part of the complex. The date reported herein provides evidence that the Pemberton Diorite Complex contains rocks of Jurassic and Cretaceous age. Alternatively, since the date is older than previous U-Pb and K-Ar dates, there could be an excess Ar component, but this is not the preferred interpretation.

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GSC 92-2 Hornblende
170.6 ± 3.3 Ma

Wt % K= 0.275
Rad. Ar= 1.912 x 10^{-6} cm^3/g
K-Ar 4341 % Atmos. Ar= 28.8

From a medium-fine grained unfoliated diorite.

(92 J/7) Collected from a west facing road cut on the east side of B.C. Highway 99, south of Pemberton (1.5 km north of GSC 92-1),

B.C.; 50°15′N, 122°52′; UTM zone 11, 510100E, 5566200N; sample PCD-10-11. Collected by T.A. Vandall and J. Baker. Hornblende separation by R.M. Friedman, T.A. Vandall and J. Baker. Interpretation by T.A. Vandall and R.M. Friedman.

This is the second sample of two collected from the Pemberton Diorite Complex (GSC 92-1). This date is in agreement (within error) with sample GSC 92-1 and is also interpreted as a cooling age following pre-Late Jurassic intrusion. In all probability these samples have been collected from the same intrusive body. The 6 Ma difference in their ages may reflect slightly different cooling histories. At this time pre-Late Jurassic and Cretaceous rocks have been identified within the northern part of Pemberton Diorite Complex and these are located in the southwest and northeast, respectively.

GSC 92-3 Whole Rock
48.2 ± 0.8 Ma

Wt % K= 2.176
Rad. Ar= 4.129 x 10^{-6} cm^3/g
K-Ar 4328 % Atmos. Ar= 3.7

From an augite basalt flow.
(92 O/9) At 1318 m (4325′) elevation, 1.65 km at 310° from elevation point 1491 m (4891′) on the ridge north of Gaspard Creek, southwestern British Columbia; 51°38′52"N, 122°29′06"W; UTM zone 10, 535640E, 5721770N; sample C90-756L. Collected and interpreted by P.B. Read.

The sample is from a sparsely amygdaloidal (2%), porphyritic 1% partly resorbed plagioclase, altered hypersthene (1%) augite basalt flow composed of 11% augite, 77% plagioclase, 4% opaque minerals, and 7% "clay minerals" of which the latter selectively replace hypersthene and fill the 2% amygdules that are 1-2 mm in diameter. The flow lies within 100 m of the base of a more than 1500 m thick succession of unnamed Eocene stratified rocks dominated by hornblende-bearing dacite in the northeastern corner of the Taseko Lakes area.

See GSC 92-6 for discussion.

GSC 92-4 Whole Rock
47.4 ± 0.7 Ma

Wt % K= 2.016
Rad. Ar= 3.761 x 10^{-6} cm^3/g
K-Ar 4329 % Atmos. Ar= 13.6

From an augite-hypersthene andesitic basalt.
(92 O/9) At 1364 m (4475′) elevation, 2.55 km at 133° from triangulation station 1581 m (5188′) on the north side of Word Creek, southwestern British Columbia; 51°40′45"N, 122°28′21"W; UTM zone 10, 536470E, 5725270N; sample C90-755E. Collected and interpreted by P.B. Read.

The sample is from a finely porphyritic augite-hypersthene andesitic basalt flow with phenocrysts of hypersthene (3%), and augite (4%) set in a matrix of 79% plagioclase (An_{51}), 1% biotite, 4% opaque minerals, and 8% augite. The sample comes from one of a string of outcrops with a moderately dipping platy jointing perhaps indicative of a dyke. It lies about 100 m above the base of a more than 1500 m thick succession of unnamed Eocene stratified rocks dominated by hornblende-bearing dacite in the northeastern corner of the Taseko Lakes area.

See GSC 92-6 for discussion.

GSC 92-5 Whole Rock
43.5 ± 1.1 Ma

Wt % K= 2.369
Rad. Ar= 4.055 x 10^{-6} cm^3/g
K-Ar 4332 % Atmos. Ar= 13.1

From an augite-hypersthene basalt flow.
(92 O/9) At 1448 m (4750′) elevation, 2.62 km at 217° from elevation point 1491 m (4891′), on the ridge crest north of Gaspard Creek, southwestern British Columbia; 51°37′14"N, 122°29′26"W; UTM zone 10, 535270E, 5718750N; sample C90-785L. Collected and interpreted by P.B. Read.

The sample is from a porphyritic (augite (1%), plagioclase (5%), hypersthene (4%)) basalt with a matrix of 86% plagioclase (An_{58-60}), 2?% opaque minerals, and 2?% "clay minerals". It lies about 200 m above the base of a more than 1500 m thick succession of unnamed Eocene stratified rocks dominated by hornblende-bearing dacite in the northeastern corner of the Taseko Lakes area.

See GSC 92-6 for discussion.

GSC 92-6 Whole Rock
50.5 ± 2.1 Ma

Wt % K= 1.722
Rad. Ar= 3.428 x 10^{-6} cm^3/g
K-Ar 4334 % Atmos. Ar= 10.5

From a fine grained olivine(?) augite basalt flow or dyke.

(92 O/10) At 1410 m (4625′) elevation close to the ridge crest north of Gaspard Creek, 5.8 km at 075° from the highway bridge across Gaspard Creek at 1036 m (3400′) elevation, southwestern British Columbia; 51°39′15"N, 122°33′28"W; UTM zone 10, 530660E, 5722440N; sample C90-780I. Collected and interpreted by P.B. Read.

The sample is from a fine grained olivine(?) augite basalt composed of 9% augite, 5% olivine? completely pseudomorphed by bowlingite all in matrix of 80% plagioclase (An_{54-57}), 6% opaque minerals, and biotite flakes. The basalt is intercalated within hornblende-bearing dacite flows and welded tuffs in the lower third of a more than 1500 m thick succession of unnamed Eocene stratified rocks dominated by dacite in the northeastern corner of the Taseko Lakes area.

Volcanic rocks of Eocene age are widespread on the west side of the Fraser River west of Fraser Fault for over 300 km from north of Lillooet in the south to north of Quesnel. In the Taseko Lakes area (NTS 92 O), basalt is widespread near the base of the Eocene succession and easily confused with basalt flows of the overlying Chilcotin Group. This dating (GSC 92-3, 4, 5, 6) allows a clear distinction between the Eocene and Miocene or later rocks. Although zeolite facies metamorphism has affected tuffaceous sediments in the upper part of the Eocene sequence (Read, 1988), its effect has been negligible in the impermeable flows except for GSC 92-5.

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GSC 92-7 Whole Rock
8.40 ± 0.27 Ma

Wt % K= 0.954
Rad. Ar= 3.121 x 10^{-7} cm^3/g
K-Ar 4331 % Atmos. Ar= 36.2

From a variolitic olivine augite basalt flow.

(92 O/9) At 869 m (2850′) elevation on the east side of the Fraser River, 2.40 km at 096° from the mouth of Churn Creek, southwestern British Columbia; 51°31′02"N, 122°14′51"W; UTM zone 10, 552220E, 5707400N; sample C90-2KAR. Collected and interpreted by P.B. Read.

The sample lies within 50 m of the base of a 100 m thick sequence of basalt flows and consists of 15% olivine, 30% augite, 51% plagioclase (An_{57-63}), and 4% opaque minerals. The rock is unaltered and free of glass but has varioles partly filled with hematite plates, slender prisms of titaniferous augite, and laths of plagioclase. The dated material was as free of varioles as possible.

See GSC 92-9 for discussion.

GSC 92-8 Whole Rock
3.20 ± 0.15 Ma

Wt % K= 0.129
Rad. Ar= 1.606 x 10^{-8} cm^3/g
K-Ar 4333 % Atmos. Ar= 90.0

From a vesicular pebble of olivine basalt.

(92 O/9) At 815 m (2675′) elevation 3.85 km at 014° from the mouth of Alkali Creek, southwestern British Columbia; 51°44′46"N, 122°20′38"W; UTM zone 10, 545300E, 5733100N; sample C90-818I. Collected and interpreted by P.B. Read.

The sample is a rounded pebble of vesicular olivine basalt composed of 16% olivine phenocrysts in a matrix of 64% plagioclase (An_{55-58}), 15% augite, 3% opaque minerals, and 2% glass of which the latter is interstitial to the plagioclase and augite. The sample lies within a metre of the top of sediments deposited in a north-draining Fraser River paleochannel and underlies the flows dated at 2.46 ± 0.14 Ma in GSC 92-9. The radiometric age of this clast provides the maximum age possible for the upper metre of sediments underlying the basalt flows and proves that the Chilcotin Group includes sediments of Pliocene age.

See GSC 92-9 for discussion.

GSC 92-9 Whole Rock
2.46 ± 0.14 Ma

Wt % K= 0.493
Rad. Ar= 4.721 x 10^{-8} cm^3/g
K-Ar 4330 % Atmos. Ar= 67.4

From a diabasic olivine augite basalt flow.

(92 O/9) At 815 m (2675′) elevation 3.77 km at 017° from the mouth of Alkali Creek, southwestern British Columbia; 51°44′53"N, 122°20′30"W; UTM zone 10, 545450E, 5733000N; sample C90-818G. Collected and interpreted by P.B. Read.

The sample comes from an olivine augite basalt composed of 50% plagioclase (An_{58-60}), 36% augite, 3% olivine, 4% opaque minerals, and 7% glass of which the latter is interstitial to the plagioclase and augite. The sample lies within a metre of the base of an 8 m thick flow which overlie dated Pliocene sediments (GSC 92-8) deposited in a north-draining Fraser River paleochannel.

In a compilation of K-Ar whole-rock ages of basalt flows from the Chilcotin Group, Mathews (1989) noted that eruptions were particularly abundant in the Middle Miocene (14-16 Ma), Late Miocene (6-9 Ma), and Plio-Pleistocene (1-3 Ma) intervals. Although previous dating (Mathews and Rouse, 1984, 1986) suggested that Miocene flows were not preserved immediately east of the Fraser River north of Big Bar Creek and south of Williams Lake, a remnant of Late Miocene basalt flows forms part of the east canyon rim between Dog and Canoe creeks (GSC 92-7). Its presence along with palynologically dated sediments of Early to Middle Miocene age immediately west of the Fraser River at Gang Ranch (Mathews and Rouse, 1984), and undated rhyolite ash and lapilli tuff north of Gang Ranch (Hickson et al., 1991) imply the existence of a north-draining Fraser River paleochannel of Miocene age subparallel to the Fraser River from Canoe to Riske creeks.

From Big Bar Creek to Williams Lake, Plio-Pleistocene flows are widespread along the Fraser. A sequence of 2-3 Ma old basalt flows extends for 20 km along the east side of the river from Dog Creek to north of Alkali Creek where it is no longer evident (GSC 92-8). Although Mathews and Rouse (1986) postulated the presence of Pliocene sediments underneath the Pliocene flows, evidence of sediments of this age was missing until the dating of a Pliocene basalt clast (GSC 92-9) from sediments beneath dated Pliocene flows. This dating indicates that a north-draining Fraser River paleochannel of Pliocene age subparallelled the present course of the Fraser.

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GSC 92-10 Whole Rock
5.8 ± 0.1 Ma

Wt % K= 4.513
Rad. Ar= 1.01 x 10^{-6} cm^3/g
K-Ar 4279 % Atmos. Ar= 22.0

From a flat-lying lava flow of sanidine-pyroxene-phyric comenditic trachyte.

(93 C/14) From an elevation of 1706 m on the south-east side of Carnlick Creek, north of Mizzen Mountain, Ilgachuz Range, British Columbia; 52°46.9'N, 125°13.7'W; UTM zone 10, 349707E, 5850093N; sample SE310785. Collected and interpreted by J.G. Souther.

The sample is from the basal flow of the Ilgachuz Range shield volcano. The age is believed to represent the original cooling age of one of the first lavas erupted from the volcano. The 5.8 Ma date on this sample is consistent with a date of 5.1 Ma (GSC 92- 11) from a flow slightly higher in the succession.

GSC 92-11 Whole Rock
5.1 ± 0.1 Ma

Wt % K= 4.061
Rad. Ar= 8.004 x 10^{-7} cm^3/g
K-Ar 4280 % Atmos. Ar= 11.2

From a flat-lying lava flow of sanidine-pyroxene-phyric comenditic trachyte.

(93 C/11) From an elevation of 1280 m on the escarpment bounding the western side of the Ilgachuz Range between Arnica Creek and Far Creek, British Columbia; 52°44.8'N, 125°25.4'W; UTM zone 10, 336426E, 4846626N; sample DG213-85. Collected and interpreted by J.G. Souther.

The sample is from a lava flow about 100 m above the base of the Ilgachuz Range shield volcano. The age is believed to represent the original cooling age of one of the early lavas erupted from the volcano. The 5.1 Ma date on this sample is consistent with a date of 5.8 Ma (GSC 92-10) from the basal flow.

GSC 92-12 Whole Rock
60.5 ± 0.7 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 6.6
K-Ar 4281 % Atmos. Ar= 12.7
(270) Ages of two heating steps + % gas:
59.6 ± 0.6 Ma (19%), 57.8 ± 0.2 Ma (75%)
Preferred age: step 2 age of 57.8 ± 0.2 Ma

From a coarse grained moderately fractured granite.
(93 C) From an elevation of 1615 m on the south-east flank of Saxifraga Mountain, central Ilgachuz Range, British Columbia; 52°43.1'N; 125°19.8'W; UTM zone 10, 342623E, 5843267N; sample SE4002B89. Collected and interpreted by J.G. Souther.

The sample is from a suite of moderately to highly deformed, locally mylonitic rocks which are exposed in isolated outcrops beneath the Tertiary lavas of the Ilgachuz Range. The relatively undeformed granite is cut by a random network of fractures which give it a cataclastic fabric. The feldspar has deformed by brittle fracturing whereas quartz shows evidence of ductile strain. The preferred age of 57.8 Ma on this sample suggests that it is from a metamorphic complex that is coeval and probably coextensive with the Eocene Tatla Lake metamorphic core complex described by Friedman and Armstrong (1988) south of Anahim Lake. The Early Tertiary thermal event that affected these rocks and reset the K-Ar dates was apparently more extensive than previously thought and could have implications regarding the petroleum potential of the Nechako Basin.

The date is the result of an Eocene thermal event which reset the cooling ages of a variety of older rocks. The protolith of this sample is believed to be a Jurassic granite. Also the sample is multi-mineralic and the first step (19% gas) could be subject to incorporation of some excess argon.

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GSC 92-13 Biotite
46.1 ± 0.9 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 5.9
K-Ar 4284 % Atmos. Ar= 27.5
(225) Ages of two heating steps + % gas:
45.7 ± 0.2 (42%), 45.8 ± 0.2 (56%)
Preferred age: 99% plateau age of 45.8 ± 0.5 Ma

From a dacite.
(93 J/13) From the east side of the ridge leading south from Mount Mckinnon approximately 300 m from north of the main east-west forest access road across the mountain, British Columbia; 54°52'10"N, 123°39'30"W; UTM zone 10, 457750E, 6080250N; sample SCB-87-470a. Collected and interpreted by L.C. Struik.

This dacite comes from a 3.5-4 m thick dyke that intrudes flattened and elongated leucogranite. The dyke has a finely crystalline margin against the leucogranite and a coarsely crystalline core (up to 8mm crystals). It dips steeply to the west-northwest, striking to 020°-025°. Faults across the dyke show centimetre scale offsets and are mainly hairline cracks.

Muller and Tipper (1969) mapped these dykes as unit 18a, and Deville and Struik (1990) mapped them as unit D, one of four suites of intrusions in sillimanite grade metasedimentary rocks of the Wolverine Metamorphic Complex. They showed that the dykes form a north-northeast-trending swarm throughout the exposures on Mount Mckinnon.

Deville and Struik (1990) interpreted the dacite of this dyke, and other dykes like it, to have formed in the upper crust during crustal extension that unroofed the Wolverine Metamorphic Complex. The finely crystalline margin of the dyke may have formed because the dacite intruded an already cool country rock. In that case the age of the biotite may be near the age of dyke intrusion. Final uplift and exposure of the Wolverine Complex would be younger than the 46 Ma dyke.

The 46 Ma age for the dacite is similar to the age of biotite from the microgranite GSC 92-14 collected 15 km to the southwest. These two dates could represent either a single period of uplift of similar levels of the crust, or a common crustal stretching and intrusion event.

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GSC 92-14 Biotite
46.6 ± 0.9 Ma

Wt % K= 6.838
Rad. Ar= 1.254 X 10^{-5} cm^3/g
K-Ar 4285 % Atmos. Ar= 22.7

From a microgranite.
(93 J/13) From a small rise south of a lake called Squawfish Lake by the British Columbia Forest Service, 9 km south and 1.5 km west of Salmon Lake; 54°46′47"N, 123°51′02"W; UTM zone 10, 445300E, 6070400N; sample SCB-89-253. Collected and interpreted by L.C. Struik.

This isolated exposure of scattered outcrops of microgranite is made up of finely crystalline (1-1.5 mm) biotite, feldspar and quartz in a altered matrix. Locally the quartz and feldspar form phenocrysts of up to 5 mm in size. Biotite forms approximately 5% of the rock. In places the rock looks like a biotite-qtz-feldspar phenocrystic rhyolite. It is undeformed.

The microgranite directly underlies agglomeratic and flow basalt in the creek valley directly southeast of Squawfish Lake. To the southeast and northwest of the microgranite are exposures of the Triassic Takla Group fragmental and flow basalts. To the west are exposures of basalt (Muller and Tipper, 1969), now interpreted as Miocene Chilcotin Group. Muller and Tipper (1969) interpreted the Squaw Fish Lake microgranite to be Paleocene? to Miocene in age, and correlated it with rhyolite dykes on Mount Mckinnon and a rhyolite flow at Carp Hill northeast of Carp Lake (Struik, 1989).

The basalt overlying the microgranite is correlated with the Miocene Chilcotin Group, because it has the same textures, macroscopic mineralogy, colour and degree of freshness, and it is flat lying, and the structurally highest basalt unit. The alteration of the microgranite is interpreted to have happened prior to the deposition of the overlying Miocene basalt, because the degree of weathering in the microgranite is not seen in the overlying basalt, and all of the exposed microgranite has the same altered texture. The microgranite, therefore, could have been exposed as a Miocene or perhaps pre-Miocene paleosurface.

The middle Eocene age for the biotite is consistent with the age assigned to this rock by Muller and Tipper (1969). The biotite age may represent a cooling age vastly different than the age of intrusion, however, because the rock is neither metamorphosed or coarse grained, its intrusive age is considered to be within several million years of the biotite age. Rhyolite dykes on Mount Mckinnon have yielded a middle Eocene biotite age very similar to this one. This is consistent with the correlation suggested by Muller and Tipper if these ages are within several million years of the intrusive age (see GSC 92-13 for related sample).

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GSC 92-15 Hornblende
29.5 ± 3.7 Ma

Wt % K= 0.183
Rad. Ar= 2.114 x 10^{-7} cm^3/g
K-Ar 4343 % Atmos. Ar= 74.9

From hornblende-feldspar-phyric andesite.
(103 B/11) From a sheeted dyke complex on the east side of Lyell Island, Queen Charlotte Islands, British Columbia; 52°42′N, 131°25.5′W; UTM zone 9, 336139E, 5841440N; sample SE360288. Collected and interpreted by J.G. Souther.

This sample is from a major dyke swarm in the Lyell Island igneous complex where plutonic, hypabyssal, and volcanic rocks are believed to be coeval and cogenetic facies of a major centre of Tertiary Igneous activity (Souther and Jessop, 1991). The 29.5 Ma age of this rock is intermediate between predominantly Eocene dates from Tertiary dykes and plutons farther south and predominantly Miocene dates from Tertiary lavas farther north. It is thus consistent with the observed shift in the locus of Tertiary igneous activity in the Queen Charlotte Islands from south to north throughout Tertiary time (Anderson and Reichenbach, 1991; Souther and Jessop, 1991).

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GSC 92-16 Whole rock
5.07 ± 0.39 Ma

Wt % K= 1:240
Rad. Ar= 2.447 x 10^{-7} cm^3/g
K-Ar 4309 % Atmos. Ar= 34.5

From a basalt.
(104 I/4) From a ridge top 9 km northwest of the confluence of Beggerlay and Moose creeks, Cry Lake map area, northwestern British Columbia; 58°04'03"N, 129°37'38"W; UTM zone 9, 4630000E, 6436200N; sample AN-77-195-3. Collected and interpreted by R.G. Anderson.

The sample is an olivine basalt, part of a sequence of massive, grey to dark grey weathering andesitic basalt to olivine basalt flows. The sequence is more than 30 m thick and occurs in a north plunging paleovalley incised into the Late Triassic Cake Hill Pluton in the Hotailuh Batholith (Anderson, in Stevens et al., 1982; Anderson, 1983).

The [late] Pliocene whole rock date is consistent with the correlation of the volcanic rocks with the Miocene-Quaternary Stikine Volcanic Belt (Souther, 1990). Porphyritic basalt of late Miocene-Quaternary age is common in the Stikine Volcanic Belt (Hickson, 1990). Near the Stikine River and north of Mount Tsaybahe (and south and west of the dated sample), six petrographically similar samples collected by Peter Read in 1982 yielded similar ages (5.4-4.6 Ma; Canadian Cordilleran Geochron File, Armstrong, 1988). The dated sample is also part of an ongoing geochemical and isotopic study of the Stikine Volcanic Belt by M.L. Bevier and her co-workers (Bevier, 1989).

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GSC 92-17 Hornblende
57.9 ± 0.9 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 0.78
K-Ar 4297 % Atmos. Ar= 19.3
(226) Ages of two heating steps + % gas:
57.6 ± 0.9 Ma (45%), 57.8 ± 0.8 Ma (51%)
Preferred age: 96% plateau age of 57.7 ± 0.8 Ma

From a foliated hornblende schist.

(104 B/12) In a saddle, 0.9 km east of Elbow Mountain, 3.2 km north-northwest of northern tip of Kakati Lake, 1070 m (3500′) elevation, western Iskut River map area, northwestern British Columbia; 56°42′15"N, 131°50′41"W; UTM zone 9, 325850E, 6287850N; sample AT-85-197-1. Collected and interpreted by R.G. Anderson.

The foliated hornblende schist sample is typical of the metabasite unit described by Porter (1992) in the southern Elbow Mountain crystalline complex. Metabasite rocks are part of the 500 m thick panel of metamorphic rocks in the complex which also include garnet-bearing metapelite, metaconglomerate, marble, calc-silicate and uncommon quartzite. The metamorphic rocks are intruded by a hornblende diorite (see GSC 92-19) dated at ca. 189 Ma (U-Pb, zircon; M.L. Bevier, unpublished data). The pluton is interpreted as late synkinematic to post-kinematic with respect to the S_1 foliation which developed synchronously with formation of the metamorphic minerals. The complex is intruded along its northern margin by Tertiary quartz monzonite of the Hyder plutonic suite; a nearby pluton of similar composition north of the Great Glacier yielded a ca. 51 Ma date (U-Pb, zircon; M.L. Bevier, unpublished data).

The hornblende schist was sampled to test the hypothesis of an Early Jurassic or older deformation and metamorphism suggested by the structure, geological relationships with the hornblende diorite and the U-Pb date for that intrusion. The preferred age of 57.7 ± 0.8 Ma for the hornblende suggests that the schist (and likely the entire Elbow Mountain complex) was thermally disturbed and its Ar systematics nearly completely reset in the Eocene, likely during the widespread intrusion of felsic intrusions of the Eocene Hyder plutonic suite.

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GSC 92-18 Hornblende

235.5 ± 3.5 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 0.38

K-Ar 4298 % Atmos. Ar= 12.0

(227) Ages of two heating steps + % gas:

258.7 ± 2.1 Ma (14%), 229.9 ± 2.4 Ma (83%)

Preferred age: step 2 at 229.9 ± 2.4 Ma

From a leucocratic hornblende monzodiorite.

(104 B/7) On the west flank of McQuillan Ridge, 3.45 km southeast of confluence of Cebuck Creek and South Unuk River, 8.4 km northeast of northeastern end of Flory Lake, 1310 m (4300 feet) elevation, southeastern Iskut River map area, northwestern British Columbia; 56°25′35"N, 130°32′02"W; UTM zone 9, 405400E, 6254375N; sample AT-85-154-3. Collected and interpreted by R.G. Anderson.

The leucocratic monzodiorite is typical of the McQuillan Ridge pluton; the massive, equigranular-hypidiomorphic, medium-grained pluton is characterized by euhedral, equant, square hornblende. The pluton intrudes well-dated Late Triassic (Carnian), ca. 226-220 Ma, Stuhini Group strata (Grove, 1986; Anderson in Hunt and Roddick, 1991, sample GSC 90-44, and M.J. Orchard and M.L. Bevier, unpublished data). U-Pb data for zircon from the sample indicates an age of ca. 226 +5/-2 Ma (M.L. Bevier, unpublished data).

The first heating step released only 14% of the gas and must contain excess argon. The preferred age (230 ± 2 Ma), obtained from the second heating step which released 83% of the gas, is in agreement with the U-Pb age for the sample's zircon and corroborates the geological relationships and the estimated duration of the Carnian stage (235-223 Ma; Harland et al., 1990).

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GSC 92-19 Hornblende

50.8 ± 1.0 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 1.2

K-Ar 4299 % Atmos. Ar= 29.2

(228) Ages of two heating steps + % gas:

54.8 ± 0.6 Ma (46%), 46.9 ± 1.3 Ma (53%)

Preferred age: step 2 at 46.9 ± 1.3 Ma

From a foliated biotite-hornblende diorite.
(104 B/12) 100 m southwest of Elbow Mountain, 1292 m (4238 feet) elevation, western Iskut River map area, northwestern British Columbia; 56°42′11"N, 131°51′30"W; UTM zone 9, 325000E, 6287775N; sample AT-85-194-3. Collected and interpreted by R.G. Anderson.

The biotite-hornblende diorite sampled provides an important youngest age for first phase deformation and metamorphism in the Elbow Mountain crystalline complex (Porter, 1992). The intrusion cross-cuts the S_1 foliation developed synchronously with formation of the metamorphic minerals. Along its margins, the diorite is intensely foliated; foliation is oriented subparallel to the northwest-trending, moderately easterly-dipping schistosity and gneissosity in the panel of metamorphic rocks (Porter, 1992). The intrusion is interpreted as late synkinematic to post-kinematic with respect to the first phase deformation and metamorphism which was dated at ca. 189 Ma (U-Pb, zircon; M.L. Bevier, unpublished data). The complex is intruded along its northern margin by Tertiary quartz monzonite of the Hyder plutonic suite; a nearby pluton of similar composition north of the Great Glacier yielded a ca. 51 Ma date (U-Pb, zircon; M.L. Bevier, unpublished data).

The hornblende from the diorite was sampled to test the hypothesis of an Early Jurassic or older deformation and metamorphism suggested by the structure, geological relationships with the intrusion, diorite and the U-Pb date for that intrusion (see also GSC 92-17). The first heating step released 46% of the gas and could contain a small amount of excess argon, a common characteristic of metamorphic hornblende. The age of 46.9 ± 1.3 Ma for step 2 (53% gas released) is the preferred of the two discordant Ar step heating ages. It suggests that the hornblende in the Early Jurassic intrusion was thermally disturbed and its Ar systematics completely reset in the Eocene, likely during the widespread emplacement of felsic intrusions of the Eocene Hyder plutonic suite; however, the age is about 4 million years younger than the nearby, well-dated Great Glacier pluton.

REFERENCE

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GSC 92-20 Biotite
49.2 ± 0.9 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.3
K-Ar 4319 % Atmos. Ar= 46.0
(238) Ages of two heating steps + % gas:
49.0 ± 0.7 Ma (34%), 48.6 ± 0.3 Ma (62%)
Preferred age: 96% plateau age of 48.7 ± 0.6 Ma

From a biotite-hornblende andesitic porphyry.
(104 B/7) On the west flank of McQuillan Ridge, 1340 m (4400 feet) elevation, 2.75 km southeast from southwest end of Flory Lake and 5.5 km northwest of confluence of Boulder Creek and Unuk River, southeastern Iskut River map area, northwestern British Columbia; 56°21′59"N, 130°38′09"W; UTM zone 9, 398950E, 6247850N; sample AT-89-75-3. Collected and interpreted by R.G. Anderson.

For interpretation see GSC 92-21.

GSC 92-21 Hornblende
52.1 ± 1.3 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 0.66
K-Ar 4320 % Atmos. Ar= 28.7
(240) Ages of two heating steps + % gas:
51.2 ± 2.7 Ma (13%), 52.4 ± 0.3 Ma (86%)
Preferred age: 99% plateau age of 52.3 ± 0.7 Ma

(104 B/7) details as for GSC 92-20

The sample is from a fresh, massive biotite-hornblende andesitic porphyry dyke which crosscuts well-dated Late Triassic, ca. 226-220 Ma, Stuhini Group strata (Grove, 1986; Anderson in Hunt and Roddick, 1991, sample GSC 90-44, and M.J. Orchard and M.L. Bevier, unpublished data). The dykes are associated with tabular, white trondhjemite intrusions (shown as limestone unit 10e in Grove, 1986) that are common east of Flory Lake.

The preferred hornblende date (52.3 ± 0.7 Ma) is discordant with that for biotite (48.7 ± 0.6 Ma; GSC 92-20) but is the best estimate for intrusion of the andesitic porphyry dyke. These Eocene dykes (and related trondhjemite?) are part of the Tertiary Hyder plutonic suite emplaced between 55-51 Ma based on U-Pb dating of plutons and the Portland and Mt. Welker dyke swarms (Alldrick et al., 1986; M.L. Bevier and J.K. Mortensen, unpublished data).

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GSC 92-22 Hornblende
171.1 ± 3.2 Ma integrated ^{40}Ar-^{39}Ar

Wt. % K = 0.52
K-Ar 4321 % Atmos. Ar = 9.0
(241) Ages of two heating steps + % gas:
169.0 ± 0.6 Ma (41%), 171.7 ± 0.5 Ma (58%)
Preferred age: 99% plateau age of 170.6 ± 1.7 Ma

From a deformed, lineated and foliated biotite-hornblende metadiorite.
(104 B/8) Just southeast of the DOC property portal, 1160 m (3805 ft) elevation, 2 km southwest of confluence of Devilbliss Creek and South Unuk River, 6.75 km south-southeast of confluence of Gracey Creek and South Unuk River, southeastern Iskut River map area, northwestern British Columbia; 56°20′22"N, 130°26′40"W; UTM zone 9, 410700E, 6244600N; sample AT-89-69-1. Collected and interpreted by R.G. Anderson.

The sample is one of a number of small deformed mafic and felsic stocks and intrusions near the DOC property. A well-foliated metadiorite pluton near Bucke Glacier (Grove, 1986) and southwest of the sample locality, is characterized by a steeply north-northeast-plunging biotite lineation and was dated at about 221 ± 1 Ma (U-Pb, zircon; M.L. Bevier, unpublished data). Biotite lineation in the sampled metadiorite plunges gently northwest. Nearby, an aplitic monzonite sill intruded and was folded with the D_1 foliation developed in metavolcanic country rocks about a moderately-plunging, north-trending fold axis. The biotite lineation in the sampled metadiorite is nearly collinear with the fold axis of the folded dyke. The folded dyke was dated at about 176 ± 4 Ma (U-Pb, zircon; M.L. Bevier, unpublished data).

Note that the ages for the two gas release steps disagree slightly with respect to their uncertainties but the difference is small (2.7 Ma). The best estimate for the hornblende cooling age at 172 ± 2 Ma is concordant with the U-Pb date for the folded dyke and the two bodies have structural elements in common. These results suggest that Aalenian deformation, intrusion and cooling was important in this part of Stikinia.

REFERENCE

Grove, E.W.
1986: Geology and mineral deposits of the Unuk River-Salmon River-Anyox area; British Columbia Ministry of Energy, Mines and Petroleum Resources, Bulletin 63, 434 p.

GSC 92-23 Whole Rock
1.32 ± 0.03 Ma

Wt % K= 1.739
Rad. Ar= 8.948 x 10^{-8} cm^3/g
K-Ar 4306 % Atmos. Ar= 59.0

From a sodic tephritic phonolite.
(104 H/8) From a nunatak in a small snowfield on a north-facing slope in Spatsizi Wilderness Provincial Park, British Columbia; 57°27.75′N, 128°27.50′W; UTM zone 9, 532500E, 6368800N; sample GAT-87-224-1. Collected and interpreted by D. Thorkelson.

This sample is from a dyke or volcanic neck that was emplaced into a folded, Late Triassic volcano-sedimentary succession (Thorkelson, 1992). The rock is a grey-weathering, sodic tephritic phonolite (22.3% ab+or, 13.4% an, 16.9% ne). No extrusive rocks related to the intrusion are known from the immediate area. The nearest Late Cenozoic extrusive rocks are the Maitland volcanic rocks, 70 km to the west. The age is interpreted as the emplacement age for this sub-volcanic intrusion.

REFERENCE

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1992: Volcanic and tectonic evolution of the Hazelton Group in Spatsizi River (104H) map-area, north-central British Columbia; Ph.D. thesis, Carleton University, Ottawa, 281 p.

GSC 92-24 Whole Rock
0.73 ± 0.03 Ma

Wt % K= 0.583
Rad. Ar= 1.665 x 10^{-8} cm^3/g
K-Ar 4307 % Atmos. Ar= 78.7

From a basalt.
(104 I) From a small knoll on the northwest side of Cry Lake, 1.8 km east of the west end of Cry Lake, Cry Lake may area (104I), British Columbia; 58°42′N, 129°14′W; UTM zone 9, 486800E, 6505500N; sample GA-89-6. Collected and interpreted by. H. Gabrielse.

The rock is a light grey weathering, vesicular basalt with amber, plagioclase phenocrysts. It is apparently flat lying.

The basalt represents the most easterly exposure of these rocks known in the Cry Lake region and was emplaced onto rocks considered to be part of Ancestral North America. The age is consistent with those for other volcanic units in the eastern Cassiar Mountains and Liard Plain to the north.

GSC 92-25 Whole Rock
4.3 ± 0.3 Ma

Wt % K= 1.203
Rad. Ar= 2.014 x 10^{-7} cm^3/g
K-Ar 4308 % Atmos. Ar= 93.5

From a basalt.
(104 O/9) From a prominent flat-topped butte 3 km south of Little Rancheria River, British Columbia; 59°43.3′N, 130°10.35′W; UTM zone 9, 433780E, 6620420N; sample JN-88-53-2. Collected and interpreted by J. Nelson.

This is from an isolated outlier of fresh, columnar-jointed basalt of the Tuya Formation (Gabrielse, 1969). The date is interpreted as the age of the flow.

The topographically-high position of this flow compared with other basalt outliers in Pleistocene valley bottoms suggests that it may be a relatively older unit in the Tuya Formation.

REFERENCE

Gabrielse, H.
1969: Geology of the Jennings River map-area, British Columbia; Geological Survey of Canada, Paper 68-55, 37 p.

GSC 92-26 Hornblende
341 ± 7 Ma

Wt % K= 0.639
Rad. Ar= 9.322 x 10^{-6} cm^3/g
K-Ar 4259 % Atmos. Ar= 76.0

From an amphibolite.
(104 I/15) On a ridge west side of Major Through valley between Rapid and Major Hart rivers, 1 km southeast of peak, elevation 2149 m, Cry Lake map area, Cassiar Mountains, British Columbia; 58°50′45"N, 128°44′00"W; UTM zone 9, 515390E, 6522707N; sample GA-89-21. Collected and interpreted by H. Gabrielse.

Sample is from a banded, medium- to coarse-grained amphibolite which in places forms layers or blocks within a feldspathic matrix. It is spatially associated with layers of ultramafic rock near the contact with a granodiorite pluton (Gabrielse, 1979).

The age could represent the age of emplacement of the granodiorite pluton or of metamorphism possibly near the base of the oceanic crust. The latter seems more probable in view of the age of GSC 92-27.

REFERENCE

Gabrielse, H.
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GSC 92-27 Hornblende
359 ± 8 Ma

Wt % K= 0.389
Rad. Ar= 6.002 x 10^{-6} cm^3/g
K-Ar 4260 % Atmos. Ar= 33.0

From an amphibolite.
(104 I/15) About 5 km west of Through Valley between Rapid and Major Hart rivers, 0.75 km ENE of peak, elevation 2157 m just north of Small Lake, Cry Lake map area, Cassiar Mountains, British Columbia; 58°47′30"N, 128°42′30"W; UTM zone 9, 516859E, 6516682N; sample GA-89-68A. Collected and interpreted by H. Gabrielse.

Sample is from a strongly foliate amphibolite associated with serpentinized alunite.

The age could represent metamorphism of rocks near the base of oceanic crust.

GSC 92-28 Muscovite
334 ± 5 Ma

Wt % K= 7.95
Rad. Ar= 1.133 x 10^{-4} cm^3/g
K-Ar 4258 % Atmos. Ar= 0.7

From a garnet-muscovite-quartz-feldspar schist.
(104 P/3) On ridge crest east of Four Mile River just south of peak, elevation 2110 m (6920 ft), McDame map area, Cassiar Mountains, British Columbia; 59°03'15"N, 129°05'00"W; UTM zone 9, 495220E, 6545878N; sample GA-89-17. Collected and interpreted by H. Gabrielse.

Sample is a garnet-muscovite-quartz-feldspar schist with some calc-silicate layers and possibly leucocratic granitic rocks. It forms part of a thrust slice within the Sylvester Allochthon.

The age is considered to be the time of metamorphism of the fault bounded sheet.

GSC 92-29 Whole Rock
0.54 ± 0.02 Ma

Wt % K= 1.588
Rad. Ar= 3.355 x 10^{-8} cm^3/g
K-Ar 4155 % Atmos. Ar= 89.6

From a olivine basalt.
(104 N/11) From along Ruby Creek, east of Atlin Lake, B.C.; 59°40.9'N, 133°20.3'W; UTM zone 8, 661700E, 5936000N; sample MLB-89-RC-1. Collected by M. Bloodgood and interpreted by J.K. Mortensen.

This sample is from a vesicular, aphyric, olivine basalt flow that is interlayered with pre-Pleistocene gold-bearing gravels of the Atlin District. The rock contains abundant small, white, quartz-rich xenoliths. It is correlated with volcanic rocks of the Stikine volcanic belt. The age is interpreted as the age of eruption of the flow.

YUKON TERRITORY
(GSC 92-30 to GSC 92-50)

GSC 92-30 Hornblende
183.6 ± 2.0 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 1.0
K-Ar 4311 % Atmos. Ar= 16.7
(233) Ages of two heating steps + % gas:
169.1 ± 3.1 Ma (20%), 188.5 ± 0.5 Ma (77%).
Preferred age: 97% plateau age of 184.5 ± 1.9 Ma.

From a hornblendite.
(105 C/13) From 14.6 km bearing 43° from confluence of Swift River and Teslin River, Yukon; 60°54.26'N, 133°40.04'W; UTM zone 8, 572293E, 6752661N; sample GGA-90-28-01B. Collected and interpreted by S.P. Gordey.

For interpretation and references see GSC 92-31.

GSC 92-31 Biotite
187.4 ± 5.2 Ma integrated ^{40}Ar-^{39}Ar

Wt. % K= 6.2
K-Ar 4316 % Atmos. Ar= 20.1
(239) Ages of two heating steps + % gas:
187.9 ± 0.7 Ma (39%), 186.8 ± 0.4 Ma (61%)
Preferred age: 99% plateau age of 187.2 ± 1.8 Ma

From a monomineralic biotite rock.
(105 C/13) From 14.6 km bearing 38° from the confluence of Swift River and Teslin River, Yukon; 60°54.71N, 133°41.06W; UTM zone 8, 571356E, 6753477N; sample GGA-90-28-04B. Collected and interpreted by S.P. Gordey.

GSC 92-30 and 92-31 were collected from a metagreenstone body northeast of the Teslin Fault, from rocks that form part of the Teslin suture zone (Stevens, 1992, unit PMgr). The greenstone body, although not mapped fully, is in the order of several kilometres in size. It is foliated along its northern margin where it is in contact with foliated marble and quartzite, but is internally massive. GSC 92-30 is an unfoliated hornblendite composed of randomly oriented fresh hornblende crystals up to 8 cm long. GSC 92-31 is from a monomineralic rock composed of randomly oriented fresh biotite flakes up to 2 cm in diameter. Both of these rocks occur in single exposures in the order of 5-10 m across, and both merge by diminution in grain size into the fine grained greenstone which surrounds them. The rocks are not dykes, and are interpreted as produced through metamorphism of the greenstone under uncertain but specific localized conditions. The ages are the same within experimental uncertainty, and are interpreted as metamorphic cooling ages. Somewhat older ^{40}Ar-^{39}Ar ages of 188-195 Ma on hornblende and white mica have been reported by Hansen et al. (1991) for Teslin suture zone rocks on strike to the northwest. Hansen et al. interpreted these ages as the time of rapid uplift of Teslin suture zone rocks upon collision with strata of the North American margin.

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GSC 92-32 Biotite
95.4 ± 1.6 Ma

Wt % K= 6.510
Rad. Ar= 2.478 x 10^{-5} cm^3/g
K-Ar 4290 % Atmos. Ar= 10.2

From a quartz feldspar biotite hornblende crystal tuff.
(105 J/04) From 10.9 km east of the north end of Orchie Lake, Yukon, 62°9.47′N, 131°38.25′W; UTM zone 9, 362590E, 6894370N; sample GGA-87-13C3. Collected and interpreted by S.P. Gordey.

For interpretation and references see GSC 92-35.

GSC 92-33 Hornblende
100.3 ± 1.9 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 0.48
K-Ar 4302 % Atmos. Ar= 18.3
(231) Ages of two heating steps + % gas:
99.9 ± 2.4 Ma (12%), 98.8 ± 0.5 Ma (86%)
Preferred age: 98% plateau age of 98.9 ± 1.1 Ma

From a hornblende porphyry.
(105 J/04) From 17.4 km southeast of the north end of Orchie Lake and 0.6 km south of Big Timber Creek, Yukon. 62°4.14′N, 131°34.79′W; UTM zone 9, 365197E, 6884340N; sample GGA-87-10G3. Collected and interpreted by S.P. Gordey.

For interpretation and references see GSC 92-35.

GSC 92-34 Biotite
96.7 ± 1.4 Ma

Wt % K= 5.270
Rad. Ar= 2.035 x 10^{-5} cm^3/g
K-Ar 4291 % Atmos. Ar= 9.0

From a porphyritic biotite-hornblende granite.
(105 K/01) From 20.8 km east-southeast of the north end of Blind Lakes, Yukon; 62°14.09'N, 132°5.13'W; UTM zone 8, 651450E, 6903545N; sample GGA-87-15J3. Collected and interpreted by S.P. Gordey.

For interpretation and references see GSC 92-35.

GSC 92-35 Whole rock
58.3 ± 1.3 Ma

Wt % K= 1.213
Rad. Ar= 2.792 x 10^{-6} cm^3/g
K-Ar 4305 % Atmos. Ar= 35.9

From an olivine basalt.
(105 J/04) From 3.6 km south-southwest of the southeast end of Marjorie Lake, Yukon. 62°2.69′N, 131°58.47′W; UTM zone 9, 344470E, 6882543N; sample GGA-87-9E3. Collected and interpreted by S.P. Gordey.

The K-Ar samples GSC 92-32, 33, 34, and 35 were collected during reconnaissance mapping in east-central Yukon (Sheldon Lake (105 J) and Tay River (105 K) map areas) to characterize the ages of plutonism and volcanism. They are the last subset of K-Ar samples of this project to be reported on; for results released earlier see Wood and Armstrong (1982), Jackson et al. (1986), and Hunt and Roddick 1988, GSC 88-41, 42; 1990, GSC 89-98 to 104; 1991, GSC 90-52 to 76; 1992, GSC 91-103, 104.

There are three main igneous suites in the region, the mid-Cretaceous South Fork Volcanics and coeval Selwyn Plutonic Suite (includes Anvil suite of Pigage and Anderson (1985)), and a bimodal volcanic suite of early Tertiary age.

GSC 92-32 is from the South Fork Volcanics, which consist largely of densely welded crystal and crystal lithic tuff preserved within several large calderas (Gordey and Irwin, 1987; Gordey, 1988). The sample comes from an isolated exposure within the wide valley of the lower Ross River. The age of this sample and the distribution of other scattered outcrops suggests that a large South Fork caldera may underlie this poorly exposed area (Gordey, 1988). The age of the sample is consistent with previous K-Ar biotite and hornblende cooling ages from the South Fork, interpreted as eruptive ages, which are generally within the range of 94-102 Ma.

Sample GGA-87-14A3 (AA92-237), from the South Fork Volcanics, the same general area as GSC 92-32 was analyzed by ^{40}Ar-^{39}Ar detailed step-heating. See Roddick et al. (1992) this volume for results and interpretation.

GSC 92-33, a hornblende porphyry, forms an isolated exposure about 40 m in diameter, in the same general region as GSC 92-32 above. It is composed of 8% hornblende (to 7 mm long) and 65% plagioclase (to 1-2 mm long) set in a fine grained matrix. It is of a composition not typical of either the Selwyn Plutonic Suite or the South Fork Volcanics. However, its age and probable intrusive(?) origin indicates it belongs to the former.

GSC 92-34 is from the Marjorie pluton of the Selwyn Plutonic Suite. This body is interpreted to intrude the South Fork Volcanics (Gordey, 1988). However, on the basis of its biotite cooling age, and ages previously reported from the adjacent South Fork (95.5 ± 3.3 Ma, Wood and Armstrong, 1982, sample 33B, p. 313; 97.1 ± 1.4 Ma, Hunt and Roddick, 1991, GSC 90-69) an age difference between the pluton and volcanics is not demonstrable within the limits of analytical uncertainty.

GSC 92-35 is from an isolated exposure of fresh olivine basalt. Its age confirms it is part of an early Tertiary bimodal volcanic suite that may be time equivalent with strike-slip displacement along Tintina Fault (Jackson et al., 1986).

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GSC 92-36 Biotite
96.3 ± 1.0 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.0
K-Ar 4312 % Atmos. Ar= 15.8
(234) Ages of two heating steps + % gas:
96.3 ± 0.8 Ma (39%), 97.0 ± 0.4 Ma (59%).
Preferred age: 98% plateau age of 96.7 ± 1.0 Ma

From a hornblende porphyry dyke.
(105 K/6) On the northern and northeastern wall of the Faro open pit mine, east-central Yukon; 62°23.1'N, 133°29.3'W; UTM zone 8, 578202E, 6917805N; sample 88-413. Collected by L.C. Pigage and interpreted by J.K. Mortensen.

See GSC 92-37 for discussion and interpretation.

GSC 92-37 Biotite
96.3 ± 1.0 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.3
K-Ar 4313 % Atmos. Ar= 8.1
(235) Ages of two heating steps + % gas:
95.5 ± 0.5 Ma (40%), 95.7 ± 0.3 Ma (59%).
Preferred age: 99% plateau age of 95.6 ± 1.0 Ma

From a hornblende porphyry dyke.
(105 K/6) On the northern and northeastern wall of the Faro open pit mine, east-central Yukon; 62°21.7'N, 133°22.6'W; UTM zone 8, 583987E, 6915384N; sample NW-DYKE. Collected by L.C. Pigage and interpreted by J.K. Mortensen.

This sample and GSC 92-36 are from coarse grained hornblende porphyry dykes that are correlated with unit 10E in the Faro area. Although the ages are interpreted as cooling ages, they are likely very close to the actual ages of intrusion. They are consistent with U-Pb and K-Ar ages for other similar dykes and plutons in this area.

GSC 92-38 Biotite
99.0 ± 1.1 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.7
K-Ar 4314 % Atmos. Ar= 11.2
(236) Ages of two heating steps + % gas:
100.7 ± 0.2 Ma (56%), 99.2 ± 0.3 Ma (42%).
Preferred age: 98% plateau age of 100.0 ± 1.0 Ma

From a granitic gneiss.
(105 F/11) East of Canol Road, southwest of the confluence of Ross River and Pony Creek, Yukon; 61°34.1'N, 133°4.7'W; UTM zone 8, 602066E, 6827396N; sample CR-459. Collected by V.L. Hansen and interpreted by J.K. Mortensen.

The sample is from a medium grained granitic gneiss which contains a well-developed mylonitic fabric. The gneiss is crosscut and the mylonitic fabrics are annealed by the Nisutlin Batholith which has given K-Ar biotite ages that range from 84-96 Ma. The age is interpreted to represent cooling after ductile deformation.

See Mortensen and Hansen (1992), this volume, for further discussion.

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GSC 92-39 Muscovite
100.6 ± 1.3 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 8.8
K-Ar 4337 % Atmos. Ar= 16.5
(259) Ages of two heating steps + % gas:
100.0 ± 0.8 Ma (47%), 100.8 ± 0.4 Ma (51%).
Preferred age: 98% plateau age of 100.4 ± 1.1 Ma

From a greisen alteration zone.
(105 K/6) On a north-south striking ridge above 1675 m (5500 ft.) elev., approximately 3 km north of Mount Mye, Yukon; 62°21.8'N, 133°5.8'W; UTM zone 8, 598500E, 6915700N; sample 889004. Collected by S.B. Ballantyne and interpreted by J.K. Mortensen.

See GSC 92-40 for discussion.

GSC 92-40 Muscovite
101.1 ± 1.1 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 8.7
K-Ar 4338 % Atmos. Ar= 10.1
(260) Ages of two heating steps + % gas:
100.3 ± 0.3 Ma (45%), 101.2 ± 0.7 Ma (53%)
Preferred age: 98% plateau age of 100.8 ± 1.1 Ma

From a greisen alteration zone.
(105 K/6) On a north-south striking ridge above 1675 m (5500 ft.) elev., approximately 3 km north of Mount Mye, Yukon; 62°21.8'N, 133°5.8'W; UTM zone 8, 598500E, 6915700N; sample 889006. Collected by S.B. Ballantyne and interpreted by J.K. Mortensen.

This sample and GSC 92-39 are from zones of strong greisen alteration adjacent to Ag-Sn-Mn-Zn bearing veins in the Cody Ridge area (Mortensen and Ballantyne, 1992). These veins cut the Mount Mye phase of the Anvil Batholith, which gives U-Pb zircon and monazite ages ranging from 109-100 Ma (Mortensen and Gordey, unpublished data). The two ages provide a precise date for the veining event. See Mortensen and Hansen (1992), for further discussion.

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GSC 92-41 Whole Rock
73.4 ± 1.3 Ma

Wt % K= 1.454
Rad. Ar= 4.231 x 10^{-6} cm^3/g
K-Ar 4096 % Atmos. Ar= 66.4

From a basalt.
(105 L/3) Natural outcrop 100 m north of Robert Campbell Highway, Yukon; 62°4′48.6"N, 135°28′22.1"W; UTM zone 8V 475300E, 6883000N; sample 166876. Collected and interpreted by Lionel E. Jackson, Jr.

Basalt, medium crystalline in texture, primarily composed of plagioclase (An 50-60%) and augite.

Sampled in order to determine if previously unmapped basalts are part of an Eocene volcanic suite or part of the ca. 65-70 Ma Carmacks Group (Grond et al. 1984). The date is compatible with the latter interpretation.

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GSC 92-42 Hornblende
67.7 ± 3.3 Ma

Wt % K= 0.940
Rad. Ar= 2.522x10^{-6} cm^3/g
K-Ar 4112 % Atmos. Ar= 8.6

From quartz-plagioclase porphyry.
(105 L/3) From a grassy slope 1700 m north of crossing of Bearfeed Creek by Robert Campbell Highway, Yukon; 62°12′04"N, 135°04′54"W; UTM zone 8V, 495750 E, 6896400 N; sample J-12688-R-6. Collected and interpreted by Lionel E. Jackson, Jr.

Isolated outcrop sampled in order to determined if Tertiary felsic volcanics are present in this area. The age is compatible with a late Cretaceous age of intrusion of the nearby Glenlyon batholith.

GSC 92-43 Whole rock
1.28 ± 0.03 Ma

Wt % K= 1.029
Rad. Ar= 5.106x10^{-8} cm^3/g
K-Ar 4168 % Atmos. Ar= 74.2

From olivine basalt.
(115 I/14) From base of cliff along northeast side of Yukon River, 5 km downstream from Fort Selkirk, Yukon; 62°48′52" N, 137°27′16"W, UTM zone 8V, 374900E, 6966900N; sample 010789-R1. Collected and interpreted by Lionel E. Jackson, Jr.

The sample was taken from the base of a basalt flow which rests upon a 1 m thick bed of lapilli. The unit was sampled in order to place a minimum age on underlying glacial deposits and to directly date a glaciation which was contemporaneous with the eruption of the basalt (Jackson, 1989; Jackson et al., 1990) This unit has been previously dated at 1.08 ± 0.05 Ma (Naeser et al., 1982).

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GSC 92-44 Whole Rock
2.08 ± 0.05 Ma

Wt % K= 1.443
Rad. Ar= 1.166 x 10^{-7} cm^3/g
K-Ar 4169 % Atmos. Ar= 83.5

From a vesicular basalt.
(115 I/14) From summit of unnamed volcanic ediface informally called Ne Ch'e Ddhawa (Jackson, 1989), 8 km southeast from Fort Selkirk, Yukon; 62°44′58"N, 137°15′40"W; UTM zone 8V, 383600E, 6959200N; sample 280689-R5. Collected and interpreted by Lionel E. Jackson, Jr.

Highly fractured dark grey vesicular basalt which contains lherzolite nodules and exotic pebbles.
See GSC 92-46 for interpretation and references.

GSC 92-45 Whole Rock
2.36 ± 0.09 Ma

Wt % K= 1.150
Rad. Ar= 1.057x10^{-7} cm^3/g
K-Ar 4170 % Atmos. Ar= 89.4

From amygdoloidal dark grey basalt containing lherzolite nodules.
(115 I/14) From a cliff along the west base of unnamed volcanic ediface informally called Ne Ch'e Ddhawa (Jackson, 1989), 7 km southeast from Fort Selkirk, Yukon; 62°47′29"N, 137°17′14"W; UTM zone 8V, 383200E, 6960500N; sample 290689-R2. Collected and interpreted by Lionel E. Jackson, Jr.

See GSC 92-46 for interpretation and references.

GSC 92-46 Whole Rock
3.92 ± 0.11 Ma

Wt % K= 1.217
Rad. Ar= 1.857x10^{-7} cm^3/g
K-Ar 4171 % Atmos. Ar= 79.0

From fine grained dark grey basalt.
(115 I/11) From a cliff along the north base of unnamed volcanic ediface informally called Ne Ch'e Ddhawa (Jackson, 1989), 8 km southeast from Fort Selkirk, Yukon; 62°44′47"N, 62°44′47"W; UTM zone 8V, 383600E, 6959200N; sample 300689-R2. Collected and interpreted by Lionel E. Jackson, Jr.

Paleomagnetic evidence indicates that GSC 92-43, 44, 45, and 46 are the same age. The last three were erupted beneath glacial ice. Since GSC 92-43 has been dated twice in the 1-1.3 Ma range, and the ages for GSC 92-44, 45, and 46 are too old (Tertiary) to be compatible with known glaciations in this region, GSC 92-44, 45, and 46 appear to be erroneously old. The error is likely the result of excess mantle-derived argon which has been frequently associated with mantle-derived xenoliths and xenocrysts which are abundant in these rocks (Dalrymple and Lanphere, 1969, p. 143-144).

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GSC 92-47 Biotite
55.3 ± 0.7 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.7
K-Ar 4300 % Atmos. Ar= 9.8
(230) Ages of two heating steps + % gas:
55.4 ± 0.2 Ma (45%), 55.2 ± 0.6 Ma (54%).
Preferred age: 99% plateau age of 55.3 ± 0.7 Ma

From a quartz-biotite schist.
(115 A/7) On ridge crest 16.7 km S76°E from the northeast corner of Dezadeash Lake, Yukon; 60°29.7′N, 136°35.5′W; UTM Zone 8, 412554E, 6707447N; sample PE-89-101B; Collected by P. Erdmer and interpreted by J.K. Mortensen.

The sample is from a sillimanite grade pelitic schist that has been correlated with either the Nisling Assemblage or the Kluane Schist. The age is interpreted to date cooling after the high grade metamorphism, and is consistent with other K-Ar and Ar-Ar cooling ages in the area. See Mortensen and Erdmer (1992), for further discussion.

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GSC 92-48 Biotite
39.1 ± 1.0 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 7.0
K-Ar 4301 % Atmos. Ar= 19.6
(229) Ages of two heating steps + % gas:
40.8 ± 1.6 Ma (44%), 38.2 ± 0.7 Ma (54%)
Preferred age: 98% plateau age of 39.4 ± 0.9 Ma

From a quartz-biotite-muscovite schist.
(115 G/1) On the east shore of Kluane Lake, 3.8 km south of the mouth of Cultus Creek, Yukon; 61°7.2′N, 138°25.9′W; UTM Zone 7, 638329E, 6779630N; sample PE-89-85. Collected by P. Erdmer and interpreted by J.K. Mortensen.

See GSC 92-49 for discussion.

GSC 92-49 Muscovite
43.1 ± 1.9 Ma integrated ^{40}Ar-^{39}Ar

Wt % K= 3.1
K-Ar 4324 % Atmos. Ar= 59.0
(257) Ages of two heating steps + % gas:
43.3 ± 3.2 Ma (57%), 43.5 ± 1.9 Ma (42%).
Preferred age: 99% plateau age of 43.4 ± 1.9 Ma

From a quartz-muscovite schist.
(115 G/1) On the east shore of Kluane Lake, 3 km northwest of the mouth of Cultus Creek, Yukon; 61°10.8′N, 138°27.8′W; UTM Zone 8, 636400E, 6785300N; sample PE-89-83. Collected by P. Erdmer and interpreted by J.K. Mortensen.

This sample and GSC 92-48 are from quartz-muscovite and quartz-muscovite-biotite schist units within the Kluane Schist. Metamorphism in this area was at lower greenschist facies. The ages are interpreted to date post-metamorphic cooling, and are consistent with other K-Ar and Ar-Ar cooling ages in the area.

See Mortensen and Hansen (1992), for further discussion.

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GSC 92-50 Whole Rock
3.05 ± 0.22 Ma

Wt % K= 1.691
Rad. Ar= 2.004 x 10^{-7} cm^3/g
K-Ar 4154 % Atmos. Ar= 94.7

From a olivine basalt.
(116 C/7) From a small borrow pit on a placer mining access road along a ridge crest on the north side of the Fortymile River, 2.8 km southwest of the Fortymile River bridge, Yukon; 64°23.6′N, 140°38.8′W; UTM zone 7, 517065E, 7140640N; sample MLB-89-362. Collected and interpreted by J.K. Mortensen.

The sample is from a small columnar-jointed olivine basalt flow that is correlated with the Selkirk Lavas in western and southwestern Yukon. At this locality the rocks contain abundant mantle xenoliths and xenocrysts, as well as rare xenoliths of underlying metamorphic country rocks. Ages obtained for Selkirk Lavas in this area range from 0.5 to 20 Ma. The age of 3.05 Ma for this sample is interpreted as the age of eruption.

DISTRICT OF FRANKLIN
(GSC 92-51)

GSC 92-51 Hornblende
1831 ± 20 Ma

Wt % K= 1.266
Rad. Ar= 1.561 x 10^{-5} cm^3/g
K-Ar 4318 % Atmos. Ar= 1.8

From a hornblende-orthopyroxene granodiorite gneiss.
(57 G) From 3 km west of Wrottesley River, 11 km from its mouth, Boothia Peninsula, N.W.T.; 71°11.5′N, 95°36′W; UTM zone 15 406464E, 7900611N; sample FS-87-10. Collected and interpreted by T. Frisch.

The host rock of the hornblende is typical of the orthopyroxene-bearing granitoid orthogneiss that forms much of the crystalline core of the Boothia Uplift on Boothia Peninsula and Somerset Island. This particular rock sample is hornblende-orthopyroxene granodiorite gneiss, weakly foliated, more or less equigranular and medium grained. Fresh, anhedral, weakly pleochroic orthopyroxene (0.7 mm), olive-green hornblende and a little brown biotite lie in a matrix of plagioclase (An_{27}; 1 mm), subordinate K-feldspar (0.6-1 mm), and quartz (0.2-1 mm). The hornblende, much of which occurs in discrete grains spatially removed from orthopyroxene, is regarded as a primary metamorphic mineral, genetically unrelated to orthopyroxene.

The rock has given a Sm-Nd model age of 3.0 Ga (unpublished work of E. Hegner) and zircons from it have yielded $^{207}Pb/^{206}Pb$ ages ranging from 2665 to 2262 Ma. Plotted on a concordia diagram, the zircons roughly define a discordia with upper and lower intercepts at 2848 +222/-142 Ma and 2267 +98/-156 Ma, respectively. Recent U-Pb dating of a variety of crystalline rocks from the Boothia Uplift indicates, besides a late Archean age for some of the gneiss protoliths, a granulite metamorphic and magmatic event at ca. 1.9 Ga and suggests another thermal event at ca. 2.3 Ga.

The hornblende age of 1831 ± 20 Ma reported here is interpreted as a cooling age related to an earlier Proterozoic thermal event (2.3 or 1.9 Ga).

DISTRICT OF KEEWATIN
(GSC 92-52 to GSC 92-55)

GSC 92-52 Illite
1335 ± 16 Ma

Wt % K= 8.207
Rad. Ar= 6.307 x 10^{-4} cm^3/g
K-Ar 4275 % Atmos. Ar= 0.1

From a intensely hydrothermally altered unfoliated granite.
(66 A/5) From diamond drill hole SW-16, sampled at a depth of 618 feet, N.W.T.; 64°19.5′N, 97°53′W; UTM zone 14, 553983E, 7133509N; sample SW-16-618. Collected and interpreted by A.R. Miller.

For interpretation and references see GSC 92-55.

GSC 92-53 Illite (hydro-muscovite?)
1347 ± 13 Ma

Wt % K= 4.816
Rad. Ar= 3.747 x 10^{-4} cm^3/g
K-Ar 4276 % Atmos. Ar= 4.2

From an intensely hydrothermally altered fine-grained unfoliated granite.
(66 A/5) From diamond drill hole SW-18, sampled at a depth of 468 feet, N.W.T.; 64°19.5′N, 97°53′W; UTM zone 14, 553983E, 7133509N; sample SW-18-468. Collected and interpreted by A.R. Miller.

For interpretation and references see GSC 92-55.

GSC 92-54 Illite
1477 ± 22 Ma

Wt % K= 7.796
Rad. Ar= 6.933 x 10^{-4} cm^3/g
K-Ar 4277 % Atmos. Ar= 0.2

From a intensely hydrothermally altered pelitic metasediment interbedded arenaceous laminae.
(66 A/5) From diamond drill hole SW-16, sampled at a depth of 177 feet, N.W.T.; 64°19.5'N, 97°53'W; UTM zone 14, 553983E, 7133509N; sample SW-16-177. Collected and interpreted by A.R. Miller.

For interpretation and references see GSC 92-55.

GSC 92-55 Illite
1344 ± 17 Ma

Wt % K= 7.120
Rad. Ar= 5.526 x 10^{-4} cm^3/g
K-Ar 4278 % Atmos. Ar= 0.3

From an intensely hydrothermally altered metapelite.
(66 A/5) From diamond drill hole SW-15, sampled at a depth of 444 feet, N.W.T.; 64°19.5'N, 97°53'W; UTM zone 14, 553983E, 7133509N; sample SW-15-444. Collected and interpreted by A.R. Miller.

This sample and GSC 92-52, 53, and 54 were collected from several diamond drill holes that were drilled to define uranium mineralization in the Southwest Grid area, Schultz Lake map area, District of Keewatin.

The most prominent metamorphosed supracrustal unit in the Schultz Lake map area is an east- to northeast-trending late Archean 2.78 Ga greenstone belt (Henderson et al., 1991) which has been informally called the Woodburn Group and Ketyet Group. In the southwestern half of the belt, which includes the study area, the greenstone belt is sediment-dominated. To the northeast and north in the Tehek Lake area, the greenstone belt is volcanic-dominated by komatiitic to basaltic volcanic rocks and interlayered greywacke-magnetite iron formation (Ashton, 1981, 1982; Henderson et al., 1991; Annesley et al., 1991).

Arenaceous and pelitic metasedimentary rocks are the lowest stratigraphic unit in the central and southwestern portion of the belt and are presumed to be in tectonic contact with layered granitic and tonalitic gneisses exposed along the southern margin of the greenstone belt. This metasedimentary unit outcrops east of 'Long Lake' and extends southwestwards into the study area and south of Judge Sissons Lake. It comprises a garnet-bearing quartzite overlain by psammitic and muscovite schists.

From drill hole intersections in the study area, garnet +muscovite and muscovite schists are in tectonic contact with granitic gneiss. These peltitic schists grade upward, uphole, into a thick metagreywacke-meta-argillite with minor interbedded oxide iron-formation that dominates the lithology in the southeastern portion of the belt. The Archean gneiss and metasedimentary units have been intruded by unfoliated ~1.9 Ga fluorite-bearing porphyritic granite and in turn by feldspar-bearing syenite dykes belonging to the 1.85 Ga alkaline magmatism in the Baker Lake Group (Gall et al., 1992).

In the study area, Thelon Formation sandstone was not intersected in drill core. However the first 100-150 metres of each sampled drill hole is intensely oxidized. These oxidized Archean metasedimentary rocks record an interval of paleoweathering prior to ~1.72 Ga Thelon Formation sedimentation. All of the above lithologies and the paleoweathered metasedimentary rocks have been overprinted by variable argillic and chloritic alteration, typical of unconformity-related uranium mineralization.

Even though hydrothermal alteration overprinted paleoweathered rocks, the intensity of argillic alteration associated with uranium mineralization and its relationship with hydrothermal chlorite is best displayed in rocks below the paleoweathered zone. Visual estimates of clay alteration made during core logging were based on the preservation of original mineralogy, texture, structure and whiteness. The four samples, two metapelitic rocks and two unfoliated granite, were chosen because the protolith was totally converted to a very fine-grained structureless aggregate of white clay. X-ray diffractometer identification of crushed whole rock powders revealed major illite or hydromuscovite and only trace quartz. Reduced quartz contents coupled with intensity argillic alteration suggests these four samples represent desilicified rocks that commonly accompany uranium mineralization. Smectite, kaolinite and chlorite can accompany the above two clay phases but were not identified in the four samples.

Three of the four ages, 1344 ± 17 Ma, 1347 ± 13 Ma and 1335 ± 16 Ma (GSC 92-55, 53, 52) are interpreted to record the time of intense argillic alteration of the Archean metasedimentary rocks and early Proterozoic granite sheets in the Southwest Grid area. Within the 2 sigma error limits, these ages are concordant with similiar intensely argillitized and desilicified metagreywacke and fluorite-bearing granite and related porphyries in the End Grid area (GSC 91-120, 121, Hunt and Roddick, 1992) and in the Kiggavik deposit, Main Zone (GSC 88-44 to 50, Hunt and Roddick, 1988). Argillic alteration is presumed to be synchronous with and may have outlasted initial uranium mineralization at 1400 Ma (Fuchs and Hilger, 1989).

South of Schultz and the eastern end of Aberdeen lakes and north of Prince Mary and Pitz lakes, the exposed Archean-early Proterozoic basement terrane contains isolated outliers of paleoweathered basement with or without overlying Thelon Formation siliciclastics. These features imply that the basement-hosted unconformity-related uranium mineralization in the central portion of the western half of the Schultz Lake map area was capped by Thelon Formation sediments and may have had extensions into the overlying sediments similiar to deposits in the Athabasca Basin. Uranium mineralization in the Schultz Lake map area is spatially associated with regional east- to northeast-trending faults with silicified breccia zones. These fault zones are the loci for intense hematitization, presumed to be related to processes related to paleoweathering and fluid migration during basin diagenesis. However these same structures can be bleached and overprinted by argillic alteration. In summary argillic alteration and accompanying uranium mineralization was focused along reactivated pre-Thelon Formation regional fault zones. In the Southwest Grid area illite ages of ~1345 Ma represent the final stage of initial argillic alteration, cooling below the closure temperature of fine grained illite.

The age of 1477 ± 22 Ma (GSC 92-54) from a metapelite is older than the other three samples and is interpreted as a mixed age derived through the recrystallization of older metamorphic muscovite and neoformation of muscovite associated with hydrothermal alteration.

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DISTRICT OF MACKENZIE
(GSC 92-56 to GSC 92-61)

Age determinations of the Indin Structural Basin

Six whole rock samples (GSC 92-56, 57, 58, 59, 60, 61; (Table 1)) from the Indin structural basin were analyzed to follow up a report of excess ^{40}Ar in the Yellowknife Supergroup meta-turbidites (Wanless et al., 1978). The samples were collected from a metagreywacke near the north end of Indin Lake, District of Mackenzie, N.W.T., NTS map sheet 86 B, 64°22′N, 114°55′W (Fig. 1, sample site "A"). See Table 1 for results. The original sample site occurs near the cordierite-staurolite isograd in the western part of the basin (Fig. 1, sample site "C") where muscovite and cordierite bearing schists were doubly folded during the Archean.

Table 1. Summary of age determinations from the Indin Structural basin with location shown in Fig. 1

Sample site A - this study						
GSC #	Sample #	K-Ar #	Age (Ma)	Wt % K	% Atmos. Ar	Rad. Ar $cm^3/g \times 10^{-5}$
92-56	B-08	2870	2523 ± 62	1.98	0.5	7.164
92-57	B-28C	2871	2069 ± 61	1.57	0.2	5.047
92-58	B-5A	2872	2365 ± 54	3.02	0.3	6.367
92-59	B-6	2873	2640 ± 58	3.09	0.4	7.801
92-60	B-33	2874	2175 ± 53	2.78	0.1	5.493
92-61	T-719D	2875	2529 ± 71	1.45	0.2	7.196

Sample site B - previous work		Age (Ma)
FYR72-366	Rb-Sr[3], wr, granodiorite	1887 ± 193
76-180	K-Ar, hb, granodiorite	1922 ± 120
76-182	K-Ar, hb, basalt incl.	2523 ± 62
Sample site C and within Indin basin - previous work		
76-176	K-Ar[1], wr - schist	3056 ± 69
76-176	K-Ar[1], wr - schist	3108 ± 69
76-176	Rb-Sr[1], wr- ms	2529 ± 50
76-177	K-Ar, ms, schist	2003 ± 46
76-178	K-Ar - diabase	2807 ± 148
76-179	K-Ar - diabase	1853 ± 52
60-48	K-Ar[2], bt, cd, schist	2167

wr = whole rock, ms = muscovite, hb = hornblende,
cd = cordierite, bt = biotite
References: [1]Wanless et al., 1978, [2]Stockwell, 1961, [3]Frith, et al 1977

The K-Ar data, sampled from other localities within the basin have ages ranging from 2069 to 2640 Ma, similar to, but not as great as the age or range of previously analyzed samples (1853 to 3108 Ma, sample site B & C, Table 1). The age of the metagreywacke schists are constrained stratigraphically by the age of the cross-cutting Indin Diabase swarm which cuts an 1887 ± 193 Ma granodiorite (whole rock, Rb-Sr, FYR72-366, Table 1) and by the age of the Yellowknife Supergroup, likely of ca. 2.67 Ga (Frith and Loveridge, 1982).

The thermal closure age was dated regionally by U-Pb methods using monazite at 2596 ± 3 Ma (Frith, 1986). Dates older than this are suspect, whereas those younger than ca. 2000 Ma may have been reset by a Proterozoic thermal-deformational event of comparable age (Frith, 1986).

Muscovite from a schist derived from the Yellowknife Supergroup turbiditic greywacke-mudstones schist (Locality "C", Fig. 1, GSC 76-177, Table 1) formed as an axial planar foliation mineral during Archean regional metamorphism and deformation. The schist is made up of muscovite, biotite and sillimanite oriented parallel to a second foliation plane (Frith, 1978). Finer grained plagioclase, quartz, chlorite and seracitic muscovite make up the groundmass for snow-balled porphyroblasts of pinitized cordierite and andalusite which overgrow the first foliation and the bedding. A second poikilitic andalusite has overgrown the earlier andalusite, probably at the expense of muscovite and cordierite. Metamorphic mineral assemblages suggest pressures of 4 Kb and temperatures of 540°C (Anderson et al., 1977).

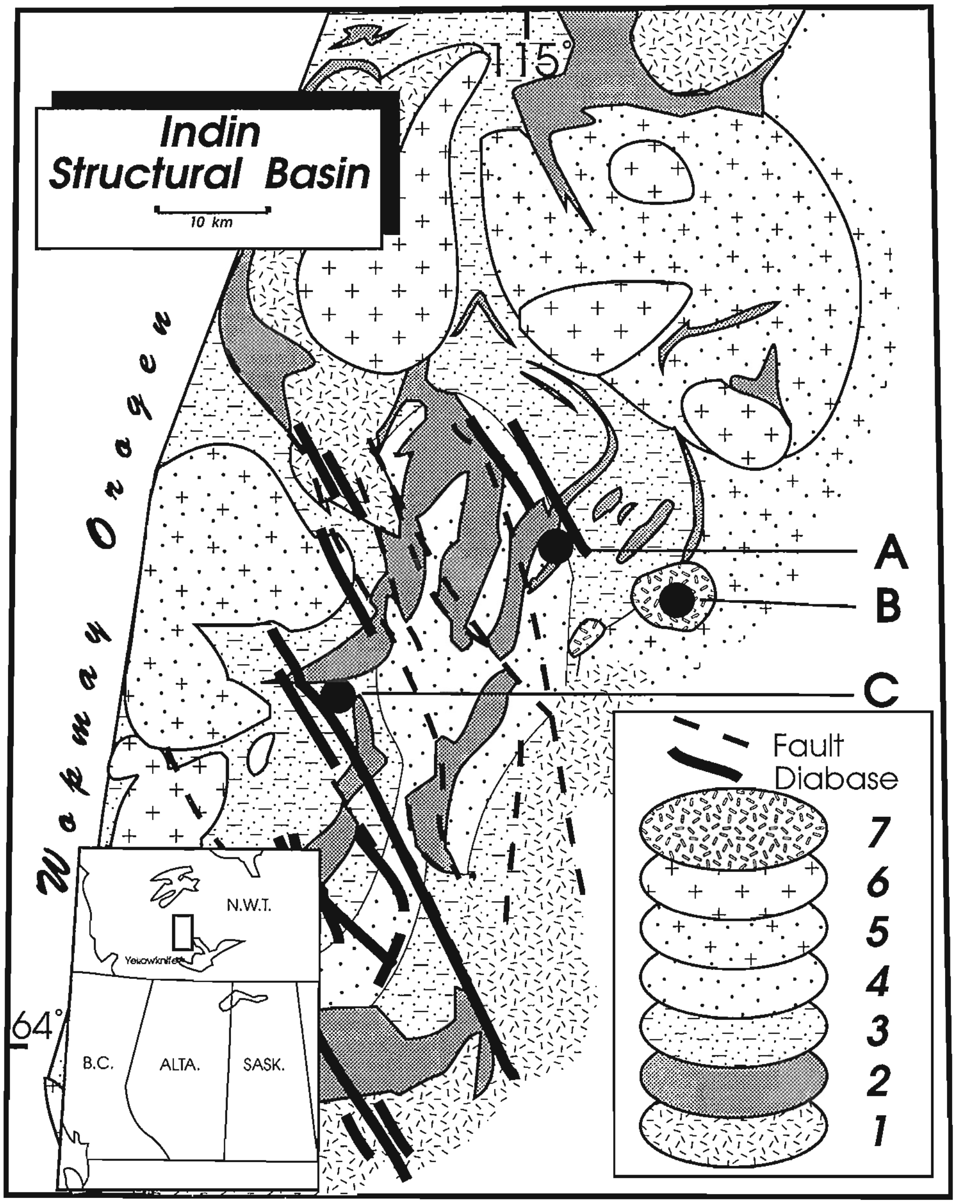

Figure 1. Sketch map of the Indin Structural Basin, western Slave Province showing: (1) the extent of the granitoid basement; (2) the Yellowknife Supergroup volcanic belts; (3) the Yellowknife Supergroup turbiditic greywacke-mudstones (YKS) above the cordierite-staurolite isograd and (4) below the isograd; (5) migmatite derived from 2 and 3; (6) plutonic granites and granodiorites; (7) Proterozoic granodiorite. Site A = this study; B = Frith et al., 1977; C = Wanless et al., 1978.

INTERPRETATION

All samples are located in the down-warped part of a structural basin formed by brittle failure along left-lateral, oblique faults. The faults are probably contemporaneous with the intrusion of the northwest trending Indin Diabase, as dykes both intrude the faults and are locally faulted by them. The data compliment earlier studies investigating the occurence of excess ^{40}Ar in rocks from the basin (Wanless et al., 1978), but suggest that the occurence, whereas not as extreme as previously realized, is more variable and more extensive. The following conclusions are drawn:

(1) post-Archean thickening and/or heating of the basin caused outgassing of ^{40}Ar which accumulated in diabase dykes and host rocks with minerals such as cordierite (Seideman, 1976) that favour ^{40}Ar entrapment;

(2) the K-Ar muscovite (GSC 76-177, Table 1) at 2003 Ma age represents the best estimate of the thermal event, as this mineral is generally insensitive to incorporation of excess ^{40}Ar. The age likely records a Proterozoic thermal blocking temperature event, caused by intrusion of contemporaneous diabase dyking;

(3) all K-Ar whole rock and hornblende age data from the region are suspect as the dated rocks may either contain excess ^{40}Ar or have only been partially reset (see GSC 76-182, Table 1);

(4) the Rb-Sr system in muscovite, with a higher closure temperature than the K-Ar system has permitted the Rb-Sr muscovite-whole rock system to "see through" the Proterozoic thermal event and record the age of development of the second foliation which approximates the peak of regional thermal metamorphism in the region (GSC 76-176, Table 1).

Finally, on a more speculative note, the tectonic conditions that gave rise to the faulting, dyking and thermal overprinting likely have their root cause in the thickening, rifting at ca. 2.2 Ga. and final compression of the Slave Province margin and westward obducted Proterozoic supracrustals at ca. 2.0 Ga. (Frith, 1991).

These samples were collected and interpreted by R.A. Frith.

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MANITOBA
(GSC 92-62 to GSC 92-64)

GSC 92-62 Biotite
1763 ± 14 Ma

Wt % K= 7.691
Rad. Ar= 8.938 x 10^{-4} cm^3/g
K-Ar 4283 % Atmos. Ar= 0.7

From a quartz eye porphyritic, hornblende-biotite tonalite, Elbow Lake pluton.

(63 K/15) Collected 460 m south of the central bay on the south side of Elbow Lake, Manitoba; 54°48′19.5"N, 100°50′52.4"W; UTM zone 14, 381225E, 6074488N; sample

WXMS-11. Collected and interpreted by J.B. Whalen.

This sample is from a slightly foliated, coarse grained, quartz eye porphyritic tonalite at the the southern end of Elbow Lake. This same sample gave a U-Pb zircon age of 1869 +20/-7 Ma. This biotite K-Ar age is interpreted as the age of a metamorphic overprinting rather than a simple cooling age. This interpretation is supported by the similar K-Ar ages obtained from Big Rat Lake pluton on the west side of Elbow Lake (GSC 92-63, 64).

GSC 92-63 Hornblende
1757 ± 21 Ma

Wt % K= 0.476
Rad. Ar= 5.504 x 10^{-5} cm^3/g
K-Ar 4335 % Atmos. Ar= 2.1

From a foliated hornblende-biotite granodiorite, Big Rat Lake pluton.
(63 K/15) From the west shore of Elbow Lake, 1.2 km west of the north end of McDougalls Point, Manitoba; 54°49′39.5"N, 100°54′35.6"W; UTM zone 14 377307E, 6077068N; sample WXMS-12. Collected and interpreted by J.B. Whalen.

See GSC 92-64 for discussion and co-existing biotite.

GSC 92-64 Biotite
1768 ± 22 Ma

Wt % K= 7.871
Rad. Ar= 9.19 x 10^{-4} cm^3/g
K-Ar 4336 % Atmos. Ar= 0.4

From a foliated hornblende-biotite granodiorite, Big Rat Lake pluton.
(63 K/15) From the west shore of Elbow Lake, 1.2 km west of the north end of McDougalls Point, Manitoba; 54°49′39.5"N, 100°54′35.6"W; UTM zone 14 377307E, 6077068N; sample WXMS-12. Collected and interpreted by J.B. Whalen.

This sample is from a moderately foliated, coarse grained hornblende-biotite granodiorite at the eastern margin of the Big Rat Lake pluton on the southwest side of Elbow Lake. This same sample gave a U-Pb zircon age of 1845 ± 3 Ma.

Similar concordant K-Ar biotite and hornblende ages obtained from this sample and GSC 92-62 suggests that these K-Ar ages record a metamorphic overprinting event rather than simple cooling ages.

QUEBEC
(GSC 92-65 to GSC 92-70)

GSC 92-65 Muscovite
1508 ± 15 Ma

Wt % K= 8.709
Rad. Ar= 7.981 x 10^{-4} cm^3/g
K-Ar 4269 % Atmos. Ar= 2.2

From a muscovite-bearing pegmatite.
(32 H /13) 400 m east of the Grenville Front, 4 km south-southeast of Lac Houde, east of Chibougamau, Quebec; 49°52′25"N, 73°55′00"W; UTM zone 18, 578100E, 5524900N; sample CMA-PKPG-89. Collected and interpreted by A. Ciesielski.

The sample is from a muscovite-rich, coarse grained pegmatite showing narrow zones of mylonitic deformation. Similar pegmatite east of the Grenville Front a few kilometres to the north contains coarse grained magnetite and muscovite. The 1508 Ma age represents a thermal age probably reset during the Grenvillian Orogeny. The pegmatite bodies are distributed along the Grenville Front and likely intruded at depth long before the uplift of the parautochthon (Ciesielski, 1988, in press). See GSC 92-67 for references.

GSC 92-66 Hornblende
1377 ± 15 Ma

Wt % K= 0.311
Rad. Ar= 2.497 x 10^{-5} cm^3/g
K-Ar 4264 % Atmos. Ar= 2.3

(32 G/9) From a garnet-bearing meta-andesitic rock.
200 m south of Lac Dollier, east of Chibougamau, Quebec; 49°42′40"N, 74°05′55"W; UTM zone 18, 5408100N, 566100E; sample CMA-99-84. Collected and interpreted by A. Ciesielski.

The sample is from a folded and lineated, garnet-bearing outcrop of basaltic and andesitic composition. The sample belongs to a middle amphibolite grade metamorphic equivalent of the Blondeau Formation, the upper part of the Roy Group east of Chibougamau (2730-2717 Ma; Daigneault, 1986). Comparable K-Ar and Ar-Ar ages on amphiboles were obtained southwest and north of the sample site; ages much older than ~1000 Ma are due to excess argon in hornblendes (Baker, 1980). Ages of ca. 1000 Ma represent a superimposed contact metamorphism induced in the Superior Province by the uplift of the Archean Parautochthon along the Grenville Front during the Grenvillian Orogeny ca. 1000 Ma (Ciesielski, 1988, in press). See GSC 92-67 for references.

GSC 92-67 Hornblende
994 ± 12 Ma

Wt % K= 0.386
Rad. Ar= 1.989 x 10^{-5} cm^3/g
K-Ar 426 % Atmos. Ar= 4.0

(32 G/9) From a garnet-bearing amphibolite.
150 m north of the Grenville Front, 3 km south of Lac Dollier, east of Chibougamau, Quebec; 49°40′30"N, 74°06′05"W; UTM zone 18, 5504750N, 565300E; sample OE-110-84. Collected by E. Ouellet and interpreted by A. Ciesielski.

The sample is from an homogeneous part of a foliated and lineated volcanogenic amphibolite showing pillow structures and layered zones of intermediate composition. The sample belongs to an amphibolite grade metamorphic equivalent of the Obatogamau Formation (2750-2730 Ma, J.K. Mortensen, pers. comm., 1991) mostly composed of basalts and crosscut by felsic dykes in the immediate region (Daigneault, 1986). Similar K-Ar and Ar-Ar ages on amphiboles were obtained a few kilometres west of the sample site (Baker, 1980). They are believed to represent cooling ages of a superimposed contact metamorphism induced in the Superior Province by the uplift of the Archean Parautochthon along the Grenville Front during the Grenvillian Orogeny ca. 1000 Ma (Ciesielski, 1988, 1990, in press).

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GSC 92-68 Hornblende
1036 ± 19 Ma

Wt % K= 0.322
Rad. Ar= 1.751 x 10^{-5} cm^3/g
K-Ar 4265 % Atmos. Ar= 1.2

(32 G/9) From a hornblende metabasite.
1 km east of Lac Dollier, east of Chibougamau, Quebec; 49°43′40"N, 74°03′00"W; UTM zone 18, 5510200N, 569000E; sample CMA-62B-84. Collected and interpreted by A. Ciesielski.

The sample is from a foliated and folded, garnet-bearing felsic portion of an exposure of generally basic to ultrabasic composition. The sample belongs to an amphibolite grade metamorphic equivalent of the Cummings ultrabasic complex (younger than 2730 Ma) intruding the Blondeau Formation, the upper part of the Roy Group (Daigneault, 1986). Comparable K-Ar and Ar-Ar ages on amphiboles were obtained 10 km southwest of the sample site (Baker, 1980). The 1036 Ma age represents a superimposed contact metamorphism induced in the Superior Province by the uplift of the Archean Parautochthon along the Grenville Front during the Grenvillian Orogeny ca. 1000 Ma (Ciesielski, 1988, 1990, in press). See GSC 92-67 for references.

GSC 92-69 Hornblende
1164 ± 13 Ma

Wt % K= 0.443
Rad. Ar= 2.816 x 10^{-5} cm^3/g
K-Ar 4266 % Atmos. Ar= 2.9

From a metasedimentary rock.
(32 G/8) Along the shore of Lac Mannard, 50 km south of Chibougamau, Quebec; 49°27′45"N, 74°21′30"W; UTM zone 18, 5478800N, 546450E; sample CMA-KK-05-85. Collected and interpreted by A. Ciesielski.

The sample is from a homogeneous but layered garnet-hornblende bearing metasedimentary rock of intermediate composition affected mainly by Kenoran deformation and younger superimposed amphibolite grade metamorphism. The sample belongs to the Caopatina Formation (2750-2730 Ma), overlying the Obatogamau Formation, which is the lower part of the Roy Group in the Chibougamau region. Comparable K-Ar and Ar-Ar ages on amphiboles were obtained in the Dollier twp. 25 km to the north; ages much older than ~1000 Ma are interpreted as being due to excess argon in hornblende (Baker, 1980). Ages ca. 1000 Ma represent a superimposed contact metamorphism induced in the Superior Province by the uplift of the Archean Parautochthon along the Grenville Front during the Grenvillian Orogeny ca. 1000 Ma (Ciesielski, 1988, in press). See GSC 92-67 for references.

GSC 92-70 Hornblende
1142 ± 15 Ma

Wt % K= 0.286
Rad. Ar= 1.771 x 10^{-5} cm^3/g
K-Ar 4268 % Atmos. Ar= 4.6

From a metabasalt.
(32 G/8) 2.5 km north of Lac Mannard, 50 km south of Chibougamau, Quebec; 49°29′40"N, 74°21′40"W; UTM zone 18, 5482200N, 546250E; sample OE-163-85. Collected by E. Ouellet and interpreted by A. Ciesielski.

The sample is from an homogeneous, garnet-bearing, metabasaltic outcrop affected by a Kenoran deformation and a younger superimposed greenschist grade metamorphism. Garnets are concentrated in chilled margins and hornblendes in the cores of pillows (Ouellet, 1988). The sample belongs to the Obatogamau Formation (2750-2730 Ma, J.K. Mortensen, pers. comm., 1991), the lower part of the Roy Group in the Chibougamau region. Comparable K-Ar and Ar-Ar ages on amphiboles were obtained in the Dollier twp. 25 km to the north; ages much older than ~1000 Ma are due to excess argon in hornblende (Baker, 1980). Ages ca. 1000 Ma represent a superimposed contact metamorphism induced in the Superior Province by the uplift of the Archean Parautochthon along the Grenville Front during the Grenvillian Orogeny ca. 1000 Ma (Ciesielski, 1988, in press).

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NOVA SCOTIA
(GSC 92-71 to GSC 92-83)

Davis Lake Pluton, Southwest Nova Scotia

The biotite and muscovite mineral separates (GSC 92-71 to 83, Table 2 and 3) were sampled from igneous granitic rocks comprising the Davis Lake Pluton (DLP) located in southwestern Nova Scotia. The samples are from the parental biotite-monzogranite of the pluton, hydrothermally altered phases of it, and from a small magmatic albitite intrusion in the centre of the pluton.

Davis Lake Pluton forms a southern extension of the much larger South Mountain Batholith (SMB) of Nova Scotia. Both the batholith and the pluton are intruded into the Meguma Terrane which is composed largely of metamorphosed sandstones, mudstones and siltstones. The Davis Lake Pluton forms a finger-like intrusion averaging 6 km in width and 25 km in length extending from the southern end of the South Mountain Batholith.

Table 2. K-Ar age dates for granitic rocks of the Davis Lake Pluton, southwest Nova Scotia

GSC #	K-Ar #	Sample	Rock* Type	Dist. from Carapace	Mineral	Age (Ma)	Wt % K	Rad. Ar (cm^3/g)	% Atmos. Ar
GSC 92-71	4192	1074	GRMG	0	Musc	294 ± 6	8.595	1.069×10^{-4}	2.9
GSC 92-72	4193	1089	GRMG	0	Musc	301 ± 8	8.192	1.045×10^{-4}	2.0
GSC 92-73	4204	1031	MUMG	0	Musc	299 ± 5	8.527	1.077×10^{-4}	3.0
GSC 92-74	4205	1083	MUMG	0	Musc	296 ± 5	7.950	9.947×10^{-5}	2.5
GSC 92-75	4200	1254	BMMG	1	Biot	271 ± 6	6.720	7.646×10^{-5}	37.3
GSC 92-76	4201	1254	BMMG	1	Musc	294 ± 4	8.169	1.014×10^{-4}	9.0
GSC 92-77	4203	1017	BIMG	3	Biot	297 ± 4	6.776	8.490×10^{-5}	1.2
GSC 92-78	4206	5006	BMMG	3	Biot	281 ± 4	7.035	8.320×10^{-5}	2.6
GSC 92-79	4207	5006	BMMG	3	Musc	301 ± 6	7.803	9.921×10^{-5}	4.8
GSC 92-80	4199	1262	ALBT	3	Biot	274 ± 10	5.938	6.841×10^{-5}	4.1
GSC 92-81	4216	1034	KMMG	10	Musc	336 ± 5	8.263	1.185×10^{-4}	2.1

* BIMG -biotite monzogranite; BMMG - biotite-muscovite monzogranite; GRMG - greisenized monzogranite; KMMG - kaolinized muscovite monzogranite; ALBT - albitite.

Table 3. Two step ^{40}Ar-^{39}Ar and integrated ages of biotite and muscovite from Davis Lake Pluton

GSC #	K-Ar #	Sample	Mineral	Rock* Type	Distance from Carapace zone	Wt % K	% Atmos. Ar	Integrated Age (Ma)	STEP 1 (Ma) + (% GAS)	STEP 2 (Ma) + (% GAS)
GSC 92-82	4304	1013	Biot	BIMG	3	6.1	4.4	271 ± 3	278 ± 1 (70)	276 ± 1 (27)
GSC 92-83	4327	0138	Musc	BMMG	4	6.5	4.8	307 ± 3	300 ± 1 (64)	320 ± 1 (35)

* see Table 2 for rock type

The northern contact of the Davis Lake Pluton with the Meguma metasediments is characterized by a hydrothermally altered carapace zone containing more localized zones of intense greisenization of which the East Kemptville tin deposit is the most extensive. The carapace zone attains a maximum thickness of 2 km from the granite/metasediment contact. Biotite monzogranite is considered to be the parental phase of the Davis Lake Pluton. Local and more regional (carapace zone) focusing of post magmatic hydrothermal fluids has resulted in the alteration of these rocks to form, in order of intensity of alteration, biotite-muscovite monzogranite, muscovite-biotite-chlorite monzogranite, muscovite monzogranite (leucocratic), and greisenized/albitized monzogranite (Geological Survey of Canada Open File 2157, 1990). One of the most noted features of this alteration sequence is the gradual replacement of biotite by muscovite (with ghosted zircons) and the gradual to complete replacement of feldspar phenocrysts by quartz-albite ± topaz assemblages.

The geochronology study in this region, of which the present data forms a part, is aimed at determining the onset and temporal extent of the hydrothermal event affecting the Davis Lake Pluton and surrounding Meguma formations.

Rb-Sr (Richardson et al, 1989; Kontak and Cormier, 1991), ^{40}Ar-^{39}Ar (Kontak and Cormier, 1991), ^{207}Pb-^{206}Pb (Chatterjee in Kontak and Cormier, 1991) and U-Pb zircon (Boyle and Roddick, unpublished data) dates for various hydrothermally altered and unaltered rocks from the Davis Lake Pluton show a wide range in ages from 240 to 385 Ma. The U-Pb zircon ^{207}Pb-^{206}Pb age of 368 Ma and the U-Pb zircon age range of 375-385 Ma fall within the modal age group for the South Mountain Batholith of 350-400 Ma. The emplacement age for the Davis Lake Pluton is still quite controversial (i.e. Pb isotopic vs U-Pb age discrepancies), but the recognition of thermal resetting(s) of the Rb-Sr and K-Ar systems is well established (Kontak and Cormier, 1991, this study).

Tables 2, 3 and 4 present K-Ar, 2-step ^{40}Ar-^{39}Ar data and the location for 13 biotite and muscovite separates from 11 different rock types in the Davis Lake Pluton.

Within the altered carapace zone of the Davis Lake Pluton the biotites are almost completely replaced by muscovite; pristine biotites, therefore, could not be obtained from this zone. The muscovites in the carapace zone (GSC 92-71, 72, 73, 74) show a very tight range of ages from 294-301 Ma. Outside of the carapace zone, and for distances up to 3 km from it, the muscovites (GSC 92-76, 79) continue to display ages in the above range. Muscovite (GSC 92-83) at the southern granite-metasediment contact, 4 km from the carapace zone, however, shows a slightly older integrated ^{40}Ar-^{39}Ar age of 307 Ma with the high temperature release at 320 Ma suggesting a response to re-heating or slow cooling (Table 3). Biotites (GSC 92-75, 77, 78, 82) from the

Table 4. Location of biotite and muscovite mineral separates from Davis Lake Pluton

GSC#	K-Ar#	Sample	Mineral	NTS	UTM Zone	UTM East	UTM North	Latitude	Longitude
GSC 92-71	4192	1074	Biotite	21A	20	282522	4884214	44°4.8'N	65°43.2'W
GSC 92-72	4193	1089	Muscovite	21A	20	284750	4885600	44°5.4'N	65°41.3'W
GSC 92-80	4199	1262	Biotite	21A	20	288243	4883702	44°4.7'N	65°38.7'W
GSC 92-75	4200	1254	Biotite	21A	20	286844	4885044	44°5.0'N	65°39.6'W
GSC 92-76	4201	1254	Muscovite	21A	20	286844	4885044	44°5.0'N	65°39.6'W
GSC 92-77	4203	1017	Biotite	21A	20	277077	4877193	44°1.0'N	65°46.8'W
GSC 92-73	4204	1031	Muscovite	21A	20	286874	4887710	44°4.2'N	65°43.2'W
GSC 92-74	4205	1083	Muscovite	21A	20	284580	4884760	44°5.4'N	65°41.4'W
GSC 92-78	4206	5006	Biotite	21A	20	282450	4879920	44°2.4'N	65°43.2'W
GSC 92-79	4207	5006	Muscovite	21A	20	282450	4879920	44°2.4'N	65°43.2'W
GSC 92-81	4216	1034	Muscovite	21A	20	294857	4882213	44°4.2'N	65°33.6'W
GSC 92-82	4304	1013	Biotite	21A	20	280600	4878896	44°1.9'N	65°44.4'W
GSC 92-83	4327	0138	Muscovite	21A	20	286959	4880708	44°3.0'N	65°39.6'W

Davis Lake Pluton display a range in ages from 271-296 Ma with an apparent decrease in age with increased alteration of the biotite monzogranite.

Taken together, the biotite and muscovite data do not show any apparent increases in age with distance from the carapace zone. The very narrow width of the Davis Lake Pluton (<10 km), the intensity and extent of alteration in the carapace zone (up to 2 km thick), the presence of a hydrothermal kaolinized breccia zone (Rushmere Zone) along the southern granite/metasediment contact, and the apparent resetting(s) of K-Ar ages for both biotite and muscovites in altered and unaltered rocks would suggest that post-magmatic thermal processes have been pervasive throughout the entire mass of the pluton. Cooling to below a muscovite closure temperature (350°C) took place at about 300 Ma whereas biotite K-Ar closure (280-300°C) took place slightly later (about 280 Ma).

The muscovite (GSC 92-81) from the kaolinized Rushmere Breccia Zone along the southern intrusive contact shows an older date (336 Ma) than other muscovite-biotite separates for the Davis Lake Pluton, indicating that this region of the pluton, which is farthest from the carapace zone (10 km), went through an earlier cooling period.

Biotite (GSC 92-80) from the small albitic intrusion in the centre of the pluton shows a K-Ar age of 274 Ma with a high associated error (± 10 Ma). Preliminary U-Pb results, although highly discordant, suggest ages greater than 350 Ma for this body indicating that K-Ar systematics have also undergone resetting in this intrusion, possibly by the hydrothermal event associated with greisenization of the northern granite-metasediment contact zone.

These samples were collected and interpreted by D.R. Boyle.

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NEW BRUNSWICK
(GSC 92-84 to 92-88)

GSC 92-84 Biotite
911 ± 16 Ma

Wt % K= 6.592
Rad. Ar= 3.037 x 10^{-4} cm^3/g
K-Ar 4344 % Atmos. Ar= 1.2

From a glacial erratic granitoid boulder.

(21 J/4) From a major concentration of glacial erratics (along a property line) in an area where erratics are otherwise scarce. The sample site is 100 m west of the north-south road leading north from Richmond Corner to the bridge over the Meduxnekead River, N.B.; 46°11.3′N, 067°42′W; UTM zone 19, 600323E 5115576N; sample PC1-91. Collected by V.K. Prest and A.A. Seaman.

Thin section reveals fresh microcline and clear quartz with scant myrmekite; good green hornblende but with some alteration or intergrowths with brown biotite.

The myriad of glacial boulders and cobbles in this location are spread over a width of 2 to 5 m and a length of at least 1 km. The boulders were obviously cleared from the adjoining fields south of the Meduxnekead River and thus may denote the loci of a former ice margin. The sudden appearance of innumerable boulders and cobbles south of the Meduxnekead River in a broad region of otherwise sparse erratics appears to denote an ice lobe position, — either an end moraine of south-flowing St. John River valley ice or a marginal moraine from east-southeast flowing ice in the Meduxnekead valley. The Precambrian age of 911 Ma is somewhat surprising but favours an incursion of an ice lobe from Maine. See also GSC 92-85.

GSC 92-85 Biotite
366 ± 11 Ma

Wt % K= 5.653
Rad. Ar= 8.914 x 10^{-5} cm^3/g
K-Ar 4345 % Atmos. Ar= 1.6

From a glacial erratic granitoid boulder.
(21 J/4) From the same bouldery concentration (boulder only about 3 m from GSC 92-84), west of the road from Richmond Corner to the Meduxnekead River bridge, and about 2 km north of Richmond Corner, N.B.; 46°11.3'N, 067°42.0'W; UTM zone 19, 600323E 5115576N; sample PC2-91. Collected by V.K. Prest and A.A. Seaman.

This granodiorite boulder has good clean, brown biotite with lesser amounts of pale green hornblende. The plagioclase is moderately altered with some zoning and minor K-feldspar.

The K-Ar date indicates a Late Devonian source area, probably in Maine, as was suggested for the Precambrian source area for GSC 92-84. Though glacial indicators denote a strong ice-flow southward from Grand Falls to the Woodstock area (Seaman, 1989) it is unlikely that such ice-flow could account for the Meduxnekead erratics; a Precambrian component in the Saint John Valley till has not been proven for more than some 15 km south of Grand Falls, and there are no obvious Devonian intrusives within the valley. On the otherhand there are eastward ice-flow indicators in the Woodstock area (Seaman, 1989) which would satisfy the concept of ice flow from Maine.

REFERENCE

Seaman, A.A.
1989: Glacial ice-flow directions in New Brunswick. New Brunswick Department of Natural Resources and Energy, Mineral and Energy Division, Plate 89-11.

GSC 92-86 Hornblende
993 ± 20 Ma

Wt % K= 1.178
Rad. Ar= 6.064 x 10^{-5} cm^3/g
K-Ar 4346 % Atmos. Ar= 1.4

From a glacial granitic boulder.
(21 I/3) Exposed in a road-side cut on Highway 112, 5.5 km east of New Canaan, near the southern edge of the Carboniferous lowland, southeastern New Brunswick; 46°04.5'N, 65°21.0'W; UTM zone 20, 318276E 5104847N; sample PC3-91. Collected by V.K. Prest.

This is a clean, fresh-looking sample with lacy feldspar intergrowths, some plagioclase, probably oligoclase and clean quartz. Mafic minerals include green hornblende with minor included apatite and scant biotite.

The several boulders both in and on the glacial till in this location — only 2 km northeast of a small outcropping of igneous rock on a branch road connecting highways 112 and 885 suggests a local source area. The igneous rock, however, has been assigned to the Devonian on the Geological Map of New Brunwick (Potter et al., 1979). The outcropping appears as a 'window' in the Carboniferous basin. A K-Ar dating of this bedrock exposure, and of another 'window' only 5 km farther east, is thus desirable; otherwise the closest Precambrian rocks lie some 40 km to the south. Thus a northward ice flow may be responsible for the granitoid boulders along this part of highway 12. It is only in recent years that northward ice flow from the Caledonia Highlands has been documented (Seaman, 1989). He noted evidence of northward flow across the Petitcodiac and Waterford areas (21 H/14 and 11) to the south of the boulder site. Details of the complex ice flow events in New Brunswick are further documented by Seaman (1991). See also the K-Ar ages on field samples taken from highway 112 west of Canaan Forks, GSC 90-113 (Hunt and Roddick, 1991) and GSC 91-165 (Hunt and Roddick, 1992), where northward transport of Devonian age boulders from a Mississippian conglomerate was suggested.

REFERENCES

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1992: 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.

Potter, R.R., Hamilton, J.B., and Davies, J.L.
1979: Geological Map of New Brunswick; Department of Natural Resources, Map number HR-1, second edition.

Seaman, A.A.
1989: Glacial ice-flow directions in New Brunswick; New Brunswick Department of Natural Resources and Energy, Mineral and Energy Division, Plate 89-11.

Seaman, A.A. (cont.)
1991: Complex glacial ice-flow events in New Brunswick with particular reference to the Oromocto Lake area, and implications for drift prospecting; New Brunswick Natural Resources and Energy, Geoscience Report 89-2, 48 p.

GSC 92-87 Biotite
426 ± 16 Ma

Wt % K= 7.922
Rad. Ar= 1.48 x 10^{-4} cm^3/g
K-Ar 4347 % Atmos. Ar= 1.7

From a granitoid glacial boulder.
(21 I/3) Newly exposed by road-work along Highway 112 near Frederiction Road corner, N.B.; 46°04′N, 065°11′W; UTM zone 20, 331139E 5103554N; sample PC4-91. Collected by V.K. Prest.

The site lies within the belt of Mississippian-age sediments comprising the northeast arm of the Caledonian Hylands flanked by the Carboniferous Lowland.

This is a definite igneous rock showing deuteric alteration. The abundant plagioclase is highly altered to saussurite with plentiful flakes of muscovite; there is some clean, zoned plagioclase and brown biotite with some showing alteration to green chlorite and epidote; the quartz is clean and unstrained.

The Silurian-age determination poses a problem as the closest Silurian bedrock lies far to the west-southwest. It is more likely that the glacial erratic was derived from a boulder within a Mississippian conglomerate; hence minimal glacial transport may be indicated.

GSC 92-88 Biotite
333 ± 14 Ma

Wt. % K= 4.77
Rad. Ar= 6.798 x 10^{-5} cm^3/g
K-Ar 4348 % Atmos. Ar= 4.0

From a granite.
(21 J/4) Collected from west side of Highway 103 between Woodstock and intersection of road east to Hartland covered bridge, N.B.; 46°15′N, 067°30′W; UTM zone 19 615627E 5122700N; sample PC5-91. Collected by V.K. Prest.

A flat slab of granite (2/3 m square and 5 to 7 cm thick) taken from among numerous boulders and cobbles from a wooded field edge, and obviously representing glacial erratics cleared from the adjoining till field.

Thin section shows moderate alteration of the feldspar, some clean albite-twinned plagioclase, with clean brown biotite and palegreen hornblende.

The granite slab must have been transported high in the ice. The closest Devonian intrusives lie some 10 km to the south (Potter et al., 1979) but northward ice-flow indicators are not known in this part of Saint John River valley (Seaman, 1989). Devonian intrusives are also known some 20 km east of the boulder site but, again, westward ice-flow indicators have not been recognized in this area. In the course of the present study of glacial erratics striae indicative of ice-flow toward N315° were observed in two locations, one of them just south of Lakeville. This ice-flow trend may indicate transport from Devonian rocks in the vicinity of Temperance Valley some 30 km to the southeast.

REFERENCES

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Labrador
(GSC 92-89 to GSC 92-95)

GSC 92-89 Biotite
1823 ± 17 Ma

Wt % K= 5.552
Rad. Ar= 6.802 x 10^{-4} cm^3/g
K-Ar 4287 % Atmos. Ar= 0.4

From a granite.

(14 E/9) From a small knoll at southeast end of pond, immediately NE of summit of Wheeler Mountain, Labrador; 57°34′N, 62°19′W; UTM zone 20, 540879E, 6380461N; sample EC-89-354. Collected and interpreted by R.F. Emsile.

This sample is a coarse grained massive, salmon pink to pink-grey granite. Microcline microperthite, and weakly-zoned plagioclase accompany large, strained to partly polygonalized quartz. Deep brown biotite, locally highly chloritized, is the chief mafic mineral.

Samples GSC 92-89 to 92-95 belong to three discrete age groups of relatively fractionated, fluorite-bearing, post-tectonic Proterozoic granitoid rocks which have been identified in the Okak Bay area of Labrador on the basis of U-Pb zircon and K-Ar geochronology. Two of the groups occur entirely within the Archean Nain Province; the third occurs largely within the Early Proterozoic Churchill Province (Torngat Orogen). None of the three episodes of granitoid magmatism can be directly linked to orogenic activity in its immediate surroundings. Although all three granite groups are broadly similar, petrologically and chemically, to the felsic rocks of the anorogenic Elsonian (~1450-1290 Ma) Nain Plutonic Suite (NPS), only one of the dated units can be correlated confidently with Elsonian magmatism. The other two groups represent significantly older anorogenic or postorogenic magmatism.

The oldest unit, the Wheeler Mountain granites of the Nain Province, has yielded a U-Pb zircon age of about 2135 Ma with K-Ar ages of 1823 ± 17 Ma and 1671 ± 23 Ma (GSC 92-89, 90). A major tectonothermal event of this age has not been previously identified in northern Labrador, and the origin of the granites is uncertain. Three small intrusions on offshore islands (White Bear, Saddle, Opingiviksuak) in the Nain Province were intruded about 1775 Ma (U-Pb zircon) and gave K-Ar ages of 1724 ± 15 Ma, 1734 ± 28 Ma, and 1589 ± 17 Ma (GSC 92-93, 92, 91). These granites may have been intruded post-kinematically into a stable Archean foreland representing a delayed effect of a preceding collisional event (Torngat Orogeny) between Nain and Churchill Province to the west. The youngest unit dated, at about 1318 Ma (U-Pb zircon) with K-Ar ages of 1322 ± 18 Ma and 1302 ± Ma (GSC 92-94, 95), is Umiakovik Lake batholith which outcrops largely within the eastern Churchill Province, and is one of the largest granitoid intrusions of the Nain Plutonic Suite. For a detailed discussion and interperation of these samples see Emslie and Loveridge (in press).

REFERENCE

Emslie, R.F. and Loveridge, W.D.

in press: Fluorite-bearing early and middle Proterozoic granites, Okak Bay, Labrador: Geochronology, Geochemistry and Petrogenesis; Lithos, in press.

GSC 92-90 Biotite
1671 ± 23 Ma

Wt % K= 7.436
Rad. Ar= 7.954 x 10^{-4} cm^3/g
K-Ar 4288 % Atmos. Ar= 0.4

From a granite.

(14 E/9) Sample taken at rockfall on hillside below peak about 3 km north of Siugak Brook, Labrador; 57°37.5′N, 62°23.7′W; UTM zone 20, 536135E, 6386911N; sample EC-89 357. Collected and interpreted by R.F. Emsile.

This sample is a coarse grained grey-pink massive hornblende-biotite granite. Microcline microperthite is the dominant feldspar; not more than about 10 percent plagioclase is present. Most quartz is highly strained but recrystallized domains are small and localized. Pale to deep olive-brown biotite is slightly more abundant than deep brownish-green hornblende. See GSC 92-89 for a detailed discussion and interperation of this sample.

GSC 92-91 Biotite
1594 ± 17 Ma

Wt % K= 7.770
Rad. Ar= 7.731 x 10^{-4} cm^3/g
K-Ar 4166 % Atmos. Ar= 0.345

From a granite.

(14 F/5) Near shore, south side of Lady Bight Harbour, Opingiviksuak Island, Labrador; 57°26.2′N, 61° 32.5′W; UTM zone 20, 587550E, 6366722N; sample EC-87-142. Collected and interpreted by R.F. Emsile.

This sample is a medium grained massive, pink-grey granite. Quartz, strained to sutured, microcline microperthite, and plagioclase are the main minerals. Pale to medium olive green biotite is accompanied by traces of bluish-green to dark olive hornblende. See GSC 92-89 for a detailed discussion and interperation of this sample.

GSC 92-92 Biotite
1734 ± 28 Ma

Wt % K= 7.821
Rad. Ar= 8.857 x 10^{-4} cm^3/g
K-Ar 4167 % Atmos. Ar= 0.3

From a granite.
(14 F/11) Southeast side of Saddle Island, high ground, Labrador; 57°36′N, 61°24′W; UTM zone 20, 595625E, 6385094N; sample EC-87-143c. Collected and interpreted by R.F. Emsile.

The Saddle Island sample is a medium grained, massive, pink to pink-grey biotite granite, locally fluorite-bearing. Thin sections show that strained quartz, microcline microperthite, and normally-zoned plagioclase are the principal minerals. Pale brownish-green to olive biotite is the mafic silicate and it shows only minor, local chloritic alteration. See GSC 92-89 for a detailed discussion and interperation of this sample.

GSC 92-93 Biotite
1724 ± 15 Ma

Wt % K= 7.826
Rad. Ar= 8.782 x 10^{-4} cm^3/g
K-Ar 4286 % Atmos. Ar= 0.5

From a granite.
(14 F/13) White Bear Island, in bay about 1 km northwest of White Point, Labrador; 57°54′N, 61°40′W; UTM zone 20, 579032E, 6418145N; sample EE-88-096. Collected and interpreted by R.F. Emsile.

This sample is a medium- to coarse-grained grey massive granite. Very similar in appearance to Opingiviksuak granite. Microcline microperthite, markedly zoned plagioclase and large quartz grains form most of the rock. Light to medium olive biotite is the main mafic silicate and chloritization is minor. See GSC 92-89 for a detailed discussion and interperation of this sample.

GSC 92-94 Hornblende
1322 ± 18 Ma

Wt % K= 0.547
Rad. Ar= 4.147 x 10^{-5} cm^3/g
K-Ar 4164 % Atmos. Ar= 1.77

From a granite.
(14 E/7) Top of promontory overlooking Umiakovik Lake from the north, Labrador; 57°24′N, 62°50.5′W; UTM zone 20, 509515E, 6361713N; sample EC-87-119. Collected and interpreted by R.F. Emsile.

This sample is a coarse grained grey to pinkish biotite-hornblende granite. Large subequant strained quartz, microcline microperthite, and patchy-zoned plagioclase are the essential minerals. Very deep brown biotite and dark olive to tan hornblende are the mafic silicates. See GSC 92-89 for a detailed discussion and interperation of this sample.

GSC 92-95 Biotite
1302 ± 19 Ma

Wt % K= 7.009
Rad. Ar= 5.198 x 10^{-4} cm^3/g
K-Ar 4164 % Atmos. Ar= 0.2

From a granite.
(14 E/7) Details as for GSC 92-94

OUTSIDE CANADA
(GSC 92-96 to 92-100)

GSC 92-96 Whole Rock
0.45 ± 0.02 Ma

Wt % K= 1.707
Rad. Ar= 2.980 x 10^{-8} cm^3/g
K-Ar 4292 % Atmos. Ar= 75.5

From a basanite.
From the Black River region east of Fort Yukon, eastern Alaska; 66.632°N, 143.176°W; UTM zone 7, 403701E, 7391848N; sample BR-1b-89. Collected by K. Wirth and interpreted by J.K. Mortensen.

See GSC 92-98 for discussion.

GSC 92-97 Whole Rock
6.26 ± 0.15 Ma

Wt % K= 1.113
Rad. Ar= 2.712 x 10^{-7} cm^3/g
K-Ar 4293 % Atmos. Ar= 69.3

From a basanite.
From a massive lava flow on the eastern side of Prindle Volcano, northeastern Tanacross quadrangle, eastern Alaska; 63.715°N, 141.634°W; UTM zone 7, 468673E, 7065214N; sample PV-14-89. Collected by K. Wirth and interpreted by J.K. Mortensen.

See GSC 92-98 for discussion.

GSC 92-98 Whole Rock
3.57 ± 0.14 Ma

Wt % K= 1.271
Rad. Ar= 1.768 x 10^{-7} cm^3/g
K-Ar 4294 % Atmos. Ar= 59.5

From a basanite.
From a massive lava flow on the eastern side of Prindle Volcano, northeastern Tanacross quadrangle, eastern Alaska; 63.670°N, 141.613°E; UTM zone 7, 469663E, 7060190N; sample PV-15-HF. Collected by K. Wirth and interpreted by J.K. Mortensen.

This sample, GSC 92-96, and 92-97 are from alkali olivine basalt centres in eastern Alaska that are thought to be broadly correlative with the Selkirk Lavas in western Yukon that regionally give ages of 0.5 to 20 Ma, and in the Yukon are termed the Selkirk Lavas. The two Prindle Volcano samples were expected to be the same age, and in view of the very well preserved constructional morphology of the volcanic edifice, were expected to be no older than 2 Ma. The data may indicate that either excess Ar is present, possibly related to the mantle and crustal xenoliths that are abundant in the flow samples, or that fractionation of the atmospheric Ar component has occurred during bakeout. The age for the Black River volcanic sample is considerably younger than K-Ar whole rock ages of about 16 Ma that have been reported previously for young basalt flows in this area, and is tentatively interpreted as the age of eruption.

GSC 92-99 Biotite
162.1 ± 2.5 Ma

Wt % K= 6.327
Rad. Ar= 4.171 x 10^{-5} cm^3/g
K-Ar 4289 % Atmos. Ar= 4.6

From a biotite quartz monzonite.
From western Liaoning Province, 3.5 km northeast of the Erdaogou gold mine, 8 km east-southeast of the Jinchanggouliang gold mine, and 2 km south of the Liaoning/Inner Mongolia border, People's Republic of China; sample MLB-89-RC-1. Collected and interpreted by J.K. Mortensen.

This sample is from a massive biotite quartz monzonite pluton that forms the local basement for a sequence of Late Jurassic subaerial felsic flows. Clasts of the pluton occur in conglomeratic interbands in the volcanic sequence. The biotite age is in good agreement with a preliminary U-Pb zircon age for the same sample (Mortensen, unpublished data).

GSC 92-100 Alunite
11.2 ± 0.3 Ma

Wt % K= 4.839
Rad. Ar= 2.112 x 10^{-6} cm^3/g
K-Ar 4339 % Atmos. Ar= 80.4

From a rhyolite ignimbrite flow.
From Lookout Rocks, elevation 580 m, at head of Ohio Creek, Thames District, Coromandel Peninsula, North Island, New Zealand; NZMS 260 T12 381520; 37°06.5'S, 175°34.1'E; UTM Zone 60, 271129E, 4085732N; sample DY-3344. Collected by K.M. Dawson.

The sample is from hydrothermally altered rhyolite ignimbrite flows and tuffs of the Late Miocene-Pliocene Whitianga Group which are underlain by Miocene Coromandel Group andesite and dacite flows (Skinner, 1986). The assemblage natroalunite-quartz-illite- sericite represents advanced argillic alteration about 400 m stratigraphically above a porphyritic quartz diorite stock of 9 to 12 Ma (ibid) at Ohio Creek. The stock contains disseminated Cu, Mo (porphyry type) mineralization in addition to previously mined epithermal Au, Ag, Te veins of the Kaiser-Sylvia mine (Merchant, 1986). The altered rhyolites also lie 500 m stratigraphically above several bonanza-type epithermal Au, Ag veins of the previously producing Thames district, interpreted to be related to Pliocene volcanic centres (ibid).

The mineral separate is a mixture of alunite and quartz. The Miocene age of the alunite alteration of rhyolite represents advanced argillic alteration developed contemporaneously with the cooling of the underlying quartz diorite stock(s) and emplacement of Cu, Mo and Au, Ag, Te mineralization at depth. Bonanza veins at Thames may also be related to these Miocene stocks rather than to Pliocene volcanic centres.

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1986: Neogene volcanism of the Hauraki Volcanic Region; Royal Society of New Zealand Bulletin, 23, p. 21-47.

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
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
GSC Paper 92-2, Report 22	92-1 to 92-100

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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74-I	80-147, 148, 149
74-K	66-99; 78-174, 175
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- DISCUSSION -

Circa 1.75 Ga ages for plutonic rocks from the Southern Province and adjacent Grenville Province: what is the expression of the Penokean Orogeny?: Discussion

K.D. Card[1]

Card, K.D., Circa 1.75 Ga ages for plutonic rocks from the Southern Province and adjacent Grenville Province: what is the expression of the Penokean Orogeny?: Discussion; in Radiogenic Age and Isotopic Studies: Report 6, Geological Survey of Canada, Paper 92-2, p. 227-228.

Davidson et al. (1992) question the existence of effects of the Penokean orogeny on Lower Proterozoic rocks of the Lake Huron region, arguing that existing geochronology does not record events of Penokean age. I would argue that although the absolute age of these deformational-metamorphic events are not well-established, there are reasonable grounds for correlating post-Nipissing (2.2 Ga), pre-Killarney (1.75 Ga) deformation and metamorphism in the Lake Huron region with the Penokean Orogeny.

It has long been recognized (e.g., Card, 1978, p. 278) that effects of the Penokean orogeny on these rocks are restricted to deformation and metamorphism and that the plutonic rocks present are either older (e.g., Creighton, Murray) or younger (e.g., Chief Lake, Killarney) than the deformational-metamorphic events. The structural style and low-grade metamorphism mark this terrane as a fold-and-thrust belt (e.g., Zolnai et al., 1984) and so the absence of synorogenic plutons cannot be taken as evidence that the region was unaffected by the Penokean orogeny.

Davidson et al. (1992) also suggest that all of the deformation and metamorphism may be pre-Penokean. As Card (1978) pointed out, the Huronian rocks have clearly been affected by more than one deformational event. Early, pre-Nipissing (2.2 Ga) deformation resulted in major folding of Huronian strata. This deformation, referred to as the "Blezardian orogeny" by Stockwell (1982), may or may not be coeval with emplacement of the Murray and Creighton plutons at 2.33-2.38 Ga. However, later deformational-metamorphic events occurred after emplacement of the Nipissing Diabase. Deformation and regional metamorphism probably began prior to emplacement of the Sudbury Igneous Complex (1.85 Ga) as there are structures and relatively high metamorphic grades (lower amphibolite facies) in Huronian rocks that are not present in nearby rocks of the Sudbury structure (Card, 1978). Events following emplacement of the Sudbury igneous complex resulted in deformation and low-grade metamorphism of the rocks of the Sudbury structure and further deformation and retrograde metamorphism of the Huronian rocks. Cumulatively, these various deformational-metamorphic events resulted in formation of an east-west fold-and-thrust belt with nodal metamorphic patterns, and no or few synorogenic granitoid intrusions. They were followed by ~1.75 Ga (Killarney, Chief Lake) and ~1.45 Ga (Croker) anorogenic magmatism. Fairbairn et al. (1969) determined Rb-Sr whole rock "ages" of metamorphosed Huronian sediments and volcanics ranging from 1800 ± 100 Ma to 2174 ± 125 Ma, with an average 1950 ± 100 Ma, which they interpreted as the age of regional metamorphism.

The Penokean orogeny in the type area of Wisconsin and Michigan, where it is clearly a composite event probably attributable to multiple terrane collisions, began by ~1.89 Ga and ended by ~1.835 Ga with post-orogenic felsic magmatism (Sims et al., 1989). There, the orogeny was followed by anorogenic magmatism at ~1.76 Ga and ~1.45 Ga. It resulted in formation of southern composite magmatic terranes with abundant synorogenic plutons and northern fold-and-thrust belts (e.g., Marquette Range fold-belt) with east-west structural trends, nodal metamorphic patterns, and few or no granitoid intrusions. The contact or suture between the southern magmatic terranes and the northern fold-and-thrust belt, the Niagara Fault, must continue eastward beneath younger cover rocks and may be represented in the Lake Huron region by the Manitoulin Island discontinuity (van Schmus et al., 1975).

In the writer's opinion, there are reasonable grounds, including similarities in age, structural trend and style, tectonic setting, and metamorphic patterns, for considering the post-Nipissing deformation and metamorphism of the pre-1.75 Ga rocks of the Lake Huron region as distal effects of the ~1.85 Ga Penokean orogeny. The alternative would require the invention of yet another name for this post-Nipissing, pre-Killarney, ca. 1.85 Ga deformation and metamorphism.

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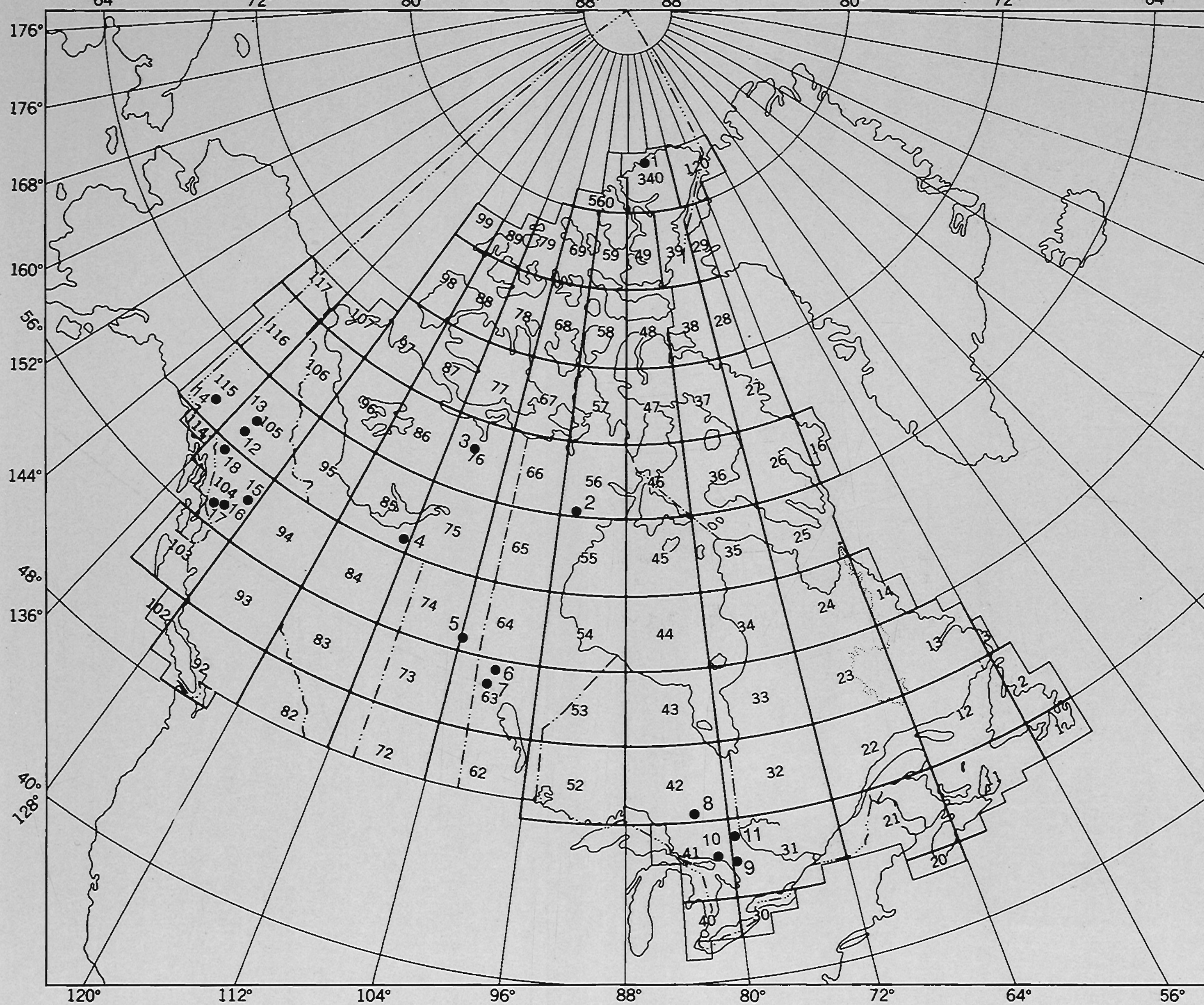
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