UNIVERSITY OF ALBERTA

ORIGIN AND EVOLUTION OF MID- TO LATE-ARCHEAN CRUST IN THE WESTERN SLAVE PROVINCE, CANADA.

by

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A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirement for the degree of Doctor of Philosophy.

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled Origin and Evolution Mid- to Late-Archean Crust in the Western Slave Province, Canada submitted by Katsuyuki Yamashita in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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Abstract

Geochemical and isotopic study of mid- to late-Archean supracrustal rocks and granitoids from the Slave province, Canada was undertaken to better understand the regional isotopic structure and tectonomagmatic evolution of this Archean craton. From the Hanikahimajuk Lake area, U-Pb ages for the pre-Yellowknife Supergroup (YKSG) basement tonalite, syn-YKSG diorite, Rim granite and post-deformation gabbro were determined at 3377.7 +1.4/-1.3 Ma, 2679.2 ± 1.1 Ma, 2676.5 +4.1/-3.6 Ma and 2574.7 ± 1.4 Ma, respectively. The YKSG mafic and intermediate volcanic rocks can be classified into back-arc (or MORB) type and island-arc type based on their trace element characteristics. However, close field proximity suggest that both of these rock types originated from the same back-arc system. The felsic volcanic rocks and Rim granite could have been generated through dehydration melting of mafic protoliths at low pressure. The geochemistry of syn-volcanic diorites indicate that they are plutonic equivalents of the YKSG volcanic rocks. The εNd_T values of YKSG volcanic rocks range from -1.2 to +2.5, indicating that the greenstone belt was, at least in part, constructed on the pre-YKSG basement, consistent with an ensialic back-arc.

From the Yellowknife area, YKSG volcanic rocks and Burwash Formation turbidites were analyzed for major/trace element geochemistry and Nd isotopes. The Nd isotopic signature of the YKSG volcanic rocks can be explained by assimilation of the pre-YKSG basement and fractional crystallization of mantle derived magma. On the other hand, the geochemical and isotopic signatures of the Burwash Formation turbidites can be explained by unmixing of detritus derived from crustally contaminated YKSG volcanic rocks and a less contaminated quartzofeldspathic (granitic?) component.

The 2.58-2.62 Ga syn- to post-deformation granitoids from the southwestern Slave province can be generated though partial melting of mafic protoliths at high (>0.8 GPa) pressure and partial melting of metasedimentary rocks, respectively. The Pb isotopic signatures of these granitoids indicate that pre-YKSG basement of >3.2 Ga was involved in their origin. The isotopic signature of the pre-YKSG component can be incorporated

through assimilation of pre-YKSG basement by YKSG magmatism or by direct input of detritus into YKSG sediments. Combined Nd-Pb isotopic modeling indicates that pre-YKSG basement comprises 10-30% of the source region of granitoids from the southwestern Slave province.

Finally, the combined Nd isotopic and U-Pb geochronological study of sediments from the Slave province indicates that early- to mid-Archean crustal evolution of the Slave province was dominated by crustal recycling, rather than addition of juvenile materials from the mantle.

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TABLE OF CONTENTS

CHAPTER PAGE

LIST OF TABLES

LIST OF FIGURES

Chapter 1

Introduction

Understanding the physical and chemical evolution of the Earth is a primary goal of earth science. An effort to reveal the mechanism of crustal evolution in the Archean is fundamental work required to accomplish this goal, because the Archean cratons are the only areas on the Earth where information on the Earth's early history is still directly preserved. The main difficulties, however, come from the fact that many of the Archean cratons have undergone complex tectonomagmatic histories and thus the geological record preserved in the craton is limited and/or highly complicated. In this respect, it is natural for progress in Archean geology to be highly dependent on the advance of analytical geochemistry, especially U-Pb geochronology and trace-element and isotope geochemistry. In this dissertation, a variety of geochemical techniques are utilized to unravel problems related to the origin and evolution of Archean Slave province, NWT, Canada. The Slave province is an attractive craton for the study of Archean crustal evolution because it hosts a variety of intrusive and extrusive rocks, ranging in age from ~2.6 to >4.0 Ga, and many of its geological features are apparently different from those of other well studied Archean cratons such as the Superior province.

Two main goals of chapter 2 are to reveal the relationship between the >2.8 Ga basement suite and the ~2.7 Ga greenstone belt in the west-central Slave province, and to constrain the possible tectonic setting in which the ~2.7 Ga greenstone belt was formed. Samples for U-Pb geochronology, major- and trace-element and Nd-Pb isotopic analyses were collected from the Hanikahimajuk Lake area of the northern Slave province. Because the contact between the pre-2.8 Ga basement and the greenstone belt is commonly obscured by younger intrusions and cover sequences or is represented by a high-strain zone (Kusky 1990, James and Mortensen 1992, Isachsen and Bowring 1994), numerous authors have argued for tectonic juxtaposition of a greenstone belt onto a sialic basement (e.g. Kusky 1989, Davis et al. 1994). However, the Nd isotopic data of the ~2.7 Ga volcanic rocks studied here shows that most, if not all, of the volcanic rocks were isotopically contaminated by the basement and thus were constructed on or near existing basement.

Geochemically, the mafic to intermediate volcanic rocks of the greenstone belt can be distinguished into two groups; one that has MORB or back-arc-like trace-element characteristics and another that has island-arc like features. However, it is well known from the study of volcanic rocks in the modern back-arc setting that trace-element characteristic of volcanic rocks produced in a back arc setting is highly variable, ranging from N-MORB to those that resemble calc-alkaline arc basalt to rhyodacite (e.g. Saunders and Tarney 1984). Therefore, given the close association of these rocks in outcrops, it is likely that these are part of the same back-arc system.

In chapter 3, clastic sedimentary rocks from the southern Slave province (near Yellowknife) are examined to address the source of the widespread ~2.66 Ga turbidites which comprise a significant portion of the supracrustal sequence in the southern Slave province. Rare earth element (REE), Th, Sc and Nd isotopic geochemistry of clastic sedimentary rocks is particularly useful in understanding the nature of the exposed crust at the time of sediment deposition, as these elements are likely transferred quantitatively to the sediments during deposition. One aspect of this study which distinguishes it from most of the previous studies on sedimentary geochemistry is that different units of the Bouma sequence were sampled separately. Through careful evaluation of geochemical and isotopic data, it is shown that the upper shale units of the turbidites generally have lower εNd_T compared with the lower sand units. This general shift in the Nd isotopic signature is most likely a result of "unmixing" of detritus derived from (1) ~2.7 Ga volcanic rocks that were isotopically contaminated by the pre-2.8 Ga basement and (2) quartzofeldspathic (granitic?) rocks that were less contaminated. The geochemical and isotopic signature of crustally contaminated volcanic rocks is more apparent in the upper shale units of the turbidites. Combined with the previous study on paleocurrent analyses of these turbidites (Henderson 1975), this model also predicts the existence of less contaminated quartzofeldspathic (granitic) rocks near or west of Yellowknife. This in turn suggests that there are regions west of Yellowknife that are not underlain by pre-2.8 Ga basement, an important conclusion for regional tectonics.

In chapter 4, the petrochemistry and Nd-Pb isotopic geochemistry of ~2.62 Ga syn-deformation and ~2.59 Ga post-deformation granitoids from the southwestern Slave province are considered. Addressing the sources of these granitoids is particularly important because it should provide insight into the nature of the deep crust before the final stabilization of the craton. Furthermore, because the late-Archean granitoids from the central Slave province have been studied in detail (Davis and Hegner 1992, Davis et al. 1994, 1996), data from the southwestern Slave province should help construct the regional isotopic structure of the Slave province.

From the results of major/trace element geochemistry, it is concluded that felsic members of ~2.62 Ga syn-deformation granitoids are produced through partial melting of mafic protoliths (garnet amphibolite or hornblende eclogite) at a pressure >0.8 GPa, whereas ~2.59 Ga post-deformation granitoids are produced through partial melting of tonalitic and sedimentary rocks. However, mafic to intermediate members of the syn-deformation granitoids are probably mantle derived as suggested by Davis et al. (1994). Using the Nd-Pb isotopic signatures of these and other granitoids from the central Slave province, isotopic modeling leads to the conclusion that a source with an ultimate mantle extraction age of >3.2 Ga is required to produce any of the granitoids from the west-central Slave province. The proportion of >3.2 Ga material may vary from 10-30% for the granitoids in the southwestern Slave province and >50% for granitoids in parts of the central Slave province. The incorporation of these ancient isotopic signatures may take place through assimilation of the basement by the ~2.7 Ga volcanic rocks (which ultimately become the source of the syn-deformation granitoids) and/or through direct input of detritus into the sedimentary rocks (source of post-deformation granitoids).

In chapter 5, the investigation of early- to mid-Archean history of the Slave province using a combination Nd model age and U-Pb detrital zircon ages of clastic sedimentary rocks was undertaken. By constructing a simple model, it is shown that the early- to mid-Archean crustal evolution of the Slave province was "dominated" (but not restricted) by crustal recycling (including partial melting and/or assimilation of pre-existing crust) rather than addition of juvenile materials from the mantle. One exception is the ~3.1 to 3.3 Ga era, when the addition of juvenile material was probably large compared to the mass of pre-existing crust.

Finally, a revised model for the evolution of Slave province which is consistent with the data collected here is presented in chapter 6. In this model, greenstone belts in the west-central Slave province represent the margin of ensialic back-arc whereas felsic dominated volcanic belts in the eastern Slave province represent the fore-arc. Subsequent collision of hypothetical crust in the eastern Slave province led to regional deformation, high T-low P metamorphism, and generation of 2.62-2.58 Ga magmatism.

In summary, this dissertation has clearly increased our knowledge on the geochemical and isotopic structure of the Slave province. However, we are still far from fully understanding the history of this craton. Further work, particularly on the eastern Slave, may reveal more.

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Chapter 2

Origin and evolution of mid-to late-Archean crust in the Hanikahimajuk Lake area, Slave Province, Canada; Evidence from U-Pb geochronological, geochemical and Nd-Pb isotopic data. *

* A version of this paper will be submitted to Precambrian Research. Co-authored by Robert A. Creaser, Joanne E. Jensen and Larry M. Heaman. Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta, T6G 2E3

Introduction

The Slave province is a small Archean craton located on the northwestern margin of the Canadian shield. The craton hosts a wide spectrum of intrusive and extrusive rocks ranging in age from ~2.6 to 4.0 Ga (Bowring e al. 1989, Villeneuve and van Breemen 1994, Bleeker and Stern 1997), making it one of the most attractive cratons to investigate the early Earth's tectonomagmatic evolution. A large number of geochronological and field oriented studies have been undertaken over the past 25 years and have resulted in several different tectonic models for the Slave province. These models can broadly be classified into two groups; (1) an intracontinental rift model, possibly followed by closure of the basin and eastward subduction of the oceanic lithosphere (Henderson 1981, Fyson and Helmstaedt 1988, MacLachlan and Helmstaedt 1995) and (2) island arc-continent accretion model (Kusky 1989, Davis et al. 1994). The existence of such conflicting models is largely due to a lack of firm understanding of the tectonic relationship between the 2.8~4.0 Ga basement and the overlying 2.6~2.7 Ga greenstone terrane. Furthermore, the possible tectonic setting in which the late-Archean greenstone terranes were developed is still under debate. Although these problems can be addressed though accumulation of high quality geochemical and isotopic data, such studies on rocks other than ~2.6 Ga late-Archean granitoids are scarce in the Slave province.

This paper presents new geochronological, geochemical and Nd-Pb isotopic data for mid- to late-Archean crust from the Hanikahimajuk Lake area in the northern Slave province (Fig. 2.1 and 2.2). The main objectives of this research are to (1) address the origin of pre-2.8 Ga basement suite, (2) determine whether the ~2.7 Ga greenstone belt was constructed on the basement, or was tectonically juxtaposed with the basement subsequent to its formation and (3) constrain the possible tectonic origin of the ~2.7 Ga greenstone belt. In addition, the results of this research will be used to evaluate the validity of the previously proposed model for the origin of the 2.6 Ga late-Archean granitoids which postulates their derivation by partial melting of crustal sources; namely mafic igneous rocks for the ~2.61 Ga syn-deformation granitoids and tonalitic and sedimentary rocks for the ~2.59 Ga post-deformation granitoids (Yamashita et al. 1998, Chapter 4). The Hanikahimajuk Lake area is an ideal location for this study because the geological framework of the area has been well established by previous workers and the results of preliminary U-Pb geochronology provides strong evidence for the existence of pre-2.8 Ga basement in this region (Gebert 1995, Jensen 1995).

Regional geology

Shown in Figure 2.1 is a simplified geological map of the Slave province (after Hoffman 1989). The Archean rocks of the Slave province are typically subdivided into three lithotectonic groups; (1) 2.8~4.0 Ga pre-Yellowknife Supergroup basement (2) 2.6-2.7 Ga Yellowknife Supergroup (YKSG) supracrustal rocks and syn-volcanic plutons and (3) late-Archean felsic granitoids that are broadly contemporaneous with the ~2.6 Ga regional deformation and metamorphism.

The pre-YKSG basement comprises heterogeneous gneisses, granitoids and supracrustal sequences which pre-date the deposition of YKSG supracrustal rocks (Henderson et al. 1982, Henderson 1985, Frith et al. 1986, Bowring et al. 1989, Lambert and van Breemen 1991, Isachsen and Bowring 1994, 1997, Villeneuve and van Breemen 1994). These rocks are found from northern to southern Slave province but have not yet been documented from the eastern Slave province; a region east of the "Nd-Pb isotopic boundary" at ~111°W (Thorpe et al. 1992, Davis and Hegner 1992).

The 2.66 to 2.71 Ga volcanic and sedimentary rocks in the Slave province have collectively been grouped as the Yellowknife Supergroup by Henderson (1970). The volcanic rocks are typically bimodal, consisting of tholeiitic basalt to andesite and calc-alkaline andesite to rhyolite. These two been termed Yellowknife-type and Hackett River-type, respectively (Padgham 1992). These volcanic rocks are accompanied by contemporaneous plutonic rocks of intermediate composition. The turbiditic sedimentary rocks of the YKSG, which makes up approximately 80% of the supracrustal sequence, are largely derived from volcanic and plutonic sources (Jenner et al. 1981). However, recent detrital zircon geochronology has also recognized an input from the pre-YKSG basement (Villeneuve and van Breemen 1994). In addition to the ~2.66 Ga turbidites, younger ~2.60 Ga sandstones and conglomerates are now recognized (Isachsen and Bowring 1994, Villeneuve et al. 1997).

The late-Archean granitoids, which cover approximately 50% of the Slave craton, are typically subdivided into 2.62 to 2.60 Ga syn-deformation granitoids and 2.60 to 2.58 Ga post-deformation granitoids. The composition of these granitoids generally change from earlier metaluminous to weakly peraluminous granodiorite-tonalite, to later weakly peraluminous monzogranite (Davis et al. 1994). The intrusion of these granitoids are contemporaneous with the 2.63 to 2.58 Ga regional deformation and high T/low P type metamorphism, ranging in metamorphic grade from greenschist to granulite facies (Thompson 1978, Relf 1992, Henderson and Schaan 1993, Henderson and Chacko 1995, Pehrsson and Chacko 1997).

Geology of the Hanikahimajuk Lake area.

The Point Lake Volcanic Belt (PLVB) is a north trending, mafic-dominated volcanic belt which runs along the western margin of the Slave province in the Point Lake - Contwoyto Lake area (Fig 2.1). It continues northward to the Napaktulik Lake volcanic belt (NLVB; Fig. 2.1). Within the PLVB, two felsic-dominated NE trending arms at approximately 375 km and 425 km north of Yellowknife are informally known as *Izok Lake volcanic belt* (ILVB) and *Amooga Booga volcanic belt* (ABVB), respectively (Bostock 1980, Gebert 1995, Jensen 1995). These two belts are important hosts for the VMS base metal deposits of the region. In this study, samples for geochronological, geochemical and Nd-Pb isotopic analyses were collected from the *Amooga Booga*, *Izok Lake* and northern *Point Lake volcanic belts*. It should, however, be remembered that these belts are continuous (Gebert 1995, Jensen 1995).

The geology of the ABVB in the Hanikahimajuk Lake area has been mapped in detail at 1:50,000 scale by Gebert (1995) and Jensen (1995). The oldest unit in the ABVB is the mixed tonalite and gabbro exposed in the vicinity of Hanikahimajuk Lake (Fig. 2.2). The preliminary U-Pb data of Gebert (1995) and Jensen (1995) have shown that the age of the tonalite is ~3.35 Ga. The age of the gabbro is uncertain. However, the field relationships indicate that there is more than one generation of gabbros, with individual units being both younger and older than the tonalites.

The YKSG supracrustal rocks in the ABVB form a NE-trending belt, generally younging towards the northwest. Although some stratigraphic details remain unresolved, there appears to be two cycles of volcanism. The early cycle is mafic which grades into intermediate and felsic volcanism. This volcanic cycle is overlain by a second cycle of mafic volcanism and subsequently by clastic sedimentary rocks (Gebert 1995, Jensen 1995). The U-Pb age of a felsic volcanic rock from the ABVB was determined to be 2694 +7/-5 Ma (Mortensen et al. 1988). The sedimentary rocks in the ABVB can be found as magnetite-bearing siltstone interbedded within mafic and intermediate volcanic rocks in the eastern ABVB, or as more extensive turbidite unit and sulfide-facies iron formation (~1 m thick) in the western ABVB. In the northwestern ABVB, a younger sequence (ca. 2.6 Ga) of conglomerates and arenites are also recognized (Gebert 1995, Jensen 1995).

Supracrustal rocks of the ABVB are bordered to the north and south by large bodies of diorite. The preliminary U-Pb ages of the syn-volcanic diorites north and south of the ABVB are 2673 Ma and 2680 Ma, respectively, placing them within the YKSG (Gebert 1995, Jensen 1995).

The pre-YKSG basement, YKSG supracrustal succession and ~2.68 Ga diorites are surrounded and intruded by late-Archean granitoids (Fig. 2.2). The preliminary U-Pb data of Gebert (1995) and Jensen (1995) indicate that at least some of these granitoids are syn- to post-deformation granitoids contemporaneous with those seen in various other areas of the Slave province such as Contwoyto Lake and Yellowknife (Van Breemen et al. 1992, Villeneuve and van Breemen 1994, Davis et al. 1994). However, the "Rim granite" which borders the volcanic rocks of the ABVB south of Hanikahimajuk Lake (Fig. 2.2) has a U-Pb age that is distinct from syn- to post-deformation granitoids in the Slave province (this study). This will be discussed later in the paper.

In order to obtain the high quality U-Pb geochronological data that is required to calculate the initial εNd values of various rock types, samples for the U-Pb geochronology were collected from the units mapped as basement tonalite, syn-volcanic diorite, Rim granite and gabbros of unknown age. In addition to these samples, one felsic volcanic rock from the Izok Lake volcanic belt was also collected for regional geochronological correlation.

Analytical procedure

All samples for the U-Pb geochronology were crushed using Jaw Crusher and Bico disk mill, and were passed through Wilfley table, heavy liquid and Frantz magnetic separator to obtain zircon (or titanite) fractions. Light fractions from the Wilfley table were saved for feldspar common Pb analyses described below. The abraded zircon fractions were spiked with a ^{235}U-^{205}Pb tracer and were dissolved in teflon bombs modified from Krogh (1973). U and Pb were extracted from the sample using anion exchange chromatography (Krogh 1973). The chemical separation of U and Pb was carried out in the U-Pb geochronology laboratory at the University of Alberta, and the total analytical blank ranged from 8 to <2 pg for Pb and 2 to <1 pg for U (generally decreasing with time). U and Pb were loaded together onto a pre-cleaned Re filament together with high purity silica-gel and H_3PO_4. Their isotopic ratios were measured on a VG 354 thermal ionization mass spectrometer in a single collector mode using a Faraday or Daly Photomultiplier detector.

The U and Pb isotopic ratios were corrected for mass discrimination (0.98 ‰/a.m.u. for Pb^+, 1.4 ‰/a.m.u. for UO_2^+) based on repeated analyses of NBS981 and U500 isotopic standards ($^{207}Pb/^{205}Pb$=0.913688+0.000141, $^{238}U/^{235}U$=0.996245±0.000374). All errors quoted on Table 2.1 were calculated using an in-housed error propagation program, and the 2σ errors for the U-Pb ages were calculated using the error propagation routine of Davis (1982).

The feldspars for common Pb analyses were purified using heavy liquid (acetone-TBE mixture), Frantz isodynamic separator and by subsequent hand picking under a binocular microscope. 24 hour leaching in 2N HCl (L1), 6N HCl (L2), 16N HNO_3 (L3) and 16N HNO_3 plus one drop of 48% HF (L4) was applied to each sample prior to the final dissolution in 4:1 mixture of 48% HF and 16N HNO_3 (R). Pb was extracted using HBr-HNO_3 media chemistry modified from Lugmair and Galer (1992). The Pb isotopic ratios were measured on a VG MM 30 mass spectrometer in a single collector mode or on a VG 354 mass spectrometer in a static mode for smaller samples. Pb isotopic compositions were corrected for mass discrimination (1.4 ‰/a.m.u. for MM30 and 1.3~2.2 ‰/a.m.u. for VG 354) based on the repeated analyses of NBS 981 standard. The reproducibilities for the $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios were 0.19, 0.20 and 0.28 ‰ for MM 30 and 0.39, 0.46 and 0.69 ‰ for VG 354 (1σ), respectively. Chemical separation of common Pb and Sm-Nd (see below) were carried out in the tracer isotope laboratory at the University of Alberta. The total analytical blank for Pb was small (~150 pg) compared to the Pb sample analyzed (~700 ng for MM30, ~150 ng for VG 354), thus no blank corrections were made.

For the Sm-Nd isotopic analyses, whole rock powders were dissolved in 5:1 mixture of 48% HF and 16N HNO_3 and were spiked with mixed ^{149}Sm-^{150}Nd spike prior to the HCl conversion. Samples were subsequently passed through cation and HDEHP chromatography to separate Sm and Nd. Nd isotopic ratios were measured using a VG 354 mass spectrometer in multidynamic mode. All ratios were normalized to $^{146}Nd/^{144}Nd$=0.7219. The value obtained for La Jolla standard during this study was

0.511848±8, and the in-house "Nd oxide" standard gave an external reproducibility of ±0.000016 (2σ). The total analytical blanks for Sm and Nd were <500 pg.

All major and trace element analyses were performed at Washington State University using XRF and ICP-MS. The details of the analytical procedure are presented in Hooper et al. (1993).

Results

U-Pb geochronology

Basement tonalite (7072)

Sample 7072 was collected from the body northwest of the Hanikahimajuk Lake (Fig. 2.2). Zircon extracted from this sample mainly consists of brown, euhedral to subhedral short prismatic grains with lesser long prismatic and equant grains. Grains containing visible cores are also recognized. For this study, zircon with no visible core and/or inclusions were selected. Multiple-grain zircon fractions showed a range of $^{207}Pb/^{206}Pb$ ages between 3.25-3.35 Ga, which do not define a discordia line (Table 2.1, Fig. 2.3a). However, when single-grain zircon fractions were analyzed, nine out of ten fractions plot along a straight line which yields an upper intercept age of 3377.7 +1.4/-1.3 Ma (Fig. 2.3a). This age is interpreted to represent the age of igneous crystallization of the basement tonalite. A small amount of scatter in the data, even with single zircon analyses, may be an effect of multiple Pb loss events. One single grain gave a significantly younger $^{207}Pb/^{206}Pb$ age of 3255 Ma (0.87 % discordance). Whether this age represents a secondary zircon growth event or is a result of early Pb loss is uncertain as U-Pb ages of other accessory minerals such as titanite are not available for this sample. However, the Th/U ratio of this zircon (0.4) is similar to those of other zircon grains and it does not exhibit the low Th/U ratio typical of metamorphic zircon (Heaman et al. 1990). For this reason, it seems more reasonable to interpret that this zircon has experienced an ancient Pb loss, but was protected from the subsequent (more recent?) Pb loss which some of the 3377.7 Ma zircon has experienced.

Syn-volcanic diorite (7078)

Sample 7078 was collected from the body of ~2.68 Ga diorite southeast of the ABVB (Fig. 2.2). Zircon from this sample consists of light brown to dark brown fragments of large crystals with abundant fractures. Judging from the shape of the fragments, the original morphology of the zircon was short to long prismatic. Four multi-grain fractions and one fraction of a large single grain define a discordia line which gives an upper intercept age of 2679.2 ± 1.1 Ma (Fig. 2.3b). This result is in accord with the preliminary U-Pb age of Gebert (1995) and Jensen (1995) and confirms that they are syn-volcanic (syn-YKSG) in age.

Rim Granite (7082=1510)

The rim granite is a granite surrounding the southern margin of the ABVB (note: sample 7082 is from the same outcrop as sample 1510 but was collected during a different field season). Zircon from this granite shows a wide spectrum in size, clarity and color. The most abundant type of zircon is colorless to yellow, clear, euhedral to subhedral grains that are magnetic at >10° tilt (1.8 A) on the Frantz isodynamic separator. Zircon fractions that are magnetic/non-magnetic at 5° tilt seem to contain higher proportion of pink to brown grains. As shown in Table 2.1 and Figure 2.3c, zircon fractions with the highest U concentration (975-2014 ppm) are the colorless clear and milky fractions. With the exception of one fraction (#17), these grains are highly discordant and give scattered $^{207}Pb/^{206}Pb$ ages ranging from 2.50 to 2.58 Ga. Compared to the clear zircon grains, pink and brown grains are typically lower in U content (94-492 ppm) and are less discordant. The $^{207}Pb/^{206}Pb$ age of these zircon fractions cluster at ~2.67 Ga.

Although the colorless clear and the pink/brown fractions show distinct range of $^{207}Pb/^{206}Pb$ age and U-Pb content, it seems probable that they were originally crystallized from the same magma since their Th/U ratios are broadly similar (>0.5) and one clear grain (#17) shows geochemical and geochronological characteristics that are similar to the purple fractions. The possible explanation for the high U content and low $^{207}Pb/^{206}Pb$ age in the clear fractions rather than the pink/brown fractions is the partial (or complete) Pb loss and annealing of high U metamict zircon during the ca. 2.6 Ga regional metamorphism. We therefore interpret the upper intercept age of discordia line defined by five near concordant pink to brown zircon fractions as the best estimate for the crystallization age of the Rim granite (Fig. 2.3d). The age 2676.5 +4.1/-3.6 Ma is significantly older than the age of syn- to post-deformation granitoids in the Slave province (2.58~2.62 Ga). Rather, the age seems to be in better agreement with the age of syn-volcanic felsic granitoids from the central Slave province (Davis et al. 1994).

Gabbro of unknown age (6161)

One gabbro of unknown age with no deformational fabric was collected from the western margin of the PLVB. This sample contains abundant (~5%) titanite which is interstitial between hornblende and clinopyroxene crystals along with alkali feldspar. Petrographical characteristics of the titanite indicate that it is a late magmatic titanite rather than metamorphic. Two fractions of brown colored titanite are concordant at 2574.7 ± 1.4 Ma (weighted mean $^{207}Pb/^{206}Pb$ age; Fig. 2.3e). This age is slightly younger than the U-Pb monazite age of the two-mica granitoids (2582 Ma) from the Contwoyto Lake area, central Slave province (van Breemen et al. 1992). Since the closure temperature for the titanite is slightly lower (~600°C) than that of monazite (700°C; see Heaman and Parrish 1991), the magmatic activity of this gabbro and the two-mica granitoids may be related to the same late-Archean tectonic event, with the age of both titanite and monazite representing a cooling age.

Felsic volcanic rock from Izok Lake volcanic belt (MSD14552)

One felsic volcanic rock from the northeast trending felsic-dominated *Izok Lake volcanic belt* (ILVB), approximately 50 km south of the ABVB, was collected to establish the regional geochronological framework of the PLVB. Zircon fractions separated from this sample mainly consists of pink, euhedral to subhedral, short prismatic grains or fragments of such grains. Four multi-grain fractions from this sample are all near concordant (<2% disc.), defining a discordia line which yields an upper intercept age of

2681 +7/-3 Ma (Fig. 2.3f). This age is nominally younger than the age of felsic volcanic rock from the ABVB (2694 +7/-5 Ma; Mortensen et al. 1988), but is within the range of typical YKSG volcanic activity (e.g. Villeneuve and van Breemen 1994, Isachsen and Bowring 1994). The similar ages of felsic volcanic rocks from the two belts (ABVB, ILVB) indicate that felsic dominated volcanic activity in the PLVB is closely related.

Major/trace element geochemistry and Nd-Pb isotopic analyses

The results for the major/trace element analyses, Sm-Nd and Pb isotopic analyses are given in Tables 2.2, 2.3 and 2.4, respectively. For convenience, the initial εNd (εNd_T) of the volcanic rocks and diorites are calculated at 2.68 Ga. The small difference between the age of syn-volcanic diorites and at least some of the volcanic rocks (2694 Ma) has a minimal effect on the calculation of the εNd_T. The εNd_T values of the sedimentary rocks and syn- to post-deformation granitoids (i.e. all granitoids except Rim granite) were calculated at 2.66 Ga and 2.60 Ga; a typical deposition age and crystallization age of the YKSG sedimentary rocks and syn- to post-deformation granitoids, respectively (e.g. van Breemen et al. 1992, Isachsen and Bowring 1994, Davis et al. 1994; see below for the details of the felsic granitoids).

The basement tonalites are characterized by low K_2O/Na_2O, Rb/Sr, and Yb_N ($_N$ refers to chondrite normalized value). These weakly peraluminous granitoids (A/CNK=1.00-1.16) are classified as tonalite to monzogranite using the classification scheme of Le Maitre (1989). The REE patterns of these rocks are highly fractionated (La_N/Yb_N=9-78) and show no negative Eu/Eu* (Fig. 2.4d). The εNd_T values of these tonalites range from -0.3 to +2.1 with TDM ranging from 3.4 to 3.6 Ga (Table 2.3). The εNd values of these rocks at 2.68 Ga (the age of the YKSG volcanism) range from -4.3 to -10.5; this value is significantly lower than those of the YKSG supracrustal rocks and syn-volcanic plutons (see below).

The mafic volcanic rocks (SiO_2=50~56%) of the YKSG can be classified into two different groups based on their rare earth element (REE) patterns (Fig. 2.4a; Jensen 1995). Those with flat REE patterns (Type I) are tholeiitic in composition and are characterized by low Th, Zr, Hf, and Ti. On the other hand, mafic volcanic rocks with an enriched light rare earth element (LREE) pattern (Type II) are calc-alkaline with higher Th, Zr, Hf, and Ti. Only one type II mafic volcanic rock was analyzed in this study, but the geochemical study of Jensen (1995) has shown that type II mafic volcanic rocks are widely represented throughout the map area, sometimes in close proximity to the type I mafic volcanic rocks.

The intermediate volcanic rocks (SiO_2=58~65%) of the ABVB have trace element characteristics similar to the type II mafic volcanic rocks. They are calc-alkaline with a LREE enriched pattern without pronounced negative Eu/Eu*. The abundance of other incompatible trace elements such as Th, Zr, Hf and Y are also broadly similar to the type II mafic volcanic rocks (Table 2.2). The felsic volcanic rocks (SiO_2=77~80%), on the other hand, have much higher REE contents and pronounced negative Eu/Eu* of 0.3-0.6 (Fig. 2.4c).

The εNd_T values of YKSG volcanic rocks are highly variable, ranging from -1.2 to +2.5 for mafic volcanic rocks, +0.1 to +1.3 for intermediate volcanic rocks and -0.6 to +0.6 for felsic volcanic rocks. These values are significantly higher than $\varepsilon Nd(2.68)$ of the basement tonalites but are lower than the value of model depleted mantle at 2.68 Ga (Fig. 2.5). The high εNd_T of some mafic volcanic rocks clearly indicate an input of magmas derived from a depleted mantle source. However, the systematic decrease in the εNd_T range from mafic to felsic volcanic rocks, suggest that these rocks were, at least in part, contaminated by the pre-YKSG basement, and the role of basement was greater in the genesis of more felsic rocks (Fig. 2.5).

The syn-volcanic diorites are tholeiitic to calc-alkaline and are classified as quartz-monzodiorite(gabbro) to quartzdiorite(gabbro) using the classification scheme of Le Maitre (1989). Although they are slightly more mafic in composition (SiO_2=46~59%) compared to the intermediate volcanic rocks, the geochemical characteristics of these rocks are generally similar to intermediate and type II mafic volcanic rocks. With the notable exception of two samples (5609 and 5719), these diorites show LREE enriched pattern with no negative Eu/Eu* (Fig 2.4e).

Although grouped here as syn-volcanic diorite, sample 5719 (SiO_2=50%) is an amphibolite sample collected from west of PLVB. It shows flat REE pattern with no negative Eu/Eu*. The REE pattern of this sample is similar to those of type I mafic volcanic rocks and thus this sample may be genetically related to type I mafic volcanic rocks rather than type II. The sample 5609 is also characterized by flat REE but with pronounced positive Eu/Eu*. The presence of large positive Eu/Eu* together with high Sr and Ca contents for the SiO_2 content (46%) likely indicate an accumulation of plagioclase in this sample. The εNd_T and TDM of syn-volcanic diorites range from -2.2 to 0.8 and 3.0 to 3.2 Ga, respectively. These values are also broadly similar to those of the intermediate YKSG volcanic rocks.

The YKSG metasedimentary rocks of the ABVB used in this study are greywacke (6066) and quartzite (6106). Both of these sediments are characterized by LREE enrichment (La_N/Sm_N=~5), relatively flat HREE (Heavy REE; Ga_N/Yb_N=1.3~1.6) and positive or negative Eu/Eu* (Fig. 2.4f). The εNd_T of these sedimentary rocks are -5.3 and -0.9, respectively.

The felsic granitoids in the Hanikahimajuk Lake area can be classified into three groups (Type I, II and Rim granite) based on their age and major/trace element characteristics. The 2677 Ma Rim granite is felsic (SiO_2=78%) and is characterized by low Al_2O_3, P_2O_5, Sr, high Nb, U, Th, ΣREE and relatively flat REE pattern with large negative Eu/Eu* (Fig. 2.6a). These geochemical features are similar to those of felsic volcanic rocks of the ABVB and low Al_2O_3 syn-volcanic granitoids (Wishbone and Gondor suite) in the Contwoyto Lake area, central Slave province (Davis et al. 1994). The εNd_T of this sample is +1.0.

Since no U-Pb ages for the felsic granitoids other than Rim granite are available, the classification of type I and II granitoids in the Hanikahimajuk Lake area is based on their geochemical similarities with the late-Archean granitoids from central and southwestern Slave province. Type I granitoids are equated with ~2.61 Ga Concession suite granitoids (central Slave province) and Defeat suite granitoids (southwestern Slave province; Davis et al. 1994, Yamashita et al. 1998). They are characterized by low

K_2O/Na_2O, Rb/Sr and fractionated REE pattern with no negative Eu/Eu* (Fig. 2.6b). These granitoids are weakly peraluminous and are classified as tonalite to granodiorite (Le Maitre 1989).

Only one type II granitoid was analyzed for major/trace element geochemistry and Nd-Pb isotopes in this study. Type II granitoids have higher Th, U, K_2O/Na_2O, Rb/Sr compared to type I granitoids and shows a LREE enriched pattern with pronounced negative Eu/Eu* (Fig. 2.6c). Geochemically, this granitoid appears to be similar to the Yamba suite granitoids in the central Slave province and Awry/Stagg suite granitoids in the southwestern Slave province (Davis et al. 1994, Yamashita et al. 1998). The εNd_T values of type I and II granitoids range from -2.2 to +0.8 with TDM of 2.9 to 3.1 Ga.

The Pb isotopic compositions of the leached feldspars from basement tonalites, syn-volcanic diorite and syn- to post-deformation granitoids are shown in Figure 2.7. The Pb isotopic signature of leachates and residues for the syn-volcanic diorite and syn- to post-deformation granitoids plots parallel to the 2.68 Ga and 2.60 Ga reference isochron, respectively, indicating that Pb in these samples has behaved as a closed isotopic system since crystallization. These Pb isotopic data generally overlap with the radiogenic Pb isotopic composition of southern and western Slave VMS deposits and syn- to post-deformation granitoids from the western Slave province (Fig. 2.7b; Thorpe et al. 1992, Davis and Hegner 1992). However, the leachates and residues of feldspars from basement tonalites do not plot along the 3.38 Ga reference isochron. The slope of the line defined by these data is broadly similar to the slope of a 2.9-3.0 Ga secondary isochron, suggesting that a post-crystallization event(s) has disturbed the Pb isotopic system of these samples. The Pb isotopic signatures of the basement tonalites must therefore be taken as the maximum value for their initial Pb isotopic signature, if not their isotopic signature at ca. 3.0 Ga.

Discussion

Geochemical and isotopic constraints for the origin of basement tonalites.

Little is known about the origin of pre-YKSG basements in the Slave province as they are exposed in limited locations, and most of the previous studies have focused on the geochronological and geographical aspects of these rocks. Although only four samples of the pre-YKSG basement were collected for this study, their major/trace element characteristics and Nd isotopic data impose important constraints on their origin. The combination of a highly fractionated REE pattern (i.e. high La_N/Yb_N), low Yb_N and lack of a pronounced negative Eu/Eu* even in the most felsic sample precludes the involvement of feldspar as a major fractionating and/or residual phase during magmatic evolution (Fig. 2.4d). Plutonic rocks with such geochemical features are found in most of the Archean granite-greenstone terranes and are thought to represent the product of dehydration melting of altered basalt (Archean TTG=tonalite, trondhjemite and granodiorite; Martin 1986, 1994). At >0.8 GPa, the residual phases after partial melting of mafic protoliths include garnet + clinopyroxene ± amphibole with little or no plagioclase (Rapp et al. 1992). The involvement of garnet rather than plagioclase as a residual phase explains the depletion of HREE and the lack of negative Eu/Eu*. The melting of mafic protoliths can take place either during the subduction of oceanic lithosphere, or during the crustal thickening where the pre-existing mafic volcanic rocks within a continental crust are carried to a depth of >30 km. Although it is difficult to distinguish between the two possibilities at this stage, the lack of evidence for the existence of large mafic volcanic successions in the Slave province prior to 2.8 Ga may favor the former. An important implication here is that regardless of which mechanism was responsible for the generation of these tonalites, a mafic crust older than 3378 Ma is required. This is in accord with the εNd_T values of these tonalites which are well below the model depleted mantle value, ranging down to -0.3.

The second important point is the similarity between the crystallization age of these tonalites and (1) the metamorphic age of the Acasta grey tonalitic gneisses (ca. 3360 Ma; Bleeker and Stern 1997), (2) the crystallization age of granitoids that intrude gneisses near the Acasta grey gneisses (3348 +43/-34 Ma; Bleeker and Stern 1997), (3) the younger metamorphic age of the Acasta banded gneisses (3382 ± 8 Ma; Bleeker and Stern 1997) and (4) the crystallization age of the Kangguyak gneisses in the northern Slave province

(3368 +5/-4 Ma; Villeneuve Pers. Comm.). This similarity in ages may suggest that the generation of ~3.38 Ga tonalites and tonalitic gneisses in the central Slave province is related to the metamorphic event which disturbed the U-Pb systematics of the Acasta gneisses. If so, it would indicate that the ~4.0 Ga Acasta gneisses and <3.4 Ga basement in the central Slave province were part of large crustal block, rather than unrelated crustal fragments tectonically juxtaposed at a later time.

Constraints for the tectonic origin of YKSG volcanic rocks and syn-volcanic diorites

Determining the origin of the YKSG volcanic rocks is problematic since many of the trace-element ratios used to discriminate tectonic environments use large ion lithophile elements (LILE) which are highly mobile during metamorphism and alteration (c.f. Wilson 1989). Furthermore, because the trace element characteristics of Archean mantle (other than REE) is not fully understood, direct application of tectonic discrimination diagrams constructed using the geochemistry of Phanerozoic volcanic rocks may not be an appropriate approach. Nevertheless, it is possible to provide some constraints on the tectonic setting of these volcanic rocks by (1) using immobile elements such as high field strength elements (HFSE) and REE and (2) assuming that the geochemical characteristics of the mantle sources (which in turn should constrain the geochemistry of the magma itself) have developed through time from primitive mantle composition early in Earth's history.

Shown in Figure 2.8 are the trace element abundances of the YKSG mafic to intermediate volcanic rocks and syn-volcanic diorites normalized to primitive mantle (PM; Taylor and McLennan 1985). On this diagram, the incompatibility of elements decreases from left to right. Th is included in this diagram because it is relatively immobile (Saunders and Tarney 1984). Also shown in this figure is the normalized trace element abundance of mafic volcanic rocks produced in different tectonic settings (Sun 1980, Chen and Frey 1983, Davies and Macdonald 1987, Hofmann 1988).

Type I mafic volcanic rocks are characterized by a relatively flat pattern at x4 to x10 PM values. They are slightly depleted in highly incompatible elements (such as Th and Ta), but show no depletion in Ta and Nb with respect to Th and Ce (Fig. 2.8a). These geochemical features are similar to modern N-type mid-oceanic ridge basalt (N-MORB) (Fig. 2.8d). A contradictory conclusion, however, is derived from Nd isotopic signatures of these rocks. Shown in Figure 2.10 are the εNd(2.68) values of YKSG volcanic rocks and basement tonalite plotted against $^{147}Sm/^{144}Nd$ ratio. On this diagram, the type I mafic volcanic rocks form a clear linear trend which points towards the data for the basement tonalite. Such a linear trends can be interpreted either as (1) an isochron which represents the age of mafic volcanic rocks and fortuitously projects to the basement values or (2) a mixing line between mafic volcanic rocks and basement tonalite. The former possibility is not considered viable because the age calculated from the slope of this line is ~750 Ma at 2.68 Ga (i.e. 3.4 Ga), which is far too old to represent the age of YKSG volcanic rocks. Mixing between the basement tonalite and type I mafic volcanic rocks, on the other hand, is supported from the observation that the εNd_T of the mafic volcanic rocks correlates strongly with the shift in geochemical characteristics (such as Nb/Th and Y/Th ratios) towards that of the basement tonalite (Fig. 2.8a and 2.9). Since the interaction between mafic volcanic rocks and basement tonalite is unlikely to take place in a mid-oceanic ridge setting, an alternative candidate for the tectonic origin of type I mafic volcanic rocks is necessary.

It is known from the study of igneous rocks associated with the spreading of back-arcs that intrusive and effusive rocks produced in such tectonic setting have trace element characteristics ranging from those of N-MORB to island-arc tholeiitic and calc-alkaline basalts (Saunders and Tarney 1984, Miyake 1994). In some cases, felsic volcanic rocks with SiO_2 ~70% are also found (Weaver et al. 1979, Miyake 1994). Such a tectonic setting seem to be the best candidate for type I mafic volcanic rocks because it can produce MORB-like volcanic rocks that are variably contaminated by the pre-existing crust. However, even this tectonic setting is not fully compatible with the geochemical data because mafic volcanic rocks produced in an "ensialic" back-arc typically have trace element characteristics intermediate between the N-MORB and island-arc basalts (e.g. moderate Ta and Nb depletion; Weaver et al. 1979, Saunders and Tarney 1984, Miyake 1994). The reason for the lack of Ta and Nb depletion in the type I mafic volcanic rocks remains uncertain at this stage. However, one possibility may be the destabilization of important HFSE hosting minerals often produced in a modern subducted slab (e.g. rutile; see Izuka and Nakamura 1995) due to a higher Archean geotherm. If the occurrence of rutile in the subducted slab is the direct reason for the depletion of HFSE in the modern island arc basalt, destabilization of such a mineral may release part of the HFSE in slab and consequently, suppress the Nb and Ta depletion in the Archean island-arc and back-arc magmatism.

Type II mafic volcanic rocks are characterized by enrichment of highly incompatible elements (Th, Ta, Nb etc.) with depletion of Nb and Ta with respect to Th and Ce, and Ti with respect to Sm and Y (Fig. 2.8b). Although none of the patterns shown in Figure 2.8d matches exactly with the pattern of type II mafic volcanic rocks, relative depletion in Nb and Ta with respect to Th and Ce, and Ti with respect to Sm and Y seems to be a characteristic feature of calc-alkaline basalts produced in a modern island-arc setting. Such a tectonic setting is appealing because it is compatible with the existence of (1) calc-alkaline intermediate volcanic rocks with similar geochemical characteristics (Fig. 2.8b) and (2) type I mafic volcanic rocks likely produced in a back-arc setting. The magnitude of Nb depletion in the type II mafic rocks is much smaller compared to that of modern island-arc basalt. This may be related to a higher Archean geotherm as previously discussed.

For the same reason previously discussed, an alternative tectonic setting for the origin of type II mafic volcanic rocks is the ensialic back-arc. In the case of Hanikahimajuk Lake area, this possibility may actually be more favorable as type I and II mafic volcanic rocks are found in close proximity in the field.

Intermediate volcanic rocks and syn-volcanic diorites are geochemically identical and show features that are similar to type II mafic volcanic rocks (Fig. 2.8c). This indicates that (1) they are likely felsic derivatives of the type II mafic volcanic rocks and (2) the syn-volcanic diorites are the plutonic equivalent of type II mafic and intermediate volcanic rocks. The observed range of εNd_T is also consistent with this hypothesis.

The felsic igneous rocks with major element composition similar to the Rim granite and the felsic volcanic rocks of the ABVB can be generated through dehydration melting of mafic rocks at relatively low pressure (<0.1 GPa), where plagioclase is a major residual phase (Beard and Lofgren 1989, Thy et al. 1990). In this scenario, however, the melt produced through dehydration melting must be further contaminated by the older basement and/or the mafic protoliths must be >300 Ma older than the felsic volcanic rocks (assuming the Sm/Nd ratio of the mafic protolith is similar to type II mafic volcanic rocks) in order to account for the lower εNd_T value of the felsic rocks. Although both of these possibilities cannot be ruled out, the latter possibility alone is unsatisfactory as a large quantity of pre-3.0 Ga mafic volcanic rocks has not been documented from the Slave province.

Alternatively, it may be argued that extreme fractionation of basaltic to andesitic magma could produce a melt similar in composition to the felsic volcanic rocks. However, such a possibility seems unrealistic because mass balance calculation an using incompatible and immobile element such as Th would indicate that a "minimum" of 65 to 85 wt. % of the original andesitic magma must be removed by fractional crystallization even to produce the rhyolite with lowest Th concentration. Evidence for such large scale accumulation of expected fractionating phases (i.e. Plag, Opx, Cpx ± Hbl) has not been reported from the Hanikahimajuk Lake area (Jensen 1995, Gebert 1995). Furthermore, the lack of samples with SiO_2 in the 65-77% range favors the possibility that these felsic volcanic rocks were generated through partial melting and contamination by basement rather than by fractional crystallization.

Source of the YKSG sedimentary rocks

The unique geochemical characteristics of these sedimentary rocks compared to most of the sedimentary rocks from the Archean greenstone belts provide critical information on the nature of the exposed crust at the time of sediment deposition. The most important feature of the Archean sedimentary rocks is the general lack of pronounced negative Eu/Eu* anomaly (Taylor and McLennan 1985, McLennan 1989). This has been interpreted as evidence for a fundamental change in the chemical composition of the upper crust at the Archean/Proterozoic boundary (Taylor and McLennan 1985). Because negative Eu/Eu* is often produced during intracrustal processes such as partial melting and fractional crystallization involving feldspars, the lack of Eu/Eu* in the Archean sediments could indicate that such processes were less important in the Archean compared to the post-Archean. The sedimentary rocks from the ABVB, however, clearly show Eu anomalies and their REE abundances are virtually indistinguishable from those of typical post-Archean sediments. Because the sedimentary rocks with "typical Archean REE signatures" are also widely available in some areas of the Slave province such as the Yellowknife area (Jenner et al 1981), the existence of sediments with Eu anomalies in the ABVB indicate that the chemical composition of the exposed crust was variable even within the 2.6~2.7 Ga YKSG greenstone belt and moreover, intracrustal processes involving plagioclase were important, at least in certain areas of the Slave province.

The second important point to notice is the lower εNd_T values (Fig. 2.5) of these sedimentary rocks compared to YKSG volcanic and syn-volcanic plutonic rocks. Since both of these sediments show negative Eu/Eu*, the sedimentary protoliths must include felsic volcanic rocks. Although a parallel shift in REE pattern may result from addition of a mafic volcanic component and/or dilution by quartz, an additional component with low HREE, low εNd(2.66) and no negative (or positive) Eu/Eu* is required to fully explain the geochemical feature of these sediments. The rock type which satisfies all these requirements is the basement tonalite. This indicates that basement tonalites were, at least in part, exposed at the time of sedimentary deposition.

Genesis of late-Archean felsic granitoids

A detailed geochemical and isotopic study of 2.62-2.58 Ga syn- to post-deformation granitoids from central and southwestern Slave province was carried out by Davis et al. (1994) and Yamashita et al. (1998), respectively. The felsic granitoids surrounding the Hanikahimajuk Lake area exhibit similar geochemical features to the granitoids in central and southwestern Slave province, and thus the genesis of these granitoids is thought to be similar (Fig. 2.6b,c).

Davis et al. (1994) suggested that the parental magmas of the ~2.61 Ga syn-deformation granitoids are derived from an enriched mantle source, and the felsic derivatives were produced through assimilation-fractional crystallization processes. However, it has been pointed out by Yamashita et al. (1998) that such processes could only account for the geochemical features of the mafic endmember of syn-deformation granitoids, and an alternative mechanism is required to produce the intermediate to felsic members. As discussed above, tonalite-trondhjemite-granodiorite melts with a steeply-fractionated REE pattern, and no Eu anomaly can be produced through dehydration melting of a mafic protolith at high pressure (>0.8 GPa) where garnet, rather than plagioclase, is a major residual phase (Martin 1986, 1994, Rapp et al. 1992). Such a mechanism, however, cannot produce the mafic portion of the syn-deformation granitoids. This led Yamashita et al. (1998) to propose that both of the above mechanisms are responsible for the generation of syn-deformation granitoids. Dehydration melting of mafic protoliths (i.e. mafic volcanic rocks of the YKSG and possibly part of syn-volcanic diorites), in this case, most likely occurred in response to their descent to a depth of >30 km during the ~2.6 Ga regional deformation and crustal thickening (Yamashita et al. 1998). The $\varepsilon Nd(2.60)$ of YKSG mafic volcanic rocks and syn-volcanic diorites (-2.9~+2.5) overlaps with the $\varepsilon Nd(2.60)$ of late-Archean granitoids (-2.2~+1.5), being consistent with this hypothesis.

The ~2.59 Ga post-deformation granitoids can be produced in various ways such as hybridization melting of basalt and biotite gneisses/metapelites, partial melting of tonalite and partial melting of metasedimentary rocks (Conrad et al. 1988, Vielzeuf and Holloway 1988, Rutter and Wyllie 1988, Patino Douce and Johnson 1991, Patino Douce 1995, Patino Douce and Beard 1995, Singh and Johannes 1996, McCarthy and Patino Douce 1997). The hybridization melting model is not viable in this case as there is no evidence for large scale mafic igneous activity in the Slave province at ~2.59 Ga. Partial melting of tonalite, on the other hand, is an ideal mechanism for the generation of post-deformation granitoids because these rocks are available throughout the Slave craton at ~2.59 Ga. However, in the Contwoyto Lake area of the central Slave province, the ~2.61 Ga tonalitic rocks with positive εNd_T values (i.e. Concession suite) are intruded by the ~2.59 Ga post-deformation granitoids with negative εNd_T values (i.e. Concession suite; Davis et al 1994). Therefore, in order to produce post-deformation granitoids through partial melting of tonalitic protoliths, ~2.7 Ga syn-volcanic and/or pre-2.8 Ga basement tonalites must also be included in the tonalitic source. The model that seems to explain well the geochemical and isotopic features of the ~2.59 Ga granitoids on a regional scale is the partial melting of greywacke and pelite (Yamashita et al. 1998). In this model, the isotopic signature of older basement is first incorporated into YKSG volcanic rocks, which in turn are weathered to form the sedimentary rocks of the YKSG. Although it has been shown that YKSG sedimentary rocks are dominated by volcanic sources (Jenner et al. 1981), direct input of basement rocks may also occur in places where there is strong evidence for the existence of exposed basement such as in the Hanikahimajuk Lake area and Point Lake area (Schärer and Allègre 1982, this study).

It is also noteworthy that the similarity in Pb isotopic signature between syn-volcanic diorite and syn- to post-deformation granitoids is consistent with the models proposed above. As previously discussed, syn-volcanic diorites are thought to be the plutonic equivalent of type II mafic and intermediate volcanic rocks. Thus if the mafic rocks were the main source of syn-deformation granitoids, the isotopic signature of the syn-deformation granitoids and syn-volcanic diorites should be broadly similar. Furthermore, if part of these mafic to intermediate rocks are weathered to form the YKSG sediments which later become the source of the post-deformation granitoids, the isotopic signature of the post-deformation granitoids should also be in the same range. Although only a few Pb isotopic data are presented in this study, their range is clearly consistent with this model.

Implications for the tectonomagmatic evolution of the Slave province.

Perhaps the most critical finding of this work is the strong evidence for the existence of pre-YKSG basement at the time of YKSG deposition. This point was brought up by several authors based on limited and/or unpublished data (Dudas 1989, Isachsen and Bowring 1994, 1997). The Nd isotopic and trace element data clearly indicate that both the back-arc type (type I) and the island-arc type (type II) magmas were contaminated by ~3.4 Ga basement rocks during the ~2.7 Ga tectonomagmatic event that produced the YKSG rocks. Although the stratigraphic relationship between the two magma types are not fully understood in the Hanikahimajuk Lake area, the existence of back-arc (or MORB) type igneous rocks overlain by island-arc type volcanic and plutonic rocks, both of which are contaminated by older basement, are also documented in the Yellowknife area of the southern Slave province (Fyson and Helmstaedt 1988, Cunningham and Lambert, 1989, Isachsen et al. 1991, Isachsen and Bowring 1994, 1997, MacLachlan and Helmstaedt 1995). Furthermore, the ~2.67 Ga volcanic rocks of the Indin Lake supracrustal belt (south of Grenville Lake area, Fig. 2.1) also show similar geochemical features (Pehrsson 1997, personal communication). Such regional similarities may indicate that the tectonic evolution of the YKSG supracrustal rocks in the west-central Slave province was similar, regardless of their latitude.

The existence of "crustally contaminated" volcanic rocks and syn-volcanic plutons forces a re-examination of the model which argues for tectonic juxtaposition of pre-YKSG basement with the "juvenile" island arc (Kusky 1989). It is clear from the data presented here that most, if not all, of the YKSG supracrustal rocks and syn-volcanic diorites were generated on or at least in the vicinity of the pre-YKSG basement.

An alternative important model for the evolution of the western Slave province is the intracontinental rift model of Henderson (1981). This model is appealing in that it can explain the "crustally contaminated" Nd isotopic signature of the YKSG volcanic rocks (Yamashita et al. 1996, 1997). However, two major problems with this model are the lack of (1) alkaline volcanism typically found in continental rift settings (Kusky 1989) and (2) a mechanism to produce volcanic and plutonic rocks with an island-arc like signature. In addition, the geochemistry of syn-deformation granitoids (Defeat granitoids) to the west of Yellowknife greenstone belt is compatible with dehydration melting of mafic protoliths, which argues for the existence of mafic igneous rocks at a depth west of Yellowknife. This model may therefore require an additional mechanism to produce MORB (or back-arc) and island-arc like volcanic rocks within and west of the Yellowknife volcanic belt simultaneously and/or subsequent to rifting.

Although we disagree with many interpretations of their geochemical data, the model that seems to fit in relatively well with the data in this study is that of MacLachlan and Helmstaedt (1995). In their model, the consolidation of late-Archean crust in the western Slave province began with rifting of pre-existing basement (Anton granite) which led to a development of oceanic crust (Chan Formation). This was followed by a decrease of extension, and the oceanic crust was subsequently overlain by arc-like volcanic rocks in response to the eastward subduction of the oceanic lithosphere (upper Kam group and Banting Group; Fig. 11 of MacLachlan and Helmstaedt 1995). However, two major points that need to be reconsidered in their model are (1) the transition from continental crust to oceanic crust was probably not abrupt as they suggested since such a mechanism will not explain the low εNd_T values of Kam and Banting Group volcanic and plutonic rocks (MacLachlan and Helmstaedt 1995, Yamashita et al. 1998) and (2) there are probably regions west of Yellowknife which are not underlain by pre-YKSG basement (evident from high εNd_T of some post-deformation granitoids and isotopic study of sedimentary rocks in the vicinity of Yellowknife (Davis et al. 1996, Yamashita et al. 1996, Yamashita et al. 1998; also see chapter 2)). The model of MacLachlan and Helmstaedt (1995) may therefore explain the tectonic evolution of a restricted area of the Slave province (i.e. region between Yellowknife and Sleepy Dragon Complex), but further modification is necessary to explain the history of the entire Slave province.

Although the geochemical and isotopic data available at this stage is clearly not enough to draw a detailed tectonic evolution model of the Slave province, it highlights some of the important points that need to be considered when constructing the model. For example, careful examination of geochronological and isotopic data suggest that the Slave craton cannot be separated simply into older western versus younger eastern Slave province. Rather, the isotopic signature of the older basement seems to lie strongly in two regions; (1) along the volcanic belt which runs from the Arcadia Bay - Napaktulik Lake (Hanikahimajuk Lake) - Point Lake - Winter Lake - Sleepy Dragon/Yellowknife and (2) in the Acasta River - Grenville Lake region (Fig. 2.1; Yamashita et al. 1998). Such complex isotopic structure demands a complex multistage evolution model or perhaps a totally new model not previously proposed. For example, if the two types of YKSG mafic to intermediate volcanic rocks (i.e. back-arc type and island-arc type) are actually both from a back arc setting as previously discussed, then the region bounded by the two sets of pre-YKSG basement may represent a single back-arc basin with the fore-arc being represented, possibly, by the intermediate to felsic dominated volcanic suite of the eastern-northeastern Slave province (i.e. Hackett River arc of Kusky 1989). This possibility, although somewhat speculative at this stage, is appealing even for the genesis of late-Archean syn- to post-deformation granitoids because tholeiitic basaltic rocks and immature sedimentary rocks which typically fill the back-arc basin are ideal sources of these granitoids. Additional geochemical, geochronological and isotopic data together with field observation from the eastern Slave province may provide further evidence for this possibility.

Conclusions

The significance of this study can be summarized as follows.

(1) The εNd_T of the 3378 Ma basement in the Hanikahimajuk Lake area ranges from -0.3 to 2.1, implying an existence of older crust. Major and trace element characteristics of these tonalites are most compatible with melting of basaltic crust at >0.8 GPa.

(2) There are two different types of mafic to intermediate volcanic rocks in the Hanikahimajuk Lake area; one with geochemical features similar to back-arc (or mid-oceanic ridge) mafic volcanic rocks and another with an island-arc like signature. It should, however, be remembered that the volcanic rocks with an island-arc like geochemical features can be generated in an ensialic back-arc setting. Therefore, two different tectonic settings (i.e. island-arc and back-arc) are not required to produce these two volcanic rock types.

(3) The geochemistry and isotopic signature of the 2679 Ma syn-volcanic diorites suggest that they are the plutonic equivalents of the volcanic rocks.

(4) The Nd isotopic signature of the ca. 2.68 Ga volcanic rocks indicate that most, if not all, of the volcanic rocks were built on, or at least in the vicinity of the basement.

(5) The geochemical characteristics of the felsic volcanic rocks indicate their derivation through dehydration melting of a mafic protolith at low pressure (<0.1 GPa). However, their Nd isotopic signature requires further crustal contamination and/or a mafic protolith of >300 Ma older than the felsic volcanic rocks.

(6) The REE patterns of YKSG sedimentary rocks indicate that the felsic volcanic rocks and basement tonalite were the major source of the sediments. The low εNd(2.66) obtained for one sample (-5.3) indicates that at least part of the basement was exposed at the time of sediment deposition.

(7) The late-Archean felsic granitoids in the Hanikahimajuk Lake area were produced through dehydration melting of mafic protolith and sedimentary (and tonalitic) rocks as suggested by Yamashita et al. (1998). The Nd isotopic signature of these granitoids is in accord with this model.

(8) Combined geochemical and isotopic studies of Hanikahimajuk Lake and Yellowknife areas demand a multistage evolution model or a completely new model to explain the tectonic history of the Slave province.

Table 2.1. U-Pb analyses of samples from the Hanikahimajuk Lake area.

Sample/ Fraction	Description[a]	Grains (#)	Weight µg	U (ppm)	Pb rad (ppm)	Pb[b] common (pg)	Th/U[c]	$^{206}Pb/^{204}Pb$[d]	$^{206}Pb/^{238}U$[e] (±1σ)	$^{207}Pb/^{235}U$[e] (±1σ)	apparent ages (Ma)[e] $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	disc.[f] (%)
(Hanikahimajuk Lake)														
Basement tonalite (7072)/														
#30	z, sp, eu, br	20	41.6	243	192	86	0.422	4732	0.6561±0.0011	25.200±0.049	3252	3316	3354	3.89
#33	z, sp, eu, br	3	8.0	102	80	12	0.378	2951	0.6565±0.0012	24.162±0.051	3254	3274	3288	1.34
#34	z, lp, eu, br	35	38.0	263	189	159	0.465	2352	0.5953±0.0010	22.126±0.043	3011	3189	3303	11.8
#35	z, sp, sb, br	22	44.5	249	186	206	0.932	2087	0.6162±0.0010	23.254±0.045	3094	3238	3327	8.79
#36	z, fr, br	40	55.2	267	191	168	0.446	3301	0.5984±0.0010	21.721±0.042	3023	3171	3266	9.31
#41	z, sp, eu, br	1	5.4	122	94	5	0.408	5177	0.6510±0.0013	23.453±0.049	3232	3246	3255	0.87
#42	z, lp, eu, br	1	1.3	274	222	7	0.360	2155	0.6817±0.0014	26.465±0.061	3351	3364	3372	0.79
#43	z, eq, eu, br	1	1.3	333	274	7	0.425	2738	0.6842±0.0014	26.709±0.062	3361	3373	3380	0.74
#44	z, dk, large	1	6.2	181	147	17	0.413	2875	0.6773±0.0012	26.370±0.052	3334	3360	3376	1.60
#45	z, sb, br	1	3.0	326	265	16	0.402	2607	0.6702±0.0013	25.932±0.055	3307	3344	3367	2.27
#46	z, sb, br	1	3.8	365	258	72	0.366	736	0.5962±0.0012	22.525±0.053	3015	3207	3329	11.8
#47	z, sb, br	1	5.5	219	168	9	0.333	5494	0.6506±0.0012	25.122±0.050	3231	3313	3363	5.01
#48	z, sb, br	1	2.8	300	231	19	0.387	1755	0.6448±0.0012	24.747±0.053	3208	3298	3354	5.51
#59	z, br	1	2.6	157	128	6	0.405	3573	0.6811±0.0013	26.365±0.055	3348	3360	3367	0.71
#60	z, br	1	3.4	134	111	23	0.406	873	0.6867±0.0016	26.720±0.069	3370	3373	3375	0.21
Syn-volcanic diorite (7078)/														
#20	z, fr	17	14.6	110	65	9	0.548	5909	0.5143±0.0009	12.947±0.026	2675	2676	2677	0.09
#21	z, eu	3	4.6	74	43	9	0.519	1222	0.5061±0.0013	12.714±0.042	2640	2659	2673	1.50
#28	z, lp, fr, p	50	50.2	57	29	12	0.636	6718	0.4356±0.0008	10.901±0.022	2331	2515	2667	15.0
#29	z, eq, fr, br	1	17.5	306	187	8	0.699	22352	0.5143±0.0010	12.977±0.027	2675	2678	2680	0.23
#40	z, eu~sb, br	27	20.9	111	65	7	0.544	9951	0.5070±0.0010	12.782±0.028	2643	2664	2679	1.59

Rim granite (7082)/														
#17	z, sp, cl	1	1.9	1089	628	10	0.542	6421	0.5015±0.0010	12.537±0.027	2620	2646	2665	2.04
#18	z, sp, br	1	4.7	94	52	9	0.574	1556	0.4743±0.0009	11.880±0.031	2502	2595	2668	7.51
#19	z, sp, mk	1	3.9	975	477	14	0.526	7233	0.4296±0.0008	9.864±0.021	2304	2422	2523	10.3
#26	z, sp, eu, cl	1	3.3	2014	935	12	0.546	13637	0.4046±0.0008	9.646±0.020	2190	2402	2586	18.0
#27	z, eq, eu, dk	1	2.2	492	276	8	0.587	4236	0.4833±0.0011	12.143±0.030	2542	2616	2673	5.95
#50	z, eq, br	1	0.9	461	297	19	0.974	727	0.5125±0.0011	12.896±0.032	2667	2672	2676	0.38
#51	z, eq, p	1	1.2	310	180	4	0.788	2540	0.4807±0.0011	12.077±0.032	2530	2610	2673	6.45
#52	z, eq, p	1	1.2	338	191	4	0.615	3008	0.4838±0.0012	12.140±0.033	2544	2615	2671	5.78
#53	z, cl	1	2.2	1178	493	6	0.514	9605	0.3680±0.0007	8.350±0.018	2020	2270	2503	22.5
#55	z, cl	1	3.6	1218	571	7	0.527	16352	0.4091±0.0007	9.552±0.019	2211	2393	2551	15.7
#56	z, cl	1	1.7	1150	506	6	0.554	8442	0.3832±0.0009	8.824±0.021	2091	2320	2528	20.2
Post-deformation gabbro (6161)/														
#13	t, fr, br	50	1265	217	140	4117	1.219	2077	0.4947±0.0032	11.714±0.076	2591	2582	2575	-0.8
#57	t, fr, br	24	424	233	157	1627	1.475	1883	0.4913±0.0009	11.633±0.023	2576	2575	2575	-0.1
(Izok Lake)														
Felsic volcanic rock (MSD 14552)/														
#1	z, sp, fr, p	21	22.0	257	125	20	0.486	9038	0.5099±0.0009	12.850±0.025	2656	2669	2678	1.01
#2	z, sp, fr, p	20	28.5	242	140	38	0.498	5812	0.5067±0.0009	12.735±0.025	2638	2660	2677	1.78
#9	z, sp, fr, p	35	38.6	234	135	80	0.497	3611	0.5054±0.0032	12.734±0.025	2637	2660	2678	1.86
#11	z, sp, fr, p	17	10.7	357	207	11	0.487	11623	0.5103±0.0009	12.869±0.025	2658	2670	2679	0.97

Note: [a] z, zircon; t, titanite; eu, euhedral; sb, subhedral; fr, fragment; lp, long prismatic; sp, short prismatic; eq, equant; dk, dark, br, brown; p, pink; cl, clear; mk, milky. All zircon abraded prior to analysis. Decay constants used are 1.55125×10^{-10} for ^{238}U and 9.8485×10^{-10} for ^{235}U (Steiger and Jäger 1977).

[b] Total common Pb in sample (initial plus blank Pb).

[c] Model Th/U ratio (estimated from the amount of ^{208}Pb).

[d] After spike and fractionation correction.

[e] After spike, fractionation, blank and initial common Pb correction. Initial common Pb estimated using model of Stacey and Kramers (1975).

Table 2.2. Major and trace element compositions of samples from the Hanikahimajuk Lake area.

Sample	1660	1754	7072	7077	3339	2917	5719	414	5609A	415
Type	1	1	1	1	2	2	2	2	2	2
(wt. %)										
SiO_2	73.31	74.38	64.88	76.00	53.68	53.52	50.49	56.27	46.28	58.59
Al_2O_3	15.09	14.56	18.35	13.32	15.15	17.97	13.33	16.18	14.44	17.13
TiO_2	0.291	0.196	0.520	0.053	1.881	0.907	2.424	1.122	3.400	0.817
FeO^*	1.86	1.35	3.36	0.54	11.81	8.07	14.98	8.69	16.12	6.80
MnO	0.033	0.019	0.061	0.009	0.192	0.125	0.327	0.145	0.258	0.115
CaO	1.9	1.99	3.51	0.50	7.11	7.95	8.60	7.22	10.17	7.42
MgO	0.78	0.50	1.76	0.07	4.96	6.15	5.47	4.56	6.75	4.20
K_2O	2.40	1.42	2.18	5.80	1.22	2.08	1.64	2.03	0.44	1.29
Na_2O	4.25	5.51	5.18	3.69	3.52	3.03	2.49	3.70	2.13	3.52
P_2O_5	0.094	0.067	0.196	0.015	0.479	0.190	0.238	0.090	0.016	0.097
(ppm)										
Ni*	13	13	20	10	58	116	31	23	0	57
Cr*	5	0	13	0	92	214	78	56	12	75
Sc*	3	0	0	0	25	24	44	24	39	15
V*	30	11	39	5	207	124	485	173	384	152
Ga*	16	17	23	13	21	19	22	21	19	18
Cu*	9	0	33	18	44	59	124	22	102	32
Zn*	52	19	69	8	140	101	151	108	142	82
U	0.85	0.44	1.53	1.56	0.86	0.61	0.74	1.18	0.02	1.00
Th	8.9	5.34	6.31	6.77	4.55	2.45	0.84	4.95	0.03	4.82
Hf	3.92	3.28	4.49	1.23	6.50	3.53	3.85	5.09	0.38	4.19
Zr*	146	127	215	47	256	130	141	203	31	167
Pb	10.22	7.66	15.67	8.35	7.78	9.19	16.77	16.00	20.19	7.30
Ba	539	340	485	807	307	302	217	439	59	318
Sr*	180	396	311	102	222	219	134	228	173	229
Rb	90.8	31.8	59.1	99.1	27.8	76.8	33.4	45.0	12.0	36.3
Cs	2.2	0.4	0.8	0.6	0.4	1.8	0.4	0.4	0.4	1.4
Nb	8.0	2.7	7.1	3.2	17.1	9.1	11.1	10.5	2.1	9.3
Y	8.42	2.61	8.35	3.96	40.16	30.58	59.70	27.28	7.51	22.18
Ta	0.74	0.51	0.83	0.74	1.01	0.61	0.82	0.90	0.32	0.98
La	31.54	24.28	25.49	4.62	36.40	19.04	12.61	26.32	1.27	25.44
Ce	49.78	39.49	41.40	7.30	71.98	40.51	35.48	50.73	2.67	48.19
Pr	4.49	3.61	4.01	0.74	8.28	5.17	4.97	5.80	0.45	5.24
Nd	14.80	11.99	14.47	2.81	33.31	22.00	23.10	23.52	2.36	20.74
Sm	2.27	1.72	2.76	0.49	8.37	5.45	7.33	5.45	0.91	4.75
Eu	0.71	0.49	0.87	0.18	2.38	1.51	2.07	1.60	0.92	1.35
Gd	1.61	0.88	2.17	0.68	7.50	5.21	8.34	4.91	1.16	4.15
Tb	0.26	0.11	0.32	0.10	1.32	0.91	1.61	0.81	0.22	0.70
Dy	1.58	0.56	1.80	0.62	7.76	5.61	10.21	5.02	1.43	4.25
Ho	0.32	0.10	0.32	0.12	1.62	1.15	2.14	1.00	0.30	0.84
Er	0.84	0.23	0.82	0.37	4.55	3.15	6.38	2.79	0.88	2.34
Tm	0.12	0.03	0.11	0.05	0.60	0.43	0.92	0.38	0.12	0.32
Yb	0.75	0.21	0.67	0.33	3.66	2.66	5.67	2.39	0.72	1.92
Lu	0.13	0.03	0.10	0.06	0.56	0.41	0.91	0.38	0.12	0.31

Notes: * Analyses performed using XRF, all others using ICP-MS; Major element analyses are normalized on a volatile free basis; FeO^*=total iron as FeO; A/CNK=molar $Al_2O_3/(CaO+Na_2O+K_2O)$; Rock type 1, basement tonalite; 2, syn-volcanic diorite; 3, felsic volcanic rock; 4, intermediate volcanic rock; 5, mafic volcanic rock; 6, metasedimentary rocks; 7, syn- to post-deformation granitoids/gabbro

Sample	2919	7078	1295	2767	1655	1666	5132	7080	1719	2484
Type	2	2	3	3	3	3	3	4	4	4
(wt. %)										
SiO_2	52.56	57.83	77.48	77.42	79.99	76.87	77.06	65.86	58.19	64.11
Al_2O_3	14.70	16.03	12.25	12.12	11.22	12.54	12.18	13.81	16.27	14.34
TiO_2	2.294	0.845	0.110	0.172	0.114	0.189	0.248	1.441	1.259	1.447
FeO^*	13.61	7.48	1.46	2.47	0.83	1.75	2.12	6.25	8.32	6.30
MnO	0.212	0.128	0.046	0.062	0.032	0.018	0.041	0.154	0.153	0.106
CaO	7.76	7.81	0.72	1.50	1.21	0.67	2.60	6.72	6.88	4.51
MgO	4.4	4.86	0.38	0.54	1.41	0.35	0.28	0.96	3.70	3.38
K_2O	1.09	1.29	6.51	1.75	1.77	3.22	1.63	0.15	0.72	0.07
Na_2O	3.04	3.61	1.03	3.96	3.42	4.38	3.79	4.36	4.24	5.40
P_2O_5	0.336	0.109	0.015	0.015	0.014	0.009	0.041	0.286	0.261	0.335
(ppm)										
Ni*	1	57	13	8	11	10	14	38	31	54
Cr*	20	96	0	0	0	0	1	86	18	64
Sc*	29	31	8	3	7	4	3	17	20	16
V*	398	147	6	5	12	6	15	170	192	170
Ga*	25	19	22	18	14	16	19	16	21	18
Cu*	46	6	8	15	2	12	22	24	23	26
Zn*	109	45	41	59	29	12	28	57	86	78
U	0.75	0.82	7.29	4.37	4.41	3.96	5.94	0.67	1.24	0.90
Th	3.08	4.21	18.54	14.49	18.15	17.02	19.57	2.24	4.60	3.67
Hf	2.78	3.84	5.49	9.99	5.71	12.13	7.60	7.74	5.48	4.92
Zr*	103	137	141	355	158	369	249	374	209	203
Pb	3.75	2.93	18.52	8.07	3.87	8.51	16.58	5.52	6.79	5.52
Ba	189	326	846	255	263	1046	760	48	222	22
Sr*	205	201	44	50	65	53	144	175	261	337
Rb	34.5	27.9	196.8	43.6	51.3	59.0	32.8	3.1	20.0	1.5
Cs	0.7	0.6	3.4	0.5	0.7	0.2	0.6	0.1	0.1	0
Nb	15.7	8.3	27.0	27.0	15.4	38.3	23.9	13.4	12.1	14.2
Y	41.96	27.52	75.65	54.21	38.01	79.71	61.51	26.01	31.08	29.45
Ta	1.25	0.74	2.30	1.92	1.42	2.55	1.88	0.89	0.86	1.00
La	21.93	17.75	41.46	57.63	38.26	88.00	63.17	17.99	27.14	25.60
Ce	47.28	38.09	85.10	111.94	71.90	185.30	118.62	37.63	52.46	52.88
Pr	5.91	4.65	10.04	12.73	7.84	20.56	12.84	4.70	5.98	6.23
Nd	26.64	19.73	40.16	49.94	29.85	76.16	49.01	20.14	24.07	26.52
Sm	7.33	4.77	10.57	11.56	6.31	17.17	11.23	5.04	6.33	6.47
Eu	2.19	1.35	1.05	2.10	0.80	2.87	1.69	1.70	1.84	2.13
Gd	7.31	4.51	10.58	9.55	5.76	13.99	9.79	4.64	5.62	5.69
Tb	1.29	0.80	2.03	1.66	1.01	2.53	1.67	0.84	1.00	0.95
Dy	7.91	4.99	12.43	9.95	6.16	15.53	10.63	4.88	5.82	5.64
Ho	1.59	1.00	2.56	2.07	1.30	3.13	2.15	1.00	1.25	1.09
Er	4.51	2.77	7.65	6.29	3.93	8.85	6.41	2.78	3.53	3.03
Tm	0.61	0.40	1.08	0.92	0.60	1.25	0.89	0.40	0.47	0.42
Yb	3.76	2.47	6.77	5.95	3.72	7.64	5.44	2.46	2.82	2.56
Lu	0.60	0.38	1.02	0.94	0.60	1.18	0.86	0.39	0.46	0.41

Abbreviations: (I), Type I; (II) Type II; (R), Rim granite; (T), turbidite; (Q), quartzite; (G), Gabbro.

Sample	2495	4259	1604	2931	268	3588	4693	5672	5697	6066
Type	4	4	5 (I)	5 (I)	5 (I)	5 (II)	5 (I)	5 (I)	5 (I)	6 (T)
(wt. %)										
SiO_2	61.46	65.16	52.31	49.29	54.50	54.33	55.75	50.58	49.73	70.88
Al_2O_3	15.54	14.38	15.12	16.47	14.80	13.33	14.43	14.55	13.54	15.12
TiO_2	1.007	0.784	0.984	0.964	0.801	2.094	1.113	1.031	1.396	0.734
FeO^*	7.73	7.18	12.56	12.52	10.61	13.22	6.78	12.84	15.08	6.01
MnO	0.125	0.143	0.256	0.218	0.179	0.159	0.233	0.212	0.232	0.040
CaO	5.68	4.39	10.42	10.19	7.41	7.52	14.51	11.07	10.03	0.82
MgO	3.93	3.05	6.48	7.70	7.80	5.75	3.11	7.10	7.08	2.60
K_2O	0.34	1.62	0.01	0.86	0.32	0.16	0.33	0.17	0.37	2.28
Na_2O	3.94	3.13	1.79	1.71	3.52	2.81	3.64	2.37	2.41	1.47
P_2O_5	0.25	0.154	0.076	0.074	0.061	0.619	0.094	0.081	0.125	0.051
(ppm)										
Ni*	28	39	93	109	50	80	73	105	64	58
Cr*	26	55	255	306	243	176	227	153	120	141
Sc*	18	21	45	41	42	40	49	51	47	15
V*	123	111	284	264	262	164	293	296	356	143
Ga*	19	18	21	17	18	20	20	19	21	19
Cu*	0	7	111	183	142	46	27	110	96	1
Zn*	141	107	107	100	73	120	52	97	107	70
U	1.61	1.85	0.08	0.09	0.60	1.12	0.24	0.08	0.12	4.53
Th	6.07	6.09	0.35	0.27	1.93	3.90	0.41	0.30	0.44	17.65
Hf	6.74	4.86	1.58	1.43	1.70	7.13	1.81	1.60	2.45	5.45
Zr*	241	182	58	55	66	300	69	62	83	189
Pb	3.90	12.76	1.73	3.41	3.25	3.69	2.35	1.17	0.64	11.08
Ba	151	478	11	77	84	18	61	34	49	465
Sr*	150	200	123	107	142	171	164	108	72	66
Rb	6.4	43.9	0.9	25.1	5.4	3.7	8.6	2.1	7.3	82.6
Cs	0.2	0.5	0	0.8	0.1	0.1	0.1	0	0	3.6
Nb	17.5	10.0	2.7	2.7	3.3	20.5	3.1	2.6	4.5	10.5
Y	40.63	27.05	22.32	21.44	22.08	41.49	23.32	23.02	31.07	25.11
Ta	1.30	0.95	0.24	0.17	0.34	1.30	0.31	0.28	0.37	1.17
La	33.09	25.41	3.57	3.38	7.92	25.87	4.78	3.4	4.97	38.18
Ce	68.27	47.39	8.39	7.87	15.51	57.03	10.44	8.09	12.32	63.49
Pr	8.2	5.39	1.27	1.16	1.91	7.18	1.52	1.26	1.86	6.38
Nd	33.80	21.64	6.41	5.93	8.56	31.82	7.51	6.35	9.37	23.53
Sm	7.91	5.08	2.32	2.20	2.62	7.95	2.67	2.39	3.44	4.76
Eu	1.90	1.37	0.96	0.80	0.97	2.72	1.01	0.93	1.13	0.89
Gd	7.34	4.60	2.94	2.84	3.14	7.72	3.21	3.16	4.08	4.14
Tb	1.26	0.83	0.58	0.56	0.61	1.29	0.61	0.60	0.84	0.72
Dy	7.66	4.95	3.80	3.76	3.90	7.88	4.19	4.13	5.91	4.48
Ho	1.53	0.99	0.84	0.81	0.85	1.63	0.92	0.87	1.25	0.91
Er	4.37	2.71	2.51	2.44	2.50	4.56	2.61	2.52	3.62	2.78
Tm	0.60	0.39	0.36	0.34	0.34	0.63	0.36	0.36	0.51	0.40
Yb	3.67	2.53	2.23	2.19	2.12	3.84	2.23	2.20	3.23	2.52
Lu	0.56	0.40	0.36	0.35	0.33	0.60	0.35	0.35	0.53	0.40

Sample Type	6106 6 (Q)	1510 7 (R)	2935 7 (I)	4171B 7 (I)	4188 7 (I)	5246 7 (I)	5698 7 (I)	1663 7 (II)	6161 7 (G)
(wt. %)									
SiO_2	81.11	78.04	74.65	73.23	71.70	73.83	67.89	73.80	53.29
Al_2O_3	8.18	11.90	14.66	15.04	15.75	14.34	16.31	14.08	7.22
TiO_2	0.273	0.084	0.190	0.202	0.288	0.192	0.587	0.253	0.971
FeO^*	3.30	1.27	1.32	1.73	2.22	1.37	3.48	1.82	7.24
MnO	0.052	0.012	0.018	0.034	0.064	0.023	0.063	0.019	0.145
CaO	3.44	0.36	0.41	2.40	2.65	1.61	3.61	1.01	14.67
MgO	1.70	0.08	0.61	1.16	0.63	0.43	1.07	0.40	12.83
K_2O	1.20	4.01	2.28	1.29	1.16	4.22	1.85	4.02	1.22
Na_2O	0.68	4.23	5.81	4.89	5.43	3.95	5.00	4.55	1.50
P_2O_5	0.061	0.005	0.051	0.026	0.104	0.046	0.147	0.057	0.922
(ppm)									
Ni*	29	10	11	16	10	11	7	14	79
Cr*	30	0	0	8	0	0	0	1	346
Sc*	12	0	0	4	0	0	3	1	47
V*	44	2	7	28	17	15	49	21	168
Ga*	11	20	20	19	22	19	22	18	16
Cu*	12	1	11	3	1	3	8	17	0
Zn*	48	5	26	45	52	38	79	21	87
U	3.27	2.50	1.74	1.54	1.00	1.28	1.50	3.10	4.59
Th	8.14	18.23	3.70	10.54	5.11	4.76	5.49	19.93	5.23
Hf	2.86	7.29	3.13	4.94	3.86	2.86	6.05	4.09	2.28
Zr*	102	176	105	195	161	106	257	144	88
Pb	4.26	5.65	3.28	29.41	10.21	15.35	9.84	18.55	8.90
Ba	273	669	329	163	214	1445	540	639	562
Sr*	84	22	134	236	212	204	423	84	413
Rb	41.4	93.1	62.5	60.6	106.9	108.4	51.1	145.4	28.5
Cs	2.1	0.3	0.9	1.0	7.2	2.8	3.1	1.9	0.2
Nb	4.4	26.7	7.1	5.1	12.2	2.4	10.0	12.7	8.2
Y	13.94	92.01	5.70	13.03	10.89	2.89	13.00	16.72	25.83
Ta	0.65	1.53	1.33	0.90	1.59	0.12	0.94	1.62	0.91
La	21.94	62.39	12.93	31.45	30.36	29.01	41.03	43.90	38.42
Ce	38.51	126.69	22.57	50.77	53.10	43.83	72.09	72.64	77.00
Pr	4.11	14.81	2.35	4.88	5.06	4.39	7.32	6.95	9.21
Nd	15.52	59.85	8.97	17.41	17.65	14.67	26.42	23.70	40.49
Sm	2.94	15.45	1.84	3.82	3.03	1.71	4.69	4.09	10.17
Eu	0.63	1.39	0.44	1.15	0.82	0.53	1.29	0.62	2.80
Gd	2.39	14.71	1.24	2.80	1.92	1.04	3.15	2.97	7.88
Tb	0.41	2.63	0.18	0.44	0.33	0.10	0.47	0.49	1.07
Dy	2.57	16.18	0.99	2.48	1.90	0.51	2.59	3.03	5.33
Ho	0.51	3.38	0.19	0.48	0.36	0.08	0.46	0.60	0.92
Er	1.42	9.58	0.51	1.28	0.93	0.22	1.20	1.70	2.37
Tm	0.20	1.33	0.08	0.17	0.13	0.04	0.17	0.24	0.30
Yb	1.25	8.06	0.49	1.03	0.76	0.27	1.02	1.47	1.85
Lu	0.20	1.22	0.08	0.17	0.11	0.05	0.16	0.23	0.28

Table 2.3. Sm-Nd isotope analyses of samples from the Hanikahimajuk Lake area.

SAMPLE	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ [a]	εNd_T [b]	TDM
(Pre-YKSG basement tonalite)						
1660	2.21	15.66	0.0855	0.510138 (8)	-0.3	3.6
1754	1.64	12.41	0.0801	0.510140 (7)	2.1	3.4
7072	2.67	15.60	0.1034	0.510600 (16)	0.9	3.5
7077	0.69	3.03	0.1385	0.511398 (14)	1.2	3.5
(Syn-volcanic diorite)						
414	5.10	24.85	0.1241	0.511244 (10)	-2.2	3.2
415	4.47	22.64	0.1195	0.511260 (8)	-0.3	3.1
2917	5.59	24.79	0.1363	0.511612 (9)	0.8	3.0
2919	6.74	28.45	0.1432	0.511746 (11)	1.1	3.0
3339	7.70	36.37	0.1281	0.511379 (10)	-0.9	3.2
5609	0.86	2.47	0.2113	0.512921 (13)	0.5	-
5719	7.03	25.02	0.1700	0.512125 (8)	-0.8	-
7078	4.63	21.26	0.1317	0.511487 (8)	-0.0	3.1
(YKSG felsic volcanic rock)						
1295	10.29	43.39	0.1434	0.511684 (5)	-0.2	3.2
1655	7.06	37.86	0.1128	0.511152 (10)	-0.1	3.0
1666	16.42	82.33	0.1206	0.511263 (8)	-0.6	3.1
2767	10.79	53.32	0.1224	0.511352 (9)	0.6	3.0
5132	10.68	53.65	0.1203	0.511293 (12)	0.1	3.0
(YKSG intermediate volcanic rock)						
1719	5.32	24.98	0.1288	0.511488 (11)	1.0	3.0
2484	5.93	27.82	0.1288	0.511439 (8)	0.1	3.1
2495	7.39	34.90	0.1280	0.511457 (9)	0.7	3.0
4259	4.82	22.86	0.1274	0.511478 (15)	1.3	3.0
7080	4.67	20.99	0.1346	0.511553 (8)	0.3	3.1
(YKSG mafic volcanic rock						
268	2.52	9.14	0.1671	0.512053 (9)	-1.2	-
1604	2.00	6.14	0.1969	0.512711 (9)	1.4	-
2931	1.97	6.05	0.1967	0.512768 (10)	2.5	-
3588	7.50	33.04	0.1372	0.511646 (9)	1.2	3.0
4693	2.52	8.18	0.1866	0.512494 (11)	0.7	-
5672	2.32	7.17	0.1956	0.512703 (9)	1.7	-
5697	3.13	9.85	0.1918	0.512631 (10)	1.6	-
(YKSG sedimentary rocks)						
6066	4.67	25.42	0.1110	0.510864 (8)	-5.3	3.4
6106	2.83	15.99	0.1070	0.511018 (12)	-0.9	3.0
(Rim granite)						
1510	16.25	68.24	0.1440	0.511755 (13)	1.0	3.1
(Syn- to post-deformation granitoids)						
1663	3.59	23.89	0.0908	0.510846 (10)	0.5	2.9
2935	1.81	9.98	0.1099	0.511190 (9)	0.8	2.9
4171	3.40	18.18	0.1131	0.511092 (8)	-2.2	3.1
4188	2.84	18.79	0.0914	0.510857 (13)	0.5	2.9
5246	1.99	16.02	0.0751	0.510513 (9)	-0.8	2.9
5698	4.50	28.30	0.0961	0.510856 (8)	-1.1	3.0
(Post-deformation gabbro)						
6161	9.33	43.46	0.1298	0.511505 (15)	0.3	3.0

Notes: [a] Normalized to $^{146}Nd/^{144}Nd=0.7219$. Numbers in (paretheses) $2\sigma m$ uncertainties.;
[b] εNd at 3.38 Ga for basement tonalite, 2.68 Ga for YKSG-volcanic rocks, syn-volcanic diorites and Rim granite, 2.66 Ga for YKSG-sedimentary rocks, 2.60 Ga for syn- to post-deformation granitoids/gabbro.
[c] TDM calcutaled using the mantle evolution model of Goldstein et al. (1984). Present day CHUR parameters are $^{147}Sm/^{144}Nd=0.1967$, $^{143}Nd/^{144}Nd=0.512638$. $\lambda_{147Sm}=6.54\times10^{-12}\ a^{-1}$

Table 2.4. Pb isotopic composition of K-feldspar and plagioclase.

Samples	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Mineral
(Basement tonalite)				
7072 (L1)*	21.833	16.714	42.030	-
7072 (L4)**	15.713	15.313	34.266	-
7072 (R)**	14.406	15.080	34.022	Plag
1660 (L1)*	18.505	15.993	42.542	-
1660 (L4)**	15.639	15.340	35.808	-
1660 (R)*	14.701	15.181	35.690	Plag
(Syn-volcanic diorite)				
415 (L1)*	22.857	16.539	39.271	-
415 (R)*	15.018	15.137	34.518	Plag
(Syn- to post-deformation granitoids)				
1663 (L1)*	19.267	15.991	38.042	-
1663 (L4)*	14.738	15.261	34.231	-
1663 (R)*	14.541	15.214	34.082	Kfs
5246 (L1)*	21.063	16.214	37.125	-
5246 (L4)*	14.299	15.027	33.825	-
5246 (R)*	14.134	14.978	33.754	Kfs

Notes: L1 =leach 1 (24 hrs in 2N HCl), L4 =leach 3 (24 hrs in 16N HNO_3+1 drop 48% HF), L4A~C = (three 20 minutes leaches in 7N HNO_3+ 5% HF combined together); Kfs=K-feldspar, Plag=plagioclase; * Analyzed on MM 30; ** Analyzed on VG 354.

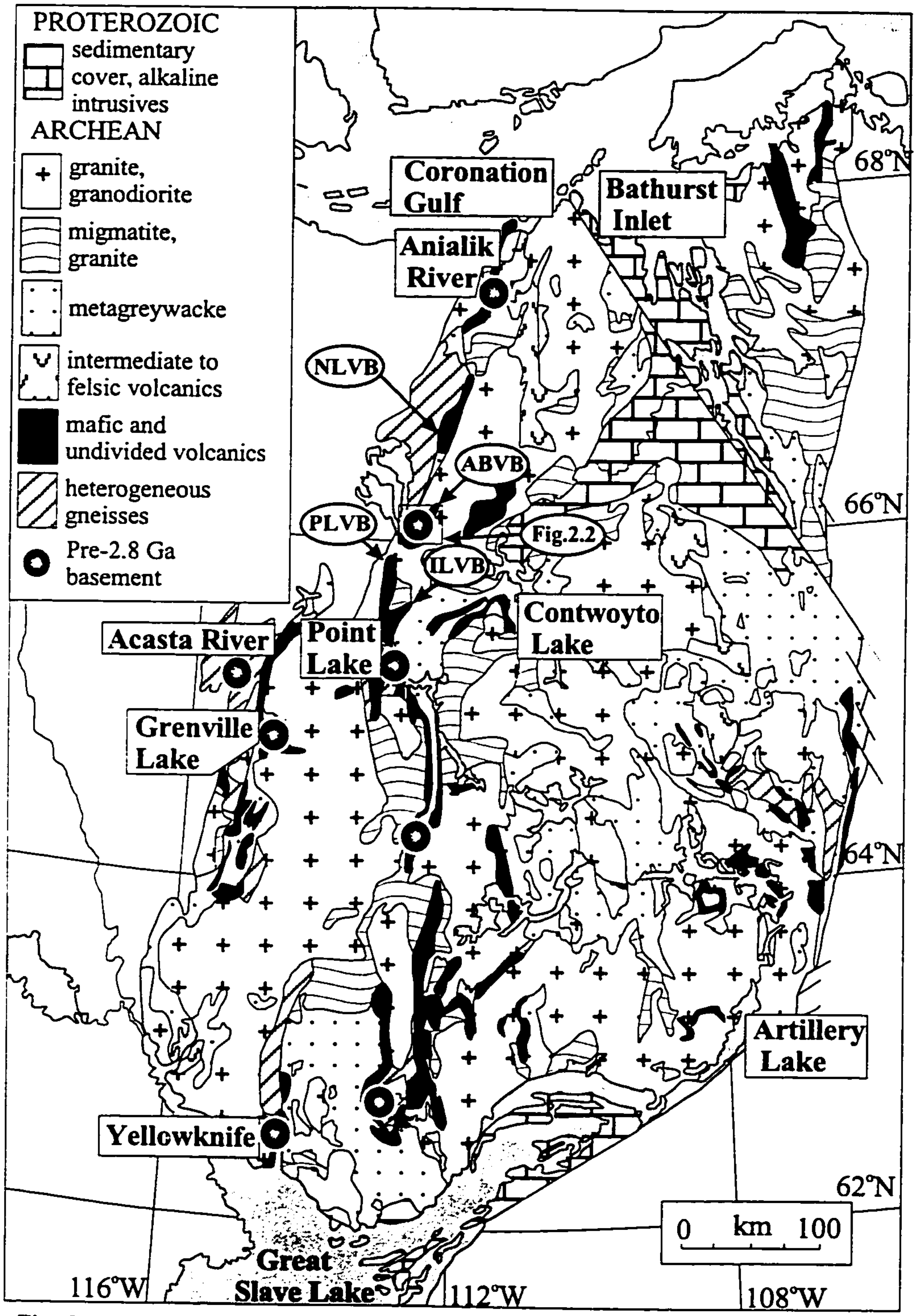


Fig. 2.1. Geological map of the Slave province (after Hoffman 1989). Abbreviations are,ILVB=Izok Lake volcanic belt,PLVB=Point Lake volcanic belt,ABVB=Amooga Booga volcanic belt, NLVB=NapaktulikLake volcanic belt.

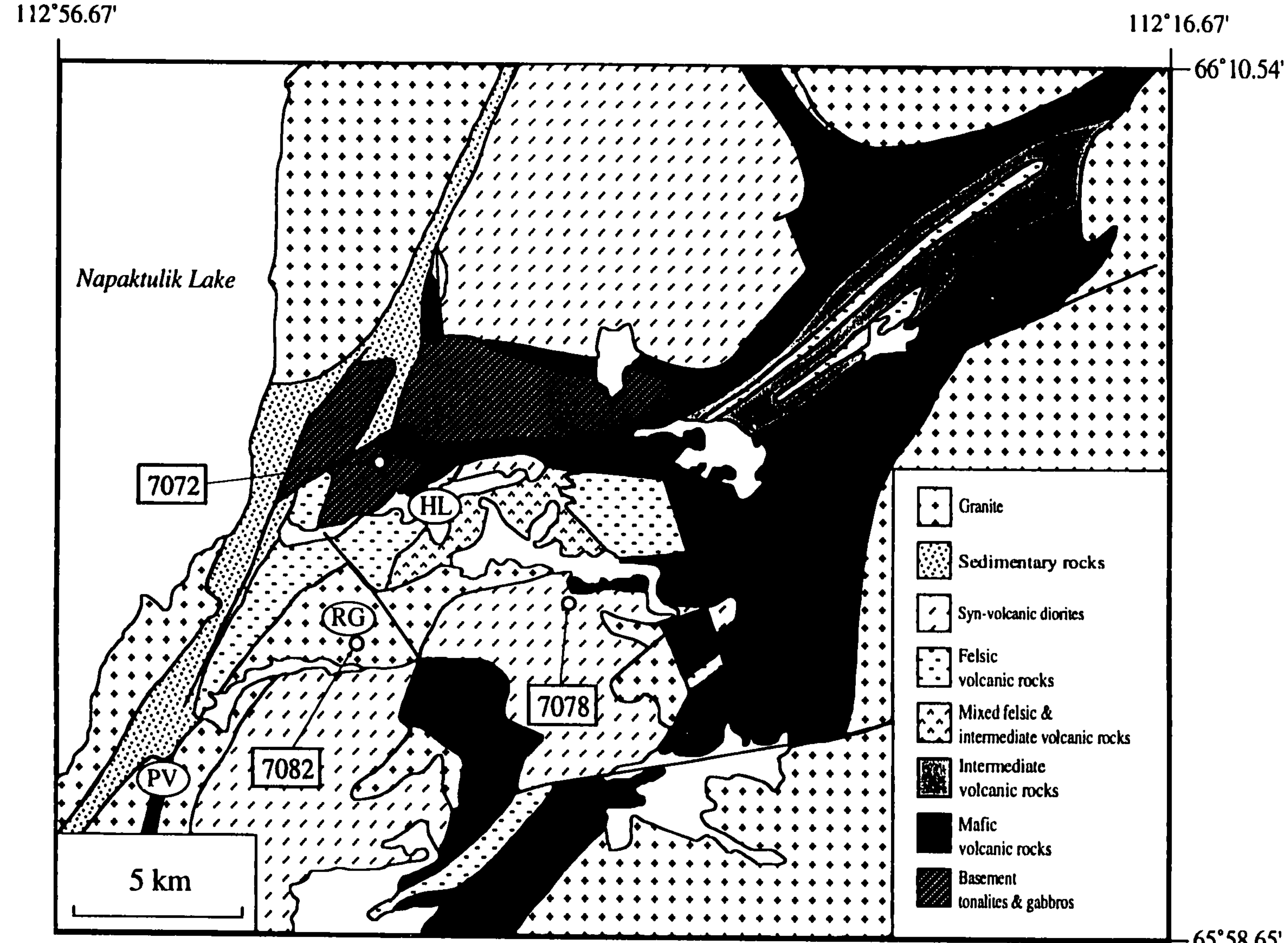


Fig. 2.2. Geological map of the Hanikahimajuk Lake area (after Jensen 1995). Abbreviations are, HL=Hanikahimajuk Lake, RG=Rim granite, PV=Point Lake volcanic belt. Also shown are the locations for the U-Pb geochronology samples (*Note: Samples 6161 and MSD14552 are located south of the map shown here).

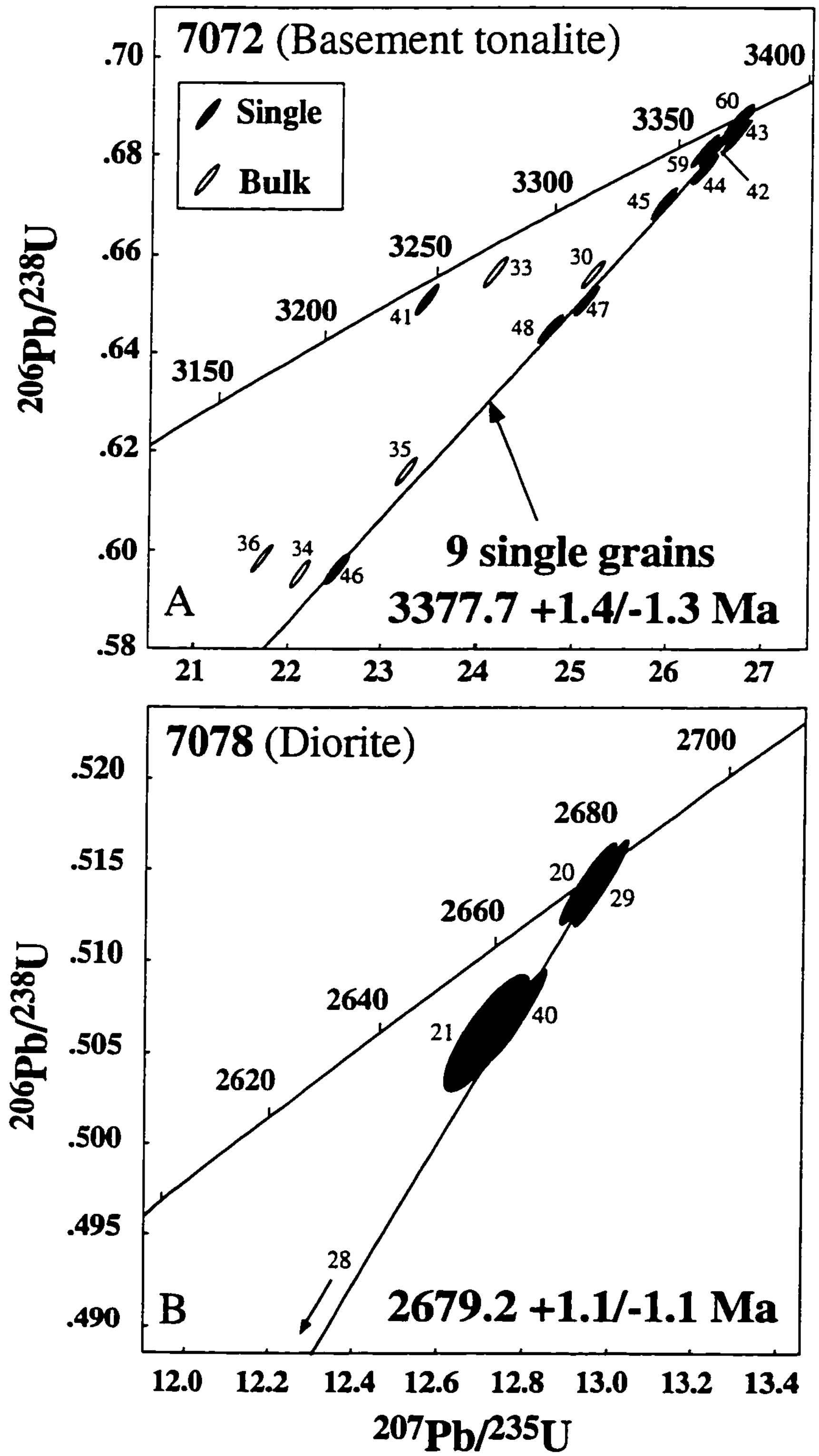
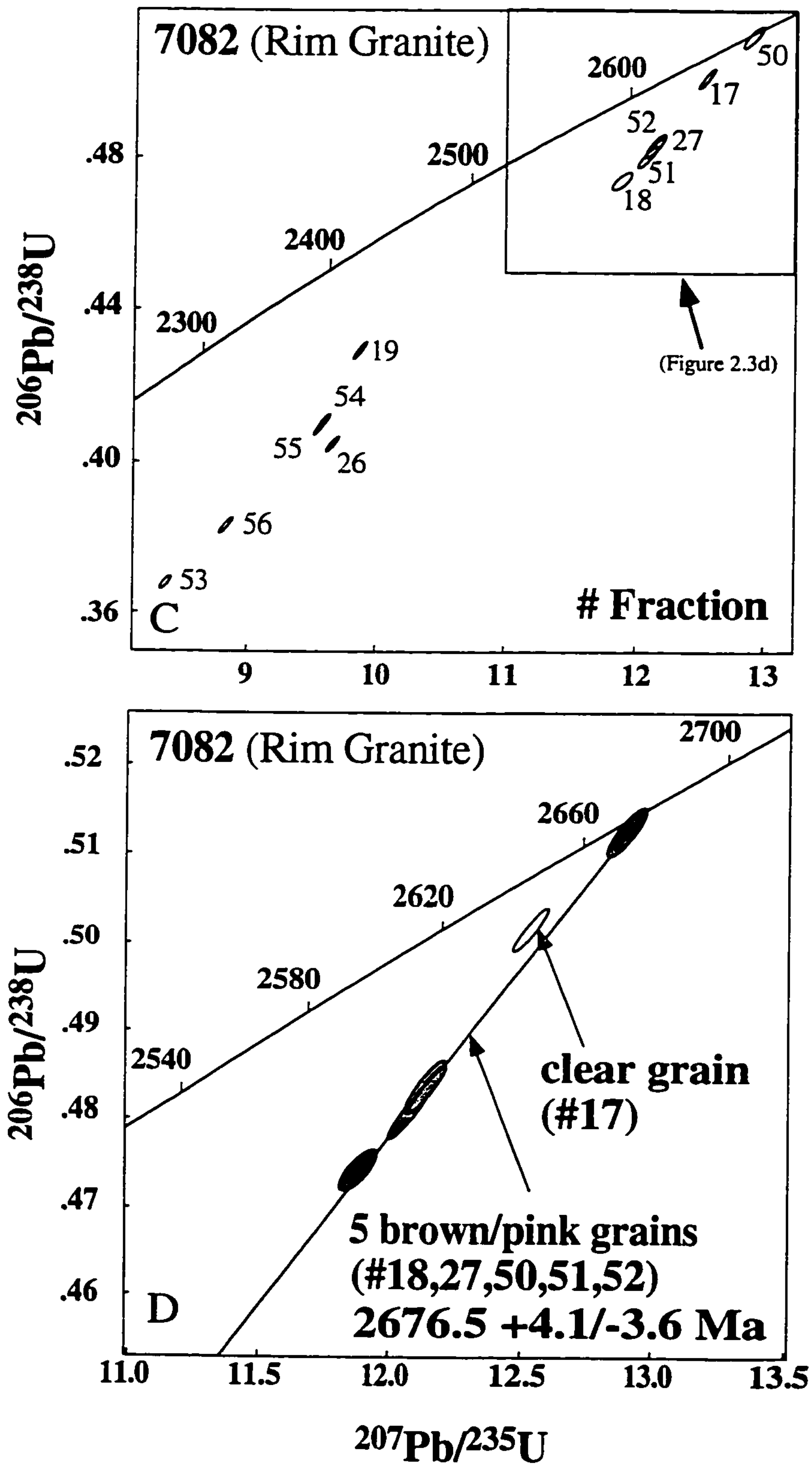
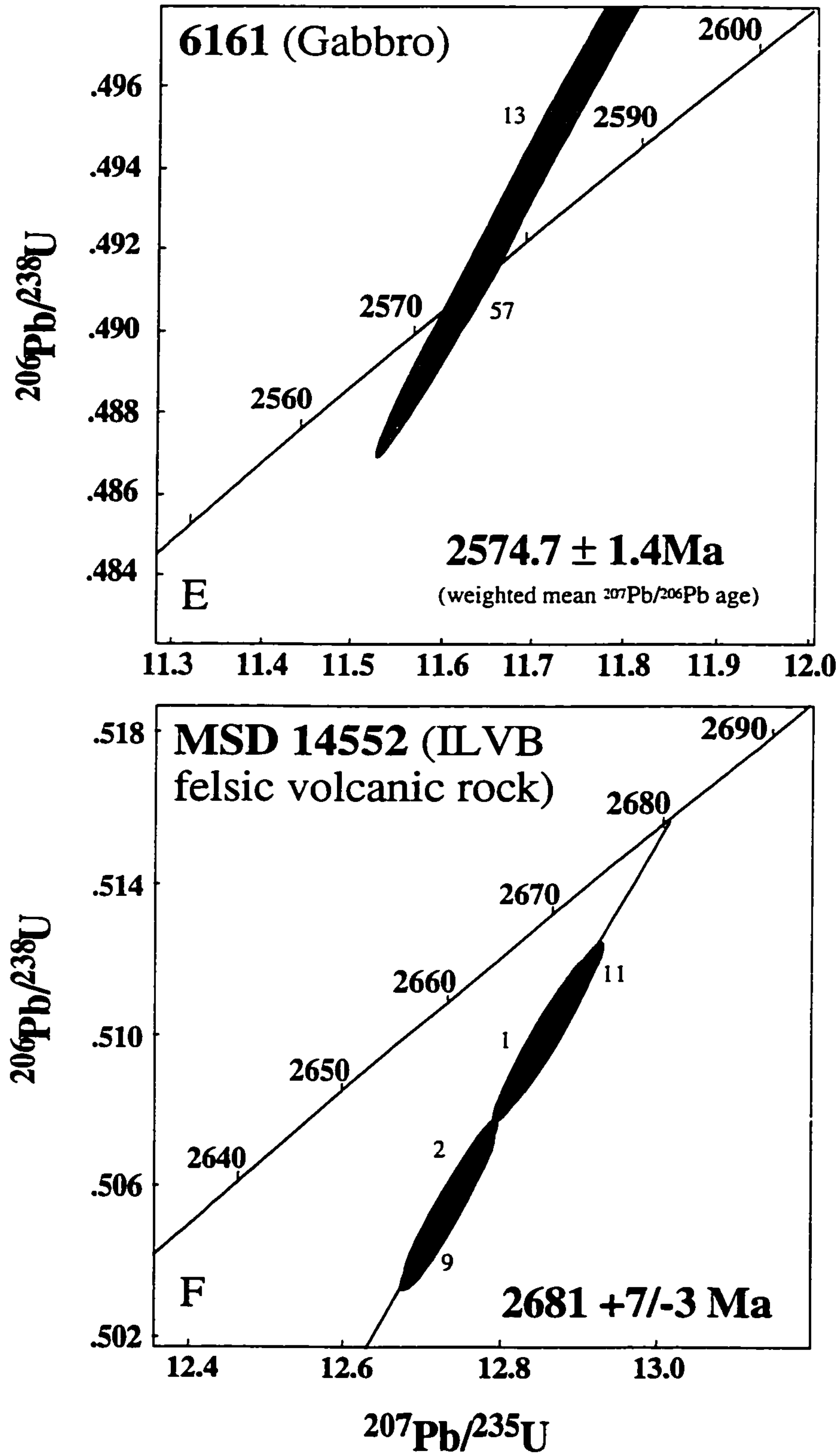


Fig. 2.3. Concordia diagrams displaying the U-Pb isotopic results for (a) pre-YKSG basement tonalite (b) syn-volcanic diorite (c,d) Rim granite (e) post-deformation gabbro and (f) felsic volcanic rocks from the Izok Lake volcanic belt. Numbers next to the ellipses represent fraction# (see Table. 2.1).



7082 (Rim Granite)
2300
2400
2500
2600
50
17
52
27
51
18
19
54
55
26
56
53
(Figure 2.3d)
C
Fraction
.36
.40
.44
.48
9
10
11
12
13
206Pb/238U
7082 (Rim Granite)
2540
2580
2620
2660
2700
clear grain
(#17)
5 brown/pink grains
(#18,27,50,51,52)
2676.5 +4.1/-3.6 Ma
D
.46
.47
.48
.49
.50
.51
.52
11.0
11.5
12.0
12.5
13.0
13.5
206Pb/238U
207Pb/235U


6161 (Gabbro)
2600
2590
13
2570
57
2560
2574.7 ± 1.4Ma
(weighted mean 207Pb/206Pb age)
E
.496
.494
.492
.490
.488
.486
.484
206Pb/238U
11.3
11.4
11.5
11.6
11.7
11.8
11.9
12.0
MSD 14552 (ILVB
felsic volcanic rock)
2690
2680
2670
2660
2650
2640
11
1
2
9
F
2681 +7/-3 Ma
.518
.514
.510
.506
.502
206Pb/238U
12.4
12.6
12.8
13.0
207Pb/235U

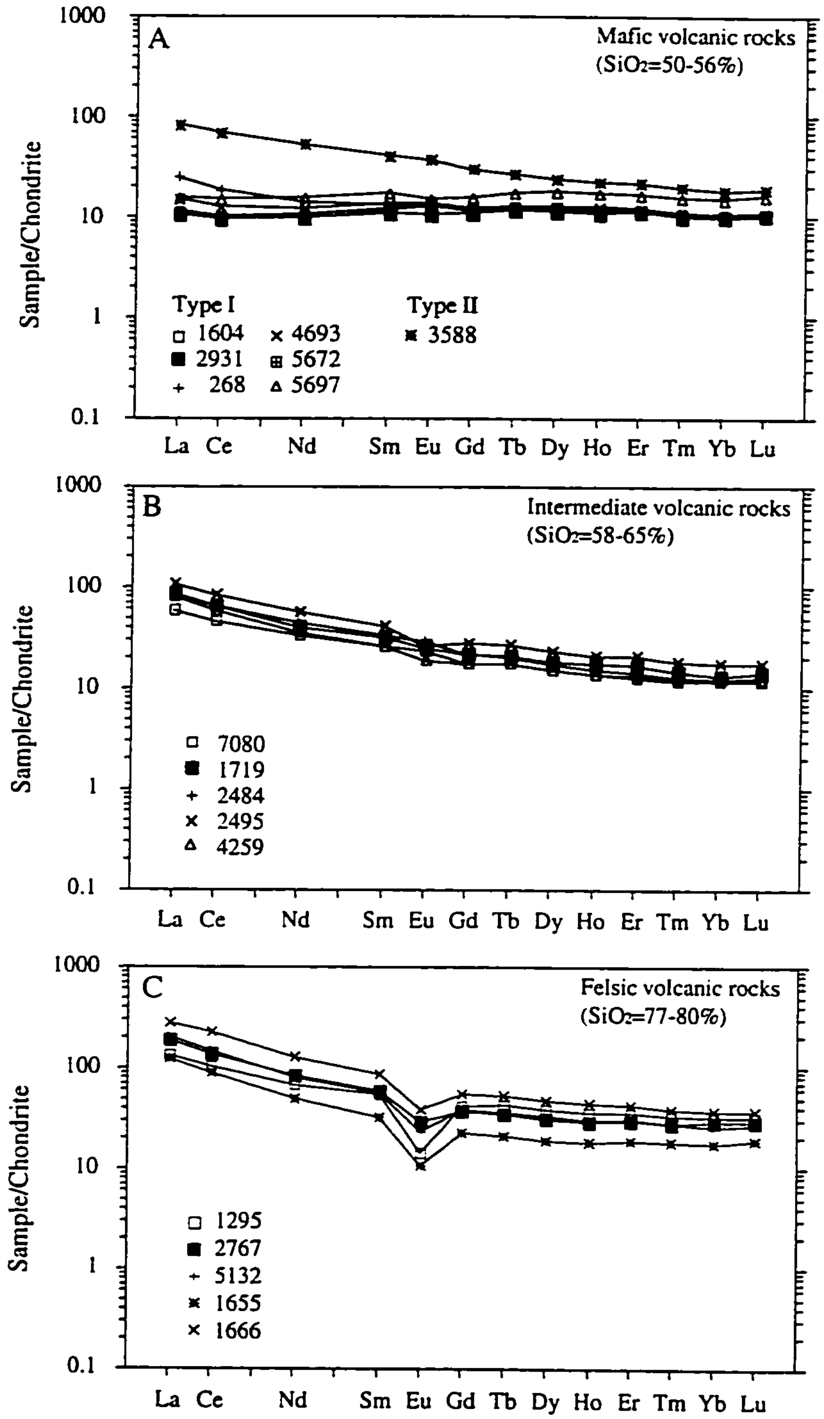
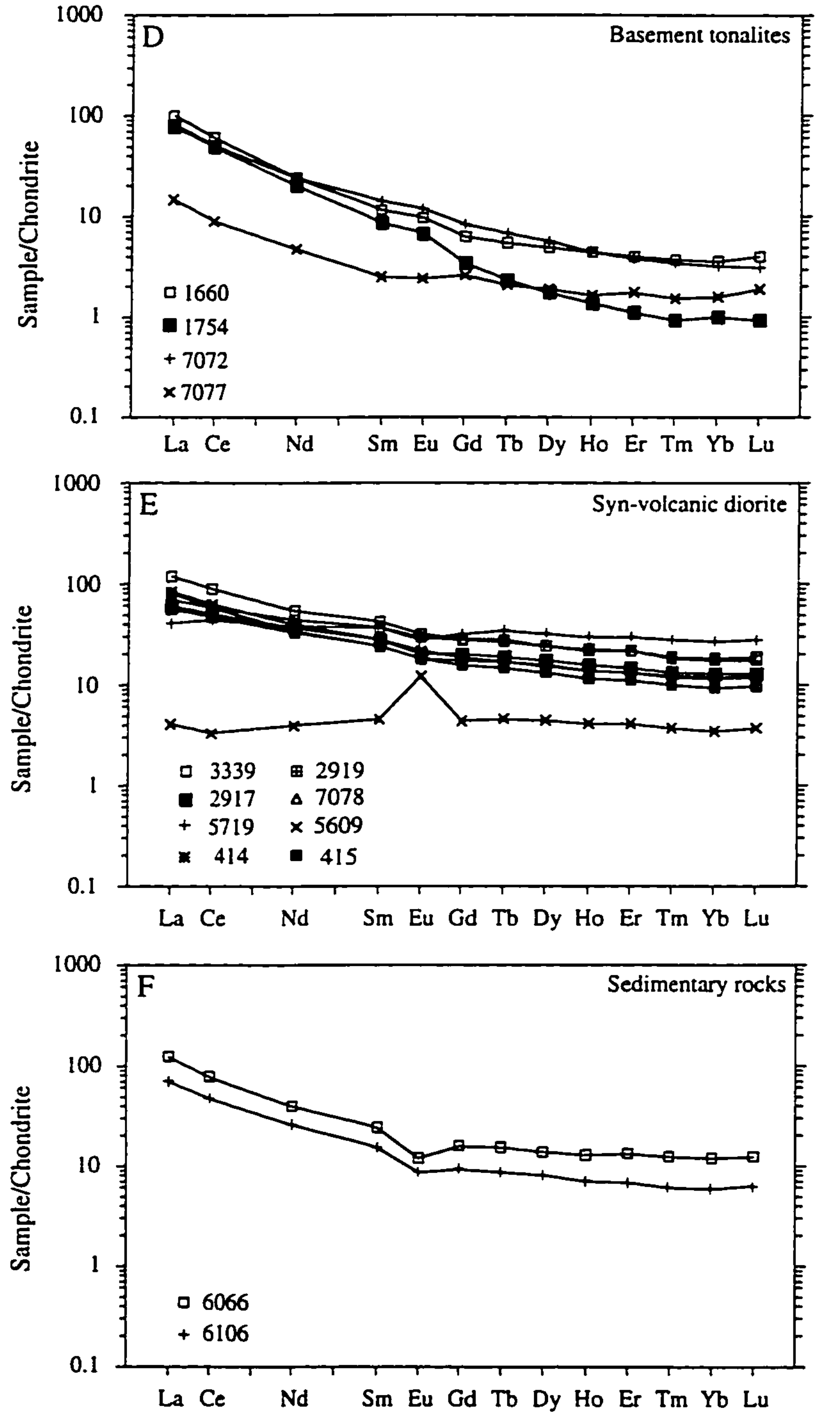
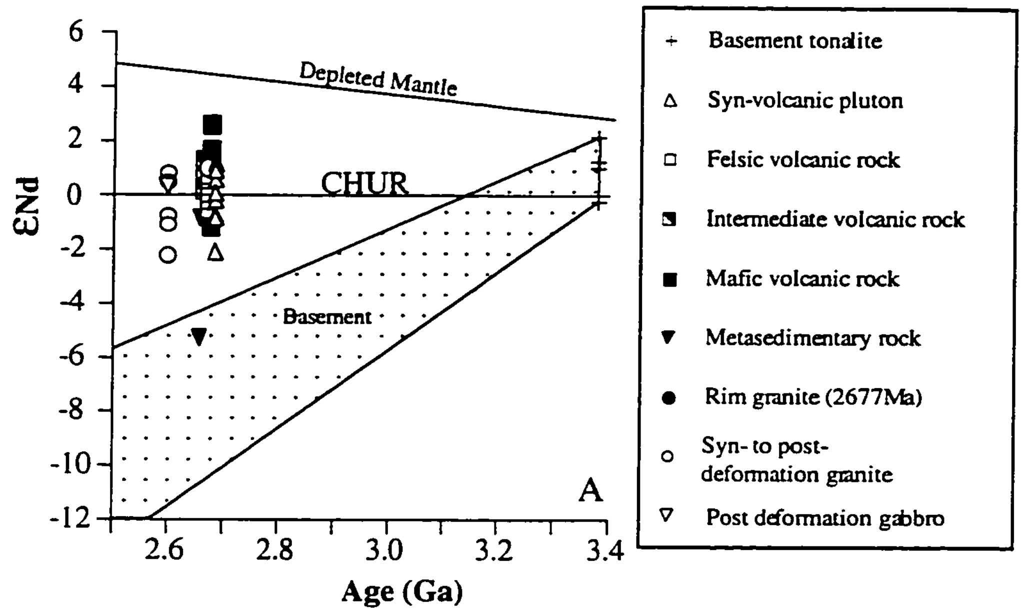


Fig. 2.4. Chondrite normalized REE patterns for (a) YKSG mafic volcanic rocks (b) YKSG intermediate volcanic rocks (c) YKSG felsic volcanic rocks (d) pre-YKSG basement tonalites (e) syn-volcanic diorites and (f) YKSG metasedimentary rocks. Chondrite values are taken from Boynton (1984).



1000
100
10
1
0.1
D
Basement tonalites
Sample/Chondrite
1660
1754
7072
7077
La Ce Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu
E
Syn-volcanic diorite
3339
2919
2917
7078
5719
5609
414
415
F
Sedimentary rocks
6066
6106



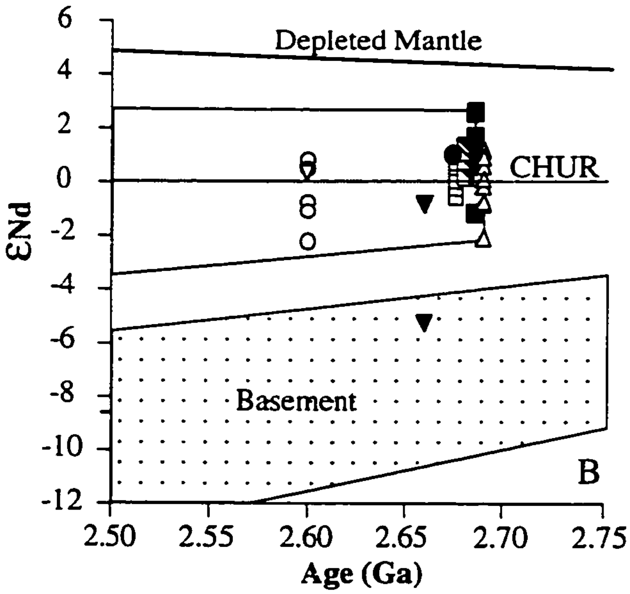


Fig. 2.5 Time versus εNd diagram for early to mid-Archean rocks in the Hanikahimajuk Lake area. Nd isotopes are shown in epsilon notation (deviation from chondritic uniform reservoir (CHUR) in parts per 10^4).

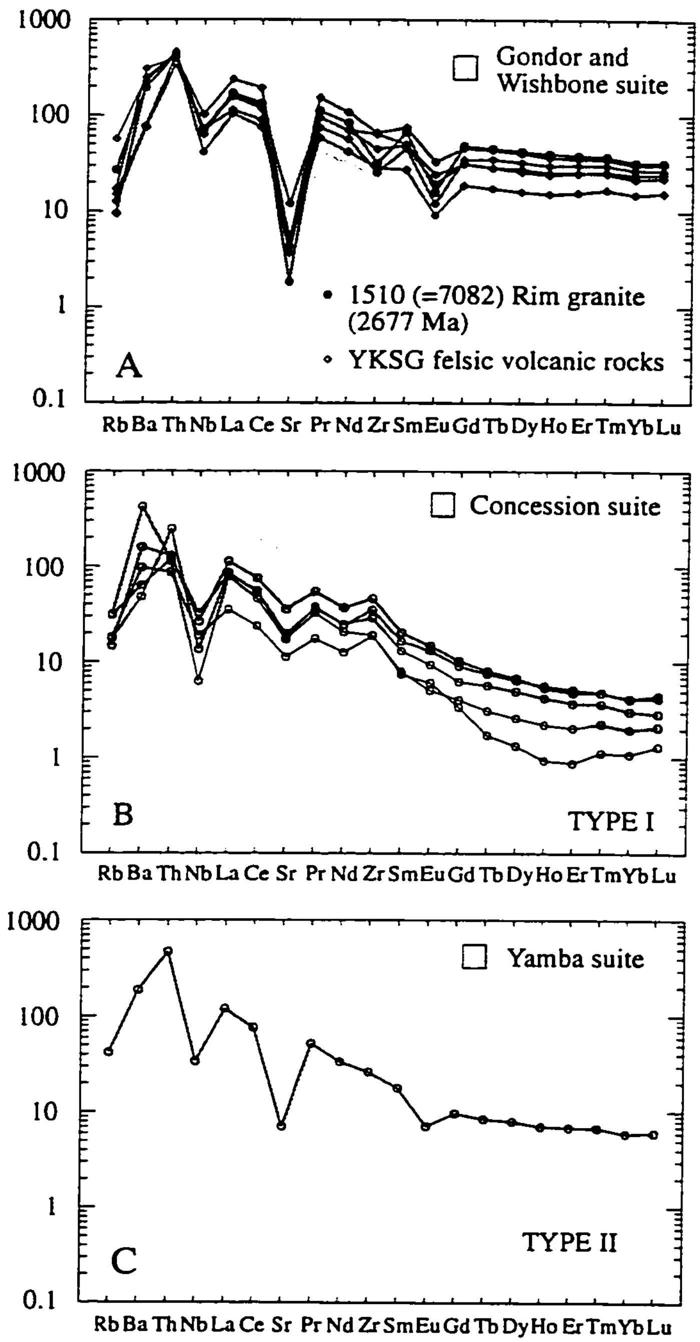


Fig. 2.6. Chondrite normalized extended REE patterns for (a) Rim granite and YKSG felsic volcanic rocks (b) Type I granitoids and (c) Type II granitoid. Also shown are patterns for Gondor, Wishbone, Concession and Yamba suite granitoids in the central Slave province (data for Gondor, Wishbone, Concession and Yamba suite granitoids are taken from Davis et al. 1994).

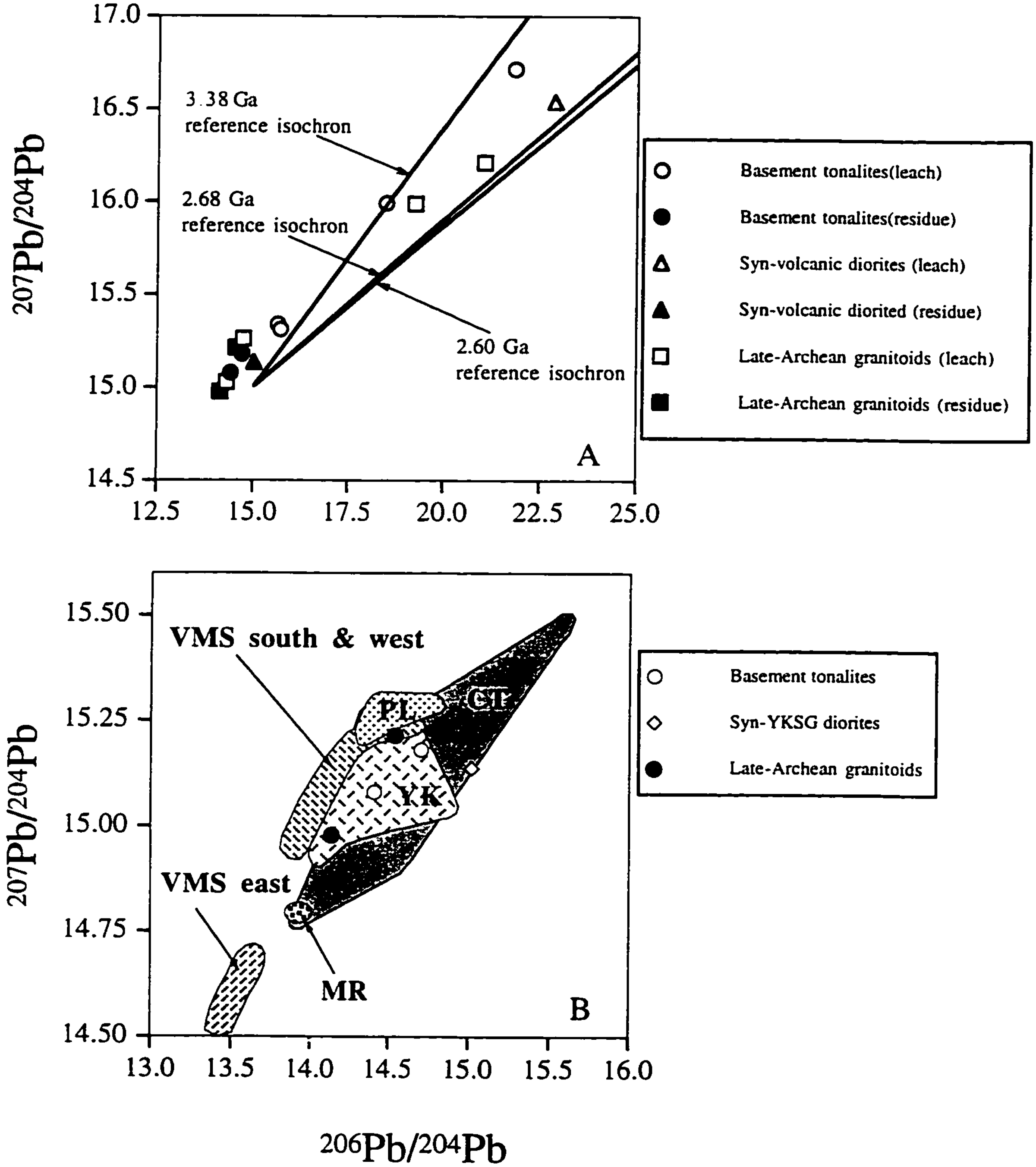


Fig. 2.7. Pb isotopic compositions of feldspars. (A) Leaches and residues of granitoids from the Hanikahimajuk Lake area. Also shown are 2.60, 2.68 and 3.38 Ga reference isochrons. (B) Residues (this study) together with fields for granitoids from the Point Lake, Contwoyto Lake and Malley Rapids. Fields for Point Lake, Contwoyto Lake and Malley Rapids are from Davis et al. (1996).

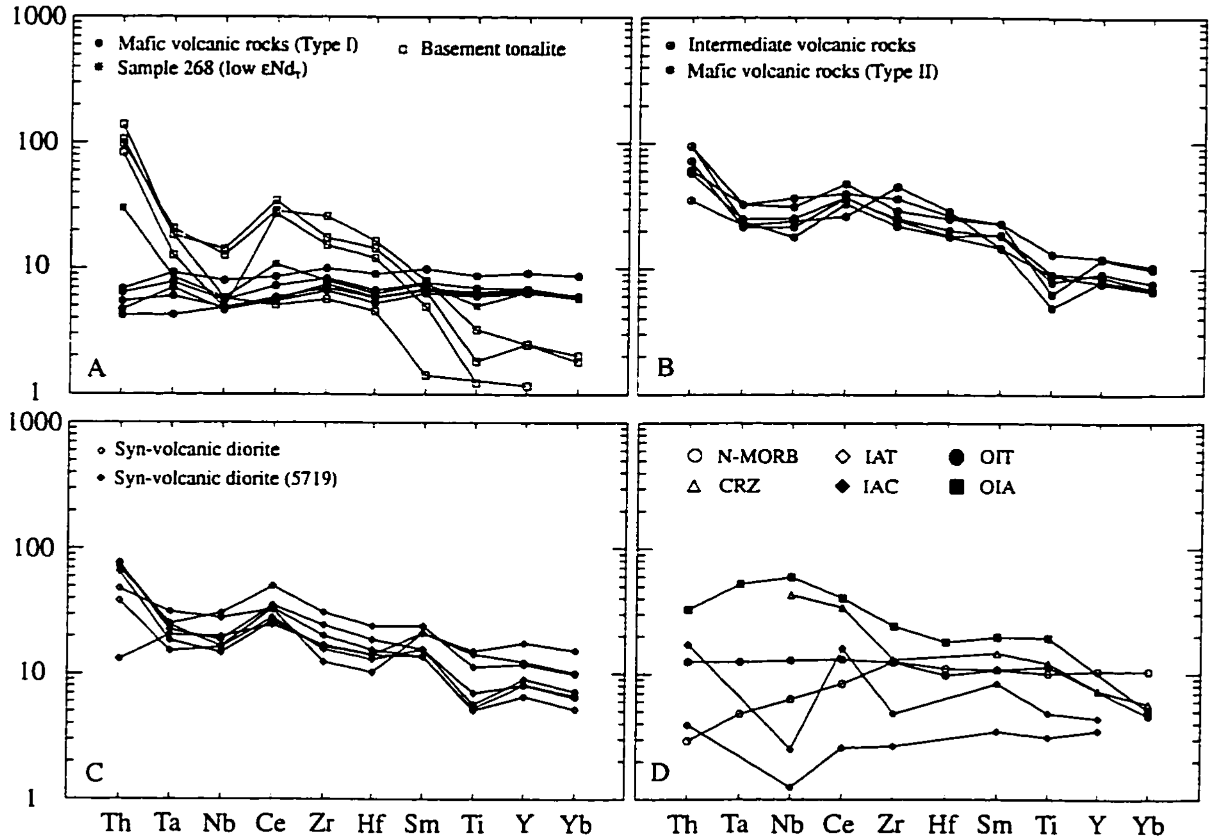


Fig. 2.8. Primitive mantle normalized abundance patterns for (a) type I mafic volcanic rocks and pre-YKSG basement tonalites (b) type II mafic volcanic rocks and intermediate volcanic rocks (c) syn-volcanic diorites and (d) representative mafic volcanic rocks from various tectonic settings. Abbreviations are, N-MORB - N type mid oceanic ridge basalt (Hofmann 1988), CRZ - continental rift zone transitional basalt (Davies and Macdonald 1987), IAT - Island arc tholeiite (Sun 1980), IAC - Island arc calc-alkaline (Sun 1980), OIT - Ocean island tholeiite (Chen and Frey 1983), OIA - ocean island alkaline (Chen and Frey 1983). Values for the primitive mantle are taken from Taylor and McLennan (1985).

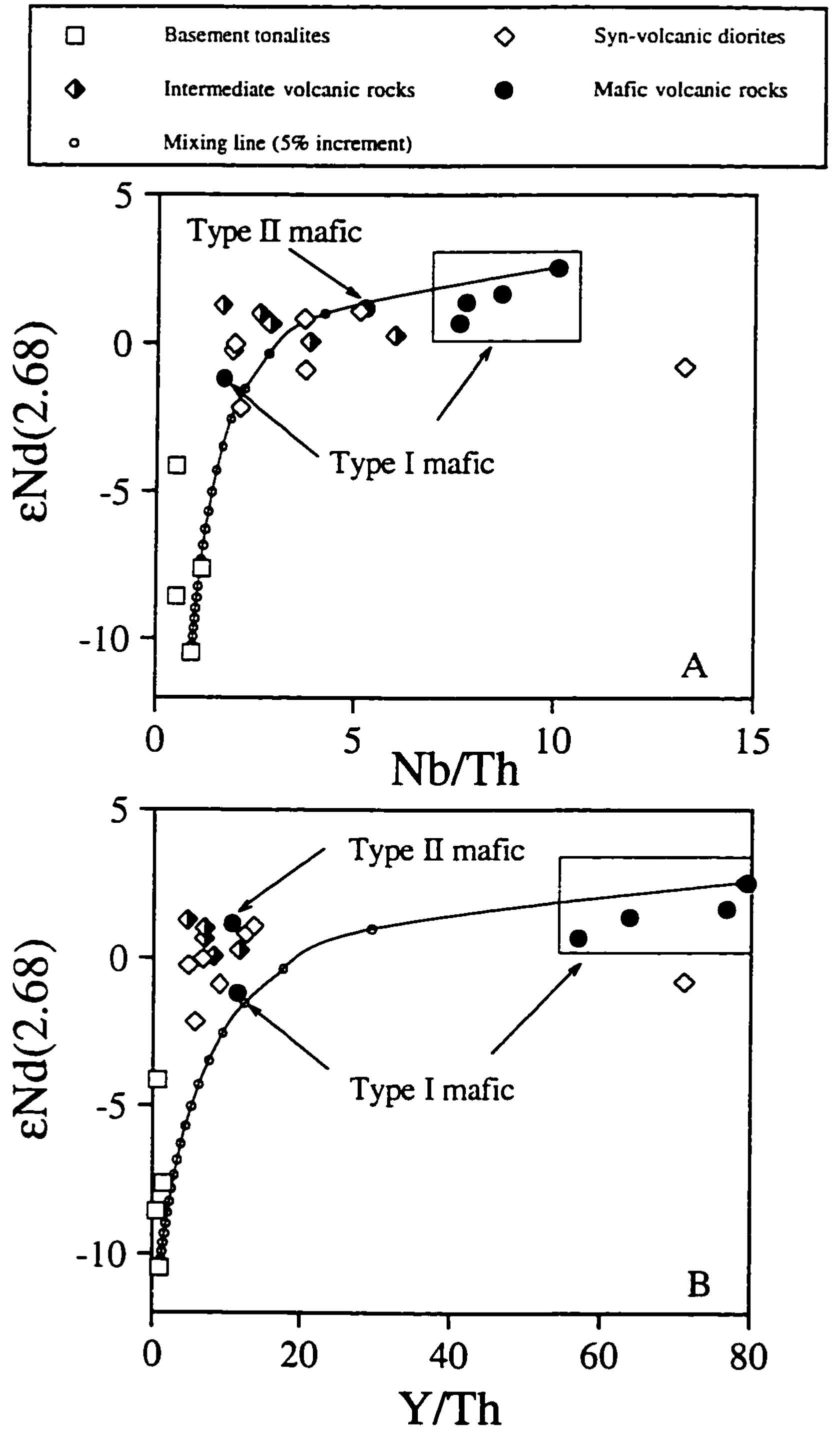


Fig. 2.9. (a) Nb/Th versus εNd(2.68) diagram and (b) Y/Th versus εNd(2.68) diagram showing the effect of crustal contamination. Type I mafic volcanic rocks with highest εNd value and basement tonalite with lowest eNd value are taken as endmembers. Type I mafic volcanic rock with lowest εNd (sample 268) lies on a mixing line. Its isotopic composition can be explained by ~15% assimilation of basement tonalite.

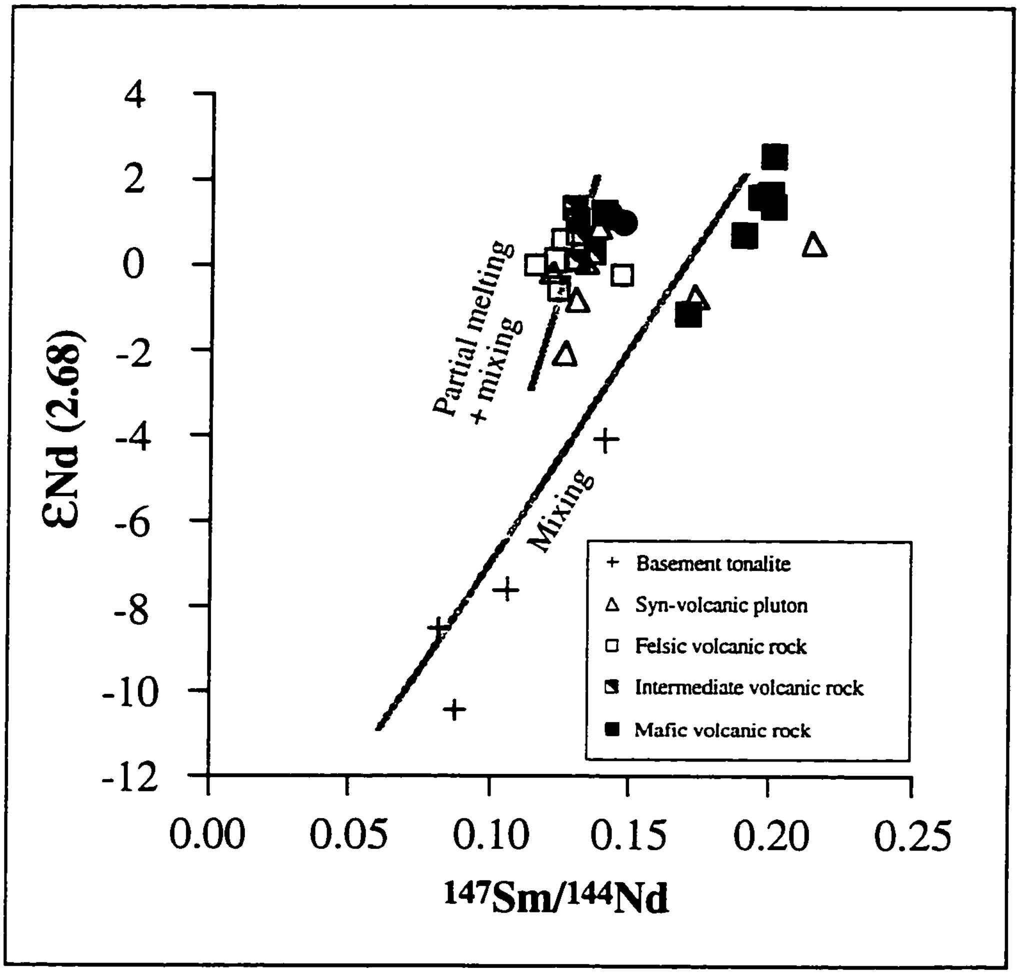


Fig. 2.10. $^{147}Sm/^{144}Nd$ versus εNd(2.68) diagram showing the data for pre-YKSG basement tonalites and YKSG mafic to felsic volcanic rocks. Also shown is mixing line between the basement and type I mafic volcanic rocks.

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Appexdix 2.1. Sample locations.

Sample	Latitude	Longitude
1660	66° 6.22'	112° 42.48'
1754	66° 5.77'	112° 41.88'
7072	66° 5.22'	112° 44.97'
7077	66° 6.19'	112° 42.82'
414	66° 7.72'	112° 42.51'
415	66° 7.83'	112° 38.55'
2917	66° 1.45'	112° 37.42'
2919	66° 1.10'	112° 37.03'
3339	66° 9.58'	112° 36.87'
5609	65° 55.82'	112° 49.08'
5719	65° 49.13'	113° 17.07'
7078	66° 3.28'	112° 38.30'
1295	66° 7.56'	112° 26.32'
1655	66° 3.66'	112° 45.31'
1666	66° 4.73'	112° 35.86'
2767	65° 59.37'	112° 45.10'
5132	65° 52.68'	113° 4.03'
1719	66° 4.98'	112° 40.52'
2484	66° 6.53'	112° 29.47'
2495	66° 6.71'	112° 27.46'
4259	65° 53.83'	113° 4.90'
7080	66° 4.05'	112° 35.34'
268	66° 5.35'	112° 43.39'
1604	66° 5.59'	112° 34.59'
2931	66° 0.67'	112° 37.59'
3588	66° 1.56'	112° 32.51'
4693	66° 2.23'	112° 52.43'
5672	65° 47.60'	113° 5.92'
5697	65° 45.50'	113° 6.84'
6066	66° 4.52'	112° 49.58'
6086	66° 6.05'	112° 47.78'
6106	66° 5.42'	112° 46.20'
1510	66° 2.66'	112° 46.10'
1663	66° 7.24'	112° 45.86'
2935	66° 0.26'	112° 35.74'
4171	65° 58.25'	112° 56.55'
4188	65° 52.24	113° 7.47'
5246	65° 52.07'	112° 45.10'
5698	65° 45.24'	113° 5.41'
6161	65° 51.68'	113° 14.33'
MSD 14552	65° 37.83'	112° 47.83'

Chapter 3

Geochemical and Nd isotopic constraints for the origin of late-Archean turbidites from the Yellowknife area, Northwest Territories, Canada.*

* A version of this paper will be submitted to Geochimica et Cosmochimica Acta. Co-authored by Robert A. Creaser. Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta CANADA, T6G 2E3

Introduction

Geochemical studies of clastic sedimentary rocks using elements such as the rare earth elements (REE), Th and Sc have proven to be a powerful tool in understanding the chemical composition of the Earth's upper crust, because these elements are likely transferred quantitatively to clastic sediments, and the efficient mixing of detritus during sedimentary processes provides the average geochemical signature of the exposed crust (Taylor and McLennan 1985, McLennan et al. 1990, McLennan and Hemming 1992). Nd isotopic analyses of clastic sedimentary rocks give an additional dimension to these studies by providing information about the average age of the sedimentary provenance (e.g. Goldstein et al. 1984, O'Nions 1984, Nelson and DePaolo 1988, McLennan et al. 1990, Zhao et al. 1992). The combination of these techniques is particularly useful in understanding the nature of Archean crust because in many cases, the original source rocks which provided the detritus to the sedimentary basin are no longer preserved.

The Slave structural province, Canada, is an Archean granite-greenstone terrane with an unusual abundance of late Archean (~2.66 Ga) turbiditic greywacke-mudstone that comprises ~80% of the supracrustal assemblage (Padgham and Fyson 1992). These turbidites are exceptionally well preserved and widespread (e.g. Henderson 1985), and were the focus of petrographical and geochemical study by Jenner et al. (1981). These rocks are the key unit in the evolution of Slave province as they provide a record of the exposed crust at the time of ~2.7 Ga crust formation/consolidation, and are also likely protoliths for some of the ~2.59 Ga granitoids generated during regional deformation and metamorphism. However, since the classic study of Jenner et al. (1981), it has been recognized that the Slave province can be divided into western and eastern Slave provinces by the presence of >2.8 Ga basement in the western Slave province, defined by Nd-Pb isotopes and U-Pb geochronology (Thorpe et al. 1992, Davis and Hegner 1992, Villeneuve and van Breemen 1994, Davis et al. 1996). The role of this basement in the 2.72-2.59 Ga crust formation/consolidation is the focus of detailed studies (Yamashita et al. 1995, 1997, 1998, Yamashita and Creaser 1996). The widespread turbidites of the Slave province therefore provide a record of exposed basement and ~2.7 Ga juvenile crust during the 2.6-2.7 Ga period but so far have not been studied in detail for the evidence for contribution of basement in their provenance. The Burwash Formation turbidites from the Yellowknife area of the Slave province are ideal samples for this study because rocks can be collected from a wide area (Fig. 3.1) and for turbidites metamorphosed to greenschist facies, different units of the Bouma sequence can be sampled separately. In addition, previous geochemical and petrographical study of these turbidites provide some constraints on the sources of these sediments (Jenner et al. 1981). Although the Nd isotopic study of different units of the Bouma sequence has been carried out on modern turbidites (McLennan et al. 1989), such studies on Archean turbidites are absent.

Three major objectives of this study are to: (1) chemically as well as chronologically constrain the sources of the Burwash Formation turbidites collected from the southern Slave province, (2) evaluate the heterogeneity of geochemical and isotopic signatures of different units of the Bouma sequence within a single turbidite package and (3) evaluate the possible change in the chemical composition of the Earth's upper continental crust at the Archean/Proterozoic transition (Taylor and McLennan 1985, McLennan and Hemming 1992).

The results of this study should constrain the role of pre-2.8 Ga basement during the 2.7 Ga tectonomagmatic event in the Archean Slave province and help answer the question on whether the basement lies beneath the entire western Slave province or is restricted to certain areas of the western Slave province.

Geological setting

Regional geology

The Slave province is an Archean craton located in the northwestern part of the Canadian shield (Fig. 3.1). It covers an area of approximately 190,000 km^2 (Henderson 1985, Fyson and Helmstaedt 1986, Hoffman 1989). Rocks of the Slave province can be classified into the following three lithotectonic units.

(1) Pre-YKSG basement.

The supracrustal rocks of the Slave province were originally grouped as Yellowknife Supergroup by Henderson (1970). However, it is now recognized that there are three distinct units of supracrustal rocks in the Slave province. The oldest supracrustal rocks comprise quartzite, oligomictic conglomerate, banded iron formation, felsic volcanic rocks and minor ultramafic rocks (Roscoe et al. 1989, Hrabi et al. 1993, 1994, Isachsen and Bowring 1997). These supracrustal rocks and >2.8 to 4.0 Ga gneisses and granitoids are found in various areas of the Slave province west of the Nd-Pb isotopic boundary at ~111°W, but have not yet been documented from the eastern Slave province (Henderson et al. 1982, Henderson 1985, Frith et al. 1986, Bowring et al. 1989b, Lambert and van Breemen 1991, Isachsen and Bowring 1994, Villeneuve and van Breemen 1994).

(2) Yellowknife Supergroup supracrustal rocks and syn-volcanic plutons.

Extensive accumulation of supracrustal rocks, dominated by greywacke-mudstone sequences, took place between 2.65-2.72 Ga (Henderson 1970, 1985, Isachsen et al. 1991, Padgham and Fyson 1992, Isachsen and Bowring 1994). These rocks are the most abundant supracrustal sequence in the Slave province and comprise ~30% of the craton (Padgham and Fyson 1992). The Burwash formation turbidites and the volcanic rocks analyzed in this study are part of the Yellowknife Supergroup, and their details are discussed in the next section. The 2.72 to 2.65 Ga syn-volcanic plutons are generally intermediate in composition (Davis et al. 1994), and are distinguished from syn- to post-deformation granitoids by a 2620-2645 Ma magmatic gap (van Breemen et al. 1992, Villeneuve et al. 1997). In addition to the 2.65-2.72 Ga supracrustal sequence, ~2.59 to

2.62 Ga sandstones and polymictic conglomerates that unconformably overlie the ~2.7 Ga supracrustal rocks are now recognized (Fyson and Helmstaedt 1986, Isachsen and Bowring 1994).

(3) Late-Archean syn- to post-deformation granitoids.

The late-Archean plutonic rocks in the Slave province range in age from 2.62 to 2.58 Ga (van Breemen et al. 1992, Villeneuve and van Breemen 1994). Together with syn-volcanic plutons, these plutonic suites comprise ~65% of the craton (Padgham and Fyson 1992). The late-Archean granitoids are typically divided into 2.62-2.60 Ga syn-deformation granitoids and 2.60-2.58 Ga post-deformation granitoids (Davis et al. 1994). The chemical composition of these granitoids changes from generally metaluminous tonalite to peraluminous granite with time (Davis et al. 1994). The generation of these granitoids is closely related to the 2.63-2.58 Ga regional high T-low P type metamorphism and deformation (Davis et al. 1994, Yamashita et al. 1998). The metamorphic grade ranges from greenschist-amphibolite-granulite facies (Henderson and Schaan 1993, Thompson 1978, Pehrsson and Chacko 1997).

Some of the models proposed to explain the tectonic history of the Slave province are (1) intracontinental rifts (Henderson 1981), (2) collision between island arc and continent (Kusky 1989), (3) rifting of pre-existing continental crust, followed by eastward subduction of oceanic lithosphere (Fyson and Helmstaedt 1988, MacLachlan and Helmstaedt 1995). These models were constructed mainly based on field studies.

Geological/geochronological framework of the Yellowknife area

The detailed geological framework of the Yellowknife area is summarized in Henderson (1985) and Helmstaedt and Padgham (1986). In the Yellowknife area, the pre-YKSG supracrustal sequence comprises quartzite (<2.92 Ga), felsic volcaniclastic rock (~2.82 Ga) and banded iron formation assemblage which unconformably overlies the ~3007-3017 Ma tonalite gneiss, 2929 Ma granodiorite gneiss and 2945 Ma mylonite gneiss of the Anton complex (Isachsen and Bowring 1997)

The YKSG volcanic rocks in the Yellowknife area are classified into mafic volcanic and intrusive rocks of the Kam Group and intermediate to felsic volcanic rocks of the Banting Group (Fig. 3.2; Henderson 1985, Helmstaedt and Padgham 1986). The Kam Group is further subdivided into lower Kam Group (Chan Formation) and upper Kam Group (Crestaurum, Townsite and Yellowknife Bay Formation), which are separated by the 2722±2 Ma Ranney chert (Helmstaedt and Padgham 1986, Isachsen and Bowring 1994, 1997). The volcanic rocks of the upper Kam Group were deposited between 2722 to 2701 Ma, as indicated from the U-Pb zircon ages of tuffaceous beds within the upper Kam Group (Isachsen and Bowring 1994, 1997). The U-Pb zircon ages of felsic volcanic rocks from the Banting Group are 2662 ± 3 and 2664 ± 1 Ma. Occurrence of zircon with cores older than 2.74 Ga in the Banting Group volcanic rocks indicate that these volcanic rocks were, at least in part, erupted though the basement (Isachsen and Bowring 1994, 1997).

The volcanic rocks of the Yellowknife greenstone belt are bordered to the east by sedimentary rocks of the Burwash Formation (Henderson 1985). These sedimentary rocks consist mainly of greywacke-mudstone pairs and are characteristic of sediments deposited by turbidity currents (Henderson 1985, Waldron and Bleeker 1997). The metamorphic grade of these turbidites ranges from greenschist to amphibolite facies. Where these turbidites are metamorphosed to greenschist facies, internal structures of the Bouma sequence are very well preserved (Henderson 1985). Where metamorphosed to amphibolite facies, the internal structures are usually lost, but the pelitic layers of the turbidite can be distinguished from the quartzofeldspathic layer by the occurrence of cordierite porphyroblasts (Henderson 1985). The depositional age of the Burwash Formation turbidite is constrained by the U-Pb age of interbedded felsic volcanic rock (2661 ± 1 Ma; Mortensen et al. 1991) at Clan Lake, approximately 50 km north of Yellowknife.

The supracrustal rocks of the Yellowknife Supergroup are intruded by syn-deformation (Defeat suite) and post-deformation granitoids (Stagg, Awry and Prosperous suite). The U-Pb ages of Defeat suite granitoids are 2621+5/-8 and 2620 ± 8 Ma, whereas the age of the Stagg suite granitoid is 2588 ± 7 Ma (Henderson et al. 1987, van Breemen et al. 1992).

The youngest supracrustal rock in the Yellowknife area is the conglomerate and sandstone of the Jackson Lake Formation. A granitic clast from a conglomerate yielded a U-Pb age of 2605 Ma, representing the maximum deposition age of this Formation (Isachsen and Bowring 1994).

The volcanic and sedimentary rocks for this study were collected mainly from areas north and east of Yellowknife (sample Y-4 is an amphibolite but it will be grouped here as volcanic rock for convenience). The sample locations are shown in Figure 3.3, and their descriptions are given in the appendix. For turbidites metamorphosed to greenschist facies, different units of the Bouma sequence (see Fig 3.4) were sampled separately. For those metamorphosed to amphibolite facies, only the lower sand unit (labeled as Ta for convenience) and the upper shale unit (labeled as Te) were collected.

Analytical methods

Major- and trace-element analyses were performed at the Washington State University using XRF and ICP-MS. The analytical method for XRF and ICP-MS analyses are described in Hooper et al. (1993). Sm-Nd isotopic analyses were carried out at the University of Alberta. Powdered rock samples were weighed into clean Teflon vials and were spiked with mixed ^{149}Sm-^{150}Nd tracer. They were dissolved in a 5:1 mixture 24N HF and 16N HNO_3 for 7 days at 180°C. Sm and Nd were separated by passing through cation and HEDEP columns (see Creaser et al. 1997 for details). Sm concentrations were measured using a single collector VG MM30 thermal ionization mass spectrometer. Nd isotopic ratios and concentrations were measured using a five-collector VG354 thermal ionization mass spectrometer operated in multidynamic mode. All Nd isotopic ratios were normalized to $^{146}Nd/^{144}Nd=0.7219$. The value obtained for La Jolla standard during this study was 0.511848 ± 8 and the external reproducibility for the University of Alberta in-housed standard (Nd-oxide standard) was 0.000016 (2σ). The analytical precisions for the

Sm and Nd concentrations are better than 0.2%. The total analytical blanks for Sm and Nd were <500 pg.

Results

Major and trace element geochemistry

Results for major- and trace-element analyses are summarized in Table 3.1. The mafic to intermediate volcanic rocks of the upper Kam group are characterized by flat or slightly light rare earth element (LREE) enriched REE pattern without negative Eu/Eu* (Fig. 3.5). One amphibolite sample from the Banting group (Y-4) also has a similar REE pattern (Fig. 3.5a). A dacite sample from the upper Kam Group (Y-5) has a LREE enriched pattern but this sample is also characterized by a lack of negative Eu/Eu* (Fig. 3.5c). This may imply that fractional crystallization of plagioclase and/or partial melting involving plagioclase as a residual phase was unimportant in the genesis of some felsic volcanic rocks in this area.

The chemical index of alteration (CIA=[Al_2O_3/(Al_2O_3+CaO*+Na_2O+K_2O)] in molecular proportions, *=CaO in silicate fraction, CIA of unweathered rocks~50: Nesbitt and Young 1982) of the Burwash Formation turbidites ranges from 52-71 (majority between 54 and 68), indicating general chemical immaturity. The upper units of the Bouma sequence always have higher CIA (Ta<Tc<Td<Te), reflecting a higher proportion of clay minerals in the upper portion of the turbidite sequence. However, these turbidites are metamorphosed from greenschist to amphibolite facies and certain major elements (particularly alkaline elements) may have remobilized after sedimentation.

Trace elements such as REE, Sc, Th and Zr are useful in identifying the provenance of Burwash Formation turbidites as they are transferred quantitatively during sedimentation (Taylor and McLennan 1985). In order to cancel the dilution of trace element concentration by quartz in the high SiO_2 units, ratios such as Th/Sc and La/Sc, instead of absolute concentrations, are often used.

In spite of a large sampling area (see Fig. 3.3), the trace element characteristics of the Burwash Formation turbidites are remarkably similar (Fig. 3.6, 3.7). With the exception of a few samples (e.g. Y-16, 17) these turbidites have REE characterized by LREE enrichment (La_N/Sm_N ~ 3.7), relatively flat HREE (Gd_N/Yb_N ~ 1.8) and very small or no Eu/Eu* (Fig. 3.6, 3.7). The enrichment relative to chondritic value is also constant at x100 chondritic value for La and x6~x10 for HREE. There is no obvious difference in the REE patterns between samples metamorphosed to greenschist and amphibolite facies, suggesting that there was very little or no significant removal or addition of REE during the 2.6 Ga regional metamorphism. A parallel shift in REE pattern between the Ta (sand) unit and Tc~Te (sand-shale mixture and shale) units of the Bouma sequence is most likely due to dilution of REE concentration by quartz in the Ta unit.

Th/Sc and La/Sc ratios of sedimentary rocks are useful in distinguishing the degree of geochemical differentiation of the sedimentary provenance because Th and La are incompatible elements whereas Sc is a compatible element during igneous differentiation. Shown in Figure 3.8 are Th/Sc vs. La/Sc ratios of volcanic and sedimentary rocks from the Yellowknife area. Most turbidites have ratios intermediate between the YKSG mafic and the felsic volcanic rocks (as well as average upper continental crust of Taylor and McLennan 1985), which is compatible with the suggestion that these turbidites are dominated by a volcanic source, likely similar to Kam and Banting Group, with minor input from a granitic source (Jenner et al. 1981). However, it is also clear from this diagram that Tc~Te units of the turbidite sequence are generally lower in Th/Sc and La/Sc ratios, suggesting a greater input from a mafic source in the Tc~Te (shale) units relative to Ta (sand) units

Figure 3.9 shows the major/trace element concentration of Burwash Formation turbidites (this study) normalized to the values of "model Burwash Formation" of Jenner et al. (1981), which is a mixture of 20% mafic-intermediate volcanic rock + 55% felsic volcanic rocks + 25% granitic rock. It can be seen from this diagram that the Ta units of turbidites are generally in good agreement with the model chemical composition predicted by Jenner et al. (1981). However, Tc~Te units have higher concentrations in most of the elements except SiO_2. Part of this is a result of dilution by quartz in the Ta unit, but large enrichment in MgO, K_2O, Ba, Ni and Cr may require an additional explanation.

Enrichment of K_2O and Ba may be explained by adsorption of these elements on clays during the weathering process (Nesbitt and Young 1984). Enrichment of MgO, Ni and Cr, on the other hand, is probably an indication of higher input from the mafic volcanic rock in the Tc~Te units as hinted from Th/Sc and La/Sc ratios.

Several important points mentioned by Jenner et al. (1981), such as lower Ca, Mn and higher Ni, Cr in the turbidite samples relative to the model composition are also confirmed. As suggested by Jenner et al. (1981), low Ca may be a result of amphibole and pyroxene breakdown during diagenesis. However, the reason for the Mn depletion remains unknown. Overall enrichment of Ni and Cr in turbidite samples may indicate a small input from an ultramafic source (Jenner et al. 1981). This may be a reasonable explanation because ultramafic flows and sills are reported locally in the Slave province (Hrabi et al. 1993, 1994).

Nd isotopic analyses

Whereas the major/trace element data of turbidites provide constraints for the geochemical characteristics of the sedimentary provenance, Nd isotopic data can provide additional information that may allow constraints of the average age of the sedimentary provenance. The results for the Nd isotopic analyses are listed in Table 3.2.

The εNd(2.66) values of the Kam and Banting group volcanic rocks show a significant range from -4.4 to +1.7. These values are much lower than the value of the model depleted mantle at this time (εNd(2.66) ~+5; Goldstein et al. 1984). Similar low εNd_T values in ~2.7 Ga volcanic rocks have also been documented from other areas of the Slave province (Yamashita et al. 1995, 1997, Yamashita and Creaser 1996), implying the existence of pre-2.8 Ga basement near the volcanic belt at the time of 2.7 Ga tectonomagmatic event in the western Slave province. Although the exposed basement in the Yellowknife area is small, the interaction of 2.7 Ga volcanic rocks with pre-2.8 Ga basement is also supported by recent geochronological work (Isachsen and Bowring 1994, 1997).

The $\varepsilon Nd_{T(=2.66Ga)}$ values of the turbidites range from -1.7 to +3.1 with TDM (depleted mantle model age) ranging from 2.8 to 3.2 Ga. Unlike the REE patterns that showed overall similarities between different samples, the εNd_T values of these turbidites are quite variable (Fig. 3.10). This is particularly the case for turbidites metamorphosed to greenschist facies that may show up to 2.7 epsilon unit difference even within a single Bouma sequence. For the greenschist-facies turbidites where both the shale (Te) and sand (Ta) units were collected, four out of five samples show lower εNd_T in the Te unit with a difference of >0.7 epsilon units (i.e. difference well outside of analytical uncertainty after propagating the errors from both $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$). This, however, is not the case for the amphibolite-facies turbidites (Fig. 3.10). Although Ta and Te units may not be from the same turbidite sequence in some samples (Y-19, 20, 22), only three out of eight turbidite sequences have a range of εNd_T values >0.7 epsilon units. There are three possible reasons for this. First is the partial resetting (or homogenization) of Nd isotopic composition on a sample size scale during the ~2.6 Ga amphibolite-facies metamorphism. The second is simply related to a sampling problem; since most of the amphibolite facies turbidites have completely lost their internal structure, only the upper shale and the lower sand units, as opposed to different units of the Bouma sequence, were sampled. The lower sand unit may contain Ta, Tb and part of Tc unit while the upper shale unit may contain Tc~Te unit. This may result in a more homogenized sampling of different units. The third possibility is the homogenization of detritus during the transportation and sedimentation. Since these turbidites are thought to have been derived from the west (Fig. 3.3; Henderson 1975), most of the amphibolite-facies turbidites are farther away from their sources. This may result in more efficient mixing of detritus between sand and shale units in the more eastern amphibolite facies turbidites. It is difficult to distinguish between these three possibilities as it is also possible that all of these processes were involved. However, the general tendency for the Ta units to have higher εNd_T values compared to the Te units seems to hold for amphibolite facies turbidites where the εNd_T values of Ta and Te units are distinguishable.

Discussion

Evaluation of heavy mineral concentration as a process for the shift in εNd_T

A clear difference in the εNd_T values between different units of the Bouma sequence may be a result of (1) heavy minerals, particularly those enriched in REE, concentrating in certain units (most likely sand) of the Bouma sequence and/or (2) slight differences in the crustal residence age of the source for different units of the Bouma sequence. Although it is difficult to distinguish between the two based only on Nd isotopic composition, combined Nd isotopic and trace element characteristics of these rocks can help resolve this issue.

Some of the important heavy minerals in sedimentary rocks that may cause a shift in the εNd_T values are apatite, allanite, monazite, and zircon (McLennan 1989). Because the densities of these minerals are higher than major rock-forming minerals such as quartz and plagioclase, they are likely to be concentrated in the lower sand unit of the turbidites if the sorting of heavy minerals takes place during the transportation and deposition of detritus.

Of the minerals mentioned above, allanite and monazite are characterized by extremely high concentration of LREE , Th and Y with pronounced negative Eu/Eu* and high Gd_N/Yb_N (Bea 1996). The La_N/Sm_N ratio (and hence the Sm/Nd ratio) of monazite is highly variable, but on average they tend to be similar or slightly higher than a typical turbidite sample (Bea 1996). If the Sm/Nd ratio of monazite is higher than typical turbidite sample, concentration of monazite may lead to slight increase in the εNd_T values (see Fig. 3.11). Concentration of monazite in the Ta units of the turbidites, however, seems unlikely for the samples analyzed in this study because the combination of high LREE, Y, Th and negative Eu/Eu* is not observed in Ta units of these turbidites. In fact, the concentration of Y and LREE is lower in the Ta unit compared to Tc~Te units for most samples where the range of $\varepsilon Nd_T > 0.7$, and the large negative Eu/Eu* that will be expected from monazite concentration is typically absent in the Ta units. Although a small negative Eu/Eu* is observed in Ta unit of some turbidite sequences, the degree of negative Eu/Eu* is the same as those of Tc~Te units of the same turbidite sequence, suggesting that the concentration of monazite, in general, is not responsible for the lower εNd_T values of the Ta units. This is further supported by the fact that high Gd_N/Yb_N (>2.0), which will result from even a small addition of monazite, is not observed in the Ta units.

Allanite has trace element characteristics very similar to monazite but with slightly lower Y and Th concentration and lower La_N/Sm_N (and Sm/Nd ratio; Bea 1996). For the same reason mentioned above, concentration of allanite in the lower unit is probably not the main reason for the higher εNd_T in Ta units. Also, since the Sm/Nd ratio of allanite is lower than typical turbidite samples, concentration of allanite in Ta will result in lower εNd_T values, which is opposite of what is observed (Fig. 3.11).

Examining the possibility of zircon concentrating in the Ta unit is particularly important because zircon is highly resistant to weathering and it may survive through several cycles of sedimentation. Zircon grains as old as ~3.5 Ga have been reported from some sedimentary rocks from the Slave province (Villeneuve and van Breemen 1994). Zircon is characterized by high Sm/Nd ratio (Bea 1996), which will lead to an increase in the εNd_T values if they are concentrated in the Ta unit of the Bouma sequence (Fig. 3.11). Two ways of detecting the anomalously high zircon concentration in the Ta units are by examining the Zr concentration and the Gd_N/Yb_N ratio (McLennan 1989). Addition of zircon in the Ta unit will show up as a high Zr concentration (typical post-Archean shales have 200±100 ppm Zr) and low Gd_N/Yb_N ratio (<1.0; McLennan 1989). There are three sets of turbidite samples (Y-15, Y-17 and Y-18) which contain unit(s) with relatively low Gd_N/Yb_N ratio (<1.2). Of these samples Y-18 has a low Gd_N/Yb_N ratio in the upper Te unit but not in the Ta unit. Y-17 has low Gd_N/Yb_N ratio in the Ta unit, but this sample has low Gd_N/Yb_N ratio in Td and Te units as well. Furthermore, Zr concentration of this sample increases from Ta to Te (Fig. 3.12). It seems, therefore, unlikely that zircon is concentrating in the Ta units of samples Y-17 and Y-18. For sample Y-15, extremely low Gd_N/Yb_N ratio may indicate a high concentration of zircon (Fig. 3.12). However, the Zr content of this sample is not much higher than other samples and thus the concentration of zircon is not fully supported. Only one sample (Y-16D) has anomalously high Zr content. This sample is also characterized by relatively low Gd_N/Yb_N ratio (~1.3). However, despite the fact that Zr content decreases abruptly from Td to Te, there is no change in the Gd_N/Yb_N ratio. This may hint toward the existence of a host(s) other than zircon being responsible for the difference in Zr content between different units of each turbidite sequence. For samples other than Y-15, 16, 17 and 18, the Gd_N/Yb_N ratios are very close to 2.0, and there are no significant differences in the Gd_N/Yb_N ratio between different units of the Bouma sequence. It can therefore be concluded that for the turbidites analyzed in this study, the concentration of zircon in the Ta unit is unlikely to be the cause of higher εNd_T in the Ta units compared to Tc~Te units.

The REE pattern of apatite is highly variable (Bea 1996), making it one the most difficult minerals to detect. Apatites from peraluminous granitoids are characterized by flat REE pattern with pronounced negative Eu/Eu* whereas apatites from metaluminous granitoids have LREE depleted, flat HREE pattern with small (~0.7) Eu/Eu*. The REE pattern of apatites from peralkaline granitoids shows a steep slope from La to Lu with La_N/Lu_N of 50-100. Apatite is also characterized by high La/Th ratio (5 to >500), so small addition of apatite into the Ta unit may result in an elevated La/Th ratio of the Ta unit. Only in three samples (Y-13, Y-18 and Y-26) is the La/Th ratio of the Ta unit significantly higher than those of Tc~Te units. Of these samples Y-13 shows no obvious difference in the εNd_T values between Ta and Te units. For samples Y-18 and Y-26, a high La_N/Lu_N ratio (which may be expected from apatite from peralkaline granitoids) or low La_N/Sm_N ratio (expected from apatite from per- and metaluminous granitoids) is absent so the possibly of additional apatite in the Ta unit is unlikely to be the cause of the higher εNd_T value. There are, however, two cases where the addition of apatite may not show up as an elevated La/Th ratio. First is when the concentration of LREE in apatite is low (as is the case with apatites from metaluminous granitoids), and second is when the La/Th ratio of the apatite is extremely low for apatite (La/Th ~ 5), such that it is indistinguishable from the La/Th ratio of a typical turbidite sample. The former possibility can be examined by looking at the Eu/Eu* of Ta units because addition of such apatite will most likely produce a negative Eu/Eu*. The latter possibility, on the other hand, cannot be detected. However, occurrence of such apatite seems to be very rare as most apatites have a La/Th ratio >10 (Bea 1996).

It can therefore be concluded that the concentration of heavy minerals in the Ta unit of Bouma sequence is not the reason for higher εNd_T values of this unit compared to Tc~Te units, and therefore this significant Nd isotope difference is interpreted to reflect unmixing of isotopically distinct detritus during sedimentation as discussed below.

Source of the Burwash Formation Turbidites

Based on the petrographical and major/trace element geochemical study of the Burwash Formation turbidites, Jenner et al. (1981) suggested that the source of these turbidites is a mixture of 20% mafic-intermediate volcanic rocks, 55% felsic volcanic rocks and 25% granitic rocks, likely similar to those exposed in the Yellowknife area. The data presented here are in good agreement with the model chemical composition proposed by Jenner et al. (1981) and support the suggestion that these turbidites are largely derived from a local volcanic-dominated source (Fig. 3.9). However, further constraints on the nature of source rock can be made by incorporating the new Nd isotopic data.

A mechanism other than heavy mineral concentration which could lead to a difference in the εNd_T values between the Ta and Tc~Te units is the unmixing of detritus derived from different sources (with different εNd_T values) during the transportation and sedimentation (McLennan et al. 1989); such a mechanism has been reported from modern deep-sea turbidites by McLennan et al. (1989). In this case, the separation of provenance components may be a result of original contrast in the grain size (i.e. fine grained material such as volcanic rock being incorporated into mud) or preferential breakdown of, for example, fine grained and/or unstable materials such as volcanic glass (McLennan et al. 1989). In either case, if the source of turbidites is a mixture of volcanic and granitic (or quartzofeldspathic) rocks, a higher proportion of material derived from volcanic rocks will be expected in the upper Tc~Te units.

Shown in Figure 3.13 are the $^{147}Sm/^{144}Nd$ and $\varepsilon Nd(2.66)$ values of volcanic and sedimentary rocks analyzed in this study, together with the $\varepsilon Nd(2.66)$ values of pre-2.8 Ga basement from other areas of the Slave province (Bowring et al. 1989, Yamashita et al. 1995, 1997). Although the basement samples from the Yellowknife area were not analyzed for Nd isotopes during this study, its existence and interaction with the ~2.7 Ga volcanic rocks has been shown by recent geochronological and isotopic study (Thorpe et al. 1992, Isachsen and Bowring 1994, 1997).

It can be seen from Figure 3.13a that the Nd isotopic composition of the volcanic rocks can be explained by a combination of fractional crystallization and assimilation of pre-2.8 Ga basement (AFC system). On the other hand, the isotopic signature of the Burwash Formation turbidites form a distinct trend (Fig. 3.13a, b). As discussed previously, the spread in the εNd(2.66) values between coarser unit (Ta) and finer units (Tc~Te) is interpreted to be the result of unmixing of detritus derived from geochemically and isotopically different sources. In this case, the low εNd(2.66) endmember is probably not the pre-2.8 Ga basement because the unmixing of detritus derived from pre-2.8 Ga basement and the volcanic rock should create a scatter of data that lies between the two. Rather, the trend defined by the isotopic signatures of Burwash Formation turbidites is a straight line with negative slope, with finer units having higher $^{147}Sm/^{144}Nd$ ratio and lower εNd(2.66) value. The isotopic composition of Tc~Te units points towards the envelope defined by volcanic rocks which are most contaminated by the older basement (Fig. 3.13b), suggesting that low εNd(2.66) endmember of the turbidite is a "crustally contaminated 2.7 Ga volcanic rock", similar to those exposed in the Yellowknife area today. This is in accord with the suggestion that detritus derived from volcanic rocks may preferentially be incorporated into the upper shale units (McLennan et al. 1989). Although a distal input of more ancient materials may be an alternative explanation for the lower εNd_T values of the Tc~Te units, the fact that Td units (e.g. sample Y-16 and 17) have lower εNd_T values compared to the Te units suggest that the pelagic material which is expected to accumulate at the upper most layer of the turbidite sequence has juvenile or high εNd_T signature. This, combined with the model of Jenner et al. (1981), leads to a conclusion that the coarser endmember (i.e. low $^{147}Sm/^{144}Nd$ and high εNd_T endmember) of the turbidite is granitic rocks that have not been severely contaminated by the older basement. This unmixing model is also compatible with the suggestion that there may be slightly higher input from the mafic source in the Tc~Te units. This is demonstrated clearly on the Th/Sc vs. εNd(2.66) diagram (Fig. 3.14).

Although no ~2.66 to 2.72 Ga syn-volcanic granitoid rocks with high εNd(2.66) values have been recognized from the Yellowknife area, it seems reasonable to assume that such rock did exist in the past because (1) high εNd(2.66) syn-volcanic granitoid rocks are actually found in other areas of the western Slave province (Davis and Hegner 1992, Davis et al. 1994, Perks 1997) and (2) some of the post-deformation (c.a. 2.59 Ga) granitoids which are thought to have derived, at least in part, from partial melting of sedimentary sources, have high εNd(2.66) value (Davis et al. 1994, Yamashita et al. 1998). Some candidates for the high εNd(2.66) endmember are the syn-volcanic granitoids of the Olga suite in the central Slave province (Davis et al. 1994) and felsic syn-volcanic plutons (and one volcanic rock) from the Wijinnedi/MacNaughton Lake areas (Perks 1997 and unpublished data). These rocks are characterized by a high εNd(2.66) values (-0.9 to +3.5) and a REE pattern without pronounced negative Eu/Eu*, which best explains the geochemical and isotopic characteristics of the Burwash Formation turbidites (Fig. 3.14).

Regional implications

The paleocurrent analyses of the Burwash Formation turbidites indicate that these sediments were transported mainly from west to east (Henderson 1975). If this is the case, the existence of high εNd_T granitic rocks that was predicted from the unmixing model is particularly important because it implies an existence of "juvenile" sialic crust in the western Slave province prior to the deposition of YKSG sedimentary rocks. Although it has been proposed that the pre-2.8 Ga basement in the Slave province is not a single sialic crust but is a collage of crustal packages with different ages (Davis et al. 1996, Yamashita et al. 1998), its distribution within the western Slave province (i.e. basement of what age is located in which part of the Slave province) is still poorly constrained. The fundamental question is whether the pre-2.8 Ga basement lies beneath the entire western Slave province or is restricted to certain areas. If the pre-2.8 Ga basement lies beneath the entire western

Slave province, it would probably be difficult to generate syn-volcanic granitoids with minimum or no sign of crustal contamination while the volcanic rocks are variably contaminated. Thus, it can be speculated that there are areas within the western Slave province that are not underlain by the pre-2.8 Ga basement. This is further supported by the relatively juvenile Nd isotopic signature of some late-Archean granitoids from the western Slave province (Chapter 4). Further Nd isotopic study of volcanic and sedimentary rocks from other areas of the western Slave province will be required to test the validity of this hypothesis.

At this point, the most important conclusion is that the pre-2.8 basement did exist in the Yellowknife area during the 2.7 Ga tectonomagmatic event and the Nd isotopic signature of this basement showed up more strongly in the upper Tc~Te units of the Burwash Formation turbidites. The isotopic signature of less contaminated quartzo-feldspathic rocks, which presumably was located farther to the west, shows up more strongly in the lower Ta unit of the turbidite. Although this example from the Yellowknife area may not be representative of all turbidite sequences, our findings warn against tacitly assuming that the mud unit (i.e. Tc~Te) of the turbidite represents the average geochemical and isotopic signatures of the exposed crust in the area. In the case of turbidite samples, careful examination of both the sand and the shale unit may be required to fully understand the geochemical and isotopic characteristics of the provenance (McLennan et al. 1989).

Implication for the geochemistry of Archean sedimentary rocks

Compared to modern turbidites, Archean turbidites are generally characterized by higher Gd_N/Yb_N, lower Th/Sc ratio and lack of pronounced Eu/Eu* (Taylor and McLennan 1985, McLennan and Taylor 1991, McLennan and Hemming 1992). These differences are interpreted as an evidence for a change in the average chemical composition of the upper crust at the Archean/Proterozoic transition (Taylor and McLennan 1985, McLennan and Taylor 1991, McLennan and Hemming 1992). The REE pattern with high Gd_N/Yb_N ratio and small or no negative Eu/Eu* is commonly seen in Archean tonalite - trondhjemite - granodiorite suites (TTG; Martin 1986, 1994) which are thought to be the product of dehydration melting of tholeiitic garnet amphibolite (or hornblende eclogite) with garnet and/or hornblende as a major residual phase (Martin 1986, 1994, Rapp et al., 1992). Wide occurrence of these rocks most likely reflects the higher geothermal gradient in the Archean compared to present day (Martin 1986, 1994). The abrupt increase in the Th/Sc ratio at the Archean/Proterozoic boundary, on the other hand, is generally interpreted as a result of the change in chemical composition of the upper crust from generally mafic (high Th/Sc) to more fractionated felsic (low Th/Sc) composition (Taylor and McLennan 1985, McLennan and Taylor 1991, McLennan and Hemming 1992).

The Gd_N/Yb_N ratio and Eu/Eu* of the Burwash Formation turbidites are clearly distinct from those of modern turbidites as shown in Figure 3.15. It is characterized by higher Gd_N/Yb_N ratio (~2.0) and very small or no negative Eu/Eu*. Turbidites with such geochemical characteristics are the most common variety of the Archean sedimentary rocks and are thought to have derived mainly from bimodal mixing of Archean tonalite-trondhjemite-granodiorite plus felsic volcanic rocks and mafic-ultramafic volcanic rocks (Taylor and McLennan 1985, McLennan 1989). Although the mixing proportion is 20% mafic-intermediate volcanic rocks and 80% felsic volcanic/plutonic rocks for the Burwash Formation turbidites, this should only result in slightly higher Gd_N/Yb_N with no difference in the Eu/Eu* as long as the plutonic source has no Eu/Eu*. While the U-Pb geochronological and Nd isotopic study of sedimentary rocks on a regional scale (i.e. central to northern Slave province) has shown that intracrustal processes such as partial melting and/or assimilation of pre-existing crust was important in the early to mid-Archean crustal evolution of the Slave province (see Chapter 5), the lack of Eu/Eu* in the Burwash Formation turbidites (which comprise a significant portion of the supracrustal sequence in the southwestern Slave province) supports the idea that intracrustal processes "involving plagioclase" were less important in the Archean compared to the present day.

The Th/Sc ratio of the Burwash Formation turbidites lies within the scatter of data defined by modern turbidite samples (Fig. 3.15). However, for a given Th concentration, the Burwash Formation turbidites tend to have slightly higher Sc concentration, indicative of a generally more mafic provenance. This may indicate that the proportion of crustal recycling relative to addition of juvenile mafic magma was slightly lower in the Archean compared to present day. McLennan and Hemming (1991) concluded that on average, approximately 15-25% of sedimentary system is open to addition of new material during the Phanerozoic. If we use their model, it seems that >50% of the Burwash Formation basin was open to the addition of new material.

Conclusions

1. The major/trace element characteristics of the Burwash Formation turbidites analyzed in this study are in good agreement with the "model chemical composition" of Jenner et al. (1981), which represents a mixture of 20% mafic-intermediate volcanic rock + 55% felsic volcanic rock + 25% felsic plutonic rock.
2. For the Burwash Formation turbidites, the REE pattern is generally similar between different units of the Bouma sequence. However, small differences in the Th/Sc ratio may indicate slightly larger input from a mafic (or ultramafic) source in the upper mud (Tc~Te) units relative to sand (Ta) unit.
3. There are distinguishable differences in the εNd_T values between different units of the Bouma sequence for most of the turbidites metamorphosed to greenschist facies. The geochemical characteristics of these turbidites exclude the possibility of heavy minerals concentrating in the coarse-grained units of the turbidite sequence. Therefore, the differences in the εNd_T values are interpreted to be the result of unmixing of detritus derived from "crustally contaminated volcanic rocks" similar to those exposed in the Yellowknife area today and "less contaminated felsic plutonic rock".
4. The spread in the εNd_T values between different units of the Bouma sequence is less obvious for turbidites metamorphosed to amphibolite facies. This may be due to resetting of Nd isotopic signature at the sample size scale, homogeneous sampling and/or homogenization of sample during distal transportation. However, where the spread in εNd_T is obvious, the general tendency for the lower units of the turbidite to have higher εNd_T value still holds.

5. The unmixing model predicts the existence of felsic plutonic rock that has not been strongly contaminated by the pre-2.8 Ga basement. Existence of such rocks suggests that there are areas within the western Slave province where the influence of the basement was small. This, in turn, suggests that the pre-2.8 basement, although present in different locations of the western Slave province, probably does not lie beneath the entire western Slave province.

6. The geochemical characteristics of the Burwash Formation turbidites (Th/Sc, Eu/Eu*, Gd_N/Yb_N) are different from those of Phanerozoic turbidites. This is in accord with the model that suggests that the average chemical composition of the upper continental crust was slightly different in the Archean.

Table 3.1. Major and trace element compositions of volcanic and sedimentary rocks from the Yellowknife area.

Sample	Y-2A	Y-2B	Y-3	Y-4	Y-5	Y-6	Y-7	Y-8	Y-9
wt (%)									
SiO_2	50.48	50.53	56.90	47.98	67.87	54.51	51.30	68.78	55.03
Al_2O_3	15.35	15.24	13.75	17.74	15.89	14.64	11.42	15.89	14.82
TiO_2	1.278	1.260	0.721	0.866	0.539	0.836	1.093	0.657	0.565
FeO^*	12.23	12.49	8.17	11.80	3.80	11.26	13.28	5.23	7.91
MnO	0.234	0.208	0.163	0.180	0.079	0.178	0.225	0.096	0.202
CaO	9.14	9.42	9.80	9.59	3.95	5.86	10.07	3.78	11.69
MgO	7.07	7.23	6.97	9.27	0.88	6.84	11.55	2.19	7.12
K_2O	1.59	0.95	0.37	0.21	2.67	0.30	0.32	0.41	0.13
Na_2O	2.52	2.56	3.09	2.27	4.16	5.49	0.63	2.89	2.47
P_2O_5	0.108	0.107	0.077	0.085	0.163	0.088	0.117	0.062	0.050
(ppm)									
Ni *	91	92	64	193	20	76	244	79	126
Cr *	147	156	209	219	18	78	850	239	424
Sc *	37	37	36	31	9	33	29	41	40
V *	324	299	244	216	73	235	252	259	219
Ga *	19	19	14	17	19	14	17	15	17
Cu *	105	101	62	70	6	0	62	113	76
Zn *	120	113	79	75	57	102	94	58	51
U	0.23	0.14	0.53	0.21	2.23	0.63	0.39	0.22	0.18
Th	0.83	0.55	1.84	1.05	10.45	2.61	1.27	0.66	0.66
Hf	2.06	2.05	1.60	1.11	3.76	2.06	2.16	1.15	0.99
Zr *	81	80	65	52	150	82	84	47	45
Pb	11.97	2.22	5.06	2.94	6.52	1.56	2.84	2.63	2.63
Ba	412	113	154	67	594	108	129	56	47
Sr *	180	110	136	261	190	117	53	81	88
Rb	54.1	30.0	7.2	4.4	58.4	3.8	5.7	12.3	2.2
Cs	0.4	0.3	0. 5	0.5	1.1	0.1	0.1	0.4	0.1
Nb	4.0	3.8	3.1	1. 8	7.3	4.4	4.3	1.7	1.5
Y	27.87	27.85	18.70	20.58	11.87	17.58	24.85	12.02	12.46
Ta	0.33	0.36	0.39	0.16	0.75	0.36	0.35	0.39	0.17
La	6.11	4.96	7.84	7.58	42.19	9.35	6.39	1.51	3.27
Ce	13.94	12.14	14.49	15.67	74.39	18.75	14.34	3.65	6.55
Pr	1.89	1.78	1.73	1.97	7.62	2.10	1.92	0.56	0.86
Nd	9.22	8.86	7.36	8.59	27.87	8.41	8.91	3.04	3.96
Sm	3.08	3.2	2.27	2.38	4.81	2.26	3.01	1.30	1.32
Eu	1.10	1.17	0.79	0.94	1.17	0.76	0.92	0.49	0.52
Gd	3.91	4.00	2.52	2.66	2.98	2.41	3.56	1.67	1.64
Tb	0.77	0.77	0.49	0.53	0.42	0.48	0.68	0.34	0.33
Dy	4.81	4.85	3.25	3.57	2.36	3.18	4.38	2.25	2.15
Ho	1.03	1.07	0.71	0.76	0.44	0.68	0.94	0.48	0.47
Er	3.10	3.08	2.12	2.35	1.22	2.00	2.77	1.42	1.42
Tm	0.44	0.42	0.30	0.33	0.17	0.28	0.38	0.21	0.20
Yb	2.69	2.69	1.84	2.19	1.06	1.74	2.42	1.32	1.25
Lu	0.42	0.41	0.28	0.35	0.17	0.27	0.38	0.21	0.19

Notes: * Analyses performed using XRF, all others using ICP-MS; Major element analyses are normalized on a volatile free basis; FeO^*=total iron as FeO; A/CNK=molar Al_2O_3/($CaO+Na_2O+K_2O$); $Eu/Eu^*=Eu_N/(Sm_N x Gd_N)^{0.5}$

Sample	Y-10 A1	Y-10 A2	Y-10 C	Y-10 E	Y-11 A	Y-12 A	Y-12 D	Y-12 E	Y-13 A	Y-13 E
wt (%)										
SiO_2	65.61	64.22	63.21	65.07	65.97	68.79	64.15	53.49	69.96	58.78
Al_2O_3	16.74	17.25	17.67	16.52	15.79	15.21	17.10	22.06	15.02	19.62
TiO_2	0.635	0.681	0.718	0.704	0.669	0.609	0.710	0.923	0.521	0.856
FeO^*	5.53	6.11	6.43	7.01	6.00	5.56	6.92	9.46	4.81	7.84
MnO	0.060	0.077	0.085	0.099	0.073	0.062	0.077	0.109	0.051	0.077
CaO	1.84	2.00	2.16	2.00	2.24	1.11	1.71	2.22	0.53	1.41
MgO	2.82	3.11	3.28	3.84	2.92	3.01	3.66	5.34	2.77	4.58
K_2O	2.72	2.58	2.51	2.05	2.27	1.68	2.25	2.99	2.79	3.26
Na_2O	3.89	3.81	3.76	2.54	3.91	3.85	3.25	3.21	3.45	3.42
P_2O_5	0.151	0.160	0.172	0.162	0.155	0.130	0.153	0.186	0.104	0.161
(ppm)										
Ni *	70	77	83	82	69	63	77	104	53	98
Cr *	142	157	167	170	153	128	152	228	105	190
Sc *	14	17	19	17	17	12	19	29	10	17
V *	121	130	134	146	118	110	140	222	98	165
Ga *	19	21	23	23	17	16	22	35	18	24
Cu *	44	63	63	57	69	61	63	41	58	50
Zn *	76	87	85	101	83	77	90	133	54	83
U	1.42	1.51	1.50	1.40	1.61	2.28	1.29	1.54	1.07	1.03
Th	6.50	6.68	6.72	6.28	7.38	6.09	5.46	7.05	4.56	5.83
Hf	3.33	3.40	3.59	3.09	3.75	2.83	2.99	3.54	2.69	3.39
Zr *	138	141	145	121	162	115	116	137	106	127
Pb	12.64	14.04	13.66	11.49	12.99	13.20	12.38	16.70	10.18	11.43
Ba	720	729	757	759	535	465	726	970	620	669
Sr *	347	377	402	352	362	202	255	363	142	246
Rb	86.3	80.3	79.0	53.9	76.1	36.5	60.3	75.3	70.6	83.0
Cs	6.0	5.2	4.7	2.2	5.6	1.1	3.1	3.0	1.3	1.7
Nb	6.2	6.7	7.0	7.2	6.4	6.1	7.1	8.2	5.3	8.8
Y	15.77	17.46	17.79	20.96	16.76	14.59	19.09	23.05	12.43	20.48
Ta	0.44	0.46	0.47	0.47	0.44	0.43	0.48	0.57	0.40	0.58
La	32.30	31.39	33.59	27.20	32.69	27.08	26.16	32.06	21.88	25.17
Ce	60.95	60.22	63.77	54.56	61.87	51.84	53.24	66.62	41.72	49.76
Pr	6.68	6.82	7.09	6.44	6.97	5.68	6.03	7.74	4.54	5.86
Nd	25.69	26.11	27.31	26.16	26.38	21.53	23.95	30.85	17.39	23.87
Sm	5.01	5.19	5.69	5.50	4.90	4.37	5.02	6.72	3.71	5.06
Eu	1.31	1.42	1.54	1.49	1.31	1.04	1.32	1.70	0.87	1.28
Gd	3.59	3.80	4.08	4.41	3.73	3.16	3.90	5.19	2.70	3.94
Tb	0.56	0.60	0.62	0.65	0.55	0.49	0.60	0.83	0.43	0.61
Dy	2.99	3.32	3.51	3.89	3.07	2.74	3.51	4.55	2.43	3.88
Ho	0.59	0.65	0.71	0.76	0.58	0.54	0.70	0.88	0.49	0.77
Er	1.63	1.79	1.89	2.07	1.61	1.51	1.97	2.42	1.40	2.20
Tm	0.22	0.25	0.26	0.28	0.23	0.21	0.26	0.35	0.19	0.31
Yb	1.40	1.58	1.64	1.81	1.50	1.35	1.59	2.15	1.20	2.02
Lu	0.22	0.25	0.26	0.28	0.22	0.20	0.24	0.34	0.19	0.30

Sample	Y-14 A	Y-15 A	Y-16 D	Y-16 E	Y-17 A	Y-17 C	Y-17 D	Y-17 E	Y-18 A	Y-18 E
wt (%)										
SiO_2	66.39	63.17	58.09	56.25	70.80	63.42	56.96	49.15	71.52	57.98
Al_2O_3	15.94	19.07	25.57	25.39	14.59	18.75	21.50	24.57	14.59	21.36
TiO_2	0.657	0.739	0.982	0.865	0.524	0.704	0.902	1.075	0.587	0.840
FeO^*	6.13	6.39	3.12	5.39	5.27	6.20	8.29	10.70	4.67	7.81
MnO	0.068	0.056	0.028	0.045	0.048	0.056	0.088	0.102	0.034	0.050
CaO	1.93	0.98	1.82	1.09	1.27	1.59	1.54	1.02	0.77	0.72
MgO	3.42	3.71	1.80	3.34	2.88	3.48	4.69	6.51	2.63	4.60
K_2O	1.60	3.40	5.83	5.83	1.65	3.06	3.68	4.59	1.66	4.43
Na_2O	3.74	2.35	2.57	1.71	2.85	2.59	2.18	2.11	3.39	2.04
P_2O_5	0.129	0.137	0.179	0.091	0.111	0.142	0.170	0.176	0.140	0.166
(ppm)										
Ni *	70	70	45	50	27	78	93	93	59	89
Cr *	154	164	87	126	114	153	211	262	153	200
Sc *	13	22	17	21	5	16	22	26	0	16
V *	121	143	147	171	84	117	180	240	95	191
Ga *	19	24	35	36	16	26	32	38	18	33
Cu *	65	65	11	8	43	52	47	35	13	14
Zn *	82	93	41	80	66	96	129	153	61	89
U	2.04	1.82	3.07	2.75	1.37	2.78	2.08	2.45	1.24	1.32
Th	7.26	7.19	13.13	11.78	6.12	12.59	9.03	9.71	5.74	6.25
Hf	3.69	3.60	8.46	5.01	3.05	3.55	3.44	4.24	4.02	3.61
Zr *	147	141	324	184	125	132	134	153	155	136
Pb	15.17	9.45	20.19	12.70	13.94	19.37	15.64	11.31	5.81	5.81
Ba	664	830	1458	1362	358	693	862	1052	500	1136
Sr *	280	162	294	202	214	262	238	167	159	142
Rb	39.2	79.7	142.3	132. 5	41. 5	82.1	97.0	125.6	37.5	79.8
Cs	2.0	3.9	6.6	6.2	2.3	4. 5	4.7	5.2	1.4	2.5
Nb	6.2	6. 9	10.2	8.5	5.5	8.4	8.9	11.0	6.3	8.3
Y	16.58	15.46	22.27	22.50	11.17	16.45	24.66	26.50	15.12	14.08
Ta	0.47	0.51	0.94	0.80	0.42	0.82	0.67	0.77	0.45	0.57
La	28.30	7.67	17.15	20.12	8.75	25.10	21.29	26.36	43.29	11.67
Ce	53.25	16.41	35.74	42.53	19.09	53.68	49.40	53.74	78.23	27.26
Pr	5.95	1.94	4.14	4.95	2.08	5.92	5.64	6.01	8.16	3.18
Nd	23.16	7.68	16.66	19.42	7.73	21.99	20.60	23.08	30.52	12.39
Sm	4.64	1.59	3.49	3.68	1.50	4.50	4.24	4.61	5.91	2.61
Eu	1.19	0.61	1.30	1.02	0.79	1.37	0.93	0.96	1.56	1.03
Gd	3.38	1.94	3.69	3.60	1.57	3.22	3.42	3.84	3.85	2.51
Tb	0.56	0.39	0.65	0.61	0.29	0.54	0.71	0.72	0.58	0.46
Dy	3.16	2.66	4.07	3.75	1.93	3.14	4.37	4.67	3.11	2.96
Ho	0.61	0.58	0.81	0.79	0.41	0.63	0.94	0.99	0.59	0.60
Er	1.69	1.82	2.32	2.26	1.23	1.75	2.75	2.80	1.56	1.78
Tm	0.23	0.27	0.34	0.32	0.18	0.23	0.40	0.41	0.21	0.26
Yb	1.51	1.68	2.20	2.21	1.19	1.50	2.59	2.67	1.38	1.74
Lu	0.25	0.27	0.38	0.38	0.18	0.24	0.40	0.44	0.21	0.28

Sample	Y-19 A	Y-19 E	Y-20 A	Y-20 E	Y-21 A	Y-21 E	Y-22 A	Y-22 E	Y-23 A-(L)	Y-23 A-(H)
wt (%)										
SiO_2	66.46	65.88	68.33	69.80	69.14	55.73	70.78	53.70	70.93	70.31
Al_2O_3	17.56	17.96	15.72	14.99	14.82	20.60	14.27	24.33	14.89	15.19
TiO_2	0.658	0.705	0.592	0.560	0.575	0.865	0.600	0.992	0.561	0.571
FeO^*	5.22	5.59	4.49	4.11	5.24	8.42	4.84	8.93	3.90	4.09
MnO	0.050	0.057	0.052	0.048	0.077	0.101	0.053	0.059	0.051	0.038
CaO	1.40	1.21	2.46	2.38	3.06	2.36	3.15	1.82	2.48	1.81
MgO	2.74	3.04	2.18	2.18	2.55	4.93	2.14	3.96	1.95	2.16
K_2O	2.97	3.18	1.68	1.70	1.40	3.20	1.59	2.99	1.54	1.67
Na_2O	2.80	2.24	4.39	4.14	3.02	3.64	2.54	3.01	3.59	4.10
P_2O_5	0.137	0.146	0.107	0.100	0.118	0.158	0.046	0.205	0.105	0.049
(ppm)										
Ni *	84	90	59	54	55	104	53	105	50	59
Cr *	151	163	130	126	126	209	121	233	115	118
Sc *	11	22	11	13	19	22	16	27	15	15
V *	107	144	97	95	103	195	103	197	94	104
Ga *	23	25	18	15	18	23	15	28	16	17
Cu *	71	44	30	29	56	33	106	26	87	32
Zn *	77	87	62	64	76	125	50	74	152	89
U	2.13	1.69	2.15	2.03	2.78	2.50	2.56	2.68	2.02	2.17
Th	6.44	5.68	6.60	6.14	6.65	6.88	8.22	7.90	6.47	6.61
Hf	4.10	2.96	3.46	3.29	3.25	3.31	3.81	4.17	3.38	3.46
Zr *	156	115	136	129	132	123	156	153	136	135
Pb	12.61	11.65	15.91	15.19	13.75	14.01	11.99	12.14	15.73	14.66
Ba	537	640	449	454	308	950	196	736	136	290
Sr *	242	204	389	362	331	254	416	263	195	197
Rb	87.8	88.6	62.4	63. 5	46.9	98.5	53. 7	96.0	36.6	51.7
Cs	7.3	5.8	5.3	5.6	7.4	11.7	10.4	14.0	2.3	3.6
Nb	6.8	7.0	6.0	5.7	6.1	8.1	6.3	9.7	5.9	6.4
Y	15.97	18.52	15.67	14.51	14.97	23.14	15.11	24.20	16.98	15.27
Ta	0.5	0.62	0.67	0.7	0.62	0.63	0.72	0.75	0.72	0.69
La	33.44	30.17	26.19	24.63	26.94	32.87	32.03	29.74	27.56	28.38
Ce	63.47	60.35	49.25	46.19	50.1	65.21	56.79	60.88	50.75	52.05
Pr	7.05	6.87	5.43	5.07	5.41	7.66	6.01	7.08	5.50	5.58
Nd	27.04	26.83	20.48	19.46	20.65	30.82	22.39	28.85	21.31	21.3
Sm	5.33	5.47	4.17	3.94	4.22	6.3	4.47	6.23	4.33	4.22
Eu	1.38	1.32	1.31	1.26	1.18	1.67	1.28	1.58	1.23	1.27
Gd	3.79	3.84	3.20	3.01	3.09	5.00	3.14	4.92	3.27	3.13
Tb	0.57	0.60	0.53	0.48	0.49	0.76	0.51	0.77	0.51	0.49
Dy	3.13	3.48	2.84	2.70	2.79	4.41	2.84	4.51	3.05	2.89
Ho	0.60	0.69	0.57	0.54	0.56	0.86	0.56	0.89	0.62	0.56
Er	1.66	1.94	1.57	1.46	1.59	2.41	1.55	2.47	1.75	1.60
Tm	0.22	0.27	0.22	0.20	0.20	0.34	0.21	0.34	0.24	0.23
Yb	1.44	1.67	1.38	1.26	1.35	2.13	1.39	2.20	1.55	1.40
Lu	0.22	0.26	0.21	0.20	0.21	0.34	0.22	0.36	0.24	0.22

Sample	Y-23 E	Y-24 E	Y-26 A	Y-26 E	Y-27 A	Y-27 E	Y-28 A	Y-28 E
wt (%)								
SiO_2	71.40	60.72	69.35	53.69	69.35	66.81	73.18	56.92
Al_2O_3	14.73	18.10	16.09	24.02	14.81	16.78	14.35	20.32
TiO_2	0.528	0.791	0.585	1.116	0.615	0.666	0.466	0.816
FeO^*	4.24	6.95	4.59	8.87	4.25	5.57	3.52	8.51
MnO	0.057	0.072	0.030	0.076	0.072	0.046	0.038	0.085
CaO	1.83	2.67	1.32	1.00	4.23	1.31	1.69	1.99
MgO	2.13	3.63	2.27	4.75	2.18	2.87	1.94	4.83
K_2O	1.64	3.10	2.35	4.41	1.63	3.06	2.22	3.42
Na_2O	3.38	3.81	3.31	1.87	2.72	2.77	2.52	2.96
P_2O_5	0.062	0.149	0.103	0.196	0.132	0.123	0.089	0.149
(ppm)								
Ni *	50	94	53	90	74	90	42	94
Cr *	102	181	129	265	150	155	101	202
Sc *	13	23	14	30	16	17	14	26
V *	82	160	98	213	109	127	78	189
Ga *	17	23	17	33	16	19	17	29
Cu *	27	4	17	34	90	35	15	57
Zn *	79	114	33	145	72	80	46	119
U	1.92	2.27	1.95	2.35	1.80	1.58	1.46	2.03
Th	5.67	7.30	4.20	7.97	5.67	5.05	4.96	6.44
Hf	2.87	3.75	3.08	4.57	4.05	3.07	2.59	3.24
Zr *	115	145	116	159	157	127	107	123
Pb	10.57	19.27	12.49	12.21	11.24	14.12	11.74	11.64
Ba	268	530	492	839	373	697	568	674
Sr *	180	334	302	184	272	290	231	258
Rb	48.3	93.7	63.7	149.2	56.8	96.4	57.9	99.9
Cs	2.9	4.5	2.4	7.7	5.3	14.4	2.8	4.3
Nb	5.9	7.6	5.9	10.7	6. 5	6.7	4.7	8.3
Y	13.72	18.85	12.13	21.06	15.78	16.98	14.62	21.85
Ta	0.61	0.65	0.57	0.87	0.76	0.64	0.34	0.63
La	25.13	33.06	21.89	27.42	31.72	40.35	21.22	33.44
Ce	46.06	63.81	41.69	55.83	58.69	75.99	40	64.82
Pr	5.01	7.09	4.54	6.68	6.46	8.15	4.41	7.45
Nd	19.19	27.52	17.63	27.52	24.51	30.73	17.09	29.94
Sm	3.94	5.65	3.63	5.91	4.85	5.70	3.68	6.14
Eu	1.20	1.65	1.21	1.29	1.41	1.58	1.16	1.50
Gd	2.79	4.20	2.65	4.46	3.28	3.89	2.97	4.58
Tb	0.45	0.65	0.40	0.68	0.51	0.59	0.46	0.73
Dy	2.59	3.58	2.29	3.95	2.94	3.30	2.66	4.20
Ho	0.51	0.70	0.46	0.78	0.57	0.65	0.53	0.82
Er	1.45	1.94	1.24	2.21	1.61	1.77	1.42	2.35
Tm	0.20	0.27	0.17	0.31	0.23	0.25	0.18	0.31
Yb	1.28	1.69	1.09	2.05	1.42	1.54	1.16	2.02
Lu	0.20	0.26	0.18	0.33	0.22	0.23	0.18	0.33

Table 3.2. Results of the Sm-Nd isotopic analyses.

Sample	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	εNd(2.66)	T DM (Ga)
(Volcanic rocks)						
Y-2A	2.92	9.60	0.1839	0.512485 (10)	1.4	-
Y-2B	2.97	9.41	0.1906	0.512617 (11)	1.7	-
Y-3	2.09	7.90	0.1601	0.511910 (8)	-1.7	-
Y-4	2.13	8.80	0.1468	0.511780 (8)	0.4	-
Y-5	4.37	28.55	0.0926	0.510823 (11)	0.2	2.9
Y-6	2.12	9.03	0.1425	0.511464 (8)	-4.4	-
Y-7	2.73	9.37	0.1762	0.512293 (8)	0.3	-
Y-8	1.18	3.22	0.2218	0.513116 (11)	0.7	-
Y-9	1.19	3.95	0.1820	0.512394 (8)	0.3	-
(Greenschist facies turbidites)						
Y-10A1	4.80	27.46	0.1057	0.511116 (8)	1.5	2.9
Y-10A2	5.01	27.80	0.1088	0.511136 (15)	0.8	2.9
1-10C	5.24	29.25	0.1082	0.511114 (8)	0.6	2.9
Y-10E	5.23	27.22	0.1161	0.511255 (12)	0.6	3.0
Y-11A	4.92	28.57	0.1042	0.511086 (8)	1.4	2.9
Y-12A	4.23	23.40	0.1094	0.511152 (11)	0.9	2.9
Y-12D	4.79	25.53	0.1135	0.511243 (15)	1.3	2.9
Y-12E	6.28	32.76	0.1160	0.511222 (14)	0.0	3.0
Y-13A	3.47	19.02	0.1103	0.511167 (11)	0.9	2.9
Y-13E	4.75	24.73	0.1161	0.511246 (8)	0.5	3.0
Y-14	4.48	25.10	0.1080	0.511132 (7)	1.0	2.9
Y-15A	1.77	8.26	0.1292	0.511363 (12)	-1.8	3.2
Y-16D	3.73	17.80	0.1268	0.511373 (9)	-0.7	3.1
Y-16E	4.06	20.53	0.1194	0.511283 (8)	0.0	3.0
Y-17A	1.53	8.31	0.1114	0.511250 (8)	2.1	2.8
Y-17C	4.29	23.83	0.1089	0.511142 (11)	0.9	2.9
Y-17D	4.26	21.94	0.1173	0.511213 (11)	-0.6	3.1
Y-17E	4.56	24.25	0.1137	0.511218 (12)	0.7	2.9
Y-18A	5.45	32.48	0.1014	0.510981 (8)	0.3	2.9
Y-18E	2.71	12.91	0.1268	0.511346 (10)	-1.3	3.2
(Amphibolite facies turbidites)						
Y-19A	4.94	28.40	0.1052	0.511189 (13)	3.1	2.8
Y-19E	5.13	28.71	0.1080	0.511122 (10)	0.8	2.9
Y-20A	4.18	22.83	0.1107	0.511159 (10)	0.6	2.9
Y-20E	3.93	21.41	0.1109	0.511145 (9)	0.2	3.0
Y-21A	4.01	22.56	0.1076	0.511133 (12)	1.2	2.9
Y-21E	6.19	32.73	0.1143	0.511262 (10)	1.4	2.9
Y-22A	4.22	24.76	0.1031	0.511039 (9)	0.9	2.9
Y-22E	5.79	29.97	0.1168	0.511294 (11)	1.1	2.9
Y-23A (L)	4.14	22.96	0.1090	0.511135 (8)	0.7	2.9
Y-23A (U)	4.05	22.95	0.1068	0.511109 (8)	1.0	2.9
Y-23E	3.74	21.05	0.1075	0.511121 (8)	0.9	2.9
Y-24E	5.42	30.22	0.1083	0.511131 (8)	0.9	2.9
Y-26A	3.69	20.34	0.1096	0.511149 (10)	0.8	2.9
Y-26E	5.65	29.23	0.1168	0.511214 (16)	-0.4	3.0
Y-27A	4.49	26.16	0.1037	0.511105 (9)	1.9	2.8
Y-27E	5.50	33.87	0.0981	0.510971 (8)	1.2	2.9
Y-28A	3.66	18.81	0.1176	0.511277 (10)	0.5	3.0
Y-28E	5.92	31.89	0.1123	0.511210 (15)	1.1	2.9

Notes: [a] Normalized to $^{146}Nd/^{144}Nd$=0.7219. Numbers in (paretheses) represent errors in 2σ.; [b] TDM calcutaled using the mantle evolution model of Goldstein et al. (1984).; Present day CHUR parameters are $^{147}Sm/^{144}Nd$=0.1967, $^{143}Nd/^{144}Nd$=0.512638.; $\lambda_{147Sm}=6.54\times10^{-12}$

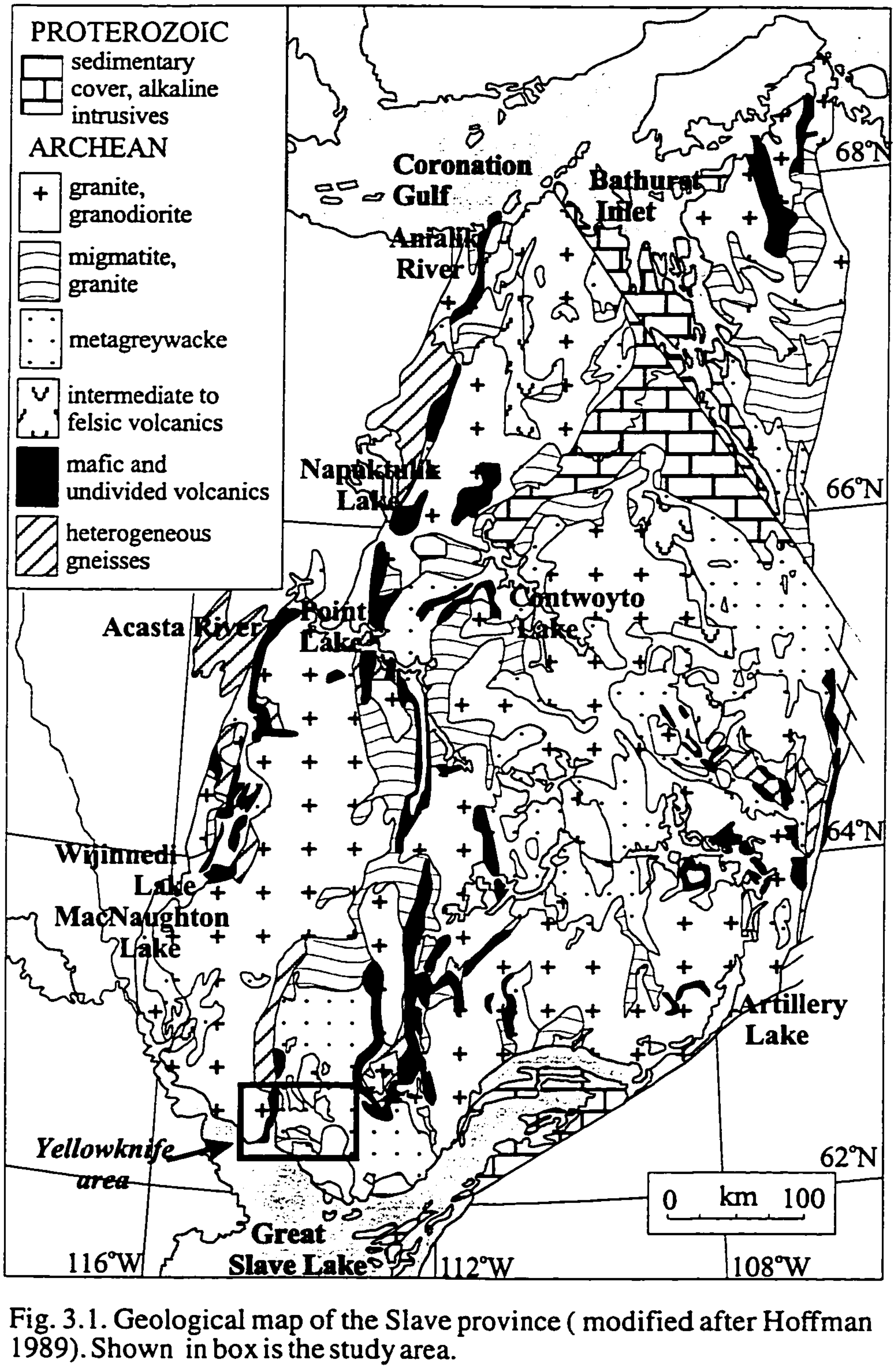


Fig. 3.1. Geological map of the Slave province (modified after Hoffman 1989). Shown in box is the study area.

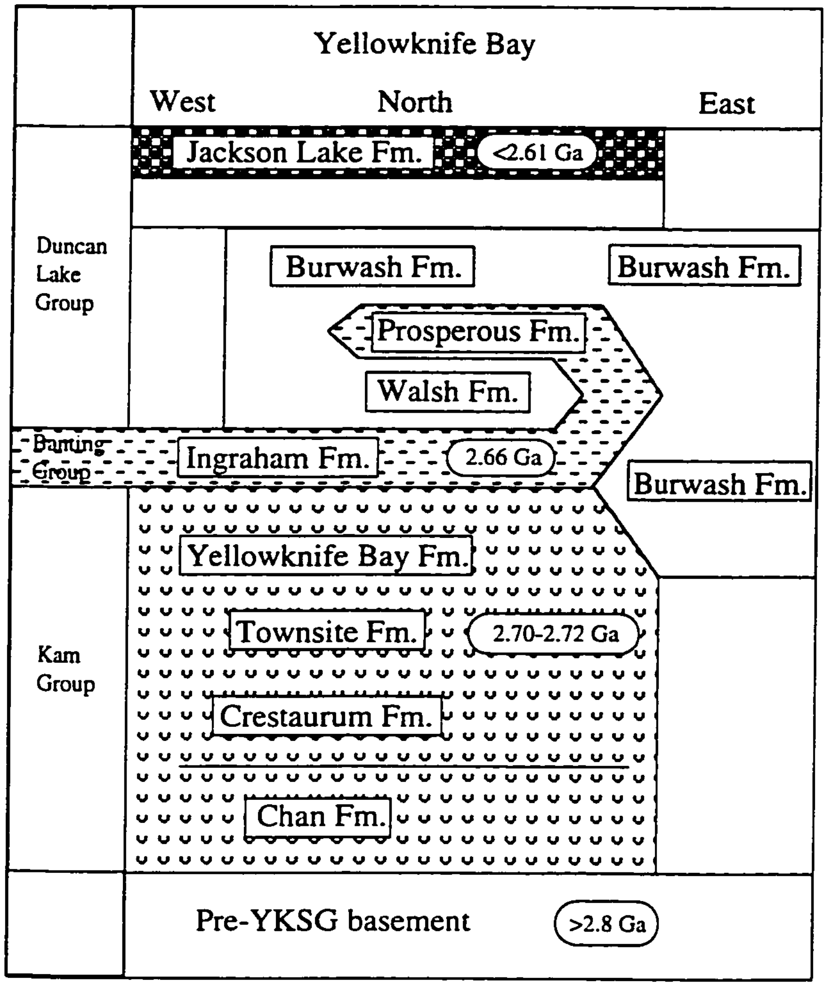


Fig. 3.2.Stratigraphic units of Yellowknife Supergroup in the vicinity of Yellowknife (after Helmstaedt and Padgham 1986). Geochronological data are from Isachsen et al (1991) and Isachsen and Bowring (1997).

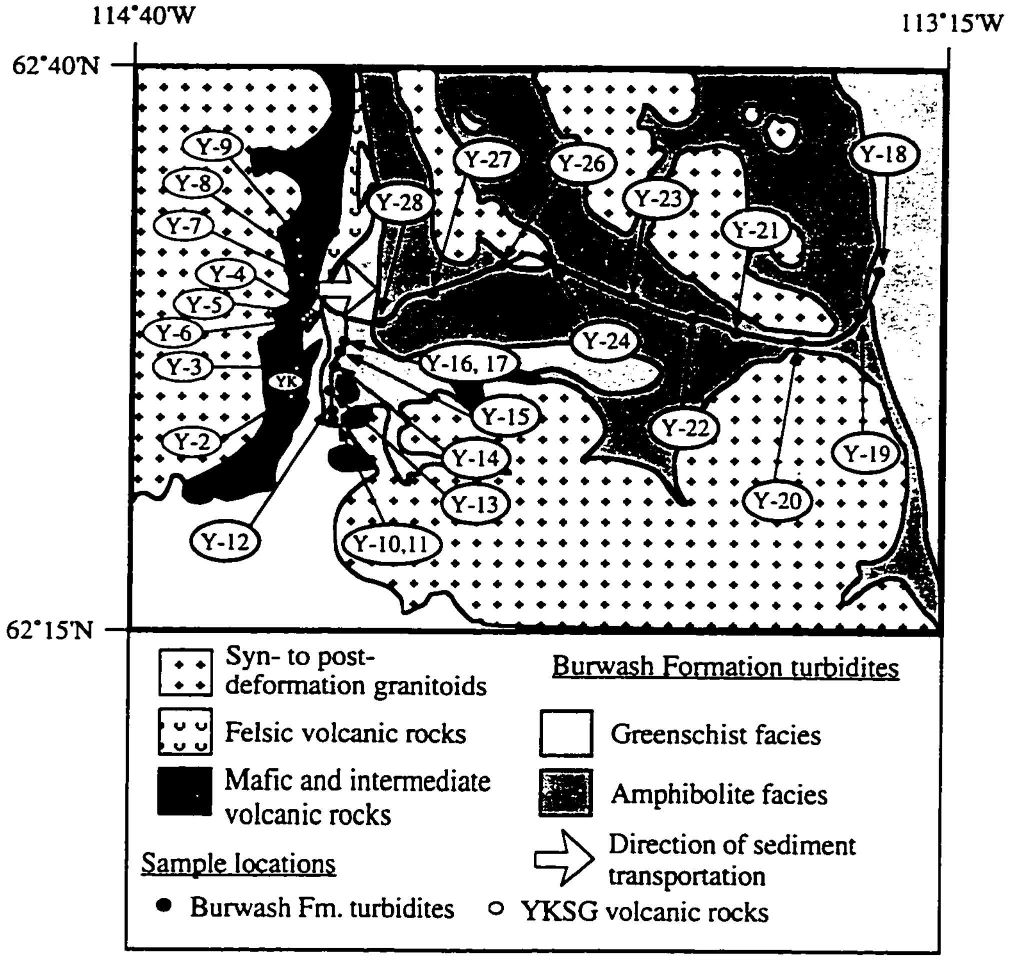


Fig. 3.3. Geological map of Yellowknife area with sample locations (modified after Henderson 1985). Also shown is the direction of sediment transportation (Henderson 1975).

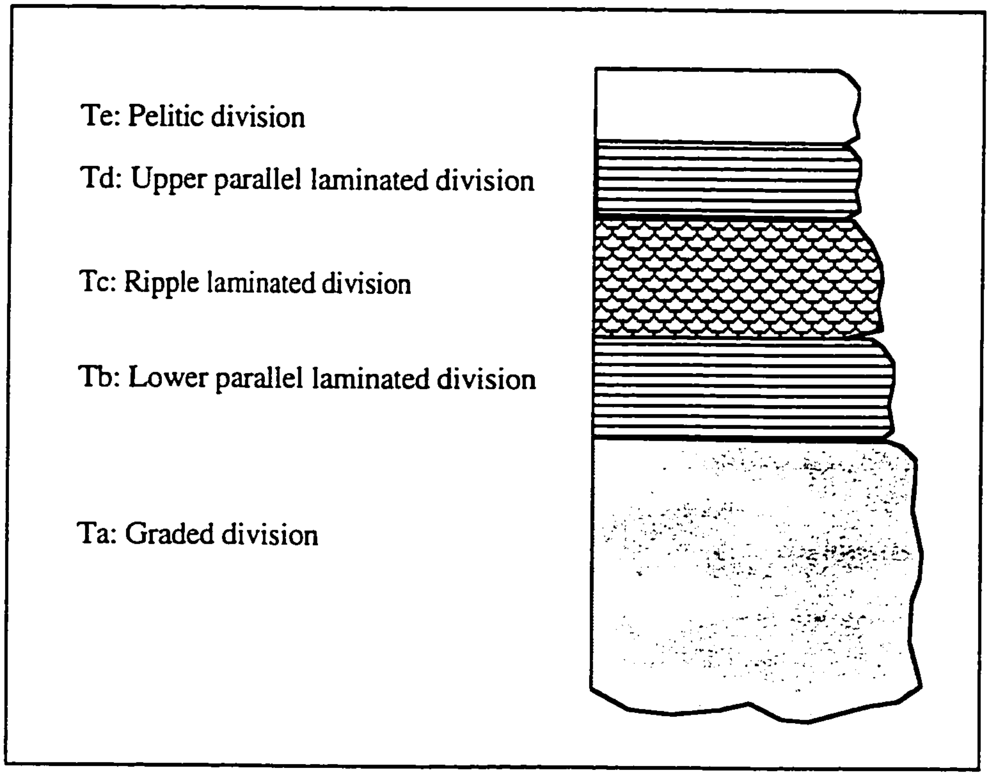


Fig. 3.4. An example of complete Bouma sequence (after Bouma 1962).

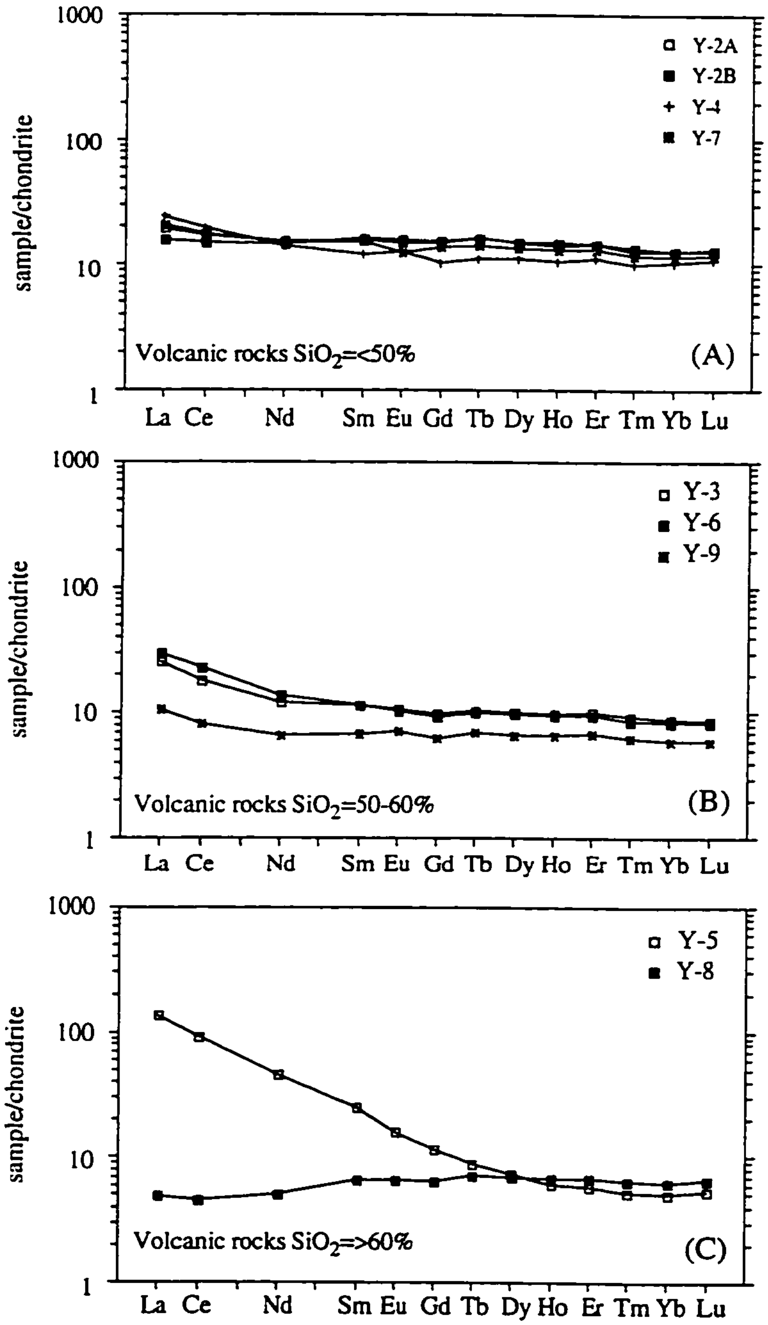


Fig. 3.5. Chondrite normalized rare earth element diagrams for YKSG volcanic rocks from the Yellowknife area. (A) samples with SiO2<50%, (B) samples with SiO2=50-60%, (C) samples with SiO2>60%. Values for the chondrite are taken from Boynton (1984).

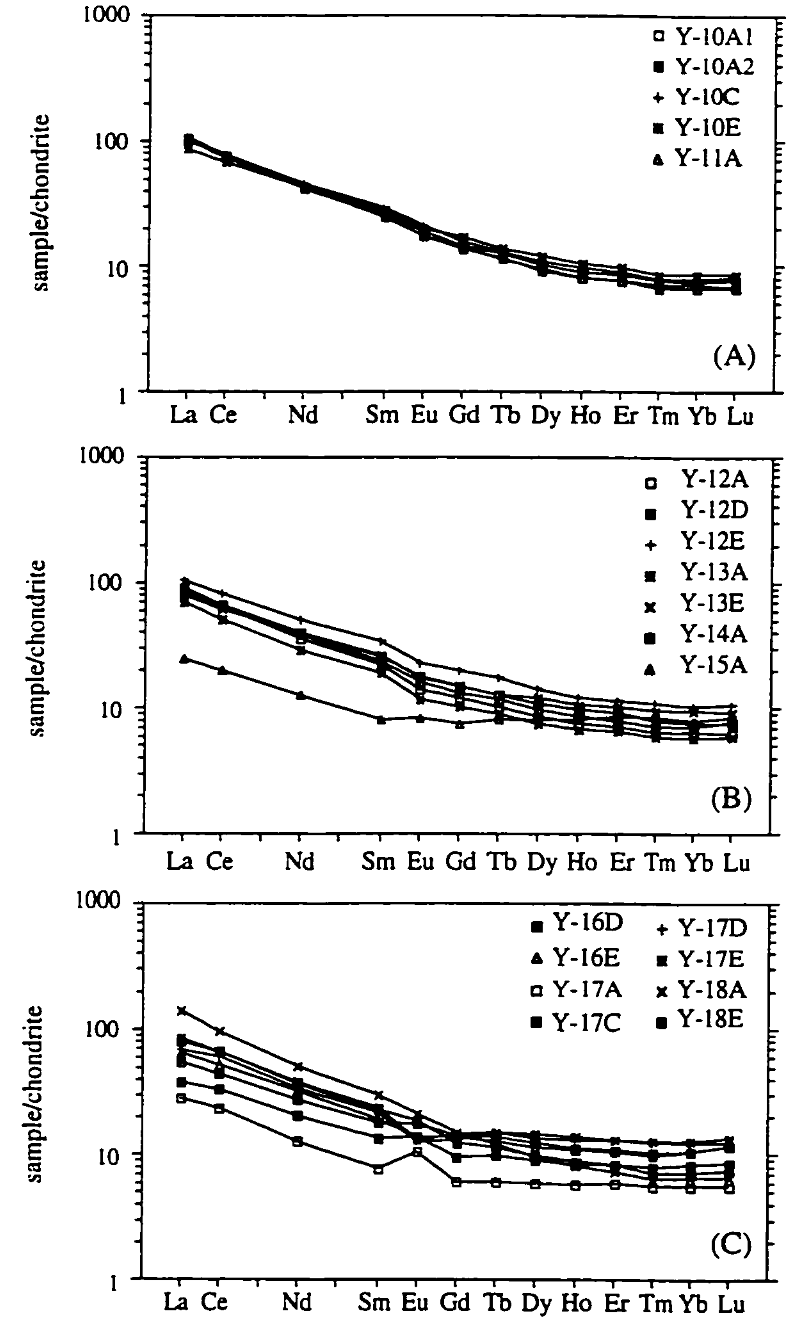


Fig. 3.6. Chondrite normalized rare earth element diagrams for the Burwash Fm. turbidites (greenschist facies). Values for the chondrite are taken from Boynton (1984).

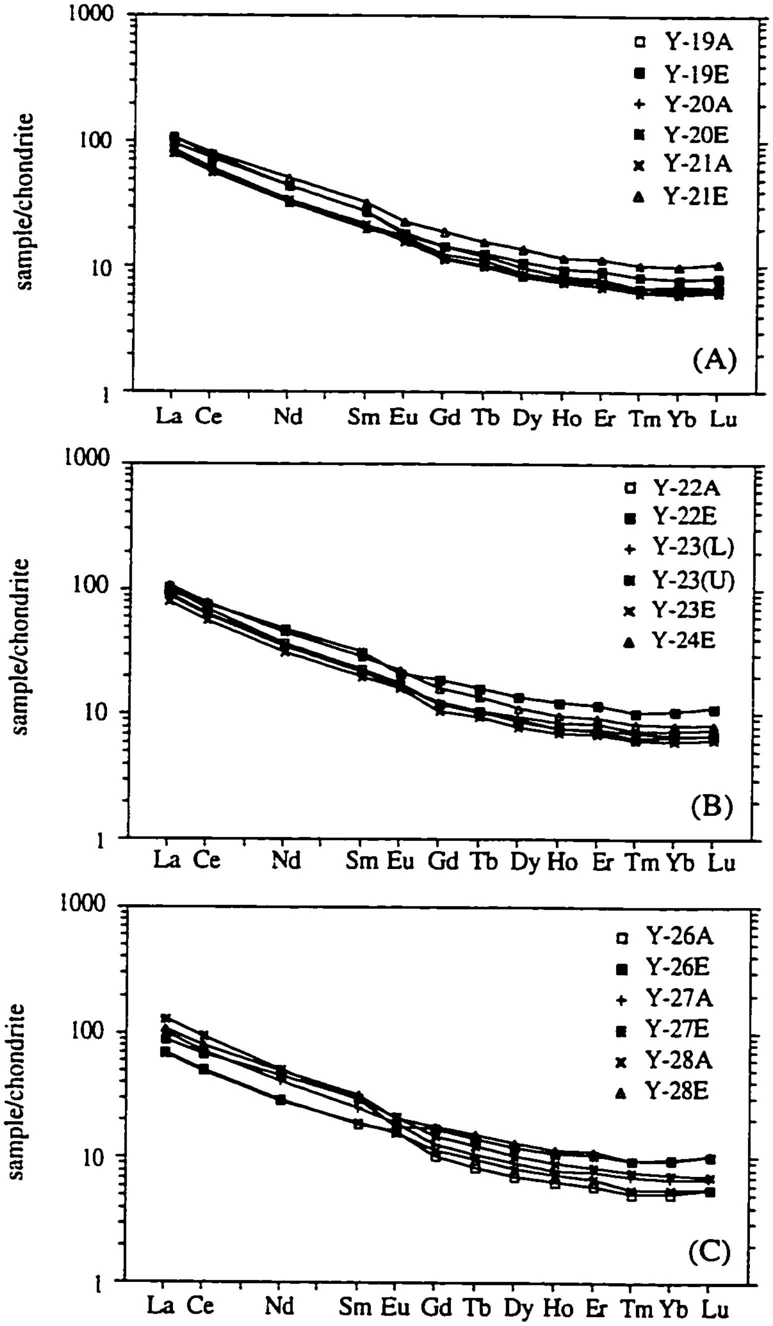


Fig. 3.7. Chondrite normalized rare earth element diagrams for the Burwash Fm. turbidites (amphibolite facies). Values for the chondrite are taken from Boynton (1984).

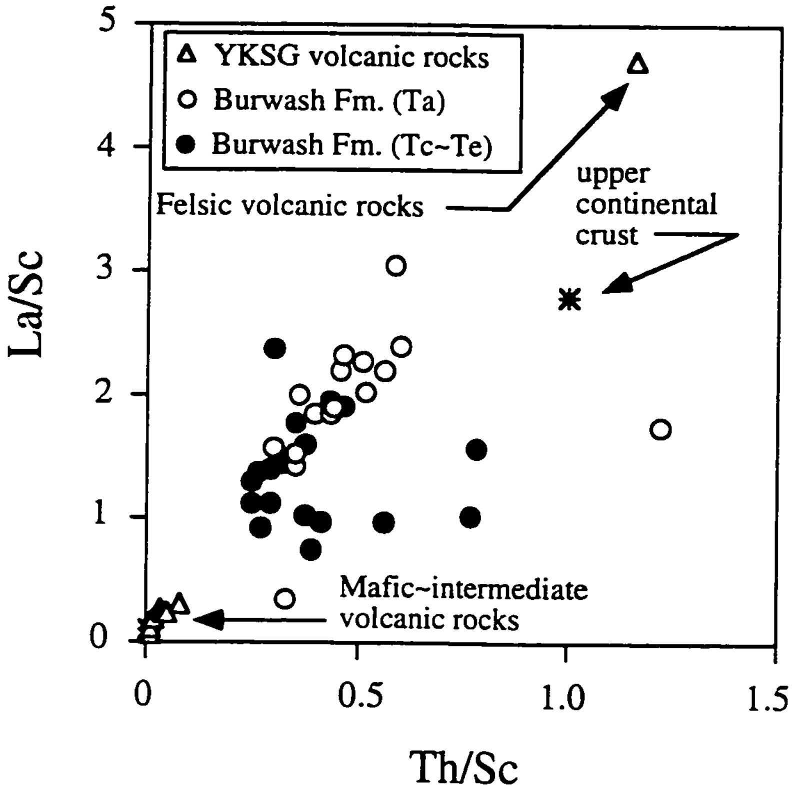


Fig. 3.8. Th/Sc versus La/Sc diagram for YKSG volcanic rocks and Burwash Fm. turbidites. Also shown is the value for the average upper continental crust of Taylor and McLennan. (1985).

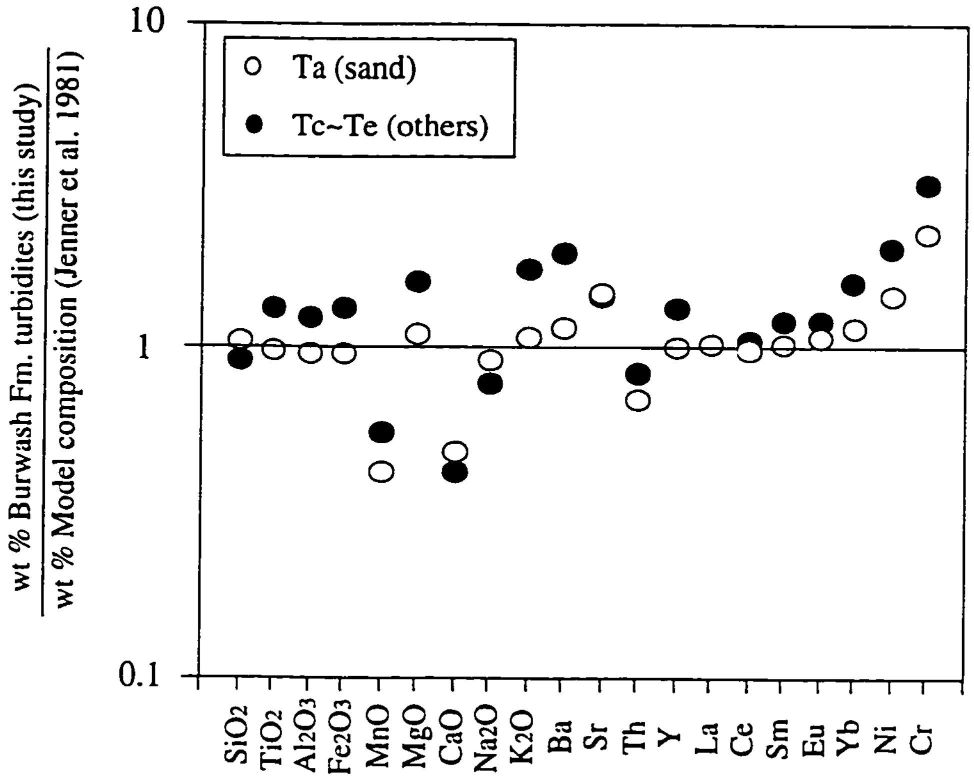


Fig. 3.9. Major and trace element abundance of the Burwash Fm. turbidites (this study) normalized to the model chemical composition of Burwash Fm. turbidites by Jenner et al. (1981). Model chemical composition is represented by 20% mafic-intermediate volcanic rocks + 55% felsic volcanic rocks + 25% granitic rocks. The average values for the sand units (Ta) and mud units (Tc-Te) are shown in open and solid circles, respectively.

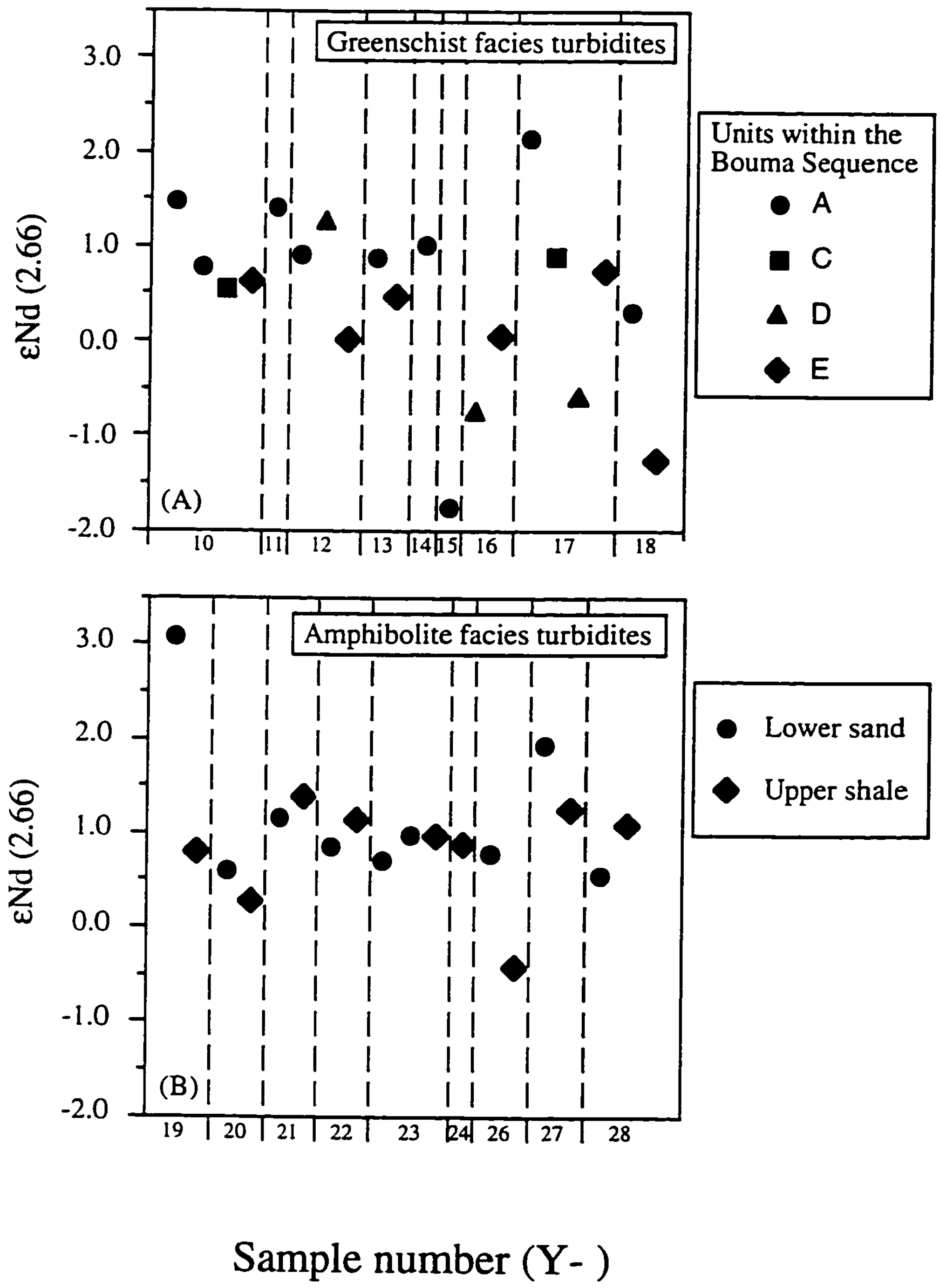


Fig. 3.10. εNdT values of the Burwash Fm. turbidites. Different units of the Bouma sequence are plotted separately.

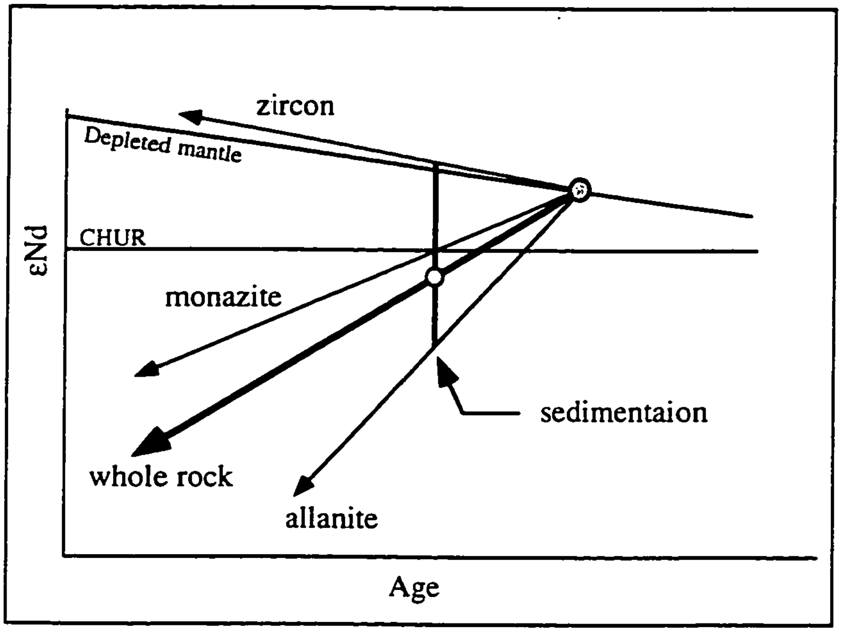


Fig. 3.11. Schematic diagram illustrating the Nd isotopic evolution of important REE hosting minerals relative to whole rock sample. Garnet is not included in this diagram because the concentration of light rare earth elements is exceptionally low (see Bea 1996) and accumulation of garnet will have minimum or no effect on the εNdT values of whole rock samples.

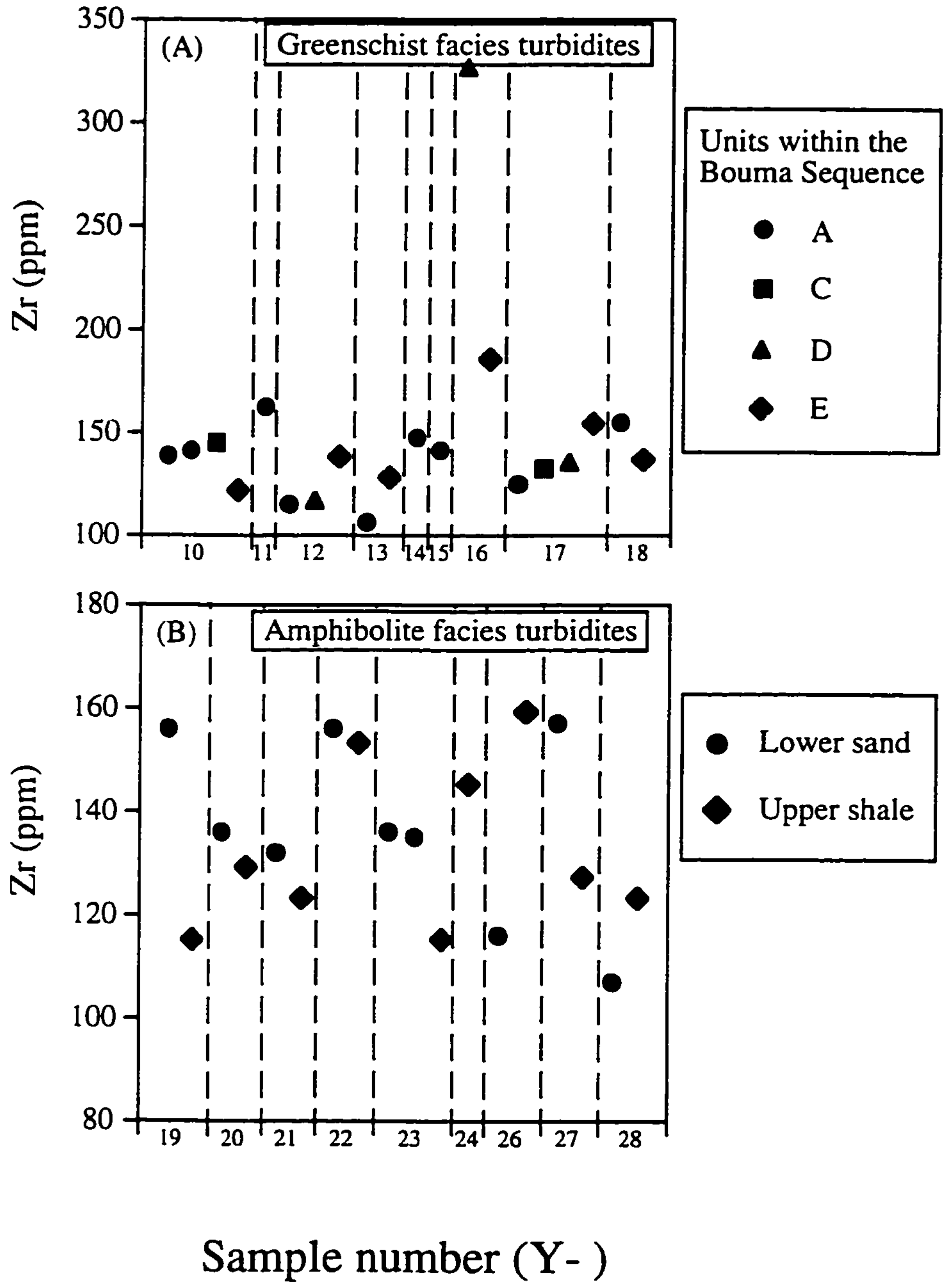


Fig. 3.12. Diagram displaying Zr (ppm) of the Burwash Fm. turbidites.

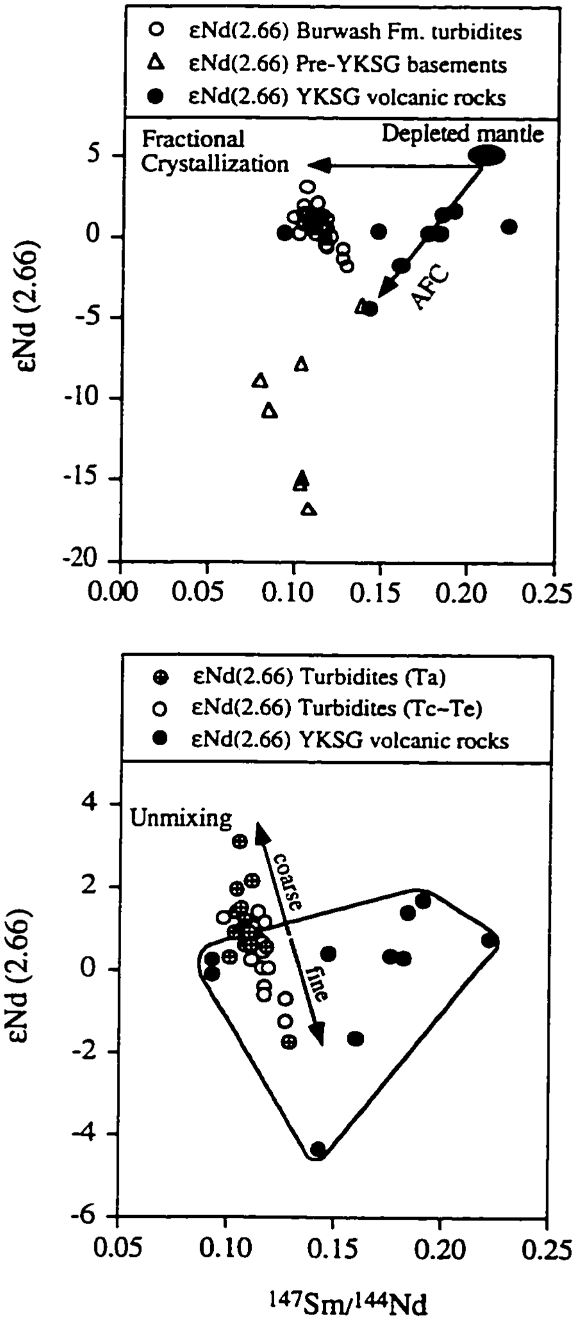


Fig. 3.13. $^{147}Sm/^{144}Nd$ versus εNd(2.66) diagram of Pre-YKSG basement, YKSG volcanic rocks and Burwash Fm. turbidites. (A) The isotopic signature of YKSG volcanic rocks can be explained by assimilation and fractional crystallization (AFC) processes of mantle derived materials. (B) The isotopic signature of the Burwash Fm. turbidites is a result of unmixing of detritus derived from crustally contaminated YKSG volcanic rocks and juvenile (or less contaminated) quartzo-feldspathic components (see text for detail).

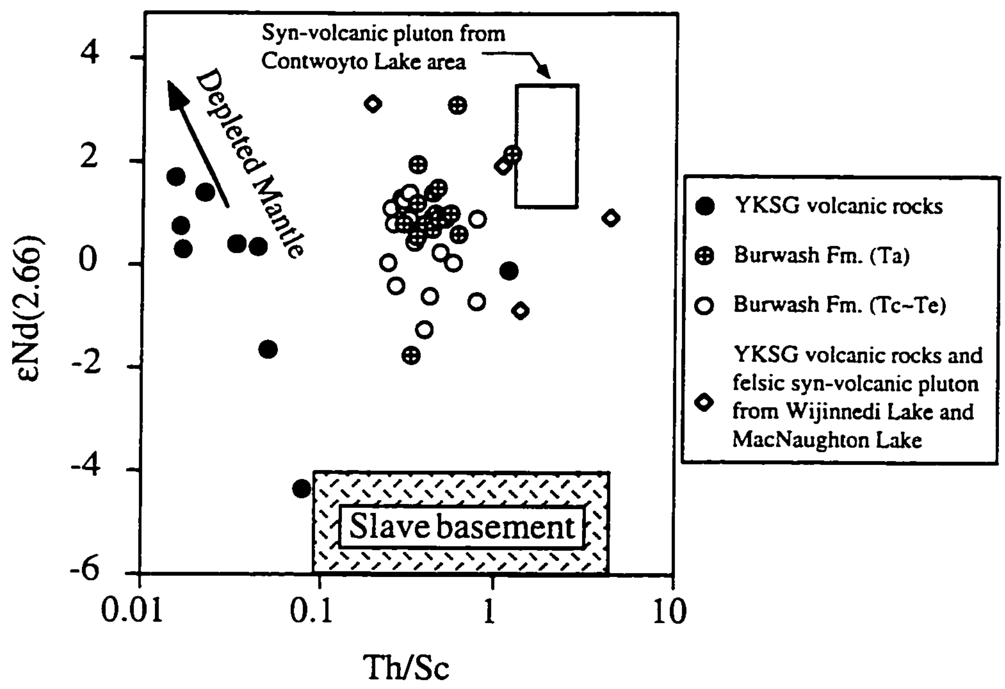
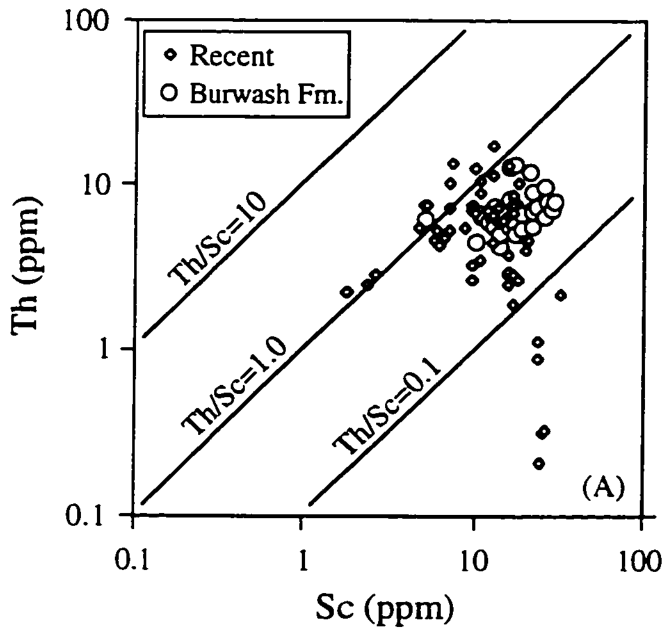


Fig. 3.14. Th/Sc versus εNd(2.66) diagram of Pre-YKSG basement, YKSG volcanic rocks and Burwash Fm. turbidites. Data for the Burwash Fm. turbidites plot in the region bordered by crustally contaminated YKSG volcanic rocks and less contaminated syn-volcanic pluton (and volcanic rocks) from the western Slave province. Data for syn-volcanic pluton from Contwoyto Lake area were taken from Davis et al. (1994).



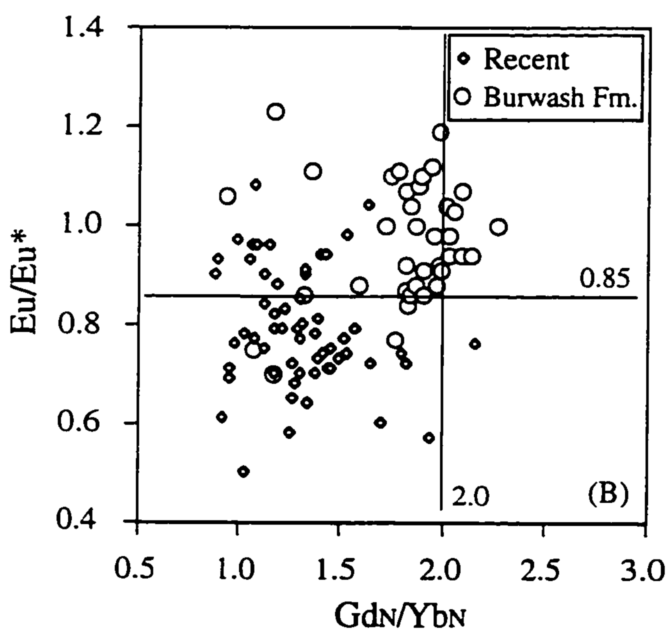


Fig. 3.15. Chemical composition of Burwash Fm. turbidites compared to modern turbidites. (A) Sc (ppm) versus Th (ppm) and (B) Gd_N/Yb_N versus Eu/Eu*. Data for the modern turbidites were taken from McLennan et al. (1990).

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<u>Appendix 3.1. (Sample locations and notes)</u>

Yellowknife volcanics

(Y-2) 62°26.89'N, 114°23.27'W

Kam group (Yellowknife Bay Formation) - A: coarse grained, B: fine grained basalt.

(Y-3) 62°28.26'N, 114°22.46'W

Kam group (Crestaurum Formation) - Fine grained basalt.

(Y-4) 62°31.32'N, 114°19.41'W

Banting group (Ingraham Formation) - Amphibolite.

(Y-5) 62°31.26'N, 114°20.94'W

Kam group (Yellowknife Bay Formation) - Andesite~dacite.

(Y-6) 62°30.76'N, 114°21.35'W

Kam group (Yellowknife Bay Formation) - Massive basalt.

(Y-7) 62°33.03'N, 114°21.94'W

Kam group (Crestaurum Formation) - Partly sheared basalt.

(Y-8) 62°33.92'N, 114°21.62'W

Cherty tuff in Kam group (interbedded with basalt and gabbros).

(Y-9) 62°34.58'N, 114°22.00'W

Kam group (Yellowknife Bay Formation) - Basaltic~andesitic volcanic rock.

Yellowknife turbidites (Greenschist facies)

(Y-10, 11) 62°25.16'N, 114°18.40'W

Y-10 A	Base of Ta
Y-10 B	Top of Ta
Y-10 C	Rippled (Tc)
Y-10 E	Silt (Te)
Y-11A	Sand (Ta unit just above Y-10E)

* 1 km north of Delta, just 20 m west of the road. Several turbidite sequences observed. Tb and Td missing. ~80 cm thick.

(Y-12) 62°26.23'N, 114°18.40'W

Y-12 A	Base (Ta)

Y-12 D Upper laminated (Td)

Y-12 E Upper silt (Te)

* Total sequence ~1 m thick. Strike 060. Bedding overturned.

(Y-13) 62°27.12'N, 114°18.55'W

Y-13 A Absolute bottom of Ta

Y-13 E Absolute top of sequence (Te)

* No Tb, Tc, Td observed. ~1 m thick sequence. Bedding overturned. Strike 050, dip 70S.

(Y-14) 62°28.35'N, 114°17.95'W

Y-14 Massive sand unit.

* ~4 m thick sand unit. No internal structure observed. Strike 010, dip 90.

(Y-15) 62°29.21'N, 114°17.51'W

Y-15 A Ta (?)

* Internal sequence not clear. Sample is "probably" base (i.e. Ta) unit.

(Y-16, 17) 62°29.38'N, 114°17.09'W

Y-16 D Td

Y-16 E Te

Y-17 A Ta (Unit just above Y-16E)

Y-17 C Tc

Y-17 D Td

Y-17 E Te

* Y-16 D~E is ~10 cm thick. Y-17 A~E is ~30 cm thick. Y-16 to Y-17 is a continuous sequence.

(Y-18) 62°32.94'N, 113°21.14'W

Y-18 A Basement sand (Ta)

Y-18 E Silt (Te), shows strong cleavage.

* End of the Ingraham road (at Tibbet lake). Turbidite sequence still visible. Bedding overturned. Strike 010, dip 70E.

Yellowknife turbidites (Amphibolite facies) - Different units of Bouma sequence cannot be sampled separately. Only "sand" and "shale" units from a single turbidite sequence were sampled.

(Y-19) 62°30.24'N, 113°23.86'W

Y-19 A Sand

Y-19 E Shale

* Big cordierites in the shale unit but none in the sand unit. Not clear which way is the top.

(Y-20) 62°29.40'N, 113°29.98'W

Y-20 A Sand?

Y-20 E Shale?

* Internal structure not preserved. Y-20 A and Y-20 E may not be from the same turbidite sequence.

(Y-21) 62°30.14'N, 113°36.81'W

Y-21 A Sand?

Y-21 E Silt?

* Bedding and other internal structure almost lost.

(Y-22) 62°30.85'N, 113°42.95'W

Y-22 A Sand?

Y-22 E Shale?

* Local pegmatites found throughout the outcrop. In places, cordierites could be ~6 cm long. Internal structure not preserved. Y-22 A and Y-22 E may not be from the same turbidite sequence.

(Y-23) 62°31.30'N, 113°48.16'W

Y-23 A upper Upper part of the sand unit

Y-23 A lower Lower part of the sand unit

Y-23 E Shale

* Bedding observed, but unable to distingush the different units of Bouma sequence. Strike 105, dip 90.

(Y-24) 62°33.43'N, 113°56.37'W

Y-24 E Shale

* Bedding observed but unable to sample the sand unit. . Strike 130, dip 60NE.

(Y-26) 62°33.27'N, 114°01.97'W

Y-26 A Sand

Y-26 E Shale

* Coarse beds, 80~100 cm thick. Y-26 A and E from same turbidite sequence. Strike 95, dip 60N.

(Y-27) 62°32.23'N, 114°08.66'W

Y-27 A Sand

Y-27 E Shale

* Clear bedding but other internal structures not observed. Cordierites may be 5~6 cm long.

(**Y-28**) 62°30.11'N, 114°14.94'W

Y-28 A Sand

Y-28 E Shale

* Internal structure well preserved but only sand and shale sampled. Bedding overturned. . Strike 130, dip 70S.

Chapter 4

Geochemical and Nd-Pb isotopic systematics of late-Archean granitoids, southwestern Slave Province, Canada: Constraints for granitoid origin and crustal isotopic structure. *

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Introduction

Considerable effort has been made over the past two and a half decades to constrain the tectonic evolution of the Slave province, an Archean craton in the northwestern Canadian shield. Some of the models proposed to date include intracontinental rifts ± closure of the basin and subsequent eastward subduction of oceanic lithosphere (Henderson 1981, Fyson and Helmstaedt 1988, MacLachlan and Helmstaedt 1995) and island arc-continent collision (Kusky 1989). These models were constructed mainly on the basis of field study, and incorporation of geochemical constraints, particularly radiogenic isotopic constraints, is still in an early stage. Of particular importance in all tectonic models is the extent of early- to mid-Archean crust (pre-2.8 Ga) and it's role in the crustal evolution of the Slave province during 2.7 to 2.6 Ga crust formation and consolidation. Recent Pb and Nd isotopic studies of 2.7 Ga volcanogenic massive sulfide deposits (VMS) and late Archean granitoids have suggested that pre-2.7 Ga basement is restricted to the western Slave province (Thorpe et al. 1992, Davis and Hegner 1992, Davis et al. 1996). However, whether this basement is a single crustal block that lies beneath the entire western Slave or whether it is restricted to certain areas of the western Slave province is still unclear. In this regard, Davis et al. (1996) proposed, in addition to the major north-south trending isotopic boundary (Davis and Hegner 1992), two east-west trending isotopic boundaries in the southwestern Slave province on the basis of limited Pb isotopic data in this area.

This study presents new major/trace element geochemical and Nd, Pb isotopic data of late Archean granitoids from the southwestern Slave province in order to better constrain the source of these granitoids and to evaluate the role of pre-2.8 Ga basement in the crustal evolution of the southwestern Slave province. Due to high Pb concentration of the crust relative to the mantle, Pb isotopes are particularly useful in tracing the "ultimate age" or "mantle extraction age" of the basement which may have contributed as a source of the granitoids. Combined Nd and Pb isotopes, on the other hand, can help determine the proportion of crustal recycling relative to the addition of juvenile crust. These results will then be combined with major/trace element geochemistry to speculate on the genesis of these granitoids as well as the mechanism by which ancient crustal signatures might be incorporated into these rocks.

Geology of the Slave province

Regional geology

Rocks of the Slave province can be classified into three lithotectonic groups; (1) pre-Yellowknife Supergroup basement, (2) supracrustal rocks of the Yellowknife Supergroup and syn-volcanic plutons and (3) syn- to post-deformation granitoids.

The oldest group comprise ~2.84 to 4.02 Ga granitoids, gneisses and supracrustal rocks which were formed prior to the deposition of Yellowknife Supergroup supracrustal rocks (Henderson 1985, Henderson et al. 1982, 1987, Fyson and Helmstaedt 1988, Hoffman 1989, Bowring et al. 1989, Isachsen and Bowring 1994, Villeneuve and van Breemen 1994, Bleeker and Stern 1997). These rocks are found from northern to southern Slave province, but recent isotopic studies indicate that the occurrence of these rocks are restricted to the west of north-south trending isotopic boundary at approximately 111° W (Thorpe et al. 1992, Davis and Hegner 1992, Davis et al. 1996).

Rocks of the Yellowknife Supergroup (YKSG) consist mainly of 2.65 to 2.72 Ga mafic to felsic volcanic and greywacke-mudstone sequences. These supracrustal rocks differ from those of many other Archean cratons in that metasedimentary rocks make up approximately 80% of the supracrustal package (Padgham and Fyson 1992, Isachsen and Bowring 1994). In addition to the 2.65 to 2.72 Ga metasedimentary rock, ~ 2.59 to 2.62 Ga sandstones and conglomerates have also been reported (Fyson and Helmstaedt 1986, Isachsen and Bowring 1994).

Plutonic rocks in the Slave province range in age from 2.58 to 2.72 Ga with an apparent "magmatic gap" between 2620-2645 Ma (van Breemen et al. 1992, Villeneuve and van Breemen 1994, Villeneuve et al. 1997). 2.72 to 2.65 Ga syn-volcanic plutons are generally intermediate in composition (Davis et al. 1994), but felsic granite as old as 2.68 Ga has also been reported (Yamashita et al. 1995, 1997, Yamashita and Creaser 1996).

The 2.62 to 2.58 Ga granitoids can be divided into 2.62-2.60 Ga syn- to late-deformation granitoids (referred to as syn-deformation granitoids hereafter) and 2.60-2.58 Ga post-deformation granitoids (Davis et al. 1994). Included in the syn-deformation granitoids are granitoids of Defeat plutonic suite (Yellowknife area), Concession and Siege suites (central Slave province) and ~2.61 Ga granitoids from the Wijinnedi Lake area (Henderson 1985, Davis et al. 1994, Villeneuve and Henderson 1998). The post-deformation granitoids include granitoids of Stagg, Awry and Prosperous suites (Yellowknife), Contwoyto and Yamba suites (central Slave province) and ~2.59 Ga granitoids from the Wijinnedi Lake and MacNaughton Lake areas (Henderson 1985, Davis et al. 1994, Perks 1997, Villeneuve and Henderson 1998). Composition of these granitoids generally changes from metaluminous biotite + hornblende granodiorite to tonalite for the syn-deformation granitoids, to peraluminous biotite ± muscovite granite for the post-deformation granitoids (Davis et al 1994), although some of the post-deformation granitoids from the Wijinnedi Lake are metaluminous to weakly peraluminous orthopyroxene bearing granitoids.

Regional deformation in the Slave Province took place between 2.63 to 2.58 Ga (van Breemen et al. 1992, Relf 1992). The peak of regional compressional deformation and metamorphism took place between 2.61 to 2.60 Ga in the central and northeastern Slave province but may have taken place slightly earlier in the southwestern Slave province (van Breemen et al. 1992). The deformation continued throughout retrograde conditions until 2.58 Ga (van Breemen et al. 1992). The metamorphic grade ranges from greenschist-amphibolite-granulite facies (Thompson 1978, Henderson and Schaan 1993, Henderson and Chacko 1995, Pehrsson and Chacko 1997).

Geochronological framework of southwestern Slave province

The geochronology of supracrustal rocks and plutonic rocks from the Yellowknife area are well established. The oldest unit is the 2.92 to > 3.41 Ga granitoid and mylonite gneisses of the Anton complex, which are overlain by <2.92 Ga quartzite (some detrital zircon are as old as >3.7 Ga) and ~2823 Ma felsic volcaniclastic rocks and banded iron formation of the Bell Lake Group (Isachsen and Bowring 1997).

The volcanic rocks of the Yellowknife Supergroup comprise the >2720 to 2701 Ma Kam Group and 2658 to 2663 Ma Banting Group (Isachsen et al. 1991, Isachsen and Bowring 1994, 1997). These volcanic rocks are surrounded to the east by sedimentary rocks of the Burwash Formation (~2.66 Ga) and to the west by syn- to post-deformation granitoids of the western plutonic complex (Henderson 1985). Although part of the Anton complex mentioned above is clearly pre-Yellowknife Supergroup in age, there are portions of the Anton complex that are 2641 +5/-4 Ma, as reported by Dudäs et al. (1990). Several ages for the syn- to post-deformation granitoids have been reported by Henderson et al. (1987) and van Breemen (1992). The ages of two samples of the Defeat suite are determined at 2621 +5/-8 Ma and 2620 Ma ± 8 Ma and the age of a granite from the Stagg suite is 2588 ± 7 Ma (Henderson 1987, van Breemen 1992). The precise U-Pb age of the Awry suite granitoids has not been reported. However, field relationship between granitoids of Stagg and Awry suite suggests that the Awry suite is younger than the Stagg suite (Henderson 1985). The youngest supracrustal rocks in the area are the <2605 Ma conglomerate and the fluvial sediments of Jackson Lake formation (Isachsen and Bowring 1994).

The U-Pb ages of samples from Wijinnedi Lake area used in this study are summarized in Villeneuve and Henderson (1998). The U-Pb ages of two syn-deformation granitoids (J-0391, J-0398) were determined to be 2605 ± 3 Ma and ~2605 Ma and the ages of three post-deformation granitoids (HBA-0505-93, J-0397, J-0400) are 2589 ± 1, 2593 +6/-4 and 2598 ± 2 Ma (Table 4.1; Villeneuve and Henderson 1998).

From the MacNaughton Lake area, none of the four samples used here have been U-Pb dated. However, a sample located < 100m from sample FL24A from the same mapped lithologic unit as FL24A and FL74A is U-Pb zircon dated at 2589 +3/-2 Ma (Perks 1997). Two other granitoids (ML20A, ML21A) are also correlated with the ~2590 Ma post-deformation magmatism on the basis of field relationships (T. Chacko, pers. comm. 1997).

Analytical procedure

Samples for this study were collected from the Yellowknife, Wijinnedi Lake and MacNaughton Lake areas of the southwestern Slave province (Fig. 4.1). The locations of these samples are given in Table 4.1.

Major- and trace-element analyses were performed at Washington State University using XRF and ICP-MS. The analytical methods are described in Hooper et al. (1993). K-feldspar and plagioclase for the common Pb analyses were separated by standard gravity and magnetic methods and hand picked to remove impurities. Samples were ultrasonicated in distilled acetone for ~10 minutes and rinsed with millipore water prior to leaching overnight in 2N HCl (L1), 6N HCl (L2), 16N HNO_3 (L3) and 16N HNO3 plus one drop of 48% HF (L4). Final dissolution was in 4:1 HF : HNO_3 (Cumming and Krstic 1987). For some samples from the Yellowknife area, residues (R) after L4 were further leached three times in 7 N HNO_3 - 5% HF mixture for 20 minutes (Housh and Bowring 1991) before final dissolution. Pb was separated using combined HBr and HBr-HNO_3 media chemistry, modified from Lugmair and Galer (1992). The blank for the entire chemical procedure was approximately 150 pg. Since the typical amount of Pb loaded onto the filaments was ~500 ng, the blank levels were negligible and thus no blank corrections were applied. Pb isotopic measurements were made on a MM30 mass spectrometer using the silica gel method at 1250 °C. Measured isotopic ratios were corrected against the recommended value of NBS 981 by Todt et al. (1996) for mass fractionation (~0.14 % a.m.u.$^{-1}$). The reproducibility of the Pb isotopic analyses is 0.19, 0.20 and 0.28 ‰ (1σ) for $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$, respectively, based on ten analyses of NBS 981 standard. For the Sm-Nd isotopic analyses, whole rock powders were spiked with mixed ^{149}Sm-^{150}Nd tracer prior to the dissolution and were passed through cation and HDEHP chromatography to separate Sm and Nd (Creaser et al. 1997). Nd isotopic ratios were measured using five collector VG354 thermal ionization mass spectrometer in multidynamic mode. All ratios were normalized to $^{146}Nd/^{144}Nd$ = 0.7219. The value obtained for the La Jolla analyzed during this study was 0.511848 ± 8 and the in-house "Nd-oxide" standard gave an external reproducibility of 0.000016 (2σ).

Results

Major and trace element geochemistry

Results of major and trace element analyses are given in Table 4.2. Granitoids of the ~2.62 Ga Defeat suite are felsic (70-74% SiO_2) and characterized by low K_2O/Na_2O, Rb/Sr and low Ni, Cr, and V. The rare earth element (REE) patterns have moderate La_N/Yb_N with low Yb_N and small or no Eu/Eu* (Fig. 4.2a). They are weakly peraluminous with A/CNK of 1.0 to 1.1, and are classified as tonalite and granodiorite (La Maitre 1989). Geochemical characteristics of these granitoids are generally similar to those of ~2.62 Ga Concession suite granitoids from central Slave province (Fig. 4.2a; Davis et al. 1994). One major difference, however, is the observed range in SiO_2 content. While the granitoids of the Concession suite have a SiO_2 content ranging from 53 to >70%, the SiO_2 content of Defeat suite granitoids analyzed in this study and those reported in Henderson (1985) is restricted to >65% and plutonic rocks of mafic to intermediate composition are rare.

Defeat suite sample Y-31 has geochemical features different to other samples of this suite. It shows high Rb and low Sr, Ba and exceptionally high Y. The REE pattern is flat with pronounced negative Eu/Eu* (Fig. 4.2b). The REE pattern of this sample is very similar to that of Y-25, a two-mica granite of ~2.59 Ga Prosperous suite. Sample Y-25 is also characterized by high Rb and low Sr, Ba, but the Y concentration of this sample is much lower than Y-31. The REE patterns of these two samples are broadly similar to that of Contwoyto suite granitoids in the central Slave province (Fig. 4.2b; Davis et al. 1994).

Granitoids of ~2.59 Ga Stagg and Awry suites are geochemically very similar to granitoids of Yamba suite in the central Slave Province (Fig. 4.2c; Davis et al. 1994). Compared to the ~2.62 Ga Defeat suite granitoids, Stagg and Awry suite granitoids have higher Th, Rb, K_2O/Na_2O, lower Sr, Ba and consequently high Rb/Sr (> 0.8). Their REE patterns are LREE enriched with pronounced negative or positive Eu/Eu* (Fig. 4.2c). These weakly to strongly peraluminous granitoids can be classified as granodiorite to monzogranite (La Maitre 1989). The range in SiO_2 is also similar between granitoids of Stagg, Awry and Yamba suite (>70%).

Syn- to post- deformation granitoids from the Wijinnedi Lake area are geochemically similar to the granitoids of the same age from the Yellowknife area (Table 4.2). 2605 Ma syn-deformation granitoids are tonalitic in composition and are characterized by low K_2O/Na_2O, Rb/Sr and lack of Eu/Eu* (Fig. 4.2d). 2589-2598 Ma post-deformation granitoids, on the other hand, are peraluminous with higher K_2O/Na_2O, Rb/Sr. They show positive or negative Eu/Eu* (Fig. 4.2e) indicating involvement of plagioclase and/or K-feldspar as accumulating, restitic and/or fractionating phases.

Two samples from the MacNaughton Lake area (ML20A, ML21A) correlated with the ~2590 Ma magmatism show broadly similar geochemical characteristics, with either positive or negative Eu/Eu* (Fig. 4.2f). Two other samples from the MacNaughton Lake area (FL24A, FL74A) show very different features, with low K_2O/Na_2O, Rb/Sr and low Ni, Cr, and V. Geochemically, these granitoids appear more similar to the syn-deformation magmatism in spite of their U-Pb age (Fig. 4.2f). Sample FL74A has high Al_2O_3, Ca and a positive Eu/Eu*, indicating accumulation of plagioclase.

Nd, Pb isotopic data

The 2.61 - 2.62 Ga syn-deformation granitoids from the southwestern Slave province have εNd_T values ranging from -1.2 to 1.1, and a very restricted range of depleted mantle model ages (TDM) of 2.9 to 3.0 Ga (Table 4.3). There is no clear difference in εNd_T

between granitoids from the various study areas. Davis and Hegner (1992) have reported Nd isotopic data of syn- to post-deformation granitoids from eastern and western Slave province in the 64°N to 66° N region (referred to as east-central and west-central plutons hereafter) and found the west-central plutons to have a significantly greater "crustal" signature than the east-central plutons. This difference was considered as evidence for the existence of pre-2.8 Ga basement in the western Slave province but not in the eastern Slave province. When compared with the east- and west-central plutons, the granitoids from the southwestern (SW) Slave province have εNd_T values slightly lower than the east-central plutons, but mostly higher than the west-central plutons (Fig. 4.3). It is also noteworthy that the majority of the syn-deformation granitoids in the Slave province have positive εNd_T, suggesting that the role of the pre-YKSG basement may have been limited during the genesis of these granitoids.

The ~2.59 Ga post-deformation granitoids from the southwestern Slave province have εNd_T values of -0.8 to +3.3 with TDM of 2.8 to 3.0 Ga (TDM of sample FL24 is not reliable because of high $^{147}Sm/^{144}Nd$). These values are significantly more juvenile relative to the ~2.59 Ga post-deformation granitoids from west-central Slave province (Fig. 4.3; Davis and Hegner 1992), indicating that their protoliths are different in terms of average crustal residence age. This is explicable if the protolith of these granitoids are (1) younger than that of west-central plutons but older than that of the east-central pluton and/or (2) the protolith is a mixture of juvenile crust and pre-YKSG basement in variable proportions (Davis et al. 1996).

Although the Nd isotopic system can provide useful information on the average crustal residence age of the granitoid's protoliths, the Pb isotopic system is much more sensitive in detecting the presence of older crust. This is because the half lives of ^{235}U and ^{238}U are shorter than ^{147}Sm, and the difference in concentration between mantle and crust is much larger for Pb compared to Nd.

The results of Pb isotopic signatures of highly leached feldspars and plagioclases are summarized in Table 4.4. For all samples where both leachates and the residue were analyzed, leachates and residues plot along a 2.6 Ga reference isochron, confirming that U, Pb in these samples behaved as a closed system since crystallization (Fig. 4.4). It is, however, difficult to completely eliminate the radiogenic components in feldspars though a leaching procedure. A small difference in Pb isotopic composition between L4 and R in most of the samples suggests that although most of the radiogenic Pb was removed through leaching, there may still be some radiogenic component in the residue and therefore the Pb isotopic composition of the residues must be treated as a maximum value for the initial Pb isotopic composition.

The Pb isotopic compositions of granitoids from the Wijinnedi Lake and MacNaughton Lake areas are slightly lower, or overlap with the least radiogenic samples from the Yellowknife area (Fig. 4.5). This small difference in the isotopic signature, which was not detected using Nd isotopes, implies an existence of isotopic variation even within the southwestern Slave province. The least radiogenic of samples from the Wijinnedi Lake and MacNaughton Lake areas approach the Pb isotopic composition of the Malley Rapids granitoids in the east-central Slave province. This difference in the Pb isotopic signature indicate that the protoliths of Wijinnedi Lake and MacNaughton Lake area granitoids may be younger than in the Yellowknife area and/or are a mixture of juvenile crust and pre-YKSG basement in differing proportions.

Discussion

Experimental constraints for the source of syn- to post-deformation granitoids

The geochemical characteristics of the syn-deformation granitoids from the southwestern Slave province (i.e. mostly felsic composition with low K_2O/Na_2O, Rb/Sr and no negative Eu/Eu*) impose important restrictions on their origin. These features are, in general, characteristic of dehydration melting of mafic protoliths such as garnet amphibolite (or hornblende eclogite) at ≥0.8 GPa (Martin 1986, 1994, Rapp et al. 1992). The expected residual phases in this case include garnet and hornblende with minor clinopyroxene and plagioclase. Subsequent fractional crystallization and/or assimilation may have taken place, but addition or removal of large quantities of plagioclase/K-feldspar is unlikely as this will produce a negative Eu/Eu*. This proposed origin contrasts with that suggested for the ~2.61 Ga Concession suite of the central Slave province by Davis et al. (1994). These authors noted that the least fractionated granitoids of the Concession suite are chemically similar to monzodiorites from southwestern Superior province ("sanukitoid suite" of Stern et al. 1989), and thus are mantle derived; more felsic granitoids are derived from these magmas by fractionation. However, these more mafic granitoids are volumetrically minor, and the dominant rock type of the Concession suite (and Defeat suite) is tonalite. Potential felsic derivatives of a "sanukitoid suite" follow a considerably potassic trend and do not include tonalite (see Fig. 4. 3 of Stern et al. 1989). We support the general model of Davis et al. (1994) for the genesis of the more mafic granitoids of the syn-deformation magmatic episode, but contend that the more felsic granitoids originate in a different manner. The simplest explanation for the origin of the syn-deformation granitoids is, therefore, a combination of two models; partial melting of an enriched mantle to produce the mafic - intermediate granitoids, and dehydrating melting of amphibolite to produce the more felsic granitoids. This is in accord with the observation that the mafic and felsic granitoids of the Concession and Defeat suites comprise discrete field units; gradational relationships that might support a genetic link between these rock types are not observed (Davis et al. 1994). Partial melting of amphibolite can be achieved if the mafic volcanic rocks, which are widely available in the Slave province at 2.6 Ga, are carried to a depth of ~30 km during ~2.6 Ga crustal thickening. Perks (1997) has described partial melting relationships in mafic granulites, interpreted to be YKSG mafic volcanic protoliths, in the MacNaughton Lake area. Partial melting of crustally-contaminated YKSG mafic volcanic satisfies the geochemical data and would also yield the observed Pb-Nd features.

The origin of sample Y-31 is not certain. However, despite the unique REE pattern, the major element characteristics of this sample (e.g. K_2O/Na_2O and A/CNK) are similar to the other samples of the Defeat suite, indicating that its source is probably not sedimentary rocks. One possibility may be the melting of basaltic rocks at shallower depth (< 0.8 GPa) where plag ± cpx ± hbl, instead of garnet, are residual phases.

There are several possible ways of producing granitoids with major element characteristics (e.g. higher K_2O/Na_2O, A/CNK) similar to the post-deformation granitoids from the southwestern Slave province granitoids. They are; (1) hybridization of high-Al basalt with partially melted biotite gneisses or metapelites (Patino Douce 1995, McCarthy and Patino Douce 1997), (2) partial melting of tonalite (Rutter and Wyllie 1988, Patino Douce and Beard 1995, Singh and Johannes 1996), (3) partial melting of greywacke at low H_2O - low T condition (Conrad et al. 1988) and (4) partial melting of metapelites (Vielzeuf and Holloway 1988, Patino Douce and Johnson 1991).

Of the four candidates mentioned above, the hybridization of basalt and biotite gneisses/metapelites is not considered viable because there is no evidence for mafic igneous activity in the Slave province at ~2.59 Ga. Dehydration melting of tonalitic rocks can also produce peraluminous melts with $K_2O/Na_2O > 1$ (Rutter and Wyllie 1988, Patino Douce and Beard 1995, Singh and Johannes 1996). The composition of the melts are granitic to granodioritic, but the melt tends to be more granitic in a system where biotite, rather than amphibole, is the main hydrous phase. This model is appealing because tonalitic rocks are widespread in the Slave prior to 2.6 Ga, but temperatures >900°C are required to generate significant amounts of partial melt. In addition, this model cannot explain the regional Nd isotopic signatures of the post-deformation granitoids if they are derived from syn-deformation tonalites, as the ~2.62 Ga Concession suite with positive εNd_T is intruded by post-deformation plutons of Yamba suite with negative εNd_T (Davis et al. 1994). Partial melting of tonalitic rocks, however, can explain the low εNd_T of post-deformation granitoids if the tonalitic protolith is a syn-volcanic plution and/or a mixture of pre-2.8 Ga basement tonalite and younger syn-volcanic and syn-deformation granitoids.

Experiments using synthetic greywackes were performed by Conrad et al. (1992). In their experiments, they showed that the melts produced at high H_2O condition are tonalitic to trondhjemitic whereas decreasing H_2O will shift the melt composition to granodioritic to granitic. The composition of the melt also changes with the temperature, and at relatively low temperature (825-850 °C) and low H_2O (H_2O~4%), a peraluminous granitic melt with K_2O/Na_2O~1 is produced. Plagioclase is present as a restitic phase which explains the Eu/Eu* of the post-deformation granitoids.

Experiments using pelitic rocks as a starting material has shown that at temperature and pressure equivalent to mid to lower crust range, the melts produced are strongly peraluminous felsic granites (Vielzeuf and Holloway 1988, Patino Douce and Johnson 1991). The $MgO + FeO + TiO_2$ of these granites are typically less than 3% and the K_2O/Na_2O values are >1. The restitic phase includes garnet, alumino silicate, quartz, ilmenite with lesser Ti-rich biotite and plagioclase at lower temperature (~900°C), and spinel, ilmenite, quartz and alumino silicate at 950 to 1050°C. Although most of the melts produced by melting of pelitic rocks have higher in $Al_2O_3/(CaO+Na_2O+K_2O)$ than the granitoids of Stagg and Awry suite, melts produced at relatively low temperature are chemically similar to the Stagg and Awry suite granitoids. Melting of pelitic may also be responsible for two-mica granites of Prosperous suite as this granite has lower $MgO + FeO + TiO_2$ and higher $Al_2O_3/(CaO+Na_2O+K_2O)$ compared to the granitoids from the Stagg and Awry suite.

Melting of metasedimentary rocks seem to explain both the geochemical and the isotopic characteristics of the post-deformation granitoids on a regional scale. In the areas where there is a clear evidence for the existence of pre-YKSG basement at the time of YKSG sediment deposition (e.g. Point Lake and Napaktulik Lake area), the εNd_T values of the post-deformation granitoids tend to be more negative or "basement like" compared to other areas where the basement was probably more limited (e.g. southwestern Slave province and eastern Slave province; Davis and Hegner 1992, Yamashita et al. 1995, 1996, 1997). Such origin is also supported by field relationships described by Perks (1997) in the MacNaughton Lake area, where granulite facies YKSG metasedimentary rocks show evidence of partial melting.

Pb isotopic constraints on the mantle extraction age of pre-YKSG component

The lack of difference in εNd_T together with the similarity in the initial Pb isotopic composition of granitoids from Yellowknife and Wijinnedi/MacNaughton Lake areas questions the existence of major east-west trending isotopic boundaries within the western Slave province suggested by Davis et al. (1996). One key question that needs to be answered in order to evaluate properly the isotopic structure of the western Slave province is whether the 3.1 to 3.4 Ga (and possibly ~4.0 Ga) basement found in other areas of the Slave province existed in the southwestern Slave province and contributed to the formation of the 2.59~2.62 Ga granitoids. Combined Nd and Pb isotopic analyses of granitoid rocks can help answer this question.

The first parameter that needs to be addressed in order to model the Pb isotopic signature of crustal rocks is the μ value (i.e. time-integrated $^{238}U/^{204}Pb$) of the mantle at the time of granite formation. This is a very difficult parameter to define because mafic and ultramafic rocks with very little or no crustal contamination are not documented in the Slave province and small amount (~1%) of crustal contamination can significantly alter the Pb isotopic signature of mantle-derived rocks. It is, however, possible to constrain the mantle μ value by using the Pb isotopic signature of the Slave province ~2.7 Ga volcanogenic massive sulfide (VMS) deposits.

Figure 4.6a shows the Pb isotope data of VMS deposits from the Slave province (Thorpe et al. 1992, Thorpe: pers. comm.). The simplest explanation for the linear trend formed by the regional VMS deposits is a mixing line between crustal and mantle components because (1) this linear trend is not a 2.7 Ga primary isochron, (2) the linear trend is unlikely to represent mixing between two crustal reservoirs as this would produce evenly distributed data rather than the observed bimodal distribution (given the difference in Pb concentrations between crustal reservoirs is small compared to crust versus mantle), and (3) there is field, geochemical and isotopic evidence for an input of mantle derived magmas during the ~2.7 Ga magmatism (e.g. MacLachlan and Helmstaedt 1995, Yamashita et al. 1995, 1997). The Pb isotopic ratio of the mantle in this case can be estimated by calculating the intercept between this mixing line and the 2.7 Ga primary isochron (Fig. 4.6a). A single stage Pb evolution curve from this intersection point to the value of Canyon Diablo (Tatsumoto et al. 1973) is thus a "first order" estimate of the single stage mantle curve. The term "first order" is used here because in reality, the U/Pb ratio of the upper mantle may change with time as more crust is extracted from the mantle and/or crustal material is recycled back into the mantle by subduction.

The μ value of the mantle calculated in this way, taking the age of the Earth as 4.559 Ga, is ~6.9. This value is significantly lower than the maximum μ of the mantle (μ=7.8) estimated by Davis et al. (1996) and determined from other Archean blocks (e.g. Dupré and Arndt 1990). However, μ=6.9 may be reasonable because (1) a single stage mantle with μ=7.8 cannot produce the linear array formed by Slave VMS deposits by mixing with more radiogenic crustal reservoir (see Fig. 4.6a) and (2) estimated μ of present-day mantle is 4.7-5.9 whereas present-day μ of silicate Earth is ~7 to 9 (Allègre et al. 1988).

An alternative method of estimating the mantle μ is to use, instead of the age of the earth, the age of bulk silicate earth, estimated to be ~4.47 Ga by Allègre et al. (1996). In this case, the slope of the 2.7 Ga primary isochron approaches that of the mixing line and therefore the mantle μ can be as high as ~8.5 (Fig. 4.6b).

The third model presented here is the modified version of the second model. In this model, the μ value of the mantle decreases with time from 4.47 Ga to the present to simulate progressive geochemical depletion of the mantle. The mantle that passes though the intercept between 2.7 Ga primary isochron defined by these evolution curves and the linear trend formed by Slave VMS deposit is the μ=9 $\rightarrow$ 6 mantle (Fig. 4.6c).

These three mantle μ values, therefore, are reasonable endmembers with which to assess the Pb isotopic composition of the crust with different μ, extracted from that mantle at different times. In Figure 4.7a, the Pb isotopic compositions of crust "at 2.6 Ga" with μ=11 ,13 and 15, extracted from the single stage mantle (μ=6.9) at 2.8, 3.1, 3.4, 3.6 and 4.0 Ga are shown. For example, point A on this diagram represents the Pb isotopic composition of 2.6 Ga crust with μ of 13, which was extracted from the mantle at 4.0 Ga. Various model crusts are connected to form a two dimensional "μ-time" mesh and are compared with the data from the Slave granitoids.

An obvious and important conclusion from Figure 4.7a is that in order to produce the Pb isotopic signature of the granitoids from various different areas of the Slave province including the southwestern Slave province, crust of older than ~3.4 Ga is required. Specifically, it requires Pb to be extracted from the mantle into the crustal rocks at this time or earlier. For this modeling, a mantle μ value of 6.9 maximizes this effect. However, the basic argument does not change even if we use the estimated age of bulk silicate earth (T=4.47 Ga) as the age of mantle reservoir with μ values of 8.5 (single stage) and 9 $\rightarrow$ 6 (Fig. 4.7b and c). In these cases, the Pb isotopic composition of the mantle at 2.6 Ga is much closer to the values of the Slave granitoids. Therefore, the crust that is required to produce the granitoids of the Slave province can be younger (Fig. 4.7b and c). However, even in these cases, the crust extracted from the mantle after ~3.2 Ga cannot produce the Pb isotopic signature of the granitoids in the Slave province unless it is isotopically contaminated by >3.2 Ga crust. This observation of extensive crustal recycling is compatible with the Nd and U-Pb isotopic studies of sedimentary rocks in the Slave province (Isachsen and Bowring 1997, Yamashita et al. 1996, 1997).

Evaluation of juvenile crustal sources using the Nd-Pb isotopic system

The next question that needs to be answered is whether the addition of juvenile (<2.7 Ga) crust has played an important role in the genesis of these granitoids (i.e. was the protolith of these granitoids nearly 100% basement, or was it a mixture of older basement and a juvenile material added from the mantle at ca. 2.7 Ga?). As previously mentioned, it is difficult to detect the signature of mantle merely from Pb because Pb isotopic signature of the mantle-derived melts are readily altered by minor crustal interaction. Combined Pb and Nd isotopes, on the other hand, can provide better constraints.

The εNd_T values of syn- to post-deformation granitoids from Davis and Hegner (1992) and this study range from 2.7 to -5.3. The TDM of these granitoids are typically 2.8 to 3.3 Ga with majority being younger than 3.0 Ga. The εNd_T of these granitoids are higher than the $\varepsilon Nd(2.6)$ of pre-YKSG basement from other areas of the Slave province, which ranges from -4.7 to -11.6 (Yamashita et al. 1995, 1997, Yamashita and Creaser 1996, unpublished data in prep. for publication). The large difference in $\varepsilon Nd(2.6)$ between syn- to post-deformation granitoids and pre-YKSG basement indicates that juvenile material, in addition to the pre-existing crust, was probably involved in the genesis of the late Archean granitoids in the Slave province. This supports the idea that the protolith of these granitoids is a "mixture" of juvenile crust and an older basement.

The discrepancy in the age of the protolith of the granitoids predicted from Pb (>3.2 Ga) and Nd (mostly <3.0 Ga) systems is explained by considering simple mass balance constraints in a model system. Shown in Figure 4.8 are examples of two component mixing in $^{207}Pb/^{204}Pb$ versus εNd space. Curve A is a mixing line between basement (20 ppm Pb and 26 ppm Nd; upper crust of Taylor and McLennan 1985) and typical mid-oceanic ridge basalt (MORB; 11.179 ppm Nd and 0.489 ppm Pb; after Hofmann 1988). Curve B is mixing between basement and mafic rocks more similar to those of upper Kam Group (~9 ppm Nd and ~2.5 ppm Pb; unpublished data, same sample as those reported in Yamashita et al. 1996), which are interpreted to be a continental margin arc volcanic rocks (Isachsen and Bowring 1997). The $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios of the basement and the mantle were taken from the most radiogenic VMS deposit from the western Slave province and the model mantle calculated previously. The $^{208}Pb/^{204}Pb$ ratio were taken from the 2.6 Ga mantle and crust of Zartman and Heines (1988) but this ratio is used only to calculate the isotopic abundance of Pb and thus has minimum effect on the result of calculation. The Nd isotopic ratio of the basement was taken from a 3.4 Ga tonalitic basement from the Napaktulik Lake area (same data reported in Yamashita et al. 1995, 1997, Yamashita and Creaser 1996) and the εNd(2.6) of the mantle were estimated to be at +5 using the $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ of present day depleted mantle (Goldstein et al. 1984). In the case of basement-MORB mixing, assimilation of ~20% crust will change the Pb isotopic signature of mantle-derived melt to a basement-like signature. However, at 20% assimilation, the Nd isotopic composition is still intermediate between mantle and basement. The change in the Pb isotopic composition is not as abrupt for the basement-arc mixing because the difference in the Pb concentration between basement and basalt is smaller. However, even in this case, ~30% assimilation will change the Pb isotopic signature of the melt to basement-like, but keep the Nd isotopic signature intermediate between the basement and the mantle.

This change in the isotopic signature of juvenile crust has an important effect on the isotopic signature of the granitoids generated by crustal melting, regardless of where the crustal component actually originates. If the "crustally contaminated" juvenile crust (i.e. volcanic rocks of YKSG and syn-volcanic plutons) is weathered to form sedimentary rocks (such as Burwash Formation turbidites in the Yellowknife area), and if these sediments are the source of the post-deformation granitoids, the isotopic signature of older basement should immediately show up in Pb but to lesser degree for Nd (see Fig. 4.8). The possibility of ca. 2.66 Ga sedimentary rocks in the Yellowknife area being derived, at least in part, from crustally contaminated volcanic rocks, is supported from the geochemical and Nd isotopic study of the sedimentary rocks (Jenner et al. 1981, Yamashita et al. 1996, 1997, Chapter 3). Using the Nd-Pb mixing model, the isotopic signature of granitoids from Yellowknife and Wijinnedi Lake areas can be explained by mixing of 10 to 30 % basement (70 to 90% juvenile crust) whereas granitoids from Point Lake area requires >50% basement. Isotopic signature of granitoids from the Contwoyto Lake area is variable, overlapping with the southwestern Slave province and the Point Lake areas (Fig. 4.8).

Regional isotopic structure of the Slave province

Geochemical characteristics of syn-deformation granitoids from the southwestern Slave province are most compatible with dehydrating melting of garnet amphibolite or hornblende eclogite at pressure ≥0.8 GPa whereas mafic to intermediate syn-deformation granitoids from the central Slave province may be mantle derived (Davis et al. 1994). In these cases, the isotopic variation of the granitoids is a function of assimilation of pre-YKSG basement by mafic protoliths or mantle derived magma.

Protoliths of the post-deformation granitoids are more difficult to specify. However, partial melting of sedimentary rocks seems to be the best candidate to explain the geochemical and isotopic characteristics of these granitoids on a regional scale. In this case, the isotopic signature of basement is first incorporated into the 2.66-2.72 Ga igneous rocks through assimilation (Yamashita and Creaser 1996, Yamashita et al. 1997), and this signature is transferred into sedimentary rocks through the weathering process. In addition to this, a direct input of detritus from the basement may take place in areas where the basement was exposed at the time of YKSG deposition. Granitoids produced by melting of these sedimentary rocks are expected to have an isotopic signature similar to its source (Fig. 4.9).

For both syn- and post-deformation granitoids, it is proposed that the major endmembers for the source of these granitoids are basement with age probably greater than ~3.2 Ga and a 2.66-2.72 Ga juvenile crust. Although it is possible that crust of <3.1 Ga (like those in Sleepy Dragon complex) was also involved, this by itself cannot produce the Pb isotopic signature of syn- to post-deformation granitoids from any part of the Slave province if this crust is mantle derived. The mixing model for the genesis of syn- to post-deformation granitoids in the Slave province eliminates the need for east-west trending isotopic boundaries within the western Slave province proposed by Davis et al. (1996). Instead, we propose that the most crustal Nd-Pb isotopic signatures in the syn- to post-deformation granitoids and the geochronological evidence of a pre-YKSG basement are found in the zone along the volcanic belts which run from Arcadia bay - Napaktulik Lake - Point Lake - Winter Lake - Sleepy Dragon complex (Fig. 4.1; Villeneuve et al. 1993, Isachsen and Bowring 1994, Villeneuve and van Breemen 1994, Yamashita et al. 1995, 1996, 1997, Yamashita and Creaser 1996, Bleeker and Stern 1997). Although there is a general tendency for the granitoids of the western Slave province to have evolved isotopic signature compared to those of the eastern Slave province, the isotopic signature of syn- to post-deformation granitoids tends to become more juvenile farther west from the "isotopic zone" mentioned above. This is most likely a result of mixing between basement and juvenile crust in various proportions, as well as the difference in the isotopic composition of the basement in different areas of the Slave province. One exception to this is the Point Lake area where the granitoids west of the isotopic zone have signatures similar to the granitoids along the zone. The Point Lake area, however, is a unique area within the western Slave province because this area is sandwiched by the a second set of pre-YKSG basement (i.e. 4.0 Ga Acasta gneiss and 3.0 Ga tonalite from Grenville Lake area; Bowring et al. 1989, Frith et al. 1986).

It is interesting to notice that the granitoids that have a Pb isotopic signature close to that of model 4.0 Ga crust are mostly from Point Lake, Napaktulik Lake and Contwoyto Lake areas. Although it remains rather speculative at this stage, this may indicate that the major influence of 4.0 Ga crust is restricted to the central region (in terms of latitude) of the Slave province. In the remaining areas of the Slave province, the basement onto which the supracrustal rocks of the YKSG built was probably younger, at >3.2 Ga to <3.8 Ga, confirming the observation that the pre-YKSG basement in the Slave province is probably a package of crustal units with different age (Davis et al. 1996).

Conclusions

Combined geochemical and isotopic study of the late Archean granitoids in the Slave province has provided important information on the ages of the basement, as well as the extent of crustal recycling. Some of the important conclusions of this study can be summarized as follows.

1. Major and trace element characteristics of the 2.62 Ga syn-deformation granitoids from the southwestern Slave province are most compatible with dehydration melting of garnet amphibolite or hornblende eclogite at a pressure ≥0.8 GPa. Low εNd_T of these granitoids compared to the model depleted mantle at 2.6 Ga is a result of crustal assimilation by the mafic protoliths.

2. The model that best explains both the geochemical and Nd-Pb isotopic characteristics of ~2.59 Ga post-deformation granitoids on a regional scale is the partial melting of metasedimentary rocks, although the partial melting of tonalitic rocks may have been important as well. In either case, the wide range of εNd_T observed in these granitoids is due to mixing of basement and juvenile crust in various proportions.

3. Mixing of basement and juvenile crust can take place in the form of assimilation and fractional crystallization system and direct mixing of detritus from the exposed basement and "crustally contaminated" 2.7 Ga igneous rocks.

4. The east-west trending isotopic boundary within the western Slave province can be eliminated because the basement of >3.2 Ga was probably involved in the genesis of late Archean granitoids across the entire southwestern Slave province. Although it is possible for the ~2.9 Ga Sleepy Dragon type basement to be involved, this by itself cannot produce the observed isotopic signature of the granitoids from both Yellowknife and Wijinnedi Lake area unless the ~2.9 Ga basement itself is isotopically contaminated by the >3.2 Ga basement.

5. In spite of the general tendency for the western Slave province crust to have radiogenic Pb isotopic signature and lower εNd_T compared to the eastern Slave province crust, it seems that the crust with strongest "basement-like" isotopic signature are in the zone which runs from Anialik river - Napaktulik Lake - Point Lake - Winter Lake and possibly Sleepy Dragon complex. In the areas west of this zone, the isotopic signature of the basement becomes weaker. This may indicate that the pre-2.8 Ga basement in the western Slave province is restricted to certain areas, rather than lying beneath the entire western Slave province.

Table 4.1. Sample locations, crystallization ages and rock types for the granitoids from the southwestern Slave province.

Sample	Long.	Lat.	Suite* / Domain**	Age (Ma)	Rock type
(Yellowknife)					
Y-1	114°11.43'	62°24.73'	Defeat	2620	granodiorite
Y-25	114°01.79'	62°34.49'	Prosperous	<2590	monzogranite
Y-29	114°30.23'	62°27.93'	Defeat	2620	granodiorite
Y-30	114°41.52'	62°28.47'	Defeat	2620	granodiorite
Y-31	114°51.66'	62°30.90'	Defeat (?)	2620?	tonalite
Y-32	115°00.55'	62°33.14'	Stagg	2590	granodiorite
Y-33	115°10.82'	62°35.29'	Stagg	2590	monzogranite
Y-34	115°17.45'	62°39.51'	Awry	<2590	monzogranite
Y-35	115°27.30'	62°41.20'	Awry	<2590	monzogranite
Y-36	115°38.51'	62°43.28'	Awry	<2590	granodiorite
Y-37	115°48.15'	62°45.44'	Stagg	2590	monzogranite
Y-38	115°57.96'	62°46.91'	Stagg	2590	monzogranite
(Wijinnedi Lake)					
J-0391	115°24.36'	63°46.31'	Ghost	2605	tonalite
J-0398	115°15.07'	63°48.98'	Ghost	2605	granitoid gneiss
HBA-0505-93	114°54.25'	63°45.98'	Ghost	2589	granitoid gneiss
J-0397	115°15.04'	63°47.83'	Ghost	2598	monzogranite
J-0400	115°15.22'	63°49.94'	Ghost	2593	monzogranite
(MacNaughton Lake)					
ML 20A	115°06.48'	63°31.50'	-	-	monzogranite
ML 21A	115°05.39'	63°30.07'	-	-	monzogranite
FL 24A	115°14.75'	63°38.94'	-	-	granodiorite
FL 74A	115°11.80'	63°38.94'	-	-	Qtz-monzodiorite

Notes: * After Henderson (1985); ** Henderson and Schaan (1993)

Table 4.2. Major and trace element compositions of syn- to post-deformation granitoids from the southwestern Slave province.

SAMPLE	Y-1	Y-25	Y-29	Y-30	Y-31	Y-32	Y-33	Y-34	Y-35	Y-36	Y-37
SiO_2	72.82	74.43	74.46	70.72	69.22	71.91	73.24	75.18	76.30	72.05	73.12
Al_2O_3	15.10	15.18	14.18	15.28	15.05	14.76	14.40	13.75	13.18	15.19	13.81
TiO_2	0.225	0.041	0.172	0.419	0.584	0.344	0.293	0.115	0.038	0.234	0.376
FeO^*	1.85	0.74	1.87	2.96	4.19	2.29	1.86	0.88	0.50	2.33	2.73
MnO	0.024	0.023	0.024	0.038	0.089	0.035	0.019	0.013	0.005	0.034	0.013
CaO	2.08	0.63	1.34	2.72	3.11	1.94	1.63	0.53	0.11	1.20	0.67
MgO	0.66	0.26	0.60	1.33	1.33	0.73	0.58	0.56	0.32	1.02	0.62
K_2O	1.78	4.46	2.75	2.45	2.05	4.01	4.20	5.83	6.86	3.98	5.65
Na_2O	5.40	3.99	4.54	3.98	4.26	3.85	3.72	3.10	2.65	3.79	2.89
P_2O_5	0.059	0.234	0.057	0.107	0.121	0.115	0.054	0.039	0.037	0.176	0.122
Ni*	11	12	5	11	11	11	9	7	8	9	10
Cr*	4	9	3	6	9	8	3	3	0	7	6
Sc*	1	8	7	8	13	11	3	2	2	6	5
V*	29	1	9	48	56	24	13	7	5	28	23
Zr*	104	17	143	172	162	233	179	111	55	120	217
Ga*	20	26	19	20	24	20	22	17	15	26	21
Cu*	4	6	7	40	9	12	4	4	10	8	11
Zn*	46	32	39	53	82	53	38	14	5	49	31
La	16.69	1.82	29.69	15.27	12.26	64.20	72.27	8.60	16.42	33.36	82.13
Ce	28.46	3.82	45.47	24.96	22.33	110.63	139.17	15.09	30.81	57.84	160.84
Pr	2.86	0.45	4.29	2.53	2.56	10.54	14.71	1.53	3.20	5.87	17.23
Nd	10.08	1.80	14.45	9.62	10.96	35.27	54.25	5.53	11.36	20.92	63.31
Sm	1.93	0.88	2.53	2.31	4.21	5.65	9.17	1.30	2.64	4.41	12.39
Eu	0.48	0.09	0.64	0.70	0.67	0.94	1.45	0.70	0.39	0.66	1.08
Gd	1.43	1.13	1.82	2.05	6.29	3.96	5.16	1.06	2.32	3.39	8.67
Tb	0.21	0.24	0.27	0.33	1.47	0.55	0.64	0.16	0.40	0.58	1.26
Dy	1.02	1.43	1.36	1.90	10.59	2.78	2.96	0.85	2.48	3.44	6.37
Ho	0.19	0.24	0.28	0.37	2.22	0.52	0.48	0.15	0.51	0.63	1.11
Er	0.46	0.65	0.76	0.98	6.61	1.41	1.19	0.40	1.41	1.52	2.54
Tm	0.07	0.10	0.12	0.15	1.13	0.22	0.17	0.07	0.21	0.22	0.33
Yb	0.45	0.63	0.75	0.85	7.78	1.33	0.99	0.46	1.23	1.35	1.78
Lu	0.07	0.09	0.13	0.13	1.19	0.22	0.17	0.09	0.19	0.21	0.24
Ba	441	24	940	428	178	925	593	686	320	381	349
Th	6.13	1.02	9.32	4.40	6.67	17.69	25.04	3.60	11.67	12.13	55.56
Nb	3.4	12.1	4.6	6.2	22.6	10.1	6.5	3.5	2.6	11.1	14.4
Y	5.24	8.20	8.15	10.21	80.77	14.82	13.37	4.53	14.62	18.27	29.01
Hf	2.26	0.76	4.07	3.74	5.52	6.22	4.08	3.71	2.40	3.41	5.66
Ta	0.86	1.86	0.43	0.50	3.12	1.02	0.69	0.36	0.20	0.89	0.94
U	1.82	4.09	3.05	1.48	5.24	3.71	2.57	1.66	5.63	7.37	19.74
Pb	14.11	24.35	11.85	18.31	21.52	24.93	19.62	25.98	32.51	16.04	24.31
Rb	59.0	266.1	63.2	63.5	228.2	154.4	129.7	128.8	168.9	143.7	185.9
Cs	2.0	12.7	1.5	1.9	8.1	1. 6	1.5	0.9	1.1	1.9	1.1
Sr*	226	18	198	208	147	149	151	131	43	150	81
A/CNK	1.03	1.21	1.10	1.08	1.01	1.04	1.06	1.11	1.10	1.19	1.14
K_2O/Na_2O	0.33	1.12	0.61	0.62	0.48	1.04	1.13	1.88	2.58	1.05	1.95
Eu/Eu*	0.88	0.28	0.91	0.98	0.40	0.61	0.64	1.82	0.48	0.52	0.32
Rb/Sr	0.26	14.78	0.32	0.31	1.55	1.04	0.86	0.98	3.93	0.96	2.30

Notes: * Analyses performed using XRF, all others using ICP-MS; Major element analyses are normalized on a volatile free basis; FeO^*=total iron as FeO; A/CNK=molar $Al_2O_3/(CaO+Na_2O+K_2O)$; $Eu/Eu^* = Eu_N/(Sm_N x Gd_N)^{0.5}$

SAMPLE	Y-38	HBA-0505-93	J-0400	J-0398	J-0397	J-0391	ML 20A	ML 21A	FL 24A	FL 74A
SiO_2	74.47	76.32	72.70	74.37	71.18	61.72	73.20	73.51	68.00	61.61
Al_2O_3	13.71	12.68	13.90	13.90	14.21	16.98	14.14	13.10	16.60	20.25
TiO_2	0.236	0.210	0.412	0.237	0.473	0.795	0.260	0.435	0.303	0.911
FeO^*	1.90	1.68	1.97	2.07	3.67	5.59	1.35	2.66	3.35	3.38
MnO	0.015	0.019	0.016	0.022	0.048	0.080	0.021	0.031	0.031	0.014
CaO	0.94	1.57	0.94	2.88	1.09	5.62	1.37	1.57	3.36	4.60
MgO	0.65	0.04	0.50	0.46	1.02	3.21	0.38	0.52	1.07	1.89
K_2O	4.30	4.59	6.74	1.57	6.11	1.80	5.56	4.90	2.59	2.50
Na_2O	3.70	2.86	2.74	4.42	2.05	3.93	3.24	3.14	4.47	4.99
P_2O_5	0.082	0.040	0.092	0.074	0.153	0.274	0.134	0.121	0.221	0.048
Ni*	12	8	9	14	16	39	12	8	6	20
Cr*	6	0	7	12	22	72	9	5	0	27
Sc*	2	0	1	0	11	13	0	1	10	4
V*	18	0	17	19	30	115	12	22	40	87
Zr*	159	484	274	219	224	180	172	335	149	107
Ga*	21	13	16	17	19	18	15	17	21	29
Cu*	9	8	3	16	10	9	2	3	11	18
Zn*	31	35	45	42	51	86	19	49	85	84
La	42.99	167.17	84.44	43.99	65.61	48.38	101.97	29.82	16.81	17.00
Ce	80.78	282.68	187.11	70.8	129.54	92.20	208.27	49.42	29.77	26.83
Pr	8.47	27.58	21.72	6.76	14.63	10.40	22.47	5.23	3.39	2.56
Nd	30.98	93.89	83.51	22.85	54.46	39.94	81.59	19.78	14.62	8.63
Sm	6.85	10.79	18.15	3.68	10.68	7.65	15.66	3.62	4.39	1.43
Eu	0.65	2.99	0.64	0.84	1.31	1.81	0.83	6.65	1.06	1.23
Gd	5.30	3.55	9.07	1.65	6.93	4.63	7.70	2.62	3.6	0.84
Tb	0.89	0.43	1.08	0.16	1.12	0.65	0.98	0.35	0.37	0.09
Dy	4.75	1.99	4.84	0.69	7.55	3.69	5.12	1.90	1.16	0.35
Ho	0.86	0.31	0.69	0.10	1.70	0.68	0.93	0.36	0.12	0.05
Er	2.08	0.70	1.26	0.22	5.00	1.74	2.57	0.87	0.21	0.11
Tm	0.29	0.12	0.14	0.04	0.66	0.24	0.35	0.13	0.03	0.02
Yb	1.62	0.71	0.76	0.26	3.59	1.39	2.20	0.91	0.19	0.09
Lu	0.23	0.14	0.11	0.05	0.51	0.22	0.33	0.15	0.04	0.02
Ba	408	3497	608	410	713	891	645	682	600	689
Th	27.70	11.70	72.85	10.61	25.56	5.64	67.67	5.41	0.51	0.40
Nb	14.5	2.1	9.9	2.7	11.2	8.3	5.1	11.8	5.3	11.0
Y	23.05	8.02	17.57	2.88	44.28	18.58	24.27	9.70	3.58	1.54
Hf	5.06	11.58	7.82	6.12	5.69	4.28	4.70	8.66	4.15	2.57
Ta	1.39	0.12	0.30	0.04	0.46	0.49	0.20	0.51	2.30	1.50
U	9.23	0.61	2.13	0.53	0.77	0.56	1.34	0.72	0.76	0.23
Pb	29.67	11.34	35.97	12.28	20.08	11.58	36.51	20.60	14.11	12.56
Rb	194.6	63.8	174.1	29.0	139.9	76. 7	134.2	115.9	43.8	91.5
Cs	1.5	0.1	0.8	0.2	1.1	1.4	0.3	0.2	0.2	0.7
Sr*	64	251	108	251	155	665	126	162	280	367
A/CNK	1.10	1.01	1.03	0.98	1.19	0.91	1.04	0.98	1.02	1.04
K_2O/Na_2O	1.16	1.60	2.46	0.36	2.99	0.46	1.71	1.56	0.58	0.50
Eu/Eu*	0.33	1.48	0.15	1.04	0.47	0.93	0.23	6.60	0.81	3.43
Rb/Sr	3.04	0.25	1.61	0.12	0.90	0.12	1.09	0.72	0.16	0.25

Table 4.3. Sm-Nd isotopic analyses for the southwestern Slave province granitoids.

Sample	Age (Ma)	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd^{a}$	εNd_T	TDM (Ga)
(Yellowknife)							
Y-1	2620	1.75	10.47	0.1012	0.511014 (10)	0.5	2.9
Y-25	<2590	0.83	1.77	0.2836	0.514082 (15)	-0.8	-
Y-29	2620	1.91	12.97	0.0891	0.510788 (7)	0.2	2.9
Y-30	2620	2.13	10.25	0.1261	0.511430 (9)	0.2	3.0
Y-31	2620	4.08	10.89	0.2263	0.513092 (15)	-1.2	-
Y-32	2590	5.73	39.45	0.0880	0.510836 (8)	1.1	2.8
Y-33	2590	7.91	53.28	0.0899	0.510894 (9)	1.6	2.8
Y-34	<2590	1.08	5.14	0.1271	0.511468 (18)	0.4	3.0
Y-35	<2590	2.46	11.56	0.1285	0.511648 (7)	3.5	2.7
Y-36	<2590	4.25	23.13	0.1112	0.511257 (8)	1.6	2.8
Y-37	2590	11.43	64.56	0.1071	0.511071 (12)	-0.7	3.0
Y-38	2590	6.53	33.72	0.1172	0.511330 (6)	1.0	2.9
(Wijinnedi lake)							
J-0391	2605	7.44	44.35	0.1014	0.511051 (13)	1.1	2.9
J-0398	2605	3.74	26.18	0.0864	0.510787 (8)	1.0	2.8
J-0397	2598	10.00	55.90	0.1081	0.511150 (13)	0.6	2.9
J-0400	2593	17.79	93.59	0.1149	0.511275 (9)	0.6	2.9
HBA-H0505-93	2589	11.64	107.2	0.0656	0.510456 (11)	1.1	2.8
(MacNaughton Lake)							
ML 20A	-	5.22	28.35	0.1114	0.511201 (6)	0.4	2.9
ML 21A	-	3.72	21.48	0.1048	0.511054 (8)	-0.3	2.9
FL 24A	-	4.52	16.31	0.1676	0.512113 (13)	-0.6	3.4
FL 74A	-	1.31	8.42	0.0942	0.510879 (9)	-0.2	2.9

Notes: [a] Normalized to $^{146}Nd/^{144}Nd=0.7219$. Numbers in (paretheses) $2\sigma m$ uncertainties.; TDM calcutaled using the mantle evolution model of Goldstein et al. (1984). Present day CHUR parameters are $^{147}Sm/^{144}Nd=0.1967$, $^{143}Nd/^{144}Nd=0.512638$. $\lambda_{147Sm}=6.54x10^{-12}$ a^{-1}

Table 4.4. Pb isotopic compositions of K-feldspar and plagioclase.

Samples	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Mineral
(Yellowknife)				
Y-29 (L4)	14.361	14.993	33.857	-
Y-29 (R)	13.993	14.921	33.740	Kfs
Y-30 (R)	14.133	14.960	33.827	Kfs
Y-31 (R)	14.134	14.994	33.873	Kfs
Y-32 (R)	14.010	14.900	33.733	Kfs
Y-33 (R)	14.057	14.981	33.970	Kfs
Y-34 (L1)	25.366	16.834	42.103	-
Y-34 (L4)	14.346	14.974	33.893	-
Y-34 (L4A~C)	14.199	14.957	33.942	-
Y-34 (R)	14.037	14.941	33.905	Kfs
Y-35 (R)	14.261	14.981	33.803	-
Y-36 (L1)	19.692	15.966	35.957	-
Y-36 (L4)	14.331	14.997	33.937	-
Y-36 (L4A~C)	14.277	14.979	33.930*	-
Y-36 (R)	14.079	14.931	33.827	Kfs
Y-37 (R)	14.380	15.046	34.117	Kfs
Y-38 (L1)	31.741	17.899	46.800	-
Y-38 (L4)	14.692	15.135	34.373	-
Y-38 (R)	14.309	15.041	34.054	Kfs
(Wijinnedi lake)				
J-0397(L1)	15.385	15.196	38.376	-
J-0397(L4)	14.156	14.981	34.244	-
J-0397(R)	14.087	14.975	34.127	Kfs
J-0400(L1)	24.603	16.700	114.28	-
J-0400(L4)	14.824	15.062	34.364	-
J-0400(R)	14.084	14.923	34.266	Kfs
HBA0505(R)	13.967	14.865	33.880	Kfs
J-0391(L1)	25.305	16.883	51.003	-
J-0391(L4)	14.250	14.942	33.902	Plag
J-0391(R)	14.057	14.899	33.831	Plag (rpt)
J-0391(R)	14.060	14.898	33.835	-
J-0398(L4)	15.348	15.086	33.943	-
J-0398(R)	14.225	14.899	33.886	Plag
J-0398(R)	14.230	14.903	33.910	Plag (rpt)
(MacNaughton Lake)				
ML 20A (R)	14.057	14.975	33.930	Kfs
ML 21A (L1)	18.308	15.745	40.714	-
ML 21A (R)	14.090	14.984	33.885	Kfs
FL 24A (R)	14.028	14.926	33.784	Kfs
FL 74A (R)	13.984	14.908	33.729	Kfs

Notes: L1 =leach 1 (24 hrs in 2N HCl), L4 =leach 3 (24 hrs in 16N HNO_3+1 drop 48% HF), L4A~C = (three 20 minutes leaches in 7N HNO_3+ 5% HF combined together); Kfs=K-feldspar, Plag=plagioclase; *:Larger error may be expected.

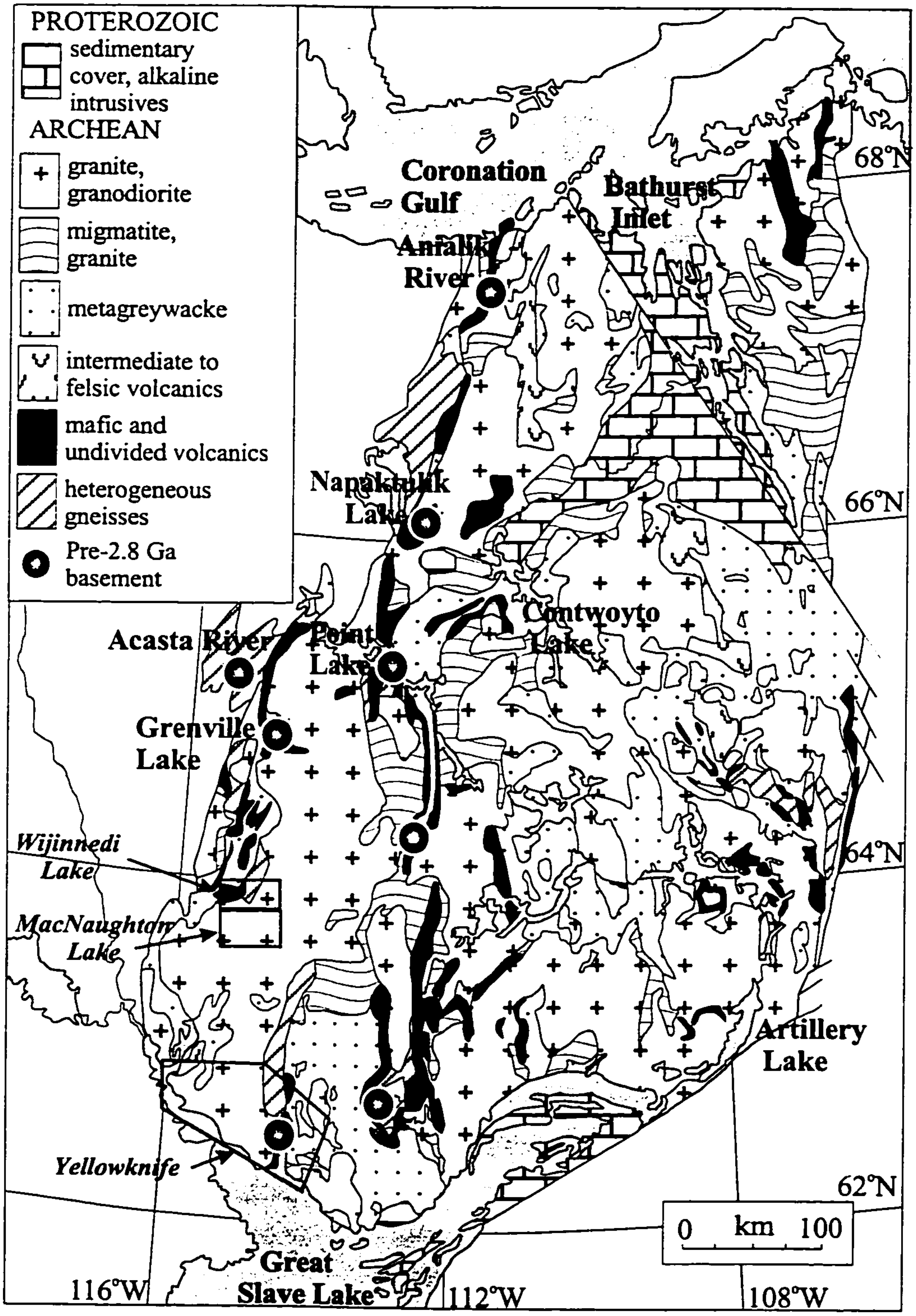


Fig. 4.1. Geological map of the Slave province showing study areas (after Hoffman 1989).

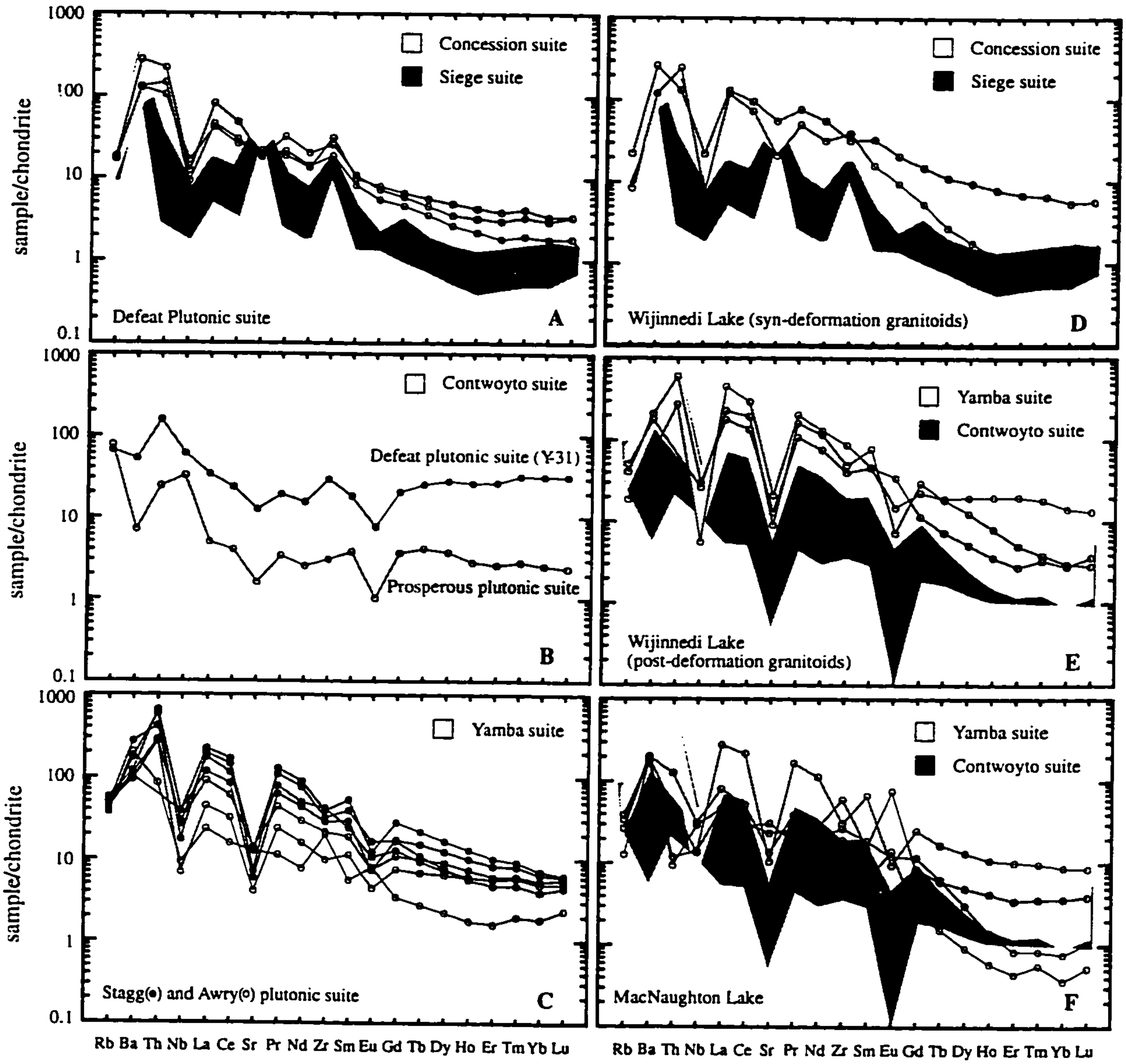


Fig. 4.2. Extended REE patterns for (a) Defeat plutonic suite (b) Defeat and Prosperous plutonic suite with flat REE pattern (c) Stagg and Awry plutonic suite (d) syn-deformation granitoids from Wijinnedi Lake (e) post-deformation granitoids from Wijinnedi Lake and (f) syn-and post-deformation granitoids from MacNaughton Lake. Data for the Siege, Concession, Contwoyto and Yamba suite granitoids used for comparison from Davis et al. (1994).

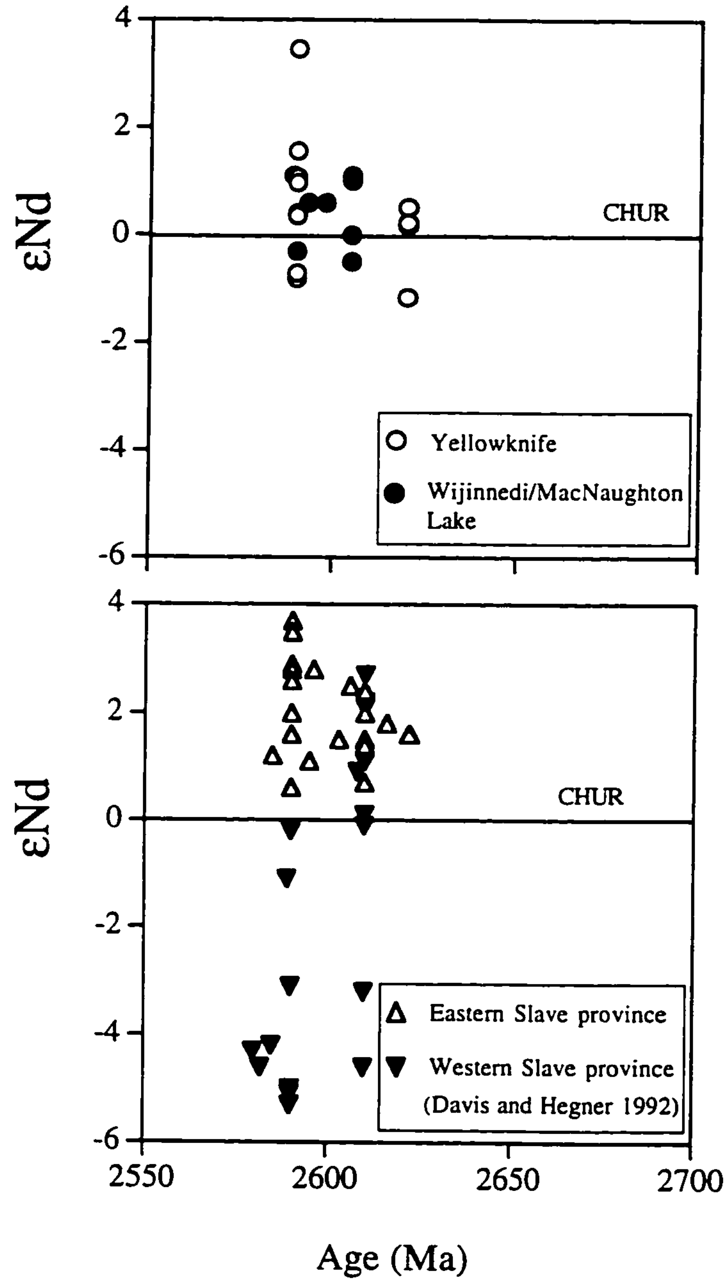


Fig. 4.3. Time vs. εNd diagram for syn- and post-deformation granitoids from the Slave province. (A) Southwestern Slave province (this study). (B) East- and west-central Slave province (after Davis and Hegner 1992).

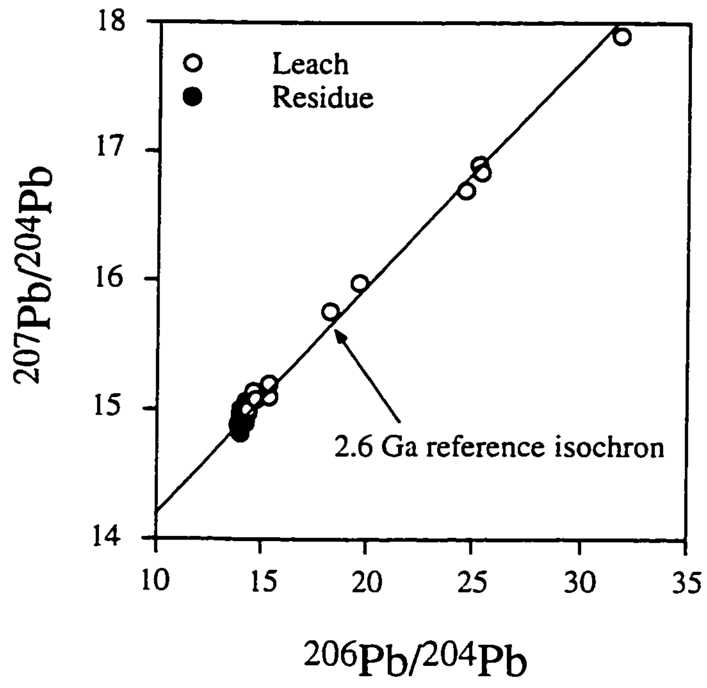


Fig. 4.4. Pb isotopic composition of feldspar leachates (open symbol) and residues (solid symbol) from SW Slave province granitoids. Leachates plot along a 2.6 Ga reference isochron, indicating that the feldspar Pb isotopic compositions has remained undisturbed since ~2.6 Ga.

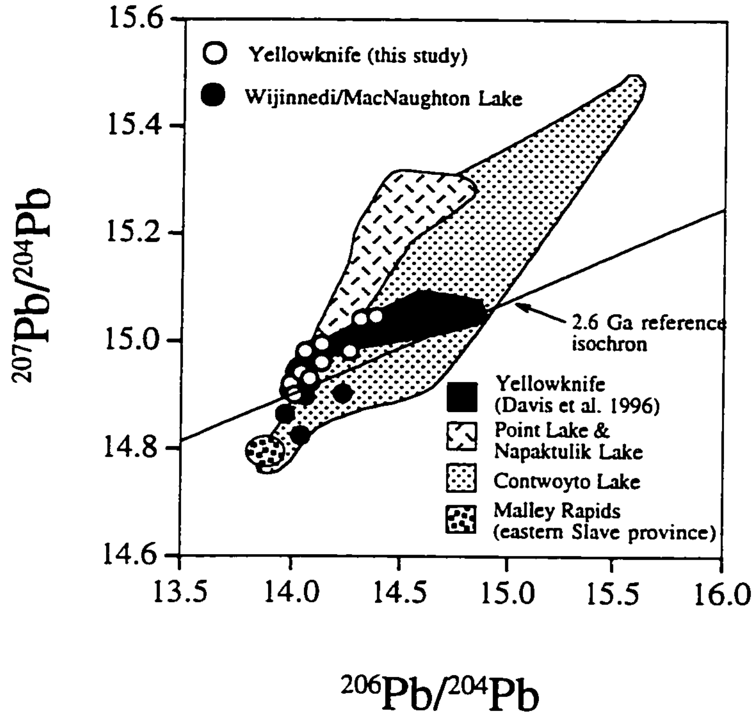
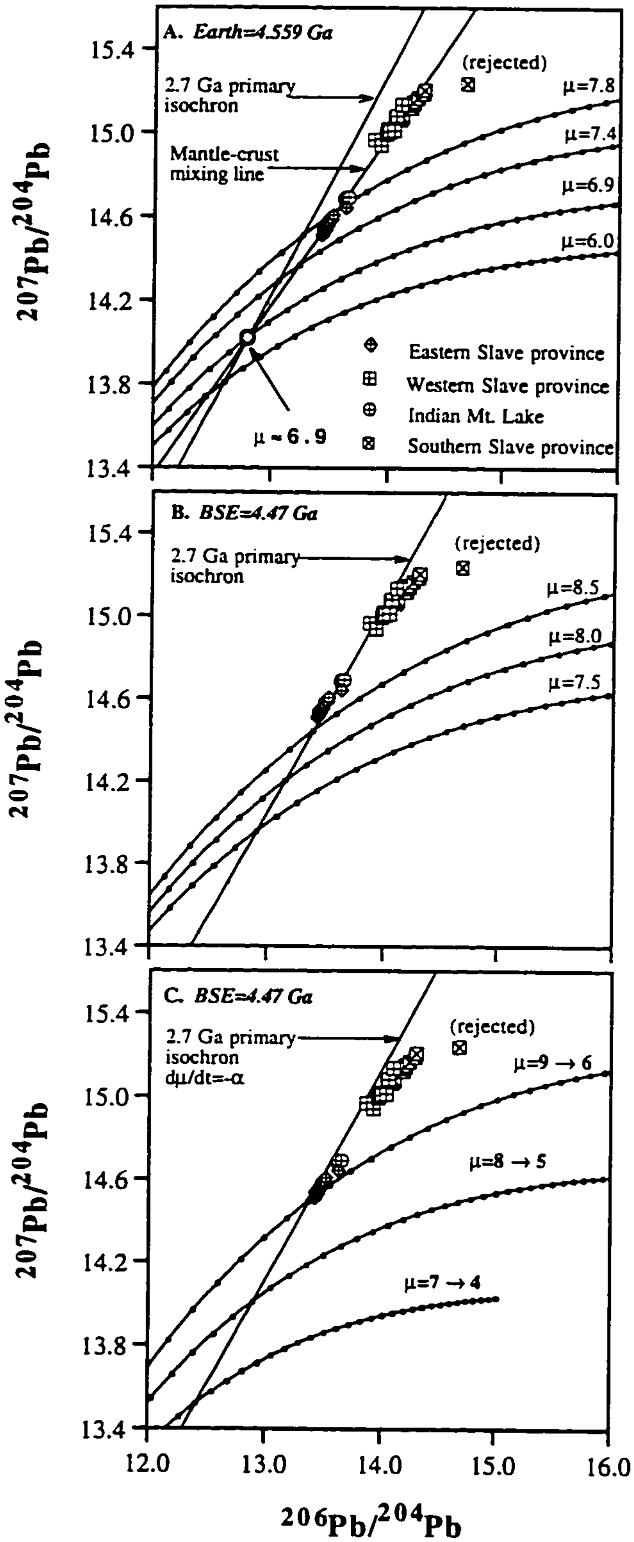


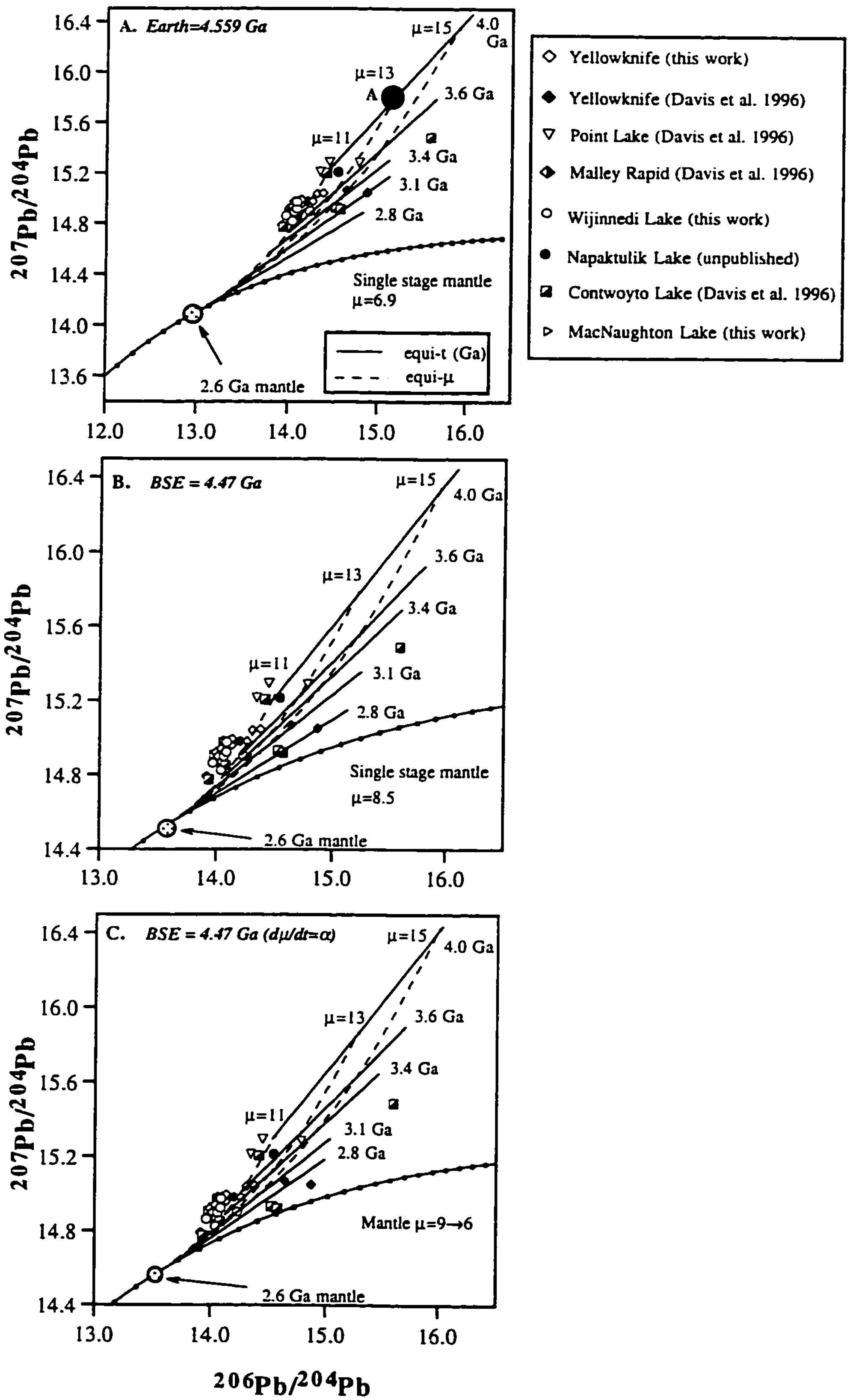
Fig. 4.5. Pb isotopic composition of feldspar from the southwestern Slave province granitoids. Fields for the Point Lake, Contwoyto Lake and Malley Rapids from Davis et al. (1996). Field for the Napaktulik Lake is from Yamashita et al. (1997).

Fig. 4.6. $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram showing the Pb isotopic compositions of Slave province VMS deposits and mantle with various μ. (A) Single stage mantle growth curves with an initial Pb isotopic composition similar to Canyon Diablo troilite (Tatsumoto 1973) and the age of the earth at 4.559 Ga (Chen and Wasserburg 1981). (B) Single stage mantle growth curve assuming the age of bulk silicate earth at 4.47 Ga (Allègre et al. 1988). (C) Mantle with continuously decreasing μ value. The age of bulk silicate earth is assumed at 4.47 Ga (Allégre et al. 1988).



A. Earth=4.559 Ga
2.7 Ga primary isochron
Mantle-crust mixing line
(rejected)
μ=7.8
μ=7.4
μ=6.9
μ=6.0
μ = 6.9
Eastern Slave province
Western Slave province
Indian Mt. Lake
Southern Slave province
B. BSE=4.47 Ga
2.7 Ga primary isochron
(rejected)
μ=8.5
μ=8.0
μ=7.5
C. BSE=4.47 Ga
2.7 Ga primary isochron
dμ/dt=-α
(rejected)
μ=9 → 6
μ=8 → 5
μ=7 → 4
207Pb/204Pb
206Pb/204Pb
15.4
15.0
14.6
14.2
13.8
13.4
12.0
13.0
14.0
15.0
16.0

Fig. 4.7. $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram showing the Pb isotopic composition of syn- to post-deformation granitoids from the SW Slave province shown with: (A) model data for crust extracted from a μ =6.9 single stage mantle (see text for detail) (B) model data for crust extracted from a μ =8.5 single stage mantle from bulk silicate Earth of 4.47 Ga (see text for detail). (C) model data for crust extracted from μ =9 $\rightarrow$ 6 mantle.


A. Earth=4.559 Ga
B. BSE = 4.47 Ga
C. BSE = 4.47 Ga (dμ/dt=α)
μ=15
μ=13
μ=11
4.0 Ga
3.6 Ga
3.4 Ga
3.1 Ga
2.8 Ga
A
Single stage mantle μ=6.9
Single stage mantle μ=8.5
Mantle μ=9→6
2.6 Ga mantle
equi-t (Ga)
equi-μ
207Pb/204Pb
206Pb/204Pb
Yellowknife (this work)
Yellowknife (Davis et al. 1996)
Point Lake (Davis et al. 1996)
Malley Rapid (Davis et al. 1996)
Wijinnedi Lake (this work)
Napaktulik Lake (unpublished)
Contwoyto Lake (Davis et al. 1996)
MacNaughton Lake (this work)

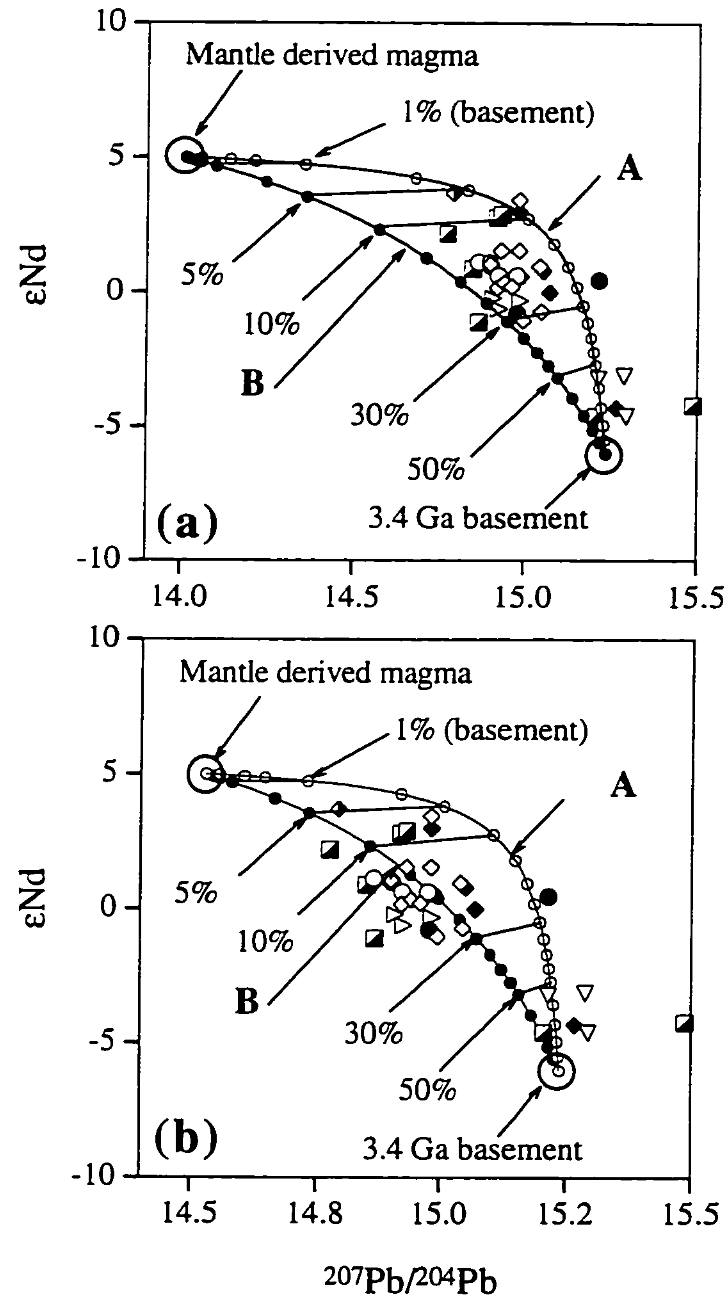


Fig. 4.8. $^{207}Pb/^{204}Pb$ vs. εNdT diagram showing the Nd-Pb isotopic signatures of the syn- to post-deformation granitoids from the Slave province. Line labelled A represents mixing between 3.4 Ga basement from the Napaktulik Lake area and typical MORB. Line labelled B represents mixing between basement used in A and basalt with Nd and Pb abundances of typical Kam Formation basalt (see text for detail). (a) $^{207}Pb/^{204}Pb$ of mantle derived magma ~14.01. (b) $^{207}Pb/^{204}Pb$ of mantle derived magma ~14.53. Symbols for the granitoids are same as Fig. 4.7.

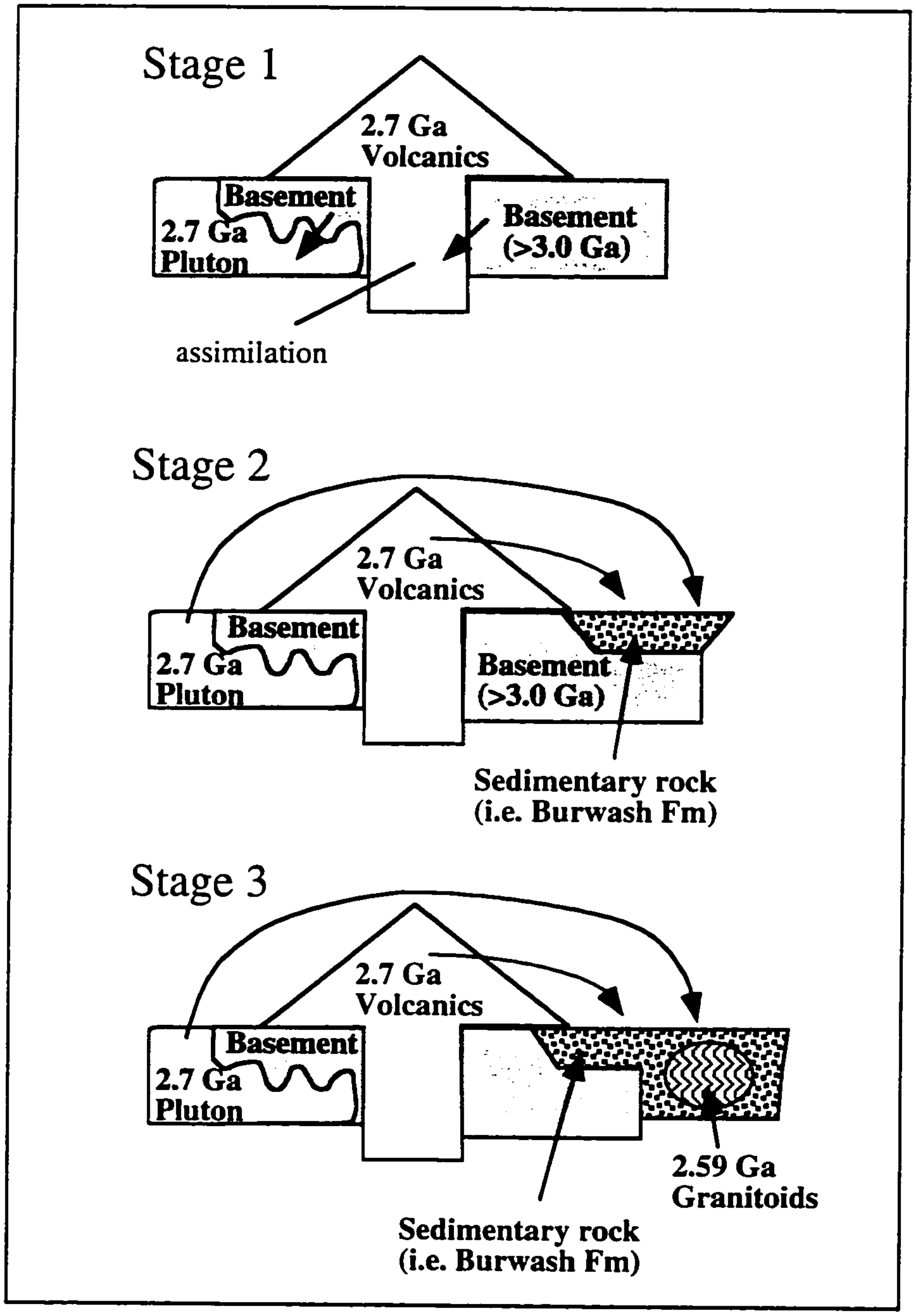


Fig. 4.9. A cartoon illustrating the mechanism for the generation of post-deformation granitoids in the Slave province. (Stage 1) The isotopic signature of the pre-2.8 Ga basement is incorporated into the ~2.7 Ga volcanic rocks (and syn-volcanic plutons) by assimilation. (Stage 2) The ~2.7 Ga volcanic rocks and syn-volcanic plutons are weathered to produce YKSG sedimentary rocks, likely similar to the Burwash Formation turbidites in the Yellowknife area (note: direct input of detritus from the pre-2.8 Ga basement may also take place at this stage, but this was probably not very important in the case of the Burwash Formation turbidites. (Stage 3) The sedimentary rocks are melted to produce the post-deformation granitoids.

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Chapter 5

Integrated Nd isotopic and U-Pb detrital zircon systematics of clastic sedimentary rocks from the Slave Province, Canada: Constraints for crustal recycling in early- to mid-Archean. *

* A version of this paper will be submitted to Earth and Planetary Sciences Letters. Co-Authored by Robert A. Creaser, Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta CANADA, T6G 2E3 and Mike E. Villeneuve, Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario CANADA, K1A 0E8

Introduction

The U-Pb ages of detrital zircon and the whole rock Nd depleted mantle model ages (TDM) of clastic sedimentary rocks are two independent "clocks" that can reveal important information on age of the sediment provenance. Combined with REE geochemistry, TDM has widely been used to investigate problems related to crustal evolution (e.g. Allègre and Rousseau 1984, Goldstein et al. 1984, O'Nions 1984, McLennan et al. 1990, McLennan and Hemming 1992) as well as the provenance monitoring of a relatively restricted area (e.g. Nelson and DePaolo 1988, Zhao et al. 1992). On the other hand, U-Pb geochronology of detrital zircon is used to constrain the maximum deposition age of the sedimentary rocks (i.e. the age of youngest zircon in a sample is taken as a maximum deposition age) or more for provenance age interpretations (e.g. Gehrels et al. 1995). The TDM value of sedimentary rock reveals the "depleted mantle" extraction age of all protoliths that contributed detritus. However, if a sediment is derived from multiple protoliths of different ages, TDM by itself cannot unravel the age of individual protoliths. In this respect, the U-Pb age of detrital zircon is much more useful. One major problem which U-Pb geochronology encounters, however, is the general lack of zircon in mafic and ultramafic rocks. Because zircon is usually present in intermediate to felsic rocks, the information which we can obtain from detrital zircon studies is clearly biased towards these rock types.

This effect may be compounded in Archean clastic sediments because the proportion of mafic crust in Archean cratons is likely higher than modern orogens (Taylor and McLennan 1985). By combining the two techniques, more reliable and less biased information on sedimentary protoliths can be obtained, but such studies have so far been scarce (Ross et al. 1991, Zhao et al. 1992).

This study presents Nd isotopic data from sedimentary rocks of the Archean Slave province, Canada, which have also been studied in detail for detrital zircon U-Pb geochronology. The main objectives of this study are to (1) investigate the relationship between T_{DM} and U-Pb detrital zircon ages for a suite of regional samples, (2) compare these data with simple model constraints for T_{DM} - U-Pb behavior, and (3) apply this to the crustal evolution of the Slave province to determine whether the early- to mid-Archean evolution of this craton was dominated by crustal recycling* or the addition of juvenile crust from the mantle (* Note: the term crustal recycling in this case is used to describe intracrustal processes, not recycling of crustal material back into the mantle). The Slave province is an ideal place for this study because the age of crust exposed in this craton extends back to ~4.0 Ga (Bowring et al. 1989, Bleeker and Stern 1997) and thus the age of the detritus potentially included in sedimentary rocks is expected to be highly variable. The results of this work should also provide a possible reason for the large volume of ancient crust not being preserved on the Earth today.

Geological setting

The Slave province is a relatively small Archean granite-greenstone terrane comprising an area of approximately 190,000 km^2 in the northwestern part of the Canadian shield (Henderson 1985, Fyson and Helmstaedt 1988, Hoffman 1989). It consists of three main lithotectonic groupings; ~2.82 to 4.01 Ga pre-Yellowknife Supergroup basement, ~2.65-2.72 Ga volcanic and sedimentary rocks of the Yellowknife Supergroup (YKSG; Henderson 1970) with syn-volcanic plutons and ~2.58 to 2.62 Ga syn- to post-deformation granitoids.

The pre-YKSG basement consists of granitoids, gneisses and supracrustal rocks which are thought to be restricted west of a north-south trending isotopic boundary at approximately 111° W (Fyson and Helmstaedt 1988, Bowring et al. 1989, Hoffman 1989, Thorpe et al. 1992, Davis and Hegner 1992, Villeneuve and van Breemen 1994, Davis et al. 1996, Bleeker and Stern 1997). Included in this group are the 3.96~4.01 Ga Acasta gneisses in the Acasta river area, 3.1 Ga tonalitic and migmatitic gneisses near Arcadia bay, 3.4 Ga tonalite from near Napaktulik Lake, 3.15 Ga Augustus granite at Point Lake area, 3.3-3.1 Ga rhyolites and a quartz arenite-orthoquartzite-oligomictic conglomerate sequence in the Winter Lake area, 2.9 Ga gneisses in the Sleepy Dragon complex and >2.93 Ga granitoid gneisses and >2.82 Ga cover sequence in the Yellowknife area (Bleeker and Stern 1997, Bowring et al. 1989, Henderson et al. 1982, 1987, Hrabi et al. 1993, 1994, Isachsen and Bowring 1994, 1997, Villeneuve and van Breemen 1994).

The 2.65 to 2.72 Ga mafic to felsic volcanic and greywacke-mudstone turbidite sequences known throughout the Slave province were grouped as Yellowknife Supergroup (YKSG) by Henderson (1970). Unlike the supracrustal sequences from most other Archean cratons, sedimentary rocks make up ~80% of the YKSG (Padgham and Fyson 1992). In addition to the 2.65 to 2.72 Ga metasedimentary rocks, ~2.59 to 2.62 Ga sandstones and conglomerates are now recognized from various different areas of the Slave province including Anialik River, High Lake, Hood River, Point Lake and Yellowknife (Fyson and Helmstaedt 1986, Hrabi et al. 1993, 1994, Henderson et al. 1994, Isachsen and Bowring 1994, Villeneuve and van Breemen 1994). The volcanic rocks of the Yellowknife Supergroup were mainly erupted during the 2.66-2.72 Ga interval, and recent Nd and Pb isotope work indicate that at least part of these volcanic rocks were erupted through the basement (Isachsen and Bowring 1994, Yamashita et al. 1995, 1997, Yamashita and Creaser 1996). The 2.65~2.72 Ga syn-volcanic plutons are also known, and at least part of them are interpreted to be the plutonic equivalent of YKSG volcanic rocks (Yamashita et al. 1995, 1997).

The 2.62~2.58 Ga syn- to post-deformation plutons are wide spread throughout the Slave province (van Breemen et al. 1992, Davis et al. 1994, Villeneuve and van Breemen

1994, Villeneuve et al. 1997). The syn- to post-deformation granitoids are generally felsic, ranging from granodiorite, tonalite to granite (Davis et al. 1994). Regional metamorphism and deformation in the Slave province took place between 2.58~2.63 Ga, with the metamorphic grade ranging from greenschist to granulite facies (Thompson 1978, Henderson and Schaan 1993, Henderson and Chacko 1995, Pehrsson and Chacko 1997).

Samples and Analytical methods

Samples of metasedimentary rocks were collected from the ~64° to ~68°N region of the west-central Slave province and include greywacke, arkose, quartzite and conglomerate (Fig. 5.1, Table 5.1). Samples of pre-YKSG basement, ~2.7 Ga YKSG turbidites and ~2.6 Ga YKSG sandstone-conglomerate sequences are all included in this study and each sample used here is classified into a specific stratigraphic category based on the results of U-Pb detrital geochronology and/or field relationship (Villeneuve 1998).

Sample size ranged from ~1 kg to ~40 kg. For the conglomerates, samples of 30~40 kg were crushed to ensure the sample homogeneity. All samples were crushed using jaw crusher and an aliquot of a few hundred grams was taken from each sample for the Sm-Nd isotope and major/trace element analyses. Aliquots for Sm-Nd isotopic and major/trace element analyses were further processed through an agate mill to obtain a very fine powder. The remaining portion from the jaw crusher was passed through standard mineral separation procedure for detrital zircon U-Pb geochronology described in Parrish et al. (1987). The U-Pb detrital zircon ages for the samples used in this study are summarized in Table 5.2. All U-Pb analyses were performed by conventional thermal ionization mass spectrometry, and the complete U-Pb data sets for these samples are available in Villeneuve (1998).

Major and trace element analyses were performed at Washington State University using XRF and ICP-MS. The details of the analytical procedures are given in Hooper et al. (1993). Analyses for the Sm-Nd isotopes were performed at the University of Alberta. Whole rock powders were weighed into clean Teflon vessels, spiked with mixed ^{149}Sm-^{150}Nd tracer and dissolved in ~5:1 mixture of 24N HF and 16N HNO_3 for 5-7 days at

160°C. Sm and Nd were separated using cation exchange chromatography and HDEHP (Di(2-ethylhexyl phosphate)) chromatography. Isotope dilution analysis for Sm was done on a single collector VG MM30 thermal ionization mass spectrometer, whereas the analyses for Nd isotopes utilized a five collector VG 354 thermal ionization mass spectrometer operated in a multidynamic mode. The measured values of $^{143}Nd/^{144}Nd$ for spiked samples of the La Jolla Nd standard and BCR-1 during the course of this study were 0.511848 ± 8 and 0.512634 ± 9, respectively. The external reproducibility of the University of Alberta "Nd oxide standard ($^{143}Nd/^{144}Nd$=0.511054)" was ± 0.000016 (2σ).

Results

Major/trace element geochemistry

Results for the major- and trace-element analyses are summarized in Table 5.3. For the fine grained sediments such as greywacke, arkose and quartzite, sample heterogeneity is usually not a problem. However, for conglomerates, there is a possibility that the aliquot taken for major/trace and Nd isotopic analyses may not be representative of the whole rock. This is particularly important for samples where the conglomerate clasts are relatively large (i.e. VN-93-04, VN-92-05, VN-92-02, HLB94B1440B). Evaluation of sediment homogeneity in terms of rare earth elements (REE) can be made by looking at its REE pattern. Shown in Figure 5.2b are the REE patterns of the conglomerate samples. In spite of their different sampling locations and different types of clasts included in these samples, the REE patterns of most of the conglomerates are very similar. Such similarities would be expected only if the sedimentary process (weathering, transportation and sedimentation) has sufficiently mixed the detritus, and the sample taken for the analyses was representative of this "well mixed" material.

All conglomerate samples except VN-92-12a are characterized by moderate LREE enrichment (La_N/Sm_N~3; $_N$ refers to chondrite normalized value), relatively flat HREE pattern (Gd_N/Yb_N =1.15~1.4) and small negative Eu/Eu* (~0.8). The combination of relatively flat HREE pattern and Eu/Eu* of ~0.8 are characteristics of sedimentary rocks from Archean high grade terranes (Taylor and McLennan 1985, McLennan 1989). The nature of the Eu/Eu* is particularly important in the Archean sediments (Taylor and McLennan 1985, McLennan 1989) because negative Eu/Eu* are typically seen in felsic rocks that have experienced intracrustal processes such as partial melting and/or fractional crystallization involving plagioclase (or K-feldspar) as a residual or fractionating phase. On the other hand, typical Archean TTG suites (tonalite-trondhjemite-granodiorite), which may derive from dehydration melting of garnet amphibolite or hornblende eclogite with garnet as a residual phase, show no negative Eu/Eu* and steep La_N/Yb_N (Martin 1986, 1994). The observed negative Eu/Eu* in the Slave province sedimentary rocks may thus imply that intracrustal processes were important, at least in certain regions of the Slave province. The La_N/Sm_N ratios of the conglomerate samples are slightly lower than that of a typical Phanerozoic sediment such as the post-Archean average Australian shale (PAAS: La_N/Sm_N ~4.33; Taylor and McLennan 1985, McLennan 1989), which is not consistent with significant input from a TTG source, as these rocks are characterized by high La_N/Sm_N (and La_N/Yb_N) ratios. The slightly lower La_N/Sm_N may therefore be explained by greater input from more mafic rocks compared to PAAS.

Only one quartz pebble conglomerate (VN-93-12A) has a distinct REE pattern. A mineralogical effect such as concentration of certain REE rich minerals may explain this feature. However the REE patterns of the important REE hosting minerals in sedimentary rocks such as monazite, allanite and zircon, are not compatible with the REE pattern of this particular sample because these minerals are characterized by pronounced negative Eu/Eu* (Bea 1996). The only REE rich mineral that may be compatible with the REE pattern of this sample is titanite. An alternate way of obtaining the REE pattern may be a bimodal mixing of typical Archean TTG (La_N/Yb_N=5 to >150; Martin 1986) and Archean mafic rocks with a flat REE pattern (e.g. Condie 1981), possibly accompanied by small addition of garnet to produce the low (Gd_N/Yb_N).

The REE patterns of greywacke, arkose and quartzite are shown in Figure 5.2a. With the exception of the quartzite sample (VN-93-20A), the REE patterns of these finer grained sedimentary rocks are generally similar. They are characterized by LREE enrichment (La_N/Sm_N=3.5~5.3), relatively flat HREE (Gd_N/Yb_N=1.3~2.0) with small or no negative Eu/Eu* (or positive Eu/Eu*). Sedimentary rocks with such REE patterns are the most common type of sediments found in the Archean granite-greenstone terrane and are suggested to represent a mixture of 50% Archean TTG and 50% mafic volcanic rock (Taylor and McLennan 1985, McLennan 1989). Jenner et al. (1981) have studied the petrography and major/trace element characteristics of the supracrustal and plutonic rocks from the Yellowknife area in the southern Slave province and concluded that sedimentary rocks of that area are made of 20% mafic-intermediate volcanic rock + 55% felsic volcanic rock + 20% granitic rocks. Although the geochemical characteristics of the fine grained sedimentary rocks analyzed in this study are not exactly the same as those studied by Jenner et al. (1981) or Taylor and McLennan (1985), the general similarities in the REE pattern suggests that they are derived mainly from a mafic (~intermediate) volcanic rock and TTG (or a volcanic equivalent). The quartzite sample (VN-93-20A), on the other hand, is characterized by higher (La_N/Sm_N=2.6), lower (Gd_N/Yb_N=1.0) and pronounced Eu/Eu* (0.5). This REE pattern, together with high Y content, can be explained by addition of HREE rich minerals with pronounced Eu/Eu* such as granulitic garnet (Bea 1996).

Nd isotopic analyses

The results of Nd isotopic analyses are summarized in Table 5.4. Taking the age of youngest zircon in each sample as the maximum deposition age, the εNd_T of each sample was calculated. In order to compare the εNd values of these samples at a given time, the εNd at 2.68 Ga (a typical age of YKSG supracrustal rock; e.g. Isachsen et al. 1991, Isachsen and Bowring 1994, Yamashita and Creaser 1996) was also calculated. The εNd_T values range from -4.9 to 2.9, and TDM from 2.9 to 3.8 Ga, respectively. The εNd(2.68) of these rocks range from -6.2 to 1.8 (Table 5.4). Unlike the result of the major/trace element analyses, there is no obvious difference in the range of εNd_T, εNd(2.68) and TDM between the conglomerates and fine grained sedimentary rocks. However, there is a clear difference in the εNd(2.68) values between the sedimentary rocks of different stratigraphic units (i.e. unit A, B and C in Table 5.1). The εNd(2.68) values of samples from unit A (pre-YKSG sedimentary rocks) are significantly lower than those of sediments from units

B and C (YKSG sedimentary rocks). Negative εNd(2.68) values of most samples from units A are clearly a result of sedimentary input from the pre-YKSG basement (supported by the range of U-Pb detrital zircon ages). Generally higher εNd(2.68) values of sediments from units B and C indicate that (1) there was little or no direct input of detritus from the basement and/or (2) the "basement like" isotopic signatures of these samples were diluted by detritus provided from younger ~2.7 Ga magmatism.

Perhaps the most interesting observation is the relation between the T_{DM} and the U-Pb detrital zircon age (Fig. 5.3). Shown in Figure 5.3 are the U-Pb ages of detrital zircon plotted against the whole rock T_{DM} for each sample. Most of the zircons have a U-Pb age that is significantly younger than the T_{DM}, but there is a positive 1:1 correlation between the age of oldest zircon and T_{DM} (i.e. U-Pb age = T_{DM}; Fig. 5.4). A similar correlation is also observed between the average age of the zircon and T_{DM}, but the average age of zircon is typically younger than the T_{DM}, and the discrepancy increases with increasing T_{DM}.

Discussion

Evaluation of model systematics

Although the T_{DM} of sedimentary rocks are generally interpreted to represent the average provenance age of sedimentary rocks, the average U-Pb age of zircon is younger than the whole rock T_{DM} in most of the samples studied here. In fact, the T_{DM} seems to be in much better agreement with the age of oldest zircon in each sample. In order to evaluate properly the relationship between T_{DM}, the U-Pb detrital zircon ages and the implications for crustal evolution of the Slave province; four different models of crustal evolution were constructed.

Figure 5.5a shows the case where the crustal evolution was dominated by addition of juvenile materials from the mantle (A and B) at two different times T1 and T2. These new crustal packages were subsequently eroded at T3 to produce the sedimentary rock (C). The expected T_{DM} of (A) and (B) is ~T1 and ~T2, respectively. The U-Pb age of zircon from (A) and (B) will be T1 and T2. If the sedimentary rock is a mixture of (A) and (B), TDM of the sedimentary rock should be in between T1 and T2, but the rock should contain zircon with U-Pb age of T1 and T2 (Fig 5.5a). In other words, the U-Pb age of the zircon should plot above and below the "concordant line" (TDM=U-Pb zircon age) on the TDM vs. U-Pb age plot. Specific Nd isotopic evolution lines for crust A-C are shown in Figure 5.6a.

Figure 5.5b, on the other hand, shows the case where the crustal evolution was dominated by recycling of pre-existing crust. In this model, a juvenile crust (A) is added to the crust at T1, and at T2, part of crust (A) is re-melted to form crust (B). In this case, the U-Pb ages of (A) and (B) will be T1 and T2, respectively. However, the TDM of both (A) and (B) should be ~T1 unless a severe fractionation of Sm/Nd takes place when the crust (B) is produced (such case will be discussed later in Fig. 5.5d). The consequence of this is that zircon in sedimentary rock (C), derived from crusts (A) and (B), should plot on or below the concordant line on the TDM vs. U-Pb age plot (Fig. 5.5b, 5.6b).

Figure 5.5c and 5.6c illustrate the case that is intermediate between the two processes mentioned above. In this case, a juvenile crust (A) and (B) is generated at T1 and T2, but the crust (B) is isotopically contaminated by crust A. It is difficult to precisely illustrate the isotopic signature of sedimentary rock (C) because the TDM of crust (B) will depend on the degree of contamination. However, the TDM of sedimentary rock is expected to be between T1 and T2 while the U-Pb ages of the zircon contained in this sample are T1 and T2. Thus, the overall position of this sample on the TDM vs. U-Pb age plot should lie somewhere in between the two cases mentioned above (Fig. 5.5c, 5.6c).

The fourth model is a modified version of the second model. In this case, the crust (A) which was generated at T1 is a mafic crust, and the crust (B) which was generated by partial melting of crust (A) at T2 is a felsic rock. Since the Sm/Nd ratio of crust (B) is expected to be significantly lower than that of crust (A), the TDM of sedimentary rock (C) will probably be in between T1 and T2 (Fig. 5.5d, 5.6d). An additional difference between this model and the second model is the lack of zircon in mafic crust (A). Although zircons can be found in most rock types (Heaman and Parrish 1991), they are particularly abundant in intermediate to felsic rocks. Therefore, the U-Pb age of zircon in sedimentary rock (C) is expected to be ~T2.

The actual sedimentary process is clearly more complicated than the above models because most of the sedimentary rocks are derived from multiple sources, each of which may have experienced the different processes mentioned above. However, if the crustal evolution was 100% controlled by addition of juvenile materials from the mantle, the age of detrital zircon from that rock should plot above and below the concordant line and the "average age" of all zircon should plot closest to the concordant line. On the other hand, as the degree of crustal recycling increases, the proportion of zircon plotting below the concordant line will increase and at nearly 100% crustal recycling, all of the zircon should plot below the concordant line and the age of the "oldest" zircon should plot on the concordant line (Fig. 5.7).

Effect of discordant zircon ages

The major artificial effect on U-Pb detrital zircon ages is the effect of Pb loss. Because the U-Pb age of zircon reported here is a $^{207}Pb/^{206}Pb$ age, a zircon that has experienced a severe Pb loss (or U addition) will have apparent U-Pb age that is younger than the true crystallization age. This can be shown by the following:

The $^{207}Pb/^{206}Pb$ ratio of a zircon which crystallized at Tc is

$$(^{207}Pb/^{206}Pb)^* = (1/137.88)\{(e^{\lambda 2Tc}-1)/(e^{\lambda 1Tc}-1)\} \qquad (1) \qquad *\text{: radiogenic}$$

where $\lambda 1$ and $\lambda 2$ are the decay constants of ^{238}U and ^{235}U respectively.

If this zircon loses a certain proportion ($0<\alpha<1$) of its radiogenic Pb at T1, the apparent $^{207}Pb/^{206}Pb$ ratio will be,

$$(^{207}Pb/^{206}Pb)' = (1/137.88)\{[\phi(e^{\lambda 2Tc} - e^{\lambda 2T1}) + (e^{\lambda 2T1} - 1)] / [\phi(e^{\lambda 1Tc} - e^{\lambda 1T1}) + (e^{\lambda 1T1} - 1)]\} \qquad (2)$$

where $\phi = 1 - \alpha$ (proportion of radiogenic Pb retained in the zircon)

Because $(^{207}Pb/^{206}Pb)' < (^{207}Pb/^{206}Pb)^*$ for zircons that have experienced a Pb loss (or U gain), the apparent Pb-Pb age of that zircon will be younger than its true crystallization age.

The difference between Tc and the apparent age is a function of ϕ and T1. An example of this is shown in Figure 5.8.

Although the % discordance of the zircon provides some ideas on the degree of Pb loss, the exact proportion of Pb that was removed from the zircon and the actual timing of the Pb loss cannot be constrained. Nevertheless, it is clear from Figure 5.8 that even for a zircon that has crystallized at 3.5 Ga, Pb loss of ~20% will have very little (<0.1 Ga) effect on the apparent age. An age difference of <0.1 Ga will have a minimum effect on the four models presented above because the accuracy of TDM is probably limited to ~0.1 Ga. We have, therefore, plotted all the zircon data in Figure 5.3, but those with high %-discordance (plotted as a solid circle) should be viewed with some caution.

Crustal evolution in the Slave province

The first conclusion that can be drawn by comparing the model on Figure 5.7 and the data on Figure 5.3 is that the pre-2.8 Ga (or pre-YKSG) crustal evolution of the Slave province was not ~100% controlled by addition of juvenile crust, but was dominated by recycling of the pre-existing crust. The crustal recycling in this case may include erosion, partial melting and assimilation of pre-existing crust during the later magmatic events. This conclusion, however, does not mean that there were no new materials added to the crust during the 2.8~4.0 Ga period. It simply precludes the direct addition of juvenile material from the mantle without significant contamination by the pre-existing crust. One petrologic model that is compatible with this conclusion is the model of Martin (1986, 1994) where the basaltic crust is partially melted to generate the TTG magma. This is in accord with the general observation that large proportion of the pre-2.8 Ga crust in the Slave province is tonalitic-trondhjemitic-granodioritic in composition (Yamashita, unpublished data in prep. for publication).

In spite of this conclusion, there does seem to be one period in the pre-2.8 Ga history of Slave craton when the addition of juvenile material played an important role. This is the interval between 3.1 and 3.2 Ga. A large portion of zircon data in this time interval plots above the TDM-U/Pb concordant line, suggesting that the proportion of newly added material from the mantle was large compared to the volume of pre-existing crust and/or the assimilation of pre-existing crust was minimum during this period. Interestingly, this period coincides with the period where there is a high frequency of U-Pb detrital zircon ages reported from the Slave province. Figure 5.9 is a histogram of all the U-Pb detrital zircon ages (produced by conventional isotope dilution method) so far reported from the Slave province (Schärer and Allègre 1982, Villeneuve and van Breemen 1994, Isachsen and Bowring 1997). Although there is some bias in the data in that a large number of zircon samples from two ~3.0 Ga sedimentary rocks of the Yellowknife area were measured (Isachsen and Bowring 1997), the general tendency for two large peaks existing at 2600~2700 Ma and 3100~3300 Ma holds even if we take this bias into account. Although the possibility of an abrupt decrease in crustal contamination at ~3.3 Ga cannot be ruled out, it seems more reasonable to accept this 3100~3300 Ma peak as evidence for a large input of juvenile (or relatively juvenile) materials into the Slave craton because a large part of pre-2.8 Ga felsic volcanic and granitoid rocks from the central Slave province have crystallization ages that fall within this range (see Bleeker and Stern 1997 and references therein). Thus, the most straightforward interpretation of the Nd and U-Pb isotopic signatures is that majority of the pre-2.8 Ga magmatic evolution of the Slave province was dominated by crustal recycling, with notable exception of 3.1~3.3 Ga interval when the input of juvenile material was much more significant (Fig. 5.10). The question on whether this 3.1~3.3 magmatic event led to a significant growth in the volume of the Slave craton, however, is uncertain because an addition of juvenile material from the mantle could be balanced by recycling of crustal material back into the mantle (Armstrong 1981, 1991).

The second large peak in the histogram is seen at 2600-2700 Ma interval. This age corresponds to the age of YKSG magmatic event (Isachsen et al. 1991, van Breemen et al. 1992, Isachsen and Bowring 1994, Villeneuve and van Breemen 1994). The lack of zircon plotting above the concordant line at ca. 2.7 Ga is related to the general lack of sedimentary rocks with T_{DM} of ~2.7 Ga. The youngest T_{DM} of sedimentary rock analyzed in this study is 2.9 Ga. This is a result of (1) contamination of YKSG magma by pre-2.8 Ga crust (Dudás et al. 1990, Yamashita and Creaser 1996, Yamashita et al. 1995, 1997) and (2) a direct input of pre-2.8 Ga crust into the YKSG sedimentary rocks. This may indicate that (1) the importance of crustal recycling has increased during the 2.7 to 3.1 Ga interval and/or (2) a large contrast in the age between the YKSG magma and pre-2.8 Ga crustal rocks began to buffer the isotopic signature (or TDM) of the juvenile material.

Use of different TDM models

For the arguments made so far, the TDM was calculated using the model of Goldstein et al. (1984). However, in order to confirm that the conclusion is not model dependent, we must also look at other models of the depleted mantle evolution. The model of Goldstein et al. (1984) assumes that the εNd value of the depleted mantle has increased linearly from εNd=0 at 4.5 Ga to εNd=+10 at present day. However, recent Nd isotope study of the early- to mid-Archean rocks has shown that the εNd value of the depleted mantle may have increased very rapidly to ~+4 at 4.0 Ga and increased steadily to ~+6 at the end of Archean (Bowring and Housh 1995). If we utilize this "concave-up" depleted mantle evolution curve, the calculated TDM will be older than those calculated using Goldstein's model because the intercept between the mantle and the crust will shift to the high TDM side (see Fig 5.11). The exact change in the TDM cannot be evaluated at this stage because the εNd value of the very late Archean depleted mantle is not fully constrained. Nevertheless, the general shift of all data on Figure 5.3 to high TDM side will not change our conclusion that the pre-2.8 Ga crustal evolution of the Slave province was dominated by crustal recycling rather than addition of juvenile material. The only change this shift may cause is that some of the zircons that were plotting above the concordant line on Figure 5.3 may no longer plot above the line. However, change in TDM for one of the samples (VN-93-12A; Fig 5.11) shows that zircon with U-Pb age > whole rock TDM cannot be fully eliminated.

The second model of depleted mantle that we must consider is that of DePaolo (1981). Unlike the model of Bowring and Housh (1995), the shape of this mantle evolution curve is "concave-down". If we calculate TDM using the model of DePaolo (1981), the TDM of each sample will shift to low TDM side as shown in Figure 5.11.

However, it is apparent from Figure 5.11 that most of the zircon will still plot below the concordant line and the general observation that the age of "oldest" zircon, rather than the average age of all zircon, correlating with the T_{DM} still holds. We can therefore conclude that the use of different models to calculate the T_{DM} does not change the arguments made.

It has been suggested by several authors that the relatively small volume of earliest crust is a result of efficient recycling of crustal material back into the mantle (Armstrong 1981, 1991, Bowring and Housh 1995). Our observation from the Slave province suggests that intracrustal processes may have been an additional (or alternate) reason for the earliest crust not being preserved on the Earth today.

Conclusions

The combined Nd isotope and U-Pb geochronological study of sedimentary rocks from the Slave province has led to the following conclusions.

(1) There is a 1:1 correlation between the T_{DM} and the age of "oldest" U-Pb detrital zircon in each sedimentary rock. There is also a positive correlation between the T_{DM} and the average age of all zircon, but the average age is typically younger than the T_{DM}. This is true for both the conglomerate and the finer grained sedimentary rocks, which are thought to represent geochemically distinct areas of the Slave province.

(2) The general observation that the majority of zircon is younger than T_{DM} suggests that the pre-2.8 Ga crustal evolution of the Slave province was dominated by crustal recycling, rather than the addition of juvenile material from the mantle.

(3) An exception is the 3.1~3.3 Ga interval. During this period, the addition of juvenile material was significant and/or the assimilation of older crust was minimum. The large peak on the U-Pb detrital zircon age histogram supports the former possibility.

(4) The lack of 2.7 Ga T_{DM} in YKSG sedimentary rock is a result of both direct input from the pre-2.8 Ga crust and contamination of 2.7 Ga YKSG magmatism by older crust.

(5) In addition to recycling of crustal materials back into the mantle, intracrustal processes are probably responsible for the lack of a large volume of early to mid Archean crust preserved in the Slave province. Since zircon can be lost during the intracrustal processes

(unless they are preserved as an inherited core), the lack of abundant ancient detrital zircon cannot be used as an evidence for the absence of ancient crust.

Table 5.1. Sample locations, rock types and stratigraphic units for the sedimentary rocks from the Slave province.

Sample	Latitude	Longitude	Rock type	Unit
VN-93-09	66°40.30' N	112°15.77'W	Coarse greywacke	B
VN-93-20A	67°43.79' N	111°37.34'W	Quartzite	A
93HSA-J0394	67°02.81' N	110°57.10'W	Turbiditic greywacke	B
HLB94B2230A	64°21.06' N	112°30.08'W	Arkose	C
KIA93W32	64°37.04' N	110°26.65'W	Coarse greywacke	B
VN-93-10	67°08.40' N	112°18.89'W	Coarse greywacke	A
VN-93-02	64°22.01' N	112°26.65'W	Conglomerate	A
VN-93-04	67°43.90' N	111°39.16'W	Conglomerate	A
VN-92-05	67°34.98' N	111°06.18'W	Conglomerate	C?
VN-92-02	67°33.26' N	111°20.78'W	Conglomerate	C
HLB94B1440B	64°23.38' N	112°26.64'W	Conglomerate	C
VN-93-12A	67°09.91' N	112°10.73'W	Conglomerate	A

Notes: Units A, pre-YKSG; B, ~2.7 older YKSG;, C, ~2.6 Ga younger YKSG.

Table 5.2. U-Pb detrital zircon ages for the sedimentary rocks used in this study.

Quartzite, greywacke, arkose

	VN-93 09	VN-93 20A	93HSA J0394	HLB94 B2230A	KIA93 W032	VN-93 10
U-Pb Age	3254 (0.14)	3243 (0.02)	2677 (2.43)	2575 (10.18)	2710 (1.11)	2950 (1.04)
(Ma)	3096 (0.73)	3200 (5.90)	2664 (0.72)	2667 (6.15)	2691 (0.02)	3142 (0.08)
	2692 (-0.27)	3205 (6.17)	2714 (0.69)	2670 (3.95)	2669 (3.06)	3158 (0.89)
	2816 (0.00)	3231 (10.15)	2718 (1.85)	3188 (6.91)	2664 (0.70)	2953 (0.38)
	2681 (0.18)	3272 (9.53)	2795 (3.26)	2962 (3.64)	2892 (1.45)	3131 (0.27)
	2703 (0.25)	3302 (5.59)	2686 (1.77)	2630 (9.23)	2697 (0.00)	3057 (-0.83)
	2739 (-0.17)	3218 (5.02)	2688 (-12.4)	2666 (1.80)	2722 (1.26)	3579 (2.26)
	2708 (8.88)	3131 (5.34)	2717 (9.00)	2642 (3.83)	2680 (0.71)	3098 (0.25)
		3151 (12.45)	2690 (-0.73)		2704 (-0.31)	2944 (1.96)
T_{max}	3254	3302	2795	3188	2892	3579
Average	2836.1	3217.0	2705.4	2750.0	2714.3	3112.4
T_{min}	2681	3131	2664	2575	2664	2944

Conglomerate

	VN-93 02	VN-93 04	VN-92 05	VN-92 02	HLB94 B1440B	VN-93 12a
U-Pb Age	2992 (2.06)	3051 (17.08)	+ see below	2748 (37.5)	2698 (0.25)	2965 (0.49)
(Ma)	2942 (1.34)	2887 (0.32)		2697 (0.23)	2955 (0.70)	2960 (3.81)
	3075 (2.43)	3213 (11.68)		2600 (0.02)	3338 (0.70)	3113 (40.12)
	2945 (2.81)	3085 (9.26)		2701 (7.50)	2833 (-1.61)	3128 (11.21)
	2949 (2.04)	3452 (8.06)		2688 (0.04)	3170 (20.40)	3297 (3.42)
	3260 (5.67)	3259 (3.49)		2956 (0.54)	3159 (-1.06)	3478 (36.99)
	2983 (-2.19)	3018 (1.57)		2776 (1.05)	3057 (0.86)	3158 (8.28)
	2941 (1.13)	2882 (1.42)		2681 (7.32)	2698 (0.15)	3128 (11.21)
				2701 (1.73)	2699 (0.35)	
T_{max}	3260	3452	2956	2956	3338	3478
Average	3010.9	3105.9	2727.6	2727.6	2956.3	3153.4
T_{min}	2941	2881	2600	2600	2698	2960

Notes: * Numbers in parentheses represent % discordance. + No data, but assumed to be equivalent to VN-92-02 (Villeneuve; personal communication).

Table 5.3. Results of the major and trace element analyses.

	VN-93 09 (GR)	VN-93 20A (Q)	93HSA J394 (GR)	HLB94 B2230A (A)	KIA93 W032 (GR)	VN-93 10 (GR)	VN-93 02 (C)	VN-93 04 (C)	VN-92 05 (C)	VN-92 02 (C)	HLB94 B1440B (C)	VN-93 12a (C)
wt (%)												
SiO_2	76.58	86.09	74.64	67.24	61.80	78.25	59.23	62.40	72.89	69.16	66.12	89.98
Al_2O_3	10.73	8.04	13.72	14.38	20.65	10.99	14.59	12.53	13.73	13.10	12.61	7.71
TiO_2	0.554	0.066	0.467	0.928	0.768	0.460	0.914	0.715	0.601	0.581	0.721	0.309
FeO^*	4.79	0.99	3.81	8.44	6.20	3.58	10.54	7.97	5.57	6.26	6.68	0.35
MnO	0.079	0.042	0.029	0.138	0.083	0.037	0.239	0.153	0.065	0.112	0.160	0.000
CaO	2.04	1.10	0.78	3.45	1.27	0.81	5.37	5.63	0.70	3.60	6.85	0.04
MgO	1.84	0.35	1.98	3.94	3.72	1.90	4.99	6.33	2.99	2.79	2.70	0.00
K_2O	2.09	2.12	1.82	0.42	3.00	1.79	1.77	1.60	3.16	1.37	1.58	0.66
Na_2O	1.24	1.20	2.67	1.05	2.39	2.13	2.25	2.59	0.19	2.91	2.49	0.94
P_2O_5	0.042	0.008	0.081	0.011	0.132	0.049	0.091	0.083	0.120	0.120	0.089	0.013
(ppm)												
Ni*	51	24	41	72	94	43	83	155	56	56	49	12
Cr*	115	36	107	151	194	145	141	337	41	110	96	118
Sc*	13	0	11	17	18	5	32	26	8	14	18	11
V*	101	0	96	98	161	70	234	168	96	112	146	62
Ga*	14	18	17	19	29	14	20	14	16	14	15	6
Cu*	64	125	18	185	75	56	51	69	55	109	80	21
Zn*	85	20	56	119	112	65	101	96	73	54	86	10
U	2.92	0.71	0.91	2.80	1.83	7.56	1.05	0.75	0.93	1.03	1.25	2.93
Th	8.37	5.55	4.21	9.48	6.23	16.85	5.06	3.45	4.64	4.11	5.75	4.83
Hf	2.33	4.12	2.81	4.38	3.58	5.12	2.51	2.03	4.80	2.90	2.88	1.99
Zr*	100	106	108	162	131	189	93	85	167	120	103	71
Pb	31.47	7.40	7.66	11.23	12.64	35.37	10.86	12.56	4.15	2.55	10.20	10.33
Ba	290	355	407	72	729	355	347	417	332	314	360	58
Sr*	87	86	240	155	231	91	102	177	35	149	118	48
Rb	107.11	47.76	51.98	7.45	100.92	56.00	43.05	47.64	66.07	36.50	40.35	21.56
Cs	19.6	1.1	2.7	0.4	14.3	1.2	1.1	1.1	1.6	1.0	1.7	1.2
Nb	5.0	12.0	5.3	6.8	8.1	7.2	5.1	5.4	11.5	5.7	5.7	2.9
Y	12.59	43.32	10.70	17.67	18.71	17.23	24.14	16.95	39.12	17.64	21.48	6.22
Ta	1.75	0.97	0.40	0.68	0.55	1.16	0.42	0.46	0.96	0.43	0.47	0.29
La	15.07	19.63	16.07	33.59	31.58	30.92	13.32	12.30	22.73	17.52	17.96	14.95
Ce	25.91	38.46	31.20	58.41	60.31	52.57	23.28	21.99	46.01	33.95	32.98	22.54
Pr	2.67	4.38	3.46	6.07	6.73	5.28	2.61	2.44	5.30	3.81	3.60	2.04
Nd	9.62	17.64	13.35	22.83	26.60	18.58	10.61	9.85	21.32	14.93	14.38	6.57
Sm	2.19	4.68	2.56	4.72	5.45	3.67	2.98	2.42	5.34	3.30	3.38	1.05
Eu	0.60	0.78	1.06	1.13	1.28	0.96	0.84	0.71	1.10	0.85	0.84	0.29
Gd	2.00	4.99	1.86	3.69	4.09	3.09	3.33	2.68	5.46	2.92	3.26	0.74
Tb	0.34	1.06	0.31	0.61	0.63	0.52	0.62	0.48	1.04	0.51	0.59	0.15
Dy	2.15	6.96	2.04	3.58	3.64	3.03	4.11	3.00	6.67	3.09	3.70	1.09
Ho	0.45	1.59	0.40	0.69	0.71	0.62	0.93	0.66	1.43	0.65	0.77	0.25
Er	1.35	4.79	1.15	1.84	1.95	1.80	2.66	1.90	4.23	1.87	2.25	0.77
Tm	0.19	0.68	0.16	0.24	0.27	0.28	0.37	0.24	0.61	0.26	0.32	0.12
Yb	1.24	4.16	0.97	1.42	1.66	1.82	2.34	1.57	3.83	1.69	1.97	0.74
Lu	0.18	0.64	0.16	0.22	0.26	0.30	0.35	0.24	0.60	0.26	0.31	0.12

Notes: * Analyses performed using XRF, all others using ICP-MS; Major element analyses are normalized on a volatile free basis; FeO^*=total iron as FeO; A/CNK=molar $Al_2O_3/(CaO+Na_2O+K_2O)$; $Eu/Eu^* = Eu_N/(Sm_N \times Gd_N)^{0.5}$; (GR: Greywacke, Q: Quartzite, A: Arkose, C: Conglomerate)

Table 5.4. Results of the Sm-Nd isotopic analyses.

Sample	Age (Ma)*	Sm(ppm)	Nd(ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ [a]	εNd_T	$\varepsilon Nd(2.68)$	TDM[b]	TDM[c]
(Quartzite, greywacke, arkose)									
VN-93-09	2680	2.40	12.00	0.1210	0.511150 (8)	-2.9	-2.4	3.27	3.16
VN-93-20A	3131	4.47	18.88	0.1433	0.511683 (8)	2.9	0.3	3.16	2.99
93HSA-J394	2664	2.56	14.47	0.1069	0.511093 (8)	0.6	1.4	2.92	2.81
HLB94B2230A	2575	4.43	24.40	0.1097	0.511000 (8)	-3.0	-1.4	3.13	3.03
KIA93W032	2664	5.14	28.15	0.1105	0.511178 (8)	1.0	1.8	2.90	2.78
VN-93-10	2944	3.54	20.39	0.1049	0.510672 (13)	-3.7	-6.2	3.45	3.37
(Conglomerate)									
VN-93-02	2941	2.84	11.45	0.1497	0.511686 (10)	-0.8	-1.9	3.46	3.31
VN-93-04	2881	2.64	11.35	0.1404	0.511318 (12)	-4.9	-5.9	3.78	3.70
VN-92-05	2600	4.95	22.94	0.1304	0.511524 (8)	0.4	1.6	2.96	2.81
VN-92-02	2600	3.28	16.54	0.1198	0.511266 (8)	-1.0	0.3	3.04	2.91
HLB94B1440B	2698	3.23	15.01	0.1300	0.511305 (14)	-2.8	-2.5	3.34	3.22
VN-93-12A	2960	0.95	6.84	0.0841	0.510448 (9)	0.2	-3.4	3.16	3.08

Notes: [a] Normalized to $^{146}Nd/^{144}Nd=0.7219$. Numbers in (paretheses) represent errors in 2σ.; [b] TDM calcutaled using the mantle evolution model of Goldstein et al. (1984).; [c]TDM calcutaled using the mantle evolution model of DePaolo (1981); Present day CHUR parameters are $^{147}Sm/^{144}Nd=0.1967$, $^{143}Nd/^{144}Nd=0.512638$.; $\lambda_{147Sm}=6.54\times10^{-12}$; * Maximum deposition age (=Youngest U-Pb detital zircon age).

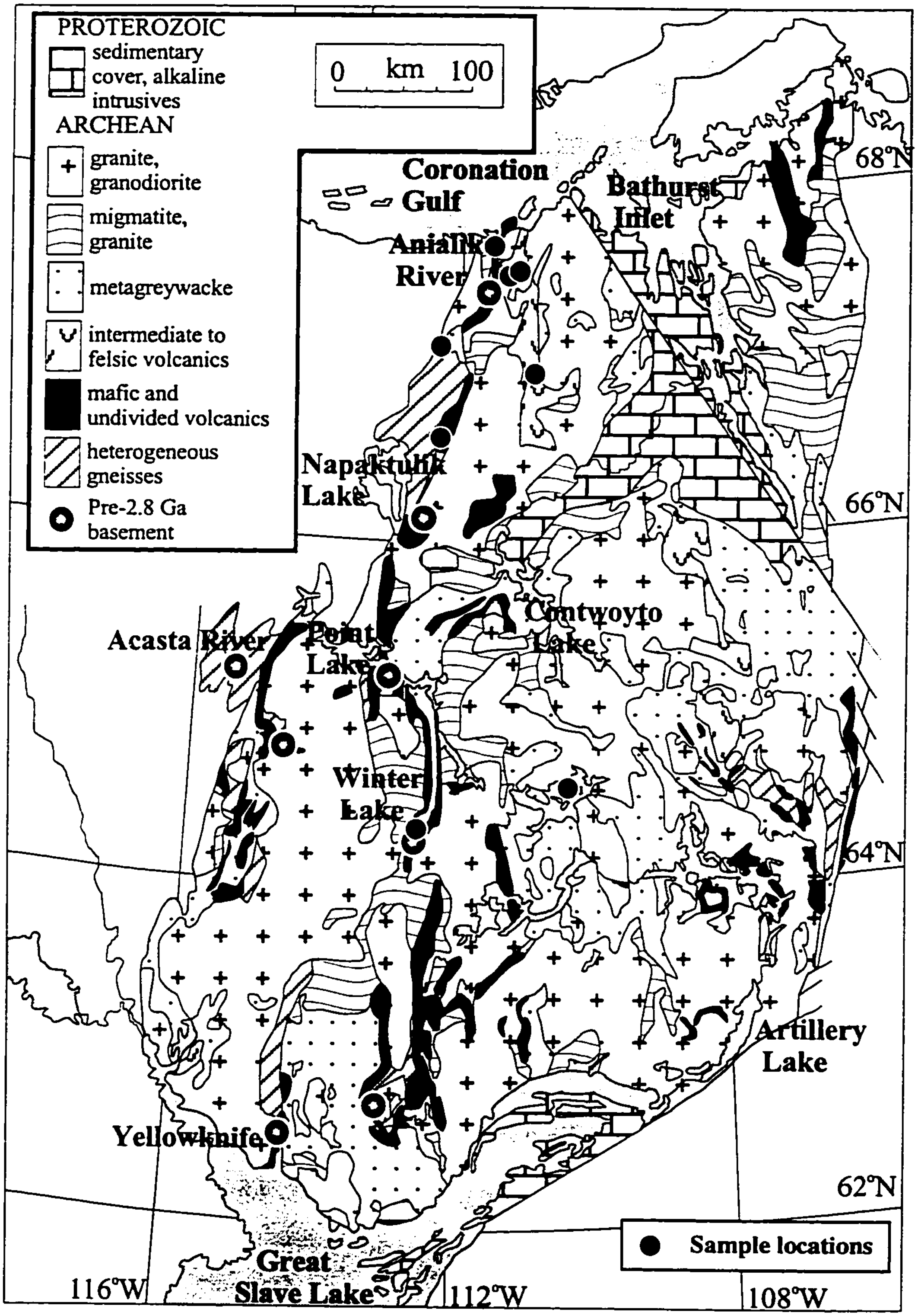


Fig. 5.1. Geological map of the Slave province (after Hoffman 1989).

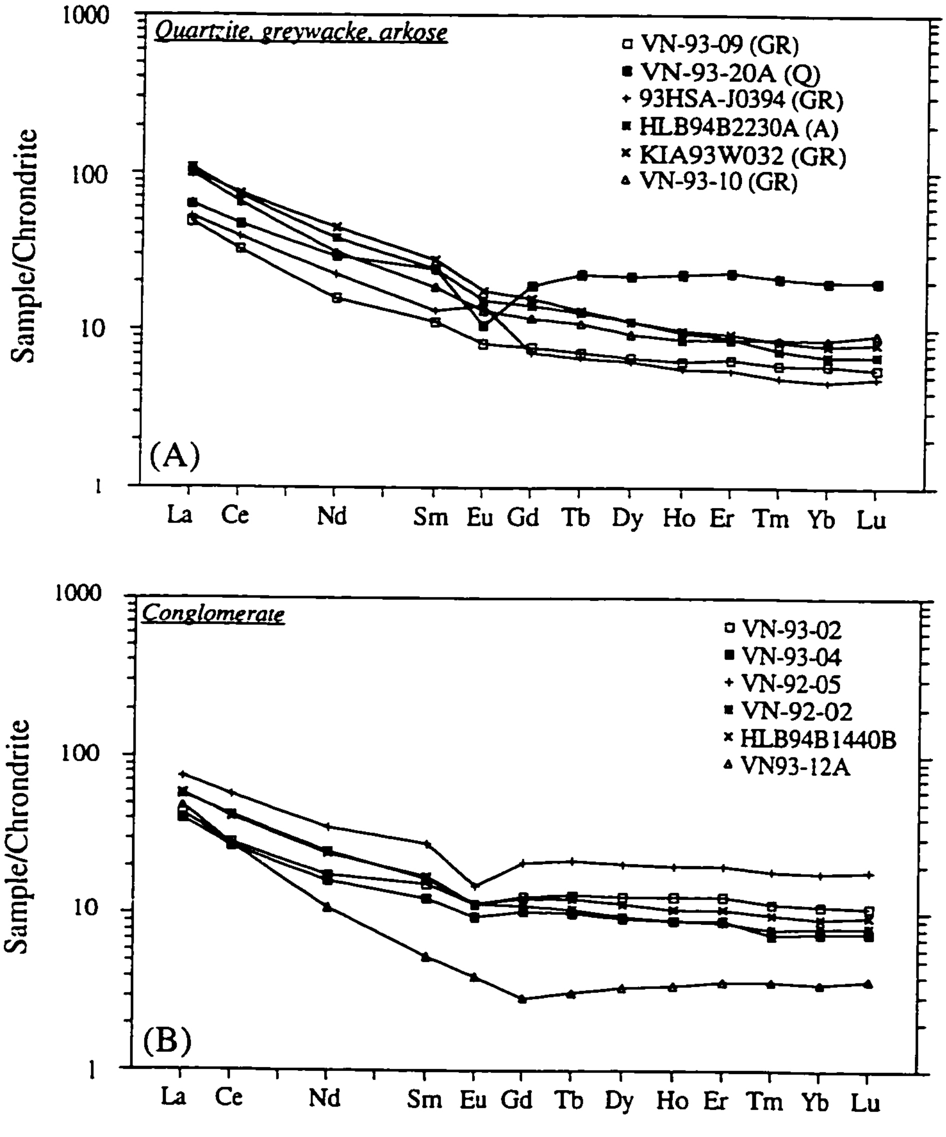


Fig. 5.2. Chondrite normalized rare earth element diagrams for the metasedimentary rocks used in this study. (A) Quartzite, greywackes and arkose. (B) Conglomerates. Values for the chondrite are taken from Boynton (1984). Abbreviations are Q, quartzite; A, arenite; GR, greywacke.

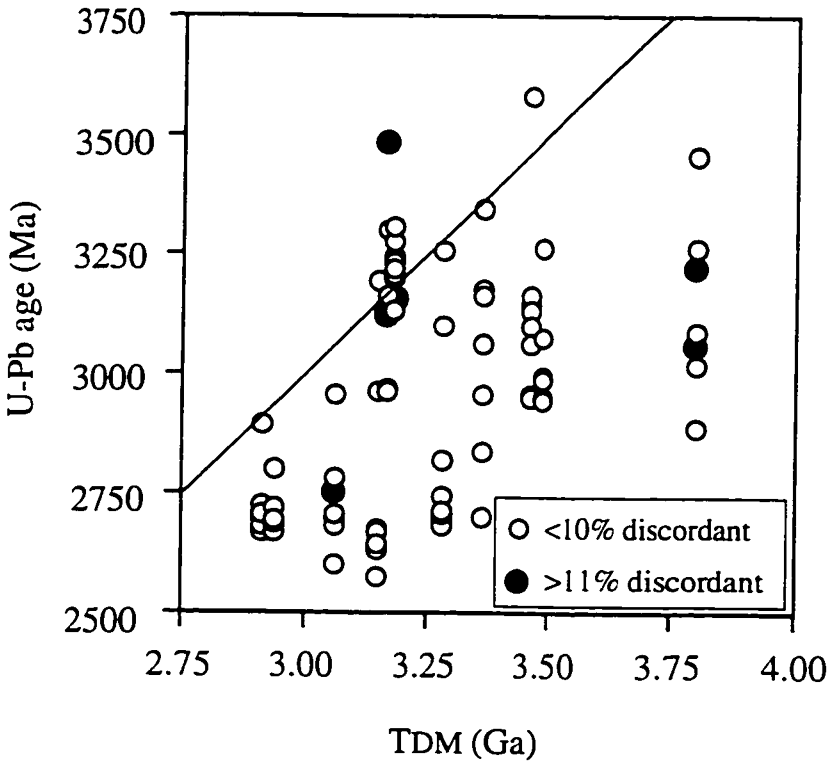


Fig. 5.3. Nd depleted mantle model age versus U-Pb detrital zircon age diagram for the metasedimentary rocks from the west-central Slave province. Also shown is the line that represents TDM=U-Pb age. TDM was calculated using model depleted mantle of Goldstein et al. (1984).

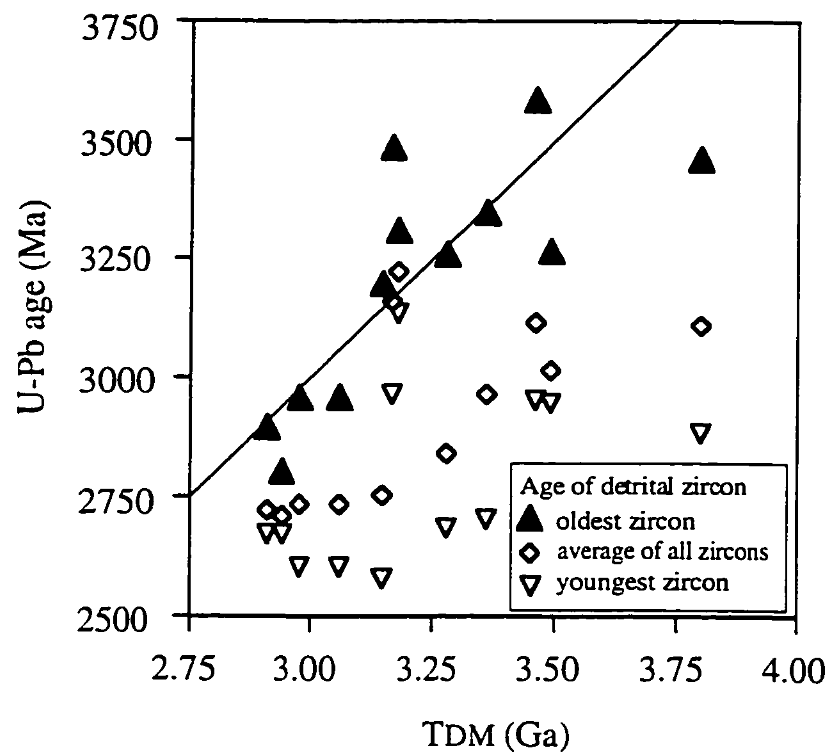


Fig. 5.4. Nd depleted mantle model age versus the oldest, youngest and the average U-Pb detrital zircon ages of each metasedimentary rock used in this study. Line represents TDM=U-Pb age. Model depleted mantle of Goldstein et al. (1984) was used to calculate the TDM.

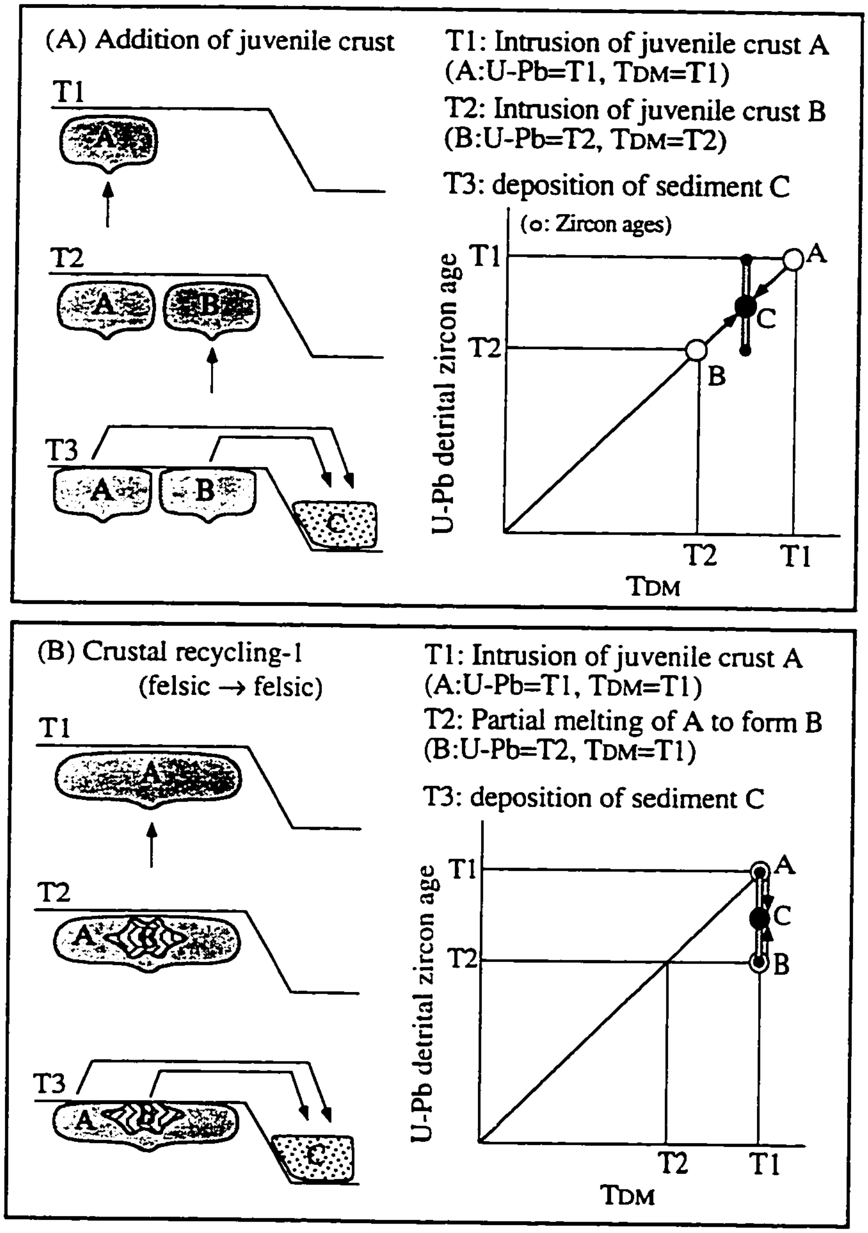
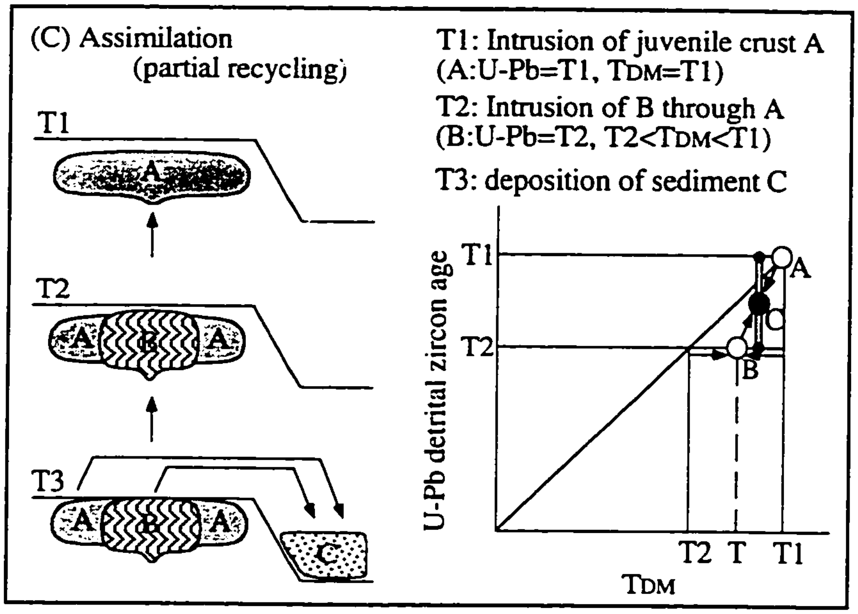
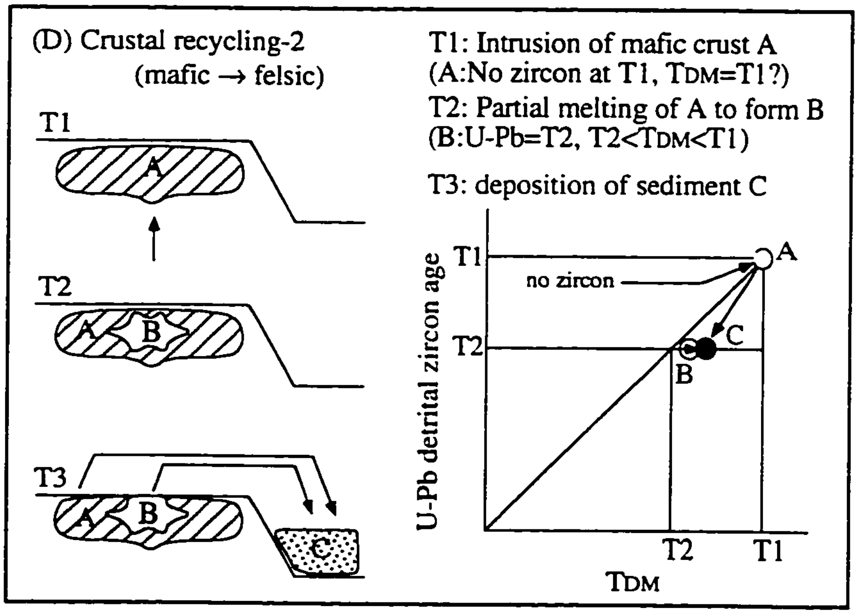


Fig. 5.5. Cartoon showing the effect of two component mixing on the TDM versus U-Pb age diagram. (A) Mixing between two juvenile crusts, (B) mixing between felsic juvenile crust and recycled crust, (C) mixing between juvenile crust and crustally contaminated juvenile crust and (D) mixing between mafic juvenile crust and felsic recycled crust (see text for detail).



(C) Assimilation
(partial recycling)
T1: Intrusion of juvenile crust A
(A:U-Pb=T1, TDM=T1)
T2: Intrusion of B through A
(B:U-Pb=T2, T2<TDM<T1)
T3: deposition of sediment C
T1
T2
T3
A
B
C
U-Pb detrital zircon age
T2 T T1
TDM



(D) Crustal recycling-2
(mafic → felsic)
T1: Intrusion of mafic crust A
(A:No zircon at T1, TDM=T1?)
T2: Partial melting of A to form B
(B:U-Pb=T2, T2<TDM<T1)
T3: deposition of sediment C
T1
T2
T3
A
B
C
no zircon
U-Pb detrital zircon age
T2 T1
TDM

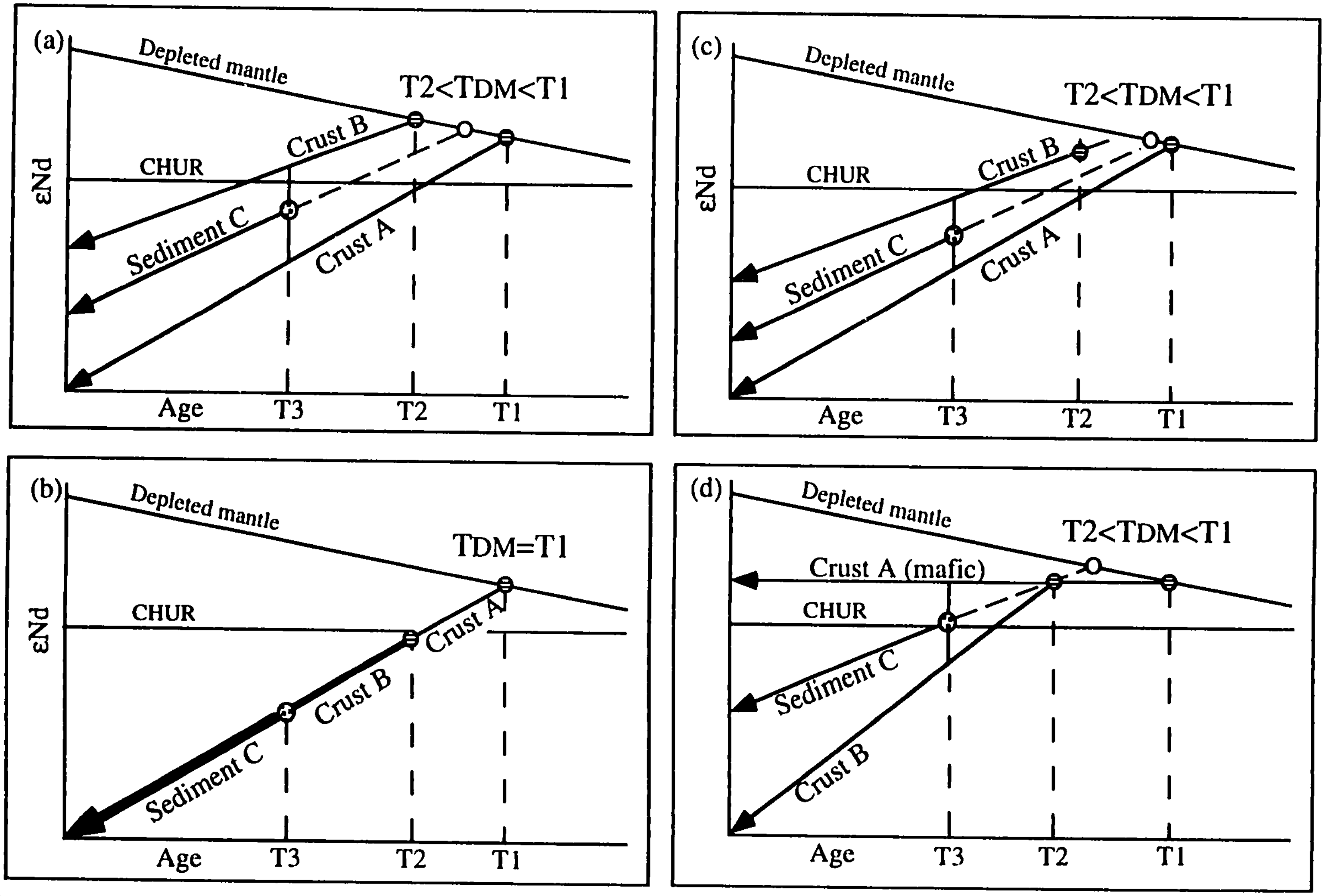


Fig. 5.6. Time versus εNd diagram for the four models described in figure 5.5. Models A-D correspond to models A-D in figure 5.5.

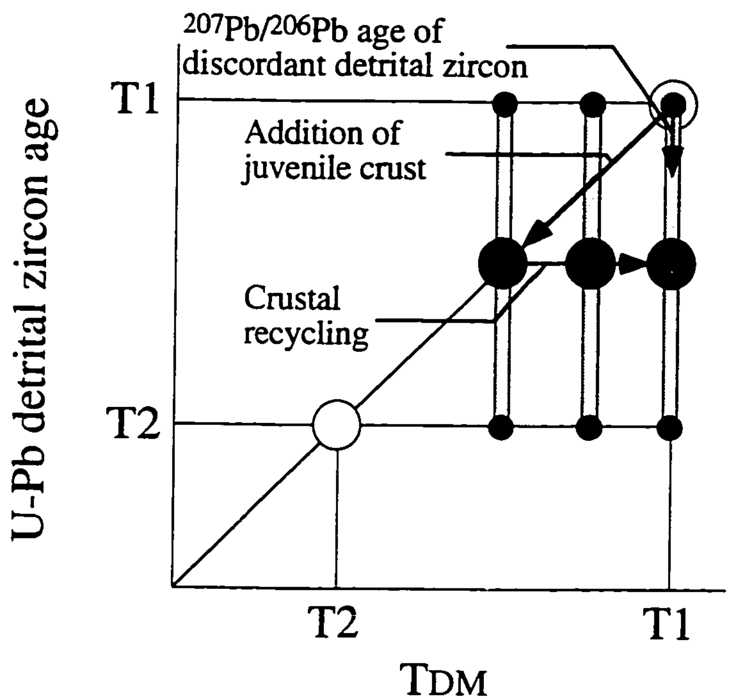


Fig. 5.7. Cartoon illustrating the effect of crustal recycling versus addition of juvenile crust on the TDM versus U-Pb age diagram

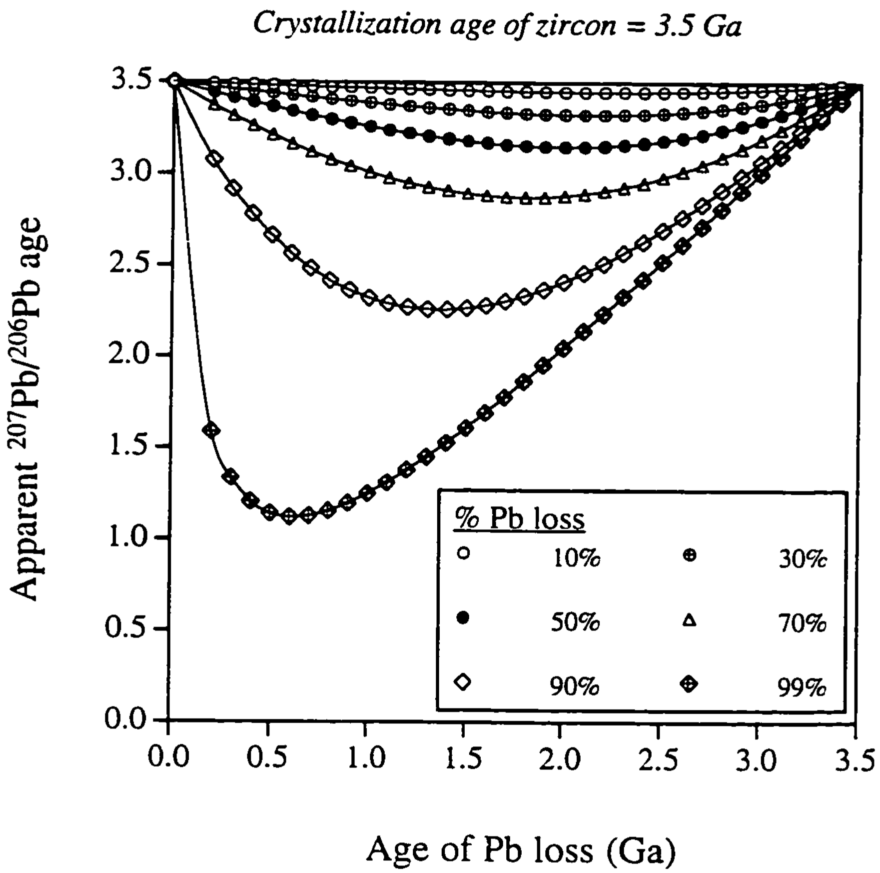


Fig. 5.8. Diagram showing the effect of Pb loss on the $^{207}Pb/^{206}Pb$ age of detrital zircon. Assumptions made are (1) crystallization age of zircon is 3.5 Ga and (2) no addition or loss of U.

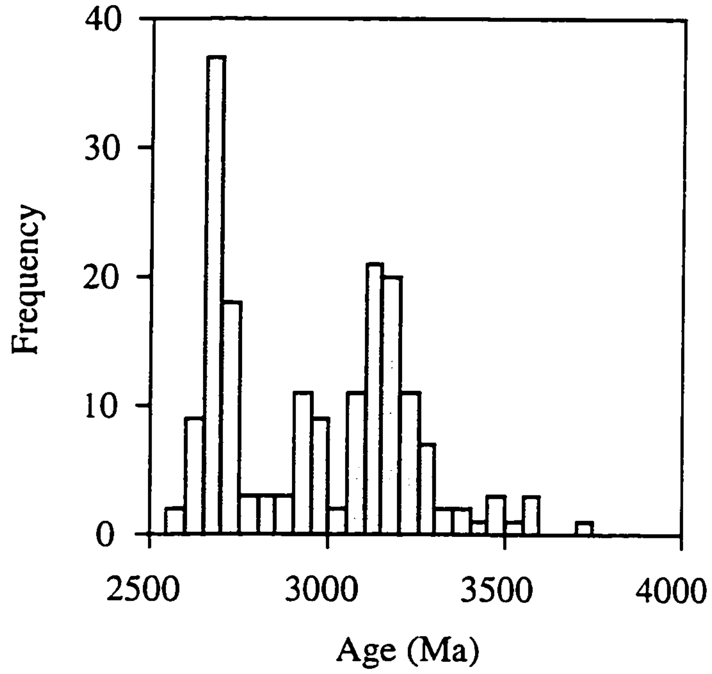


Fig. 5.9. Histogram showing the U-Pb detrital zircon ages from the Slave province. U-Pb data are taken from Schärer and Allègre (1982), Villeneuve and van Breemen (1994), Isachsen and Bowring (1997) and Villeneuve (1998).

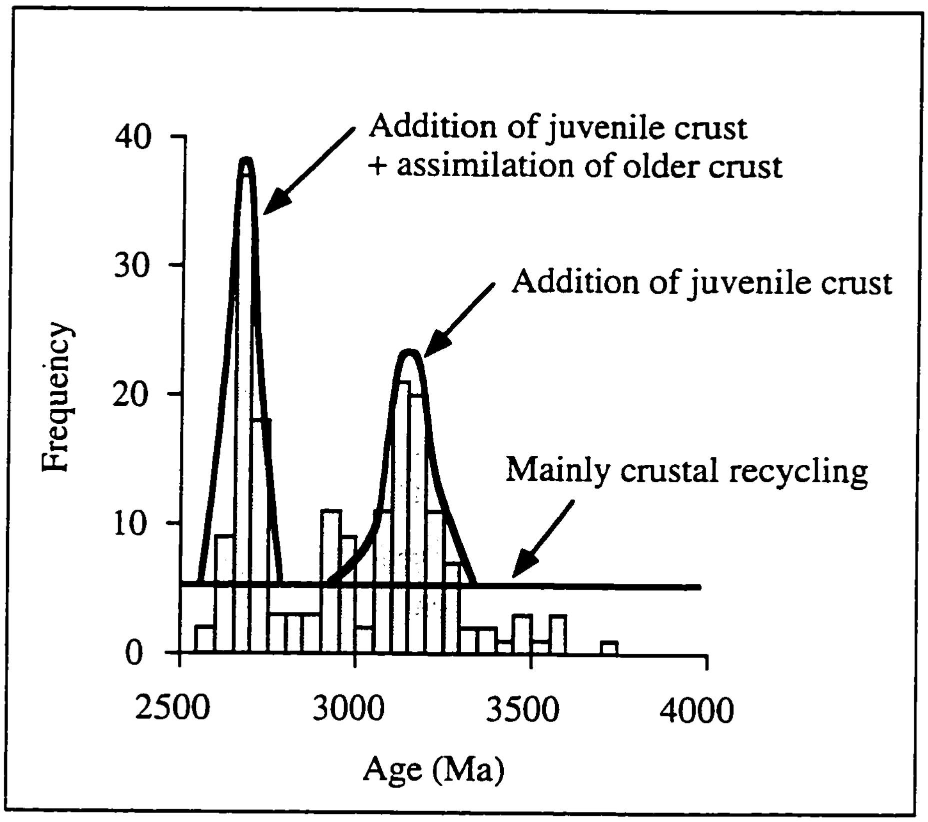


Fig. 5.10. Histogram showing the U-Pb detrital zircon ages from the Slave province and the possible mechanisms for the crustal evolution during the 3.1-3.4 Ga and 2.6-2.7 Ga intervals. U-Pb data are taken from Schärer and Allègre (1982), Villeneuve and van Breemen (1994), Isachsen and Bowring (1997) and Villeneuve (1998).

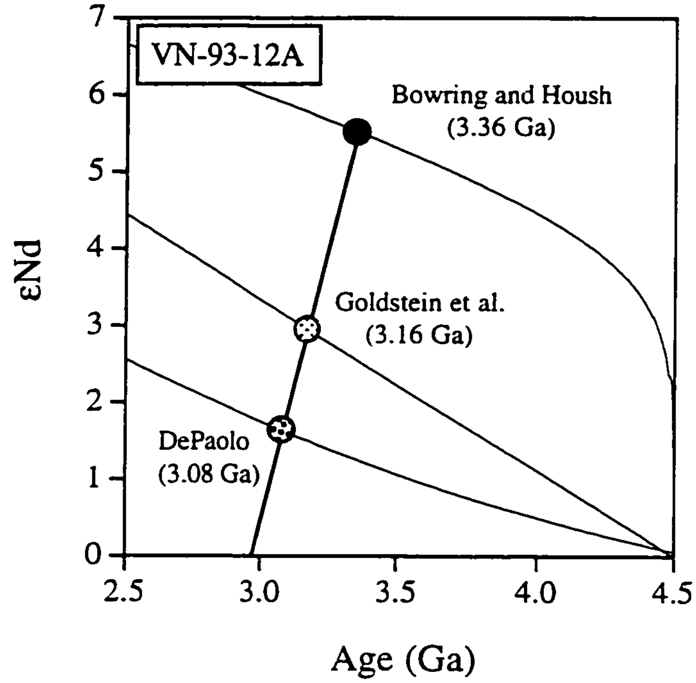


Fig. 5.11. Time versus εNd diagram illustrating the differences in the calculated T_{DM} using the models proposed by DePaolo (1981), Goldstein et al. (1984) and Bowring and Housh (1995). Sample VN-93-12A was used in this diagram as an example. The evolution curve for the depleted mantle proposed by Bowing and Housh (1995) was approximated using the equation $0.5x\ln(4.5-T)+(4.5-T)+4.3$; where T=age in Ga and T>4.5.

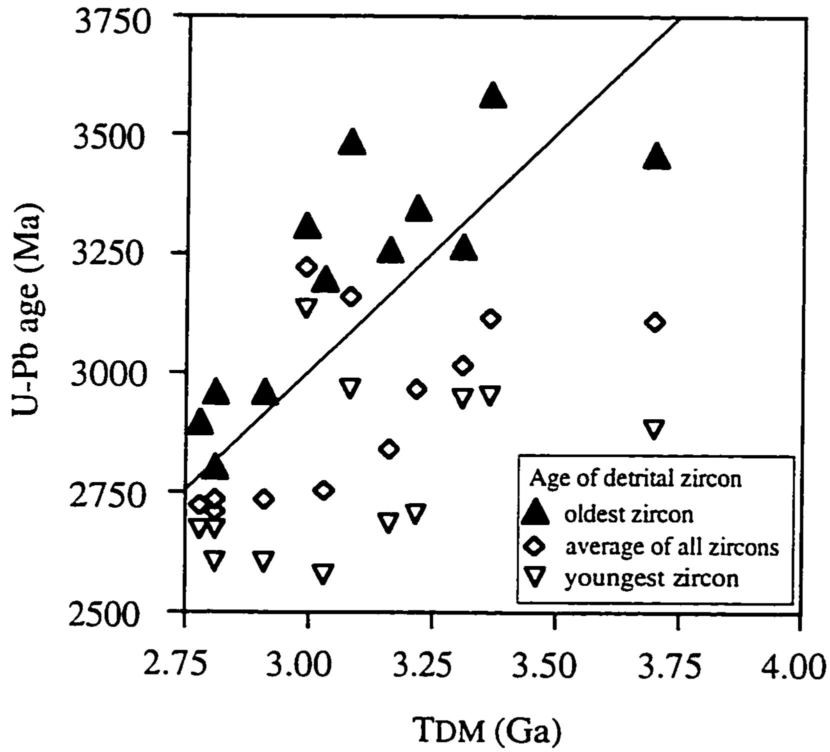


Fig. 5.12. Nd depleted mantle model age versus the oldest, youngest and the average U-Pb detrital zircon ages of each metasedimentary rock used in this study. Line represents TDM=U-Pb age. Model depleted mantle of DePaolo (1981) was used to calculate the TDM.

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Chapter 6

Concluding remarks

The Archean Slave province in the northwestern Canadian shield is generally classified into (1) pre-2.8 Ga basement, (2) 2.66-2.72 Ga supracrustal assemblage of the Yellowknife Supergroup (YKSG) with syn-volcanic plutons and (3) ~2.6 Ga late-Archean granitoids. Although this craton has received considerable attention among geologists since the discovery of the 4.0 Ga Acasta gneisses, detailed geochemical and isotopic studies involving all of these units had not been carried out prior to this study. In this dissertation, the geochemical and Nd-Pb isotopic features of igneous and sedimentary rocks from the Slave province were examined in detail to better understand its origin and evolution.

In chapter 2, the major/trace element abundances, Nd-Pb isotopic signatures and U-Pb ages of pre-YKSG basement tonalites, YKSG supracrustal rocks, syn-YKSG diorites and late-Archean granitoids from the Hanikahimajuk Lake area, northern Slave province, were determined. The age of pre-YKSG basement in the Hanikahimajuk Lake area was determined at 3378 Ma; this age coincides with the age of the metamorphic overprint of the Acasta gneisses, and the age of tonalitic gneisses in the northern-most Slave province (i.e. Kyagguyak gneisses). This similarity suggests that these pre-2.8 Ga crustal blocks were, at least at one stage, a single crustal block rather than fragments of unrelated crustal blocks that were later juxtaposed. The εNd_T of the basement tonalites range from -0.3 to 1.2, implying a role of >3.4 Ga crust in their genesis. The trace-element characteristics of these tonalites are most compatible with their derivation through partial melting of mafic protoliths at a pressure >0.8 GPa. The mafic YKSG volcanic rocks in the Hanikahimajuk Lake area can be classified into two groups based on their trace-element abundances; back-arc (or MORB) type and island-arc type. However, because the geochemical features of magmas produced in a modern back-arc setting are highly variable, it is likely that both of these rock types originated from the same back-arc system. The geochemical features of intermediate volcanic rocks and 2679 Ma syn-volcanic diorites are similar to those of the island-arc type mafic volcanic rocks, suggesting that they are fractionated derivatives of these rocks. The geochemical characteristics of felsic volcanic rocks and 2677 Ma felsic granite (Rim granite), on the other hand, are more compatible with their generation through partial melting of mafic protoliths at relatively low pressure, followed by assimilation of 3.4 Ga basement. The εNd_T of YKSG volcanic rocks and syn-volcanic diorites range from -2.2 to +2.5, indicating that these rocks were, at least in part, constructed on or in the vicinity of the 3378 Ma basement. Furthermore, the εNd_T of YKSG sedimentary rocks suggests that part of the basement was exposed at the time of ~2.66 Ga YKSG sediment deposition. Although no U-Pb ages for the late-Archean felsic granitoids in the Hanikahimajuk Lake area were determined, the major and trace element characteristics of these granitoids are similar to those of ~2.6 Ga syn- to post-deformation granitoids from central and southwestern Slave province. The $\varepsilon Nd(2.6)$ of these granitoids are broadly similar to the $\varepsilon Nd(2.6)$ of YKSG supracrustal rocks and syn-volcanic diorites, suggesting they are derived from a local source. This possibility was further examined in chapter 4.

In chapter 3, the major/trace element and Nd isotopic study of the ~2.66 Ga YKSG turbidites and volcanic rocks from the southern Slave province (near Yellowknife) was undertaken. Unlike most of the previous geochemical studies on clastic sedimentary rocks, different units of the Bouma sequence (or at least lower sand and upper shale units) were sampled separately. The εNd_T of the YKSG volcanic rocks ranges from -4.4 to +1.7. The low εNd_T of some samples clearly indicate that these volcanic rocks were, at least in part, erupted through the pre-YKSG basement. The εNd_T of turbidites ranges from -1.8 to +3.1 with TDM ranging from 2.8 to 3.2 Ga. There is no difference in the ranges of εNd_T between turbidites metamorphosed to greenschist facies and amphibolite facies. However, there is a general tendency for the upper shale units of the turbidites to have lower εNd_T values compared to the lower sand units. Through careful inspection of their trace element abundances, it is concluded that this shift in εNd_T is not an artifact of REE hosting heavy minerals concentrating in the lower sand units of turbidites but rather, is a result of "unmixing" of detritus derived from sources with slightly different average crustal residence ages. Previous petrographical and petrochemical study of these turbidites has shown that the source of these turbidites is a mixture of 20% mafic-intermediate volcanic rocks + 55% felsic volcanic rocks + 25% granitic rocks (Jenner et al. 1981). The geochemical data presented in chapter 3 are generally consistent with this model. However, by evaluation of the Nd isotopic signatures, it is further revealed that these turbidites are a mixture of YKSG volcanic rocks which were variably contaminated by pre-YKSG basement and quartzofeldspathic (granitic) rocks that were less contaminated. The proportion of detritus from the crustally contaminated YKSG volcanic rocks is generally higher in the upper shale units. Because these sediments are thought to have derived from the areas near or west of Yellowknife, existence of detritus derived from less contaminated quartzofeldspathic rocks would indicate that there are areas near and/or west of Yellowknife that are not underlain by pre-YKSG basement.

In chapter 4, detailed Nd-Pb isotopic and petrochemical study of 2.58-2.62 Ga syn- to post-deformation granitoids from the southwestern Slave province was undertaken. By comparing the trace-element data with recent results of high-pressure experiments, it is concluded that most of the syn-deformation granitoids in the SW Slave province are compatible with an origin through partial melting of mafic protoliths. However, study of syn-deformation granitoids from the central Slave province has also shown that some of the syn-volcanic granitoids (e.g. mafic granitoids) are mantle derived (Davis et al. 1994). Thus, the simplest explanation for the origin of syn-deformation granitoids is the combination of these two mechanisms. The ~2.59 Ga post-deformation granitoid, on the other hand, are derived mainly through partial melting of juvenile greywacke and pelites although tonalitic rocks may have also been an important source. Through Pb isotopic modeling, it is shown that the sources for both syn- and post-deformation granitoids must include some component from the pre-YKSG basement, and the ultimate age (or the mantle extraction age) of the pre-YKSG component must exceed 3.2 Ga (more likely ~3.4 to 3.8 Ga). The incorporation of a pre-YKSG signature can take place through (1) assimilation of pre-YKSG basement by YKSG volcanic rocks and (2) through direct input of detritus from the basement during sediment deposition. Furthermore, a combined Nd-Pb isotopic modeling has shown that granitoids from the southwestern Slave province contains 10-

30% basement component whereas the granitoids from Point Lake and Contwoyto Lake areas may contain >50%.

In chapter 5, the relationship between the U-Pb detrital zircon ages and Nd depleted mantle model ages of sedimentary rocks collected from various locations of the western Slave province was evaluated. This relationship was then applied to investigate the early- to mid-Archean crustal evolution of the Slave province. Through simple modeling, it is shown that the crustal evolution of the pre-2.8 Ga Slave province was dominated by crustal recycling (including assimilation and partial melting of pre-existing crust) rather than addition of juvenile materials from the mantle. A notable exception to this is the period between ~3.1 to 3.3 Ga, when the addition of juvenile materials was large enough to buffer the isotopic signature of the basement. The role of crustal recycling (especially crustal assimilation) seems to have increased in the post-2.8 Ga history of the Slave province. This is consistent with the presence of a basement-like signature in most if not all of the YKSG supracrustal rocks and late-Archean granitoids studied in chapters 2 to 4.

Finally, it is emphasized that through this dissertation, I have shown that none of the existing models successfully explain the entire tectonic history of the Slave province. Clearly, a more complicated or perhaps a completely new model is required. Shown in Figure 6.2 is a model that may explain the tectonic history of the Slave province. This preliminary model was constructed to explain the following specific results of this dissertation.

(1) The generation of ~3.4 Ga basement in the northern Slave province is related to the tectonomagmatic event which disturbed the U-Pb systematics of the ~4.0 Ga Acasta gneisses.

(2) Two types of mafic volcanic rocks found in the Hanikahimajuk Lake area (i.e. back-arc type and island-arc type) can both be generated in a same ensialic back-arc system.

(3) These two types of mafic volcanic rocks are also found in the Yellowknife area (southern Slave province) and Indin Lake area (western-most Slave province, Fig 6.1; Pehrsson, pers. comm.).

(4) The mafic volcanic rocks from the Hanikahimajuk Lake and Yellowknife areas are clearly contaminated by pre-2.8 Ga basement (no Nd isotopic data is available from the Indin Lake volcanic belt).

(5) The Nd isotopic systematics of the ~2.66 Ga turbidites in the Yellowknife area suggest that there are areas west of Yellowknife not underlain by the pre-2.8 Ga basement.

(6) The Nd-Pb isotopic signature of syn- to post-deformation granitoids show that the strongest basement-like signatures are found in two areas of the Slave province; (*i*) along the volcanic belt that runs along the central Slave province from Anialik River - Hanikahimajuk Lake - Point Lake - Winter Lake - Sleepy Dragon Complex - Yellowknife, and (*ii*) in the Acasta River and Grenville Lake area.

(7) Most of the 2.60 - 2.62 Ga syn-deformation granitoids are generated through partial melting of mafic protoliths whereas 2.58 - 2.60 Ga post-deformation granitoids are produced through partial melting of greywacke to pelite (and possibly tonalitic rocks).

(8) Major crust forming events in the Slave province took place between 3.3-3.1 Ga and 2.7-2.6 Ga.

Furthermore, the present model takes into account the following points made by previous researchers.

(1) The volcanic belt in the east-northeastern Slave province is felsic dominated while the volcanic belt in the central-western Slave province is mafic dominated (Padgham 1985).

(2) There is a gap in igneous activity between 2625-2640 Ma (Villeneuve et al. 1997).

(3) The existence of crust older than 2.8 Ga has not been documented in the eastern Slave province (Thorpe et al 1992, Davis and Hegner 1992, Davis et al. 1996).

(4) The regional metamorphism and deformation in the Slave province took place between 2.63-2.58 Ga (van Breemen et al. 1992).

In the present model, the consolidation of the Slave craton begins with the generation of Acasta gneisses at ~4.0 Ga (stage 1). A major tectonomagmatic event which disturbed the U-Pb systematics of the Acasta gneisses produces the 3.4 Ga tonalites (and tonalitic gneisses) in the northern-central Slave province. The tectonic setting in which these TTGs (tonalite-trondhjemite-granodiorite) were produced remains unconstrained. However, the continuous northwest to southeast younging generation of TTGs and felsic volcanic rocks from the Acasta River (~4.0 Ga) to Anialik River/Hanikahimajuk Lake area (~3.4 Ga) to Point Lake/Winter Lake area (~3.3-3.1 Ga) to Sleepy Dragon/Yellowknife area (~2.9 Ga; Bleeker and Stern 1997 and references therein) may hint towards the possibility of north-northwestward subduction of mafic oceanic crust (Stage 2).

After a "brief" period of quiescence, a west verging subduction of oceanic crust begins at ~2.72 Ga. This subduction created a felsic dominated fore-arc in the eastern Slave province and mafic dominated back-arc in the west-central Slave province (Stage 3). Although the exact geometry of this island-arc is not known, it may have been similar to the present-day Japanese tectonic setting; an intermediate-felsic dominated fore-arc and mafic-dominated back-arc system, with a remnant of continental block(s) (in part Precambrian) in the back-arc side and further west in the Korean peninsular (e.g. Lee 1987, Kimura et al. 1991). It is also noteworthy that the size of the present day Japanese arc plus the sea of Japan is broadly similar or larger than the entire Slave province.

At ~2.64 Ga, a collision of hypothetical crust in the eastern-most Slave province terminates the YKSG volcanic and plutonic activity in the Slave province (Stage 4). Whether this hypothetical crust still remains in the eastern Slave province is not known. However, the collision of such hypothetical crust will explain the time gap in igneous activity between syn-volcanic plutons and subsequent syn- to post-deformation magmatism related to closure of the back-arc basin and crustal thickening. Following the collision of hypothetical crust, regional deformation and metamorphism lead to the generation of ~2.61 Ga metaluminous to weakly peraluminous granitoids, mostly through partial melting of pre-existing mafic crust but also through addition of mantle-derived magma (Stage 5). Finally, at ~2.59 Ga, peraluminous post-deformation granitoids are produced through partial melting of sedimentary (and tonalitic) rocks, which are also widely available in the Slave province (Stage 6). The pronounced shift in the geochemistry of granitoids from metaluminous to peraluminous is in accord with the general stratigraphy of a back-arc system, that is mafic igneous rocks overlain by sedimentary rocks such as greywacke and pelites. If crustal thickening leads to a production of syn- to post-deformation granitoids, the mafic rocks on the lower part of the stratigraphy will melt first and this will be followed by a melting of sediments on the top. The existence of volcanic rocks overlain by sedimentary rocks in the Yellowknife area of the Slave province is supported by a recent seismic profile (Clowes 1997).

As previously mentioned, this model is still preliminary and somewhat speculative. However, it is compatible with many geochemical and Nd-Pb isotopic features seen in rocks studied during this dissertation as well as the regional geological features documented in the literature. Perhaps the most important work that needs to be done is to look at the structure of the YKSG supracrustal rocks in the eastern Slave province because if this model is correct, we should find some structures that resemble a modern accretionary prism. Furthermore, major/trace element geochemical and Nd isotopic characteristics of the YKSG volcanic rocks from the eastern Slave province must be investigated to see if these rocks have the geochemical features compatible with their origin in an island-arc setting.

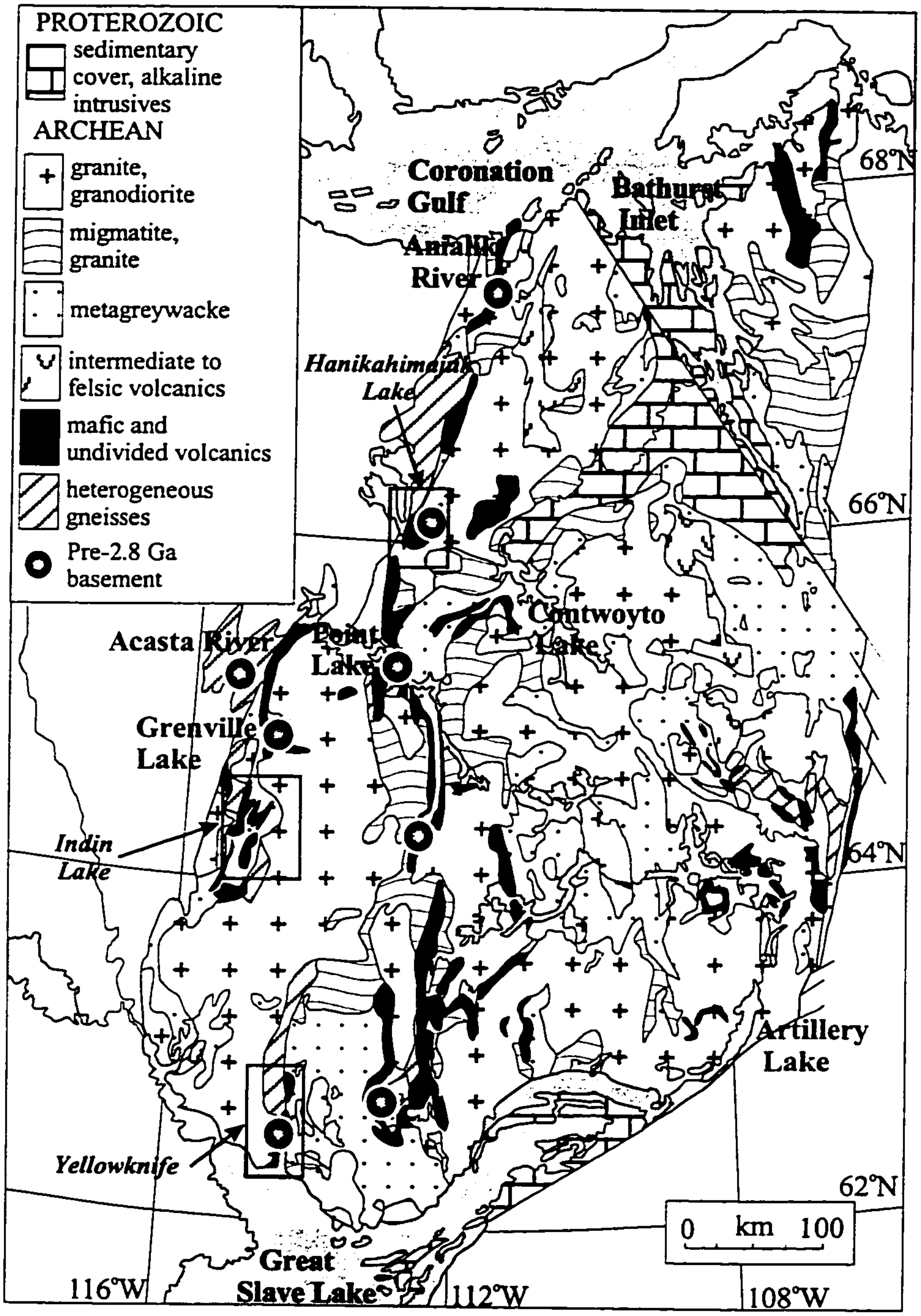
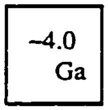
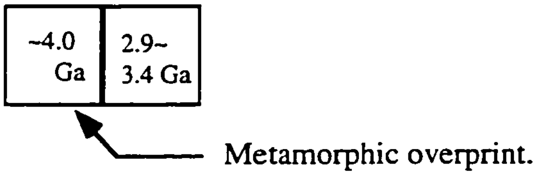


Fig. 6.1. Geological map of the Slave province (after Hoffman 1989).

(Stage 1) ~**4.0 Ga** Generation of the Acasta gneisses (unknown tectonic setting).



(Stage 2) ~**2.9-3.4 Ga** Generation of basement in the central Slave province (unknown tectonic setting).



(Stage 3) ~**2.7 Ga** Developement of fore-arc in the eastern Slave province and ensialic back-arc in the western Slave province.

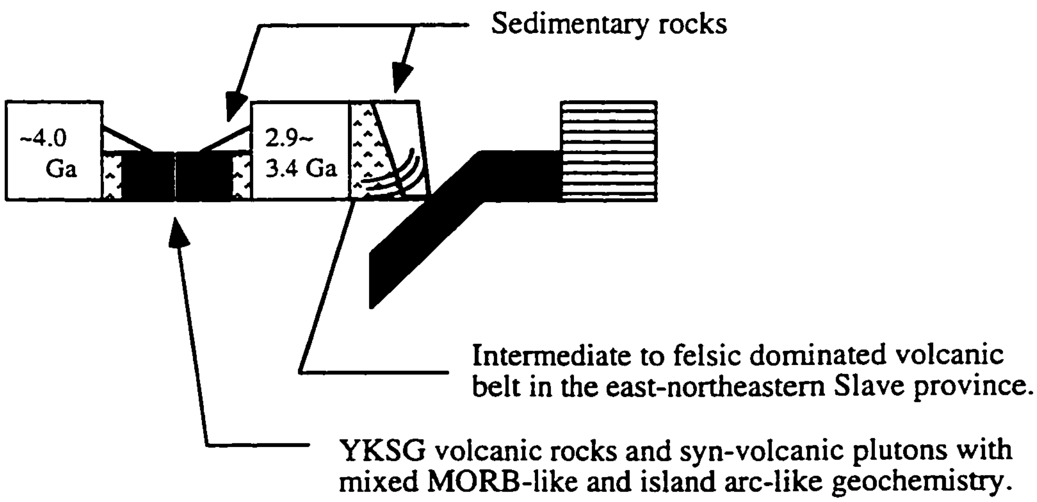
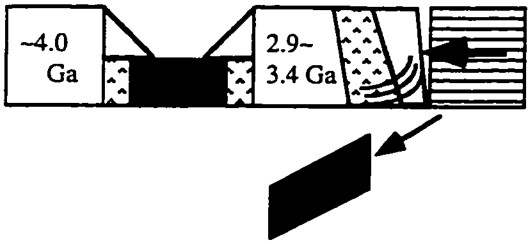
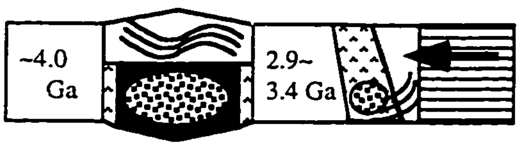


Fig. 6.2. Cartoon illustrating the tectonic evolution of the Slave province.

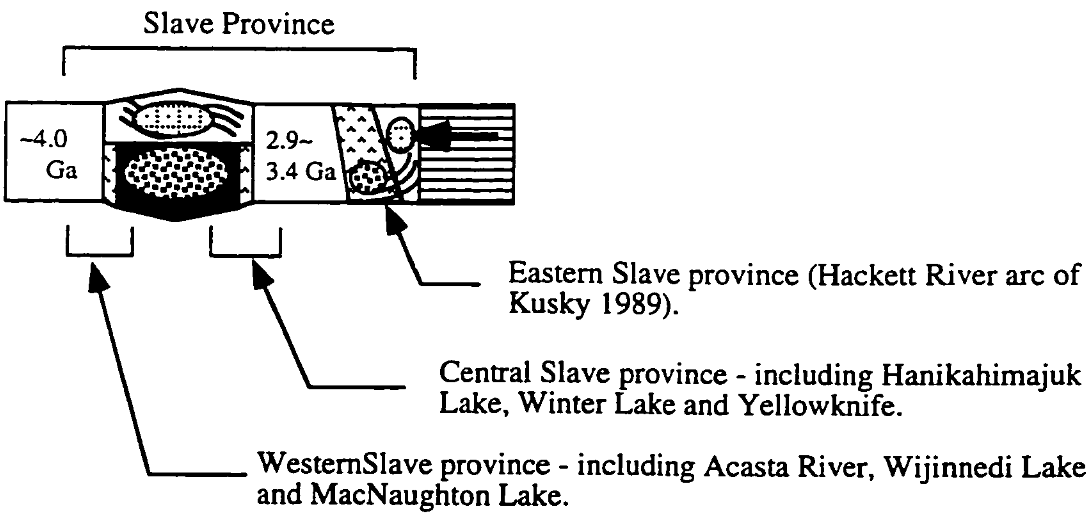
(Stage 4) ~**2.64 Ga** Collision of hypothetical crust to terminate the YKSG volcanic activity.



(Stage 5) ~**2.62 Ga** Crustal thickening accompanied by melting of mafic rocks to produce syn-deformation granitoids.



(Stage 6) ~**2.59 Ga** Partial melting of sedimentary rocks to produce post-deformation granitoids.



Note: The size of each crustal block and the orientaion of different rock types within the two granite-greenstone blocks are not representative of their actual size/orientation.

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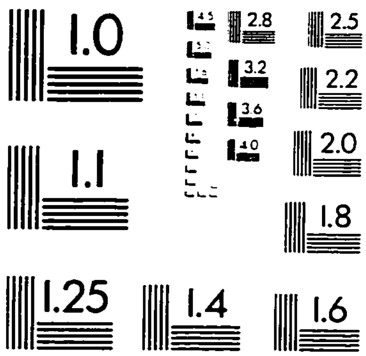
Mineral Development Subsidiary Agreement. Geological Survey of Canada, Open file 2484: pp. 179-184.

IMAGE EVALUATION TEST TARGET (QA-3)



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