GEOLOGICAL SURVEY OF CANADA
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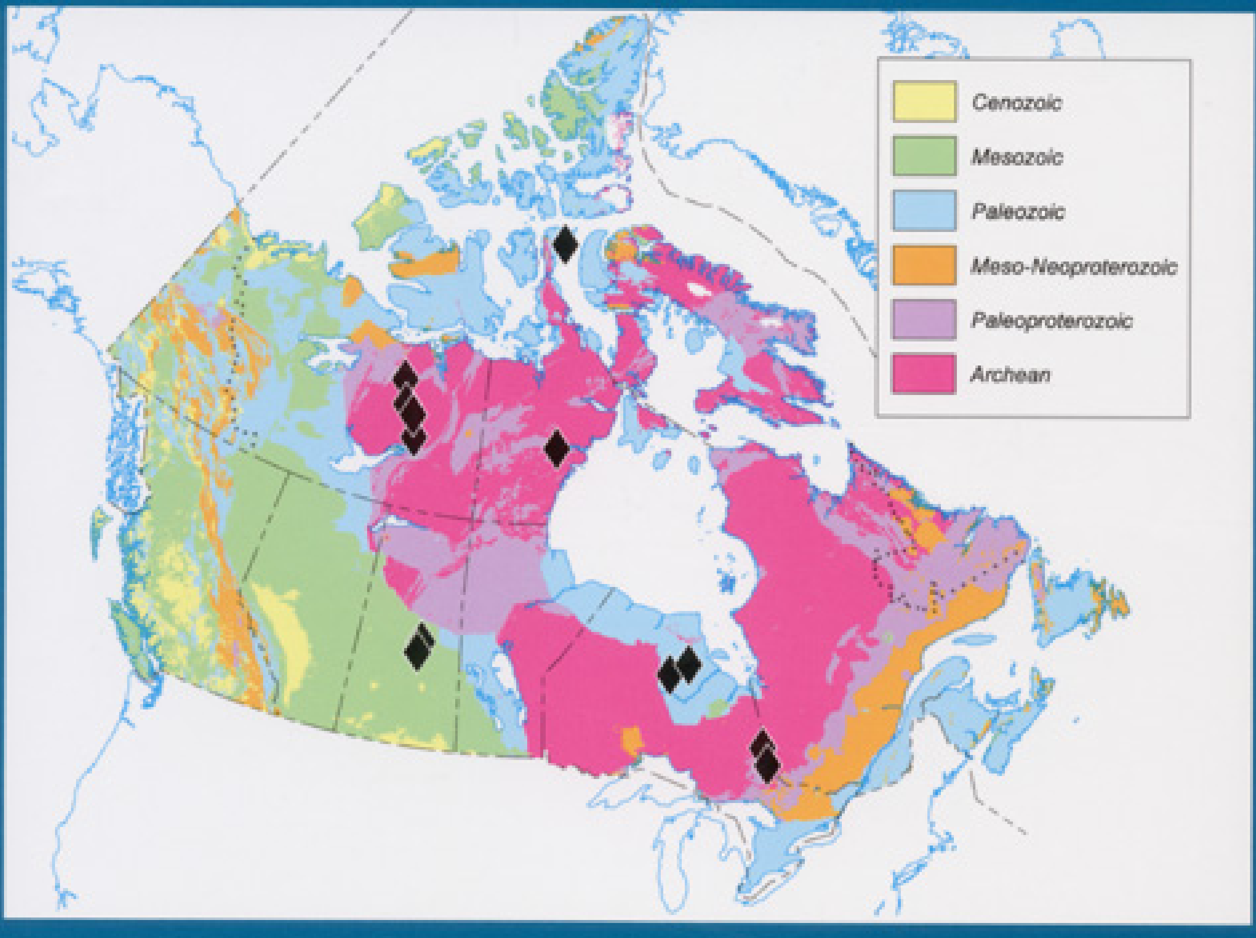
# SEARCHING FOR DIAMONDS IN CANADA

Edited by
A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson

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Canada

NATMAP CARTNAT

Contribution to Canada's Slave Province National Geoscience Mapping Program/ Contribution au projet de la Province des Esclaves du la programme national de cartographie géoscientifique du Canada

**Cover Description:**

Black diamonds indicate the locations of primary diamond-bearing rocks in Canada plotted on a simplified bedrock geological map derived from the Geological Map of Canada (Wheeler et al., in press). Primary diamond-bearing rocks occur where deep-seated magmas rose from depths >150 km through the old stable nucleus of the North American continent. Occurrences are in small intrusions within Archean or Paleoproterozoic rocks of the Canadian Shield, or in similar intrusions in adjacent cover sequences underlain by Archean or Paleoproterozoic basement.

**Wheeler, J.O., Hoffman, P.F., Card, K.D., Davidson, A., Okulitch, A.V., Sanford, B.V., and Roest, W. (comp.)** in press: Geological Map of Canada; Geological Survey of Canada, Map 1860A, 1:5 000 000 scale map.

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**1996**

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## PART 3: GEOPHYSICAL EXPLORATION AND GEOGRAPHIC INFORMATION SYSTEM (GIS) APPLICATIONS

# SEARCHING FOR DIAMONDS IN CANADA

## Foreword

**A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson**

Diamonds! "The Great Canadian Diamond Rush" north of Yellowknife (McNellis, 1993) revives for this generation a sense of the excitement and dreams of the Klondikers of the last century. Until recently, most Canadians thought of diamonds only as exotic and treasured jewels, appreciated for their rarity and brilliance, but of little direct economic interest to Canada. Prospective areas of the Canadian Shield were largely ignored, even though for almost thirty years geologists have known that diamond deposits are closely associated with the old stable nuclei of continents (cratons). Diamonds originate in the Earth's mantle at depths >150 km and most are stored in distinctive source rocks that make up part of the stable mantle root beneath Archean (>2500 million years old) and Proterozoic (2500 to 570 million years old) cratons. The two most important diamond source rocks are peridotite and eclogite, and each rock type contains a characteristic suite of minerals that are key indicators for diamond exploration. Primary diamond deposits occur where kimberlite and lamproite magmas erupted, since these deep-seated magmas provide a medium to sample diamond-bearing source rocks and transport diamonds and associated indicator minerals to surface. Economic diamond-producing fields occur on most Archean cratons worldwide, with the notable exception of Archean cratons in Canada, such as the Superior, Slave and Nain provinces. However, intense exploration activity throughout Canada, since the late 1980s, has located numerous diamond-bearing kimberlites in the Slave Province near Lac de Gras, north of Yellowknife, and additional kimberlites have been discovered in Alberta, Saskatchewan, Manitoba, Ontario, and Quebec.

Surprisingly, before the 1990s, diamonds were almost absent from Canadian folklore and mineral history. Jacques Cartier's men mined "diamonds" at the mouth of Rivière du Cap-Rouge in 1541, but their treasure turned out to be worthless quartz. This episode gave Quebec's Cap Diamant its name, and the story survives in the saying "faux comme des diamants du Canada". Early this century, reports by officers of the Geological Survey of Canada suggested microdiamonds were recovered from chromitite lenses in the Tulameen complex, British Columbia (Camsell, 1911) and from chromite ore mined at Black Lake, Quebec (Dresser, 1913). Although the Tulameen "microdiamonds" were later shown to be synthetic periclase formed by laboratory heating of the rock samples, recent work in Morocco, Spain, and Tibet has documented the association of diamonds with similar tectonically emplaced ultramafic rocks (Davies at al., 1993; Bai et al., 1993).

J.J. Brummer (1978) culled meagre data from a wide range of sources to provide a remarkably comprehensive overview of the early history of "Diamonds in Canada". He noted that W.H. Hobbs (1899) first raised the possibility of diamond sources in Canada, based on discoveries of diamonds in glacial drift south of the Great Lakes. Although a 33 carat alluvial diamond was discovered near Peterborough, Ontario before 1920, and some finds were reported in Saskatchewan and Quebec in the late 1940s and early 1950s, significant diamond exploration did not begin in Canada until the 1960s, when indicator mineral surveys were conducted in Ontario by mining companies, the Ontario Department of Mines and the Geological Survey of Canada. Satterly (1949) recognized the first Canadian kimberlites in Michaud Township, north of Kirkland Lake and, by the late 1960s, several other kimberlites and a few diamonds had been discovered. The history of recent diamond discoveries in Canada has yet to be written, but many of the most enthusiastic explorationists, active over more than 25 years, now find themselves drawn to the Barrenlands near Lac de Gras, hoping to be among the first to bring Canadian diamonds to world markets.

Discovery of world-class diamond deposits depends on determined mineral exploration aided by a reliable and comprehensive geological database. Papers in this volume provide a snapshot of the spectrum of geoscience information available to assist diamond exploration in Canada. Maps provide elegant and ready access to data acquired and interpreted by the Geological Survey of Canada (GSC), Provincial Surveys, and University-based researchers. Mapping by the GSC now integrates traditional geological, geophysical, geochemical, and surficial surveys with specialized Geographic Information System (GIS) techniques, several of which have important applications to diamond exploration. This volume provides background for several of the national databases maintained by the GSC, as well as summaries of specific areas of diamond-related research and short reviews of GSC research relevant to diamond exploration. Readers seeking further information are encouraged to contact the authors, whose addresses are listed at the end of the volume. In addition, the GSC has published a bulletin that reviews the use of various indicator minerals and mineral assemblages as important aids in diamond exploration in Canada (Fipke et al., 1995).

Reports contained in this volume were submitted during the period December 1994-June 1995, and have been reviewed by GSC staff, but have not undergone rigorous scientific review. Thanks are extended to the many scientists who contributed to this volume, and to O.J. Ijewliw, R. Lacroix, D. Paul, S. Scully, M. Sigouin, K. Venance and T. West, all of the GSC, for assistance with figure production. W.C. Morgan undertook the technical editing. Preliminary corrections, compilation and layout were completed by A. Anand with assistance from N. Devine, C. Bélanger, L. O'Neill and C. Plant (all of the GSC). Printing of this volume was funded by the GSC's Mineral Resources and Continental Geoscience divisions.

Sadly, two scientist who contributed to this volume died in 1995. On February 23, Chris Roddick died in a skiing accident in Vermont, tragically cutting short a scientific career in isotope geoscience characterized by imagination, enthusiasm, and curiosity. His contribution to the Geochronology Laboratory of the GSC is commemorated in the introduction to the 1995 Radiogenic Age and Isotope Studies report (Parrish, 1996). Chris leaves a rich legacy and is deeply missed. Marianne Mareschal, a leading scientist in Canada's LITHOPROBE project, passed away on July 11, 1995, after a long and courageous struggle with cancer. She was a member of the organising committee for the Precambrian'95 conference in Montréal and a dedication to her is published in the Program and Abstracts volume for Precambrian'95. Her kindness and energy touched all whom she met, and she will be missed by many. Most of all, her direction and vision in combining seismic and electromagnetic experiments for the study of cratonic roots, which was her last major research effort, will bear fruit for many years to come. A foundation, established in her name, will be used to provide a student bursary in geophysics at the Ecole Polytechnique de Montréal. Contributions should be forwarded to: Fonds Marianne Mareschal, Department de genie minéral, Ecole Polytechnique de Montréal, Montréal, CP 6079, Succ"centre ville", Montréal H3C 3A7, Canada.

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# PART 1: GEOLOGY, PETROLOGY, AND GEOTECTONIC CONTROLS

## Introduction

**A.N. LeCheminant and B.A. Kjarsgaard**

*LeCheminant, A.N. and Kjarsgaard, B.A., 1996: Part 1: Geology, petrology, and geotectonic controls - Introduction; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 5-9*

Diamonds, though extremely rare, are widely distributed and have been discovered in unconsolidated and consolidated sediments, diverse igneous rocks of upper mantle origin, mantle xenoliths, ophiolites, ultra high pressure metamorphic rocks, meteorites and impact structures. Of these, only diamond-bearing kimberlite and lamproite, and their derived placer and paleoplacer deposits, have proven to be economically viable. Before 1960, >80% of all diamonds were recovered from secondary deposits; by 1990, increased diamond production from kimberlite and lamproite pipes reduced this to <25% (Levinson et al., 1992). Nearly two thirds of present world production by weight comes from only five pipes located in Australia, Botswana, Russia, and South Africa (Boucher, 1995).

Kimberlites occur in continental shield areas, and economic kimberlites are associated with the stable parts of Archean (>2.5 Ga) cratons. Lamproites, on the other hand, typically intrude remobilized Archean cratons or Proterozoic orogenic belts. For example, the Argyle deposit in Australia, the world's most productive diamond mine, occurs in a lamproite pipe within the Proterozoic Halls Creek mobile belt. Locations of diamond-bearing rocks in Canada are shown in Figure 1, along with the locations of many commercially important kimberlites and lamproites. This figure also shows the worldwide distribution of Archean and Proterozoic cratons based on a generalized geological map of the world, now available digitally (Kirkham et al., 1994, 1995).

In Canada, Archean and Proterozoic structural provinces are well known and geological maps provide a rigorous framework for examining the distribution of kimberlites and for testing models of diamond formation. Precambrian rocks of the North American craton have been subdivided into numerous crustal domains with distinctive ages, structural trends, and geophysical characteristics (Stockwell, 1961; Hoffman, 1989). Archean cratons, which underlie a large part of the exposed Canadian Shield, were welded together by Proterozoic orogens, many of which continue in the sub-surface beneath adjacent Phanerozoic sedimentary cover (e.g. Trans-Hudson orogen - Fig. 1; Percival, 1996). The contrast between exposed Archean shields and adjacent thinly covered Proterozoic orogens is one characteristic that hints at significant differences between Archean and Proterozoic processes of lithosphere formation and preservation (Durrheim and Mooney, 1994). The continental lithosphere, made up of the crust and part of the upper mantle, is a strong persistent layer relative to the underlying convecting asthenosphere. Seismological studies indicate that the lithosphere is thicker beneath Archean cratons, and contains regions of cool lithospheric mantle within the diamond stability field (i.e. depths >150 km; Durrheim and Mooney, 1994; Grand, 1994; Polet and Anderson, 1995). The first three papers in this volume provide a brief summary of the complex history of Archean cratons and flanking Proterozoic orogens in Canada, and contain key references to maps and reports on the Precambrian geology of Canada.

In this volume, the geology of kimberlites and lamproites is covered in summary papers, accompanied by short papers on analytical methods and specific kimberlite fields in Canada. For comprehensive summaries of the petrology of kimberlites, lamproites, and lamprophyres the reader is referred to books by Mitchell (1986; 1995), Mitchell and Bergman (1991) and Rock (1991). Extensive additional information can be obtained from the Proceedings Volumes for the five International Kimberlite Conferences published up to 1994, and in the Abstracts Volume for the Sixth International Kimberlite Conference (*Proceeding Volumes, 1979-1994; Abstracts Volume, 1995).

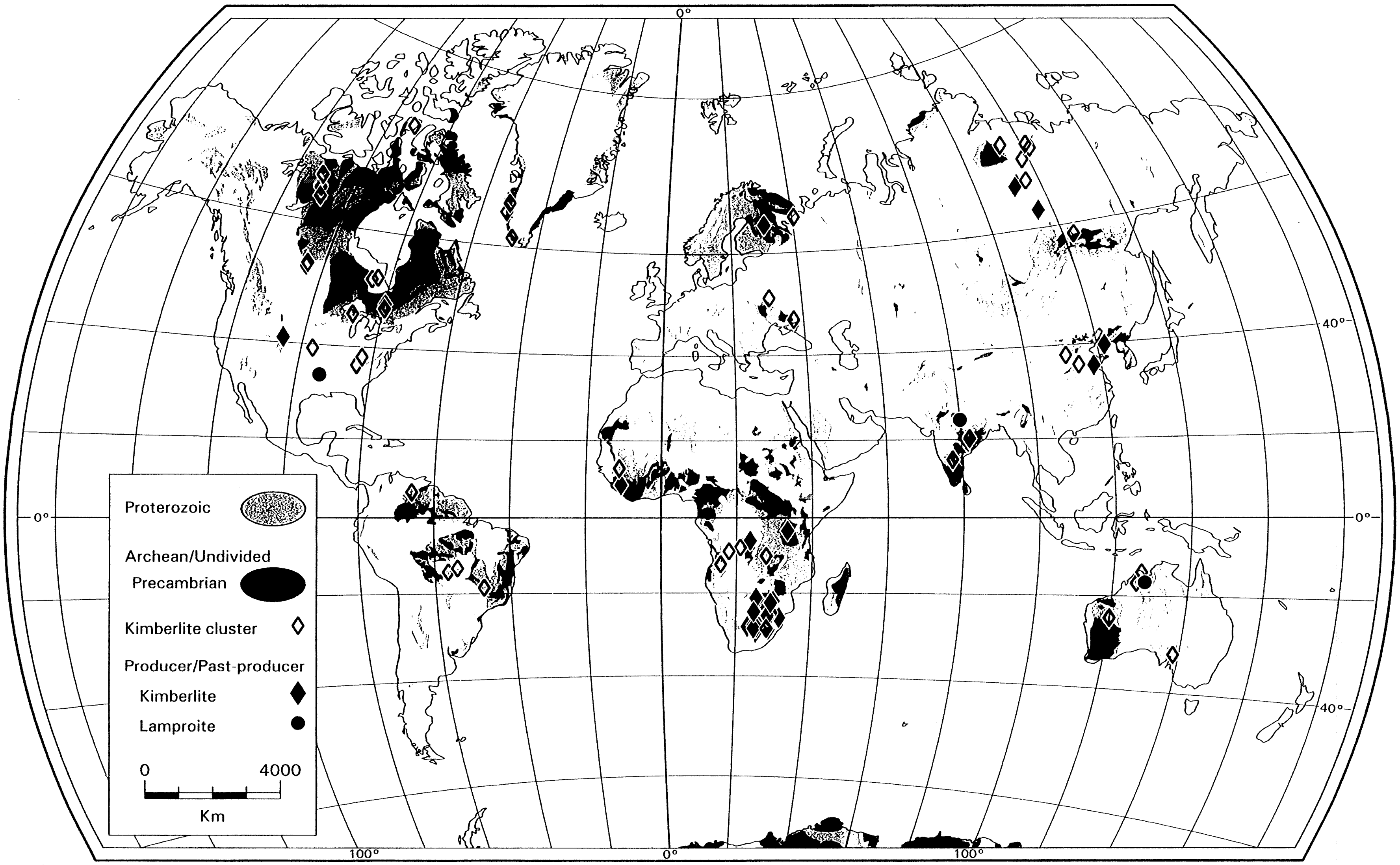


*Figure 1. Locations of diamond-bearing kimberlites in Canada, and worldwide locations of kimberlite clusters and primary diamond producers and past-producers. The distribution of Archean and Proterozoic rocks is derived from a digital geological map of the world compiled by Kirkham et al. (1994, 1995).*

Kimberlites resemble and can be spatially associated with other alkaline rocks, some of which originate at depths >150 km, and are therefore potentially diamond-bearing. Diamonds have been reportedly recovered from lamprophyre diatreme breccias in the Canadian Cordillera, from lamprophyre dykes and diatremes south of Baker Lake, N.W.T., and from the Ile Bizard diatreme breccia in Quebec. These diamond-bearing rocks and several other alkaline intrusions in Canada are described in a series of short papers.

Kimberlite, lamproite and other alkaline magmas originate as small volume melts of deep-seated origin. The triggering mechanism for generation of kimberlites is unknown, and correlation of kimberlite magmatism with mantle plumes, with flexure of the lithosphere, or with specific plate tectonic processes has not been adequately demonstrated worldwide (Mitchell, 1986; Haggerty, 1994). Furthermore, no viable theory has accurately predicted the location of kimberlite fields within a craton, although many fields have preferred orientations that suggest pre-existing structural controls are important. Lamproites and lamprophyres originate from shallower sources than kimberlites, and magmas are derived from partial melting of subcontinental lithospheric mantle that has a long and complex metasomatic history (Mitchell and Bergman, 1991). Kimberlites have chemical and isotopic signatures suggesting they originated from sources in the asthenosphere, although the magmas can be modified by interaction with metasomatized regions in the overlying continental lithosphere (Haggerty, 1994; Ringwood et al., 1992; Tainton and McKenzie, 1994).

Much of our knowledge of the deep crust and upper mantle is inferred indirectly from geophysical methods. Mantle and crustal xenoliths and xenocrysts transported to surface by deep-seated magmas are actual samples of the mantle and deep crust, and can be used to test the geophysical models and study the lithosphere beneath Canada. Pilot studies provide a glimpse of the rich potential provided by xenolith suites in kimberlites and related rocks, discovered as a result of the continued success of diamond exploration across Canada. Age determinations on diamond inclusions and on primary minerals in diamond-bearing kimberlites indicate that diamonds are xenocrysts in the kimberlite magma. Kimberlites act only as transportation agents, bringing diamonds and mantle xenoliths from within the diamond stability field to the surface. In general, diamonds are disseminated throughout the kimberlite host, although xenoliths of diamond-bearing source rocks, such as eclogite and rare peridotite, are known.

Of unique interest in some kimberlites are down-dropped blocks of country rocks. Emplacement processes involved in near-surface diatreme and crater formation produced kimberlites containing numerous country rock fragments, some of which are the only preserved evidence for stratigraphic units that have now been removed by erosion. Papers on fossil-bearing xenoliths recovered from Lac de Gras kimberlites summarize new and surprising information about the age and geological setting of pipe formation.

Mafic magmatic events generated by mantle plumes and rifting are potentially destructive to the cool diamond-bearing roots of continental lithosphere (Helmstaedt and Gurney, 1994). Large-scale mafic dyke swarms are a surface record of these thermal events, and their age, source, and distribution provide evidence about possible selective destruction or preservation of diamond-bearing mantle roots.

Conceptually, diamond formation and preservation is linked to areas of thick and cool continental lithosphere which extend into the diamond stability field in the mantle (depths >150 km). However, diamonds apparently unrelated to such old cratonic nuclei have been discovered associated with Phanerozoic collisional orogens, such as in New South Wales, Eastern Australia (Barron et al., 1994), and diamonds occur in other off-craton localities. Diamonds have been recovered from alluvial sources proximal to tectonically emplaced ultramafic massifs in collisional orogens, and microscopic diamonds have been discovered in situ in fault-bounded ultra high pressure metamorphic massifs. In addition, microscopic diamonds occur in carbon-bearing rocks at impact sites worldwide and, to date, no systematic search has been made of the twenty-six known impact structures in Canada. Three papers in this volume summarize these associations and outline the Canadian context. Although none of these unusual diamond occurrences worldwide has proven to be of major economic importance, interest is piqued by rocks from ophiolites, such as the Beni Bousera ultramafic massif, that has been interpreted to have initially contained up to 15% diamond (Nixon et al., 1991).

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Extended Abstracts Volume-Sixth International Kimberlite Conference, Novosibirsk, Russia August 1995, 707p.

# Archean cratons

**J.A. Percival**

*Percival, J.A., 1996: Archean cratons; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 11-15.*

Cratons are blocks of continental crust that have not been significantly reworked since their formation. In the Canadian Shield, three main cratons of Archean age, the Superior, Slave and Nain, form nuclei onto which younger Precambrian provinces were added in the Proterozoic (Fig. 1). In addition, parts of the Rae and Hearne structural provinces consist of Archean cratonic rocks that were variably reworked in the Paleoproterozoic.

The Nain Province of north-central Labrador and southern Greenland contains a record of complex Archean history from 3.9 to 2.5 Ga. Early Archean (3.9-3.6 Ga) supracrustal and plutonic gneissic components were structurally interleaved with younger Archean terranes between 2.78 and 2.5 Ga (Nutman et al., 1989) to form a composite craton. High-grade metamorphism affected parts of the Nain Province at 3.6, 2.8, and 2.7 Ga prior to marginal thermotectonic reworking during Paleoproterozoic orogeny (Hoffman, 1989).

Characterized for the most part by rock units 2.75 to 2.68 Ga in age, the Superior Province also contains older (3.5-2.8 Ga) components in the northwest and southwest. The craton consists of several distinct lithotectonic subprovince types: granite-greenstone, metasedimentary gneiss, high-grade gneiss, and plutonic complexes (Card, 1990). In the northwest, WNW-trending greenstone belts ~3.0 Ga in age are overlain and intruded by rocks of ~2.7 Ga. The older block acted as a nucleus onto which east-trending belts (Fig. 2) were accreted between 2.73 and 2.70 Ga as Andean-type magmatic arcs formed on its margins (Williams et al., 1992). Recent interpretations (Card, 1990; Williams et al., 1992) regard Superior Province greenstone belts as collages of juvenile oceanic deposits including abyssal plain, island arc, and oceanic plateau material, whereas the metasedimentary belts represent deformed turbidite fans shed off tectonically active highlands. Komatiitic rocks occur in most granite-greenstone subprovinces and are common in the Abitibi belt. Several deformation episodes have been recognized in most parts of the Superior Province, related to multiple accretionary events and late dextral transpression (Percival and Williams, 1989; Card, 1990; Williams et al., 1992). Archean lamprophyres were emplaced in the transpressive regime (Stern, 1996). Seismic anisotropy in the mantle lithosphere oriented parallel to crustal structures may have been inherited from late Archean tectonism (Silver and Chan, 1988, 1991).

High-grade gneisses include ~3.5 Ga rocks in the Minnesota River Valley, 2.68 Ga granulites of the Ashuanipi complex, and several deep-crustal slices exposed in the Kapuskasing and Pikwitonei uplift structures of Paleoproterozoic age (e.g., Percival and West, 1994). The Kapuskasing structure appears to have localized alkaline magmatism at 1.88 and 1.15 to 1.0 Ga.

A large area in the northeastern Superior Province has northerly structural and aeromagnetic trends that reflect lithotectonic domains of dominantly plutonic character. Sparse tonalite, as old as 3.1 Ga, and supracrustal relics (2.8-2.72 Ga) occur within granodiorite and granite emplaced between 2.73 and 2.69 Ga (Percival et al., 1994). A gradual transition from easterly to northerly structural trends occurs in north-central Quebec, which, coupled with geochronological similarity between the east-west and north-trending domains, links the two structurally distinct parts of the province.

The Slave Province consists dominantly of late Archean (2.7-2.6 Ga) supracrustal and plutonic rocks (McGlynn and Henderson, 1972; Henderson, 1981; Padgham, 1992), with inliers of older (4.0-2.8 Ga) gneiss and younger sedimentary outliers (King and Helmstaedt, in press). The Acasta gneiss, with components as old as 4.0 Ga, is exposed in the western part of the province. The presence of an ancient component is reflected in Nd isotopic values of supracrustal and plutonic rocks, which are evolved in the west and relatively primitive in the east (Davis and Hegner, 1992). Belts of volcanic rock are overlain by widespread turbiditic greywacke which makes up the bulk of the Yellowknife Supergroup (Henderson, 1970); minor, stratigraphically distinct conglomerate and sandstone sequences constitute pre- and post-Yellowknife Supergroup assemblages (King and Helmstaedt, in press). Volcanic sequences, for the most part 2.71 to 2.65 Ga, include tholeiitic mafic-felsic packages, which are common in the west, and calc-alkaline intermediate series, which are more abundant in

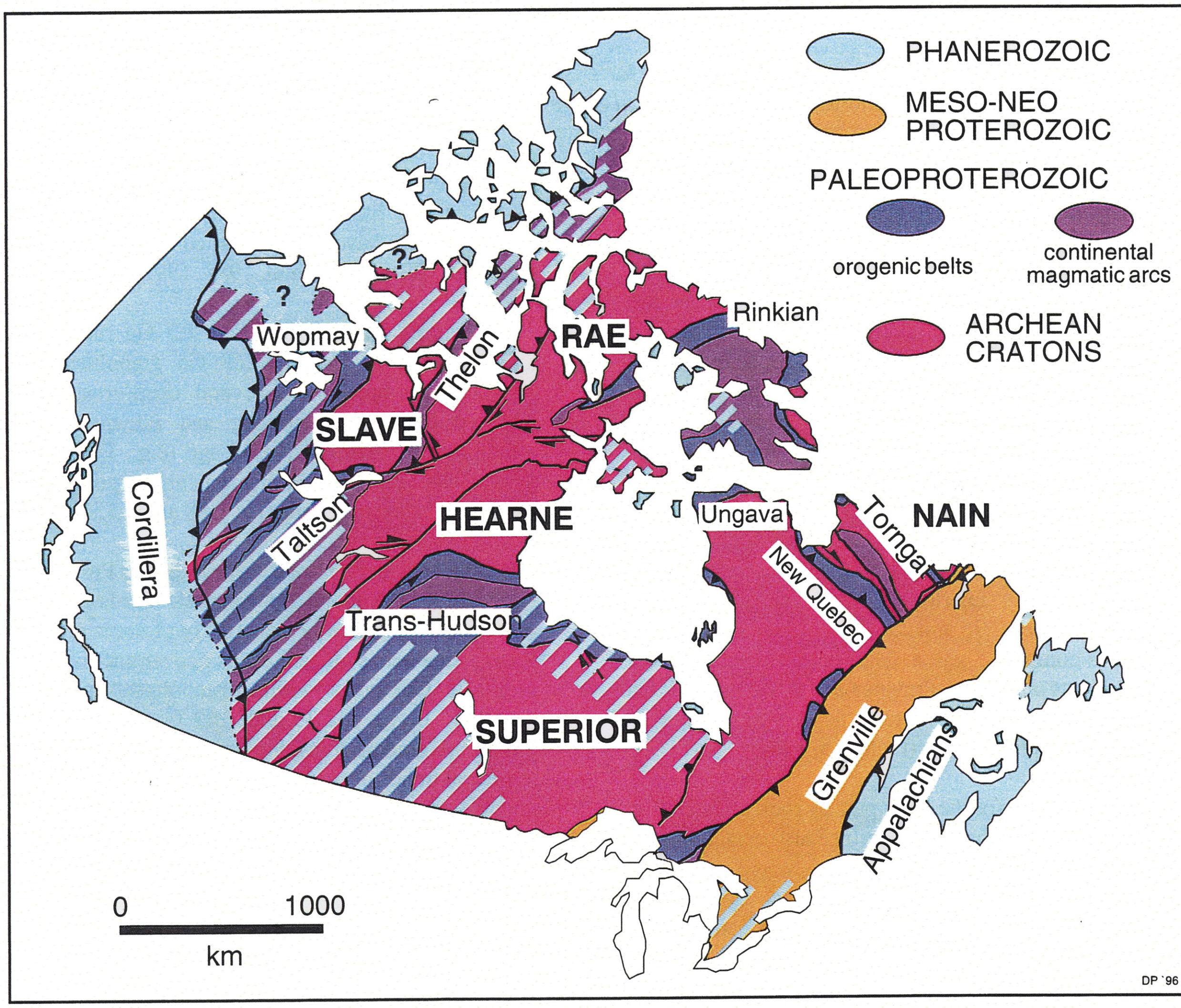


*Figure 1. Precambrian tectonic elements of the North American craton in Canada, modified after Hoffman (1988, 1989) and Ross et al. (1995). Proterozoic platformal cover is not illustrated (compare to Fig. 2). Blue diagonal stripes represent areas where the Precambrian basement is covered by Phanerozoic platformal cover. Upper case names are Archean provinces; lower case names are Proterozoic and Phanerozoic orogens.* ***Note added in proof:*** *New mapping and preliminary geochronology in southern Baffin Island (St-Onge et al., 1996) indicates that much of Meta Incognita Peninsula is underlain by Paleoproterozoic rocks, possibly a northward continuation of the Ungava orogen, rather than by Archean rocks of the Rae Province as previously interpreted. This map shows this new interpretation; Fig. 2 and other maps in this volume show southern Baffin Island as primarily an Archean craton.*

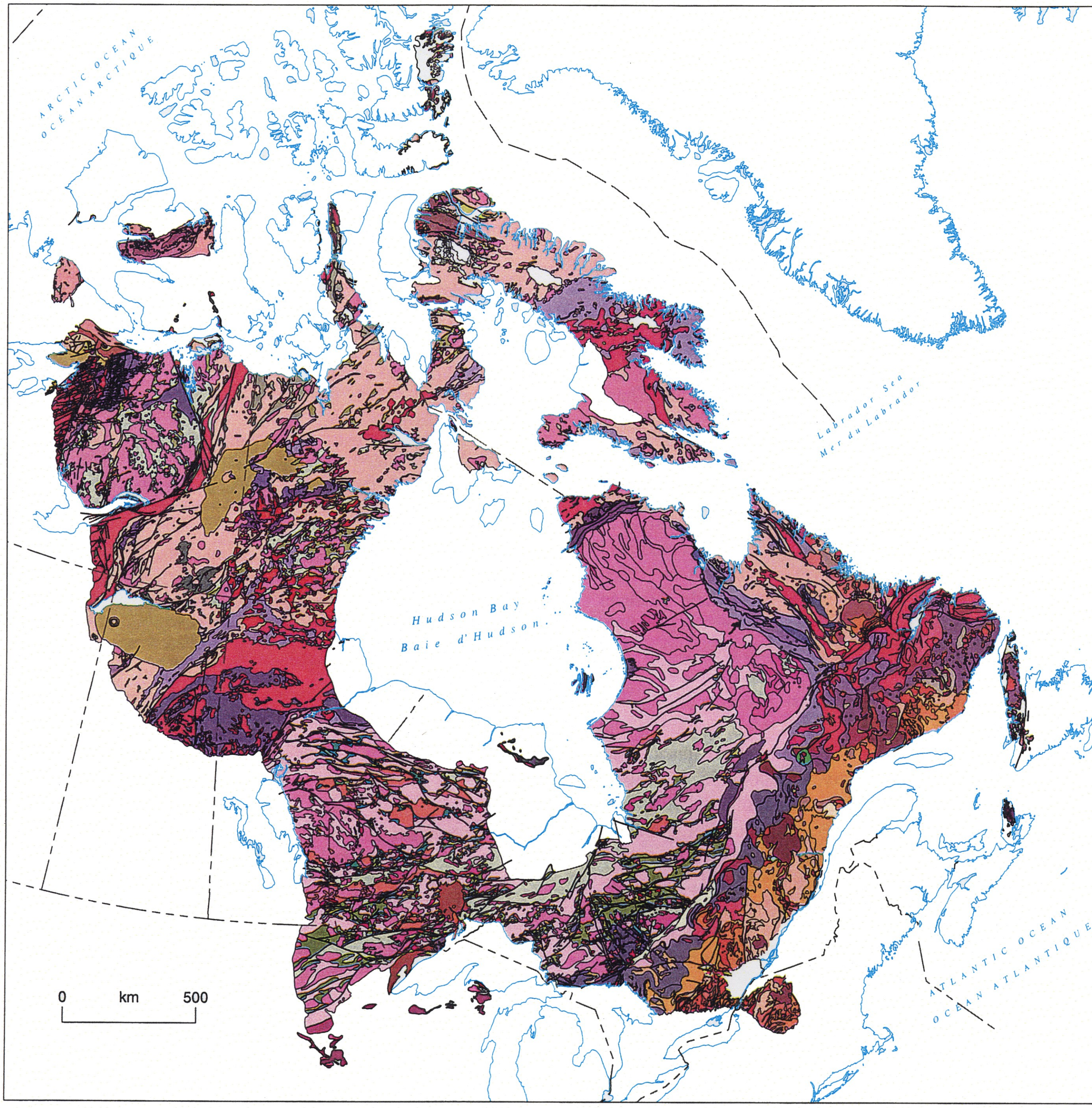


*Figure 2. Bedrock geological map of the Canadian Shield simplified from the Geological Map of Canada (Wheeler et al., in press). Archean cratons (shades of pink and green, and grey) and reworked Archean rocks (salmon and greenish brown) are separated by Paleoproterozoic orogenic belts (purple and red) and partly covered or flanked by Meso-Neoproterozoic rocks (mustard brown and orange). Structural provinces are identified on Fig. 1.*

the east. Rare komatiites and volcanics of 3.12 Ga have been recently documented (Hrabi et al., 1994). Turbiditic greywackes were derived from dominantly felsic sources 2.65 to >2.8 Ga (Villeneuve and van Breemen, 1994). In addition to the ancient gneisses, granitoid units include synvolcanic plutons and 2.58 to 2.62 Ga calc-alkaline and peraluminous suites (Davis and Hegner, 1992). Several Archean deformation episodes give the province its dominantly northerly structural grain and complex internal geometry related to early thrusting, later polyphase folding, and late faulting. Low-pressure, high-temperature metamorphism characterizes large parts of the province (Thompson et al., 1996). Seismic anisotropy in the mantle may have formed through tectonic processes during the Archean (Silver and Chan, 1988).

The Rae and Hearne structural provinces (formerly parts of the Churchill Province) consist dominantly of Archean rocks, but differ from cratons sensu stricto in their variable but widespread Paleoproterozoic reactivation. The Hearne Province (Hoffman, 1989, 1990; Fig. 1) consists mainly of high-grade Archean gneisses of 3.5 to 2.60 Ga age, but includes the Kaminak block, a 2.7 to 2.65 Ga granite-greenstone terrane. The extent of Proterozoic reworking in the Hearne Province is evident in the folded and thrust geometry of inliers of the Paleoproterozoic Hurwitz Group, although the Kaminak block appears to have retained much of its Archean character. Extensive ultrapotassic magmatism and younger rapakivi granites and rhyolites have ages in the range 1.85 to 1.74 Ga (Peterson and LeCheminant, 1996). The Rae Province (Hoffman, 1990) is exposed in three segments, in the western N.W.T., Quebec-Labrador, and the Arctic Islands (Fig. 1). It consists of Archean rocks with ages in the range 3.3 to 2.6 Ga and infolded thrust remnants of Paleoproterozoic cover sequences. Archean supracrustal rocks include 3.0 to 2.8 Ga komatiite-quartzite sequences as well as calc-alkaline volcanic rocks (2.9-2.6 Ga) and associated sediments. Deformation effects, and mid-amphibolite-facies metamorphism of cover sequences, as well as the presence of 1.85 Ga ultrapotassic rocks and 1.76 Ga rapakivi granites, suggest that the Rae Province was also significantly reworked in the Paleoproterozoic.

The Archean cratonic blocks of the Canadian Shield all have a significant 2.7 to 2.6 Ga history of metamorphism and magmatism (Percival, 1994), suggesting the possibility that a late Archean supercontinent may have existed (Williams et al., 1991). The supercontinent was fragmented and parts re-assembled in the Paleoproterozoic as Laurentia (Hoffman, 1989). Archean blocks in other cratons with a common 2.7 to 2.6 Ga history may have originated as parts of a single late Archean supercontinent.

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# Paleoproterozoic Orogenic Belts

**M.R. St-Onge and S.B. Lucas**

*St-Onge, M.R. and Lucas, S.B., 1996: Paleoproterozoic Orogenic Belts; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 17-24.*

## INTRODUCTION

Paleoproterozoic (2.5-1.6 Ga) orogenic belts in Canada preserve a record of rifting, drifting, and eventual collisional amalgamation of the principal Archean provinces. As summarized by Hoffman (1988, 1989), the zenith of Paleoproterozoic orogenesis was 2.1-1.8 Ga, the interval during which much of the core of the North American craton was assembled. A hallmark of these elongate (several hundreds of kilometres) orogenic belts (Fig. 1; Percival 1996, this volume) is that collectively they contain tectonostratigraphic assemblages developed at all stages of their tectonic history (initial rift to terminal collision), and preserved at all metamorphic grades. Across the Paleoproterozoic orogens, assemblage types include continental rift, continental (passive) margin, ophiolite, intraoceanic arc and continental arc, back-arc basin, and syn- and post-collisional basin (Fig. 1).

The forelands to these orogens include the Slave, Superior, Wyoming, and Nain cratons (Fig. 1; Percival, 1996, this volume). The 'Churchill Province', as defined by Stockwell (1961), comprises Archean (>2.5 Ga) and Paleoproterozoic (2.5-1.6 Ga) rocks that form a vast hinterland to the Paleoproteozoic orogens that bind the North American continent together. Recognition that the 'Churchill Province' actually represents a collage of Archean microcontinents welded together along Paleoproterozoic collision zones led to the suggestion by Hoffman (1988, 1989, 1990a) that it be subdivided into two fundamental entities in northwestern Canada: Rae Province and Hearne Province. Although recent studies have indicated that the boundary between the Rae and Hearne provinces in the northwestern shield (Snowbird line; Hoffman, 1988, 1990a) might comprise an intracontinental transform fault of Neoarchean age (Hanmer et al., 1994), other datasets point to significant displacements along segments of the boundary zone during the Paleoproterozoic (Ross et al., 1995). Late Paleoproterozoic (ca. 1.7 Ga) intracontinental sedimentary basins occur in the Churchill Province hinterland to the Trans-Hudson (Athabasca basin) and Taltson-Thelon (Thelon basin) orogens. The Paleoproterozoic orogenic belts are also covered by Phanerozoic platformal cover of the Western Canada sedimentary basin, the Hudson Bay Basin, Hudson Strait/Ungava Bay, and the High Arctic. This contrasts with the general absence of platformal cover on Archean cratons (i.e. the Canadian Shield), and led Hoffman (1990b) to suggest that the shield is underlain by an anomalously thick, low temperature mantle root of presumed Archean age.

## PALEOPROTEROZOIC OROGENS

A brief overview of the following orogens is presented from east to west: Torngat , New Quebec, Ungava , Trans-Hudson , Taltson-Thelon, and Wopmay. For more information on these, as well as other Paleoproterozoic belts not described in the text, the reader is referred to the references listed in the text and Hoffman (1988, 1989).

### *Torngat Orogen:*

The Torngat Orogen (Fig. 1; Van Kranendonk and Ermanovics, 1990; and Fig. 1; Percival 1996, this volume) separates the Nain Province from a proposed southeastern extension of the Rae Province (Hoffman, 1990a). The orogen has four principal lithotectonic components from west to east: Lac Lomier complex, Tasiuyak complex, Killinek suite, and the Four Peaks domain (Wardle and Van Kranendonk, in press). The Lac Lomier complex contains a mixture of Archean gneisses contiguous with those in the Rae Province, which are infolded with supracrustal gneisses and intruded by 1.87 Ga tonalite to granodiorite plutons (Wardle and Van Kranendonk, in press). The supracrustal rocks are thought to be correlative with those in the Lake Harbour Group (Ermanovics and Van Kranendonk, 1990). The Tasiuyak complex consists of a distinctive metasedimentary gneiss that extends the length of the Torngat Orogen and marks the boundary between rocks clearly associated with the Rae Province and those tied to the Nain craton. The gneiss is interpreted to be derived from turbidites and may represent an accretionary wedge (Wardle and Van

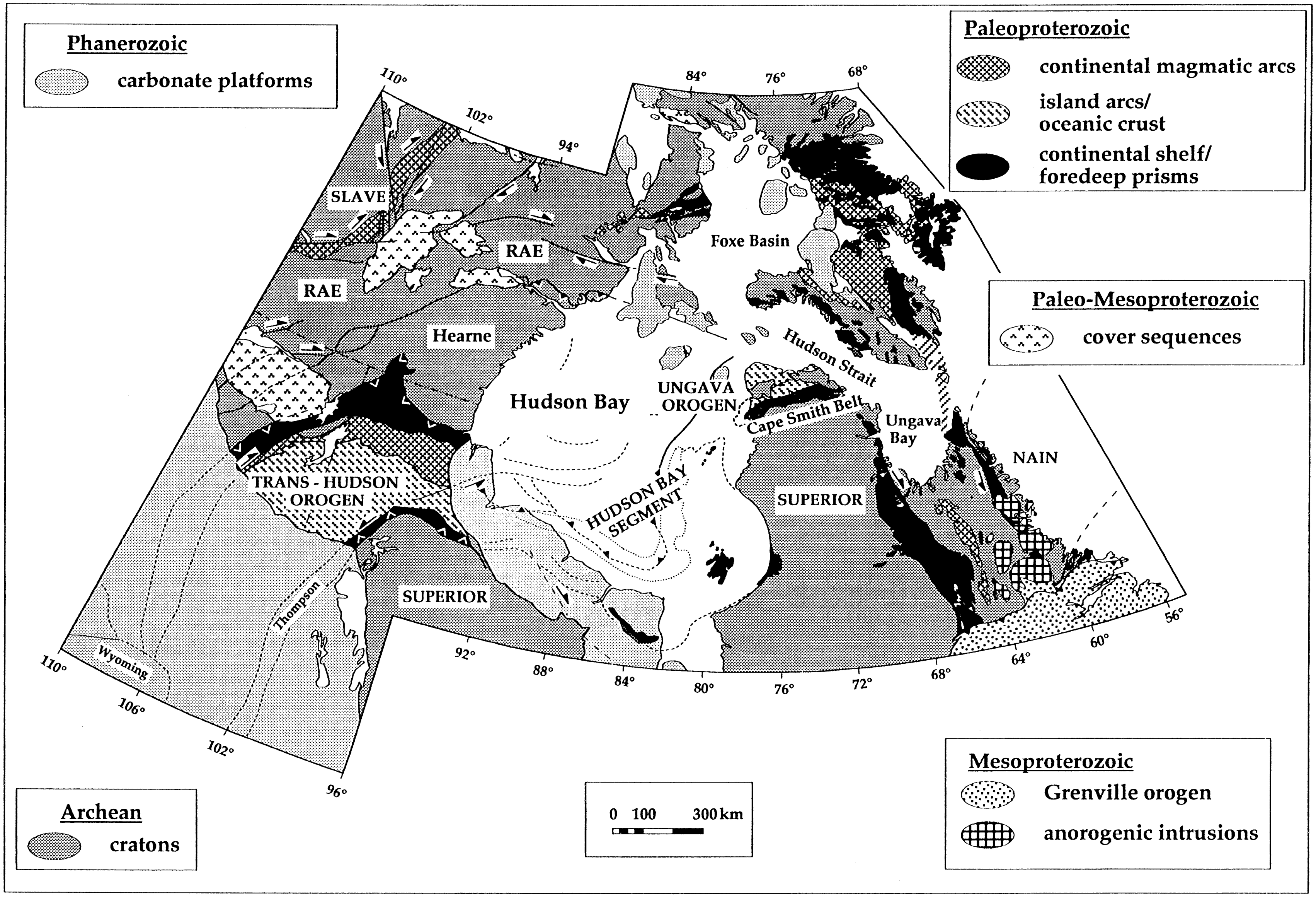


*Figure 1. Tectonic map of the Paleoproterozoic orogenic belts forming the western, northern and eastern margins of the Archean Superior Province ( after Hoffman, 1989).*

Kranendonk, in press). The Killinek suite is predominantly composed of 1.91 to 1.86 Ga calc-alkaline plutonic rocks that intrude pelitic gneisses (Scott and Machado, 1995), and has been interpreted as the roots of a magmatic arc (Wardle et al., 1992). The Four Peaks domain is composed largely of Archean gneisses that were intruded on their western margin by Paleoproterozoic anorthosite and calc-alkaline plutons (Wardle et al., 1994).

The Torngat Orogen displays an overall doubly vergent profile formed by west-verging folds in the Lac Lomier complex and adjacent parts of the Rae Province (west side), and east-verging thrusts and folds in the Torngat foreland of the Nain Province craton (east side of the orogen). The orogen's axial zone, the Tasiuyak complex is characterized by vertical structures and forms the locus for the sinistral Abloviak shear zone. The doubly vergent profile of the Torngat Orogen is attributed to sequential deformation events between 1.87 and 1.71 Ga that result from the convergence and collision of the Nain Province with the southeast Rae Province (Wardle and Van Kranendonk, in press).

### *The New Quebec Orogen:*

The New Quebec Orogen (Fig. 1; Percival 1996, this volume) is a foreland fold-and-thrust belt composed predominantly of low- to medium-grade sedimentary and volcanic rocks which separate the Superior Province from the southeast Rae Province. Four tectonostratigraphic zones characterize the New Quebec Orogen. Rocks of the autochthonous Chiok zone and the (par-) autochthonous Meleze-Schefferville zone (Clark and Avramtchev, 1990) each comprise two sedimentary/volcanic cycles. The lower cycle includes a sequence of arkose, shale and dolomite that rests unconformably upon Superior Province basement and represents the transition from a terrestrial to shallow-shelf environment during initial rifting of the Superior craton. The upper cycle disconformably overlies the lower cycle and includes a sequence of quartzite, iron formation, turbiditic shale-sandstone and arkose and alkalic lavas dated at 1879 Ma (Wardle et al., 1990). The Baby-Howse zone includes lower cycle sedimentary rocks, interstratified with basalts and intruded by gabbro sills (2.17 Ga; Rohon et al., 1993), that are overlain by upper cycle iron-formation, greywacke and ca. 1.88 Ga (Machado, 1990) mafic intrusive/extrusive rocks (cf. Skulski et al., 1993). To the east lies the Rachel zone, characterized by an amphibolite-facies volcano-sedimentary package and several complex domes and refolded nappes of Archean basement (Moorhead and Hynes, 1990). The Rachel zone may represent either the distal part of the Superior Province passive margin sequence (Poirier et al., 1990) or a fore-arc accretionary wedge (van der Leeden et al., 1990).

The New Quebec Orogen is a west-verging fold-and-thrust belt. Its western foreland is thin-skinned and roots down to the east into a thick-skinned belt which is associated with basement slices (Boone and Hynes, 1990; Moorhead and Hynes, 1990). Metamorphic grade rises progressively from greenschist facies in the west to amphibolite, and locally granulite facies in the east (Poirier et al., 1990).

### *Ungava Orogen:*

The Ungava Orogen (Fig. 1; Percival, 1996, this volume) is an arc-continent collisional belt (Lucas et al., 1992; St-Onge et al., 1992, 1995). Preserved within the orogen are: (1) (par)-autochthonous rocks of the Archean Superior Province; (2) autochthonous and allochthonous sedimentary and volcanic units (Povungnituk and Chukotat groups) associated with Paleoproterozoic rifting of the Superior Province craton; and (3) allochthonous crustal assemblages interpreted as an ophiolite (Watts Group), a fore-arc clastic apron (Spartan Group), and a magmatic arc (Narsajuaq arc and Parent Group). The amphibolite-grade ophiolite comprises crustal units only, and includes pillow basalt, sheeted dykes, massive gabbros and mafic-ultramafic cumulates (Scott *et al.*, 1992). Rift margin units yield ages that span ca. 2.04 to 1.92 Ga (Parrish, 1989; Machado *et al.,* 1993). A gabbroic layer in the mafic cumulates of the ophiolite was dated at 2.00 Ga (Parrish, 1989). Plutonic and felsic volcanic rocks of the magmatic arc range in age between ca. 1.90 - 1.82 Ga, whereas collisional granites have ages that cluster at ca. 1.80 Ga (Parrish, 1989; St-Onge et al., 1992; Machado *et al.,* 1993; R.R. Parrish, pers. comm., 1994).

Contractional deformation in response to collision of arc and oceanic assemblages with the north-facing Superior Province margin led to the development of the Cape Smith Thrust Belt, which may have initially formed at ca. 1.87 Ga but culminated by ca. 1.80 Ga (Lucas and St-Onge, 1992). Within the thrust belt, the Paleoproterozoic tectonostratigraphic assemblages were progressively juxtaposed, internally imbricated and translated southward across the Archean crystalline basement (Lucas, 1989). Following arc-continent collision, both the Paleoproterozoic thrust belt and its footwall basement were deformed into regional-scale folds during two post-thrusting folding episodes (St-Onge and Lucas, 1990; Lucas and Byrne, 1992). As a result of the thick-skinned folding events and subsequent exhumation, autochthonous footwall basement is exposed in two large tectonic windows in the orogen's internal zone.

### *Trans-Hudson Orogen:*

The Trans-Hudson Orogen, which flanks the western margin of the Superior Province (Fig. 1; Percival 1996, this volume) comprises four principal tectonic domains (Lewry and Collerson, 1990): (1) a narrow eastern foreland, the Thompson Belt (Bleeker, 1990); (2) a broad collage of dominantly juvenile arc and oceanic terranes, the Reindeer Zone, that structurally overlies an 'exotic' Archean block exposed in three small basement windows (Lewry et al., 1990); (3) an Andean-type continental margin batholith, the Wathaman-Chipewyan Batholith (Meyers et al., 1992); and (4) a broad, reworked northwestern hinterland, the Cree Lake Zone of the Hearne craton (Bickford et al., 1994). The overall tectonic history of the Trans-Hudson Orogen can be viewed in terms of the two major stages. The first stage involved convergent margin (intraoceanic) tectonics and magmatism, and occurred between 1.92 and 1.83 Ga; the second involved intracontinental tectonics following complete consumption of ocean basin(s) at ca. 1.83 Ga, and was associated with convergence between bounding Archean cratons that lasted until ca. 1.69 Ga (Lucas et al., in press). Generation of arc and back-arc crust within the Reindeer Zone occurred between 1.92 and 1.87 Ga (Bickford et al., 1990; Gordon et al., 1990; David et al., 1995; Stern et al., 1995a, 1995b) and was followed by intra-oceanic accretion at ca. 1.88 to 1.87 Ga (Lucas et al., in press). The interval from 1.87-1.83 Ga was marked by development of post-accretion arcs (e.g., 1.86-1.85 Ga Wathaman Batholith; see also Lucas et al., in press), and by continental sedimentation (Ansdell et al., 1992) and back-arc basin development (Ansdell et al., 1995). Terminal collision occurred between 1.83 and 1.80 Ga (Bickford et al., 1990; Gordon et al., 1990), and involved the Reindeer Zone and trailing Hearne craton, the 'exotic' Archean block and the Superior craton. Post-collisional deformation continued until ca. 1.69 Ga (Fedorowich et al.,1995).

In 1991, LITHOPROBE acquired over 1100 km of seismic reflection data across the orogen from the Superior Province to the Hearne Province (Lewry et al., 1994; Lucas et al., 1994; White et al., 1994). Seismic images reveal a broadly symmetric crustal reflection pattern about a crustal culmination in the west-central Trans-Hudson Orogen, with reflective zones dipping eastward and westward from near surface to the lower crust. The reflection Moho is well-defined for over 500 km across the western portion of the orogen, with a local crustal root located beneath the crustal culmination. Within the root, crustal thickness increases from 36 to 45 km. The culmination is interpreted as an image of the 'exotic' Archean block that is exposed in the basement windows within the Reindeer Zone (Lewry et al., 1994). Modelling of seismic refraction data for Trans-Hudson Orogen indicates long wavelength variations in crustal thickness (<40 to >50 km) that are consistent with those imaged in the seismic reflection profiles (Nemeth et al., in press). The uppermost mantle shows increasing velocity with depth, reaching 8.55 km/s at 140 km (Z. Hajnal, pers. comm., 1995).

One implication of the recent geological and geophysical work in the Trans-Hudson Orogen is that much of the lithosphere underlying the internal zone may be either Archean, or derived from Archean-aged lithosphere (Snyder et al., in press). Isotopic study of mantle inclusions in kimberlites that sample the sub-crustal lithosphere beneath Trans-Hudson Orogen represents an important test of this hypothesis.

### *Taltson-Thelon magmatic zone:*

The Taltson-Thelon magmatic zone (Fig. 1; Percival 1996, this volume) separates the Rae Province in the east from the Slave Province and the Paleoproterozoic Buffalo Head Terrane in the west (Thériault, 1992). The Thelon magmatic zone is a 1.9 to 2.0 Ga belt of highly deformed dioritic to granitic plutons (van Breemen et al., 1987a, b). It is inferred to comprise both a continental margin arc, resulting from eastward subduction of oceanic crust beneath the Rae Province, and a syn- and post-collisional magmatic arc (Hoffman, 1988, 1989). Eastward subduction is inferred by Hoffman (1989) on the basis of preservation of an eastward-thickening foredeep on an east-facing shelf (Tirrul and Grotzinger, 1990). The Taltson magmatic zone is the less deformed southern extension of the Thelon zone (Thériault, 1992). Three chronologically and petrologically distinct intrusive suites are recognized (Thériault, 1992): the 1.99 Ga Deskenatlata suite, the 1.96 - 1.93 Ga Slave suite, and the 1.94 Ga Konth suite. The older portion of the Taltson zone is viewed as a pre-collisional continental arc predating two younger suites of collisional granites (Thériault, 1992). Deformation of both the Taltson and Thelon magmatic zones appears to be the result of collision of Archean (e.g., Slave Province) and/or Proterozoic microcontinents with the western margin of the Rae Province (Fig. 1).

### *Wopmay Orogen:*

Wopmay Orogen is the westernmost Paleoproterozoic orogen in the Canadian Shield, flanking the west side of the Slave Province (Fig. 1; Percival, 1996, this volume). The eastern portion of the orogen comprises autochthonous and allochthonous units of the Coronation Supergroup, a shelf-rise and succeeding foredeep prism deposited on the western margin of the Slave Province (Hoffman et al., 1988). Dating of volcanic ash beds indicates that shelf subsidence began at ca. 1.97 Ga (Hoffman, 1989). Destruction of the shelf-rise margin, possibly resulting from collision of the Slave margin with the Hottah arc terrane (see below) at about 1.91 Ga, involved: (1) flexure of the shelf and deposition of the foredeep sedimentary rocks, (2) development of an east-verging, thin-skinned, fold-and-thrust belt, (3) emplacement of 1.90 to 1.88 Ga mafic-felsic plutons (Hepburn Batholith) in off-shelf sedimentary rocks; and (4) eastward translation and imbrication of the Coronation Supergroup sedimentary prism and the still-hot Hepburn Batholith onto the Slave margin concurrent with low- to high-grade metamorphism (Hoffman et al., 1988).

The Hottah terrane and Great Bear magmatic arc occur in the western portion of the orogen (Hildebrand et al., 1987). The older (1.95 - 1.91 Ga) Hottah arc terrane was strongly deformed between 1.91 and 1.90 Ga, possibly due to collision with the Slave continental margin. Establishment of east-dipping, dextral-oblique subduction following Hottah-Slave collision led to generation of the younger 1.88 to 1.86 Ga Great Bear arc. The calc-alkaline Great Bear arc was principally built on Hottah terrane, but also onlaps and intrudes the deformed continental margin to the east (Hoffman, 1989). Termination of Great Bear arc magmatism was followed by thick-skinned folding and dextral shearing of the arc and the thrust belt to the east. Finally, Wopmay orogen was transected by a system of conjugate transcurrent faults, accomplishing east-west shortening and north-south extension, sometime between 1.84 and 1.66 Ga (Hoffman, 1989).

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# The Mesoproterozoic Grenville orogen

**A. Davidson**

*Davidson, A., 1996: The Mesoproterozoic Grenville orogen; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 25-27.*

The Grenville Province in the southeastern part of the Canadian Shield is part of a much larger orogen that extends in the subsurface across the North American craton as far southwest as Texas and Mexico (Rankin et al., 1993), and which is represented in Scandinavia by the Sveconorwegian Province (Gower et al., 1990). Its northwest margin, the Grenville Front, is a pronounced crustal-scale fault (e.g. Green et al., 1988), marked at the surface by zones of southeast-dipping mylonite which truncate the structural trends that characterize the older, neighbouring Archean cratons and Paleoproterozoic orogens (Fig. 1a and 1b; Percival, 1996, this volume). Relatively deep levels of orogenically thickened crust are exposed in its interior, witnessed by the prevalence of high-grade metamorphic rocks including large tracts of granulite and local relics of eclogite. The Grenville orogen is widely interpreted as the end result of continent-continent collision whose effects terminated at ~1.0 Ga (e.g. Windley, 1989, 1993; Davidson, 1995).

In terms of regional structure, the Grenville orogen in Canada (Fig. 1) is a complex collage of north- to northwest-verging, crust-scale thrust sheets containing, in a broad sense, progressively southeastward-younging rocks that have been carried on low angle ductile shears toward pre-Grenvillian Laurentia (Rivers et al., 1989; Hoffman, 1989). Within the province are terranes of 1), reworked rocks representing the adjacent Superior craton and its marginal Paleoproterozoic orogenic belts (Makkovik, Torngat, and New Quebec in the northeast and Penokean in the southwest), 2), late Paleoproterozoic orogens whose equivalents lie mainly outside the confines of the exposed Shield (Labradorian in the northeast, Yavapai-Mazatzal in the southwest), and 3), Mesoproterozoic supracrustal and plutonic rocks that can be related more directly to the development of the Grenville orogen itself. In sequence, the Mesoproterozoic rocks record the break-up of a continent assembled at ~1.6 Ga (1.5-1.35-Ga igneous suites), formation and closure of one or more (small?) oceanic basins in the central and southwest parts of the province (1.45-1.23-Ga calc-alkaline suites), and rift-type sedimentation and igneous activity elsewhere (1.33-1.24 Ga), that was followed by north- to northwest-directed thrust stacking of thick crustal slices, and then by pervasive within-plate magmatism (1.18-0.95-Ga anorthosite and A-type granite suites). The thickened crust was subject to extension in the late stages of supercontinent assembly, leading to uplift and unroofing at ~1.0 Ga.

There is no evidence for Archean crust except near the front adjacent to the Superior craton (Fig. 1). Much of the province contains reworked mid-Paleoproterozoic and early Mesoproterozoic crust, the latter dominating the southeast half (Dickin and McNutt, 1989, 1990; Dickin and Higgins, 1992). New crust was added from mantle sources following post-collision thickening; addition through early Grenvillian subduction-related processes appears to have been very minor. The Grenville Province thus has an orogen-parallel age-zonation: reworked Archean to mid-Proterozoic rocks (>1.8 Ga) are restricted to the northwest side; mid- to late-Mesoproterozoic supracrustal and plutonic rocks (1.3-0.9 Ga) are limited mainly to the southeast.

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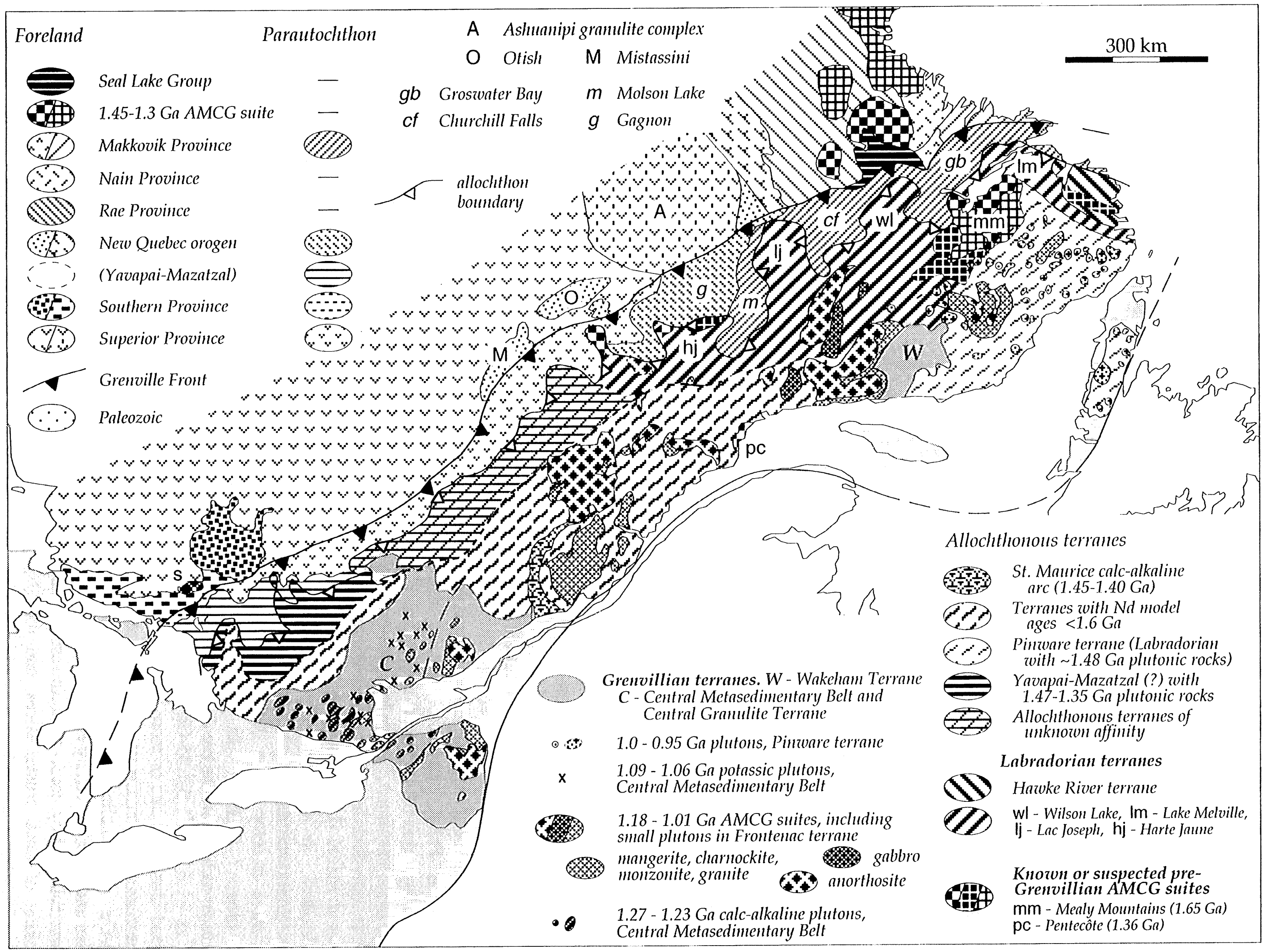


***Figure 1**. Principal tectonic elements of the Grenville Province in Canada (Davidson, 1995).*

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# Kimberlites


## B.A. Kjarsgaard



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## INTRODUCTION

Two types of kimberlite have been recognized (Wagner, 1914; Smith, 1983): 'basaltic' or 'Group I' kimberlite, hereafter termed kimberlite, and; 'micaceous' or 'Group II' kimberlite, hereafter termed orangeite (Mitchell, 1991; 1995). These latter rocks have only been recognized in South Africa, and therefore will not be discussed in greater detail. In kimberlites, diamonds occur mainly as sparsely dispersed mantle-derived xenocrysts and in mantle xenoliths contained within the kimberlite matrix. Age determinations on diamond inclusions and their kimberlite hosts (Kinney and Meyer, 1994) are consistent with other evidence suggesting that kimberlite-derived diamonds (specifically macro- diamonds >1 mm) are xenocrysts. A number of assumptions are made in diamond age determinations, as the inclusions are not syngenetic (except zircon). Note that ultramafic (or P-type) diamond inclusions are utilized for Nd model age determination; the inclusions are pooled from several diamonds. Eclogite diamond inclusions are utilized for Sm-Nd, Ar-Ar or U-Pb studies; the inclusions are, in general, from single diamonds (see Kinney and Meyer, 1994 for further discussion). Kimberlites act as transportation agents only, bringing diamonds or diamond-bearing mantle xenoliths from within the diamond stability field ( >4.5 GPa or 150 km depth, for normal continental geotherms) to the surface.

Economic quantities of diamond are mainly found in kimberlite diatremes. Kimberlites with preserved crater plus diatreme facies rocks are much rarer, but in a few cases represent important high grade and high tonnage deposits (e.g. Orapa and Jwaneng, Botswana). In kimberlite pipes, the grades and qualities of diamond vary considerably. Approximately 1% of all kimberlite pipes worldwide are economic. There are approximately 5000 kimberlites worldwide, with 50 mined at some time or another. At present, 20 mines are active and 15 of these are major producers. Currently, diamonds recovered from kimberlites and related secondary deposits account for >90% (monetary value) of all diamonds mined worldwide. Lamproite and related diamond deposits are the other important producer (see Peterson, 1996).

## KIMBERLITE: DEFINITION AND CLASSIFICATION

Kimberlite is a volatile-rich ultrabasic rock that has an enriched incompatible (Sr, Zr, Hf, Nb, REE) and compatible (Ni, Cr, Co) element signature. Kimberlites often appear hybrid in nature, as they may contain mantle xenoliths, mantle xenocrysts, and macrocrysts (a nongenetic term for large crystals of unknown origin, 1 - 20 cm in size), plus crustal xenoliths in a matrix crystallized from a kimberlite magma.

The following definition of kimberlite has been adapted and modified from Clement et al. (1984) and Mitchell (1986). Kimberlites are $CO_2$- and $H_2O$-rich rocks that have a distinctive inequigranular texture due to the presence of large, rounded, anhedral macrocrysts (i.e., megacrysts and xenocrysts) plus euhedral to subhedral phenocrysts set in a finer grained groundmass. Macrocrysts include minerals derived from disaggregated mantle xenoliths plus the kimberlite megacryst suite of minerals (olivine, Mg-ilmenite, Ti-Cr-pyrope garnet, clinopyroxene, phlogopite, enstatite and zircon). Primary matrix minerals include second generation euhedral olivine phenocrysts/microphenocrysts, and one or more of the following: spinels, ilmenite, perovskite, monticellite, apatite, phlogopite-kinoshitalite micas, carbonates, and primary serpentine. Diopside (microcrystalline) is rarely observed in crustally contaminated diatreme and hypabyssal facies rocks; it forms at the interface of xenoliths with kimberlite matrix. Commonly, macrocrysts and both early- and late-formed matrix minerals (e.g. monticellite) are replaced by deuteric serpentine and calcite.

The diverse mineralogy and associated mineral chemistry of kimberlites are reflections of the unusual major and trace element composition of these rocks. In this respect, combined petrographic, mineral chemistry and whole-rock geochemical studies can usually discriminate kimberlites from other rock types of similar mineralogy (e.g., alnöite, aillikite and other lamprophyres) and magmatic style (Mitchell, 1986). Chemical zoning trends observed in minerals such as phlogopite (plots of $Al_2O_3$ - FeO and $Al_2O_3$ - $TiO_2$) and spinel (reduced and oxidized spinel prism plots) can be particularly useful in constraining the identification of an unknown rock type (Mitchell, 1986).

Discrimination between kimberlites and ultramafic lamprophyres is, however, exceptionally difficult and there are a number of Canadian examples where petrologists disagree upon the classification of an unknown rock, for example Ile Bizard (compare Mitchell, 1983 with Raeside and Helmstaedt, 1982). Other examples include the Batchelor Lake (or Le Taq) 'kimberlite' (Watson, 1955), which is considered an ultramafic lamprophyre (Mitchell, 1986; Kjarsgaard, unpublished data) and the newly discovered diamondiferous Kyle Lake pipe in the James Bay Lowlands which is at present unclassified (Robinson, 1995). Drawing strict and exact classification boundaries between kimberlites and ultramafic lamprophyres (e.g. Fig. 4 of Scott-Smith, 1994) is not entirely feasible (Fig. 1). Part of this problem lies in the misconception that ultramafic lamprophyre magmas are generated at lower pressures ($\approx$3.0 GPa) than kimberlites (Mitchell, 1994; Scott-Smith, 1994). However, some lamprophyres must be generated at pressures >4.5 GPa, in order to be consistent with xenolith studies, experimental data, and the occurrence of diamonds in these rocks (Janse, 1994).

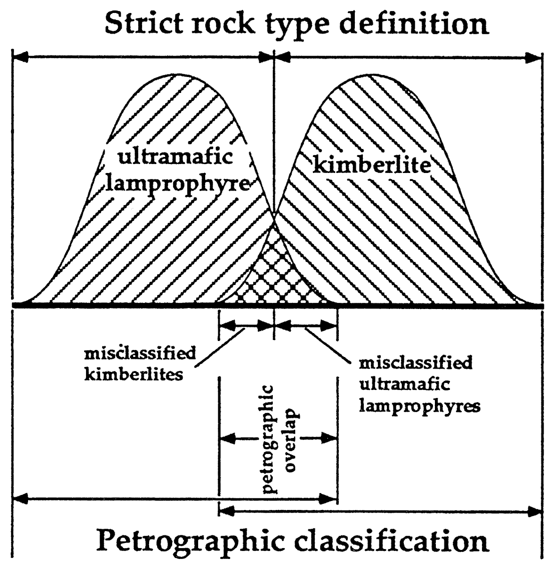


*Figure 1. Petrographic variation in rock types classified as ultramafic lamprophyres and kimberlite, illustrating petrographic overlap (horizontal axis). Petrographic overlap is the occurrence of the same mineral assemblage in two rocks which are strictly defined as being petrologically distinct. Note that in the area of petrographic overlap, the rock type definitions, and samples might be misclassified. Modified from Scott-Smith (1994).*

## A GENETIC MODEL FOR KIMBERLITE

Kimberlites occur in a restricted tectonic setting and are observed only in continental shield regions older than 1.5 Ga (Clifford, 1966). The initiation of kimberlite magmatism is deep seated, and magma generation is poorly understood. Correlation of this magmatism with hotspots or plate tectonic processes (transform fault extensions, subduction zones) cannot be satisfactorily demonstrated on a worldwide basis (Mitchell, 1986). Kimberlite magmas were thought to form by the partial melting of carbonated peridotite source regions at 5.0 to 6.0 GPa (Eggler, 1989). However, Ringwood et al. (1992) have proposed an alternate model in which kimberlite magma is generated by partial melting at much higher pressures (10.0 to 16.0 GPa). Ultra-high pressure majorite garnets that occur as inclusions in diamonds (Moore and Gurney, 1985) and in mantle xenoliths (Haggerty and Sautter, 1990) are consistent with kimberlite magma formation at depths of at least 300 km (10.0 GPa).

The range in diamond contents of kimberlites is dependent upon the amount of diamond-bearing mantle material entrained by the ascending magma, the proportions of various mantle lithologies sampled (e.g. eclogite and peridotite; eclogite often contains higher modal diamond content) and the degree to which resorption and mechanical sorting of this entrained material occurs during transport to the surface. Kimberlites probably ascend through the mantle at substantial velocities (10 - 30 km/hr; Eggler, 1989) by crack propagation processes. Near the surface, vent velocities of a few hundred km/hr may be possible, due to rapid $CO_2$ degassing from the magma. Highly explosive near-surface volcanism is consistent with the formation of kimberlite diatremes and tuff rings as well as the entrainment of large amounts of angular crustal material.

A number of distinct morphological types of kimberlite pipes have been recognized (this study). The first, probably best known type, is the classical 'South African pipe' (Clement, 1975). These occur as cone-

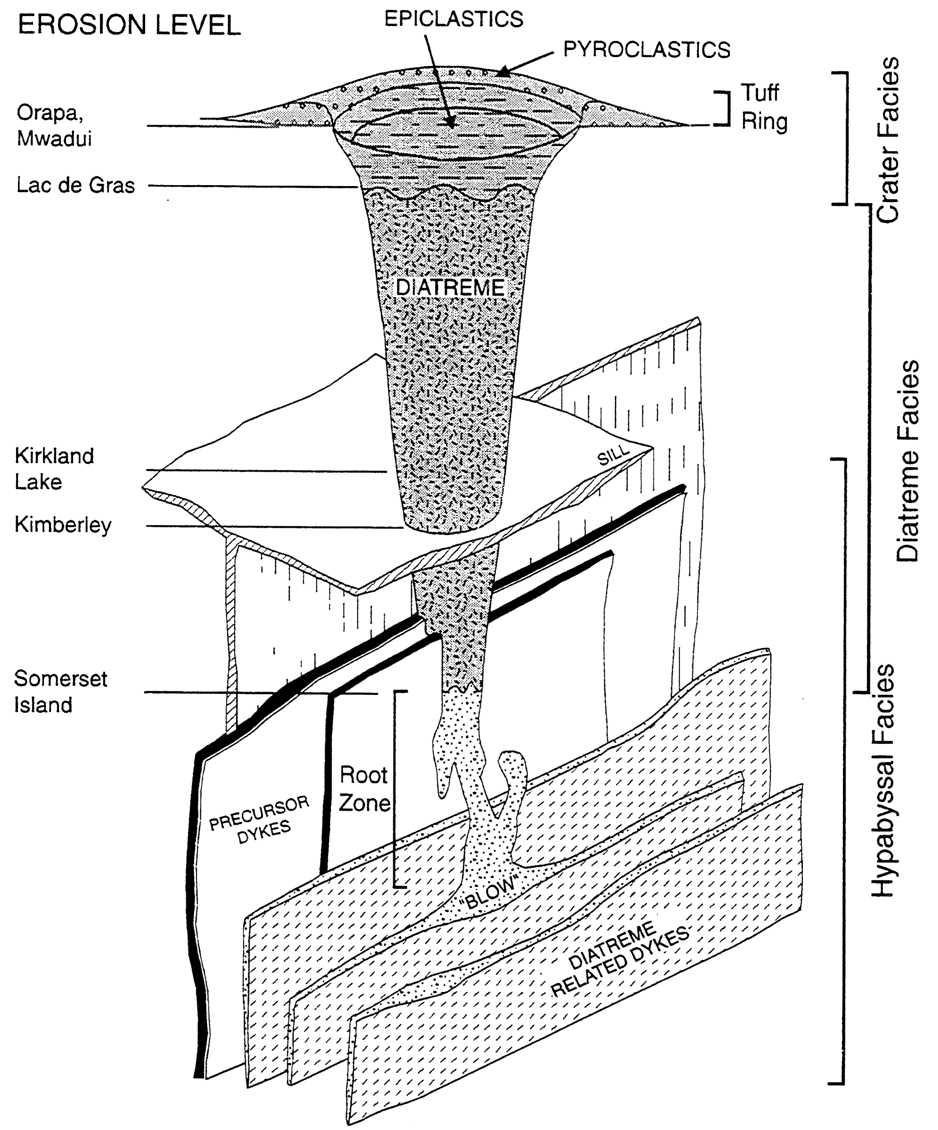


***Figure 2.** Generalized model of the classic South African type of kimberlite magmatic system, showing crater, diatreme, and hypabyssal facies rocks. Crater facies rocks consist of pyroclastic (tuff ring) and resedimented volcaniclastic rocks (crater infill); diatreme facies rocks consist mainly of tuffisitic kimberlite breccias; hypabyssal facies rocks are found in the root zone of the diatreme and consist of "blows" (enlarged dykes), dykes, and sills. Also shown are the present erosion levels of some representative economic and Canadian kimberlites. Modified from Mitchell (1986).*

shaped diatremes (Fig. 2), with steeply dipping (75 to 85°) country rock contacts. Rarely, crater facies rocks are preserved at the top of the diatreme (Fig. 2). Crater facies rocks consist of kimberlite pyroclastics (i.e. tuff ring) plus resedimented volcaniclastics within the crater. Crater walls dip inward at angles ranging from 25 to 75°. Diatreme facies rocks consist mainly of tuffisitic kimberlite breccias, which are relatively uniform compared to crater or hypabyssal facies rocks. This homogeneity is envisaged to result from the mixing processes involved in diatreme formation (Mitchell, 1991). With increasing depth, kimberlite diatremes grade into root zones (Fig. 2), consisting of kimberlite dykes, blows, and sills. Several distinct intrusive phases of hypabyssal kimberlite can occur in the root zone of a kimberlite pipe.

Subsequent erosion of South African type pipes (Fig. 2) produces differing levels of exposure (e.g. crater-, diatreme-, or hypabyssal-facies kimberlite), and variable pipe geometries. However, other distinct morphological variations of kimberlite pipes have recently been recognized (Fig. 3). These include 'miniature' versions of the classic South African type (containing crater-,

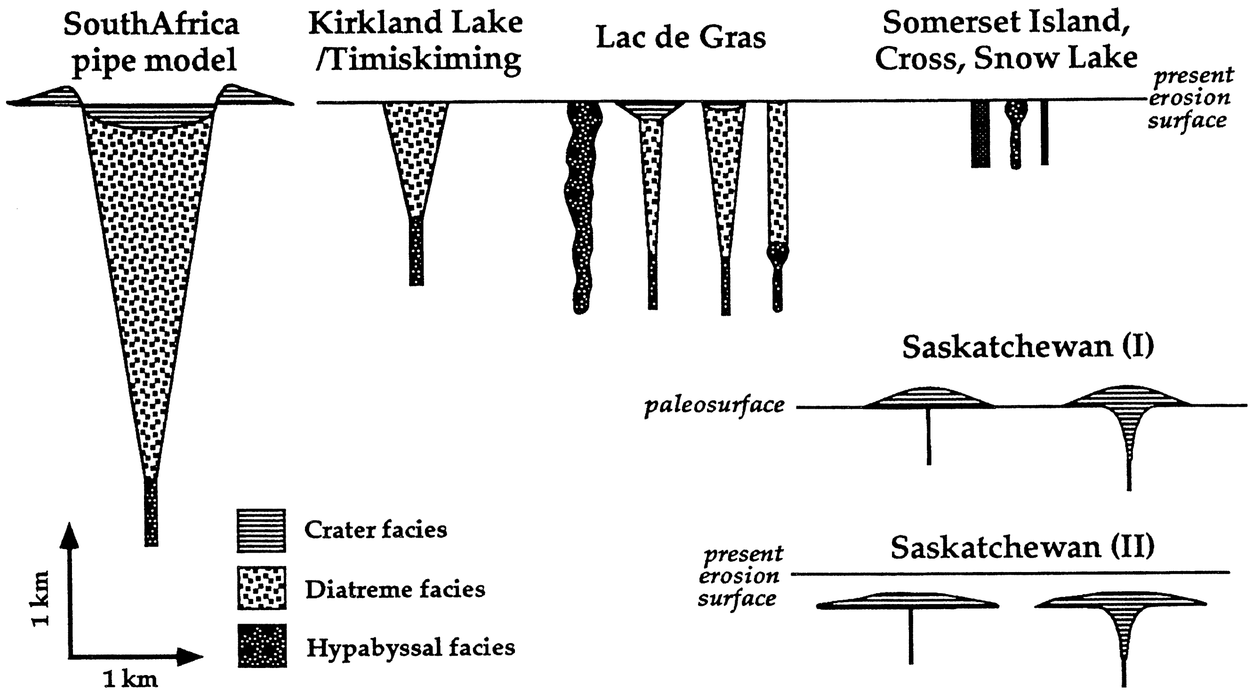


***Figure 3.*** *Variation in morphology of kimberlite pipes in Canada, as compared to the classic South African type pipe. Note that kimberlite pipes in Kirkland Lake/Timiskiming (diatreme facies) and Somerset Island/Cross/Snow Lake (hypabyssal facies) are interpreted as South African type pipe which have been variably eroded. In contrast, many of the Lac de Gras area pipes are 'miniature' versions of the classic South African type pipe, a morphology which is also observed in Yakutia. The central Saskatchewan kimberlites form tephra cones, with or without associated underlying vents (Saskatchewan I). Subsequent marine transgression, reworking and burial produces the present day 'pancake' or sheet-like morphology (Saskatchewan II).*

diatreme-, and hypabyssal-facies rocks). Numerous pipes in the Lac de Gras field (Fig 2.; Kjarsgaard, 1996a, this volume) as well as pipes in Yakutia (Mitchell, 1995) have this 'miniature' geometry. Furthermore, kimberlite pipes which do not contain diatreme facies kimberlite, but consist only of crater facies rocks (pyroclastic kimberlite and associated resedimented volcaniclastic kimberlite), forming tephra cones or tephra cones with associated underlying vents (Fig. 3) are recognized in central Saskatchewan (Fig. 1; Kjarsgaard, 1996c, this volume) and in Zaire. Crater wall contacts in these pipes are horizontal to shallowly (0 - 35°) dipping. At Fort à la Corne, the occurrence of bedded, kimberlite juvenile lapilli tuffs are interpreted to have formed by strombolian and vulcanian-type eruptions (Kjarsgaard, 1995, 1996c). This passive to mildly explosive style of volcanism contrasts with the highly explosive style inferred for the formation of kimberlite diatremes. Furthermore, the occurrence of rare, glassy kimberlite lapilli tuffs (products of subaerial fire fountaining) implies the existence of small volume, transient, upper crustal kimberlite magma chamber(s), which has important implications for diamond resorption.

## KIMBERLITES IN CANADA

The locations of kimberlites in Canadian is shown in Figure 4. Recent exploration activity for kimberlites in the Slave Province has resulted in the discovery of more than 100 pipes, most of which are in the central or southern part of the province (Kjarsgaard, 1996a). The Somerset Island kimberlite field is described in more detail by Kjarsgaard (1996b). Victoria Island, possibly underlain by a northern extension of the Slave Province, has been the site of recent exploration activity, with promising till indicator mineral anomalies (Northern Miner, 1995). The Crossing Creek kimberlite, in the southern Canadian Rocky Mountains, has been described by Hall et al. (1989) and Ijewliw and Pell

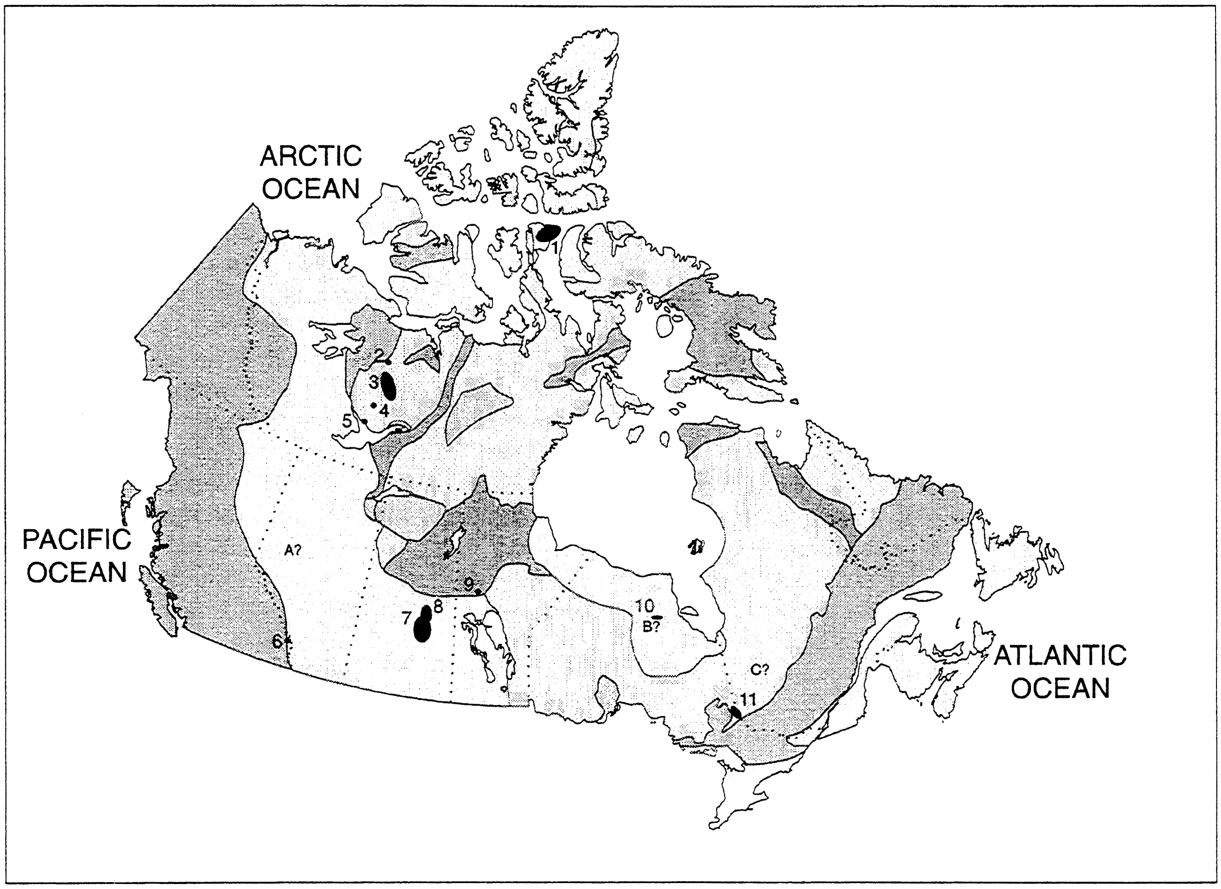


*Figure 4. Location map of kimberlites in Canada: 1) Somerset Island cluster; 2) Ranch Lake; 3) Lac de Gras province; 4) Cross Lake; 5) Dry Bones Bay; 6) Crossing Creek; 7) Fort à la Corne field; 8) Candle Lake cluster; 9) Snow Lake - Wekusko; 10) Attawapiskat cluster; 11) Kirkland Lake field. Probable kimberlites are: A? = Grande Prairie; B? = Kyle Lake; and C? = Batchelor Lake (Le Taq)*

(1996). Kimberlites in the Canadian prairies (Alberta; plus the Candle Lake and Fort à la Corne pipes in central Saskatchewan) are described by Kjarsgaard (1996c; in press). Relatively little is known about the kimberlites in the Wekusko Lake area of Manitoba, located to the east of the Snow Lake Cu-Zn mining camp. Schulze (1996) provides an update on kimberlites in the Kirkland Lake/Lake Timiskaming field (Archean Superior Province). At present, little is known about the kimberlite pipes in the Attawapiskat field (James Bay Lowlands, Ontario). It is surprising that no kimberlite fields have been discovered in the northern Quebec portion of the Superior Province. Similarly, one might expect kimberlites to exist in the Archean Nain craton of Labrador.

## EXPLORATION MODEL

Economic kimberlites are found in old (>2.5 Ga) stable cratons characterized by thick crust and low geothermal gradients. Various methods are used to locate kimberlites, depending upon local conditions (e.g. type of country rock, climate and overburden). The main exploration techniques used are: 1) indicator mineral sampling (heavy mineral separates from stream sediment, soil and till sampling); 2) remote sensing (LANDSAT, air photo interpretation); 3) geophysical surveys (magnetic, gravity, electrical, radiometric, seismic profiling); and, 4) geochemical. Biogeochemical methods have also been utilized. Atkinson (1989) provided a recent general summary of exploration techniques.

As kimberlites are rare rocks which generally do not outcrop well, exploration methods must be capable of finding a hidden target. Although the geophysical signature of a kimberlite is not unique, it is unusual and can be discerned by low-level aeromagnetic and EM surveys. Numerous kimberlite pipes in the Archangel region (Russia) and the central Kalahari in Botswana were located by aeromagnetic surveys (Atkinson, 1989). This technique is most effective in areas with a uniform and low magnetic background. Combined aeromagnetic and electromagnetic surveys have been used with a high success rate in the Lac de Gras area, NWT. A recent, comprehensive review of geophysical techniques as applied to kimberlite exploration can be found in Urquart and Hopkins (1993).

The unique mineralogical signature of kimberlites enables the application of indicator mineral sampling exploration techniques. The identification of resistant minerals that can indicate the potential presence of a kimberlite has been widely and successfully applied as an exploration technique in South Africa, Yakutia, and Canada. However, it is extremely important to note that these so-called kimberlite indicator minerals are also found in many other rock types that do not contain diamonds. Kimberlite indicators include minerals derived from the kimberlite (spinel, olivine, ilmenite, and perovskite), all the macrocryst minerals (olivine, spinel, low-Cr Ti-pyrope, Mg-ilmenite, Cr-diopside, enstatite, and zircon), as well as minerals from disaggregated mantle xenoliths (olivine, enstatite, Cr-diopside, chrome pyrope garnet, Cr-spinels, pyrope-almandine garnet, omphacitic pyroxene, and diamond). In Canada, application of the indicator mineral method to stream sediment sampling is problematic due to Quaternary glaciation. However, success in locating kimberlite pipes has been obtained by esker and till sampling in the Lac de Gras and Kirkland Lake areas (Ward et al., 1996 and McClenaghan, 1996). A combination of alluvial and stream sediment sampling for heavy minerals, coupled with ground magnetics, was utilized in the discovery of pipes in the Attawapiskat kimberlite cluster (A.J.A. Janse, pers. comm., 1994).

Because diamond is a rare mineral in kimberlite, (0-1.4 ppm), a subset of the kimberlite indicator minerals, termed 'diamond indicators' is used to indicate the potential presence of diamond in these rocks. This is based on studies of silicates and oxide inclusions in diamond and minerals from diamond-bearing mantle xenoliths (Gurney, 1989). Specific diamond indicator minerals (with xenolith-paragenesis type in brackets) include subcalcic Cr-pyrope (garnet-bearing harzburgite/dunite source rock), Cr-pyrope garnet (garnet-bearing lherzolite source rock), high Cr-Mg chromite (chromite-bearing harzburgite/dunite source rock), and high Na-Ti pyrope-almandine garnet (eclogite source rock). It is important to note that these minerals (and xenoliths) are not definitive of kimberlite volcanism, as they can be observed in other rock types of deep-seated origin (e.g., ultramafic lamprophyres). Furthermore, these minerals are not an infallible indicator of the presence of diamond in kimberlite; the Skerring (Australia) and Zero (South Africa) kimberlites both contain subcalcic Cr-pyrope garnet, but lack diamonds (Gurney, 1989).

A new diamond exploration tool, the 'Ni in garnet thermometer', has recently been developed (Griffin et al., 1989) for studies concerning diamond exploration and genesis. Griffin (1990) suggested that the Ni thermometer is a "rapid, cheap and powerful tool for evaluating diamond prospects" and, furthermore, "is independant of the presence or absence of G10 garnets and appears in fact to be a more robust and reliable indicator of potential diamond grade". The geothermometer is based on the observation that the Ni content of garnet (in equilibrium with olivine) is strongly temperature dependent, and involves the measurement of low concentrations (5 - 130 ppm) of Ni in garnet by PIXE ('proton microprobe'; e.g., Campbell et al., 1996). The use of the Ni thermometer in assessing diamond potential relies on converting a temperature estimate to a depth estimate. This is accomplished by obtaining a pressure estimate from the intersection of an assumed geothermal gradient with the temperature obtained by Ni thermometry; garnets having estimated pressures greater than 4.5 GPa fall in the 'diamond window'. However, Kjarsgaard (1992), noted that a major drawback of this technique is the problem of rigorously constraining the geothermal gradient, a point later acknowledged by Griffin and Ryan (1993).

A further problem with Ni thermometry, as discussed by Kjarsgaard (1992), lies in the empirical calibration of this geothermometer. Kjarsgaard (1992) suggested that the Ni thermometer could be re-calibrated using the BKN thermometer of Brey and Köhler (1990) for reference temperature. Data from Kjarsgaard (1992), shows good congruence with the new experimental calibration of Canil (1994), as shown in Figure 5. The original calibration line of Griffin et al. (1989), lies at a much shallower slope than the calibration suggested by Canil (1994)(Fig. 5). This difference in slope of the regression line has important implications for applying Ni thermometry to diamond exploration (e.g. Fig. 9 of Canil, 1994). The two different calibrations of the thermometer overlap only in the range 1050 to 1150 °C (Campbell et al., 1996). Furthermore, the disparity in Ni calibration, noted above, suggests that the application of Zn in chromite geothermometry and also 'fictive garnet Cr barometry' (Griffin and Ryan, 1993) is open to question, as they both rely on the validity of the Ni thermometer calibration.

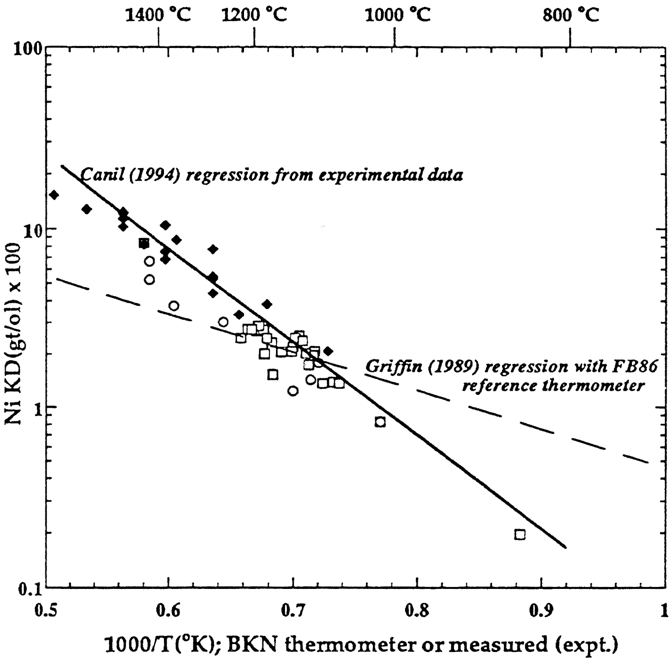


***Figure 5.*** *Comparison of data sets for Ni partitioning between Cr-pyrope garnet and olivine versus temperature. Solid diamonds and associated regression line (solid line) are from the experimental study of Canil (1994). Open squares represent data from Somerset Island lherzolite xenoliths (Kjarsgaard, 1992) with temperature determined by the recommended thermometer (BKN) of Brey and Köhler (1990); Ni in garnet for the Somerset xenoliths determined by micro-PIXE at the University of Guelph (Kjarsgaard, 1992; Campbell et al., 1996). Dashed line labeled "Griffin (1989)" represents the original regression line (which utilized the FB86 thermometer for reference temperature) employed to calibrate the Ni thermometer (Griffin et al., 1989). Open circles are data for peridotite xenoliths from Griffin et al. (1989), with temperature recalibrated utilizing the BKN thermometer.*

## SUMMARY

Kimberlite pipes in Canada show many similarities to those from other parts of the world. Most kimberlites form carrot-shaped diatremes, however, the central Saskatchewan pipes have a flattened cone-shaped morphology. In Canada, kimberlite exploration which utilizes the indicator mineral 'pathfinder' technique must take into consideration Quaternary glacial processes. Airborne geophysical methods have been shown to be a viable exploration method in Canada. The application of 'diamond indicator mineral' chemistry studies to suggest the diamond potential of a kimberlite is a valuable tool; the Lac de Gras area pipes have a higher proportion of subcalcic chrome pyrope ('G10') garnets as compared to other kimberlite fields in Canada. This observation is consistent with the diamond potential of the Lac de Gras area, where permits have been applied for to mine five kimberlite pipes.

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# Isotopic age determinations of kimberlites and related rocks: Methods and applications.

**W.J. Davis, R.R. Parrish, J.C. Roddick, and L.M. Heaman**

*Davis, W.J., Parrish, R.R., Roddick, J.C. and Heaman, L.M., 1996: Isotopic age determinations of kimberlites and related rocks: Methods and applications; in Searching forDiamonds in Canada, A.N. LeCheminant, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 39-42.*

## INTRODUCTION

Establishing the age of individual kimberlite intrusions and clusters is key to understanding both regional and local geotectonic controls on kimberlite formation and emplacement. Kimberlite intrusions range in age from the Mesoproterozoic to the Eocene, with the vast majority being Phanerozoic. A number of peak periods of kimberlite activity are recognized worldwide (Fig. 1), and kimberlites within a particular field or craton may have different ages, indicating episodic magmatism on a local as well as global scale. Accurate determination of kimberlite emplacement ages within a particular area is necessary to establish the age range of magmatism and the number of intrusive episodes. Obtaining accurate ages can determine if kimberlites of a particular age group are more, or less, diamondiferous and the relationship of each of the age groups to local structural and tectonic controls and regional geological setting.

Comparatively few dates are available for kimberlites and related rocks in Canada, and many kimberlite fields have only one or two dated intrusions (Fig. 1). Ages range from 1.1 Ga for the Bachelor Lake ultramafic lamprophyre in Quebec (Alibert and Albarede, 1988) to 52±1.2 Ma for kimberlites in the Lac de Gras area, N.W.T. (R.O. Moore quoted in Northern Miner, 1993; Fig. 1). Detailed geochronological studies have only been reported for the Kirkland Lake area (L. Heaman, unpublished data, reported in Brummer et al., 1992), where kimberlite magmatism occurred over 8 Ma between 151 and 159 Ma. Additional detailed geochronological studies are required to better place Canadian kimberlite occurrences in their local, regional, and global geological context.

The following sections summarize the four principal isotopic dating methods currently used to obtain precise intrusion ages: Rb-Sr analyses of phlogopite ± whole rocks; K-Ar or Ar-Ar analyses of phlogopite; U-Pb analyses of perovskite; and U-Pb analyses of kimberlitic zircon. Because each of the methods has certain analytical constraints (outlined below) it is recommended that more than one method be used wherever possible. A detailed review of these methods and additional references can be found in Allsopp et al. (1989).

## Rb-Sr METHODS:

Phlogopite is present in many kimberlites and other alkaline rocks and can yield a precise intrusion age because it has elevated Rb-Sr ratio and consequently a high abundance of radiogenic Sr, even for relatively young intrusions. Phlogopite occurs both as a fine-grained groundmass phase and/or as coarser phenocrysts and/or macrocrysts in many kimberlites. Mineral separation and cleaning of groundmass phlogopite is difficult, and in many cases it may not be possible to eliminate inclusion of older xenocrystic mica, which results in erroneously old ages (Allsopp et al., 1989). Phlogopite macrocrysts are comparatively simple to separate, and analyses can be made on a single grain eliminating problems of mixing material of different ages. The principal limitation is alteration of the phlogopite and the presence of Sr-rich phases such as calcite along cleavage planes within the phlogopite. Although acid leaching techniques have been developed to try to overcome these problems (e.g. Brown et al., 1989), the most reliable results are obtained on non-altered grains. Examples of Rb-Sr phlogopite ages determined for Canadian alkaline rocks include the Cross kimberlite, southeastern B.C. (240-250 Ma, Smith et al., 1988), Sturgeon Lake kimberlite, Saskatchewan (98 ± 1 Ma , Hegner et al., 1995) and a minette from the Sweet Grass Hills area of southern Alberta (50.3 ± 0.5 Ma, Davis and Kjarsgaard, 1994).

## K-Ar and Ar-Ar METHODS:

K-Ar methods have been successfully used to date kimberlite intrusions (e.g. Kirkland Lake, Wanless et al., 1968), but this technique has limitations. It has

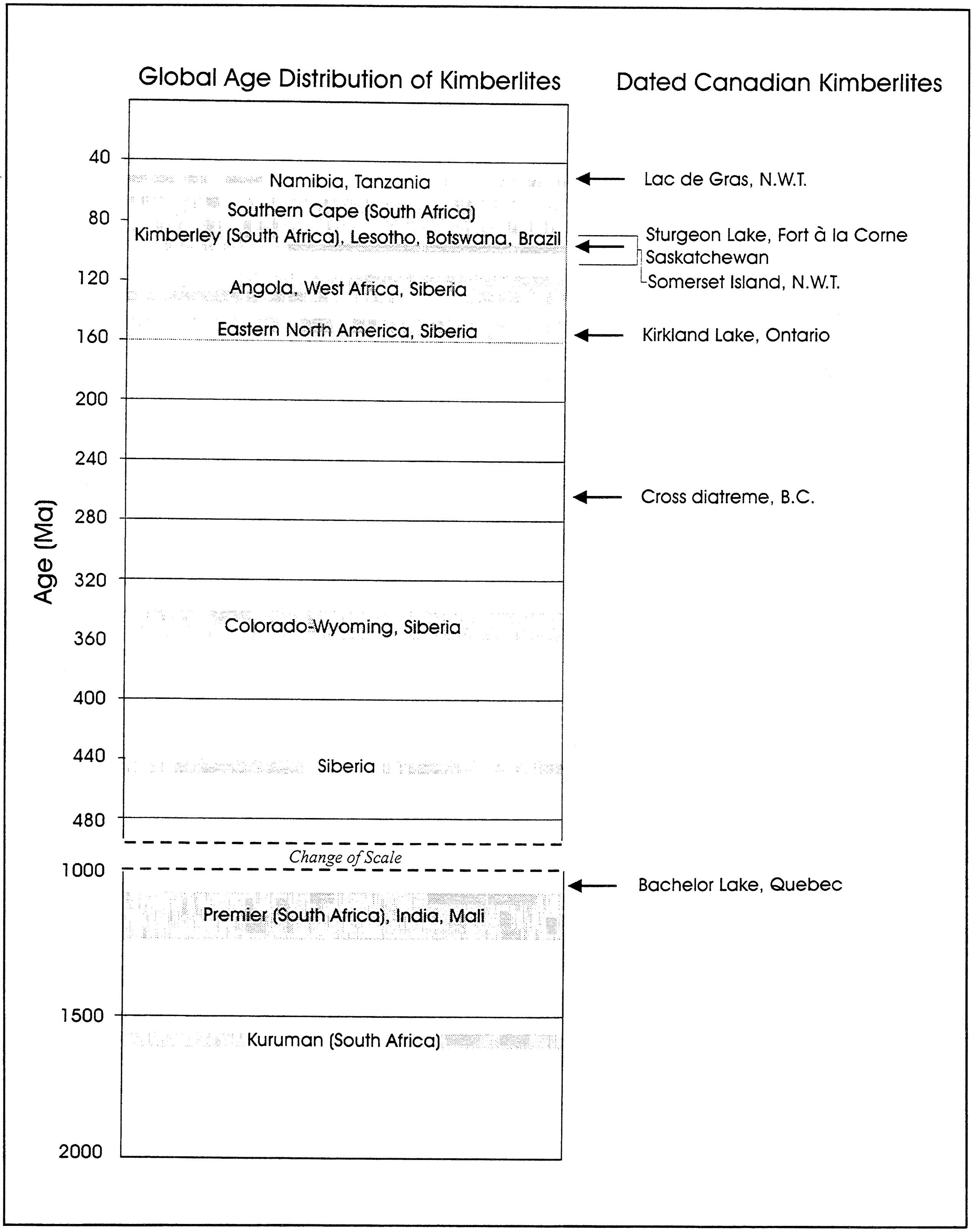


***Figure 1.*** *Comparison of the ages of dated Canadian kimberlites to the age of kimberlite provinces worldwide (after Kirkley et al., 1992). References to Canadian occurrences are given in the text.*

been shown that macrocrystic phlogopite commonly contains excess argon yielding older and incorrect $^{39}Ar$-$^{40}Ar$ ages (e.g. Allsopp and Roddick, 1984). This problem is less evident in analyses of groundmass phlogopite, however, the problem of contamination with older mica is a potential difficulty that can also produce anomalously old ages (Zartman et al., 1967). As with the Rb-Sr technique, obtaining fresh material is paramount to obtaining a reliable age.

## U-Pb METHODS:

### *Perovskite*

U-Pb analyses of perovskite, a common groundmass mineral in some kimberlites, has been successfully used to date kimberlites at Kirkland Lake and Somerset Island (Heaman, 1989; Smith et al., 1989; B.A. Kjarsgaard and L.M. Heaman, unpublished data), as well as in South Africa (Kramers and Smith, 1983). Widespread use of this method has been limited by the extremely small grain size of perovskite in most kimberlites and the resultant difficulty in separating the mineral. In addition, kimberlitic perovskite typically contains high abundances of common Pb (up to 80% of total Pb) necessitating the determination of the initial common Pb composition of the intrusion and application of a large common Pb age correction or isochron treatment of the data. In spite of these complications, precise ages can be obtained using this technique.

### *Zircon*

U-Pb analyses of kimberlitic zircon was pioneered by G.L. Davis and has been widely used to date kimberlites from the main diamond producing fields of the world (see Allsopp et al., 1989, for references). Zircons are extremely rare in kimberlites but are recovered in the heavy concentrates with diamond (Kresten and Fels, 1975). Individual grains are large (up to 20 mm) and typically have grey chalky coatings of fine grained baddeleyite ± diopside resulting from desilisification of the zircon in the mantle (Kresten, 1975; Heaman and LeCheminant, 1993). Kimberlitic zircons have characteristically low trace element contents (i.e. <10 ppm U), are believed to form within the mantle as part of the macrocryst suite (Moore et al., 1992), and are included as xenocrysts within the kimberlite during intrusion (Davis, 1977). Although kimberlitic zircons are xenocrystic in origin, most appear to retain radiogenic Pb only after extraction from the mantle during eruption. In certain cases, however, kimberlitic zircons record an age significantly older than the intrusion age, indicating that they may retain Pb for significant periods of time prior to eruption (Kinny et al., 1989).

A new application of U-Pb dating methods is the use of the ion microprobe to determine the age of zircon included within diamond (e.g. Kinny and Meyer, 1994). Although zircons included within diamonds during their growth are extremely rare they offer one of the few opportunities to obtain a maximum formation age of diamond. The installation of the SHRIMP ion microprobe at the GSC laboratory in Ottawa in 1995 will allow this type of study to be done in Canada. The ion probe may also be used to determine the $^{206}Pb$-$^{207}Pb$-$^{208}Pb$ isotopic composition of sulphides included in diamonds which can yield model dependant age information (Rudnick et al., 1993).

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# Fossils from diamondiferous kimberlites at Lac de Gras, N.W.T. : Age and paleogeography

**W.W. Nassichuk and D.J. McIntyre**

*Nassichuk, W.W. and McIntyre, D.J., 1996: Fossils from diamondiferous kimberlites at Lac de Gras, N.W.T. : Age and paleogeography; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 43-46.*

## TIME OF EMPLACEMENT OF KIMBERLITE PIPES

Diamond exploration in the Slave Province has led to the discovery of Mesozoic (Cretaceous) and Tertiary (Paleocene) fossils preserved within xenoliths of sedimentary strata in the crater and diatreme facies of 13 kimberlite pipes. Samples from most of these pipes in the Lac de Gras area were provided by the BHP-Dia Met Joint Venture, but samples from a few other pipes closer to Yamba Lake (Fig. 1) were provided by the Mill City-Tanqueray Joint Venture. The authors first indicated that fossil dinoflagellates, pollen, spores, and teleost fish remains ranging in age from latest Early Cretaceous (Albian) to early Tertiary (Paleocene) were recovered from pipes in the Lac de Gras area (report in the Northern Miner, Sept. 20, 1993). Thus, emplacement of the pipes had to postdate the youngest fossils in the pipes and

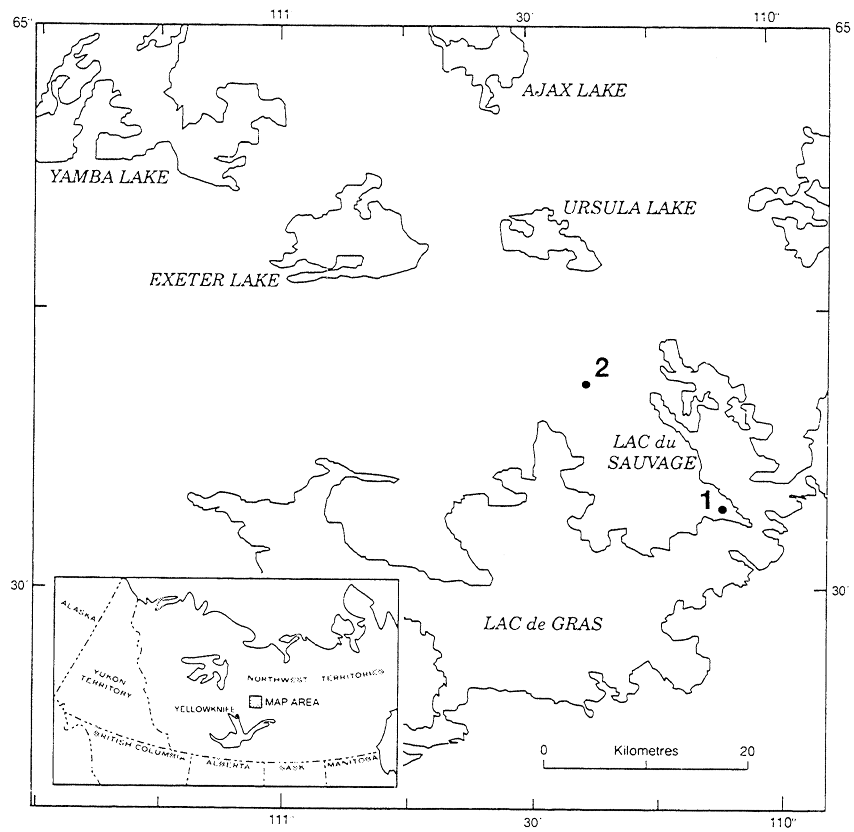


*Figure 1. Map of Lac de Gras area where fossils have been recovered from kimberlite pipes. The position of two of the pipes containing fossils that were referred to by Nassichuk and McIntyre (1995) are: 1) BHP-Dia Met Point Lake pipe, 2) BHP-Dia Met "Hawk" pipe.*

probably occurred between late Paleocene and early Eocene time according to the time scale of Harland et al. (1989). In the same Northern Miner report, R.O. Moore from BHP Minerals Canada indicated that rubidium-strontium data from mica (phlogopite) and whole-rock samples from a pipe in the Lac de Gras area suggest eruption 52±1.2 million years ago; that is, in early Tertiary (Eocene) time. Accordingly, paleontological and rubidium-strontium data for the time of emplacement (age) of the pipes are mutually supportive.

King and McMillan (1975) reported Mesozoic nannofossils from a nonkimberlitic breccia at Ford's Bight, Labrador. They concluded that the breccia is, in part, a diatreme emplaced during an episode of rifting between Labrador and Greenland that began in Jurassic time.

Paleozoic conodonts and brachiopods of undetermined Ordovician to Devonian age have been recognized in xenoliths within kimberlites east of Kirkland Lake, Ontario (H. Lee, pers. comm., 1994). Wanless et al. (1968) determined a K-Ar age of 151±8 Ma (Upper Jurassic) for phlogopites in a kimberlite dyke in the Upper Canada Mine, some 16 km east of Kirkland Lake (see also Lee, 1968). Elsewhere in the Kirkland Lake area, detailed geochronological studies have shown diatremes to range in age from 155 to 160 Ma (L.M. Heaman, unpublished data reported *in* Brummer et al., 1992). Paleozoic conodonts, corals, and thelodonts, possibly of Ordovician or Silurian age, are known from xenoliths within the Sloan kimberlite pipe in Colorado, which is thought to have been emplaced during Devonian time.

Numerous kimberlites have been discovered in Saskatchewan, but we are not aware of fossils having been recovered from xenoliths within them. Gent (1992) indicated that the Saskatchewan kimberlites penetrated Lower Cretaceous strata and that they are probably of latest Early Cretaceous (Albian) or slightly younger Cretaceous age.

Fossilized wood of early Tertiary (Paleocene) and possibly older (Cretaceous) ages is common in kimberlite pipes in the Lac de Gras area. Wood and other fossils are also rather common in kimberlites in the eastern part of the Siberian Platform. Brakhfogel (1984) reported diverse brachiopods, pelecypods, rugose and tabulate corals, ostracodes, tentaculitids, trilobites, bryozoans, acritarchs, spores, pollen, and plant parts, ranging from Ordovician to Permian in a number of pipes. He also reported abundant Lower Cretaceous carbonaceous wood (*Araucariopitys* sp.) from some pipes. Milashev and Shulgina (1959) described a Late Jurassic to Early Cretaceous belemnite (*Pachyteuthis*? sp.) from an exposed pipe in Siberia, and concluded that Siberian kimberlites occurred in two groups; one was emplaced prior to Late Permian time and the other during Cretaceous time. Additional data on the time of emplacement of other kimberlites elsewhere in the world were reported by Brummer et al. (1992).

## PALEOGEOGRAPHY

Nassichuk and McIntyre (1995) recorded marine Cretaceous and nonmarine Tertiary (Paleocene) fossils from kimberlites in the Lac de Gras area, and those fossils, some of which are shown in Figure 2, facilitate rather precise biostratigraphic correlations with successions elsewhere in northern and western Canada. Further, the newly discovered fossils are essential for refining Cretaceous and Tertiary paleogeography in northern Canada. Prior to discovery of kimberlites in the area, no data had ever been presented to prove that Mesozoic and Cenozoic rocks were deposited above Precambrian rocks in the Slave Province. Indeed, the nearest Cretaceous exposures are several hundred kilometres to the west of Lac de Gras. Now, it is clear that much of the Slave Province must have been covered by a veneer of Cretaceous and lower Tertiary strata, long since eroded away by river systems and glacial action. An important idea initiated by Bell (1895) and expanded upon by McMillan (1973) is that from late Paleocene to Pliocene time a major anastomosing river system carried sediments from the eastern Cordillera and Canadian shield areas (including the Slave Province) eastward to the Labrador Sea. New data supporting this concept of easterly flow in Tertiary time were provided by Duk-Rodkin and Hughes (1994).

The discovery of marine strata of uppermost Early Cretaceous (Albian) to latest Cretaceous (Maastrichtian) age in the Slave Province necessitates revision to paleogeographic reconstructions of the Western Interior Seaway. Nassichuk and McIntyre (1995) briefly discussed Late Cretaceous connections between the Seaway and the Sverdrup Basin in the high Arctic and concluded, on the basis of fossils from Lac de Gras, that the Seaway was considerably wider than shown by previous data.

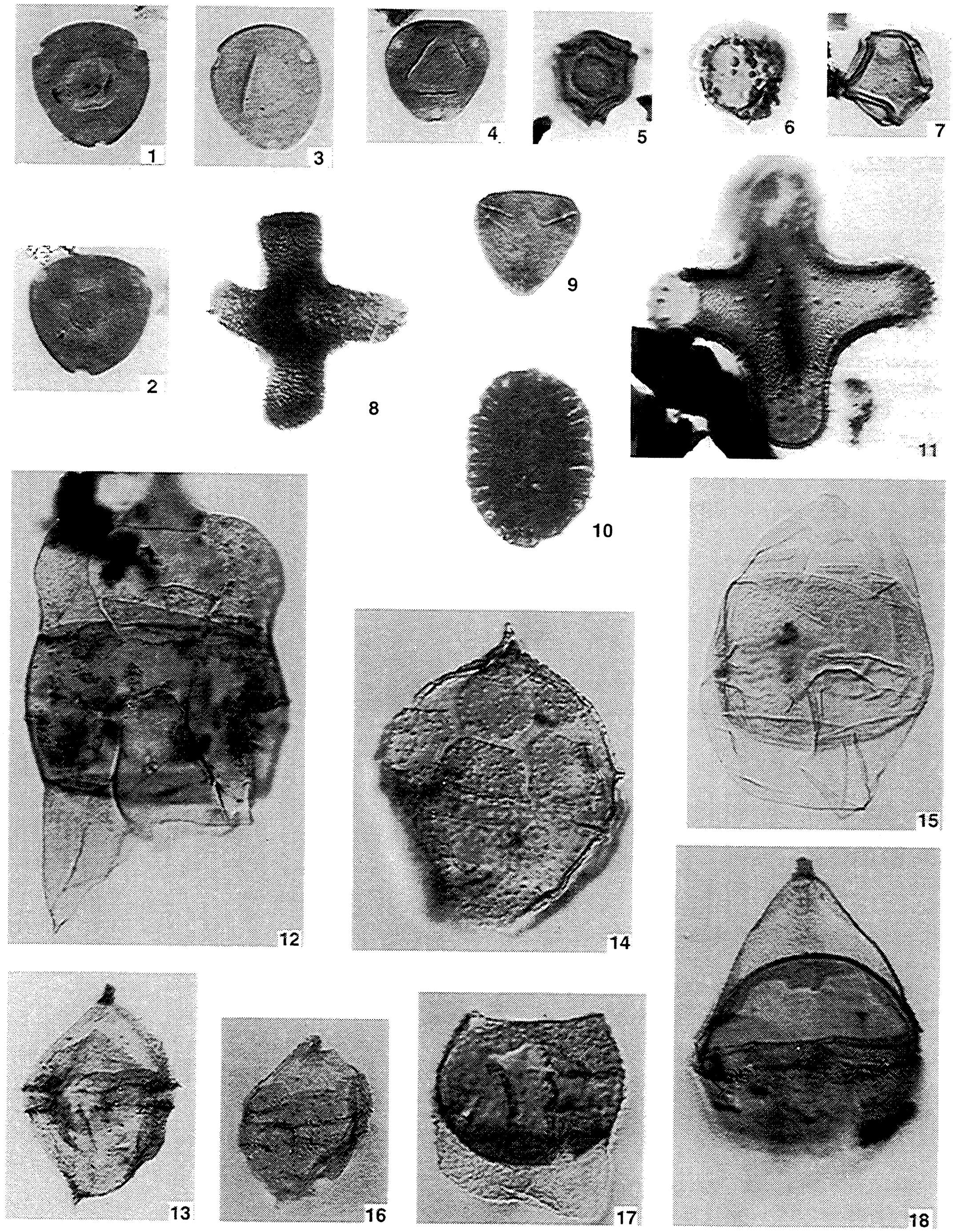


*Figure 2. Representative pollen and dinoflagellates from xenoliths in kimberlite pipes, Lac de Gras area. The slides containing the illustrated specimens are available at GSC-Calgary. Access to the specimens is provided by the GSC type number, shown in square brackets. Magnification factor for all specimens - 550x.*

*1-7: Paleocene*
1 ***Momipites anellus*** *Nichols & Ott. [GSC 109689]*
2 ***Momipites ventifluminis*** *Nichols & Ott. [GSC 109690]*
3 ***Carya*** *sp. [GSC 109691]*
4 ***Caryapollenites veripites*** *(Wilson & Webster) Nichols & Ott. [GSC 109692]*
5 ***Paraalnipollenites alterniporus*** *(Simpson) Srivastava. [GSC 109693]*
6 ***Pistillipollenites mcgregori*** *Rouse. [GSC 109694]*
7 ***Alnus sp***. *[GSC 109695]*

*8-11: Maastrichtian*
8 ***Aquilapollenites conatus*** *Tschudy. [GSC 109696]*
9 ***Myrtipites scabratus*** *Norton. [GSC 109697]*
10 ***Wodehouseia spinata*** *Stanley. [GSC 109698]*
11 ***Aquilapollenites augustus*** *Srivastava. [GSC 109699]*

*12-14: Campanian*
12 ***Chatangiella ditissima*** *(McIntyre) Lentin & Williams. [GSC 109700]*
13 ***Laciniadinium biconiculum*** *McIntyre. [GSC 109701]*
14 ***Ginginidinium ornatum*** *(Felix & Burbridge) Lentin & Williams. [GSC109702]*

*15: Turonian*
15 ***Williamsidinium banksianum*** *Lentin. [GSC 109703]*

*16-18 Cenomanian and Late Albian*
16 ***Ginginidinium evittii*** *Singh. [GSC 109704]*
17 ***Ovoidinium verrucosum*** *(Cookson & Hughes) Davey. [GSC109705]*
18 ***Luxadinium propatulum*** *Brideaux & McIntyre. [GSC 109706]*

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# Tools of investigation: The electron microprobe and scanning electron microscope

**J.A.R. Stirling and G.J. Pringle**

*Stirling, J.A.R. and Pringle, G.J., 1996: Tools of investigation: The electron microprobe and scanning electron microscope; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 47-53.*

## INTRODUCTION

Mineral chemistry has been essential for the development of petrogenetic models used for diamond exploration (Mitchell, 1995), and the chemistry of indicator minerals recovered from glacial deposits is a key exploration guide for diamond-bearing kimberlites in Canada (DiLabio, 1996). Mineral compositions can be determined qualitatively on a scanning electron microscope (SEM) equipped with an energy dispersive X-ray analyzer. The electron microprobe is the most frequently used analytical instrument for quantitative major and trace element analyses by energy-dispersive or wavelength dispersive X-ray spectrometry, and the proton microprobe is used for high precision trace element studies (Griffin and Ryan, 1995). In order to properly interpret or compare microprobe results it is necessary to understand all the inherent analytical constraints that are introduced by the choice of analytical scheme and by the nature of the sample. This paper discusses the impact of analytical constraints on the validity of indicator mineral analyses used for diamond exploration and for petrogenetic studies.

## DISCUSSION

### *Inherent analytical constraints*

Samples must be polished to the highest quality possible and kept free of dirt and oils. In the case of grain mounts the epoxy must be free of air bubbles and, for efficiency, the grains should be arranged in an orderly grid. Deviation from these conditions can adversely effect the quality of the analyses.

When planning the analytical approach and interpreting microprobe data it is absolutely essential to have detailed discussions with the analyst. The conditions used during analysis have a major affect on the confidence that can be placed on a given value, and can significantly affect the time required to complete the work.

At the GSC, analytical schemes have been developed for particular requirements. A major element analysis scheme ("SILPYROX", Table 1) will not necessarily work for trace element analysis ("DIAMOND", Table 2). For example, when analyzing a garnet it is possible to obtain a calculated minimum detection limit (MDL, Table 3) of 0.004 wt% for $Na_2O$ when using the "DIAMOND" trace element analysis scheme. This requires a beam current of 200 nA (nanoamperes) and a peak counting time of 100 seconds, whereas, in the case of the "SILPYROX" scheme for major element analysis the MDL for $Na_2O$ is only 0.04 wt.%. If the trace element analytical conditions were applied to a less stable mineral, like feldspar, elements like the alkalis would become unstable, and the results would be meaningless. Sodium is a prime example of an element where different schemes are essential for each mineral.

Element interferences must be considered. For example, in analyzing ilmenite the MDL calculated for $V_2O_5$ may be about 0.05 wt%, however, interference from Ti can result in an apparent concentration of up to 0.2 wt% $V_2O_5$. To compensate for this, a peak overlap correction must be applied at the time of analyses. A similar problem occurs with interference between Na and Zn. A mineral with high Zn, such as the zinc spinel, gahnite, will have an apparent $Na_2O$ concentration of as much as 10 wt%. At the GSC, we have developed peak overlap corrections for the Cameca SX50 microprobe; they are a specialized option that can be invoked when required for each analytical scheme.

### *Diamond exploration*

It is well known that minerals such as garnet and chromite, associated with diamond-bearing kimberlites, have a unique chemistry (Griffin and Ryan, 1995; Gurney and Zweistra, 1995). For indicator mineral studies (Ward et al., 1996; Garrett and Thorleifson, 1996; McClenaghan, 1996), polished blocks of hand-picked

**Table 1:** The analytical conditions and standards for the major element scheme "SILPYROX" which is used to analyze most garnets, pyroxenes, and olivines.

**CALIBRATION DATA**

| ELEMENT | SPC | XTAL | POS. | +BG OFFSET | -BG. OFFSET | BG SLOPE | PK-BG C/s/nA | SIGMA | PK_TIM sec. | % REQ. ACCUR. | BG_TIM ms |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Na | 1 | PCO | 26851 | 2000 | -2000 | 0.00 | 1414.96 | 0.8 | 20 | 0.1 | 10000 |
| K | 3 | PET | 42746 | 512 | 0 | 1.00 | 193.63 | 0.1 | 20 | 0.1 | 10000 |
| Fe | 4 | LIF | 48082 | 512 | 0 | 1.00 | 146.12 | 0.6 | 20 | 0.1 | 10000 |
| Mg | 2 | TAP | 38506 | 912 | 0 | 1.00 | 943.54 | 0.2 | 20 | 0.1 | 10000 |
| Si | 2 | TAP | 27737 | 1531 | 0 | 1.00 | 972.57 | 0.1 | 10 | 0.1 | 5000 |
| Ca | 3 | PET | 38379 | 531 | 0 | 1.00 | 213.78 | 0.3 | 10 | 0.1 | 5000 |
| Mn | 4 | LIF | 52204 | 631 | 0 | 1.00 | 47.54 | 3.1 | 10 | 0.1 | 5000 |
| Ti | 3 | PET | 31427 | 831 | 0 | 1.00 | 414.45 | 0.3 | 10 | 0.1 | 5000 |
| Cr | 4 | LIF | 56876 | 531 | 0 | 1.00 | 41.64 | 0.4 | 10 | 0.1 | 5000 |
| Al | 2 | TAP | 32467 | 1431 | 0 | 1.00 | 1064.40 | 0.4 | 10 | 0.1 | 5000 |
| Ni | 4 | LIF | 41167 | 531 | 0 | 1.00 | 116.95 | 6.3 | 10 | 0.1 | 5000 |

**Abbreviations:**
SPC = spectrometer number; XTAL = spectrometer crystal; POS. = position of spectrometer; +BG OFFSET = positive background offset; -BG OFFSET = negative background offset; BG SLOPE = background slope; PK-BG C/s/nA = peak minus background in counts/second/nanoamperes; PK_TIM Sec. = peak counting time in seconds; % REQ. ACCUR. = percent required accuracy for peak counting time; BG_TIM ms = background counting time in milliseconds.

**STANDARD DATA**

| ELEMENT | STANDARD | WEIGHT FRACTION | LINE | kV | Beam Current nA |
|---|---|---|---|---|---|
| Na | NACL7 | 0.3930 | Ka | 15.0 | 10.0 |
| K | KBR7 | 0.3290 | Ka | 15.0 | 10.0 |
| Fe | MAG1 | 0.7236 | Ka | 15.0 | 10.0 |
| Mg | MGO1 | 0.6032 | Ka | 15.0 | 10.0 |
| Si | QTZ1 | 0.4674 | Ka | 15.0 | 30.0 |
| Ca | WOL1 | 0.3432 | Ka | 15.0 | 30.0 |
| Mn | RHOD1 | 0.2854 | Ka | 15.0 | 30.0 |
| Ti | RUT1 | 0.5895 | Ka | 15.0 | 30.0 |
| Cr | CHR1 | 0.2504 | Ka | 15.0 | 30.0 |
| Al | COR1 | 0.5290 | Ka | 15.0 | 30.0 |
| Ni | NIS4 | 0.6457 | Ka | 15.0 | 30.0 |

**STANDARD COMPOSITION**

| STANDARD | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| NACL7 | Na | 0.3930 | Cl | 0.6070 | | | | | | | | |
| KBR7 | K | 0.3290 | Br | 0.6710 | | | | | | | | |
| MAG1 | Fe | 0.7236 | O | 0.2764 | | | | | | | | |
| MGO1 | Mg | 0.6032 | O | 0.3968 | | | | | | | | |
| QTZ1 | Si | 0.4674 | O | 0.5326 | | | | | | | | |
| WOL1 | Ca | 0.3432 | Fe | 0.0030 | Mn | 0.0012 | Si | 0.2399 | O | 0.4127 | | |
| RHOD1 | Mn<br>Na | 0.2854<br>0.0010 | Al<br>K | 0.0003<br>0.0007 | Fe<br>Zn | 0.0131<br>0.0574 | Si<br>O | 0.2178<br>0.3722 | Mg | 0.0052 | Ca | 0.0469 |
| RUT1 | Ti | 0.5895 | Fe | 0.0050 | Nb | 0.0050 | O | 0.4005 | | | | |
| CHR1 | Cr<br>Mn | 0.2504<br>0.0015 | Al<br>Ni | 0.0762<br>0.0012 | Fe<br>Si | 0.2985<br>0.0011 | Mg<br>O | 0.0434<br>0.3211 | Ti | 0.0054 | V | 0.0012 |
| COR1 | Al | 0.5290 | O | 0.4710 | | | | | | | | |
| NIS4 | Ni | 0.6457 | Fe | 0.0010 | S | 0.3533 | | | | | | |

**Table 2:** The analytical conditions and standards for the trace element scheme "DIAMOND" which is used to analyze trace elements in most garnets, pyroxenes, and olivines.

**CALIBRATION DATA**

| ELEMENT | SPC | XTAL | POS. | +BG OFFSET | -BG. OFFSET | BG SLOPE | PK-BG C/s/nA | SIGMA | PK_TIM sec. | % REQ. ACCUR. | BG_TIM ms |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Na | 1 | PCO | 26847 | 1000 | -1000 | 0.00 | 1601.61 | 0.5 | 100 | 0.1 | 5000 |
| Si | 2 | TAP | 27735 | 1500 | 0 | 1.00 | 1301.34 | 0.1 | 5 | 0.1 | 5000 |
| Ke | 3 | PET | 42746 | 1000 | 0 | 1.00 | 294.60 | 0.1 | 10 | 0.5 | 3416 |
| Fe | 4 | LIF | 48082 | 1050 | 0 | 1.00 | 324.54 | 0.2 | 10 | 0.1 | 5000 |
| Al | 2 | TAP | 32470 | 1000 | 0 | 1.00 | 1374.43 | 0.1 | 5 | 0.1 | 5000 |
| Ca | 3 | PET | 38380 | 2000 | 0 | 1.00 | 367.72 | 0.4 | 5 | 0.1 | 5000 |
| Mn | 4 | LIF | 52207 | 1200 | 0 | 1.00 | 420.37 | 1.0 | 10 | 0.5 | 2421 |
| Mg | 2 | TAP | 38509 | 1000 | 0 | 1.00 | 1124.46 | 0.2 | 5 | 0.1 | 5000 |
| Ti | 3 | PET | 31425 | 1000 | 0 | 1.00 | 765.32 | 0.2 | 10 | 0.1 | 5000 |
| Cr | 3 | PET | 26198 | 1000 | 0 | 1.00 | 362.02 | 0.4 | 10 | 0.1 | 5000 |
| Ni | 4 | LIF | 41165 | 1000 | 0 | 1.00 | 495.25 | 0.7 | 100 | 0.5 | 2063 |
| Zn | 4 | LIF | 35622 | 1000 | 0 | 1.00 | 398.88 | 0.1 | 10 | 0.1 | 5000 |
| Nb | 3 | PET | 65412 | 2000 | 0 | 1.00 | 125.08 | 0.5 | 10 | 0.5 | 5000 |
| Co | 4 | LIF | 44427 | 1000 | 0 | 1.00 | 504.44 | 0.2 | 10 | 0.5 | 2025 |
| V | 4 | LIF | 62188 | 800 | 0 | 1.00 | 248.34 | 0.4 | 10 | 0.1 | 5000 |

Abbreviations: provided in Table 1.

**STANDARD DATA**

| ELEMENT | STANDARD | WEIGHT FRACTION | LINE | kV | Beam Current nA |
|---|---|---|---|---|---|
| Na | NACL7 | 0.3930 | Ka | 20.0 | 200.0 |
| Si | QTZ1 | 0.4674 | Ka | 20.0 | 200.0 |
| K | KBR7 | 0.3290 | Ka | 20.0 | 200.0 |
| Fe | MAG1 | 0.7236 | Ka | 20.0 | 200.0 |
| Al | COR1 | 0.5290 | Ka | 20.0 | 200.0 |
| Ca | WOL1 | 0.3432 | Ka | 20.0 | 200.0 |
| Mn | MN | 1.0000 | Ka | 20.0 | 200.0 |
| Mg | MGO1 | 0.6032 | Ka | 20.0 | 200.0 |
| Ti | RUT1 | 0.5895 | Ka | 20.0 | 200.0 |
| Cr | CHR1 | 0.2504 | Ka | 20.0 | 200.0 |
| Ni | NI | 1.0000 | Ka | 20.0 | 200.0 |
| Zn | ZN | 1.0000 | Ka | 20.0 | 200.0 |
| Nb | NB | 1.0000 | Ka | 20.0 | 200.0 |
| Co | CO | 1.0000 | Ka | 20.0 | 200.0 |
| V | V | 1.0000 | Ka | 20.0 | 200.0 |

**STANDARD COMPOSITION**

| STANDARD | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION | Element | WEIGHT FRACTION |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| NACL7 | Na | 0.3930 | Cl | 0.6070 | | | | | | | | |
| OTZ1 | Si | 0.4674 | O | 0.5326 | | | | | | | | |
| KBR7 | K | 0.3290 | Br | 0.6710 | | | | | | | | |
| MAG1 | Fe | 0.7236 | O | 0.2764 | | | | | | | | |
| COR1 | Al | 0.5290 | O | 0.4710 | | | | | | | | |
| WOL1 | Ca | 0.3432 | Fe | 0.0030 | Mn | 0.0012 | Si | 0.2399 | O | 0.4127 | | |
| MN | Mn | 1.0000 | | | | | | | | | | |
| MGO1 | Mg | 0.6032 | O | 0.3968 | | | | | | | | |
| RUT1 | Ti | 0.5895 | Fe | 0.0050 | Nb | 0.0050 | O | 0.4005 | | | | |
| CHR1 | Cr<br>Mn | 0.2504<br>0.0015 | Al<br>Ni | 0.0762<br>0.0012 | Fe<br>Si | 0.2985<br>0.0011 | Mg<br>O | 0.0434<br>0.3211 | Ti | 0.0054 | V | 0.0012 |
| NI | Ni | 1.0000 | | | | | | | | | | |
| ZN | Zn | 1.0000 | | | | | | | | | | |
| NB | Nb | 1.0000 | | | | | | | | | | |
| CO | Co | 1.0000 | | | | | | | | | | |
| V | V | 1.0000 | | | | | | | | | | |

**Table 3.** Comparison of results from triplicate analyses of the same garnet crystal using the major element scheme "SILPYROX" and the trace element scheme "DIAMOND"

**"SILPYROX"**

| OXIDES | SAMPLE #3 PT1 (wt %) | SAMPLE #3 PT2 (wt %) | SAMPLE #3 PT3 (wt %) | MEAN (wt %) | 95% STD | ERROR % | MDL (wt %) |
|---|---|---|---|---|---|---|---|
| $Na_2O$ | .01 | .00 | .00 | .00 | .00 | 282.84 | .04 |
| $K_2O$ | .02 | .00 | .01 | .01 | .02 | 184.52 | .05 |
| FeO | 8.66 | 8.85 | 8.52 | 8.68 | .27 | 3.10 | .13 |
| MgO | 20.79 | 20.92 | 20.71 | 20.81 | .18 | .86 | .04 |
| $SiO_2$ | 41.71 | 41.45 | 41.22 | 41.46 | .39 | .95 | .02 |
| CaO | 4.68 | 4.70 | 4.68 | 4.69 | .01 | .30 | .04 |
| MnO | .35 | .38 | .28 | .34 | .09 | 25.53 | .09 |
| $TiO_2$ | .34 | .32 | .35 | .34 | .02 | 7.11 | .05 |
| $Cr_2O_3$ | 1.01 | .95 | 1.02 | .99 | .06 | 5.92 | .09 |
| $Al_2O_3$ | 22.92 | 22.72 | 22.61 | 22.75 | .25 | 1.11 | .02 |
| NiO | .00 | .00 | .00 | .00 | .00 | N/A | .13 |
| TOTAL | 100.50 | 100.30 | 99.40 | 100.07 | .95 | .95 | N/A |

**"DIAMOND"**

| OXIDES | SAMPLE #3 PT1 (wt %) | SAMPLE #3 PT2 (wt %) | SAMPLE #3 PT3 (wt %) | MEAN (wt %) | 95% STD | ERROR % | MDL (wt %) |
|---|---|---|---|---|---|---|---|
| $Na_2O$ | .03 | .04 | .03 | .033 | .004 | 13.09 | .004 |
| $K_2O$ | .00 | .01 | .01 | .005 | .007 | 141.42 | .008 |
| FeO | 8.65 | 8.71 | 8.69 | 8.683 | .049 | .56 | .014 |
| MgO | 20.60 | 20.63 | 20.70 | 20.645 | .085 | .41 | .009 |
| $SiO_2$ | 41.39 | 41.43 | 41.12 | 41.312 | .274 | .66 | .008 |
| CaO | 4.63 | 4.64 | 4.60 | 4.624 | .031 | .67 | .007 |
| MnO | .34 | .36 | .35 | .350 | .019 | 5.38 | .015 |
| $TiO_2$ | .33 | .34 | .33 | .331 | .010 | 2.98 | .010 |
| $Cr_2O_3$ | 1.01 | .99 | .99 | .998 | .013 | 1.33 | .009 |
| $Al_2O_3$ | 22.45 | 22.44 | 22.53 | 22.475 | .083 | .37 | .008 |
| NiO | .01 | .01 | .02 | .015 | .004 | 28.80 | .005 |
| ZnO | .01 | .00 | .02 | .010 | .017 | 172.05 | .023 |
| $Nb_2O_5$ | .07 | .00 | .02 | .029 | .058 | 198.42 | .036 |
| CoO | .02 | .01 | .01 | .013 | .008 | 62.81 | .013 |
| $V_2O_3$ | .04 | .04 | .05 | .042 | .012 | 29.93 | .031 |
| TOTAL | 99.57 | 99.64 | 99.48 | 99.564 | .125 | .13 | N/A |

N/A = Not applicable.
MDL = Minimum detection limit as calculated by the method discussed by Harris (1990).

minerals such as ilmenite, spinel, pyroxene, and garnet are prepared. Each block may contain several hundred grains and a complete project can involve several thousand grains, requiring hundreds of hours of microprobe time. In order to accomplish this task in the shortest possible time certain compromises must be made.

Specialized sample holders were developed at the GSC to accommodate polished thin sections and polished blocks in the SEM (Walker and LeCheminant, 1989). Backscattered electron images (BSE) at low magnification are prepared with the SEM, and these images are used as maps when the grains are selected for microprobe analysis. The location of each grain is selected "off line" using a computer-controlled stage on an optical microscope and the co-ordinates are transferred to the computer which controls the microprobe. An experienced operator can log 700 points/hour.

To optimize the time required for analysis, decisions must be made as to what elements are to be analyzed and the detection levels required. It is possible to analyze each grain for major and trace elements, however, this would require at least twice as much time as our recommended procedure. We currently suggest that the optimum setup for major element analysis be used initially for each grain. The resulting data can then be processed to identify potential indicator minerals. In order to recognize diamond indicator compositions in a large data set, several types of sorting must be done on the data. First, the results are imported into a data-base program, then criteria are established to sort the data into mineral groups, such as garnet or pyroxene. The mineral stoichiometry is then calculated permitting minerals in each group to be further classified into mineral species. A program called "MINREP" (Pringle,1995) has been developed to calculate stoichiometry and weight percent end members (Table 4). Data are then re-entered into the data-base, and criteria are established to identify and group potential indicator grains. In many cases, trace element determinations are then required to confirm an indicator grain. These grains are re-analyzed using conditions optimized for trace elements.

At this stage, only one analytical point has been determined, generally at the centre of each grain, and there are many factors which could produce a non-representative analysis. Typically, the number of grains which have been identified as indicators is small. Consequently, it is practical to examine the BSE image of each grain to ensure that it is correctly identified, and that the analysis is not from a zoned or polymineralic grain, or an artifact caused by a grain boundary, poor sample preparation, or the presence of an inclusion. When the geologist is satisfied about the identity and condition of the grain, then it can be re-analyzed for major and trace elements at optimum operating conditions, at more than one point if required.

### *Petrogenetic studies*

Polished thin sections are usually used for petrogenetic studies of kimberlites (Kjarsgaard, 1996), although when the geologist is interested in a particular mineral, such as garnet or chromite, a polished grain mount may be prepared from a mineral concentrate. A thorough petrographic study is done using an optical microscope, before the sample is examined with the SEM. Qualitative analyses on the SEM complement mineral identifications made using a petrographic microscope, since proper classification of many kimberlites and related rocks requires accurate identification of minor, fine-grained or altered matrix minerals (Mitchell, 1995).

The SEM has more imaging capabilities than the microprobe; particularly at low and high magnifications. BSE images can be used as maps for selecting spot locations for quantitative analysis, and if the BSE image shows complex zoning patterns it can be used to select points for analysis to determine the composition of these zones. After the sample is thoroughly examined with the SEM, quantitative analysis can be performed on selected minerals with the microprobe. Since the number of analyses required for a petrogenetic study are fewer than in a till indicator mineral project, the time constraints are different. The selection of material to be analyzed is more rigorous and the analytical scheme used for each analysis may be more complete. For example, the analysis of major and trace elements may be made at the same time with more complex and time consuming analytical schemes. The location of points for analysis may be selected off line or interactively depending on the nature of the study. The results of the study are processed with the "MINREP" program and made available digitally for input into programs used to calculate parameters such as the pressures and temperatures of mineral formation.

## CONCLUSIONS

The scanning electron microscope and electron microprobe provide a powerful combination for characterizing mineral parageneses and determining chemical compositions. To achieve the best results on the microprobe, the operating conditions must be set for the problem at hand. It is essential to consult with an experienced microprobe operator when designing

**Table 4.** Calculation of the microprobe analyses into mineral stoichiometry, and weight and molecular percent end members. Note that the FeO and $Fe_2O_3$ have been calculated from the analyzed total Fe proportioned as required by the stoichiometry.

**"SILPYROX"**

| Oxides | weight % | Numbers of ions on the basis of 24 O | | |
|---|---|---|---|---|
| $SiO_2$ | 41.46 | Si | 5.875 | |
| $TiO_2$ | .34 | Ti | .036 | 5.911 |
| | | | .000 | |
| $Al_2O_3$ | 22.75 | Al | 3.799 | |
| $Cr_2O_3$ | .99 | Cr | .111 | |
| $Fe_2O_3$ | 2.52 | $Fe^{3+}$ | .269 | 4.179 |
| FeO | 6.41 | $Fe^{2+}$ | .760 | |
| MgO | 20.81 | Mg | 4.395 | |
| MnO | .34 | Mn | .041 | 5.910 |
| CaO | 4.69 | Ca | .712 | |
| $Na_2O$ | .00 | Na | .000 | |
| $K_2O$ | .01 | K | .002 | |
| TOTAL | 100.32 | | | 16.000 |

***GARNET END MEMBERS***

| | weight % | molecular % |
|---|---|---|
| Almandine | 14.81 | 12.86 |
| Andradite | 8.03 | 6.83 |
| Grossular | 2.50 | 2.40 |
| Pyrope | 69.37 | 74.39 |
| Spessartine | .79 | .69 |
| Uvarovite | 3.26 | 2.82 |
| TOTAL | 98.77 | 99.99 |

**"DIAMOND"**

| Oxides | weight % | Numbers of ions on the basis of 24 O | | |
|---|---|---|---|---|
| $SiO_2$ | 41.312 | Si | 5.892 | |
| $TiO_2$ | .331 | Ti | .035 | 5.927 |
| | | | .000 | |
| $Al_2O_3$ | 22.475 | Al | 3.778 | |
| $Cr_2O_3$ | .998 | Cr | .113 | |
| $Fe_2O_3$ | 2.474 | $Fe^{3+}$ | .266 | 4.157 |
| FeO | 6.457 | $Fe^{2+}$ | .770 | |
| MgO | 20.645 | Mg | 4.388 | |
| MnO | .350 | Mn | .042 | 5.917 |
| CaO | 4.624 | Ca | .707 | |
| $Na_2O$ | .033 | Na | .009 | |
| $K_2O$ | .005 | K | .001 | |
| TOTAL | 99.704 | | | 16.001 |

***GARNET END MEMBERS***

| | weight % | molecular % |
|---|---|---|
| Almandine | 14.91 | 13.04 |
| Andradite | 7.89 | 6.75 |
| Grossular | 2.43 | 2.34 |
| Pyrope | 68.82 | 74.28 |
| Spessartine | .81 | .71 |
| Uvarovite | 3.30 | 2.872 |
| TOTAL | 98.15 | 99.99 |

methodologies for the study, and to understand the analytical constraints inherent in the results when interpreting or comparing microprobe data, particularly if trace elements analyses are involved.

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# Slave Province kimberlites, N.W.T.

**B.A. Kjarsgaard**

*Kjarsgaard, B.A., 1996: Slave Province kimberlites, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 55-60.*

## INTRODUCTION

In September 1991, the first kimberlite in the Slave Structural Province (Fig. 1) was discovered in the Lac de Gras area. This discovery culminated a decade-long search that initially followed kimberlite indicator minerals in trunk eskers 'up ice' to delineate a highly prospective area. The discovery sparked a staking rush, said to be the largest in Canadian history. At present, there are 75 confirmed kimberlite pipes (with over 100 rumoured to exist), of which 35 are diamond-bearing (Schiller, 1994). The locations of a number of these pipes are shown in Figure 2, which illustrates that most of the pipes are in the central or southern regions of the Slave Province. Exploration continues over most of the region, and recent till sampling for indicator minerals suggests that there is high potential for kimberlites to be discovered in the northern part of the Slave Province (e.g. the Coppermine area; Jennings and Barker, 1995).

## GEOLOGY OF THE ARCHEAN SLAVE PROVINCE

The Archean Slave Province has been the subject of extensive bedrock mapping and related scientific studies. However, its overall tectonomagmatic evolution remains controversial, for instance compare the variations in the intracratonic models of Henderson, (1981), Thompson, (1989) or Padgham, (1992) with the subduction-type models of Fyson and Helmstaedt, (1988), King et al., (1992) or Kusky (1991). Improved understanding of Slave tectonomagmatism requires additional data on the absolute timing of metamorphism and deformation. As is typical of Archean cratons, a wide variety of lithologies of different age occur within the Slave Province. At the surface, 65% of the rocks are granitoid and 35% supracrustal (Fyson and Padgham, 1993). However, more than half the surface area consists of Late Archean granitoids (2.70 to 2.55 Ga; van Breemen et al., 1992; Davis et al., 1994). Most of the supracrustal rocks in the Slave Province have been dated at 2.715 to 2.655 Ga, although both earlier (3.15 Ga) and younger (<2.615 Ga) supracrustal rocks have been recognized (Villeneuve and van Breemen, 1995). Isotopic studies (Nd: Davis and Hegner, 1992; Pb: Thorpe et al.,1992; Davis et al., in press) suggest that the Slave Province can be subdivided into two domains (Fig. 1). In the western half, which contains inliers of 4.0 to 2.8 Ga gneisses, granitoids and supracrustals, the late Archean granitoids have mixed juvenile (Late Archean; <2.7 Ga) plus recycled Early to Mid Archean components. Late Archean (<2.7 Ga) granitoids in the eastern half, however, have isotopic signatures consistent with juvenile mantle extraction ages (<2.8 Ga). However, on the basis of new Pb isotopic data (Davis et al., in press), it may be more useful to consider the Slave Province in terms of multiple regions. Post-Archean magmatism in the Slave Province is limited to minor alkaline continental intraplate volcanism (e.g. Big Spruce Lake complex; Currie, 1976) and intrusion of diabase dyke swarms (LeCheminant et al., 1995). No relationship has been discerned between the distribution of kimberlites and Precambrian basement geology; kimberlites have been discovered in both the western and eastern domains of the Slave Province (Fig. 2).

## KIMBERLITE GEOLOGY

Little public information is available about the Slave Province kimberlite pipes. Kimberlites in the Lac de Gras area form typical carrot-shaped diatremes identical in shape to those known in South Africa (M. Kirkley, oral presentation, Yellowknife Geoscience Forum, December, 1994), with steep (≈80°) country rock/diatreme contacts. Crater, diatreme, and hypabyssal facies kimberlite have been identified. Although the Lac de Gras area pipes are similar in geometry to diatremes in South Africa, the majority of the pipes discovered so far are quite small, with surface areas of <10 ha (M. Kirkley, op. cited). These small kimberlite pipes (Fig. 3; Kjarsgaard, 1996, this volume) are however, quite similar to a number of the economic pipes in Yakutia. Bedrock mapping by the Geological Survey of Canada in the Lac de Gras area (Kjarsgaard and Wyllie, 1993, 1994; Kjarsgaard et al., 1994a, b) resulted in the discovery of two previously unknown kimberlites north of Paul Lake. Two kimberlite pipe emplacement ages are known at present for the Lac de Gras area: Cretaceous (73 to 75 Ma, U-Pb perovskite; L. Heaman, pers. comm., 1995) for the C-13 pipe, and

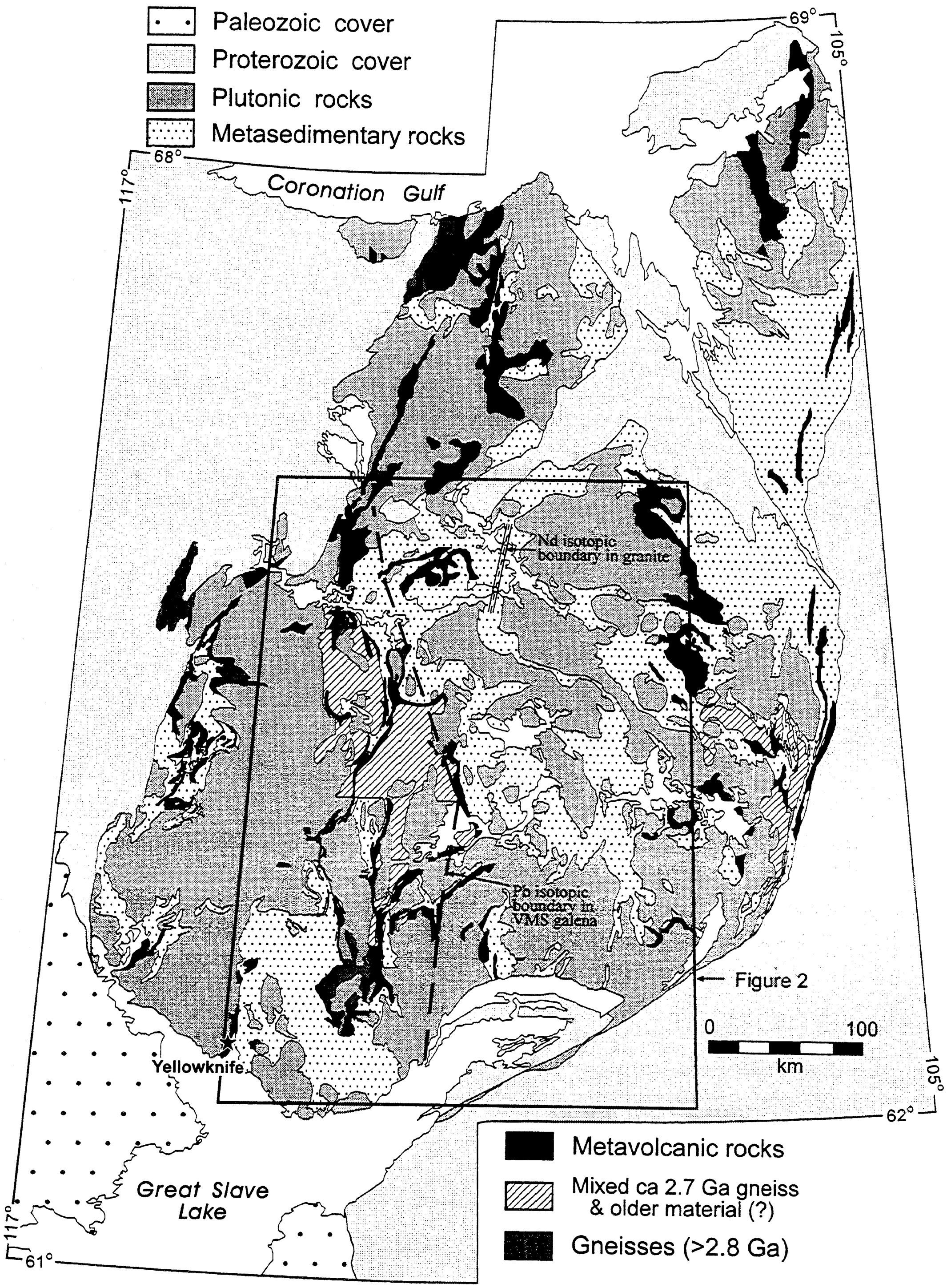


*Figure 1. Simplified geology of the Slave Province. Heavy line labelled 'Pb' is the Pb isotope line of Thorpe et al., (1992) which subdivides the Slave province into eastern and western domains. Line labelled 'Nd' is the Nd isotopic line of Davis and Hegner (1992). Box is the enlarged area shown in Figure 2.*

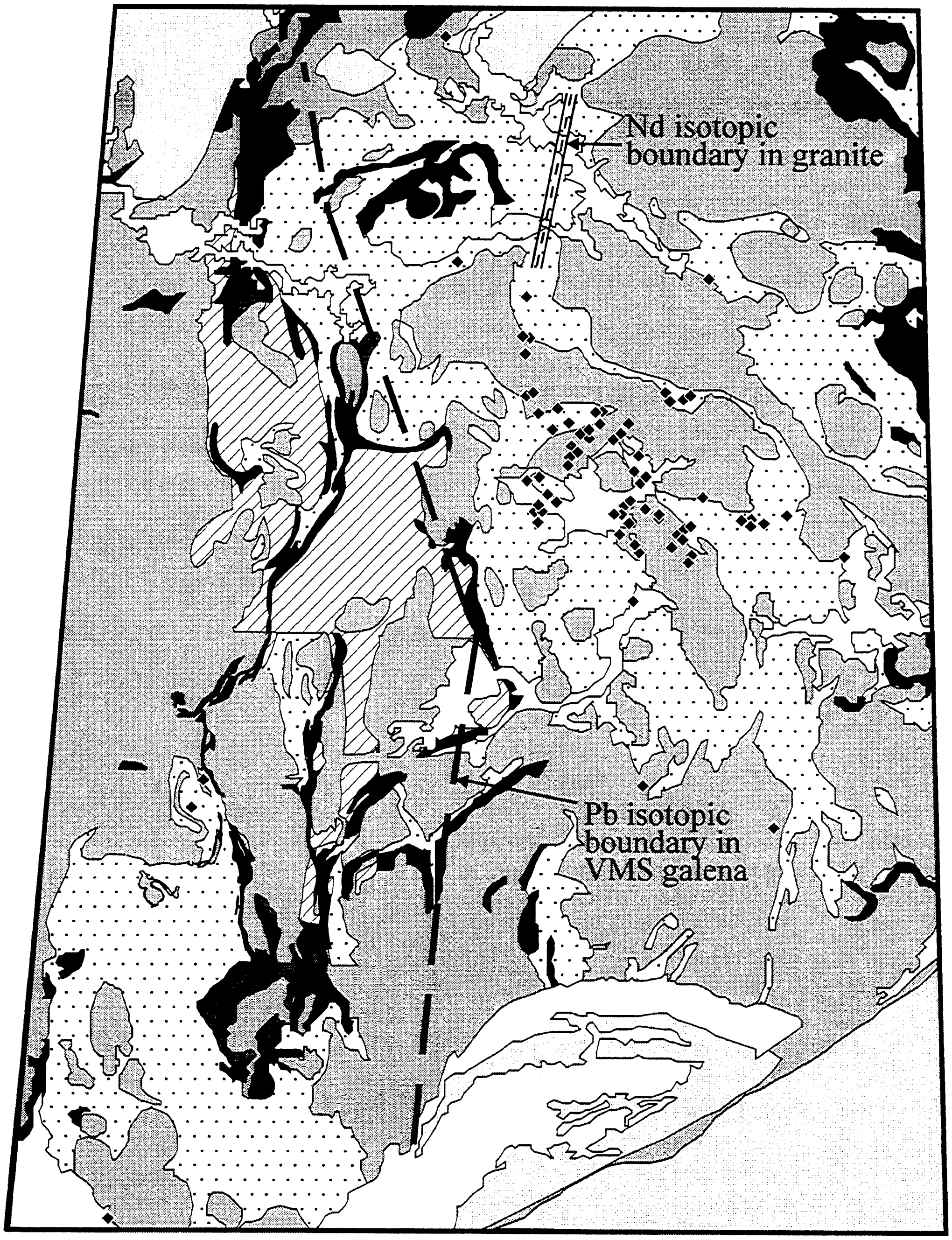


*__Figure 2.__ Distribution of kimberlites in the Slave Province. Kimberlite locations from Pell (1995).*

Eocene (52 ± 1.2 Ma, Rb-Sr phlogopite/whole-rock isochron) for an unspecified pipe on the BHP/Dia Met property (Northern Miner, 1993). W.J. Davis and B.A. Kjarsgaard (unpublished data) established a Lutetian (middle Eocene) age by the Rb-Sr (2 phlogopite separates plus whole-rock) method for a kimberlite on the BHP/Dia Met property.

The distribution of kimberlite pipes in the central Slave Province (Fig. 2) is similar to that observed in other kimberlite fields i.e. all the pipes follow a main trend (NNW), with pipe clusters orthogonal to the main trend (NNE and ENE). Age determinations on Slave Province kimberlites illustrate there are at least three distinct emplacement periods (Ordovician, Cretaceous and Eocene). Kjarsgaard and Heaman (1995) noted that the central Slave Province kimberlites form a Type 3 field (i.e. kimberlite fields of different age in the same region; Mitchell, 1986), a phenomenon also observed in Yakutia and South Africa (two areas notable for their economic kimberlite pipes).

Heavy mineral concentrate studies of tills (e.g. Ward et al., 1996) in the Lac de Gras area are consistent with the occurrence of macrocrysts found in kimberlites (Kjarsgaard, 1996). Petrological studies (B.A. Kjarsgaard, unpublished data) indicate that the diatremes examined have petrographic assemblages (two generations of olivine and phlogopite, groundmass monticellite, serpentine, calcite, perovskite, spinel, apatite, and kinoshitalite mica) typical of kimberlite. Whole-rock major- and trace-element geochemistry (B.A. Kjarsgaard, unpublished data) on a number of Lac de Gras area kimberlites as well as Sr isotopic studies (W.J. Davis and B.A. Kjarsgaard, unpublished data) are also consistent with a kimberlite signature.

## ECONOMIC POTENTIAL

Economic evaluations (utilizing mini-bulk samples, >10t) have been undertaken on a number of kimberlite pipes in the Lac de Gras area. Of these pipes, a subset have undergone further bulk sampling (>100t). Assessment of diamond parcels from the BHP/Dia Met kimberlites which have been mini-bulk or bulk sampled contained 6 to 33% of gem quality stones, with stone values in the range of 26 to 130 US$/c. A number of these stones are in the 1 to 3 carat size, which is quite promising considering that the bulk samples were not particularly large. Data from the bulk samples have been plotted as grade (c/100t) versus stone value (US$/c) in Figure 3, to give an ore value in US$/t. This ore value has then been plotted versus tonnes of ore to 120m depth

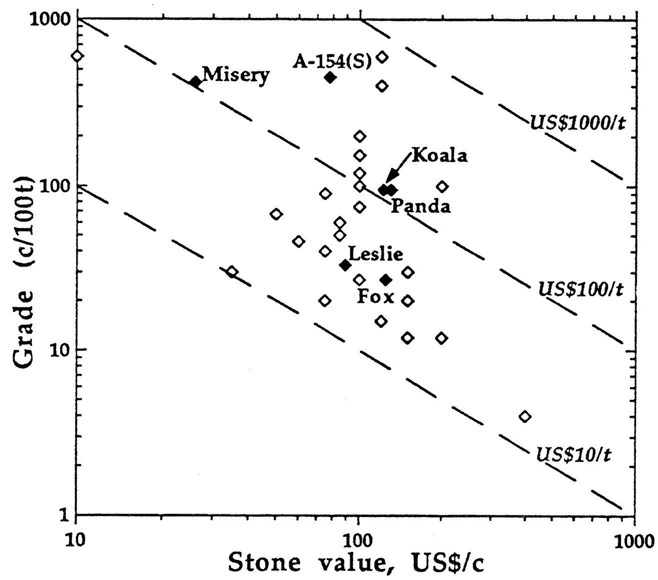


*Figure 3. Kimberlite ore value (US$/t) as determined by grade (c/100t) multiplied by diamond value (US$/c) for a number of economic kimberlite pipes worldwide (open diamond symbols; data from Janse, 1993) as compared to the Northwest Territories pipes (solid diamond symbols; data from various press releases).*

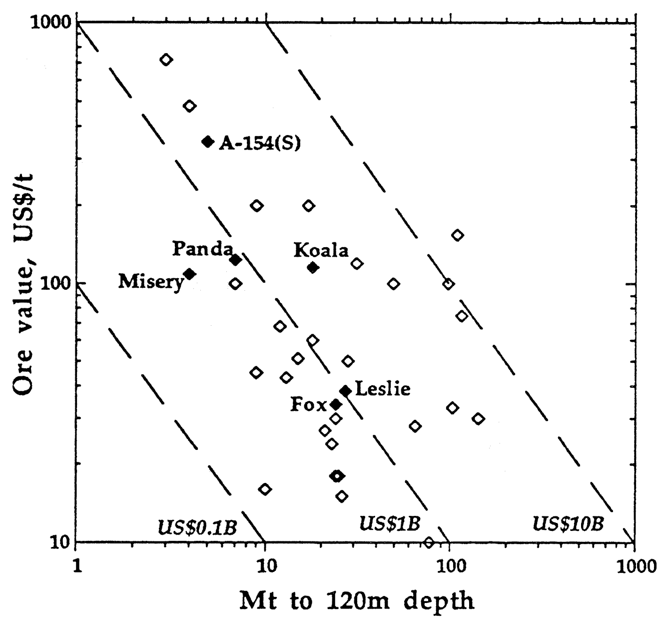


*Figure 4. 'In ground kimberlite pipe value' (in US$B to a depth of 120 m) as determined by deposit size (Mt to a depth of 120 m) multiplied by average value per tonne (US$/t) for a number of economic kimberlite pipes worldwide (open diamond symbols; data from Janse, 1993) as compared to the Northwest Territories pipes (solid diamond symbols; ore value from Figure3; tonnage to 120 m depth estimated by the author).*

(based on the estimated size of the pipe in ha) to produce an 'in ground' value in US$B to 120 m depth (Fig. 4). It is apparent from Figures 3 and 4 that five pipes on the BHP/Dia Met property are comparable to mined kimberlites worldwide. It should also be noted that the very high grade (450c/100t) and good average carat value (54US$/c) of the Aber/Kennecott A154(S) pipe is also similar to mined kimberlites worldwide.

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* Contribution to the Canada-Northwest Territories Mineral Initiatives (1991-1996), an initiative under the Canada-Northwest Territories Economic Development Cooperation Agreement.

* Contribution to the Geological Survey of Canada's Slave Province National Geoscience Mapping (NATMAP) Program.

# Somerset Island kimberlite field, District of Franklin, N.W.T.

**B.A. Kjarsgaard**

*Kjarsgaard, B.A.., 1996: Somerset Island kimberlite field, District of Franklin, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 61-66.*

## INTRODUCTION

Igneous breccias of basic affinity were first noted in the early 1960's during bedrock mapping of Somerset Island (Blackadar and Christie, 1963; Blackadar, 1967). During subsequent exploration work on Somerset Island, H. Neale recognized that these rocks resembled kimberlites. Subsequent petrographic and mineralogical studies on these samples by Mitchell and Fritz (1973) and Clarke and Mitchell (1975) established the Peuyuk 'diatreme' as a bona fide kimberlite. Industry exploration for diamonds during the 1970s resulted in the discovery of additional kimberlites on Somerset Island (Fig. 1). Bulk sampling at this time resulted in the recovery of five small diamonds (net weight 0.297 c) from a 174.7t sample (grades of < 1c/100t). Renewed industry exploration and government studies in the 1990s has resulted in the discovery of additional kimberlites on Somerset Island (Pell 1993, 1995); the field is now interpreted to extend farther east to account for the discovery of a kimberlite on Brodeur Peninsula, Baffin Island.

## AGE AND GEOLOGICAL SETTING

The kimberlites of Somerset Island form a belt trending approximately northeast (Fig. 1). Localization of kimberlite magmatism appears to have been controlled by three distinct fracture sets (trending north; northeast and northwest: Mitchell, 1975), which developed in the Precambrian basement as a result of three phases of Cornwallis folding (Brown et al., 1969). These fracture sets are exposed as major lineaments in the Precambrian Boothia terranes on the western side of Somerset Island (Mitchell, 1975). Radiometric age determinations by Frisch and Hunt (1993) of Precambrian basement rocks from Somerset Island range from 2.48 to 1.71 Ga. The Nd model ages of 3.0 to 2.2 Ga suggest the presence of juvenile Proterozoic rocks as well as Proterozoic crust with incorporated Archean material.

Cretaceous intrusive ages have been determined for the Ham kimberlite (88 Ma: Heaman, 1989) and Georgia pipe (105 Ma: Smith et al., 1989) utilizing U-Pb perovskite dating. Smith et al. (1989) also dated the Elwin Bay kimberlite by this technique, but the interpreted age (27 - 36 Ma) was considered spurious. An errorchron age of 100 Ma (Rb-Sr phlogopite) was determined for the Batty Bay K10 ('Tunraq') kimberlite by Smith (cited in Smith et al., 1989). These radiometric ages are consistent with the kimberlites intruding the Paleozoic cover rocks (Cambro-Ordovician to Upper Silurian sedimentary rocks; Stewart, 1987).

## KIMBERLITE GEOLOGY

Detailed mineralogical studies have been undertaken on several Somerset Island kimberlites, including: *Peuyuk* (Mitchell and Fritz, 1973; Clarke and Mitchell, 1975; Mitchell and Clarke, 1976); *Elwin Bay*, Mitchell (1978a); *Tunraq* (Mitchell, 1979); *Jos* (Mitchell and Meyer, 1980);and *Ham* (Jago and Mitchell, 1985). Re-investigation of the Somerset Island kimberlite field undertaken in July 1990 by B.A. Kjarsgaard and T.D. Peterson included sampling of all the known kimberlite bodies (Fig. 1), as well as detailed mapping of the Batty Bay complex (Fig. 2). From this field work, and subsequent petrographic studies it was recognized that the Somerset Island kimberlites consist principally of hypabyssal facies kimberlite (Kjarsgaard and Peterson 1991). Dykes occur at Jos, Ham, and Batty Bay, and enlarged fissures (blows) are found at Ham, Nord, Ouest, Peuyuk, Elwin Bay, and Batty Bay. Rare diatreme facies (pelletal lapilli-bearing) and transitional hypabyssal/diatreme facies kimberlite occurs at the Batty Bay pipe. This suggests the current exposure level of the Somerset Island pipes is at the kimberlite root zone level (hypabyssal and lowermost diatreme facies rocks; Fig. 3; Kjarsgaard, 1996, this volume). Detailed mapping of the surface expression of the Batty Bay pipe has shown that it is highly irregular, consisting of at least seven petrographically distinct types of kimberlite. These observations are consistent with root zone morphology (Mitchell, 1986)

The mineralogy (Mitchell and co-workers, see previous; B.A. Kjarsgaard, unpublished data) and zoning

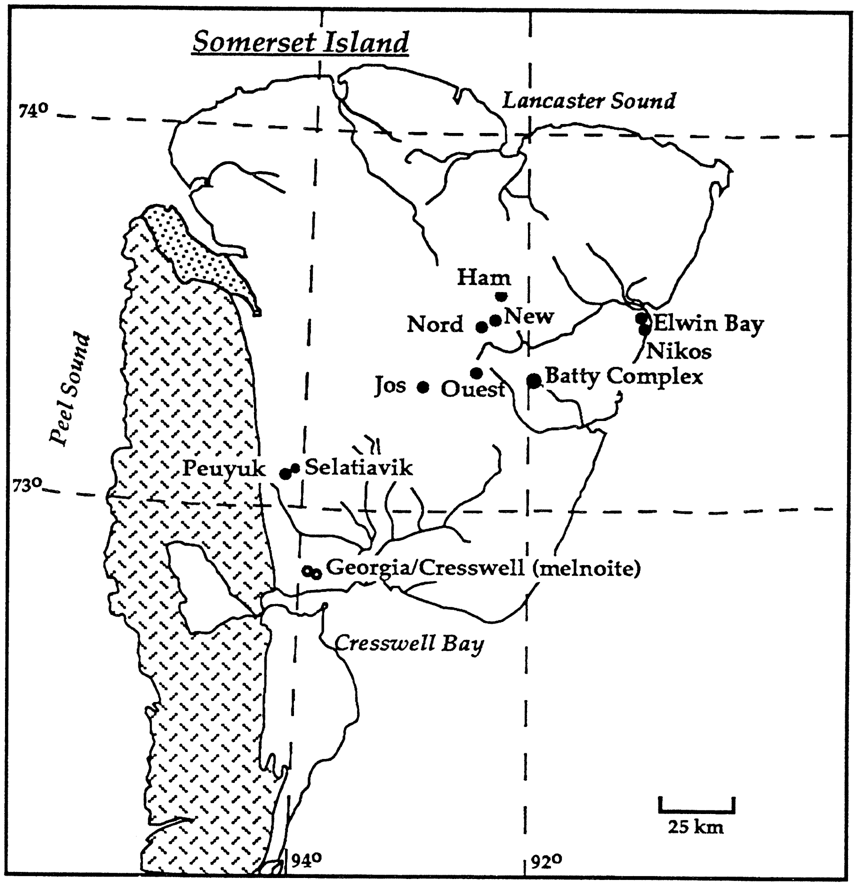


***Figure 1.*** *Location map of the Somerset Island kimberlite field. Regional geology after Stewart (1987): white= Paleozoic cover; stipple= Neoproterozoic cover; cross hatch= Precambrian basement. Kimberlite pipe locations from Kjarsgaard and Peterson (1992); Nikos and New pipe locations from Pell (1995). Note that Georgia and Cresswell pipes are not kimberlites (see text).*

trends of specific minerals are typical of kimberlite. Kimberlite minerals observed include: two generations of olivine and phlogopite plus groundmass serpentine, calcite ± dolomite, spinel, perovskite and apatite, ± Ba-phlogopite - kinoshitalite mica ± monticellite ± pyrite. Whole-rock major and trace element geochemistry (Kjarsgaard, 1993 and unpublished data) of the Somerset Island kimberlites are typical of kimberlite (Smith et al., 1985). A subset of the samples (carbonate-rich) have anomalously high incompatible element signatures, interpreted as a result of fractionation. Combined petrographic, geochemical and concentrate studies (Kjarsgaard, 1992) on samples from Cresswell/Georgia preclude the classification of these samples as kimberlites, and they are best classified as ultramafic lamprophyres (melnoites).

## MANTLE XENOLITHS

Mantle xenoliths have been found at a number of the Somerset pipes. The mantle xenoliths are dominantly spinel, spinel + garnet, and garnet lherzolite with rare garnet dunite and harzburgite (Mitchell, 1987; Kjarsgaard and Peterson 1992; Kjarsgaard, 1992). Xenoliths which show cryptic (e.g. high-Ti garnet lherzolite) and modal (e.g. phlogopite- and rutile-bearing garnet lherzolite) metasomatism are also present in the xenolith population. A single eclogite xenolith has been reported to date (Kjarsgaard, 1992). Using the textural nomenclature scheme of Harte (1977), the Somerset xenoliths are mainly classified as coarse equant to coarse tabular types, although porphyroclastic, mosaic porphyroclastic and disrupted porphyroclastic types are also found.

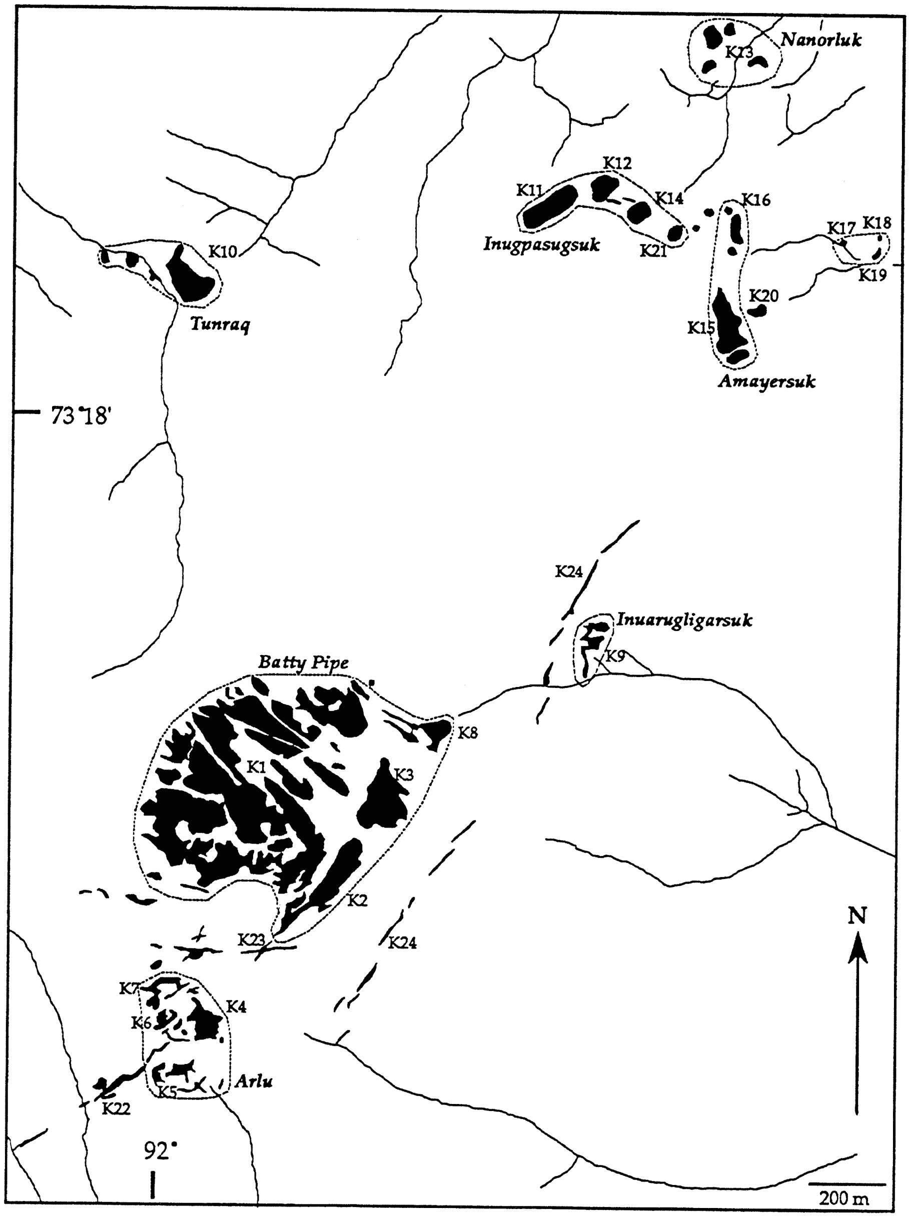


*Figure 2. Detailed map of the surface expression of the Batty Bay kimberlites (solid black areas) showing the surface expression of the complex (based on outcrop, frost-heaved blocks and regolith). Adapted from Kjarsgaard and Peterson (1992). Dashed lines group the kimberlites as named by Mitchell (1976).*

Geothermobarometric studies on the xenoliths have been reported by Mitchell (1977, 1978b, 1987), Jago and Mitchell (1987), and Kjarsgaard and Peterson (1992; Fig. 3). Mitchell (1987) reported maximum P-T conditions of 39 kb at 1146 °C, that is, not within the P-T stability of diamond. In contrast, Kjarsgaard and Peterson (1992) found garnet lherzolite xenoliths derived from depths consistent with diamond stability (53kb at 1234 °C). The xenolith data in P-T space can be interpreted as defining a geotherm, as suggested by Mitchell (1987) and Kjarsgaard and Peterson (1992). Both of these studies are in good agreement, suggesting a steady state paleogeotherm slightly hotter than 44m/Wm$^2$ (Fig. 3).

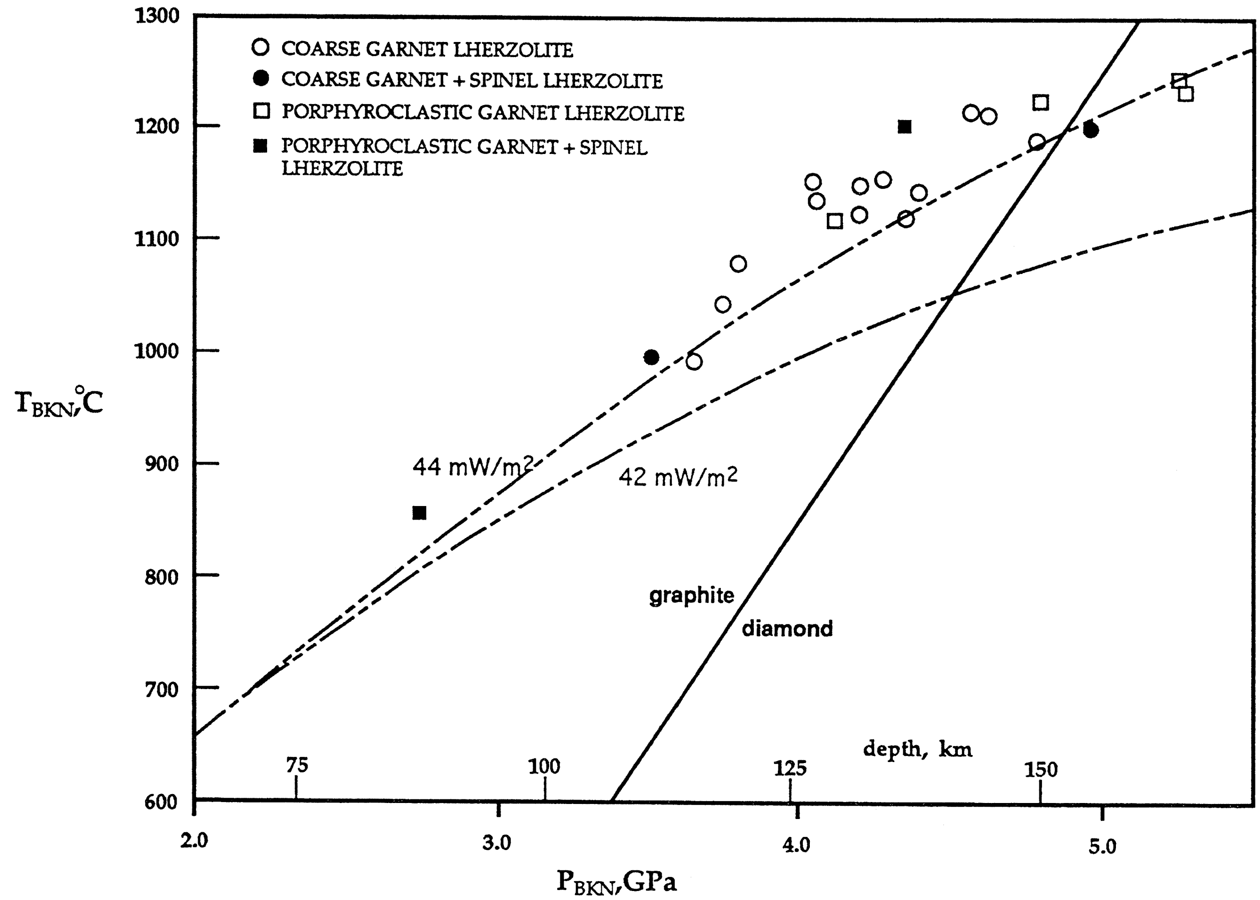


*__Figure 3__. Pressures and temperatures of equilibration of mantle xenoliths from the Batty Bay kimberlites, illustrating a geotherm slightly hotter than 44mW/m². Adapted from Kjarsgaard and Peterson (1992).*

## HEAVY MINERAL CONCENTRATES

Heavy mineral concentrate data have been reported by Fipke (1989), Kjarsgaard (1992, and unpublished data), Shulze (1993), and Gurney (cited in Helmstaedt, 1993). These studies recognized minerals associated with mantle xenoliths and the megacryst suite. Xenolith-derived minerals are dominantly from spinel and garnet-bearing lherzolites. Garnet dunite/harzburgite and eclogite xenoliths are rare, consistent with the heavy mineral concentrate data and the paucity of diamonds in the Somerset Island kimberlites (Kjarsgaard, 1992).

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# Prairie kimberlites


## B.A. Kjarsgaard



*Kjarsgaard, B.A., 1996: Prairie kimberlites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 67-72.*


## CENTRAL SASKATCHEWAN KIMBERLITES

In 1988, the first kimberlite in Saskatchewan was discovered at Sturgeon Lake (SL1 pipe; Fig. 1), 50 km northwest of Prince Albert. A second discovery was made in 1989, 15 km farther to the northwest (SL2 pipe; Fig. 1). No diamonds were recovered from preliminary bulk sampling of the SL1 body (Letendre, 1989a), although the SL2 body is reported to be diamondiferous.

The discovery of the SL1 kimberlite led to increased exploration for kimberlite pipes in central Saskatchewan. Tracts of ground were subsequently staked in the Fort à la Corne area (85 km east of Prince Albert; Fig. 1) in the late summer/fall of 1988 on the basis of existing GSC aeromagnetic surveys (e.g. GSC Map 7743G) and additional industry airborne and ground geophysical surveys. Drilling in the summer of 1989 confirmed the existence of kimberlite in the Fort à la Corne area. Lehnert-Thiel et al. (1992) provided a concise summary of the exploration and drilling activity in that area until 1991. More recent exploration work at Fort à la Corne has delimited a number of pipes in spatially distinct clusters within this kimberlite field. By 1994, there were 44 confirmed (drilled) pipes and more than 30 additional geophysically defined drill targets (Scott-Smith et al., 1994). Reported diamond grades for the Fort à la Corne kimberlite pipes range from 0 to < 23c/100t based on sample weights of < 10t (Northern Miner, 1995).

Continuing exploration in central Saskatchewan has resulted in the discovery of five kimberlites to the northeast of Candle Lake (Fig. 1). At present, it is unknown if this cluster of pipes represents an extension of the Fort à la Corne field or is part of a different kimberlite field.

### *Geology of the subsurface basement*

In central Saskatchewan, preliminary geological interpretations of the subsurface crystalline basement were made by Green et al. (1985) utilizing geophysical data. In the Fort à la Corne - Prince Albert - Candle Lake region, the basement is interpreted to be the southerly extension of the Glennie Domain (Green et al., 1985). The Glennie Domain is one of the lithostructural subdivisions of the internal (Reindeer) zone of the Paleoproterozoic Trans-Hudson Orogen (Lewry and Collerson, 1990) and is characterized by island arc volcanogenic (supracrustal) successions separated by reworked granitoids and granitoid gneisses (Van Schmus et al., 1987; McNichol et al., 1992). However, rare Archean interlayered basement gneisses (dated at >2800 and 2500 Ma) occur in tectonic windows within Proterozoic rocks (Chiarenzelli, 1989). The supracrustal rocks are subdivided into volcanogenic (volcanic, synvolcanic intrusives, volcaniclastics and associated sediments), arkosic, and pelitic-semipelitic assemblages. U-Pb age determinations on various Proterozoic rock types from the Glennie Domain indicate that magmatism occurred during the interval 1893 to 1710 Ma (summarized in McNichol et al., 1992)

Collerson et al. (1988, 1989) provided new isotopic constraints from drill-core samples of subsurface basement rocks in central Saskatchewan. Crystalline rock types observed in the Fort à la Corne area include banded iron formation, phyllite, schist, and ortho- and paragneiss, consistent with known lithological associations observed in the exposed Glennie Domain to the north. At present the only published U-Pb (zircon) age is 1786±4 Ma on an intermediate granulite gneiss from the subsurface to the southeast of Fort à la Corne (SASK-8; Collerson et al., 1988). Nd model age determinations for three central Saskatchewan subsurface Glennie Domain samples are 3.12, 2.98, and 2.35 Ga (Collerson et al., 1989). The 2.98 Ga Nd model age for SASK-8, coupled with its 1786 Ma crystallization age, indicates reworking of Archean material. In summary, aeromagnetic and gravity data, isotopic studies and lithological assemblages suggest that the crystalline basement through which the kimberlites erupted is geologically similar to the Glennie Domain, as exposed on the shield; a collage of dominantly juvenile Paleoproterozoic rocks plus reworked Archean material that is interpreted to overlie an Archean basement terrane (Lucas et al., 1993).

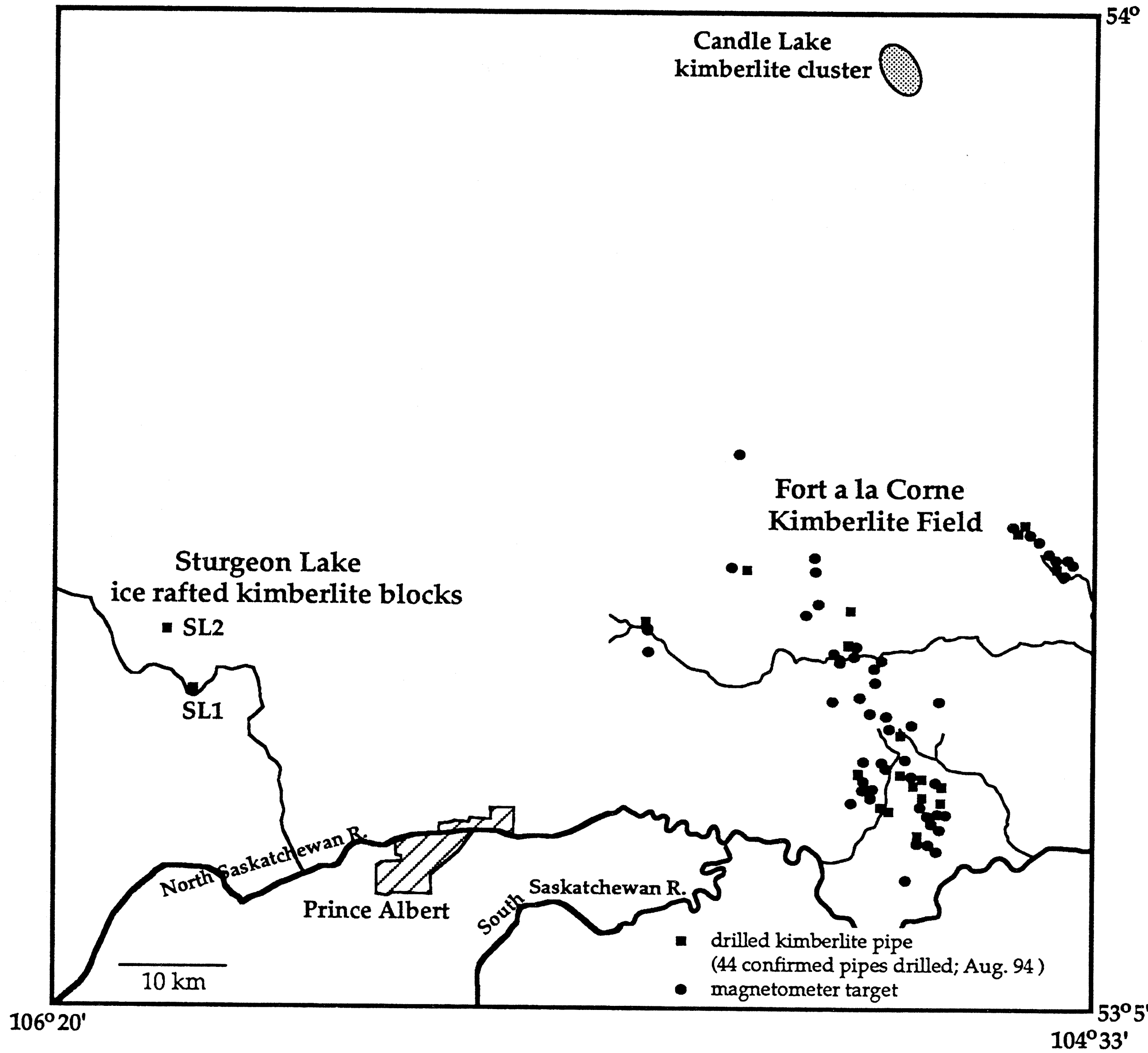


*Figure 1. Central Saskatchewan kimberlites, illustrating the positions of the ice rafted Sturgeon Lake kimberlites (SL1 and SL2), the pipes of the Fort à la Corne kimberlite field, and the Candle Lake kimberlite cluster (stippled oval).*

## *Geology, Sturgeon Lake kimberlites*

Drilling results from the SL2 body (Fig. 1) reported by Letendre (1989b), indicate glacial till occurs both above and below the kimberlite intersection in two separate drill holes. The interpretation of the drill results is that the SL2 body is a glacially transported kimberlite block. Similarly, the SL1 body (Fig. 1) has been interpreted as an ice rafted block (Kjarsgaard, 1995). This ice rafting model is consistent with the SL1 body outcropping in an area of >120 m of glacial overburden. Geophysical studies of the SL1 pipe, which indicate this kimberlite is a thin, rootless, puck shaped body (Urquart and Hopkins, 1993) also support this idea. The SL2 kimberlite has been dated at 98.0 ± 1.0 Ma (5 point Rb-Sr phlogopite/whole-rock isochron; Hegner et al., 1995)

Heavy mineral concentrates from the SL1 kimberlite consist of olivine, phlogopite, Cr-diopside, enstatite, garnet, spinel, ilmenite and zircon (Fipke, 1989); Kjarsgaard, 1994), consistent with this body being a kimberlite. Garnet xenocrysts, with compositions similar to those observed as diamond inclusions in South African and Yakutian kimberlite pipes, have been recovered from concentrate; garnets of both subcalcic harzburgite/dunite paragenesis and high $Na_2O$ (>.08 wt%) eclogite paragenesis are observed in minor proportions. Compositions of spinels from concentrate, however, lie outside the range of diamond inclusion spinels.

### *Geology, Fort à la Corne pipes*

The Fort à la Corne kimberlites consist only of crater facies rocks (hypabyssal feeder dykes have not been observed) and their geological interpretation is problematic. Originally thought to have formed as a result of subaqueous eruptions (Gent, 1992; Nixon et al., 1993), more recent studies (Scott-Smith et al., 1994; Scott-Smith, 1995; Kjarsgaard, 1995; Kjarsgaard et al., 1995) indicate these kimberlites formed by subaerial processes. However, even within this subaerial eruption framework, two different models are proposed. Scott-Smith et al. (1994) interpreted that the volcanism occurred over at least 25 million years (from >119 to 91 Ma), consisting of a precursor phase from >119 to 100 Ma, followed by the main eruptive phase from 97.5 to 91 Ma. However, no palynological or paleontological data are presented by Scott-Smith et al. (1994) in support of their model, and it is difficult to assess the quality of the Rb-Sr phlogopite ages (96 - 94 Ma) because analytical techniques, data, and error estimates are not quoted. The main eruptive phase is suggested by Scott-Smith (1995) to be maar type (phreatomagmatic), with the pancake shaped form of these bodies a consequence of subsequent infilling of the crater by pyroclastics.

Recently, detailed studies of a complete drill core (stored at GSC Calgary, core archival no. 21250) from the Smeaton area (geophysical target no. 169) of Fort à la Corne have been undertaken by the GSC (Kjarsgaard et al., 1995). The core (141 m length @ >99% recovery) penetrated upper Albian marine shales of the Westgate Formation, 87 m of varying types of kimberlite, and middle Albian nonmarine sediments of the Cantuar Formation (Fig. 2). Studies undertaken include sedimentology, volcanology, mineralogy, geochemistry, palynology, micropaleontology, and radiometric dating. This multidisciplinary approach supports a model which contrasts with that proposed by Scott-Smith (1995) in terms of both timing and mode of emplacement. Kjarsgaard (1995) suggests that the major period of kimberlite volcanism at Fort à la Corne occurred at late Albian times (98.5 to ca. 103 Ma), and that all kimberlite volcanism occurred during the Albian (98.5 ± 0.5 Ma to 112 ± 1.0 Ma; using the time scale of Obradovich, 1993). Kimberlites from the main phase of volcanism form conformable, airfall deposits on terrestrial sandstones of the Cantuar Formation, Mannville Group (middle Albian) and occur below marine shale of the upper Westgate Formation. The stratigraphic position of the kimberlites below the Fish Scales Formation (which marks the boundary between the Albian and Cenomanian) suggests the volcanism occurred prior to 98.5 Ma (Fig. 2). The 'pancake shape' of the pipes is interpreted to be a result of

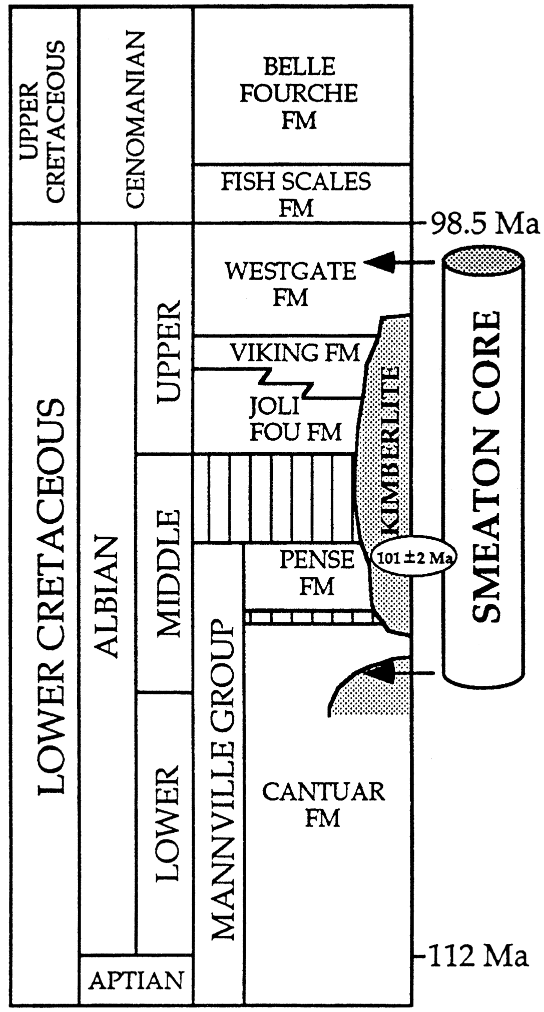


*Figure 2. Stratigraphic section of Cretaceous rocks in the Fort à la Corne area, showing stratigraphic position of the early and main period of kimberlite volcanism from Smeaton drill core no. 21250. Vertical pattern represents no time-rock record. The 101 ± 2 Ma is a U-Pb perovskite age for a subaerial lapilli tuff from the Smeaton drill core (L.M. Heaman and B.A. Kjarsgaard, unpublished data). Adapted from Kjarsgaard (1995).*

the formation of Fort à la Corne kimberlite tephra cones which were subsequently reworked in a marine environment and bevelled off during transgression associated with deposition of the marine Westgate shales (Kjarsgaard, 1995). Combining results from the geological investigations of Scott-Smith et al. (1994), and those of Kjarsgaard (1995) and Kjarsgaard et al. (1995), it is suggested here that there are two 'end-member' kimberlite morphology types at Fort à la Corne;

kimberlite tephra cones and kimberlite vent/tephra cones (Fig. 3; Kjarsgaard 1996, this volume). The age of the precursor kimberlites is not as well constrained as the main eruptive phase, but is interpreted to be younger than 108 Ma on the basis of stratigraphy.

Heavy mineral concentrates from Fort à la Corne kimberlites (Lehnert-Thiel et al., 1992; Schulze; 1993; B.A. Kjarsgaard, unpublished data) contain both P-type and E-type garnets with compositions similar to those from diamond inclusions in South African and Yakutian kimberlite pipes, although garnets with P-type diamond inclusion signatures are not common.

## ULTRAMAFIC PIPES IN ALBERTA

In Alberta, industry hydrocarbon drilling activity has provided core samples of the subsurface basement. Ross et al. (1991) and Villeneuve et al. (1993) utilized U-Pb age and Nd model age determinations coupled with regional geophysical data to interpret the geology of the crystalline basement. Identification of basement domains of Archean and Paleoproterozoic age (Ross et al., 1991) is geologically favourable for the occurrence of diamondiferous kimberlite and lamproite pipes. On this basis, vast tracts of land in Alberta have been staked for diamond exploration (Fig. 3). Dufresne et al. (1994) recently provided an in depth analysis of the diamond potential of Alberta.

Currently, ultramafic pipes have been identified in the Grande Prairie area of northwest Alberta, where they intrude sedimentary rocks of the Western Interior Platform (Stott and Aitken, 1993). The basement geology in this region consists of Paleoproterozoic gneissic rocks of the Chinchinga Domain (a prominent gravity low), formed at 2.17 to 2.08 Ga (U-Pb zircon; Villeneuve et al., 1993). Nd model ages range from 2.46 to 2.68 Ga (Thériault and Ross, 1991; Villeneuve et al., 1993), indicating a mixed juvenile plus Archean origin for Chinchaga rocks. Preliminary information available about the pipes in the Grande Prairie area suggests they are dominated by crater facies material, similar to those at Fort à la Corne, Saskatchewan (B.H. Scott-Smith, oral presentation; CIM Meeting, Vancouver, 1994). Petrographic studies (Kjarsgaard, in press) classify these rocks as olivine-rich juvenile lapilli tuffs. Preliminary results from whole-rock geochemistry and mineralogy studies suggest the Grande Prairie pipes are ultramafic lamprophyres, or they are crustally contaminated kimberlites (Kjarsgaard, in press). Additional work (in progress) is required to resolve this problem.

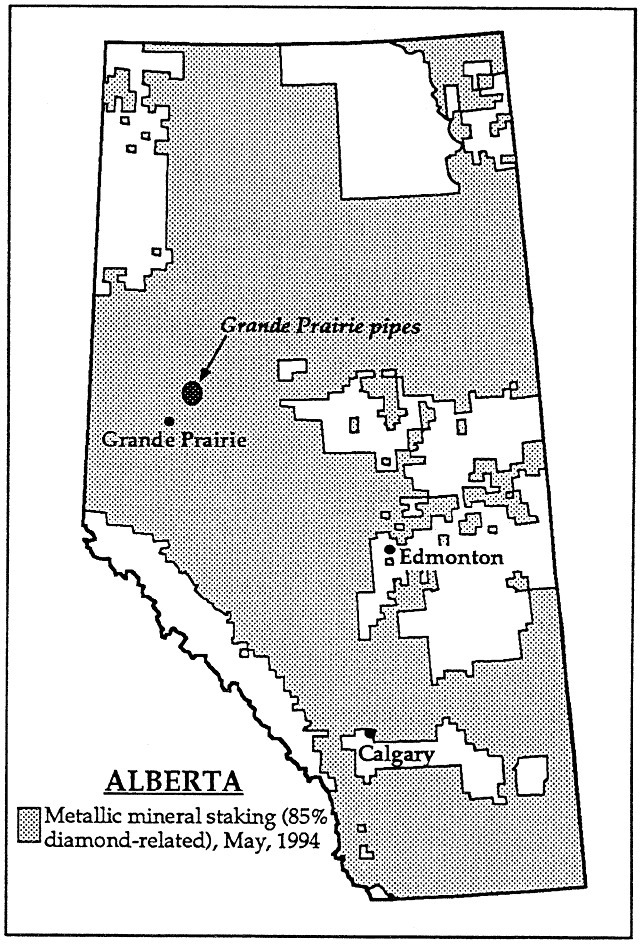


*Figure 3. Extent of staking in Alberta, also showing the region in which ultramafic lamprophyre/ kimberlite pipes have been discovered in northwest Alberta. Note that metallic mineral staking includes diamonds, and that approximately 85% of the land shown as staked in this figure is related to diamond exploration (modified from Dufresne et al., 1994).*

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# Kimberlites in the vicinity of Kirkland Lake and Lake Timiskaming, Ontario and Québec

D.J. Schulze

*Schulze, D.J., 1996: Kimberlites in the vicinity of Kirkland Lake and Lake Timiskaming, Ontario and Québec; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 73-78.*

## INTRODUCTION

Kimberlites have been known to occur near Kirkland Lake, Ontario for almost 50 years (Satterly, 1948). Exploration for diamond-bearing pipes in this region, which continues to the present time, has resulted in the recognition of more than 29 kimberlite dykes and diatremes, many of which contain diamonds. The history of these discoveries and the geology of many of these occurrences has been documented by Brummer et al. (1992 a,b).

The kimberlites seem to occur in two separate clusters, one north and east of Kirkland Lake and the other, to the south, near Lake Timiskaming (Fig. 1). Kimberlites in the Kirkland Lake (KL) cluster intrude Archean rocks of the Abitibi belt, and those in the Lake Timiskaming (LT) cluster intrude Huronian Supergroup metasedimentary rocks and diabase sills, which overlie Archean basement (Bennett et al., 1993). Most kimberlites are deeply buried under glacial sediments (Brummer et al., 1992b; McClenaghan, 1996).

Jurassic emplacement ages (155-159 Ma) were determined by the U-Pb method on matrix perovskites by Heaman (1989; and quotation in Brummer et al., 1992b), in agreement with an earlier K-Ar age of 151 Ma (Lee and Lawrence, 1968). Xenoliths of Paleozoic limestone are common in kimberlites in the Kirkland Lake cluster, and were probably the cover rocks at the time of kimberlite eruption.

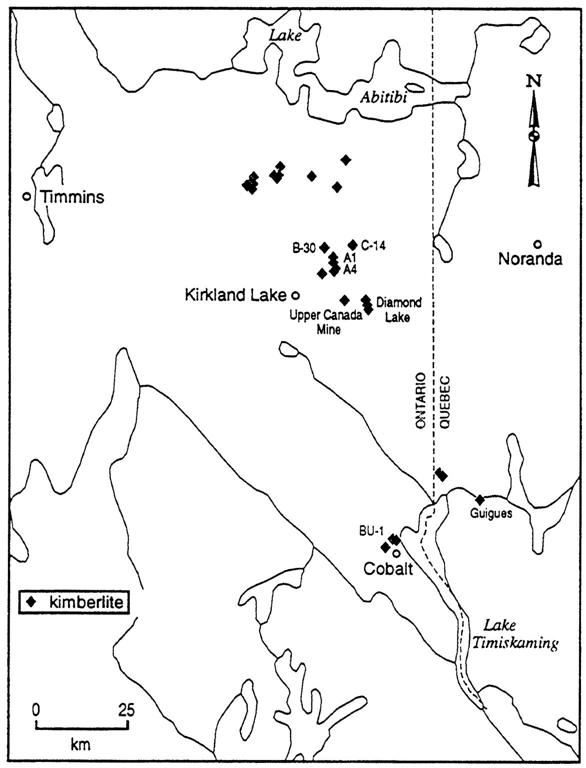


*Figure 1. Locations of known kimberlite bodies in the Kirkland Lake-Lake Timiskaming region. Locations from Brummer et al. (1992b), McClenaghan (1993) and Zalnieriunas and Sage (1995).*

## KIMBERLITE STRUCTURE AND PETROLOGY

The six bodies in the Lake Timiskaming cluster are thought to be kimberlite diatremes, or pipes, as are most of those in the Kirkland Lake cluster. Several thin dykes also occur in the KL cluster, such as the Upper Canada Mine kimberlite described by Lee and Lawrence (1968), and most of the kimberlites to the northwest of the main KL cluster are dykes encountered in gold mining and exploration activities. The Diamond Lake pipe is cut by a dyke that extends into the surrounding Timiskaming Group greywacke. Pipes described by Brummer et al. (1992b) are typically 100 to 300 m in diameter at their present upper levels, and the largest, the B-30 (or Nickila Lake) pipe, measures 220 by 350 m and has an estimated surface area (beneath cover) of 5.8 ha (Brummer et al., 1992b).

The dominant rock type in the diatremes is heterolithic tuffisitic (or volcaniclastic) kimberlite breccia, an intrusive breccia with abundant (>15%) xenoliths larger than 4mm. Kimberlite pellets are common. Xenoliths are primarily of Paleozoic carbonate rocks and Archean metavolcanic rocks in the Kirkland Lake kimberlites, and diabase fragments are common in kimberlites from the Lake Timiskaming cluster. Garnet granulites, inferred to be lower crustal fragments, are present, although uncommon (Davis and Moser, 1996), as are xenoliths of mantle-derived peridotite and eclogite.

Hypabyssal kimberlite (non-fragmental) occurs as xenoliths (autoliths) in many of the breccias, and as dykes or irregularly shaped intrusive bodies in the deeper parts (>300 m) of some drill holes (e.g. C-14 and A- 1). This suggests that these depths correspond to the transition from diatreme facies to the root zone of the kimberlites (Mitchell, 1986) and, therefore, that the bulk of the pipes represent the mid to lower portions of diatremes. There is no evidence for preservation of the crater facies of these pipes, in contrast to those in the Slave Province (Kjarsgaard, 1996).

Little petrographic information on Kirkland Lake kimberlites has been published. Arima et al. (1986) examined a hypabyssal fragment in tuffisitic breccia from the B-30 (Nickila Lake) kimberlite, and concluded that it was related to micaceous kimberlite (i.e., the Group II kimberlites of Smith, 1983). Many other characteristics of these kimberlites are inconsistent with a Group II classification, however, including the abundance of ilmenite in all pipes examined by me, (except C-14, where ilmenite is present, but uncommon), the presence of high-temperature deformed garnet peridotite nodules (Meyer et al., 1993; Vicker and Schulze, 1994), Cr-poor garnet and clinopyroxene megacrysts and matrix monticellite (D.J. Schulze, unpublished data), the composition of matrix spinels from the Upper Canada Mine dyke (Mitchell, 1978, 1986), and the abundance of matrix perovskite.

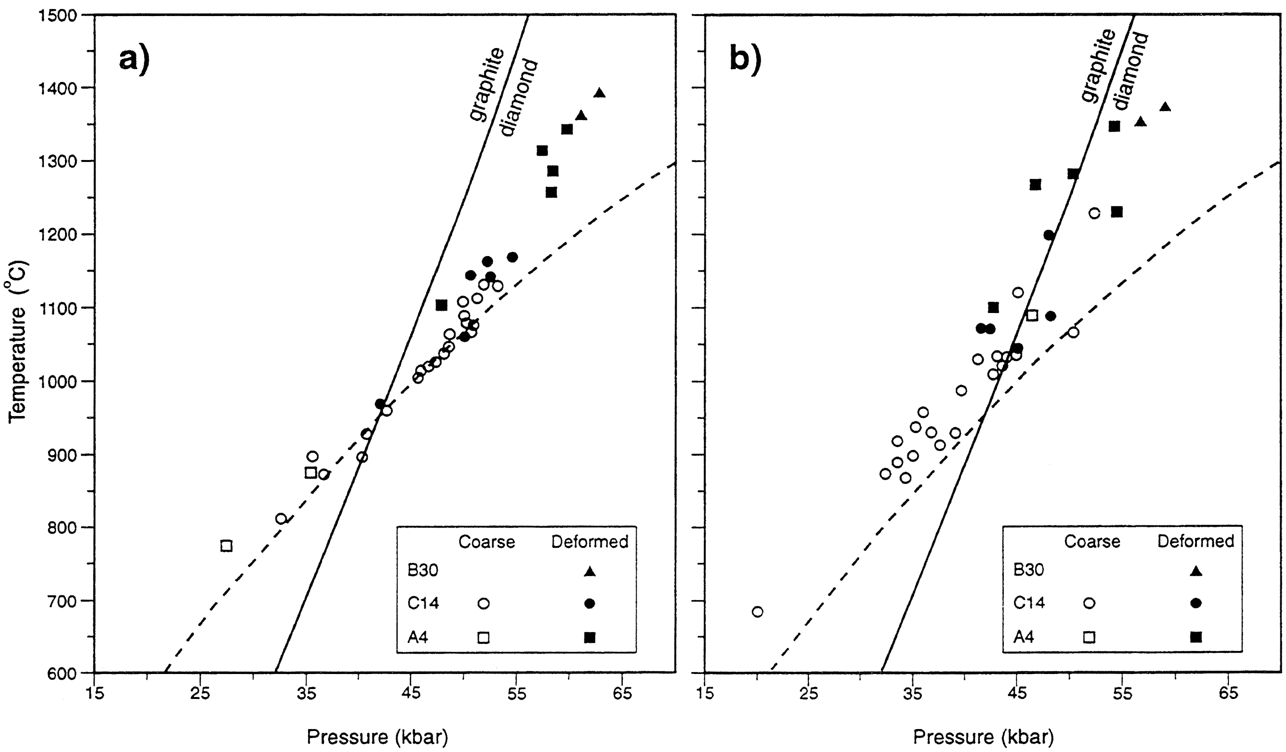


*__Figure 2.__ Equilibration conditions of garnet peridotite xenoliths from Kirkland Lake kimberlites A-4, B-30, and C-14 (P.A. Vicker, unpublished data). Solid line represents graphite-diamond transition of Kennedy and Kennedy (1976), and dashed curve represents a subcontinental steady state geothermal gradient corresponding to a surface heat flow of 40mW/m² (Pollack and Chapman, 1977); a) Temperatures calculated with the Fe-Mg exchange between garnet and olivine calibrated by O'Neill and Wood (1979) and pressures calculated using data from MacGregor (1974); b) Temperatures and pressures calculated by the methods of Brey et al. (1990).*

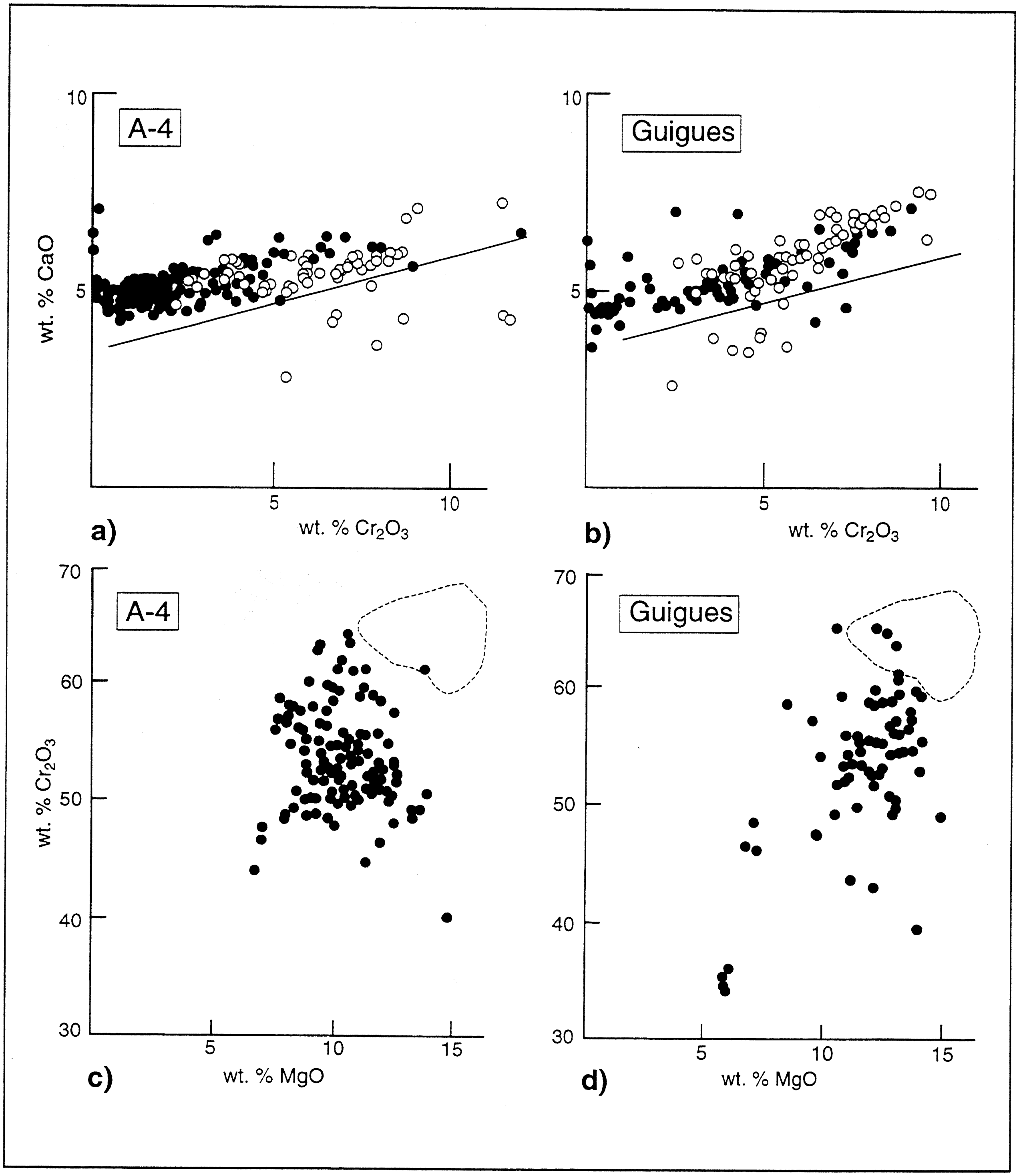


*Figure 3.* *Compositions of garnet (a,b) and chromite (c,d) xenocrysts from kimberlite pipes A-4 and Guigues (D.J. Schulze and P.F.N. Anderson, unpublished data). In garnet diagrams, solid circles are a representative population of all garnets in small splits of concentrate, and open circles indicate garnets selected on the basis of their purple colour, in a search for more G10 garnets. The solid line represents the lower CaO limit of garnets from Kirkland Lake lherzolites (P.A. Vicker, unpublished data), and is similar to the G9-G10 division of Gurney (1984). In the chromite diagrams, the dashed line outlines the field of chromite inclusions in diamonds (Fipke, 1989).*

## MANTLE INCLUSIONS

Mantle-derived material in these kimberlites includes diamonds, garnet and spinel peridotites, eclogites, Cr-poor megacrysts, and a variety of xenocrysts. Although many of the pipes have been reported to contain diamonds, the only available data on diamond grade have been published by Brummer et al. (1992b). They reported diamond contents for A-4, B-30, C-14 (KL cluster), and BU-1 (LT cluster) to be uniformly low, with a maximum value at C-14 of 2 carats per 100 tonnes.

Garnet peridotite xenoliths are dominantly lherzolites. Garnet harzburgites are rare, and a single low-Ca garnet harzburgite has been identified. Eclogites are common only in A-4, and have not yet been studied in detail. Garnet peridotites have been recovered and analyzed from C-14, A-1, A-4 and B-30, all in the KL cluster (Meyer et al., 1993, 1994; Vicker and Schulze, 1994; P.A. Vicker, unpublished data). Coarse, undeformed varieties are most common and have apparently equilibrated at conditions corresponding to those on a steady-state geothermal gradient, extending into the diamond stability field (Fig. 2).

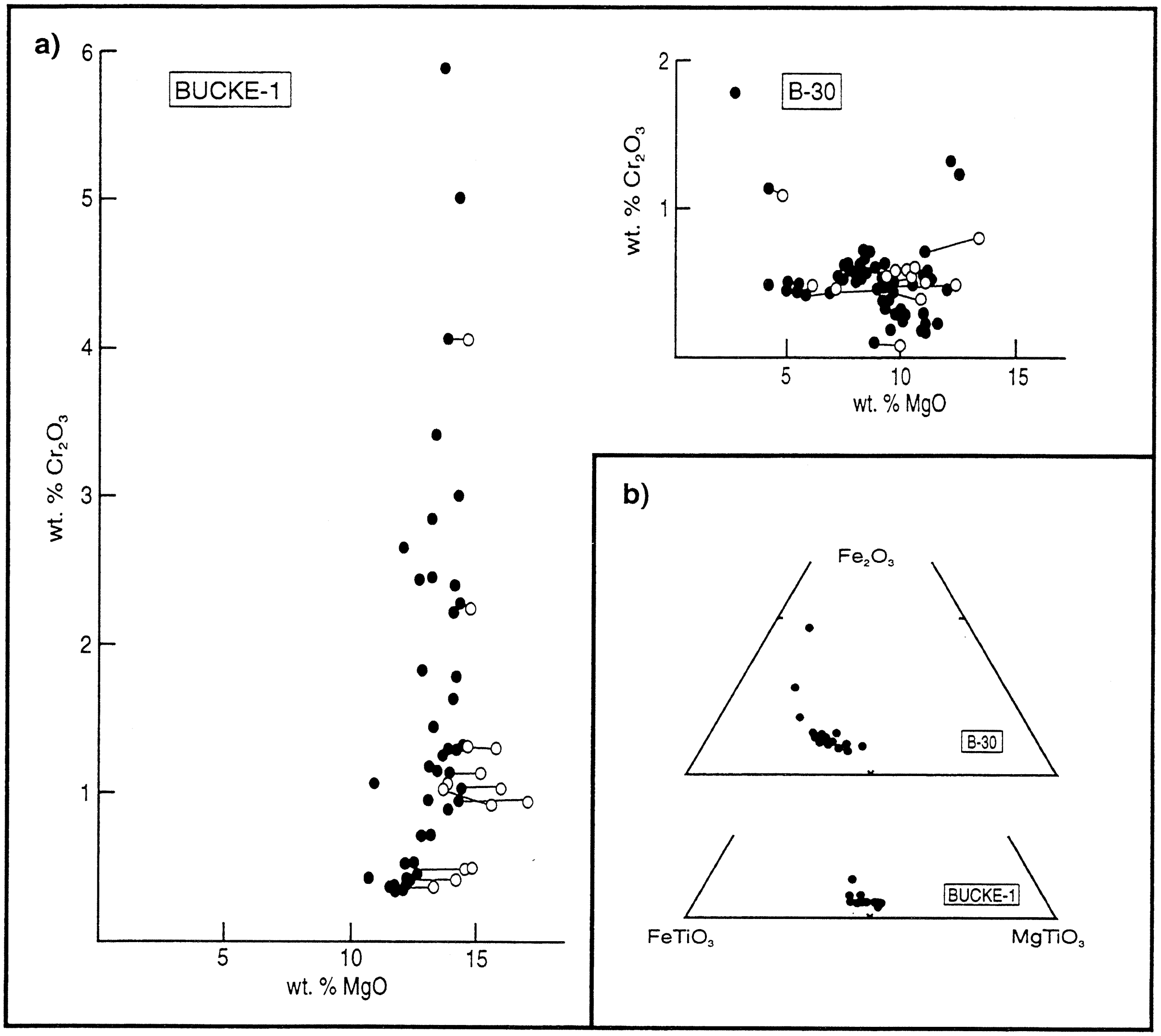


*Figure 4. Ilmenite populations from Bucke-1 and B-30 kimberlites (Schulze et al., in press); a) MgO-$Cr_2O_3$ variations in ilmenites, with core compositions represented by solid circles and rim compositions by open circles; b) Ilmenite end-members $FeTiO_3$-$MgTiO_3$-$Fe_2O_3$ (mole%) calculated on the basis of ilmenite stoichiometry.*

Some high-temperature peridotites have equilibration temperatures above such a geothermal gradient and contain Cr-pyrope garnets with rims enriched in $TiO_2$ and FeO, suggestive of high-temperature metasomatism (Vicker and Schulze, 1994), a recognized feature of high-temperature garnet peridotites, worldwide (e.g., Smith and Boyd, 1987). Metasomatic replacement of garnet by diopside + phlogopite is common in peridotites from C-14.

Composition of Cr-pyrope and chromite xenocrysts are consistent with the presence of diamonds, but also with the very low reported diamond grade of these pipes (Schulze and Anderson, 1992, 1994, unpublished data; Meyer et al., 1994). Analyses of garnet and chromite populations from A-4 (KL cluster) and Guigues (TL cluster) are shown in Figure 3. Although, from a diamond exploration viewpoint, these are the most encouraging xenocryst populations documented to date from the KL-LT kimberlite province, the data do not suggest significant diamond grade in these pipes, as only a few low-Ca pyropes (G10's) and high-Cr chromites corresponding to diamond inclusion minerals have been identified.

Each pipe in this region seems to have a compositionally distinct ilmenite population. Ilmenite populations from BU-1 and B-30 are shown in Figure 4. BU-1 ilmenites are enriched in MgO and $Cr_2O_3$ and are reduced ($Fe_2O_3$ component to a maximum of 8%). In contrast, B-30 ilmenites are typically lower in MgO (to as low as 2.6 wt% MgO), with a negative correlation between MgO and $Cr_2O_3$ for MgO-poor grains, and are much more oxidized (to 48 mole% $Fe_2O_3$). Both pipes have few diamonds, however, and ilmenite populations seem to have little relationship to diamond contents of the kimberlites.

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# Lamproites

**T.D. Peterson**

*Peterson, T.D., 1996: Lamproites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 79-86.*

## INTRODUCTION

Other than kimberlites, lamproites are the only primary sources of diamond which form economically exploitable bodies. But since lamproite intrusions generally have low diamond grade and/or quality, their potential value is less. Nevertheless, colourless and coloured stones from Australia- mostly red, champagne, and yellow- have found a niche in the gem market (Jaques et al., 1986). The lamproite-hosted Argyle mine in northern Australia, the world's largest producer of diamonds, yielded over 35 million carats in 1991 (over 90% industrial stones). Three other lamproite mines, in India, west Africa, and the United States, have supported local economies with modest diamond production, which is striking considering that lamproites are rare; there are at present less than 40 known fields (Fig. 1). In addition, some petrologists consider Group II kimberlites (or *orangeites*), some of which are major gem producers, to be a variety of lamproite peculiar to the South African craton (e.g., Mitchell, 1994; Tainton, 1994). Lamproites do occur in Canada, and have been peripheral targets during the recent phase of exploration activity. A rich source of microdiamonds has been discovered west of Hudson Bay in a dyke that resembles lamproites in several respects (MacRae et al., 1996).

Like kimberlites, lamproites are restricted entirely to continents, but there are few other similarities. Lamproites are more complex rocks than kimberlites, and they exhibit a very wide range in mineral assemblage, texture, and composition. They occur as dykes, volcanic vents, lava flows and lava lakes, small plutons, and a variety of pyroclastic rocks. Individual centres commonly have one or more geochemical peculiarities not shared by other lamproite bodies. Their isotopic compositions indicate that their source regions had a prolonged history within subcontinental lithospheric mantle. In addition to being a potential source of diamonds, lamproites yield unique and vital information on the long-term history of continental lithosphere.

## DEFINITIONS, TYPES, AND CLASSIFICATION

Attempts at categorizing lamproites and related rocks have always produced controversial results; witness the latest report of the IUGS subcommission on alkaline rocks (Woolley et al., 1996). There is broad agreement that lamproites are distinguished from a host of similar rocks (minettes, leucitites, kamafugites, and others) on the basis of high weight percent (wt. %) $K_2O/Na_2O$ (typically >3), high atomic K/Al (usually >0.8, often >1), low CaO (typically <6 wt. %), and high concentrations of incompatible elements. Some suites have very high $TiO_2$ up to and exceeding 6 wt. % (e.g., Ellendale, Australia and Smoky Butte, Montana). High contents of Ti, Zr, and Nb produce a wide variety of accessory minerals such as perovskite, priderite, shcherbakovite, and wadeite, which are considered diagnostic of lamproites (Mitchell and Bergman, 1991).

Phlogopite is nearly ubiquitous as a phenocryst phase in lamproites; other common phenocrysts are chromite, olivine, clinopyroxene, apatite, and leucite. High-Ti potassian richterite is a common and diagnostic groundmass phase, and may form large poikilitic grains, or euhedral crystals that project into residual patches rich in carbonate minerals plus priderite and others. Sanidine, mostly restricted to the groundmass, is Ba- and Fe-rich, and Na-poor (usually $>Or_{90}$). Compositional trends in all minerals indicate increasing peralkalinity with Fe and Ti enrichment, which is most clearly seen in late Fe-Ti micas that are strongly depleted in Al (tetraferriphlogopite). Diopsidic clinopyroxene is Al-poor (<1 wt. %), often with green Na-rich rims. These are features (including the diagnostic accessory mineral suite) which are often shared by orangeites (see Mitchell [1985] and Mitchell and Bergman [1991] for a summary of lamproite mineralogy). Common minerals in alkaline rocks which exclude a given suite from the lamproite clan are nepheline, melanite, and melilite.

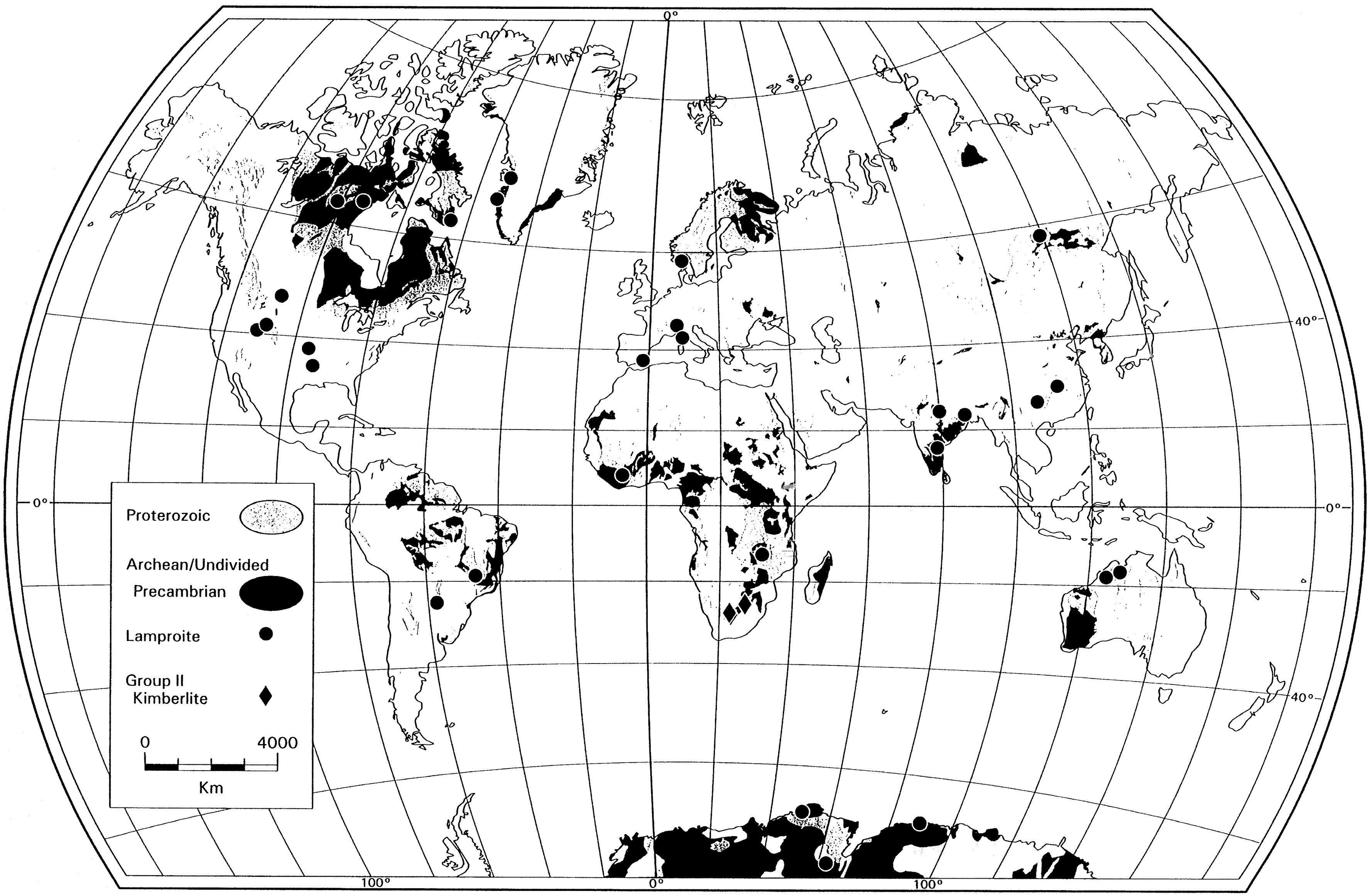


***Figure 1.*** *Locations of prominent lamproite centres and fields. The distribution of Archean and Proterozoic rocks is derived from a digital geological map of the world compiled by Kirkham et al. (1994, 1995).*

It is important to note that the term lamproite refers to a group of rocks, some of which may be related by differentiation and have widely varying $SiO_2$ contents. For example, olivine lamproites, which are the only important diamond source, have silica contents of about 35 to 45 wt. %; leucite lamproites have up to 60 wt. % $SiO_2$ and both can be found in contact in the same vent (e.g., Ellendale 5: Jaques et al., 1986). Glass in rapidly chilled lamproites may contain >60 wt. % $SiO_2$; quartz is a groundmass phase and may even occur in lamproite melts crystallizing at high mantle pressures (Mitchell, 1992; Peterson and LeCheminant, 1993). It is uncertain that primary lamproites (in the sense of, derived from the mantle with little evolution during ascent) need be in equilibrium with olivine or even be Mg-rich (Mitchell and Bergman, 1991). Most authors consider that lamproites originate in veins in mantle rocks composed of a host of volatile-rich (phlogopite, amphibole, carbonate) and incompatible-element enriched phases (such as apatite and rutile) (e.g., Foley, 1991; Waters, 1987). Different degrees of partial melting, total pressure, and local modal variations in source rock mineralogy, could conceivably produce a wide spectrum of primary lamproite melts.

Attempts have been made to manage the difficulties of distinguishing lamproites by developing multielement discriminant diagrams, in which definitive compositional characteristics are recognized statistically and combined algebraically into two variables (Foley et al., 1987; Rock, 1991). While these are of broad interest, they can misclassify individual suites and should be used cautiously. Identification of lamproites should be made on the basis of all available mineralogical and geochemical criteria, with the realization that some suites may resist classification. For example, the lamproites of Murcia-Almeria, Spain (Venturelli et al., 1984) lack most of the diagnostic accessory mineral suite, have phenocrystic orthopyroxene (unique among lamproites), and have relatively low $K_2O/Na_2O$ and high $Al_2O_3$. Nevertheless, they are considered lamproites because groundmass micas have the required compositional zoning, and other characteristic features are present (e.g., high Ba and Rb; presence of K-Ti richterite).

A host of rock names based on local geography has been rejected in favour of a mineralogical classification scheme (Woolley et al., 1996). When naming lamproites, one should refer to the 2 or 3 dominant phases present; e.g., an olivine phlogopite hyalolamproite is dominated by olivine, phlogopite, and glass. A textural distinction is made by Mitchell and Bergman (1991) for rocks where the phlogopite mainly occurs as poikilitic groundmass plates (so-called 'madupitic' lamproites).

## MODE OF INTRUSION

Lamproites occur in virtually all volcanic and extrusive forms. In this, they differ from kimberlites, which occur almost exclusively as dykes, sills, and diatreme and crater facies rocks. There are significant differences between kimberlite and lamproite volcanism (Smith and Lorenz, 1989). Kimberlites, which are essentially carbonated alkali peridotites, exsolve $CO_2$ rapidly during ascent, and can develop diatremes, characterized by complex root zones consisting of intersecting dykes, in nearly any type of country rock. Lamproite magmas, though volatile-rich, are peralkaline and maintain high solubility of volatiles at all pressures. Therefore, explosive lamproite vents are generally restricted to phreatomagmatic centres. Explosive emplacement occurs from the point where magma contacts the water table, such as at the Ellendale field (Stachel, 1992). Phreatomagmatic lamproite vents typically have flaring tops (so-called champagne-glass shaped), filled with inward-dipping beds of olivine-rich lapilli tuffs. These may be intruded by a central plug of differentiated lamproite relatively rich in leucite.

The emplacement site of lamproite intrusions is strongly controlled by local crustal discontinuities. These may be large-scale faults related to basin formation, as at Argyle, or they may be second-order structures related to such faults, as is the case for the diamondiferous Lissadell Road dykes near Argyle (Deakin and White, 1991). A dramatic Canadian example of fault-controlled ultrapotassic volcanism is given by the dykes and basins of the Christopher Island Formation, Northwest Territories (Peterson and LeCheminant, 1996).

One consequence of the relatively slow ascent rate of lamproites through the crust compared to kimberlites is that they contain few mantle xenoliths and xenocrysts. Diamonds in lamproites are more often resorbed and graphitized than in kimberlites, probably due to prolonged interaction with the magma. This makes exploration for lamproites more difficult than it is for kimberlites, since indicator mineral suites are both less voluminous and less diagnostic.

## DIAMONDS, XENOCRYSTS, AND XENOLITHS

Although there is no way to distinguish individual diamonds found in lamproites from those found in kimberlites, it is recognized that diamond populations of lamproites are distinctive. In general, lamproite-hosted diamonds are: more resorbed and graphitized; more often coloured (yellow and brown are common); smaller; and more likely to contain an eclogitic mineral inclusion suite. Hall and Smith (1985) contended that lamproite diamond size distribution is log-normal. Peridotitic (P-type) mineral inclusions in diamonds from Argyle and Ellendale (Jaques et al., 1989) have compositions that overlap those in kimberlite-derived diamonds, but in general they are richer in Fe, and chrome pyropes are not subcalcic. Eclogitic (E-type) clinopyroxene inclusions are anomalously rich in K, consistent with models suggesting that the formation of many E-type diamonds carried by lamproites was related to potassic metasomatism of the lamproite source region.

Sm-Nd model ages of E-type Argyle diamond inclusions, which average 1580 Ma (Richardson, 1986) postdate orogenesis (ca. 1.8 Ga) but predate eruption of the lamproite (ca. 1180 Ma). These data may indicate that the E-type diamonds form during postorogenic cooling of metasomatized mantle. The carbon source for E-type diamonds is interpreted by some to be subducted sediments (e.g., Eldridge et al., 1991). Many petrologists conclude that lamproite source regions form as a result of subduction-related metasomatism (e.g., Nelson et al., 1986), a viewpoint strongly supported by the enriched Sr and (especially) Nd isotopic signatures of lamproites. The relationship between some lamproites and the diamonds they transport to the surface is therefore almost certainly more direct than between kimberlites and diamonds, which helps to explain the exceptionally high grade of Argyle and perhaps some other ultrapotassic rocks (e.g., the Akluilâk dyke: MacRae et al., 1996).

Mantle xenoliths in olivine lamproites are dominantly dunites, with rare harzburgites and lherzolites (Mitchell and Bergman, 1991). These are commonly altered to talc, serpentine, and other hydrous phases (e.g., Peterson and LeCheminant, 1996). A distinctive characteristic of several diamondiferous olivine lamproites is the presence of so-called dog-tooth olivines (Fig. 2), which consist of rounded mantle xenocrysts with faceted overgrowths of magmatic olivine.

## THE DISTRIBUTION OF DIAMONDIFEROUS LAMPROITES

Unlike kimberlites, lamproites are not concentrated within Archean cratons (Fig. 1). Most lamproites are emplaced within mobile belts where there is evidence for previous plate subduction, but the timing of emplacement relative to formation of the belt and later reactivation is variable. Young lamproites of the Mediterranean region,

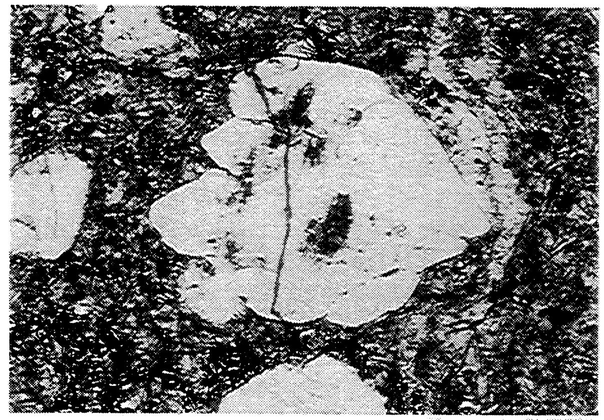

***Figure 2.*** *Dog-tooth olivine in olivine lamproite from Ellendale Pipe #11 (GSC specimen PHA-93-E11-1). Field of view=2 mm. Clear, pyramidal olivine clearly has overgrown resorbed and partly altered olivine.*

where subduction-related volcanism still continues, postdate terminal collision by only 10 to 50 million years. The Argyle and Ellendale lamproites of northern Australia, dated at 1180 and 20 Ma respectively, were emplaced during reactivation of a 1.8 Ga mobile belt. Only in the Wyoming and Churchill provinces of North America do lamproites and related rocks clearly intrude Archean lithosphere, but this lithosphere was extensively reworked during the Paleoproterozoic.

There are no known Archean lamproites; most are Proterozoic and others range in age from Paleozoic to Recent. One lamproite volcano has been active during the last 50000 years (Gaussberg, Antarctica: Sheraton and Cundari, 1980). Most diamondiferous lamproites have Proterozoic emplacement ages. The following summary of occurrences is by no means exhaustive, but lists the most important lamproite intrusions in the context of diamond exploration.

### *Africa*

As was noted above, Group II kimberlites (orangeites) of South Africa much more closely resemble lamproites than they do Group I kimberlites, or any other rock type. Their ages are restricted to between 200 and 100 Ma (Skinner, 1989). Important diamond producing centres include Finsch (110 Ma), New Elands (127 Ma), and Star (age unknown).

The Bobi lamproite (Ivory Coast) is a 2 km long dyke (ca. 1.4 Ga) in west Africa (Mitchell and Bergman, 1991). Diamonds are present in economic proportions, and alluvial stones occur downstream of the intrusion although, because the region is rich in alluvial diamonds, it is uncertain if the lamproite is a significant source of those diamonds.

The Kapamba lamproites, Zambia (Scott-Smith et al., 1989; ca. 220 Ma) are a series of dykes and pipe-like intrusions. Diamonds at subeconomic concentrations are found mainly in olivine lamproite tuffs, but have also been recovered from leucite lamproite flows. The Kapamba rocks have exceptionally high Na and Al for diamond-bearing lamproites.

### *Asia*

Diamond-bearing lamproites occur in both India and China. The Mahjgawan lamproite (1140 Ma) of northern India, was intensively exploited for diamonds from 1937 to 1964 (Mitchell and Bergman, 1991). The grade was about 10 ct/100t. The mine was economic largely due to low labour costs.

The Zhenyuan field of the Yangtze craton, China, contains some early Phanerozoic lamproitic rocks with diamond grades up to 25 ct/100t (the Maping No. 1 sill: Andi et al., 1994). However, the small size of the intrusions renders these occurrences uneconomic.

### *Australia*

The world's largest source of natural diamonds is the Argyle AK1 lamproite pipe which, except for its size, is typical in all respects of diamondiferous lamproites. Altered olivine lamproite tuff is well mixed with up to 50% quartz sand derived from the quartzite country rocks. A central phase has a higher proportion of juvenile volcanic material. The elongated (2 km x 0.5 km) pipe intruded along a fault which was the site of an aquifer in the quartzite, promoting a violent phreatomagmatic eruption. The high-grade portion of the pipe under production late in 1993 had a grade of 2000 ct/100 t (S. Deakin, pers. comm. 1993). A high proportion of the diamonds are resorbed and graphitized; the largest stone recovered weighed 41.2 carats.

Nearby, the Miocene Ellendale field contains a small number of marginally economic vents that are not in production. This field and the Argyle mine are notable for producing coloured stones, mostly red, pink, yellow, and green. Both occurrences are within ca. 1.8 Ga Proterozoic belts on the southern edge of the Kimberley craton.

### *South and North America*

Lamproitic and related ultrapotassic flows and dykes occur in the Alto Paranaíba igneous province of south-central Brazil (Gibson et al., 1995), and also in Paraguay (Presser, 1994). None are known to contain diamonds, although alluvial diamonds are widespread in the area. The Prairie Creek lamproite, Arkansas (a.k.a. Murfreesboro; 100 Ma) was mined in the early twentieth century for diamonds, producing about 10000 ct from breccia phases over 15 to 20 years of intermittent operation (Mitchell and Bergman, 1991).

In Canada, Proterozoic ultrapotassic rocks are found within the Dubawnt Supergroup, Keewatin District, NWT. Many rocks in this very extensive province resemble lamproites, and a dyke near Gibson Lake has yielded a large number of microdiamonds (MacRae et al., 1996). A single octahedral microdiamond was recovered from a volcaniclastic breccia pipe at Outlet Bay, Dubawnt Lake (Peterson and LeCheminant, 1996). Lamproite dykes with unaltered leucite occur near Iqualuit on Baffin Island (Hogarth and Peterson, 1996). Strongly potassic, tuffisitic breccias are preserved as highly altered lenses in the Canadian Cordillera near Golden, B.C. (McCallum, 1994; Ijewliw and Pell, 1996). Rb-Sr ages indicate these are Paleozoic intrusions, predating Tertiary thrusting and metamorphism. Recovery of 4 small diamonds from 2 intrusions (the Jack and Mark 1 "pipes") has been reported, but these may reflect contamination, and identification of these intrusions as lamproites is highly uncertain.

### *Other*

Lamproites and related rocks occur throughout much of Europe, although no diamond-bearing centres have been reported (Mitchell and Bergman, 1991). It may be expected that lamproites are present within Russia but no definitive descriptions have yet been published. The Gaussberg volcano, Antarctica, is Quaternary to Recent (10000-50000 years) and relatively undissected. Very fresh, glassy lamproites occur at this site and are the youngest lamproites yet described (Sheraton and Cundari, 1980).

## EXPLORATION

From the above summary and discussion, it is likely that exploration for diamondiferous lamproites will be most successful in Proterozoic mobile belts, particularly where thick, poorly consolidated sedimentary sequences overlie major fault zones (e.g., northern Australia: Jaques et al., 1986). The history of diamond exploration in the region of Argyle and Ellendale (Jaques et al., 1986; G. Boxer, pers. comm. 1993) indicates that the best indicator mineral (aside from diamond itself) is magnesiochromite, which is a heavy, resistant mineral that appears as phenocrysts in most high-Mg lamproites. Xenocrystic garnets, chromites, chrome diopsides, and ilmenites are much less abundant because of the poor preservation of mantle xenoliths in lamproites.

Aeromagnetic surveys were highly successful in locating individual pipes within the Ellendale field (Jaques et al., 1986). Flight line spacings were typically 50 to 100 m. Due to their extreme composition, outcropping lamproites may be readily detected by radiometric or geochemical surveys.

In Canada, lamproitic rocks with diamond potential have been positively identified only in the eastern Northwest Territories (MacRae et al., 1996; Hogarth and Peterson, 1996). The Dubawnt Supergroup ultrapotassic igneous province, which extends from northeastern Hudson Bay through the central Churchill Province and possibly as far south as northern Saskatchewan, is particularly attractive for diamond exploration. This province transects numerous Proterozoic fault systems and sedimentary basins. Although Proterozoic mobile belts have not been unequivocally recognized in the region, they do surround it, and there is strong geochemical evidence for Proterozoic subduction-related enrichment of Churchill Archean lithospheric mantle (Peterson and LeCheminant, 1996).

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# The relation of diamond-bearing rocks to other alkaline rocks

**K.L. Currie**

*Currie, K.L., 1996: The relation of diamond-bearing rocks to other alkaline rocks; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 87-90.*

## INTRODUCTION

Among primary igneous rocks, diamonds occur in mafic to ultramafic alkaline rocks (kimberlites, lamproites, minettes) which are lamprophyric in the broad sense of Rock (1987). In general, lamprophyres contain several generations of mafic phenocrysts (olivine, pyroxene, amphibole, biotite, phlogopite), lack quartz or feldspar phenocrysts, and have high volatile contents expressed by phases such as amphibole, biotite, phlogopite, carbonates, and sulphates. Diamond-bearing rocks have passed through pressure-temperature conditions at which diamond is stable, and travelled to near-surface conditions quickly enough to metastably preserve some of the diamonds. Equilibrium subcontinental geotherms require depths of 150 to 200 km (45-55 kbar at 1050-1200°C) to stabilize diamond, but in a subduction zone diamond may form in the cold subducting slab at depths of 80 to 90 km (22-25 kbar at 300-400°C) and persist for millions of years (Barron et al., 1994). Most alkaline rocks originate at depths of greater then 100 km (compare Menzies, 1987), and hence can potentially entrain and conserve diamond in xenoliths or xenocrysts if ascent and emplacement are sufficiently rapid. Understanding the distribution and emplacement of alkaline rocks helps to divide potentially diamondiferous varieties from those which are not.

## DISTRIBUTION OF ALKALINE ROCKS IN CANADA

The alkaline rocks of Canada, including lamprophyres, occur in belts, of which nine were distinguished by Currie (1976; Fig.1). These belts are limited geographic areas within which alkaline rocks have been observed to occur. Although they have some links to igneous provinces and to specific tectonic events, they cannot be simply correlated with either. Belts of alkaline rocks typically exhibit polycyclic magmatism with nearby occurrences differing in age by hundreds of millions of years. The range of composition within each belt, typically from silicate-free carbonatite through ultramafic to mafic-free alkaline rocks, suggests that all the alkaline rocks cannot be derived from the same mantle source. In some cases specific occurrences, or groups of occurrences, can be linked to tectonic events such as continental rifting. However, belts contain occurrences significantly older or younger than the supposed tectonic cause. According to current petrogenetic theory, most alkaline magmatism results from partial melting of metasomatized mantle (Bailey, 1987). The polycyclic nature of belts of alkaline rocks can be explained by the attachment of substantial volumes of such metasomatized mantle to continental crust. Whenever this source is tapped by a suitable tectonic event, alkaline magmatism results. Metasomatism is a small-scale and heterogeneous process. The diversity of alkaline rocks can be explained by tapping of different combinations of lithospheric and asthenospheric sources.

The nature of some possible tectonic "triggers" can be illustrated by three different tectonic settings. The Ottawa-St. Lawrence belt, the Red Wine-Gardar belt (Labrador and Greenland), the Coldwell-Nagagami belt, and possibly the Kapuskasing High belt are in continental rifted regions, a classic setting for polycyclic alkaline magmatism. The Ottawa-St. Lawrence belt originated during Neoproterozoic continental break-up, and was reactivated during Jurassic-Cretaceous opening of the Atlantic Ocean. Alkaline magmatism occurred during both episodes, and possibly also in mid-Paleozoic time (Bon Conseil, 430 Ma, Currie, 1976). Alkaline magmatism also occurs associated with continental collision zones, commonly within local rifts. The Cordillera, Grenville, and Circum-Ungava orogens have well developed alkaline suites along craton margins. Age determinations show that the age of some alkaline rocks overlaps the age of deformation, but polycyclic magmatism occurs, for example the Ice River complex, emplaced at 368 Ma (Parrish and Armstrong, 1987), occurs in the midst of Eocene and younger alkaline rocks in the Cordillera. A third type of alkaline suite results from upwelling of magma from great depth (continental intraplate magmatism), either in tightly constrained, moving plumes such as the Monteregian-White Mountains magma series (Sleep, 1990) of the

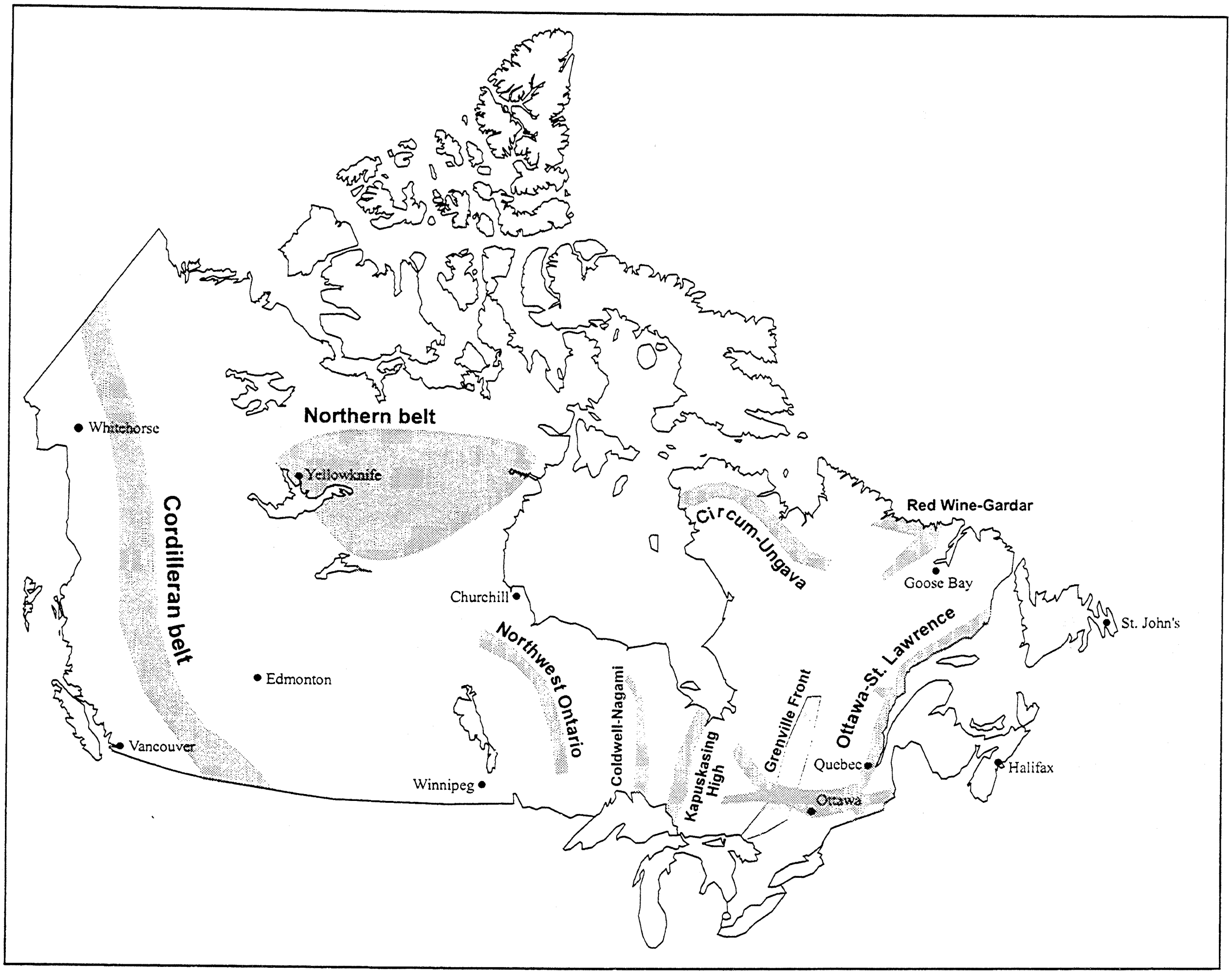


*Figure 1. Distribution of alkaline rocks in Canada (revised from Currie, 1976).*

Ottawa-St. Lawrence belt, or in more diffuse areas such the Northern belt (Fig.1). Plume magmatism is relatively short-lived and not polycyclic, but continental intraplate magmatism generally is polycyclic. The causes of such magmatism remain speculative, but are associated with thermal and mechanical processes in the mantle, possibly triggered or modulated by more superficial phenomena such as thermal blanketing by supercontinents.

## PETROGENESIS

The diversity of alkaline igneous rocks results from complex combinations of igneous fractionation, mixing, assimilation, and metasomatism. Mafic to ultramafic alkaline lamprophyres, including kimberlites, lamproites, and minettes are a component of belts of alkaline rocks where the ultramafic carbonatite and nephelinite carbonatite families (Currie (1976), and to a lesser extent the alkali basalt family, occur. The unusual chemical composition of these rocks probably results from melting of variably metasomatized mantle protoliths that are highly enriched in potassium and other incompatible elements (Ba, Rb, Nb, REE) under a variety of P-T conditions. The chemistry (and mineralogy) of mantle-derived alkaline ultramafic rocks is extremely diverse, with each occurrence having some unique characteristics, presumably because each derives from a small-volume mantle source which has undergone a unique metasomatic history. The nomenclature of such rocks is presently in a state of flux (compare Woolley et al., 1996; LeMaitre et al., 1989). True kimberlites seem to be restricted to continental intraplate magmatism, whereas lamproites occur on the continental edge of collision zones (Peterson, 1996). However, any potassic (molecular K>Na) mafic to ultramafic rock should be considered potentially diamond-bearing.

Many alkaline ultramafic rocks can be eliminated as possible sources of diamond by mineralogical criteria. Potentially diamond-bearing rocks are usually rich in olivine, clinopyroxene, phlogopite, and carbonate.

Rocks with primary calcic plagioclase (alkali basalt family) have never been reported to contain diamonds. Other primary minerals not known to be compatible with diamond include nepheline, leucite, feldspathoids (hauyne, sodalite), zeolites, and melilite, although leucite is not uncommon in rocks associated with diamondiferous lamproite, and some of these minerals may occur as secondary alteration. Potassium feldspar and diamond cannot coexist stably, but diamond-bearing lamproites and related rocks contain sanidine or orthoclase (see Peterson, 1996; MacRae et al., 1996), presumably formed during ascent and quenching. Kimberlites and some lamproites are unique among alkaline rocks in containing modal orthopyroxene, a mineral which cannot crystallize from alkaline magmas at low pressure (Edgar, 1987). Clinopyroxenes and phlogopite of diamondiferous rocks are commonly Cr-rich, rather than Ti-rich as in chemically similar rocks crystallizing at lower pressures. Garnet xenocrysts in diamond-bearing igneous rocks commonly contain >0.04 weight % $Na_2O$ (Barron et al., 1994).

Diamond-bearing rocks are never directly associated with central alkaline complexes. However, ultramafic alkaline dykes with chemical compositions overlapping those of kimberlite and minette are relatively common, particularly among dyke suites associated with carbonatite complexes (compare Yoder, 1986). Central complexes imply existence of a relatively shallow magma chamber, wherein magma resides for a significant period, and relatively slow cooling of the magma. Any diamonds originally present would be converted to graphite.

The physical form of an intrusion contributes to its diamond potential. All known diamond-bearing igneous rocks form dykes, sills, plugs, vents, diatremes, and associated breccias. Lavas are known only for minettes and lamproites. Lamprophyric intrusions with outcrop areas larger than 5 $km^2$ are unknown. The necessity of rapid transit to the surface in order to preserve diamond ensures that most diamond-bearing rocks will be brecciated or associated with breccias, signifying rapid ascent, whether $CO_2$-driven, as in the case of kimberlites, or water-driven as is probable for lamproite and minette. Deformation and metamorphism decrease the diamond potential of a body. Diamonds are unlikely to survive greenschist or higher grade metamorphism at oxygen fugacities normally obtainable within the crust. Eclogite metamorphism at extreme pressures and moderate temperatures is a special case where supracrustal basaltic rocks have passed through the diamond stability field at relatively low temperatures during subduction (compare Berman, 1996). Diamond-bearing eclogites are not primary igneous rocks, although they may serve as a source of diamonds in younger igneous rocks. The age of a body relative to other local alkaline rocks is not relevant to its diamond potential, and examples are known where diamonds occur in nearby rocks of radically different age (Argyle and Ellendale, see Peterson, 1996).

## SUMMARY

Diamond-bearing igneous rocks occur associated with other alkaline rocks in well-defined regions (belts of alkaline rocks). Any volatile-rich potassic ultramafic alkaline rock within such a region should be considered potentially diamond-bearing. The terminology of such ultramafic lamprophyric rocks is presently confused and confusing. Rocks associated with central alkaline complexes, whether in the complex or its associated dyke swarm, have no potential for diamond, even though they may have appropriate chemistry. Rocks containing primary plagioclase, nepheline, leucite, melilite, feldspathoid, or zeolite may also be eliminated from consideration. Dykes, sills, plugs, and diatremes are the most prospective candidates, particularly if they are associated with breccia. The age of the body relative to nearby alkaline rocks is not a reliable diagnostic criteria, and sites should be checked for several ages of potential hosts.

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# Diatreme breccias in the Cordillera

**O.J. Ijewliw and J. Pell**

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## INTRODUCTION

Since the 1970s, diatreme breccias in eastern British Columbia and the Northwest Territories have generated exploration interest due to their presumed similarity to kimberlites and lamproites, the possible carriers of diamonds (Pell, 1986; Godwin and Price, 1987).

Most Cordilleran diatremes are east of the Rocky Mountain/Tintina trench-system and are aligned in a NNW-trending belt (Fig. 1). Lithologies forming these breccia pipes include alkaline basalts, alkaline and ultramafic lamprophyres and a few occurrences of true kimberlites. Microdiamonds have been reported from several of these pipes, including the Mountain diatreme, NWT (Godwin and Price, 1987), Lens Mountain diatreme, BC (Jack Claim), the Valenciennes River diatreme, BC (Mark Claim) (Northcote, 1983a,b; Dummett et al., 1985; George Cross News Letter, Jan. 23, 1990) and the Ram 5 kimberlite, British Columbia (George Cross News Letter, Nov. 24, 1994).

## LITHOLOGY, HOST ROCKS, AND AGE OF EMPLACEMENT

In general, the diatremes are recognized in the field by their brecciated texture, moderate to high degree of foliation, and characteristic alteration colours of brown, red, or green, which are distinct from the predominately buff to grey limestones and shales which they intrude. Some weather recessively; others are more resistant than the host rocks. Due to alteration by deuteric or metasomatic processes, combined with the effects of low-grade regional metamorphism, lithological classifications are often difficult, and classifications based on petrographic observations may be more reliable than whole-rock geochemistry. Lithological designations are based, in part, on the work of Rock (1987).

The Mountain diatreme in the Mackenzie Mountains, Northwest Territories (64°14'N, 129°28'W), intrudes Upper Cambrian to Middle Ordovician carbonates. It has a central core of dark green breccia containing rebrecciated and carbonated-cemented breccia and a massive dyke phase that is surrounded by a marginal rusty weathering breccia. An epiclastic reworked tuff ring is also present (Godwin and Price, 1987). Breccia fragments consist of carbonate country rock xenoliths and autoliths cored with carbonate or phlogopite pseudomorphs after olivine and pyroxene. The diatreme was originally considered to be of kimberlitic affinity due to the presence of microdiamonds, picroilmenite, pyrope, and chrome diopside (all xenocrysts), although mineralogically and geochemically it is not a typical kimberlite (Godwin and Price, 1987). Recent work (Goodfellow et. al., 1995) suggests that the Mountain diatreme is an alkaline basalt. Original K-Ar and Rb-Sr geochronological studies yielded ages from 425 to 445 Ma (Godwin and Price, 1987). However, a more recent evaluation of the data from which these dates were derived, using newer, more appropriate constants suggests that an Early Devonian age (circa 400-410 Ma) is more likely (R.L. Armstrong, pers. comm. 1988).

The northernmost pipe in British Columbia, the Kechika River diatreme in the Kechika Ranges of the Cassiar Mountains (58°42'N, 127° 30'W) is the only documented occurrence west of the Rocky Mountain Trench (Pell, 1994). It is related to a suite of alkaline rocks which include syenite, trachyte, malignite, related tuffs and agglomerates, and carbonatite dykes, that are hosted by Silurian carbonates and metamorphosed to greenschist facies. Field relationships such as interbedding of agglomerates and sedimentary strata suggest that these igneous rocks are similar in age to the host strata (Silurian or slightly younger). The diatreme comprises heterolithic tuffisitic breccias with quartzite, carbonate and syenite xenoliths. Quartz xenocrysts, rare chrome spinels, juvenile and vesiculated glass lapilli, and crystal fragments of potassium feldspar and phlogopite occur in a matrix of carbonate minerals, potassium feldspar and muscovite (Pell, 1994). Alteration precludes specific classification of the diatreme: mineralogically it bears some affinity to alkaline basalts; geochemically these rocks have an affinity to alkaline lamprophyres (Pell, 1994).

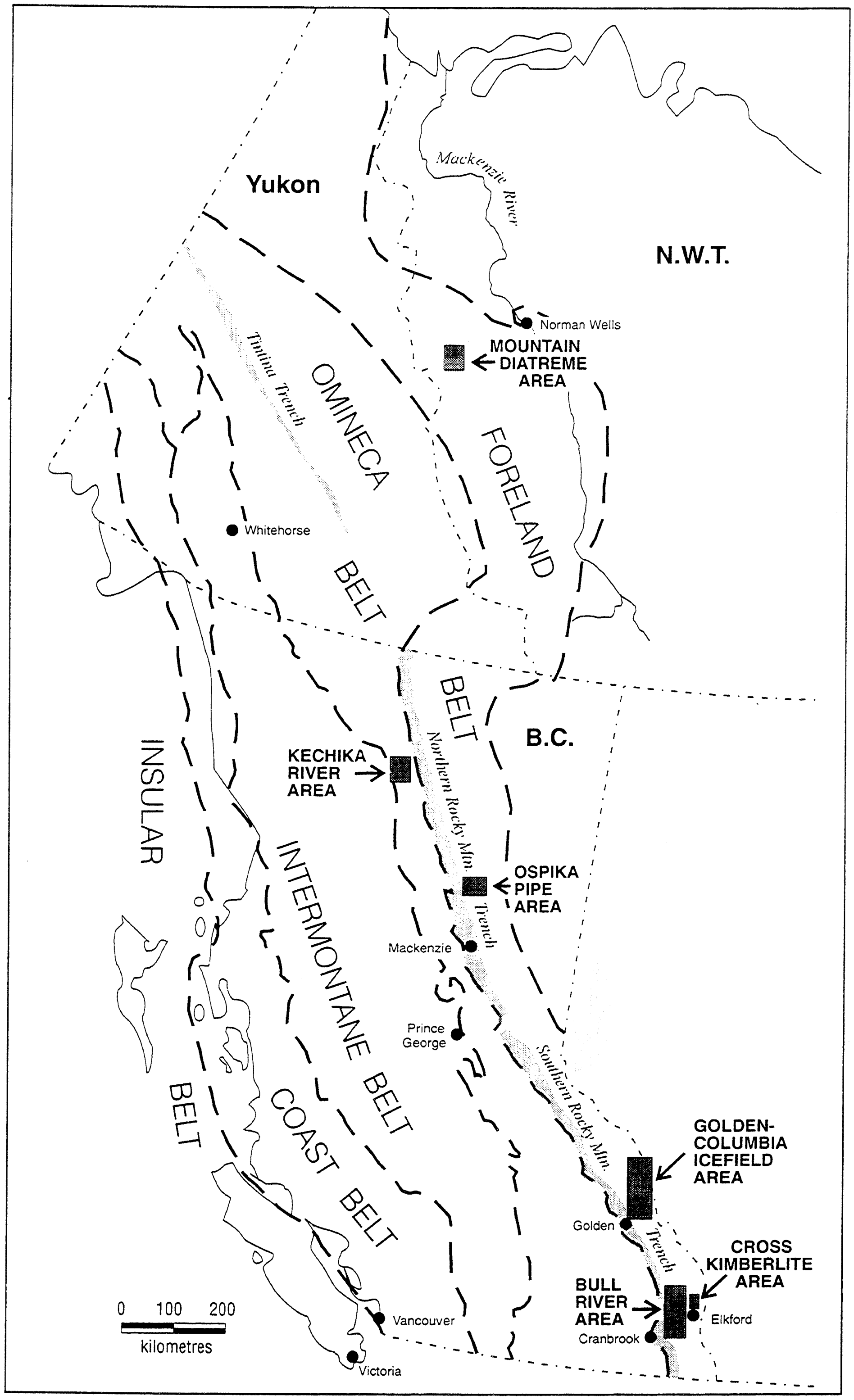


*Figure 1. Distribution of breccia diatremes in the Cordillera of western Canada (modified after Pell, 1994).*

The Ospika pipe is in the Rocky Mountains, northeastern BC, east of Williston Lake (56°27'N,123°45'W), a few hundred metres from the large Aley carbonatite complex. It intrudes Ordovician carbonate rocks and has yielded a Rb-Sr age of $334 \pm 7$ Ma (Pell, 1994). The Ospika pipe is a complex diatreme that comprises at least five distinct breccia and massive phases. The breccias contains fragments of sedimentary rocks, some with reaction rims, rare cognate xenoliths, and globular segregation structures. Phlogopite, titaniferous augite, rare altered olivine and bright green diopside occur in a fine grained carbonate matrix. Petrography and whole-rock geochemistry indicate an ultramafic lamprophyre, specifically aillikite, for the Ospika pipe (Pell, 1994).

Diatreme breccias and dykes hosted by Upper Cambrian carbonate rocks occur in at least five different areas near Golden, southeastern BC (51°41'N, 116°57'W to 52°05'N, 117° 23'W), Bush River, Mons Creek, Valenciennes River, Lens Mountain, and Campbell Icefield. Within this region, some lithological variation has been observed, but alteration often precludes unequivocal classification. In the Bush River area, diatremes and dykes contain abundant fragments of sedimentary and crystalline rocks; olivine pseudomorphs, biotite, spinel, and rare, fresh plagioclase occur in a dusty carbonate/chlorite matrix. Based on mineralogy these rocks have been classified as olivine kersantites, a calc-alkaline lamprophyre (Ijewliw, 1991). Geochemical results, however, point to a more alkaline affinity (McCallum, 1994; Pell, 1994). Diatremes and dykes in the Mons Creek and Valenciennes River areas are commonly very altered and comprise olivine pseudomorphs, clinopyroxene, biotite, and spinels in a groundmass of calcite/chlorite, biotite, plagioclase, and serpentine. They have been classified as camptonite, an alkaline lamprophyre (Ijewliw, 1991). Geochemical data, although ambiguous, in general substantiate an alkaline lamprophyre designation (McCallum, 1994; Pell, 1994). It has been suggested that a diatreme at Lens Mountain is lamproitic, based on the chemical composition of minerals recovered from concentrate (xenocrysts) and whole-rock geochemical analyses (McCallum, 1994). Similarities between the Lens Mountain diatreme and those in the Mons Creek and Valenciennes River areas suggest that this is not an unequivocal classification. The HP Pipe, south of the Campbell Icefield, is the least altered of the Golden diatremes. It consists of at least five breccia phases and crosscutting dykes. Fragments of marmorized limestone clasts, quartzite, altered plutonic rocks, and cored autoliths, megacrysts, and phenocrysts of clinopyroxene, melanite garnet, biotite, spinel, and apatite, occur in a groundmass of calcite, chlorite, serpentine, talc, and pyrite (Fig. 2). Based on mineralogy, this rock is an aillikite, a carbonate-rich ultramafic lamprophyre (Ijewliw, 1991). This classification is supported by geochemical analyses (Pell, 1994). The HP pipe and dykes from the Bush River area have yielded isotopic ages around 400 to 410 Ma (Pell, 1994).

***Figure 2.*** *Chrome diopside megacryst in core of dark green autolith with white marmorized limestone clasts from the HP Pipe in the Golden cluster, British Columbia.*

Forty or more breccia pipes and related dykes occur in the Bull and Elk River areas, southeastern British Columbia (49°30' to 51°32'N, circa 115° 15'W). They differ from the Golden diatremes by being free of obvious hydrous minerals such as biotite. Most are tuffisitic and contain fragments of sedimentary country rocks, altered plutonic rocks, some cognate xenoliths, altered vesicular glass lapilli, and altered clinopyroxene, olivine, calcite, and spinel, in a groundmass of carbonate, chlorite, talc, and minor plagioclase. Locally, bedded crater-facies volcaniclastic sediments and some volcanic flows are present within these diatremes. Although the high degree of alteration makes classification difficult, mineralogically and geochemically these rocks show affinities with alkaline basalts (Pell, 1994). The age of these pipes can only be inferred from stratigraphic evidence, which suggests that there were two periods of emplacement, one in the Late Ordovician, circa 450 Ma, and the other in the Early Devonian, circa 400 Ma (Helmstaedt et al., 1988; Pell, 1994).

The Cross kimberlite, near Elkford in southeastern BC (50°05'N, 114°59'), intrudes Permian carbonates in a fault block east of the Bull River/Elk River pipes. It is a multiphase intrusion containing country rock fragments, peridotite xenoliths, and cored autoliths. Minerals include serpentinized olivine (Fig. 3), phlogopite, partially altered garnets, some pyropes with kelyphitic rims and spinel in a groundmass of calcite and serpentine (Ijewliw, 1987; Hall et al., 1989; Hall, 1991; Pell, 1994). Geochemical analyses corroborate the kimberlite designation (Pell, 1994). The Cross kimberlite has been dated using Rb-Sr at 245 Ma (Grieve, 1982; Smith et al., 1988). Until recently it was the only true kimberlite discovered in British Columbia; however, in 1994, four additional kimberlites were reportedly found nearby (George Cross News Letter, Nov. 24, 1994).

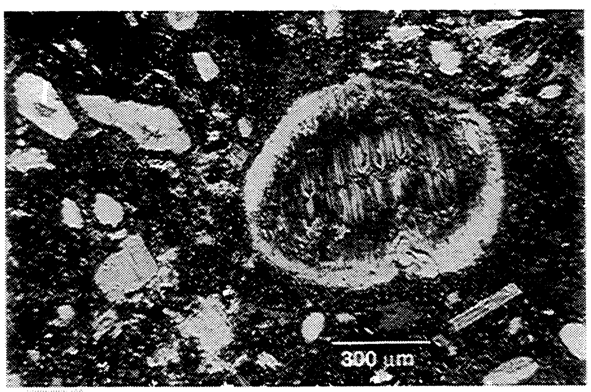


***Figure 3.*** *Photomicrograph of serpentinized olivines from the Cross kimberlite, British Columbia (field of view is 1.8 mm).*

## CONCLUSIONS

Diatreme breccias in British Columbia and the western Northwest Territories define a NNW-trending belt that lies mainly east of the Rocky Mountain/Tintina trench system. The breccias were emplaced into sedimentary strata of the Cordilleran passive continental margin, of western North America. Magmatism occurred at various times, in particular during the Late Ordovician (circa 450 Ma), Early Devonian (circa 400 Ma), Devono-Mississippian (circa 345 Ma), and Permian (circa 245 Ma). The intrusions were metamorphosed, deformed, and transposed eastward, along with their host rocks during the Cretaceous Columbian orogeny. A diverse group of lithologies are represented by these diatremes, ranging from alkaline basalts through alkaline and ultramafic lamprophyres to kimberlites proper. Some microdiamonds have been discovered and exploration interest continues.

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# Ultrapotassic rocks of the Dubawnt Supergroup, District of Keewatin, N.W.T.

T. D. Peterson and A.N. LeCheminant

*Peterson, T.D. and LeCheminant, A.N., 1996: Ultrapotassic rocks of the Dubawnt Supergroup, District of Keewatin, N.W.T.; in Searching forDiamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 97-100.*

## INTRODUCTION

Between 1.84 and 1.7 Ga, the central Churchill Province was overlain by sandstone, conglomerate, and volcanic rocks of the Dubawnt Supergroup (Fig. 1). The oldest volcanic rocks (Christopher Island Formation [CIF] 1.84 Ga) are ultrapotassic, and sufficiently resemble some diamondiferous lamproites that they are valid targets for exploration. To date, one unconfirmed microdiamond occurrence has been reported from a CIF diatreme breccia at Dubawnt Lake (Northern Miner, 1993; J.Davis, pers. comm., 1994) and a microdiamond, embedded in the rock matrix, was discovered in 1993 in a CIF lamprophyre dyke southeast of Baker Lake (MacRae et al., 1994). Subsequent sampling of this dyke in 1994 revealed a remarkably high concentration of microdiamonds (MacRae et al., 1996).

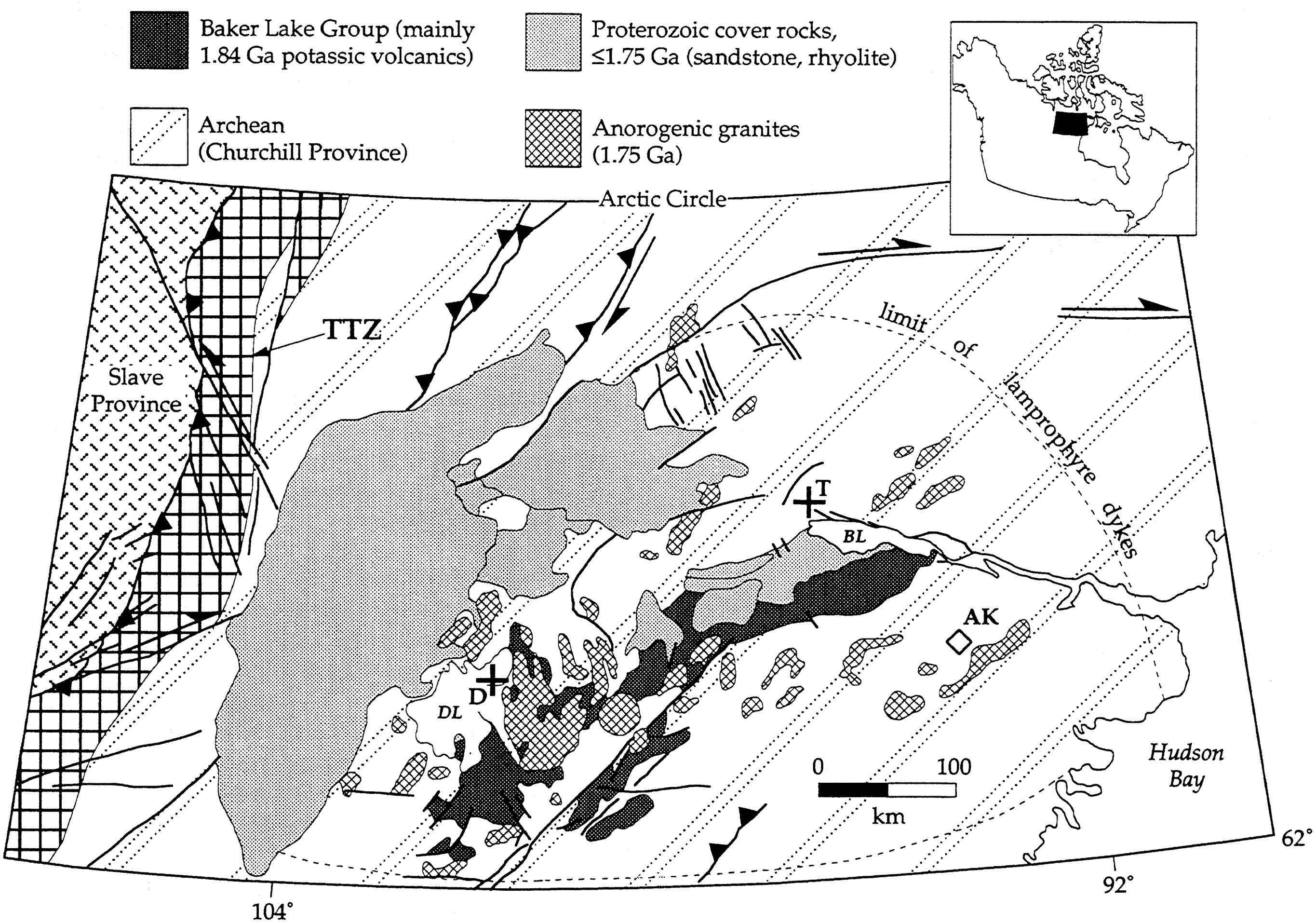


*Figure 1. Geological map of the central Churchill Province showing rocks of the Dubawnt Supergroup (Baker Lake Group and other Proterozoic cover rocks). TTZ=Thelon Tectonic Zone, BL=Baker Lake, DL=Dubawnt Lake. The approximate outer limit of lamprophyre dykes is from LeCheminant et al. (1987). Symbols and letters indicate locations of the Dubawnt Lake diatreme (+D), the Thelon River lamprophyre dyke (+T) and the Akluilâk diamondiferous lamprophyre dyke (◊AK).*

## GEOLOGICAL SETTING

The CIF is exposed in several fault-bounded basins extending ENE from southern Dubawnt Lake to Baker Lake (Fig. 1). Most of the basins are elongated, with a normal fault on the NW edge, but a large triangular basin southeast of Dubawnt Lake developed by E-W compression and N-S extension (Peterson, 1994). The large volume of volcanic rocks (>4000 $km^3$) is unmatched by any other ultrapotassic rock province in the world. Many volcanic facies have been recorded, including lapilli tuffs and diatremes rich in country rock fragments.

Within a region at least 600x400 km surrounding the basins, phlogopite- and clinopyroxene-rich lamprophyre dykes intrude predominantly Archean country rocks (LeCheminant et al., 1987). Volcanic rocks of the CIF, and the lamprophyric feeder dykes, mainly have mineral assemblages typical of minettes (Peterson and LeCheminant, 1993). Phlogopite compositions partly overlap those of minettes, although the compositional arrays follow lamproite and Group II kimberlite trends (Fig. 2). Geochemically, many of the flows and dykes closely resemble Mediterranean-type lamproites (Peterson et al., 1994), which typically erupt within orogenic zones 10 to 50 million years after terminal collision (Mitchell and Bergman, 1991). Such rocks have high Al and Na, and depletions in Ti and Nb, relative to lamproites which erupt 100 million years or more after collision (such as Ellendale and Argyle in northern Australia). The CIF dykes and flows have near-constant $\epsilon_{nd}$ near -8, and wide variation in $\epsilon_{Sr}$ (-40 to +200); Pb is extremely nonradiogenic (Peterson et al., 1994). These features are common to many lamproite suites, but not to minettes.

## MANTLE XENOLITHS AND XENOCRYSTS

Although crustal xenoliths are abundant in CIF breccias and dykes, mantle-derived peridotitic xenoliths and xenocrysts have only been identified from rare, altered fragments rich in talc and carbonate. Phlogopite-rich (glimmerite) xenoliths are locally abundant (Peterson and LeCheminant, 1993). Phlogopite in glimmerites follows the evolutionary trend of Group II (micaceous) kimberlites (Fig. 2) and contains inclusions of calcite, strontianite, sodian chromian diopside, and spinel ranging in composition from magnesiochromite to chromite with significant Zn and Mn. The glimmerite xenoliths and their carrier magmas are interpreted to be genetically related.

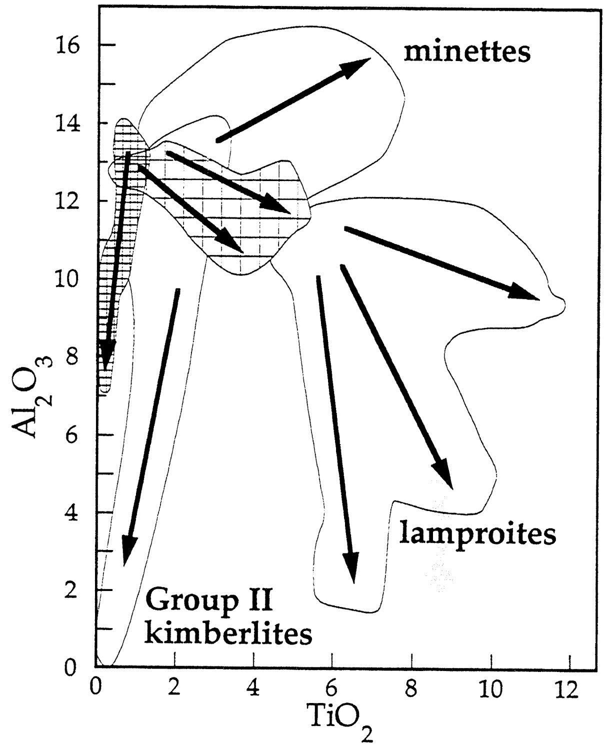


*Figure 2. Composition of phlogopites in CIF lavas and lamprophyre dykes (coarse hatch) and in glimmerite xenoliths (fine hatch) compared to phlogopites from other ultrapotassic rocks. Arrows indicate compositional trends from high-Mg to low-Mg micas. Fields from Mitchell and Bergman (1991) and Peterson and LeCheminant (1993).*

Glimmerites are also found in kimberlites, and have been interpreted as samples of lamproite dykes or veins in lithospheric mantle (e.g., Waters, 1987). The mineralogy of the glimmerite xenoliths unfortunately does not permit precise pressure estimates.

Heavy mineral processing of samples from the Dubawnt Lake diatreme yielded abundant (Na, Cr)-rich clinopyroxene and magnesiochromite, with minor eclogitic and peridotitic (lherzolitic and harzburgitic) garnet, and high-(Mg, Cr) picroilmenite (Chisholm, 1993). The magnesiochromites overlap the compositional field of diamond inclusion chromites (as do chromites in the glimmerite xenoliths), but form part of the normal compositional trend of lamproitic spinels. Probable xenocrystic spinel has only been identified in a lamprophyre dyke exposed on the Thelon River (Fig. 1). This spinel has an aluminous chromite core similar in composition to spinel within a harzburgite nodule from a kimberlite in Lesotho (Carswell et al., 1979). The symplectite reaction/overgrowth rind consists of (Zn, Mn)-rich chromite, sodian chromian diopside, apatite, and chromian grossular (Fig. 3). These spinels may record subduction-driven metasomatism of oceanic lithosphere (e.g., Schulze, 1986) containing Mn- and Zn-rich sediments, or may indicate reaction of depleted continental lithosphere with rising ultrapotassic magmas and/or fluids.

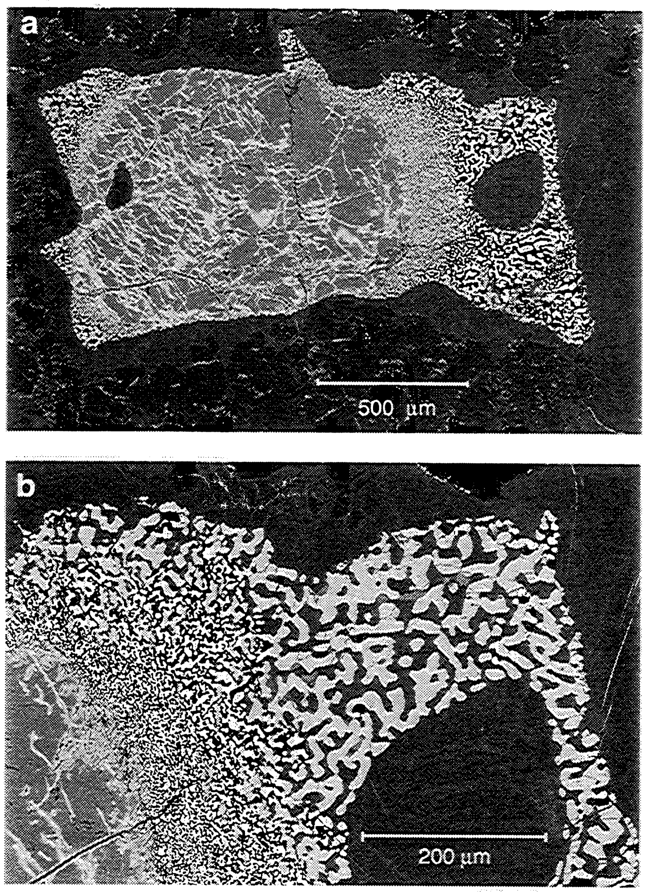


***Figure 3.*** *a) Back-scattered electron image of a serpentinized ultramafic xenolith (black) from the Thelon River lamprophyre dyke, which contains a rounded xenocryst of aluminous Cr spinel (grey), partly rimmed by a coarsening- outward symplectite consisting mainly of chromite (white) and clinopyroxene (black), with minor garnet and apatite. The core spinel and the symplectite are enclosed by an irregular thin rim of Cr-rich, high Mg phlogopite (dark grey).* ***b****) Symplectite close-up. Mn and Zn-bearing Cr-rich spinel (white) is intergrown with Na-Cr-rich clinopyroxene (dark grey). Minor Cr-rich grossular garnet and apatite (medium grey) occur within the symplectite.*

## TECTONIC IMPLICATIONS

Eruption of the CIF at 1.84 to 1.83 Ga occurred after subduction beneath the western margin of the Churchill Province at 2.0 to 1.9 Ga (Thelon orogeny) and beneath the southeastern margin at 1.86 to 1.85 Ga (Trans-Hudson orogeny) (St.-Onge and Lucas, 1996). Isotopic data are consistent with the transport, by fluid metasomatism, of subducted sedimentary components into the source region (Peterson et al., 1994). The dominant source material for the ultrapotassic magmas, based on geochemical constraints, was probably phlogopitized harzburgitic mantle.

There is much room to speculate on the possible relationship between the CIF and diamonds. Eruption of the CIF may have been triggered by subduction-related metasomatism, with a widespread volcanological expression due to ongoing internal deformation of the Churchill Province. The large volume of magma, and possible associated strong heating of the lower lithosphere, are not favourable tectonic conditions for diamond preservation. However, the CIF magmas may have formed in, or been channeled into, shear zones, leaving large blocks of lithosphere relatively unaffected. Dykes and diatremes outside the major shear zones and basins could conceivably have sampled deep, cool, Archean lithospheric mantle containing diamonds older than 1.84 Ga.

Subduction-related enrichment of the sub-Churchill mantle may have resulted in the formation of Proterozoic diamonds. Eclogitic diamonds, which probably form during or after subduction of carbon-bearing oceanic lithosphere (Eldridge et al., 1991), are concentrated in lamproites relative to most kimberlites. Many eclogitic diamonds yield Proterozoic mineral inclusion isochron ages (Richardson et al., 1990), and most diamondiferous lamproites are Proterozoic. Diamonds formed within Churchill lithosphere at 1.84 Ga could have been sampled by magmas postdating subduction-related metasomatism. It is therefore noteworthy that the diamondiferous Akluilâk dyke (MacRae et al., 1996) geochemically resembles the stratigraphically youngest rocks at Dubawnt Lake, which are also the most lamproitic in the section (Peterson et al., 1994).

The CIF erupted within the interior of an Archean terrane, so these models require horizontal underplating of subducted lithosphere at a substantial distance (>300 km) from the site of subduction. However, lamproites throughout the Wyoming-Churchill provinces and western Greenland consistently yield isotopic (Carlson and Irving, 1994) and geochronological (Rudnick et al., 1993) evidence for subduction-related metasomatism during the Paleoproterozoic, and for contributions from subducted sediments. The simplest interpretation is that pockets of phlogopitized peridotite developed in sub-Churchill lithospheric mantle by subduction-related enrichment processes during the assembly of Laurentia, between 2.0 and 1.8 Ga.

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# The diamondiferous Akluilâk lamprophyre dyke, Gibson Lake area, N.W.T.

**N.D. MacRae, A.E. Armitage, A.R. Miller, J.C.Roddick, A.L. Jones, and M.P. Mudry**

*MacRae, N.D., Armitage, A.E., Miller, A.R., Roddick, J.C., Jones, A.L., and Mudry, M.P., 1996: The diamondifersous Akluilâk lamprohyre dyke, Gibson Lake area, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 101-107.*

## INTRODUCTION

The Akluilâk lamprophyre dyke (akluilâk meaning "richest" in Inuktitut) hosts a unique and rich concentration of microdiamonds. Located approximately 120 km northwest of Rankin Inlet, N.W.T. (see Fig. 1 of Peterson and LeCheminant, 1996), this is the first confirmed multidiamond occurrence in the central Churchill Province. A single microdiamond (280 µm diameter) was discovered in a hand sample from the dyke in 1993 (Armitage et al., 1994a; MacRae et al., 1994); processing of a 22 kg bulk sample from the same location taken the following summer yielded > 1500 diamonds, and 2 macrodiamonds (>500µm; Northern Miner, 1995a; MacRae et al, 1995). Similar diamond recoveries were obtained from a second bulk sample collected and studied independently (C. Fipke, pers. comm., 1995), while a third independently studied sample yielded 6680 diamonds, including 3 macrodiamonds from 7.8 kg of rock (Northern Miner, 1995b). This paper presents the preliminary chemistry and petrography of the dyke.

## REGIONAL GEOLOGY

The Akluilâk dyke intrudes metavolcanic and metasedimentary rocks of the Archean Gibson-MacQuoid Lake greenstone belt (Armitage et al., 1994b, 1995; Fig. 1; see also Fig. 1 in Peterson and LeCheminant, 1996). Lamprophyre dykes up to 2 m wide are common in the area and trend 340 to 355°. These dykes vary considerably in character, but are typically phlogopite-phyric and contain rounded xenoliths of country rock,

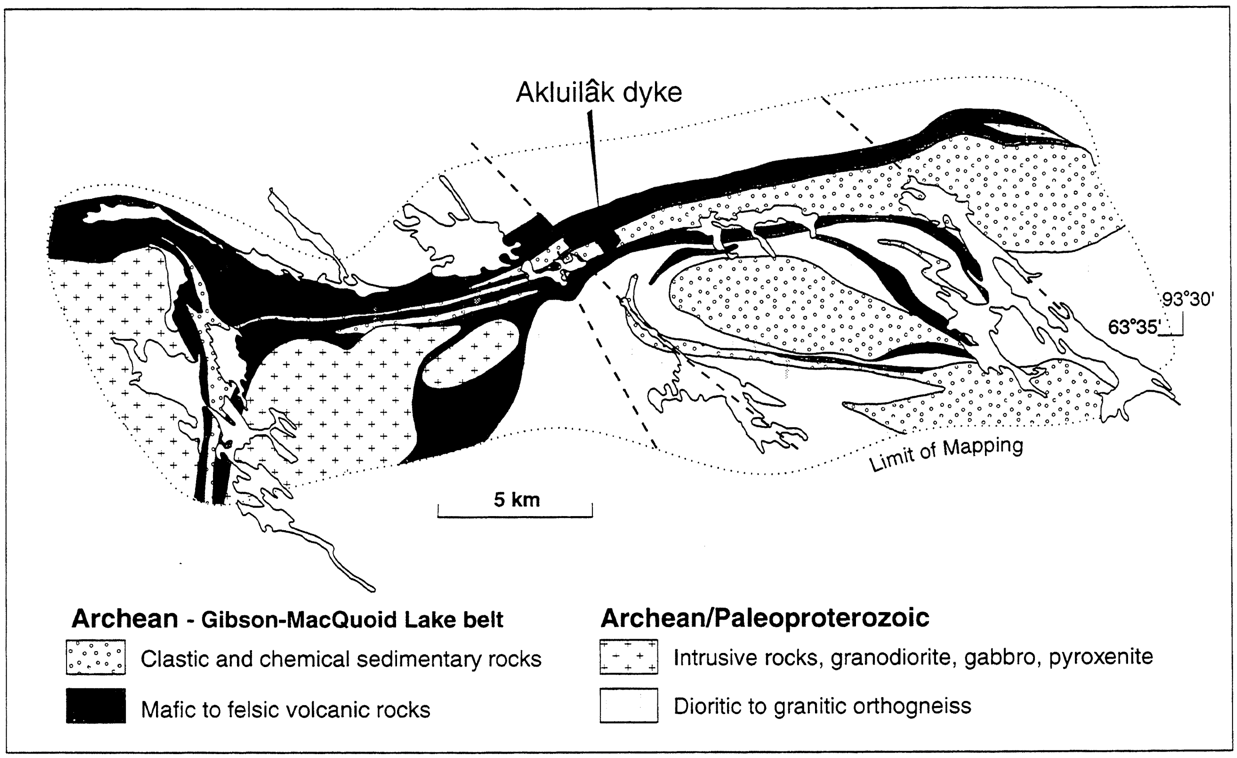


*Figure 1. Generalized geological map of the Gibson-MacQuoid Lake map area (after Armitage et al., 1995) showing the location of the Akluilâk dyke.*

including granite, orthogneiss, and metasediments. The dykes in this area have been correlated with the ca 1.84 Ga Christopher Island Formation (CIF) of the Dubawnt Supergroup (Armitage et al., 1994b).

The CIF consists primarily of potassic to ultrapotassic subaerial flows and pyroclastic rocks that are preserved in a series of continental basins in the Dubawnt Lake to Baker Lake region (Gall et al., 1992; Peterson et al., 1994; Peterson and Rainbird, 1990; LeCheminant et al, 1987); areally extensive dykes of equivalent composition, in part, are interpreted to represent feeder dykes to the CIF (Blake, 1980; LeCheminant et al., 1987; Gall et al., 1992). The CIF constitutes one of the most voluminous and oldest ultrapotassic sequences described (Peterson, 1994; Rock, 1991).

## FIELD OCCURRENCE AND PETROGRAPHY

The Akluilâk dyke strikes 350°, dips nearly vertically, and is 1.5 m wide at the diamond discovery site. Approximately 175 m to the north, it pinches to 30 cm width and is covered by glacial debris; south of the discovery site, it is also covered, but some 2 km along strike, an apparently identical 2.5 m wide dyke, with the same orientation, appearance, and bulk chemistry was traced for 3 km.

The discovery dyke weathers dark grey to black with a 'knobbly' poikilitic texture, and contains sparsely distributed rounded to ovoid xenoliths of granitic gneiss and gabbro, none of which show reaction envelopes. It is dominated by 1 to 5 mm orthoclase oikocrysts enclosing abundant biotite, apatite and calcite, with accessory titanite, rutile, ilmenite, and zircon. Larger oikocrysts locally show microperthitic albite at grain margins. Interstitial to the oikocrysts are aggregates of biotite, calcite, titanite, and apatite. Fine grained (<100 µm) quartz and albite are sparsely interspersed as thin selvages between biotite laths. Total mode of major minerals is 42% biotite, 40% orthoclase, 10% calcite, and 7% apatite. Mineralogically, this rock contains virtually none of the primary phases which characterize lamproites (Mitchell and Bergman, 1991), except for abundant apatite. In addition, the presence of primary albite - present in trace quantities in the Akluilâk dyke - is a criterion for exclusion from the lamproite group (Mitchell and Bergman, 1991).

Several microdiamonds were noted in thin section, and they are readily recovered in the >3.2 S.G. heavy mineral separate. The diamonds are primarily yellow-brown, but include a few pale yellow and pale green crystals. Larger grains have partial to complete thin coatings of graphite. Primary habits of octahedra (Fig. 2), macles, and cubes are all present, but most show the transitional tetrahexahedroidal form (Fig. 2) which develops during resorption.

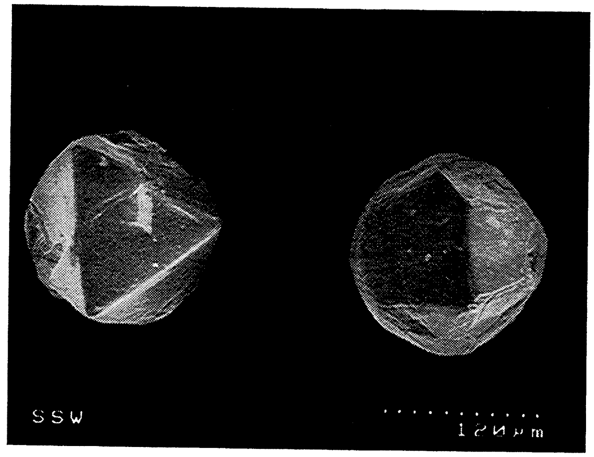

*__Figure 2.__ Scanning electron micrograph of octahedral (left) and tetrahexahedroidal (right) microdiamonds from the Akluilâk dyke (325D).*

## CHEMISTRY

Major element whole-rock chemistry for the Akluilâk dyke is presented in Table 1. For comparison, we list in Table 1 major element data for average lamproite and average minette and the composition of a lamproitic dyke rock (K89/8) from the CIF in the Dubawnt Lake region (Peterson et al., 1994), which shows similarities to 325D. Both 325D and K89/8 are ultrapotassic and peralkaline, but only K89/8 is close to being perpotassic. Of the chemical criteria noted by Mitchell and Bergman (1991) as characteristic of lamproites, 325D meets slightly more than half, including appropriate ranges of Niggli *mg* and Niggli *k* values. CIPW calculations show 325D to be strongly leucite normative in contrast to average lamproite and to K89/8.

In terms of trace elements, 325D has high concentrations of Ba and Sr (9480 ppm and 1650 ppm, respectively) - typical of diamondiferous host rocks (Mitchell and Bergman, 1991), but low Cr and Ni contents (both 87 ppm). REEs are very similar for both 325D and K89/8, showing strong LREE enrichment, and are higher than average minette (Fig. 3).

**Table 1.** Major element composition of Akluilâk dyke and comparable rocks

| | 325D | Average lamproite | Average minette | K89/8 |
|---|---|---|---|---|
| $SiO_2$ | 43.49 | 51.1 | 51.5 | 49.95 |
| $TiO_2$ | 1.08 | 4.1 | 1.0 | 1.48 |
| $Al_2O_3$ | 14.88 | 7.6 | 12.6 | 11.03 |
| $Fe_2O_3$ | 1.35 | 7.8 | 7.6 | 3.61 |
| FeO | 5.1 | | | 2.55 |
| MnO | 0.27 | 0.1 | 0.13 | 0.1 |
| MgO | 4.71 | 11.4 | 8.0 | 5.5 |
| CaO | 8.88 | 4.8 | 7.9 | 9.98 |
| $Na_2O$ | 0.73 | 0.65 | 2.0 | 1.38 |
| $K_2O$ | 9.53 | 7.3 | 6.0 | 9.16 |
| $P_2O_5$ | 2.79 | 1.3 | 1.2 | 2.75 |
| $H_2O$ | | 3.2 | 2.2 | 1.5 |
| $CO_2$ | | 0.5 | 1.8 | 4.2 |
| LOI | 4.46 | | | |

325D analyses by XRF from XRAL Analytical Laboratories; FeO by wet titration. Average lamproite and average minette are from Rock (1991); total Fe is expressed as Fe2O3. K89/8 is from Peterson et al. (1994). All values in weight percent.

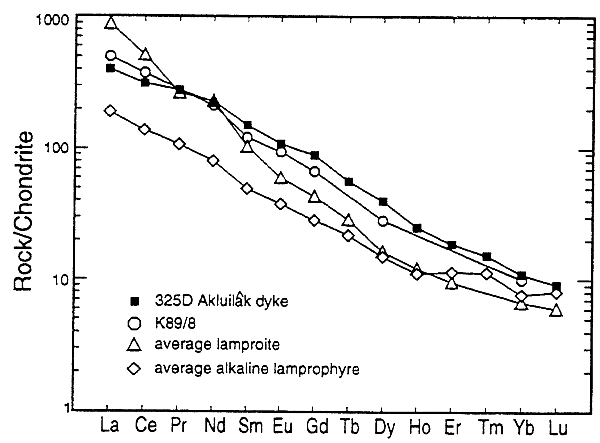


*Figure 3. Chondrite-normalized whole-rock REE abundances plot for the: Akluilâk dyke (325D); K89/8 (Peterson et al., 1994); average lamproite and average alkaline lamprophyre (Rock, 1991). REE normalizing values after Taylor and McLennan (1985)*

Microprobe analyses have been performed on biotite, alkali feldspar, carbonate, apatite and most accessory phases. Biotites are unzoned and uniform in composition, with high FeO (av. 16.2 weight %) and $Al_2O_3$ (av. 15.3 weight %), and moderate $TiO_2$ (av. 2.0 weight %), F and BaO (av. 0.5% and 0.18%, respectively). On a $TiO_2$ versus $Al_2O_3$ variation diagram (Fig. 4), 325D biotites plot just at the edge of the field for micas from minettes and Roman Province lavas, and overlap slightly with aluminous micas from the Leucite Hills, Wyoming. They differ from the Leucite Hills mica, however, in having much higher Fe-content.

Apatites occur as large (>2 mm) phenocrysts, containing abundant calcite and/or monazite inclusions, and as microphenocrysts, with very few inclusions. All are fluorapatites and most show a slight but consistent SrO zoning from 0.3% SrO cores to higher 0.5% rims.

## AGE OF DYKES

Published ages for CIF alkaline rocks are relatively limited; a possible upper constraint is the U-Pb zircon age of 1850 +30/-10 Ma obtained from a quartz syenite body thought to be related to CIF magmatism (Tella et al., 1985). A lower limit of 1753 +3/-2 Ma was established from a U-Pb zircon emplacement age of a crosscutting pluton (Loveridge et al., 1987). Roddick and Miller (1994) obtained a $^{40}Ar/^{39}Ar$ age of 1825 ± 12 Ma from hornblende separated from a plutonic suite genetically linked to the CIF in the Dubawnt Lake area. Apatite containing monazite inclusions has been separated from the Akluilâk dyke. Using the Pb-Pb isochron method, the apatite separate yielded an age of 1832±28 Ma with MSWD = 0.0101 (J.C. Roddick, unpublished data). The primary contribution of this paper is the confirmation that the Akluilâk dyke is related to CIF magmatism.

## COMPARISON TO WESTERN AUSTRALIAN DIAMOND PROVINCE

In view of this unusual diamond occurrence, the tectonic setting of the CIF invites comparison of the Churchill Province to the diamondiferous magmatism in the tectonically similar Halls Creek Mobile Zone of Western Australia. From a crustal-scale perspective, the ultrapotassic rocks in both regions are contained in regional structural corridors.

The Halls Creek Mobile Zone is a NE-striking tectonic corridor dominated by strike-slip motion, and borders the eastern margin of the inferred Archean

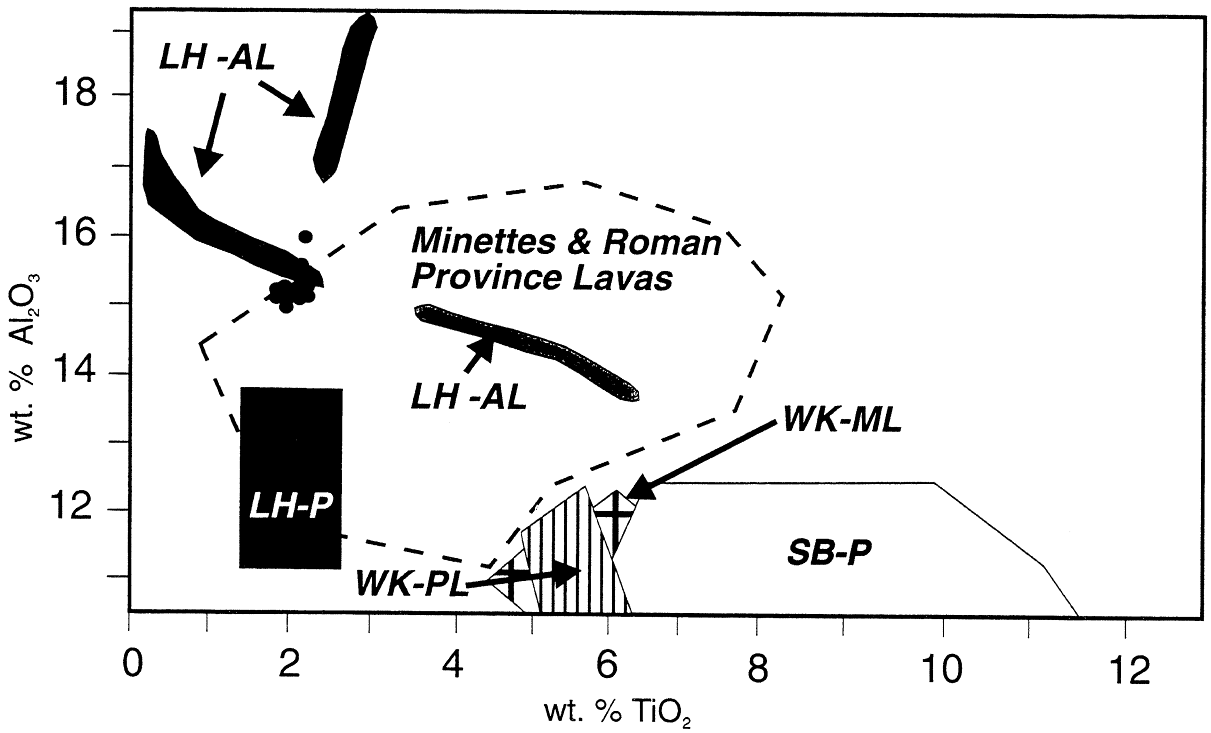


***Figure 4.*** *Variation of $Al_2O_3$ versus $TiO_2$ for micas in lamproites compared to those from minettes and Roman Province lavas (modified from Mitchell and Bergman, 1991). Analysis of biotites from the Akluilâk dyke are plotted as solid circles [SB-P = Smoky Butte phlogopite; WK-PL = West Kimberley phlogopite lamproites; WK-ML = West Kimberley madupitic lamproites; LH-Al = Leucite Hills aluminous micas; LH-P = Leucite Hills phenocrysts].*

Kimberley Block. This structural zone underwent Proterozoic thermotectonism at ca 1920 and 1850 Ma (Page and Williams, 1988), the latter orogeny associated with indentation of the Archean Kimberley Block into northwestern Australia (White and Muir, 1989). The Argyle olivine lamproite diatreme, dated at 1178 ±47 Ma (Jacques et al., 1989a; Boxer et al., 1989) postdates the 1850 Ma orogeny by almost 700 Ma and is associated with Mesoproterozoic reactivation of major fault systems within the mobile zone.

In the central Churchill Province, a series of coalesced strike-slip basins having a combined east to northeast strike length greater than 400 km are filled with CIF volcanic rocks and interbedded continental sediments (Blake, 1980; LeCheminant et al., 1979a,b; Tella et al., 1981; LeCheminant et al., 1981). These strike-slip basins formed as part of a crustal-scale fault array that transects the Churchill Province; the faults are related to 2.0 to 1.8 Ga indentation and accretion of marginal microcontinents to the Archean Keewatin hinterland (Hoffman, 1990). Ultrapotassic melts generated in metasomatized lithospheric mantle migrated through and onto the crust via the translithospheric fault zones (Peterson and Rainbird, 1990; Peterson et al., 1994). Intracratonic ultrapotassic volcanism began at ca 1840 to 1830 Ma (Roddick and Miller, 1994) and coincides with the transition at ca 1.83 Ga from convergent margin to intracratonic tectonics in the juvenile terrane of the Trans-Hudson orogen (Stern et al., 1993; Stern and Lucas, 1994).

## DISCUSSION

Currently, kimberlite and lamproite are widely recognized as potential host rocks for diamonds. Generation of a lamprophyric magma is suggested to be limited to depths above about 125 km (e.g. Mitchell and Bergman, 1991), thus above the diamond-graphite univariant curve. There are no properties apparent in either the diamonds or the Akluilâk lamprophyre dyke (we use the term lamprophyre, as recommended by Mitchell (1994), to be a mafic porphyritic igneous rock lacking felsic phenocrysts) to suggest that the primary magma was not generated below the critical diamond stability level. Peterson et al. (1994), in a discussion of the CIF lamprophyres, noted their similarities to minettes and to lamproites. They suggested that dyke sample K89/8, similar in some respects to 325D, may have crystallized from magmas parental to magmas feeding stratigraphically higher, more lamproitic and more felsic flows. Of the available analyses, K89/8 and 325D are, so far, unique in having very high P, high primary carbonate content, and high REE.

The distinctive mineral assemblage in the olivine-phlogopite volcaniclastic rocks and dykes that comprise the Argyle pipe identifies them as lamproites, as does the heavy mineral suite, dominated by chrome spinel and diamond with rare chrome pyroxenes and garnet. Octahedra, dodecahedra, and macle diamond morphologies are recognized in Argyle diamonds, and these possess a colour range dominated by brown, even though yellow, white, and grey hues are present. As well, these diamonds are strongly frosted and rounded, and resorbed dodecahedra are common (Jacques et al. 1989b). The absence of a suite of lamproitic or kimberlitic indicator minerals in the Akluilâk dyke may, in part, be a consequence of the limited sample volume examined to date. Nevertheless, it accentuates the uniqueness of this discovery and the importance of using diamond as a pathfinder mineral within boundaries defined by the distribution of ultrapotassic dykes and volcanic rocks in the central portion of the western Churchill Province (Fig. 1; Peterson and LeCheminant, 1996, this volume).

## ACKNOWLEDGMENTS

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# Lamproite dykes of southeast Baffin Island

**D.D. Hogarth and T. D. Peterson**

*Hogarth, D.D. and Peterson, T.D., 1996: Lamproite dykes of southeast Baffin Island; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 109-110.*

## INTRODUCTION

A broad belt of ultrapotassic rocks, ranging in age from 1.84 Ga to 1 Ma, extends southwest to northeast across the interior Archean terranes of Laurentia from Wyoming to Greenland (Peterson et al., 1994). The eastern end of this ultrapotassic "superprovince" lies in western Greenland, where the 1.24 Ga Sisimiut lamproite dykes are located (Scott, 1981). Until recently, a gap in this belt existed in the area of the northeast Churchill Province. Dykes bearing remarkable similarity to the Sisimiut lamproites, and also to the Christopher Island Formation of the central Churchill Province (Peterson and LeCheminant, 1996) have now been identified near Iqaluit, Baffin Island.

## DESCRIPTION OF THE DYKES

Narrow (≤2m) dykes of phlogopite- leucite-olivine lamproite intrude Archean anorthosites and mafic orthogneisses along the shores of Napolean Bay, Baffin Island (62°53' N, 65°21' W). One dyke was traced for over 2 km inland. The dykes have prominent chilled borders and multiple internal chill zones, and were emplaced at shallow crustal levels. Unaltered leucite phenocrysts are abundant; many of these underwent subsolidus decomposition to kalsilite plus sanidine (Fig. 1).

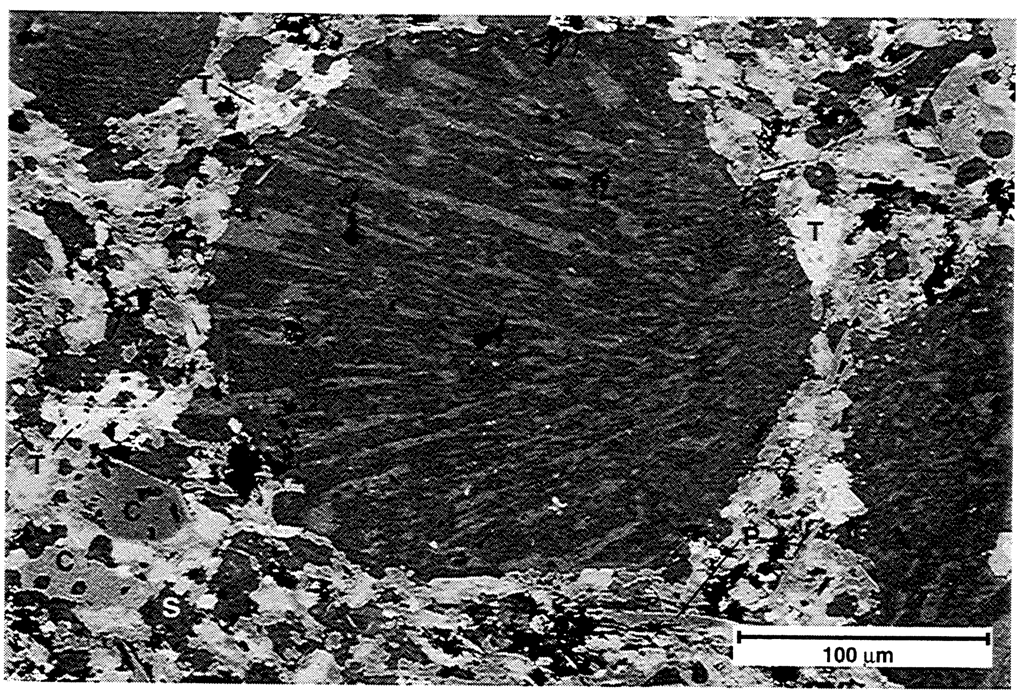


*Figure 1. Backscattered electron image of a decomposed leucite phenocryst; the image has been digitally enhanced to increase the contrast between kalsilite (light grey) and sanidine (dark grey). Surrounding groundmass phases include titanite (T), clinopyroxene (C), phlogopite (P), and sanidine (S).*

The rocks are remarkably unaltered for their age; a K-Ar date (phlogopite) of 1.24 Ga has been obtained (J.C. Roddick, pers. comm. 1995), and confirmed by Rb-Sr isochrons on whole rock and phlogopite separates. These are the first true lamproites to be described in Canada (Hogarth and Peterson, 1995).

Phenocryst phases are olivine ($Fo_{87\text{-}90}$), phlogopite, leucite, and diopside. Carbonate, sanidine, titanite, and amphibole are prominent in the groundmass and in segregations. Olivine is coated and partly replaced by phlogopite. Diagnostic lamproite minerals recognized in the dykes include K-Ti richterite, perovskite, shcherbakovite, and priderite. The potassian sulphide djerfisherite, previously unreported in lamproites, is also present. Phlogopite phenocryst composition trends follow those of high-Ti lamproites, and groundmass micas are similar to those in Group II kimberlites (Mitchell and Bergman, 1991). Whole-rocks have high molecular K/Al (0.9), wt% $K_2O/Na_2O$ (8), and Nb (>100 ppm).

The Napolean Bay lamproites are precisely the same age as the Sisimiut dykes, and share an identical, diagnostic accessory mineral suite (R.H. Mitchell, pers. comm. 1995) as well as similar geochemistry (e.g., high $CO_2$ contents). However, their Nd isotopic composition is very different. At 1.24 Ga, average $\varepsilon_{nd} = -8$, whereas Sisimiut has average $\varepsilon_{nd} = -27$ (Nelson, 1989). The Nd isotopic composition and average $\varepsilon_{Sr}$ (+42) are identical to those of the Christopher Island Formation. The Napolean Bay dykes therefore present an enigmatic link between Greenland and the Churchill Province, in which major/trace element geochemistry, and isotope geochemistry, are decoupled in the source regions of ultrapotassic magmas.

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# Sweet Grass minettes, Alberta

**B.A. Kjarsgaard and W. J. Davis**

*Kjarsgaard, B.A. and Davis, W.J., 1996: Sweet Grass minettes, Alberta; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 111-114.*

## INTRODUCTION

Potassic rocks, known variously as the Milk River or Sweet Grass intrusives, were discovered in the Milk River area of southern Alberta and northern Montana by G.M. Dawson in 1873 and 1874 during his assessment of the geology and resources of the 49th Parallel (British North American Boundary Commission, Montreal, 1875; cited in Dawson, 1884). The small intrusive masses and dykes in Alberta were described by Dawson (1884) as "mica-traps" and considered to be subordinate intrusions of the Sweet Grass Hills igneous complex to the south. Subsequent mapping of the geology of southern Alberta in the 1920s by Williams and Dyer (1930) and in the 1930s by Russell and Landes (1940) led to the discovery of additional outcrops in the Milk River area. All the sites were characterized as minette intrusions, with the exception of 'porphyritic andesite' at the westernmost locality. More recently, Currie (1976) described a sample from Black Butte. The mineral assemblage observed (biotite + augite + potassium feldspar + olivine) and the lamprophyric texture noted (Currie, 1976), is consistent with that of a minette. Recently, these potassic rocks have been reinvestigated as part of the 1992-95 Canada-Alberta Agreement on Mineral Development (Kjarsgaard, 1994; Kjarsgaard and Davis, 1994; Davis and Kjarsgaard, 1994; Davis 1994; Ross et al., 1994; Kjarsgaard, in press).

## GEOLOGY AND PETROLOGY

Potassic rocks outcrop in six areas in southern Alberta (Fig. 1). Most of the occurrences are thin (1 - 3 m wide) dykes, although much larger, irregular shaped bodies and a vent complex have been described by Kjarsgaard (1994). Recent high-resolution geophysical surveys are strongly suggestive of the presence of related dykes intruding the Cretaceous cover (Ross et al., 1994). The region of potassic magmatism may therefore extend northward to the Lethbridge area. At present, however, the verified northern limit of the Montana Alkaline Province (69 - 27 Ma; Marvin et al., 1980) is the outcrops in the vicinity of the Milk River, as shown in Figure 1.

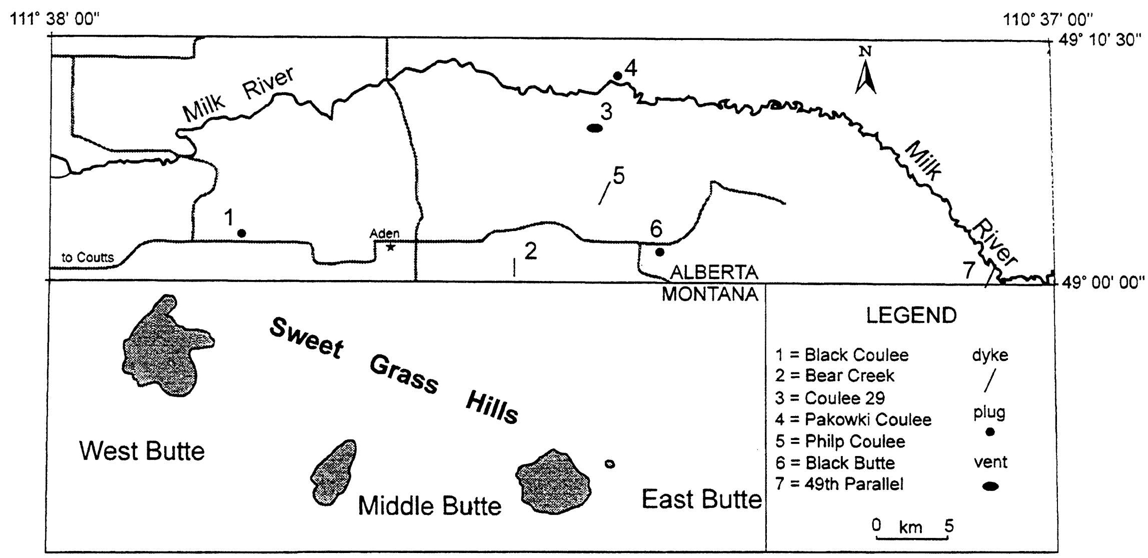


***Figure 1.*** *Location map of the Sweet Grass intrusives in the Milk River area of southern Alberta in relation to the Sweet Grass igneous complex (light grey hatching) in Montana. Thin lines denote minette dykes; solid circles are irregular shaped plugs; solid ovoid is a minette vent complex (adapted from Davis and Kjarsgaard, 1994).*

A Rb-Sr phlogopite/whole-rock isochron age of 50.3 ± 0.5 Ma reported for an olivine minette dyke from the Coulee 29 locality (Davis and Kjarsgaard, 1994) is identical, within error, to a previously published K-Ar age of 49.2 ± 2.5 Ma (recalculated) for a minette from Black Butte (Baadsgaard et al., 1961). These data confirm an Eocene age for the Sweet Grass minettes in southern Alberta and are within error of the 50 to 54 Ma age determined by Marvin et al. (1980) for samples from the Sweet Grass Hills in Montana. This magmatism is synchronous with the peak of magmatic activity occurring in the Montana Alkaline province (Missouri Breaks, Eagle Buttes, Sweet Grass Hills, and the Bearpaw and Highwood Mountains) at 49 to 55 Ma (Davis and Kjarsgaard, 1994).

Petrographic studies presented by Kjarsgaard (1994; in press) suggest these rocks are best classified as minettes, confirming the observations of earlier workers. Four petrographic groups of minette samples have been identified, based on modal petrography. These are:

1) alkali olivine minette, consisting of olivine + phlogopite + diopside+ Cr-spinel phenocrysts in a groundmass of mica (phlogopite-biotite$_{ss}$) + salite + sanidine + titanomagnetite + apatite + calcite ± analcime;
2) phlogopite minette, consisting of phlogopite (cumulate-rich) + diopside + olivine + Cr-spinel phenocrysts in a groundmass of mica (phlogopite-biotite$_{ss}$) + salite + sanidine + titanomagnetite + apatite + calcite ± analcime;
3) alkali minette, consisting of phlogopite + augite + Cr-spinel (± rare olivine) phenocrysts in a groundmass of mica (phlogopite-biotite$_{ss}$) + salite + sanidine + titanomagnetite + apatite + calcite ± analcime, and;
4) peralkaline minette, consisting of rare phlogopite + salite phenocrysts in a groundmass dominated by phlogopite-biotite$_{ss}$ and sanidine with calcite + apatite + NaK-Ti amphibole + aegirine augite + titanomagnetite ± analcime.

In general, mafic minerals in the olivine-free minettes have lower magnesium numbers. In all rock types, phlogopite and clinopyroxene phenocrysts are strongly zoned, showing normal, reverse and oscillatory zoning. Coarse grained xenoliths (clinopyroxenite, phlogopite clinopyroxenite, and glimmerite) within minette have varying modal proportions of the mineral assemblage mica (phlogopite or biotite) + clinopyroxene (diopside or salite) + apatite ± K-feldspar ± dolomite. This assemblage is the same as that observed in the host minettes (with the exception that calcite is present in the minettes and dolomite in the xenoliths), and the xenoliths are considered cognate. Minerals in the cognate xenoliths show minor or no compositional zoning, in contrast to those in the minettes.

Whole-rock geochemical and mineral chemistry studies (mica) also confirm the petrographic classification of these rocks as minettes (Kjarsgaard, 1994). Phlogopite-biotite$_{ss}$ mica, a ubiquitous phase in the Sweet Grass minettes and associated cumulate xenoliths, exhibits a wide compositional range. Microprobe analyses of zoned mica phenocrysts (core to rim) and homogeneous groundmass crystals from minette, plus coarse-grained mica from cognate xenoliths are illustrated in Figure 2. The salient points to be observed on the diagrams are: 1) compositions of mica from cognate xenoliths overlap those from the core and middle of zoned phenocrysts, although $Al_2O_3$ contents are slightly higher in the phlogopite clinopyroxenite and clinopyroxenite xenoliths; 2) zoning trends show increasing $FeO_{total}$ and $TiO_2$ at constant to decreasing $Al_2O_3$; and; 3) most, but not all compositions, fall in the known range for micas from minettes (c.f. Mitchell and Bergman, 1991). Importantly, a subset of samples have alumina levels which are similar to those observed in some lamproites (Fig. 2A, B). However, it should be noted that combined $FeO_{total}$ - $TiO_2$ -$Al_2O_3$ relations of micas from lamproites and the Sweet Grass minettes are not similar, thus allowing these two rock types to be discriminated. The trend of decreasing $Al_2O_3$ in the Sweet Grass micas is suggested here to be typical of alkaline and peralkaline minettes; it is also observed, but not well developed, in other highly alkaline minettes (e.g. Bohemian alkali minettes; Mitchell and Bergman, 1991).

## DISCUSSION

Classification of the Sweet Grass intrusions as minette (more specifically, alkali minette) indicates these rocks could potentially be diamond-bearing, as diamonds have been recovered from an alkali minette dyke in the Gibson Lake area, NWT (MacRae et al., 1996). In considering the diamond potential of these rocks, it is important to note that the Sweet Grass minettes intrude the Archean Medicine Hat Block (U-Pb zircon ages range from 2612 - 3278 Ga; Villeneuve et al., 1993). However, it has been suggested by Davis (1994) that the lower crust in the Medicine Hat Block was thermally reworked in the Paleoproterozoic, synchronous with a metasomatic event in the mantle. This mantle metasomatic event may have affected diamond stability and preservation which, in part, could explain why diamonds have not been recovered from these rocks.

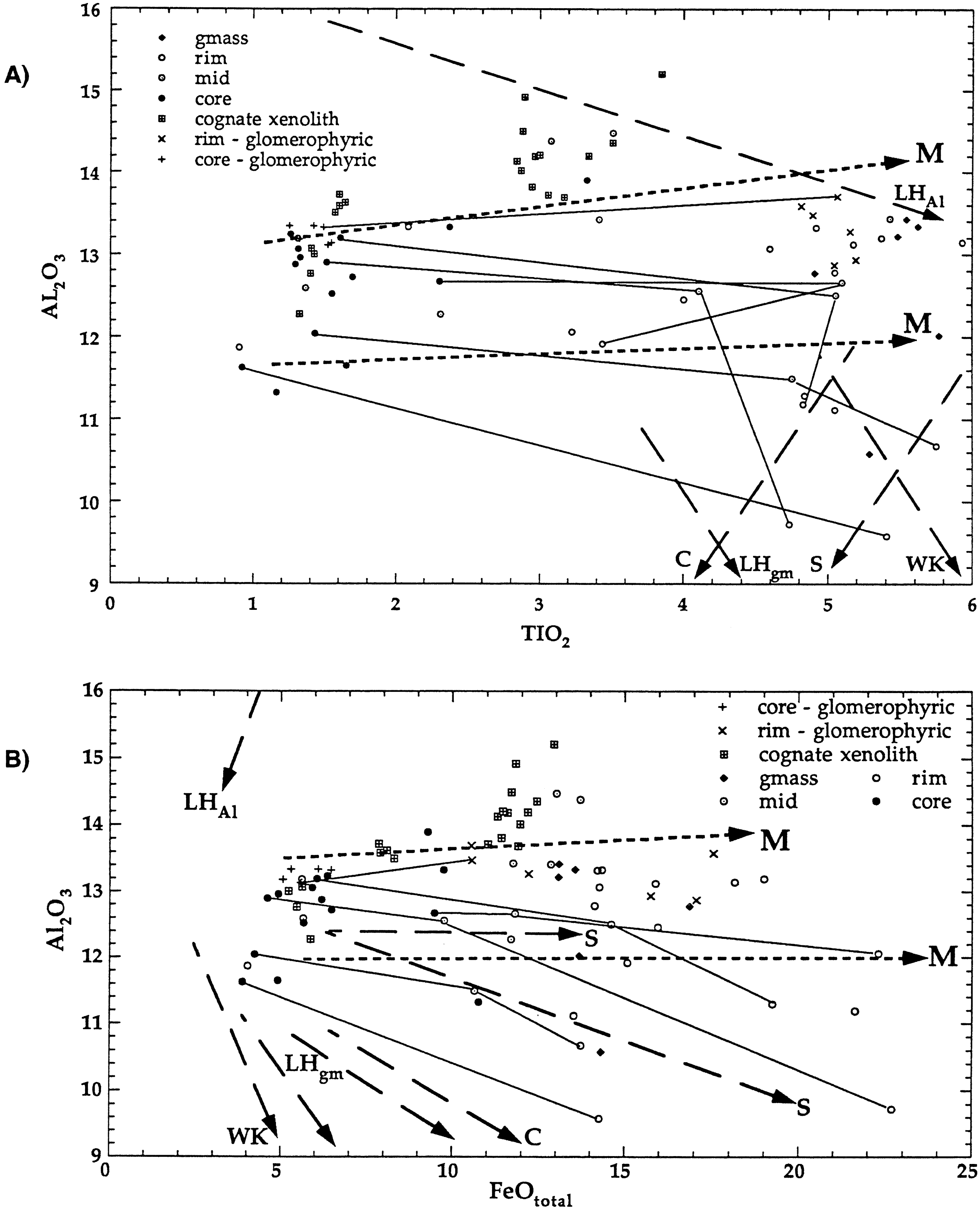


*Figure 2. Microprobe analysis of micas from the Sweet Grass minettes: **A)** $Al_2O_3$ versus $TiO_2$ ; **B)** $Al_2O_3$ versus $FeO_{total}$ - both diagrams are wt%. Solid lines join analyses from zoned phenocrysts. Minette differentiation trend (Mitchell, 1986) indicated by short dashed lines labelled M. Representative lamproite differentiation trends (Mitchell and Bergman, 1991) indicated by long dashed lines labelled as follows: WK=West Kimberley; C=Chelima; S=Sisimiut; $LH_{gm}$=Leucite Hills groundmass; $LH_{Al}$=Leucite Hills aluminous biotite high pressure phenocrysts.*

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* Contribution to the Canada-Alberta Agreement on Mineral Development (1992-1995), a subsidiary agreement under the Canada-Alberta Economic and Regional Development Agreement.

# Superior Province lamprophyres

**R.A. Stern**

*Stern, R.A., 1996: Superior Province lamprophyres; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 115-116.*

Archean and post-Archean lamprophyre dykes and related intrusions are a volumetrically minor and relatively poorly understood component of the Superior Province. The Archean lamprophyre dykes have U-Pb and Sm-Nd crystallization ages between 2.67 and 2.70 Ga (e.g. Barrie, 1990; Stern and Hanson, 1992), and were emplaced diachronously across the Superior Province. The dykes that have been studied in greatest detail intruded the low-grade greenstone-granite terranes (e.g. Abitibi, Wabigoon subprovinces), but less well known lamprophyre dykes are also present within high-grade plutonic terranes (e.g. Minto Block). The isotopic ages of the dykes and their field relationships demonstrate that they were emplaced during the waning stages of magmatic activity and transpressional accretionary tectonics that led to the present east-west orientation of lithostructural elements of the Superior Province (Wyman and Kerrich, 1989). Some of the lamprophyres, notably those in the Abitibi Subprovince, are spatially associated with fault graben and are temporally associated with alkalic volcanism and molasse sedimentation (e.g. Timiskaming Group). Most dykes are spatially associated, and some are genetically related, with late- to post-kinematic monzonite-syenite-(carbonatite) and diorite-granodiorite-(pyroxenite)-(gabbro) plutonic bodies (Stern and Hanson, 1992).

The Archean lamprophyre dykes are compositionally and mineralogically most similar to Rock's calc-alkaline lamprophyres (Rock, 1984), occurring as minettes, kersantites, and spessartites, although ultramafic lamprophyres have also been reported (e.g. aillikite of Barrie, 1990). The lamprophyres have broadly 'arc' trace element signatures, including low $TiO_2$, Nb, and Ta abundances, high K, Th, Ba, Rb, Sr, and LREE contents, and correspondinly high LREE/HFSE, LILE/LREE ratios. An important characteristic of the Archean lamprophyres is that they are isotopically juvenile. For example, the dykes have initial epsilon Nd values of +1 to +3, implying derivation from long-term light REE depleted mantle, with little or no contribution from significantly older crustal sources. These data, in combination with the field and temporal relationships, suggest an origin for the Archean lamprophyres by melting of subduction-metasomatized depleted arc mantle in an overall subduction-related, transcurrent to transpressional regime (Wyman and Kerrich, 1989; McCall et al., 1990; Stern and Hanson, 1992).

Post-Archean lamprophyre dykes are also present in the Superior Province, but few in-depth studies exist. In contrast with the Archean lamprophyres, which are dominantly calc-alkaline, the post-Archean ones are typically ultramafic or alkaline. Some examples include the 1650 Ma (Rb-Sr) ultramafic lamprophyre dykes (McKellar Harbour dykes) near Marathon, Ont. (Platt et al., 1983), the ultramafic lamprophyre dykes from Opapimiskan Lake area, Sachigo Subprovince (Wyman and Kerrich, 1989), the Wawa area alkaline lamprophyre dykes (monchiquites; Wyman and Kerrich, 1989), and the 128 Ma ultramafic lamprophyre dykes and sills at Coral Rapids, Ontario (Edgar et al., 1994). The available geochemical data (e.g. Wyman and Kerrich, 1989; Edgar et al., 1994) show that the lamprophyres are geochemically distinct from the Archean ones in lacking depletion in HFSEs relative to the REEs and HFSEs. Their geochemistry is indicative of their intraplate setting, in contrast to the plate-margin setting inferred for the Archean dykes.

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# Alnöites and related rocks, Monteregian Hills alkaline igneous province, Québec

**J. H. Bédard and A.N. LeCheminant**

*Bédard, J.H. and LeCheminant, A.N., 1996: Alnöites and related rocks, Monteregian Hills alkaline igneous province, Québec; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 117-121.*

## INTRODUCTION

The Monteregian Hills alkaline province is composed of an WNW-trending chain of plutons and associated dykes and sills. This chain of intrusions crosscuts the Appalachian structural grain (Fig. 1), and it has been proposed (Philpotts, 1978; Bédard, 1985) that the trend reflects the presence of the Ottawa-Bonnechère graben in the Grenville basement, which is buried beneath Paleozoic platform sediments and northwesterly overthrusted Appalachian allochthons. The Cretaceous Monteregian magmatism is a young event in a succession of Mesozoic alkaline igneous events in Québec and New England associated with different stages in the opening of the Atlantic Ocean (McHone and Butler, 1984; Bédard, 1985).

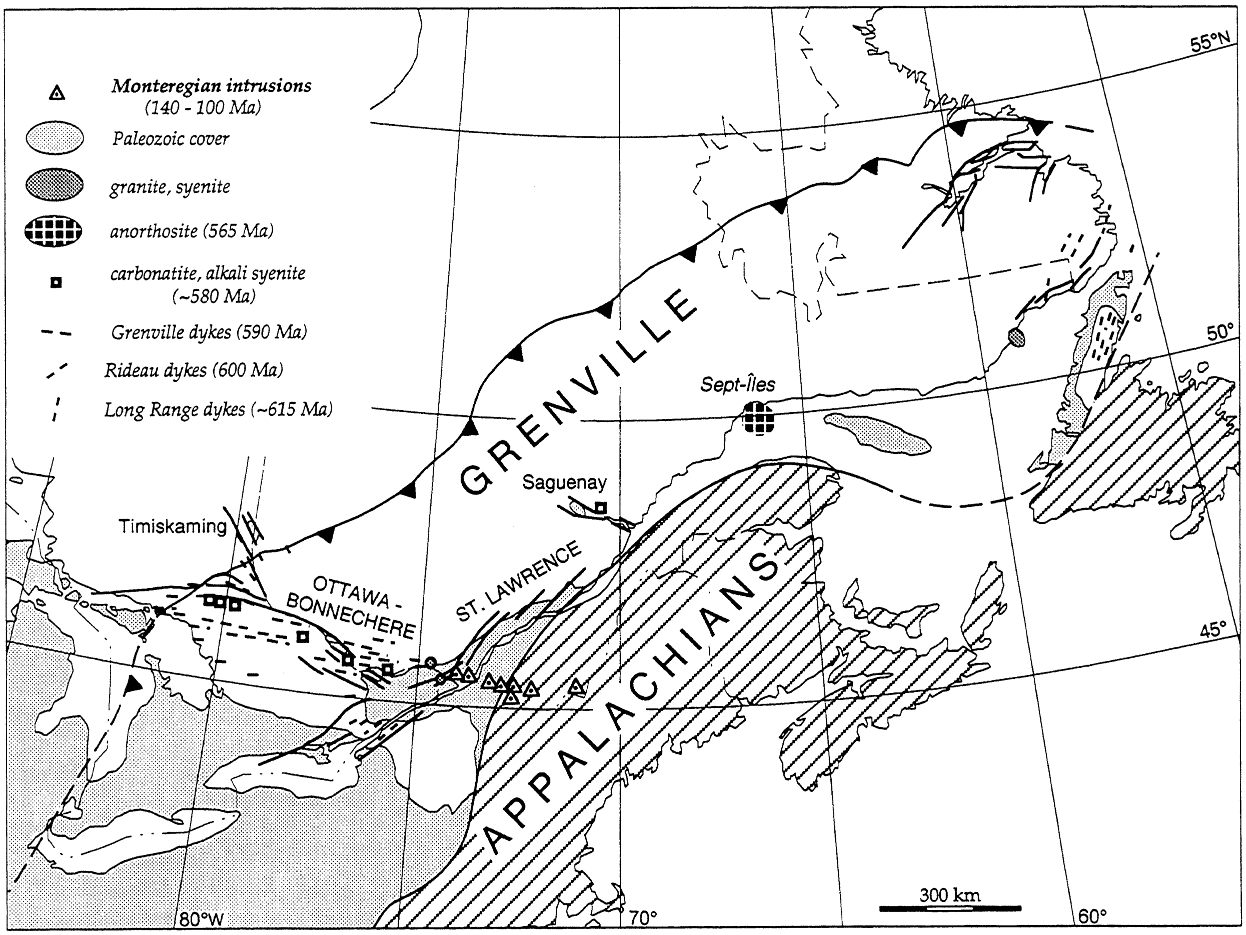


*__Figure 1.__ Monteregian intrusions and their relationship to Neoproterozoic rifts and 0.6 Ga igneous and sedimentary rocks in the Grenville Province (from Davidson, 1995). The St. Lawrence rift, related branches, and local alkaline igneous complexes originated during opening of the Iapetus Ocean (rift faults are shown as solid lines, diabase dyke swarms are represented by dashed lines). Monteregian igneous activity signalled Cretaceous renewal of alkaline magmatism in the rift (see also Currie, 1996), and was accompanied by reactivation of pre-existing rift faults during formation of the North Atlantic basin.*

The westernmost Monteregian intrusions comprise a heterogenous assemblage characterized by undersaturated rocks (Gold et al., 1986), including the Oka carbonatite-ijolite complex, alnöitic dykes and diatremes, and dykes and sills of monchiquite and camptonite. Breccia dykes and diatremes indicative of explosive igneous activity are restricted to the region from Montreal to Oka (Gold et al., 1986). In the central Monteregians, the plutonic complexes consist of alkali pyroxenite, yamaskite, olivine gabbro to essexite, feldspathoidal diorite to monzonite, syenite and foyaïte (Philpotts, 1974; Eby, 1987). Hypersthene and quartz-bearing gabbros and syenites are subordinate, and are crustally contaminated variants of the main, undersaturated lineages (Gold, et al., 1986). Silica-oversaturated rocks become dominant in the eastern intrusive complexes, reflecting greater amounts of crustal contamination (Eby, 1985a).

Ultramafic lamprophyres, mainly alnöites, are restricted to the western end of the Monteregian province. Other lamprophyres, primarily monchiquite and camptonite, occur throughout the province (Hodgson, 1968; Eby, 1985b), and there is no obvious relation between geographic location and relative proportion of these two rock types. The petrology and mineralogy of Monteregian lamprophyres has been considered in detail by Bédard (1988, 1994) and Bédard et al. (1988).

## ALNÖITES AND RELATED ROCKS

Alnöites and related rocks represent the deepest level of generation of Monteregian magmas (Raeside and Helmstaedt, 1982). Intrusions near Montreal include dykes and diatreme breccias, such as Île Bizard and Île Cadieux (Raeside and Helmstaedt, 1982; Harnois et al., 1990; Harnois and Mineau, 1991). Alnöites contain phenocrysts of phlogopite + clinopyroxene ±olivine, with Al-Mg rich Ti-magnetite microphenocrysts and mixtures of glass, phlogopite, melilite, monticellite, perovskite, melanite garnet, apatite, chlorite, carbonate, and serpentine in the groundmass (Bédard, 1994). Breccias carry a wide variety of xenoliths and xenocrysts (Gold et al., 1986), some of which are of mantle origin. Pyroxene phenocrysts are strongly zoned and can contain high pressure cores, a consequence of polybaric crystallization and differentiation during ascent from the upper mantle (Bédard et al., 1988).

The Île Bizard diatreme breccia, the only reported diamond-bearing Monteregian intrusion, shares many characteristics with melilite-bearing alnöites, but the breccias contain groundmass calcite and do not carry melilite. For this reason, the intrusion was termed an aillikite by Eby (1985b) and Gold et al. (1986). Other authors reported the Île Bizard diatreme has affinities to both kimberlites and alnöites (Raeside and Helmstaedt, 1982), however the merits of the kimberlite assignment were vigorously debated by Mitchell (1983) and Raeside and Helmstaedt (1983).

Chemically, alnöites are strongly undersaturated potassic rocks with $K_2O>Na_2O$ (wt%), and high normative nepheline (average (av.) 8%), leucite (av. 10%), and larnite (av. 11%). The Île Bizard aillikites are also potassic, strongly undersaturated rocks, but have lower total alkali contents and higher MgO, $SiO_2$, $H_2O$ and $CO_2$ than do alnöites (Table 1). All these rocks plot in the ultramafic lamprophyre field on the MgO/CaO vs $SiO_2/Al_2O_3$ diagram of Rock (1991, his Fig. 5.5). Alnöites are generally poorer in $SiO_2$ and $TiO_2$, and richer in CaO, MgO, $P_2O_5$ and $K_2O$ than camptonites and monchiquites. Chemical variations within the alnöite suite appear to be due to a combination of accumulation and fractionation (Bowen, 1922), superimposed on primary source region heterogeneity. Analyses plotted by Bédard (1994) on the Ol-Di-Ne pseudo-liquidus diagram of Sack et al. (1987), suggest that evolution of the alnöites and Île Bizard aillikites was principally controlled by olivine fractionation, and that these are distinct, though closely related, suites.

## DIAMONDS AT ÎLE BIZARD ?

Ten small diamonds were recovered in 1968 from 22 $m^3$ of rock quarried from the Île Bizard diatreme (Brummer, 1978; Raeside and Helmstaedt, 1982). Five of these diamonds (see Fig. 12 in Brummer, 1978) were donated to the National Mineral Collection by the Canadian Rock Company, Ltd, a subsidiary of Anglo-American, and are stored at the GSC in Ottawa. The diamonds were recovered from a heavy mineral separate processed in South Africa, and although contamination has been suspected (Brummer, 1978) the small size and poor quality of the diamonds is consistent with a non-economic source in the Île Bizard breccia.

A wide range of ultramafic xenoliths, both mantle cumulates and residues of partial melting from the upper mantle, have been recovered from Île Bizard breccias (Marchand, 1970; Raeside and Helmstaedt, 1982; Harnois et al., 1990). Phlogopite and amphibole-bearing nodules, perhaps derived from metasomatically-altered mantle, have also been reported (Marchand, 1970; Raeside and Helmstaedt, 1982; Bédard et al., 1988). Pressure and temperature estimates for garnet lherzolite and garnet websterite xenoliths are outside the diamond stability field, and range up to almost 1000°C at

**Table 1.** Normalized average analyses (Bédard, 1994)

| weight % | alnöite | (n) | Île Bizard | (n) |
|---|---|---|---|---|
| $SiO_2$ | 34.6 | (27) | 35.7 | (12) |
| $TiO_2$ | 2.6 | (27) | 2.4 | (12) |
| $Al_2O_3$ | 9.4 | (27) | 6.8 | (12) |
| $Fe_2O_3$ | 6.1 | (27) | 6.1 | (12) |
| FeO | 6.2 | (27) | 6.1 | (12) |
| MnO | 0.25 | (26) | 0.21 | (5) |
| MgO | 16.6 | (27) | 22.2 | (12) |
| CaO | 18.8 | (27) | 17.8 | (12) |
| $Na_2O$ | 1.8 | (27) | 0.25 | (12) |
| $K_2O$ | 2.2 | (27) | 1.2 | (12) |
| $P_2O_5$ | 1.6 | (27) | 1.2 | (8) |
| $H_2O$ | 2.6 | (20) | 8.4 | (9) |
| $CO_2$ | 4.0 | (19) | 6.6 | (8) |
| **ppm** | **alnöite** | **(n)** | **Île Bizard** | **(n)** |
| S | 2521 | (4) | - | |
| Cr | 723 | (17) | 782 | (5) |
| Ni | 359 | (11) | 429 | (5) |
| Cu | 89 | (6) | - | |
| Zn | 106 | (11) | - | |
| Ba | 2371 | (14) | 1209 | (1) |
| Rb | 53 | (12) | - | |
| Sr | 1560 | (12) | - | |
| Y | 38 | (10) | - | |
| Zr | 424 | (10) | 371 | (1) |
| Nb | 215 | (11) | 135 | (1) |
| Co | 57 | (14) | - | |
| V | 200 | (8) | - | |
| Sc | 23.4 | (14) | 231.2 | (5) |
| La | 235 | (14) | 195 | (5) |
| Ce | 417 | (17) | 97 | (5) |
| Nd | 180 | (14) | 44.4 | (5) |
| Sm | 23.8 | (14) | 22.2 | (5) |
| Eu | 6.54 | (14) | 11.4 | (3) |
| Gb | - | | 6.6 | (4) |
| Tb | 1.54 | (13) | 5.8 | (1) |
| Ho | 1.67 | (3) | 2.2 | (1) |
| Tm | 0.88 | (2) | 2.62 | (5) |
| Yb | 1.96 | (14) | 4.7 | (4) |
| Lu | 0.26 | (14) | - | |
| Hf | 6.0 | (13) | - | |
| Ta | 13.1 | (13) | - | |
| Th | 28.9 | (13) | 11.2 | (1) |
| U | 1.9 | (3) | 2.7 | (1) |

pressures equivalent to depths of about 100 km (Raeside and Helmstaedt, 1982, Schulze, 1996).

Baddeleyite ($ZrO_2$) xenocrysts have been recovered from Île Bizard. The unique, octahedral crystals were enclosed within titanomagnetite megacrysts, and are interpreted to have crystallized in the mantle as tetragonal $ZrO_2$, at temperatures above 1170°C (Heaman and LeCheminant, 1993). The U-Pb systematics of the baddeleyite xenocrysts are unusual and may record the timing of metasomatic events in the mantle source regions for the Monteregion intrusions. The rare Zr-rich garnet, kimzeyite, occurs as xenocryst cores in groundmass Ca-Ti garnets in alnöites from both Île Bizard and Husereau Hill, near the Oka carbonatite, and may also be useful for U-Pb geochronology (Villeneuve and LeCheminant, 1990).

## SUMMARY

Emplacement of the Cretaceous Monteregian intrusions occurred after opening of the Atlantic Ocean. The diversity of alkaline magmatism suggests a heterogeneous source, with ascent contolled by pre-existing faults of the Ottawa-Bonnechère graben. Magmatism was localized by this paleorift in the underlying Grenville Province and may be related to reactivation of deep-seated faults during continental breakup and formation of the North Atlantic basin. Alnöites and associated rocks occur in diatreme breccias near Montreal and represent the deepest level of generation known for Monteregian magmas. Diamonds reportedly recovered from the Île Bizard diatreme remain puzzling, since contamination cannot be ruled out and mantle xenoliths do not record P-T conditions in the diamond stability field.

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# Lamprophyric dykes in Labrador: Summary of occurrences and their significance to diamond exploration

**B. Ryan**

*Ryan, B., 1996: Lamprophyric dykes in Labrador: Summary of occurrences and their significance to diamond exploration; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 123-127.*

## INTRODUCTION

Lamprophyric dykes, some apparently having kimberlitic affinities, are known from several widely scattered locations along the Labrador coast, mainly from the Archean Nain craton and adjacent Proterozoic orogens (Fig. 1). There is a paucity of quantitative geochronological data for these intrusions, but geological relations indicate as many as three widely separated periods of magmatism, viz. Paleoproterozoic, Late Neoproterozoic-Early Paleozoic and Mesozoic.

The regions in which the lamprophyric dykes are known in Labrador have been recently examined for the possible existence of kimberlite intrusions. In light of the discovery of diamondiferous kimberlites in the Slave Province, Labrador is also viewed as offering suitable environments for exploration. No diamondiferous intrusions are known in Labrador at this time. This may reflect the geological reality, but could also reflect the lack of prospecting. Given the discovery of diamond-bearing intrusions in the once contiguous Greenland Shield (Larsen, 1991; Fig. 1), it is possible that Labrador could also be a region where diamondiferous intrusions are present. Recent regional surveys for critical indicator minerals of diamondiferous intrusions by private industry (cf. Danielson, 1994) and the Newfoundland and Labrador Department of Natural Resources (Ryan and McConnell, 1995) may provide the impetus for a focussed examination of Labrador from this perspective. The following summary outlines the history of investigation of these rocks and some of their salient characteristics.

## AILLIK BAY AREA

Lamprophyric dykes are most profusely distributed in the Aillik Bay region, north of Makkovik. There they are tectonically situated in the Paleoproterozoic Makkovik Province, where they have intruded reworked Archean gneisses, deformed and recrystallized supracrustal rocks of the Paleoproterozoic Aillik Group, and Paleoproterozoic granitoid rocks. The time of intrusion of these dykes is equivocal, but Foley (1989) has surmised that they represent igneous activity in the Late Precambrian-Early Paleozoic, and in the Mesozoic. These dykes were first recognized by E. H. Kranck who, during a pioneering study of the Labrador coast in 1937 and 1939, recorded hornblende-rich lamprophyres and olivine + biotite-bearing alnöitic dykes from the Aillik Bay area. He was so impressed by the carbonate-rich aspect of some of the alnöitic dykes that he proposed a new name for them - aillikite - and postulated that the carbonate was a primary magmatic phase (Kranck, 1939). Subsequently, he provided a petrographic subdivision of the lamprophyres, noting alkaline, subalkaline and ultrabasic varieties. The subalkaline varieties that he described are most certainly part of the Kokkovik dioritic intrusions, a suite of generally shallowly dipping, hornblende + plagioclase-bearing rocks having lamprophyric affinities (Ermanovics, 1993); these are much richer in silica (> 50% $SiO_2$) than the other rocks addressed here. Among the alkaline and ultrabasic lamprophyres, Kranck (1953) recorded monchiquite (a pyroxene + mica rock) and alnöite (an olivine + mica rock). He speculated that many of the ultrabasic dykes were intruded in a partly crystalline state, thus having a feature in common with kimberlites. King (1963) confirmed the lamprophyric character of many of the dykes on the Aillik Peninsula and presented further descriptions of the field and microscopic character of the intrusions. King described the monchiquite dykes as comprising biotite phenocrysts in a matrix of poikilitic to euhedral salitic pyroxene, laths of biotite, magnetite, olivine, apatite and carbonate. The alnöites, or aillikites, he described as biotite- and olivine-phenocrystic rocks having a groundmass composed of biotite, olivine, pyroxene and carbonate.

Hawkins (1976) provided a detailed account of the field and petrographic character of the dykes of the Aillik Bay area. He referred to them as kimberlites, carbonatites and lamprophyres. Hawkins followed the definition of kimberlite given by Mitchell (1970), essential to which is two generations of olivine and phlogopite

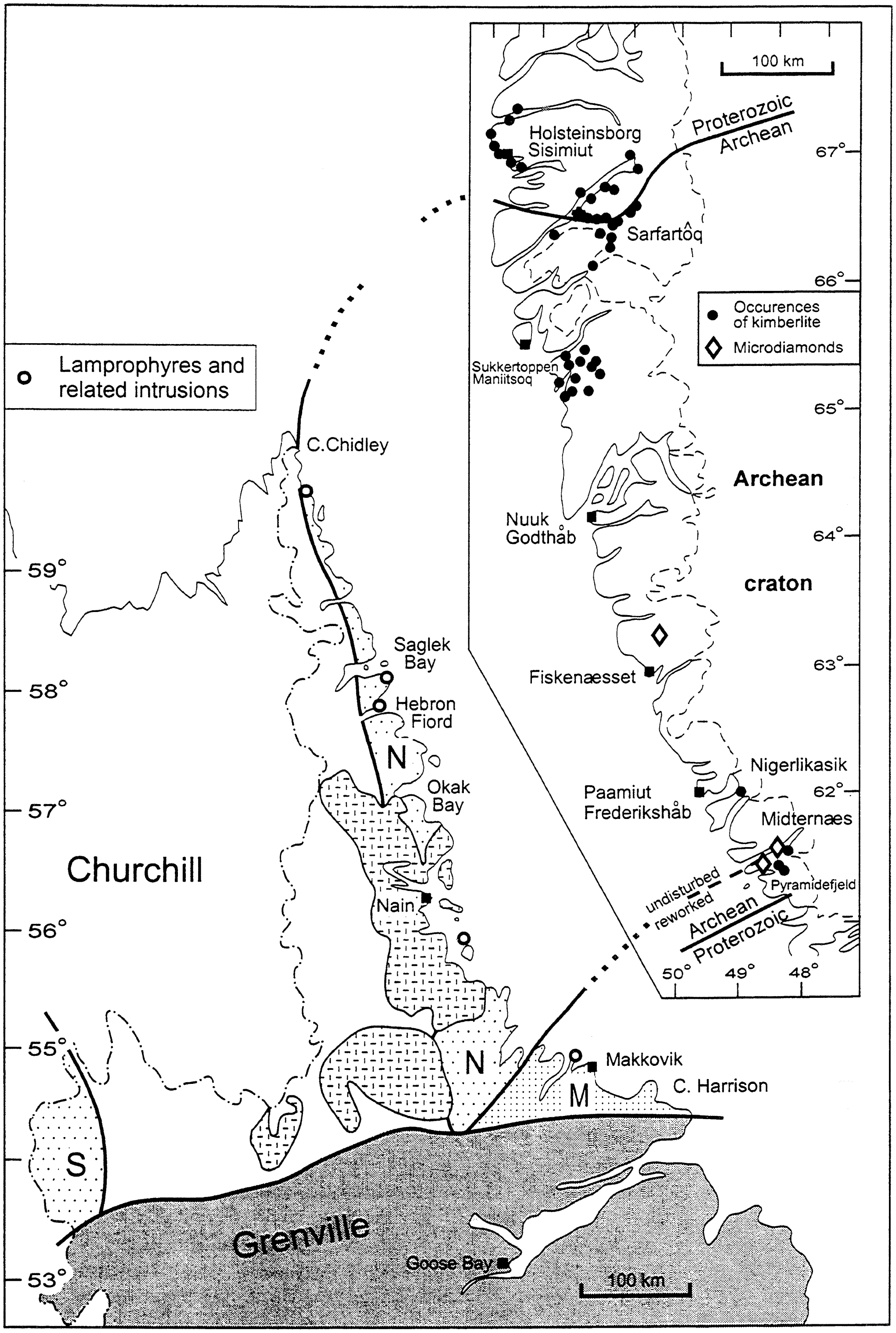


*__Figure 1.__ Side-by-side comparison of areas of known alkaline intrusions in Labrador and western Greenland. Ornamented areas of Labrador are: Archean crust of the Nain (N) and Superior (S) provinces (stippling), reworked Archean crust and Paleoproterozoic rocks of the Makkovik Province (smaller/less dense stippling), and Mesoproterozoic anorogenic intrusions (dashed pattern). Unornamented areas of Labrador are accreted orogens of Proterozoic age, the one between the Nain and Superior province containing significant amounts of reworked Archean rocks. Map of Greenland is taken from Greenland Minex News, July, 1993.*

thus encompassing some of the alnöitic rocks of previous workers. Carbonatite was the name he applied to all rocks with over 50% carbonate. He used the term alnöite for those kimberlite-like lamprophyric dykes that have pyroxene and abundant melilite. For the other lamprophyric dykes, Hawkins followed the nomenclature set out by Williams et al. (1955): minettes (orthoclase-bearing and having biotite as the dominant mafic) and monchiquites (feldspar-free rocks having alkali pyroxene or amphibole as the chief mafic).

The Aillik Bay dykes were also subjected to close scrutiny by Foley (1982). He re-evaluated the compositional spectrum of the Aillik dykes in light of contemporaneous discussions of lamprophyre terminology (Rock, 1987), and revised the nomenclature for the intrusions. Although he, like Hawkins (1976) referred to some of the intrusions as kimberlites, he applied the name sannite to the alkaline lamprophyres. He preferred sannite to minette and monchiquite (used by Hawkins) because he demonstrated that nepheline is present, the silica content is low, and the clinopyroxene has sodic (acmitic) rims. Within the sannites he recognized an olivine-rich subgroup (up to 35% olivine), but generally the sannites are characterized by Ti-rich salitic clinopyroxene and lesser olivine and poikilitic phlogopite in a matrix of pyroxene, biotite, magnetite, K-feldspar, apatite, nepheline, analcime, carbonate, rutile, and opaques; leucocratic ocellar structures are common (Foley, 1984). Foley's use of kimberlite for the carbonate-rich olivine+mica-porphyritic rocks was superseded by the re-introduction of the term aillikite by Malpas et al. (1986), thus recognizing these rocks as a subdivision of the ultramafic lamprophyres. The aillikites are described by Malpas et al. (op cit) as having phenocrysts of olivine and phlogopite in a groundmass of carbonate, apatite, mica, magnetite, and perovskite; aillikites lack feldspar and feldspathoids. Glimmerite nodules are locally abundant in aillikite dykes.

The association of alkaline and ultramafic lamprophyres in the Aillik Bay area has been used by Malpas et al. (1986) to suggest that these intrusions are related to a nephelinite-carbonatite intrusive centre situated slightly offshore, a centre that may have been generated indirectly by kimberlitic magmatism.

## NAIN AREA

A suite of deformed and variably recrystallized, amphibole-rich and carbonate-rich lamprophyric dykes has intruded a Paleoproterozoic monzonitic pluton and surrounding Archean gneisses on several small islands 30 km southeast of Nain (Ryan, 1992). These alkaline intrusions pre-date the ca. 1300 Ma Nain Plutonic Suite, and may be related to the igneous episode that produced the monzonite; they are, therefore, not related to the Aillik Bay intrusions.

The lamprophyre dykes of the Nain area are dark green to black-weathering, amphibole- and mica-rich, and locally carbonate-bearing. Many are foliated and folded, and disrupted along strike. These dykes are generally less than 2 m wide, and locally divide into several branches along strike. The carbonate-bearing ones are layered (from multiple intrusion ?) and are similar to the aillikites of the Aillik Bay area. The mineral assemblage of the lamprophyre dykes is variable, and indicates that alkaline and ultramafic types are present. Some of the "amphibolite" dykes contain granular to subhedral brown amphibole (barkevikite) as the major constituent, along with lesser amounts of granular to subhedral clinopyroxene, interstitial aggregates of plagioclase, and small flakes of orange biotite. The carbonate-bearing dykes are composed of calcite, variably serpentinized olivine, clinopyroxene, phlogopite, and opaque oxides. Other dykes are mica+clinopyroxene rocks, some of which contain serpentinized olivine. A few dykes contain a green spinel and a strongly pleochroic yellow humite-group mineral. A limited number of chemical analyses (Cadman et al., 1995) augment the petrography to date, and confirm that both alkaline and ultramafic lamprophyres occur within the suite.

## SAGLEK FIORD AREA

The Saglek-Hebron area in the northernmost part of the Labrador has several occurrences of rocks described by Collerson and Malpas (1977) and Bridgwater et al. (1990) as kimberlite dykes and glacially derived kimberlitic boulders. Brummer (1978) has noted that an unpublished map of Collerson's shows two kimberlite dykes on Big Island in Saglek Fiord, and several "diatremic breccia pipes" in the vicinity of Saglek. Collerson and Malpas (1977) described the kimberlite dykes, which are emplaced into early Archean gneisses, as containing olivine megacrysts and ovoid ultramafic fragments (micaceous dunite and glimmerite) in a fluidal-textured groundmass that contains phlogopite, diopside, perovskite, calcite, ilmenite, and melilite. The latter mineral, according to Mitchell (1991) is absent from both kimberlites and lamproites, thus rendering the Saglek intrusions questionable as true kimberlites. Melilite is, however, common in ultramafic lamprophyres (Rock, 1987), suggesting that the Saglek intrusions are members of this group instead. The diatremic rocks of the Saglek area appear to be equivalent to basaltic breccia dykes in the vicinity of Hebron Fiord (Ryan and Martineau, 1992), and between Hebron and Okak Bay (Ermanovics et al., 1989). These dykes are characterized by abundant angular to rounded fragments of surrounding gneiss and fine-grained mafic rock in a chloritized matrix that locally exhibits evidence of quenching. These breccia dykes do not appear to have alkaline affinities; they are probably rapidly injected volatile-rich members of a Paleoproterozoic diabase dyke swarm that fed now-eroded volcanic rocks like those of the nearby Mugford Group (Ermanovics, et al., 1989).

Bridgwater et al. (1990) reported the existence of a 2 m wide, dark, kimberlitic dyke in gneisses near Hebron village that is biotite-rich and contains olivine-rich nodules and garnetiferous xenoliths.

## OTHER LAMPROPHYRIC DYKES

In addition to the above, several other occurrences of lamprophyric dykes are known from Labrador. Wardle et al. (1993) recently reported the discovery of several biotite+olivine lamprophyre dykes within Archean and Paleoproterozoic gneisses near Cape Chidley at the northernmost tip of Labrador. Taylor (1979) has identified isolated lamprophyric dykes at several widely separated locations along the coast and the interior, suggesting that there are more intrusions of this type than presently known.

## DISCUSSION

The above summary outlines the general characteristics and distribution of lamprophyric and kimberlite-like dykes in Labrador. Malpas et al. (1986) and Collerson and Malpas (1977) concluded that the Aillik Bay intrusions and the Saglek intrusions, respectively, were generated at depths greater than 100 km. The macrocryst and inclusion assemblage in both areas is not encouraging for proposing that the magmas were derived from within the stability field of diamond, thus lowering the potential of these rocks as hosts to diamond. This does not, however, eliminate the areas containing such dykes from being targets for exploration because the controls on diamond formation and distribution are not yet completely understood. For example, until recently, lamproites and lamprophyres were not considered as viable hosts to significant quantities of diamond, the chief target of exploration companies being kimberlites. This original narrow outlook has been re-evaluated lately in light of the development of the rich, lamproite-hosted Argyle diamond deposit in Western Australia (Bergman, 1987), the occurrence of diamonds in nephelinitic and alkali basaltic rocks in New South Wales (Barron et al., 1994), and the recent discovery of abundant microdiamonds in a lamprophyre dyke in the District of Keewatin (McRae et al., 1996).

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# Ultramafic xenoliths and xenocrysts in kimberlite and alnöite: Windows to the upper mantle

**D. J. Schulze**

*Schulze, D.J., 1996: Ultramafic xenoliths and xenocrysts in kimberlite and alnöite: Windows to the upper mantle; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 129-133.*

## INTRODUCTION

Kimberlites and certain other alkaline rocks, such as minettes, lamproites, and alkaline basalts, contain inclusions of ultramafic rock that represent pieces of the Earth's upper mantle accidentally incorporated into the rising magmas. These invaluable fragments of otherwise inaccessible portions of the interior of the Earth are our best source of actual upper mantle material available at the Earth's surface for study. Such samples are the only material that can be used to constrain or guide geophysical modelling for depths within the Earth that cannot be drilled for confirmation (e.g., LITHOPROBE, see Jones et al., 1996). Furthermore, the compositions of some of these xenoliths and xenocrysts also provide information on the diamond potential of their host rocks.

Within Canada, ultramafic xenoliths are known from alkaline basalts, ultramafic lamprophyres, and kimberlites. Canadian occurrences of spinel-facies peridotite xenoliths in alkaline basalts were reviewed by Mitchell (1987) and Canil (1989). Extensive mantle xenolith suites in alkaline volcanic complexes of the Canadian Cordillera have been used to study variations in the composition and physical properties of the upper mantle from the edge of the North American craton across the accreted terranes of the Cordillera (Francis, 1987; Ji et al., 1994). These suites are not considered further here, as the spinel-facies xenoliths are of relatively shallow derivation and are unlikely to contain information on diamond occurrence and genesis. This paper focuses on deeper garnet-facies peridotite occurrences in Canada, as they may contain information on the diamond potential of their host rocks.

Kimberlites and lamproites are the only transport agents known to carry economic concentrations of diamond to the Earth's surface. Most of our knowledge of the primary mantle source rocks of diamond has come from study of syngenetic mineral inclusions in natural diamonds, supplemented by information obtained from diamond-bearing mantle xenoliths (see review by Gurney, 1989). The dominant mantle diamond source rock is thought to be harzburgite or dunite containing minor amounts of low-Ca garnet and/or high-Cr magnesiochromite. Garnet lherzolite is a minor mantle diamond source and eclogite typically is also a minor source, although the latter host can be locally abundant and very significant. In Siberia, the peridotite-suite of minerals overwhelmingly dominates the diamond inclusion population, with a peridotite:eclogite ratio of approximately 99:1 (Yefimova and Sobolev, 1977). In southern Africa, the peridotite-suite also dominates, although there, approximately one-quarter of mineral inclusions in diamond belong to the eclogite suite (Hawthorne et al., 1978). Some mines, such as Premier and Orapa, are in fact dominated by diamonds of the eclogite suite (Gurney et al., 1984, 1985).

Gurney (1984) showed that in the dominant peridotite inclusion suite in diamonds, approximately 85% of Cr-pyropes have lower CaO contents than those of garnets from garnet lherzolites, and termed the low-Ca Cr-pyropes "G10" garnets. Chromites included in diamonds have restricted, and high, $Cr_2O_3$ and MgO contents (see compilations by Fipke, 1989 and Gurney and Moore, 1993).

Texturally and chemically eclogites are divided into Groups I and II (MacGregor and Carter, 1970). Diamonds are associated with Group I eclogites (e.g., Robinson et al., 1984) which are characterized by garnets with elevated $Na_2O$ and $TiO_2$ contents and clinopyroxenes that are $K_2O$-rich (McCandless and Gurney, 1989; Gurney and Moore, 1993). It is these low-Ca Cr-pyropes, high-Cr chromites, and $Na_2O$-rich eclogite garnets that can potentially indicate the presence of diamond in a kimberlite of unknown diamond content.

## MANTLE SIGNATURES FROM DIAMOND-BEARING ROCKS IN CANADA

Diamonds, or diamond indicator minerals, such as G10 garnets, have been reported as xenocrysts in mantle-derived rocks from several Canadian localities.

Significant occurrences are reviewed below, approximately in the order in which they were discovered or their diamond content recognized. For additional information on specific locations, the reader is referred to other papers in this volume.

### *Île Bizard, Québec*

The first diamonds purported to have been found in a "primary" host rock in Canada were described from the Île Bizard alnöite near Montreal, Québec (Brummer, 1978; Bédard and LeCheminant, 1996). Xenoliths of garnet peridotite, spinel peridotite, garnet-spinel peridotite, garnet pyroxenite, websterite and glimmerite have been described from this body by Marchand (1970), Raeside (1978) and Raeside and Helmstaedt (1982). No diamond indicator minerals were found by Fipke (1989). Equilibration conditions of garnet lherzolite and garnet websterite xenoliths are significantly shallower than the diamond stability field (Fig. 1). The presence of diamonds at Île Bizard remains an enigma.

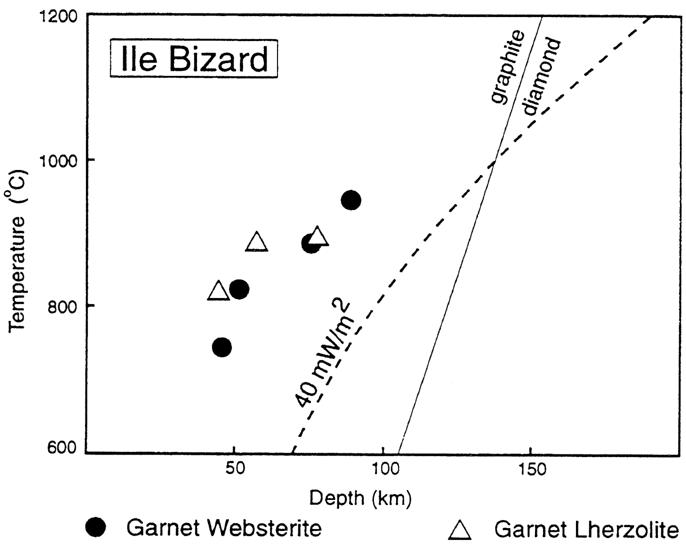


***Figure 1.*** *Equilibration conditions of garnet lherzolite (circles) and garnet websterite (triangles) xenoliths from Île Bizard. Data from Raeside (1978) and Schulze (unpublished data). Reference curves are the graphite-diamond transition of Kennedy and Kennedy (1976) and a subcontinental geothermal gradient corresponding to a surface heat flow of 40 mW/m$^2$ (Pollack and Chapman, 1977).*

### *Somerset Island, N.W.T.*

Trace quantities of diamonds have been recovered from kimberlites on Somerset Island, N.W.T. (Pell, 1995; Kjarsgaard, 1996a). Low-Ca garnet harzburgites (Kjarsgaard and Peterson, 1992) and marginally G10 garnets (Fipke, 1989; Kjarsgaard and Peterson, 1992; Schulze, 1993) are known from the Batty pipe, and garnet compositions and abundances are consistent with the trace diamond abundance. The equilibration conditions of garnet lherzolite indicate an elevated geothermal gradient at the time of eruption (Mitchell, 1987; Kjarsgaard and Peterson, 1992; Kjarsgaard, 1996a), and the deepest samples are consistent with the presence of diamonds in some of these pipes (Kjarsgaard, 1996a).

### *Cross diatreme, British Columbia*

A single kimberlite body (the Cross diatreme) has been recognized in the Rocky Mountains of British Columbia, although numerous other non-kimberlite alkaline rocks are also known ( Ijewliw and Pell, 1996). Garnet peridotite xenoliths have been recognized only in the Cross pipe, which also contains spinel peridotites. Two fresh xenoliths of garnet lherzolite have been analyzed that yield equilibration conditions outside of the diamond stability field (Hall, 1991). Although low-Ca Cr-pyropes are present in the heavy mineral xenocryst suite (Hall, 1991), no diamonds have been reported from this kimberlite. A few diamonds have been recovered, however, from the nearby Jack diatreme (Fipke, 1989; Ijewliw and Pell, 1996).

### *Kirkland Lake, Ontario*

Several kimberlites in the Kirkland Lake kimberlite cluster (Schulze, 1996) contain trace quantities of diamonds (Brummer et al., 1992). Although not known to be of economic grade, the indicator mineral signature and garnet peridotite equilibration conditions indicate this cluster sampled appropriate mantle depths and compositions to account for the presence of diamond. Low-Ca pyropes and high-Cr magnesiochromites occur in most pipes in this region, but abundances are low. Garnet lherzolites, some of which show evidence of metasomatic replacement of garnet by diopside + phlogopite, dominant the mantle xenolith suite and record equilibration conditions extending well into the diamond stability field (Schulze, 1996; P.A. Vicker, unpublished data). Only one low-Ca garnet harzburgite has been documented and eclogites are common only in the A-4 pipe, but have not yet been studied in detail. Eclogite garnet xenocrysts are uncommon, and are dominantly from Group II eclogites (D.J. Schulze, unpublished data) that are typically non-diamondiferous (e.g., McCandless and Gurney, 1989).

### *Attawapiskat, Ontario*

A large cluster of kimberlites (at least 19 bodies; Zalnieriunas and Sage, 1995), some of which contain diamonds, occurs in the Attawapiskat region of northern Ontario. The kimberlites intrude Paleozoic sedimentary rocks; one pipe is exposed at surface and the others have been drilled. The mantle xenolith population includes coarse and deformed garnet, garnet-spinel and spinel peridotites, plus rare eclogites, in addition to garnet, diopside, and ilmenite megacrysts. As yet, no isotopic or chemical data are available for the xenoliths. Evidence of a pre-eruption metasomatic event exists as diopside-phlogopite clots in peridotites that appear to represent garnet replacement, analogous to the metasomatism noted in garnet lherzolite xenoliths from Kirkland Lake. Ilmenite macrocrysts from one body range in MgO content from 5 to 12 wt% in cores, within a relatively restricted range in $Cr_2O_3$ (1.4 to 1.8 wt%); hematite component is in the range 6 to 23 mole % $Fe_2O_3$ (Schulze, et al., in press), consistent with the presence of diamond.

### *Fort à la Corne, Saskatchewan*

Tens of kimberlite bodies have been discovered in the Sturgeon Lake and Fort à la Corne areas of Sakatchewan, and some have low to moderate diamond contents (Kjarsgaard, 1996b). Eclogite and garnet peridotite xenoliths, including a low-Ca garnet dunite, have been recovered from the kimberlites, but no chemical data have been published for the xenoliths (B.A. Kjarsgaard, pers. comm., 1995). A few low-Ca (G10) garnets have been identified (Fipke, 1989; Lehnart-Thiel et al., 1992; Schulze, 1993), and garnet populations are similar to those in the Somerset Island kimberlites. Both group I and II eclogite garnet xenocrysts have been recognized (Schulze, 1993). Macrocrysts of magnesian ilmenite in one kimberlite range to high $Cr_2O_3$ values (0.2 to 5.3 wt% $Cr_2O_3$ at 10 to 15 wt% MgO), with a small range in hematite component (9 to 16 mole % $Fe_2O_3$; Schulze, et al., in press), consistent with the presence of diamonds and even more reduced than the Attawapiskat ilmenites.

### *Slave Province, N.W.T.*

Although all indications are that many kimberlite pipes near Lac de Gras (Pell, 1995; Kjarsgaard, 1996c) have significantly higher diamond contents than any other kimberlites known from North America, almost no data are available on the mantle sampled by these pipes. Xenoliths of garnet peridotite and eclogite, and Cr-diopside megacrysts occur in kimberlites drilled by BHP, which also contain "diamond indicative" high-Cr chromites and low-Ca Cr-pyropes in their heavy mineral concentrates (M. Kirkley, pers. comm., 1995). Data obtained for lherzolite-derived garnet-clinopyroxene xenocryst pairs by Dynes (1994) show equilibration temperatures consistent with sampling of the diamond stability field for the diamond-bearing Ranch Lake pipe (ca 1050 -1150 °C).

## CONCLUSIONS

Kimberlites provide "windows" to the upper mantle that afford us glimpses into small regions of the subcratonic lithosphere, essentially equivalent to drill holes, with poor recovery, to >100 km. Work to date has allowed information to be gleaned from a few "drill holes" at specific localities in Canada, where diamond-bearing source rocks have been tapped. Additional information will come from recently discovered Slave Province, Attawapiskat, and Saskatchewan kimberlites. Perhaps the most complete examination of the subcratonic lithosphere will come from the many kimberlites that have sampled the mantle beneath the central Slave Province, allowing us to create a three-dimensional model of a diamond-bearing part of the craton, one that seems to have the highest concentrations of diamonds in Canada.

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# Geochronological and petrogenetic studies of lower crustal xenoliths entrained in kimberlites and alkaline rocks

**W.J. Davis and D. Moser**

*Davis, W.J. and Moser, D., 1996: Geochronological and petrogenetic studies of lower crustal xenoliths entrained in kimberlites and alkaline rocks; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 135-137.*

## INTRODUCTION

Understanding of the tectonic processes by which Precambrian cratons formed and evolved has increased tremendously over the last two decades, and these processes have been linked to the presence of diamonds in the subcratonic lithosphere (Helmstaedt and Gurney, 1984; Helmstaedt and Schulze, 1989). For the most part, tectonic models are based on geological and geochronological studies of exposed upper crustal rocks, since comparatively little information about the evolution of the lower continental crust of the cratons is available. A more comprehensive understanding of how cratons form and their subsequent thermal history requires detailed petrogenetic and geochronological studies of the lower crust, in combination with geophysical observations. Although rare, pieces of the lower crust may be entrained as xenoliths in kimberlites and other alkaline intrusions during their ascent, and thus provide unique samples of the lower crust for direct study.

Lower crustal xenoliths typically have mineral assemblages of the granulite facies indicating crystallization at pressures > 0.6 GPa. Generally, xenoliths are dominated by mafic rock types (pyroxenites ± garnet), although felsic varieties, including metasedimentary rocks, are locally important (Rudnick, 1992). The dominantly mafic composition of the lower crust determined from xenolith studies contrasts with generally felsic compositions of exposed granulite terrains (Rudnick, 1992), but is supported by seismic velocity studies which indicate faster and therefore more mafic crust below approximately 30 km depth (Holbrook et al., 1992). Most well studied xenolith suites are from Proterozoic or younger crust and few samples are available from Archean cratons (Rudnick, 1992). For this reason, less is known of the lower crust beneath the Archean cratons than that beneath younger terranes.

Petrological, geochemical and isotopic studies of lower crustal xenoliths have been used to infer the origin, time of formation, and thermal history of the lower crust (Rudnick, 1992 and references therein). The origins of the xenoliths are, in many cases, difficult to establish. Some may be tectonically buried upper crustal rocks, some the residues of partial melting events that produced granites in the upper crust, whereas others may be mafic intrusions and cumulates of underplated mafic magmatism. One of the key pieces of information for xenolith studies is high-precision geochronological data. U-Pb dating of zircon and other accessory minerals such as titanite and rutile can be used to establish the time of formation and/or the time of metamorphism of the lower crust. This information is essential to examine relationships between the time of formation of upper and lower crust and to evaluate processes such as tectonic or magmatic underplating. In some cases, a temporal link between metasomatic events in the mantle and thermal/metasomatic events in the lower crust may be established (Amelin et al., 1994; Carlson and Irving, 1994; Davis, 1994). In addition, the extent of thermal reactivation of the lower crust during younger tectonic events such as intrusion of mafic dyke swarms (LeCheminant et al., 1996) and formation of marginal collisional orogenic belts can be evaluated.

## RECENT CANADIAN STUDIES

Samples of lower crustal xenoliths from the Canadian Shield have become available as a result of recent successes in diamond exploration. A number of multi-isotopic (U-Pb, Sr, Nd) studies of lower crustal xenoliths from the Precambrian cratons are currently underway, including samples from the Superior Province (Kirkland Lake area; Moser and Heaman, 1994), the Grenville Province of Quebec (Amelin et al., 1994), the Archean Medicine Hat Block in southern Alberta and northern Montana (Davis, 1994), and the Yamba Lake - Lac de Gras area of the central Slave Province (W.J. Davis, unpublished data). In all of these regions the lower crustal xenoliths are dominated by mafic compositions. Initial results of these studies are summarized below.

Lower crustal granulite xenoliths from kimberlites in the Kirkland Lake area of the Superior Province experienced two major periods of metamorphic zircon growth, one at 2.58 Ga and one at ca. 2.49 Ga (Moser and Heaman, 1994). The older age indicates that the lower crust underwent high-grade metamorphism almost 100 Ma after low-grade regional metamorphism of the rocks exposed at surface (see also Krogh, 1993). This lower-crustal event is coeval with episodes of hydrothermal alteration in ore deposits of the Kirkland Lake area. The younger period of zircon growth at 2.49 Ga is interpreted to reflect increased heat flow associated with Matachewan rifting along the southern margin of the Superior Province and deposition of the Huronian Supergroup.

In the Archean Medicine Hat Block of southern Alberta, lower crustal rocks experienced a significant granulite metamorphic event at 1.7 to 1.8 Ga (Davis, 1994), more than 1 Ga after formation of the Archean tonalitic crust which characterizes the block (Villeneuve et al., 1993). The younger metamorphism in the lower crust is temporally related to Proterozoic deformational events on the margins of the block resulting from collisional assembly of Laurentia between 1.8 and 1.7 Ga. The time of granulite metamorphism in the lower crust can also be linked to the time of a metasomatic enrichment event in the lithospheric mantle source of the minette intrusions that carry the xenoliths. (Kjarsgaard and Davis, 1996; W.J. Davis and B.A. Kjarsgaard, unpublished results). These results indicate significant reworking of the Archean lithosphere in the Paleoproterozoic.

In the Grenville Province, Amelin et al., (1994) suggested that mafic lower crustal xenoliths from beneath the Central Metasedimentary Belt of Québec formed at ca. 1.3 Ga, and may represent underplated cumulates from a major episode of tholeiitic volcanism recognized in the Central Metasedimentary Belt (see also Corriveau et al., 1996). The lower crust was thermally reactivated at the time of emplacement of the host lamprophyre dyke at ca. 1.08 Ga.

## SUMMARY

Petrogenetic and geochronological studies of lower crustal xenoliths are essential to develop comprehensive models of the development and evolution of Precambrian cratons. Many lower crustal xenoliths are extremely small (<2 cm), and contain few zircons. The zircons may exhibit complex growth histories. The capability of the Geological Survey of Canada to carry out these types of geochronological studies has been enhanced by the arrival of the SHRIMP ion microprobe.

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# Insights on minette emplacement and the lithosphere underlying the southwest Grenville Province of Québec at 1.08 Ga

**L. Corriveau, D. Morin, M. Tellier, Y. Amelin, and O. van Breemen**

*Corriveau, L., Morin, D., Tellier, M., Amelin, Y., and van Breemen, O., 1996: Insights on minette emplacement and the lithosphere underlying the southwest Grenville Province of Québec at 1.08 Ga; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 139-142.*

## INTRODUCTION

Lower crustal and mantle xenoliths in the Rivard minette dyke are a unique probe of the lithosphere underlying the southwestern Grenville Province in the Mont-Laurier area, Québec, at 1.08 Ga (Fig. 1). Research on the dyke by the GSC and an industry partner, KWG Ressources, combined with GSC regional surveys, give the mineral industry a framework to establish a diamond exploration strategy in the southwestern Grenville Province (Corriveau and Jourdain, 1993; Corriveau and Madore, 1994; Corriveau et al., 1994, 1996a,b; Corriveau and Leblanc, 1995; Davidson, 1996; Hetu and Corriveau, 1992, 1995; Tellier et al., 1994, 1995). The research also provides insights into the mechanism and rate of ascent of minette magmas at deep crustal levels.

## RIVARD MINETTE DYKE

The 1.08 Ga Rivard dyke (>200m long and <1.7m wide; Fig. 2) is a N-trending, steeply dipping, multiple intrusion breccia that contains abundant xenoliths. It was emplaced about 100 million years after regional metamorphism at mid-crustal depth in a quartzofeldspathic gneiss complex. A second intrusion breccia is exposed along strike 8 km to the south (Fig. 1). As no pre-existing fault zones were observed in the field, the distribution of the two dykes suggests that regional fractures formed at the time of emplacement. The host lamprophyre is a minette with biotite and clinopyroxene phenocrysts set in a groundmass of alkali feldspar, amphibole, apatite, biotite, clinopyroxene, plagioclase, quartz, sulphides and titanite. It is ultrapotassic ($K_2O/Na_2O$ in weight percent (wt.%) = 2.2), contains 5 wt.% $K_2O$, 50 wt.% $SiO_2$, 6-9 wt.% MgO, and is enriched in large ion lithophile elements (3000 ppm Ba, 95 ppm Ce, La/Yb = 20). The dykes are coeval with a 450 km long belt of 1.08 Ga K-rich alkaline plutons characterized by geochemical signatures typical of island arcs (Corriveau, 1990; Corriveau et al., 1990; Corriveau and Gorton, 1993). The arc signature may be interpreted isotopically in terms of a 1.3 Ga subduction-related mantle enrichment event (Corriveau and Amelin, 1994).

## XENOLITHS IN THE RIVARD DYKE

The minette breccia contains 40-70% rounded to angular xenoliths (Fig. 2). Xenoliths of garnet-, aluminous spinel- and phlogopite-clinopyroxenites and websterites (Mg# 56-92; up to 3400 ppm Cr; 3-18 wt.% $Al_2O_3$), and mafic granulites have Nd and Pb model ages of ca. 1.9 Ga. Garnet-clinopyroxene Sm-Nd and Pb-Pb ages cluster around 1.08 Ga, the time of xenolith extraction from the lower lithosphere. Younger Rb-Sr ages between 630 and 840 Ma are indicative of post-crystallization resetting in garnet. Dunite and wehrlite xenoliths contain olivine ($Fo_{90-94}$) and aluminous spinel, and have unusually low Ni (ca. 40 ppm) and Cr (ca. 125 ppm) contents. The xenoliths record maximum P-T conditions of ca. 25 kbar and 1050°C, and lie on a high geothermal gradient outside the diamond stability field. Initial isotopic values of $\epsilon_{Nd}$(1.08 Ga) and $^{206}Pb/^{204}Pb$ for clinopyroxene are similar to those of contemporaneous potassic plutons, but initial $^{87}Sr/^{86}Sr$ values are distinctly higher (Amelin et al., 1994; Corriveau and Amelin, 1994). This isotopic difference indicates that none of the xenoliths studied represent the source of the K-rich alkaline magmas. The apparent absence of ultra high-pressure xenoliths and the presence of cognate clinopyroxene-phlogopite xenoliths suggest that magma did not ascend directly from source regions, but first ponded in the upper mantle. Thus, the 25 kbar pressure recorded by the xenoliths provides only a minimum depth of origin (ca. 75 km) for this type of magma.

## DISCUSSION

Rate of magma ascent was calculated from the size of ultramafic xenoliths in the Rivard dyke, and from estimates of magma viscosity, density and volatile content as well as effective viscosity (Morin and Corriveau, in press). At an early stage of ascent, the minimum velocity of the magma was on the order of 1 to 10 km/hour.

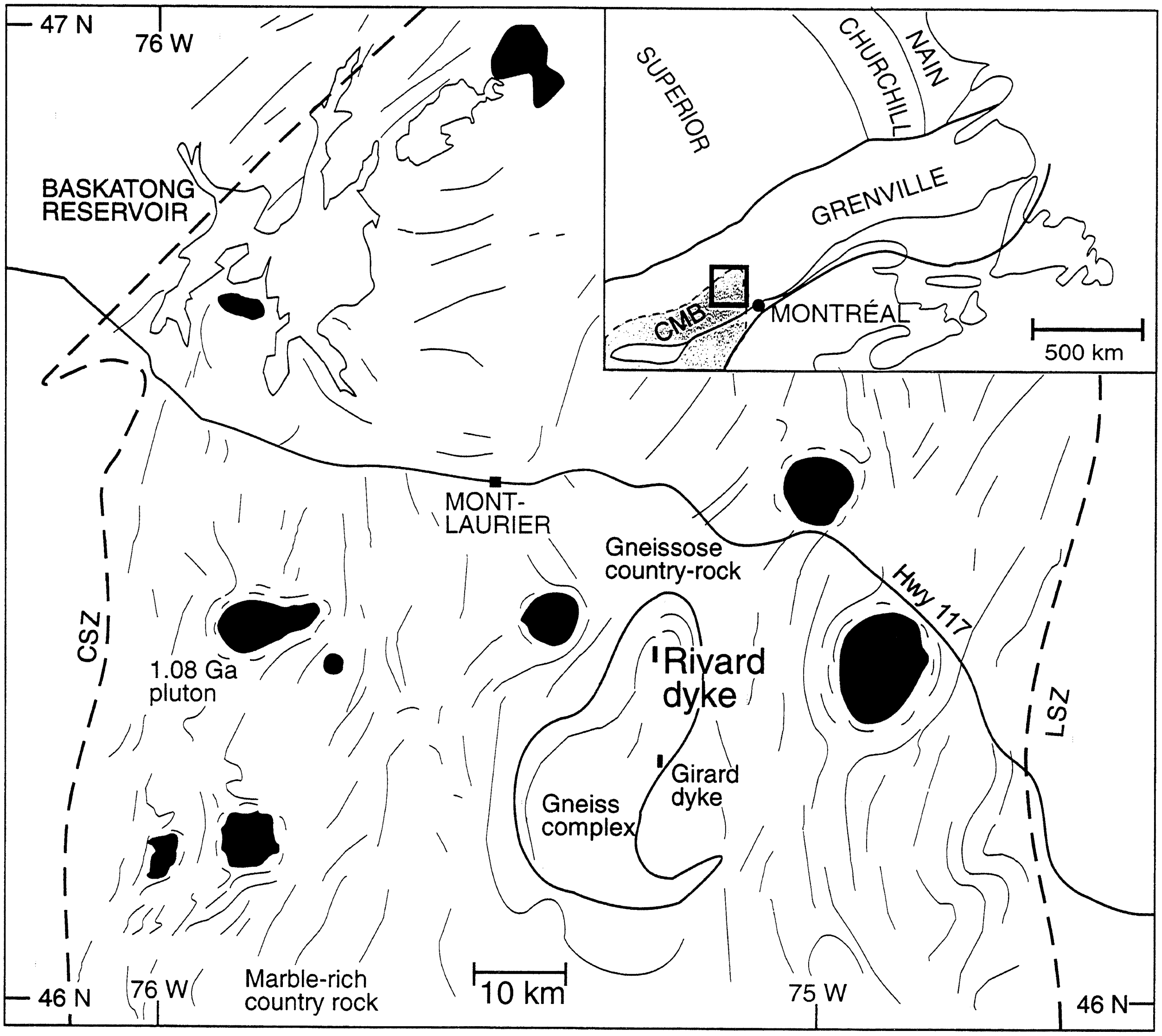


***Figure 1.*** *Geological and structural setting of 1.08 Ga K-rich alkaline magmatism in the Mont-Laurier area, Québec. Plutons (black) are spatially associated with marble-rich host rocks, whereas, the minette breccia dykes occur in a gneiss complex. LSZ is the Labelle shear zone (Martignole and Corriveau, 1991) and CSZ is the Cayamant shear zone (Sharma et al., 1993). Inset map shows the location of the Mont-Laurier area in the Central Metasedimentary Belt (CMB) of the Grenville Province.*

During ascent, the magma had the capability of fragmenting its wall rocks, as indicated by the extensive xenolith suite transported. Incorporation of xenoliths increased the effective viscosity of the liquid-solid mixture to the extent that it became too viscous to rise, and the magma solidified at mid crustal levels (Morin and Corriveau, in press). The dykes appear to be restricted to gneisses, whereas, coeval plutons occur in marble-bearing country rock (Fig. 1). Country-rock rheology influenced the mode of magma ascent and the nested plutons crystallized as a result of pooling of magma within marble-rich crust (Corriveau and Leblanc, 1995).

Regional metamorphism occurred at 1.19 Ga in the southwestern Grenville Province and was followed by a widespread 1.16 Ga magmatic event. Metamorphism is attributed to a pre-1.19 Ga arc-continent-microcontinent collision (Corriveau et al., 1995a,b). This model is consistent with the presence of mylonitic xenoliths in the

*Figure 2. The Rivard minette dyke and its xenoliths. Dyke width is 1.5 m. Most small xenoliths are ultramafic rocks. Xenoliths of locally derived gneisses are large and angular.*

Rivard dyke, which indicate that the crust sampled at depth was already dissected by major shear zones at 1.08 Ga. Gabbro and anorthosite xenoliths in the dyke may be derived from layered intrusions like the 1165 Ma intrusions of the Mont-Laurier area. Based on the xenolith suite and regional information, such as the distribution of 1.16 Ga plutons, it is proposed that the minette magma ascended through reworked Paleoproterozoic and Mesoproterozoic rocks at the eastern margin of the Superior Archean craton. Due to continuing northwest-directed thrusting onto the craton after dyke emplacement (Martignole and Pouget, 1994), upper parts of the lithosphere sampled by the dyke lie directly beneath the present-day surface exposure, but lower slices are likely now farther to the east.

Analyses of heavy mineral concentrates recovered from tills and from the Rivard dyke have not revealed any indicator minerals typical of those carried by diamond-bearing kimberlites. Gamma-ray spectrometric and magnetic airborne survey were also conducted (Hetu and Corriveau, 1992, 1995), but no anomalies are associated with the potassic stocks and dykes in the survey areas.

Research on the Rivard dyke suggests that primary diamond sources older than 1.08 Ga are unlikely in the southwestern Grenville Province of Québec. However, the mantle underlying the Mont-Laurier area was highly enriched at 1.3 Ga, and since the present-day geothermal gradient is much lower than a billion year ago, more recent alkaline magmas, if present, may have diamond potential.

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# Fossils as indicators of thermal alteration associated with kimberlites

**A.D. McCracken, D.K. Armstrong[1], D.C. McGregor**

*McCracken, A.D., Armstrong, D.K., and McGregor, D.C., 1996: Fossils as indicators of thermal alteration associated with kimberlites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 143-145.*

## INTRODUCTION

Fossils have long been the standard for determining the geological age of Phanerozoic sedimentary strata, but now they are also being used extensively for geothermometry (Poulton, 1989). Increasing temperature causes progressive and irreversible changes in organic material. In two groups of microfossils, conodonts (tooth-like fossils composed of apatite; Fig. 1) and acritarchs (organic-walled unicellular organisms of largely unknown origin), the changes are primarily in colour.

Colour changes in conodonts have been known for some time, but it was the work of Epstein et al. (1977) that showed the colour change was time and temperature dependent. Epstein et al. (ibid.) produced an index of colour changes, the Conodont Alteration Index (CAI) from 1 (pale yellow) to 5 (black). Through laboratory experimentation and examination of field collections, they correlated these changes with temperature ranges, with CAI 5 corresponding to 300 to 400°C. Additional experimentation has added new indices to the CAI scale, the highest value, CAI 8, corresponds to temperatures

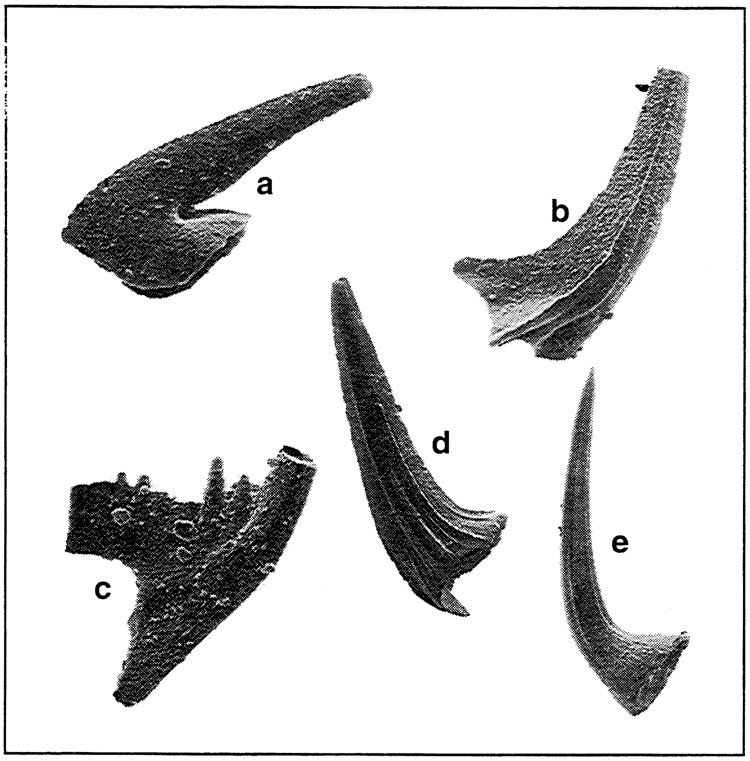


*Figure 1 Representative Lower Ordovician conodonts from the Beauharnois Formation, GSC loc. O-106029, Grande-Ile, Québec. (a)* ***Drepanoistodus angulensis*** *(Harris), GSC 113159, x 50. (b)* ***Acodus delicatus*** *(Branson & Mehl), GSC 113139, x 50. (c)* ***Oepikodus communis*** *(Ethington & Clark), GSC 113163, x 70. (d)* ***Scolopodus subrex*** *Ji & Barnes, GSC 113170, x 22. (e)* ***Colaptoconus quadraplicatus*** *(Branson & Mehl), GSC 113153, x 32.*

*1 Contribution by D.K. Armstrong is by permission of the Director of the Ontario Geological Survey.*

>600°C (Rejebian et al., 1987). Colour changes in acritarchs (AAI) are also tied to ranges of temperature (Legall et al., 1981).

Regional studies of CAI have been used to assess the hydrocarbon potential of sedimentary strata (Epstein et al., 1977; Utting et al., 1989), and to document differences in the geological histories of Mississippi Valley type lead-zinc deposits (Sangster et al., 1994).

Research on CAI has been directed towards refining the indices and other characters with both experimental and field data, and applying these in local and regional geological interpretations. Burnett (1988) investigated surface microtexture of conodonts and noted that diagnostic changes in faunal composition were due to heat related stress (i.e., differential thermal expansion). For example, the total number of conodonts recovered per kilogram of rock decreases towards a dyke because of thermal destruction, and the percentage of more robust or stress-resistant forms increases.

## CONTACT METAMORPHIC EFFECTS

Field testing of conodont alteration around localized intrusions shows that the alteration aureoles have limited extent. The Fitzroy lamproite in the Canning Basin of Western Australia consists of a number of plugs, dykes, and sills. The CAI value is high (8) at the contact of a small volcanic plug (less than 20 m surface outcrop), but it diminishes to the regional value of CAI 1 just one metre away (Nicoll, 1981). The CAI pattern around the 30 m thick Holy Island quartz dolerite dyke in northeast England ranges from CAI 5 at 7.5 m from the dyke to CAI 3 at 22.5 m, decreasing to regional values at 55 m (Burnett, 1988).

One of the effects of contact metamorphism is the variability of CAI within a sample - conodonts in roof pendants have a wide range of CAI values (Rejebian et al., 1987). Rejebian et al. (ibid.) discovered that the uniformity or variability of CAI values within a sample and surface microtexture can help to distinguish grades and environments of metamorphism, and that hydrothermal activity may also affect a conodont's appearance.

## REGIONAL METAMORPHIC EFFECTS

Regional studies of CAI have been used to interpret proximity of strata to intrusive bodies. Aldridge (1984) found CAI values in Silurian conodonts of the Oslo region increased towards outcrops of Permian igneous rocks. Nowlan and Barnes (1987a) produced a number of maps for eastern Canada showing CAI values and found several instances where high CAI values were in regions known to contain igneous intrusions. In these two studies, the intrusions are younger than the rocks containing the fossils. Older intrusive bodies may also influence CAI in younger strata. For example, in the Pennines of northern England, CAI 4 values were found in Carboniferous strata overlying Devonian granite batholiths, but CAI 1.5 occurred in strata above the basement metamorphic rocks. Burnett (1987) suggested that the anomalous CAI was related to greater thermal conductivity of the granites compared to the metamorphic rocks.

An interesting view of two eastern Canadian CAI anomalies is that they are the trace of mantle hot spots. Legall et al. (1981) suggested this cause for elevated CAI in the region from Ottawa to Montréal (Ottawa-Bonnechère graben to Monteregian intrusives). Nowlan and Barnes (1987a) interpreted the same for CAI values in northern Newfoundland, and noted that the thermal effects of hot spots could be important in the generation of localized hydrocarbons. Nowlan and Barnes (1987b) thought that hot spot traces could be used to define narrow belts for exploration of diamond-bearing kimberlites.

## NORTHERN ONTARIO KIMBERLITES

Regional studies of CAI and other fossil thermal indicators rely heavily on existing data originally compiled for age determinations. Thermal histories are thus better developed in areas of more detailed collecting. Microfossil faunas, and their thermal alteration, are not as well documented from northern Ontario, in part because of limited exposure of sedimentary strata in the Hudson Bay Lowlands and the few outliers on the Canadian Shield. From limited data, the background CAI in northern Ontario is low (1-1.5). However, one anomaly is known from a glacially transported kimberlite xenolith found near Larder Lake. These few conodonts were found to have a value of CAI 3 (Nowlan, 1987). Diamond exploration drill core from northeastern Ontario is now being studied for regional geological correlation. Carbonate samples, including xenoliths, from various cores are being processed for conodonts and acritarchs. These fossils will provide age determinations and help establish regional CAI/AAI values. This work is in its preliminary stage - conodonts close to kimberlite have been found altered, but the changes, especially in surface microtexture, have yet to be investigated.

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# Thermal data from petrographic analysis of organic matter in kimberlite pipes, Lac de Gras, N.W.T.

**L.D. Stasiuk and W.W. Nassichuk**

*Stasiuk, L.D. and Nassichuk, W.W., 1996: Thermal data from petrographic analysis of organic matter in kimberlite pipes, Lac de Gras, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 147-149.*

## INTRODUCTION

Diverse assemblages of Cretaceous and Tertiary fossils have been recovered from xenoliths in kimberlite pipes that intrude Archean rocks in the Slave Province near Lac de Gras (Nassichuk and McIntyre, 1995; 1996). Phanerozoic strata are absent at the surface and the fossils are the first evidence that marine sediments were deposited in the Slave Province from uppermost Early Cretaceous (Albian) to latest Cretaceous (Maastrichtian) time. Slightly younger Tertiary (Paleocene) spore and pollen fossils indicate deposition in a nonmarine, possibly lacustrine environment. During late Paleocene or early Eocene time, kimberlites erupted to the surface penetrating the Cretaceous and Tertiary strata and blocks of country rock fell into the crater facies and became incorporated with kimberlitic rocks as xenoliths (Nassichuk and McIntyre, 1995).

Fragments of fossilized wood have been recovered from boreholes penetrating the crater and diatreme facies of eight pipes. Wood is particularly common from depths <50 m in the crater facies, where it appears fresh. Wood from depths >120 m in the diatreme facies has a more coalified appearance. We have analyzed five samples of wood and two samples of coaly black shale and mudstone from four different pipes (Fig. 1) to determine the petrographic character and thermal history (coal rank) of the organic matter. In particular, we are interested in the maximum temperature that the wood and other organic matter were exposed to following original deposition. To achieve that end, we analyzed polished, block and particulate samples using incident light microscopy to assess the organic matter or maceral composition and to determine the percent reflectance in oil (% Ro) for coal

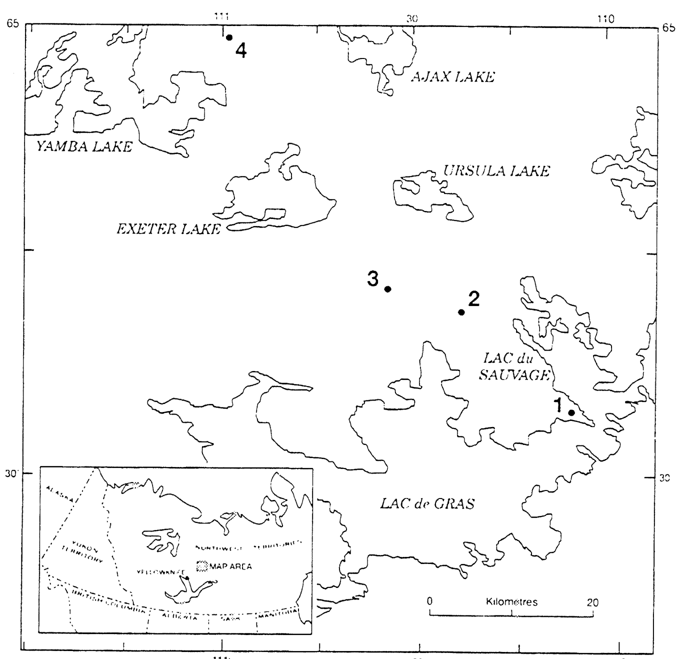


*Figure 1. Map of Lac de Gras area showing position of kimberlite pipes containing fossils that were referred to by Nassichuk and McIntyre (1995) and Stasiuk and Nassichuk (1995). 1) BHP-Dia Met Point Lake pipe, 2) BHP-Dia Met "Hawk" pipe, 3) BHP-Dia Met Koala pipe, 4) Mill City Tanqueray Torrie pipe. Samples were provided through the courtesy of the BHP-Dia Met Joint Venture in the Lac de Gras area and the Tanqueray-Mill City Joint Venture of Yamba Lake.*

rank assessment. Our data is of considerable importance in evaluating temperatures in the crater and diatreme facies of kimberlite pipes in the Slave Province shortly after emplacement.

## PETROGRAPHIC AND REFLECTANCE DATA

Larger (up to 12 cm long) fragments of fossil wood are dominated by the equivalent of coal huminite macerals (i.e. those organic remains which have been derived from lignin and cellulose of terrestrial vascular plants). Well preserved vascular plant cellular structure (e.g. ray cells, vascular bundles) was observed in one diatreme facies sample (Torrie pipe; Fig. 1) and in all crater facies samples. The low-reflecting, peatified to lignitic woods from the crater facies also contain a significant amount of cellulose, as expected by their fresh, unaltered appearance. A sample from the diatreme facies in the Point Lake pipe (Fig. 1) is significantly more coalified in appearance than other samples and does not have the same details of vascular plant morphology preserved. Two samples from the Hawk pipe (Fig. 1) are shales or mudstones that contain microscopic, detrital macerals dispersed within the mineral matrix. All three maceral groups (inertinite>huminite>liptinite) are present in these samples. Inertinite is dominated by detrital inertodetrinite particles, many of which have frayed margins suggesting significant reworking in the paleoenvironment. Liptinite macerals are important in both of the Hawk samples. These liptinites are dominated by pollen with lesser amounts of sporinite, cutinite, alginite, traces of dinoflagellates, oxidized plant resins, and exsudatinites (fluorescing, oily, solid hydrocarbons).

Reflectance in oil (%Ro) was determined for several types of huminite macerals (see Stasiuk and Nassichuk, 1995). Reflectance values for samples from the Koala and Torrie pipe samples range from 0.19 to 0.26%Ro. Some of the woody huminites have an anisotropy typical of cellulose. This is a compositional characteristic typical of huminite in a very early stage of diagenesis and a very low level of "coalification". The Point Lake and Hawk samples have higher reflectance values ranging from 0.33 to 0.44%Ro. A bar chart comparing the reflectance of huminite for all samples reveals two distinct classes (Fig. 2). The Hawk and Point Lake diatreme samples form a high-reflectance group (Group B) while the Koala and Torrie, mainly crater samples (2 out of 3), form a low-reflectance group (Group A).

## DISCUSSION

### *Rank-thermal history*

The huminite macerals from the diatreme and crater facies do not show any direct evidence for a pyrometamorphic event (i.e., temperatures >250°C). This is not surprising since it has been noted that country rock in the Bachelor Lake kimberlite breccias, South Africa, show no pyrometamorphic effects and were probably emplaced at low temperatures (Clement, 1975). The presence of unaltered, primary asphaltic bitumen (exsudatinite) without flow textures in the Hawk pipe samples also suggests that the maximum temperature was less than 104°C, the temperature at which micro-flow begins in asphalt bitumens under atmospheric conditions (Jacob, 1984). The low reflectance of Group A crater facies samples (Fig. 2) is typical of peatified woods which have only been buried to very shallow depths. Based on a well established time-temperature-%Ro relationship, for a burial period of 50 million years and a %Ro of 0.20, the maximum average temperature for Group A samples must have been less than 30°C (e.g., Bostick, 1979).

The range of reflectance values for most Group B diatreme samples (Fig. 2) fall within the range of reflectance for coals of lignitic to sub-bituminous "C" rank (Stach et al., 1982). Since the diatreme facies samples have higher reflecting huminite than the crater facies samples, a different thermal history must be invoked. Based on current data, the most likely thermal history scenario for the coaly material in the diatreme is a "normal" burial history, where virtually no thermal effect due to kimberlite emplacement affected the wood during or after incorporation into the breccia. In this case, all of the coalification postdates the kimberlite. Thus, to achieve a 0.40 to 0.45 %Ro coalification level, an average maximum temperature of 50°C would be required over an effective heating interval of 50 million years (i.e. for an average geothermal gradient of 30°C/km and a burial depth of 1 km; e.g. Bostick, 1979). If the diatreme huminites are older (e.g., Albian) than the crater facies huminites (e.g., Paleocene), then the lignite to sub-bituminous level of coalification could have been achieved by burial prior to incorporation into the kimberlites.

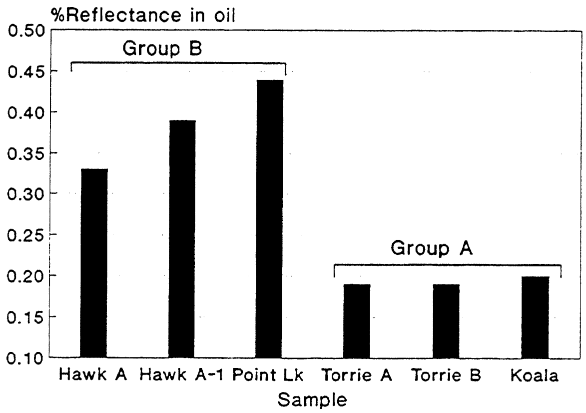


*Figure 2. Bar diagram illustrating the low-reflectance (Group A) and high-reflectance (Group B) huminite populations. All of Group B samples are from the diatreme facies; three of four in Group A are from the crater facies.*

## CONCLUSIONS

Further investigation is required to better constrain the depositional history of the diatreme huminite and more accurately assess the conditions by which lignite to sub-bituminous levels of coalification were achieved. In particular, a precise geological age of the crater and diatreme fossilized woods is essential. At this stage in the investigation, however, it can be concluded that the kimberlite intrusion had a negligible effect on the organic constituents trapped within the crater and upper diatreme stratigraphic level of the pipe.

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# Thermal evolution of the lithosphere in the central Slave Province: Implications for diamond genesis

**P.H. Thompson, A.S. Judge, and T.J. Lewis**

*Thompson, P.H., Judge, A.S., and Lewis, T.J., 1996: Thermal evolution of the lithosphere in the central Slave Province: Implications for diamond genesis; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 151-160.*

## INTRODUCTION

Thermal conductivity and radiogenic heat production in the crust strongly influence the escape of heat from the interior of the earth and hence the temperature distribution in the lower crust and lithosphere (Pollack and Chapman, 1977; Morgan, 1984). These parameters together with the present terrestrial heat flow form the essential components of numerical models constructed to explain the thermal evolution of the Slave Province, its metamorphic history and, of particular interest, the time at which the lithosphere became thick and cool enough to contain diamonds. Although there is much discussion still on the nature of the lithosphere/asthenosphere boundary (Anderson, 1995), for the modelling described here we assume a thermal boundary (1300°C) related to the peridotite solidus. This simplifying assumption does not consider the complications introduced by the effects of water, carbon dioxide, and increasing pressure on the solidus. Nor does it address compositional variations in the mantle.

Thermal data for the Slave Province are limited. Prior to Thompson et al. (1995b), the only published measurements of heat flow, heat production and thermal conductivity in the craton were from a single site near Yellowknife (Lewis and Wang, 1992). The heat flow of 50 $mW/m^2$ is somewhat higher than the average (42 $mW/m^2$) for the Archean Superior Province (Drury, 1991).

Charles Fipke's discovery of diamond-bearing kimberlites just east of the Winter Lake - Lac de Gras map area (Fig. 1) indicates that the lithosphere was at least 150 km thick 52 Ma ago (Northern Miner, 1993) at the time of intrusion of kimberlite pipes. On the other hand, Thompson et al.(1995a) suggested that high geothermal gradients evident from metamorphic rocks in the Winter Lake-Lac de Gras map area preclude formation of diamonds beneath the Slave Province during orogenesis at 2.6 Ga. Furthermore, numerical thermal models with a reasonable fit to Late Archean geological history in the Slave Province (Thompson and Lucazeau,1989) start and end with a thermal lithosphere 60 to 70 km thick in the time period 2.7 to 2.4 Ga. Preliminary calculations based on our new thermal data (Thompson et al., 1995b) support the hypothesis that, in this case at least, diamonds are not "forever" and may have formed no earlier than the Mesoproterozoic. Moreover, our model which considers the role of spatial and temporal variations in crustal heat production suggests that the thickness and temperature of the lithosphere within the diamond stability field varies with surface heat flow, and hence diamonds may not be distributed equally everywhere beneath the Slave Province. The concept of a strong relationship between surface heat flow, crustal heat production and temperature with depth in the continental lithosphere, is not new (Pollack and Chapman, 1977; Pollack et al., 1993; Stein, 1995). In this paper we extend the concept to embrace changes in temperature with time in a stable craton due to a natural reduction in heat production in rocks.

## GEOLOGICAL SETTING

The Slave Province of the Canadian Shield is distinguished from classic Archean "granite-greenstone" terranes by the high proportion of sedimentary rocks in the supracrustal sequence, abundant highly evolved granitoid rocks, and the limited occurrence of ultramafic igneous rocks (McGlynn and Henderson, 1972; Henderson, 1981; Padgham,1990). Located 250 km north of Yellowknife, the Winter Lake-Lac de Gras map area (11 000 $km^2$, Thompson et al.,1995a; Thompson and Kerswill, 1994) is typical of central Slave Province geology, containing parts of two major lithotectonic elements. Thin discontinuous volcanic belts (2.7 Ga; Villeneuve, 1993) separate a domain dominated by recrystallized, migmatitic and gneissic granitoid rocks from one dominated by metamorphosed greywacke-mudstone. The heterogeneous granitoid rocks are interpreted to be, for the most part, older than the

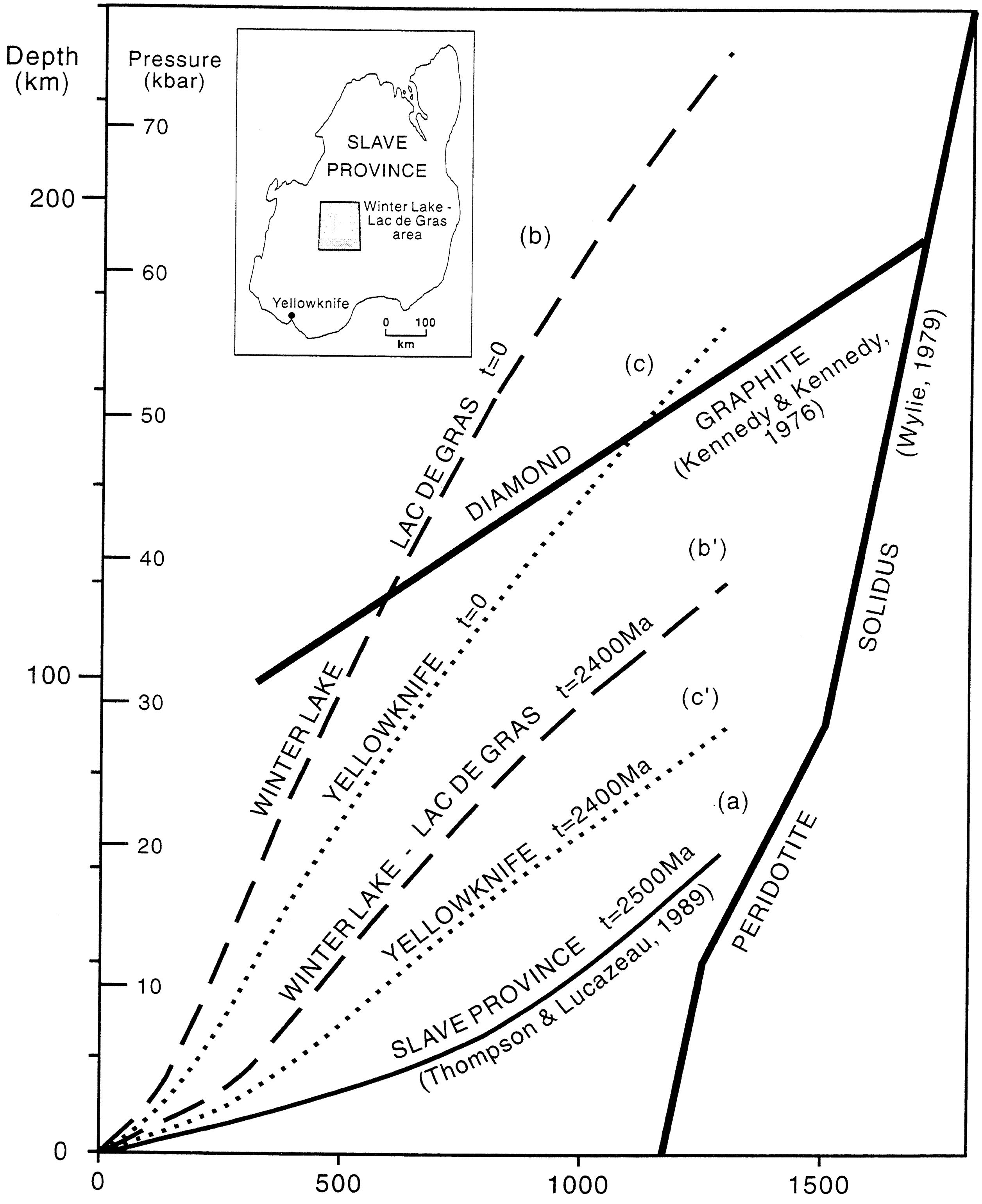


***Figure 1:*** *Pressure(depth)-Temperature diagram illustrating calculated geothermal gradients, stability field for diamond (Kennedy and Kennedy, 1976), and peridotite solidus (Wyllie, 1979). Gradient* ***(c)*** *is based on the data at a single site near Yellowknife where surface heat flow was measured. Gradient* ***(b)*** *represents an average derived from data from 69 sample sites distributed across the Winter Lake - Lac de Gras map area (11 000 km²) where as yet no heat flow measurements have been made. In the range 0 to 900 °C, geothermal gradient* ***(a)*** *from Thompson and Lucazeau (1989) is similar to that derived from metamorphic data by Thompson et al.(1995a) time = 2600 Ma. Inset indicates location of study area in the Slave Province, northwestern Canadian Shield.*

supracrustal sequence (Yellowknife Supergroup of Henderson, 1970). Younger granitoids (2.6 Ga) intrude both domains. Diamond-bearing kimberlites occur close to the map area, with the major Lac de Gras field lying just to the east (Pell, 1995; Kjarsgaard, 1996).

## THERMAL PROPERTIES OF SLAVE PROVINCE ROCKS

As part of the Slave Province NATMAP project (Thompson et al., 1995a), additional rock samples were collected for measurement of thermal conductivity and radiogenic heat production. The new data are presented together with a summary of laboratory methods in Thompson et al. (1995b). Thermal conductivity ranges from 2.3 W/m•K in mafic rocks to 4.9 W/m•K in granitic rocks. Conductivity parallel to planar foliation or layering is up to 1.6 times higher than that perpendicular to these features. At the surface, dips of planar structures are generally >45°. Radiogenic heat production varies with rock type from 0.1 to 0.8 $\mu W/m^3$ in mafic metavolcanic rocks and tonalitic gneiss to 8.0 to 15.8 $\mu W/m^3$ in the youngest granites. Metasedimentary rocks, including twenty samples from other sedimentary domains in the Slave Province, are relatively low (average = 1.2 $\mu W/m^3$), as are the main components of the older granitoid suite (0.4 to 2.0 $\mu W/m^3$) that is interpreted as basement beneath the supracrustal rocks. Uncertainties of less than plus or minus five percent are associated with heat production and thermal conductivity measurements. Radiogenic heat production in rocks decreases with time through a natural decay of the radioelements (Van Schmus, 1995). At the time of intrusion of "younger" granitoids in the Slave Province (2620-2580 Ma; van Breemen et al., 1992), heat production of each of the rock units was almost double present day measurements.

### *Present thermal setting*

Diamonds tend to occur in tectonically stable Precambrian shield areas, primarily Archean cratons, where the lithosphere at present is relatively thick (>150 km), and geothermal gradients low (Nyblade and Pollack, 1993; Gurney, 1989). Modern studies using seismic tomography confirm the presence of lithosphere 200 to 250 km thick beneath areas of Precambrian shield such as the Slave Province (Silver and Chan, 1988; Montagner, 1994). In the absence of nonconductive heat transfer within the lithosphere, the increase of temperature with depth in the lithosphere is a function of its thermal conductivity, heat derived from radioactive decay within the lithosphere, and heat coming from the underlying asthenosphere. Higher heat flow at the Earth's surface corresponds to a higher geothermal gradient; that is, a larger temperature change through the lithosphere, and often accompanies a thin or thinner lithosphere (Pollack and Chapman, 1977; Stein and Stein, 1992).

Thompson et al. (1995b) calculated present day temperature profiles with depth below the surface for a site near Yellowknife (gradient **c**, Fig. 1) and the Winter Lake-Lac de Gras area (gradient **b**, Fig. 1) using the following equation and assuming purely conductive heat transfer within the lithosphere and crust;

$$T = T_o + qz/k - Az^2/2k$$

T = temperature at depth z, $T_o$ = surface temperature

q = surface geothermal flux (summation of crust and mantle contributions)

k = thermal conductivity

A = radiogenic heat production (per unit time per unit volume)

The model parameters derived from and discussed in Thompson et al. (1995b) are summarized in Table 1. A thermal conductivity of 3.5 W/m•K was calculated for the surface rocks. Dependence of conductivity on temperature takes the form, $k = k(T{=}0°)/(1 + cT)$, where T is temperature and c = 0.002 for the crust and 0.001 elsewhere. The base of the lithosphere is assumed to correspond to the 1300°C isotherm (i.e., lithosphere/asthenosphere boundary is thermal in nature), the crust to be 35 km thick (Barr, 1971; Braile et al., 1989), and values for heat production within the more mafic middle and lower crust and the ultramafic upper mantle taken as 1.0, 0.15, and 0 $\mu W/m^3$, respectively. Recent heat flow and heat production measurements in the Basin and Range Province of the U.S. (Lachenbruch et al., 1994) appear to confirm the assumption that radioactivity of rocks decreases with depth, at least within the crust. Upper, middle, and lower crust are assumed to be 10, 10, and 15 km thick. Heat production (2.3 $\mu W/m^3$) inferred for the upper crust from granodiorite sampled at the single Yellowknife site (measured surface heat flow, 50 $mW/m^2$), is probably higher than the average for the region because metasedimentary rocks east of Yellowknife (Thompson et al., 1995b) yielded half that value. The relative proportions of rock units across 11000 $km^2$ on the present erosion surface and the average heat production of each unit were used to derive the average value (1.7 $\mu W/m^3$; Thompson et al., 1995b) for upper crust in the Winter Lake-Lac de Gras area. Since

**Table 1.** Model parameters estimated and assumed by Thompson et al. (1995b).

| Parameters | Yellowknife | | Winter Lake-Lac de Gras | |
|---|---|---|---|---|
| | t = 0 | t = 2400 Ma | t = 0 | t = 2400 Ma |
| Thickness (km) | | | | |
| upper crust | 10 | 10 | 10 | 10 |
| middle crust | 10 | 10 | 10 | 10 |
| lower crust | 10 | 15 | 15 | 15 |
| Heat Production ($\mu W/m^3$) | | | | |
| upper crust | 2.3 | 4.2 | 1.7 | 3.0 |
| middle crust | 1.0 | 2.0 | 1.0 | 2.0 |
| lower crust | 0.15 | 0.30 | 0.15 | 0.30 |
| crustal average | 1.0 | 1.9 | 0.8 | 1.6 |
| lithospheric mantle | 0 | 0 | 0 | 0 |
| Thermal Conductivity surface rocks (W/m•K)* | 3.5 | 3.5 | 3.5 | 3.5 |
| Upper boundary, top of crust (°C) | 0 | 0 | 0 | 0 |
| Lower boundary base of lithosphere (°C) | 1300 | 1300 | 1300 | 1300 |
| Surface heat flow ($mW/m^2$) | 50 | 91 | 40 | 73 |
| Derived mantle heat flow (base of lithosphere) | 13 | 21 | 11 | 18 |

* conductivity (k) = k(T=0°)/(1 + cT), where T = temperature and c = .002 for crust and c = .001 for mantle.

heat production in crustal rocks is a major component of surface heat flow, a lower surface heat flow value of 40 $mW/m^2$, more typical of Archean shield, was assumed. Whereas the temperature profile below the present surface calculated for Yellowknife is derived from data at a single point, the profile for the Winter Lake-Lac de Gras area represents an average obtained from ten rock units with a range of heat production and conductivity data. The higher heat production at the Yellowknife site is within the range measured for granitoid rocks of the Winter Lake-Lac de Gras area (Thompson et al., 1995b).

These preliminary results showing how present day average crustal heat production and, hence, surface heat flow may vary across the Slave Province, are consistent with anomalies tens to hundreds of kilometres wide that were mapped by Darnley et al. (1986) using airborne methods. Our model illustrates how such changes could affect the thickness of lithosphere within the diamond stability field. In other words, the probability of a kimberlite picking up diamonds on its way through the lithosphere may not, if our model is correct, be everywhere the same.

## THERMAL EVOLUTION OF THE SLAVE LITHOSPHERE

Past geothermal gradients can be reconstructed from a combination of measurements of the present thermal parameters and determination of the metamorphic history of rocks (Richardson, 1970; Thompson, 1977; Royden et al., 1980). Typical of low pressure metamorphism in the Slave Province (Thompson et al., 1995a), geothermal gradient **(a)**, Figure 1, is much higher than the present values, implying a lithosphere too thin for diamond stability. It represents the thermal regime Thompson and Lucazeau (1989) derived from numerical experiments for 2500 Ma, just after the main orogenic event in the Slave Province. Presumed to have developed in thinning continental lithosphere during basin formation, the high gradients necessary to explain regional low pressure metamorphism and generation of syntectonic magmatism are interpreted to have persisted during moderate crustal overthickening and into the early stage of exhumation of the crust back to normal thickness (Thompson,1989). The numerical modelling best represented the geological record when the initial lithospheric thickness (1300 °C) was less than 100 km. Such thin warm lithosphere precludes formation of diamonds beneath the Slave Province in the Late Archean (Fig. 1). Thin lithosphere at 2.6 Ga is also required by a hypothesis of lithospheric delamination, an alternative proposal to explain late granite magmatism and metamorphism in the Slave Province (Davis et al., 1994). Thin lithosphere, persisting from the Archean into the Proterozoic, is further supported by Grotzinger and Royden (1990) who argued the case on the basis of a flexural analysis of the Proterozoic Kilohigok sedimentary basin in the northern Slave Province. Hoffman (1990) pointed out that these results are in direct disagreement with arguments supporting an Archean age for deep lithospheric roots . He proposed other basins as a test for the Grotzinger and Royden model, and suggested that models for progressive thickening of the lithosphere need consideration. The results of our modelling provide another line of evidence for some lithospheric roots having formed since the end of the Archean.

Geothermal gradients **(b')** and **(c')** calculated for 2400 Ma (Fig. 1) correspond to gradients **(b)** and **(c)** at the present time. The differences are caused solely by the radioactive decay of U, Th, and K. We assume, to set boundary conditions for simple models of thermal conduction in the crust and lithosphere, that the present erosional surface has not changed significantly, which does have geological support (McGlynn and Henderson, 1972; Nassichuk and McIntyre, 1995; 1996), and that uplift and erosion related to Archean orogenesis had ceased 2400 million years ago. Also, we assume that Proterozoic orogenic events around the margins of the craton did not disrupt the thermal regime of the central Slave Province. Although diabase dyke swarms intruded the Slave craton at 2.23, 2.21, 2.02 and 1.27 Ga (LeCheminant et al., 1996), associated thermal effects are interpreted to be localized on the craton margins and to have little overall impact on the lithospheric thermal regime in the central Slave. As time before present increases to 2400 Ma (Fig. 1), the geothermal gradients increase and lithospheric thickness decreases, approaching that associated with gradient **(a)**. Eventually, the geothermal gradients for both areas reach asthenospheric temperatures without intersecting the diamond stability field, thereby suggesting a maximum age for growth of diamond beneath the Slave craton.

Depth-time diagrams (Fig. 2a, 2b) show this time dependence of lithospheric growth based on our proposed model for thermal evolution of the lithosphere over 2400 million years. The model results are very sensitive to crustal heat flow and heat production. This is demonstrated by comparing Figures 2a and 2b using data for Yellowknife and Winter Lake-Lac de Gras with surface heat flows of 50 and 40 $mW/m^2$ (derived mantle flux of 13 and 11 $mW/m^2$), respectively. Calculated values for 2400 Ma are 91 and 73 $mW/m^2$ at the surface (21 and 18 $mW/m^2$ for mantle flux). Derived present day mantle heat flow is comparable to values of 10 to 14 $mW/m^2$ obtained for the Superior Province (Guillou-Frontier et al., 1994).

If the estimated average crustal heat production and surface heat flow are at the upper end of craton range as at the Yellowknife locality (Fig. 2a), the amount of present day lithosphere within the diamond stability field is less than in the Winter Lake-Lac de Gras area (Fig. 2b), thereby reducing the chance of a kimberlite picking up diamonds during ascent. Moreover, the time at which the base of the lithosphere entered the diamond stability field occurs later, implying a younger maximum age for diamond formation. If average heat flow is closer to that assumed for Winter Lake-Lac de Gras and more typical of Archean shield (Fig. 2b), the lithosphere is thicker, and the present stability field of diamond in the lithosphere extends 25 km closer to the surface, and the oldest possible age of diamond growth is ca. 1850 Ma. Our modelling indicates that the decrease in average crustal heat production since the end of the Archean, can account for the thickening that is required to change the thin Archean lithosphere, implied by low pressure/high temperature metamorphism and by numerical models, to one thick enough to contain diamonds by the time kimberlites were intruded at 52 Ma.

## DISCUSSION

While there is widespread agreement, based largely on the interpretation of seismic data, that relatively thick lithospheric roots occur beneath Archean cratons today, how and when they formed and the absolute thickness remain under discussion. For example, while Jordan (1988) derived a thickness of 300 to 400 km from shear wave velocity anomalies, Anderson (1990) estimated magnitudes of 150 to 200 km based on a rapid reduction of shear wave velocities at these depths. Root thickness varies beneath Archean components of the Canadian Shield (Grand, 1987). Polet and Anderson (1995) indicated that variation of root thickness from one craton of similar age to the next may represent "different snapshots in time of transient roots". Moreover, they suggested that "permanent" roots beneath old cratons may be quite small, and that, roots induce cold downwellings in the underlying asthenosphere that increase their apparent depth. Our study indicates that continental crustal heat production could be a factor causing lithospheric thickness variations laterally beneath a craton

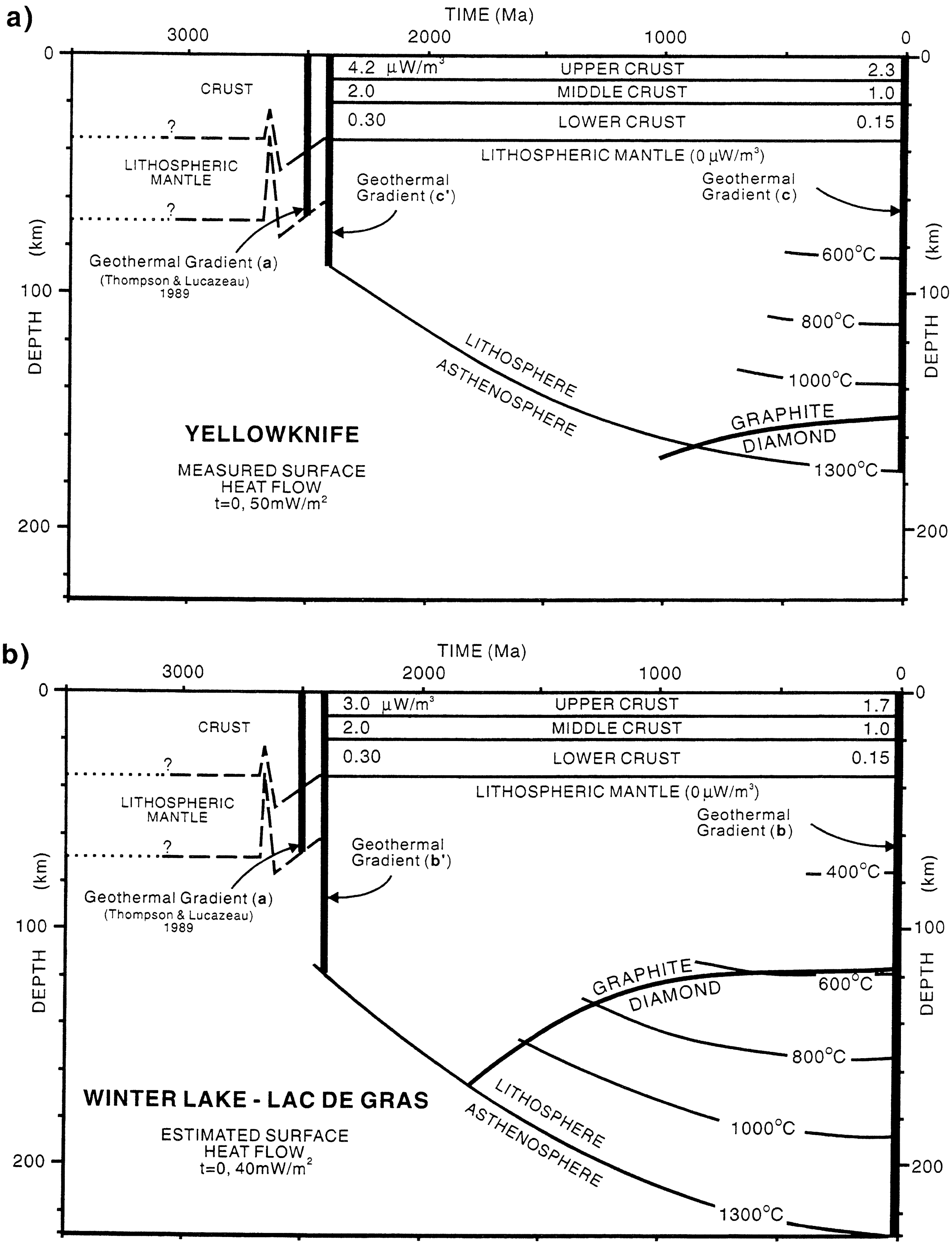


***Figure 2:** Depth-time diagrams representing the thickening of the lithosphere with time as radiogenic heat production in the crust decreases since the end of the Archean Era, 2500 Ma ago. Thermal evolution between 2700 and 2400 Ma is from Thompson and Lucazeau (1989). Geothermal gradients at time = 0 (today) and for time = 2400 Ma are from Figure 1. Diamond stability field from Kennedy and Kennedy (1976). a) Yellowknife locality; b) Winter Lake - Lac de Gras map area. Although there is a significant discrepancy in lithospheric thickness between the geologically constrained forward modelling and our model calculated back in time, both approaches point to thin lithosphere at the end of the Archean.*

as well as progressive thickening of the lithosphere since the end of the Archean.

Mantle xenoliths found in kimberlites indicate that a significant proportion of the lithospheric mantle beneath Archean cratons is depleted in basaltic components relative to the asthenosphere (Ringwood, 1966; Pearson et al., 1995). A lithosphere of this composition would be more refractory and more buoyant than adjacent asthenosphere and its seismic velocity would be relatively high. Jordan (1988) assumed that the full extent of the seismic anomaly, down to depths of 300 to 400 km, represented chemically-depleted lithosphere. Polet and Anderson (1995) inferred, in some cases at least, a chemical root of 150 to 250 km and attributed the rest of the seismic anomaly to low temperatures in the asthenosphere. Thermobarometric data (Finnerty, 1989; Carlson, et al., 1994) indicating that mantle xenoliths are derived from maximum depths of 150 to 200 km perhaps support the latter interpretation.

While workers such as Jordan (1988) proposed formation of the "deep" lithospheric root at the same time as the overlying Archean crust and in the course of an extensive partial melting event in the mantle, or during periods of compressive orogenesis, others have suggested a progressive thickening by underplating of subducted oceanic lithosphere (Helmstaedt and Schulze, 1989; Gurney, 1990; Abbot, 1991). Ringwood (1989) proposed a lithosphere thickened by diapirs of melt rising from megaliths of oceanic lithosphere that had settled to depths of 400 to 500 km. In most cases, mantle xenoliths and inclusions in diamonds from individual kimberlites yield a range of model ages; for example 1.0 to 3.2 Ga in South Africa (Richardson et al., 1990) and 0.4 to 3.2 Ga in Siberia (Pearson et al., 1995). Early formation of a mantle root in the Archean requires a long subsequent history of metasomatic events to explain this wide range of ages. For a root thickening with time, at least some of the age variation may record both the history of accretion and the heterogeneity of the resulting lithospheric mantle.

If the lithosphere thickened slowly downward during the last 2500 million years (Fig. 2), it has incorporated material by basal accretion that was part of the relatively mobile asthenosphere for different lengths of time. Our model implies that the age of the lithosphere should decrease with depth as progressively younger asthenosphere cools and accretes to the lithosphere. That is, lithospheric mantle xenoliths and diamonds found in kimberlites should have a range of ages with higher pressure xenoliths likely to be younger. Furthermore, an accretionary root should reflect the chemical heterogeneity of the mantle (and subducted components within it) through which the Slave craton has moved during the last 2500 million years. In this case, the upper, "permanent" part of the zone of anomalous seismic velocities may reflect both the relatively cool thermal regime indicated by our model and the average composition of a heterogeneous root.

The simple thermal model presented here is obviously not the complete solution to the complex problem of the origin of subcratonic lithosphere which is both thermally and chemically controlled. However, a number of interesting implications arise from this study. Preliminary results based on two modelling approaches, back-modelling the current thermal state and forward modelling of Archean crustal evolution for the period 2.7 to 2.4 Ga are consistent with the lithosphere beneath the Slave Province being considerably thinner 2400 million years ago, perhaps half of what it is now. This suggestion of a thin Archean lithosphere is supported by other arguments (Grotzinger and Royden, 1990; Thompson, 1989; Thompson et al., 1995a). Also, the model implies that formation of diamonds in the lithosphere beneath the Slave Province before that time, appears unlikely, and remains so before ca. 2.0 Ga.

These preliminary results raise a question about the distribution of diamondiferous lithosphere beneath the Slave Province. Is it possible that the heat flow in some parts of the Slave crust is high enough for the lithosphere never to have been thick enough to contain diamonds? If this is the case, kimberlites passing through this thin lithosphere would be barren. Additional measurements of heat production and of heat flow across the entire Slave Province are needed. This can be done relatively inexpensively by using the exploration drill holes, although specialized borehole measurement techniques are required to combat the presence of permafrost in these northern locales (Judge,1973).

A history of low pressure regional metamorphism and abundant magmatism synchronous with deformation of volcano-sedimentary belts similar to that preserved in the Slave Province occurred 500 to 600 million years earlier in the Archean craton of southern Africa (Saggerson and Turner, 1976; Cahen et al., 1984). A useful extension of our study would be to determine if the surface geology and thermal parameters in the Kaapvaal craton are consistent with a lithosphere less than 120 km thick prior to 2.7 to 2.8 Ga as indicated by Helmstaedt and Schulze (1989) or with one that was 225 km thick at that time (Pearson et al., 1995). The current lithosphere beneath the Kaapvaal craton could be modelled, and the results related to diamond distribution and other geological information from the region.

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# Mafic magmatism, mantle roots, and kimberlites in the Slave craton

**A.N. LeCheminant, L.M. Heaman, O. van Breemen, R.E. Ernst, W.R.A. Baragar, and K.L.Buchan**

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## INTRODUCTION

The lithosphere is thicker beneath Archean cratons, and contains cool lithospheric mantle roots that extend into the diamond stability field (Durrheim and Mooney, 1994; Polet and Anderson, 1995). Periodically, mafic magmas intrude the cratons to form extensive flood basalts, layered intrusions, and diabase dyke swarms. Such large magmatic outbursts can be generated by hot mantle plumes, causing rifting and continental break-up (White and McKenzie, 1989; Storey et al., 1992; Coffin and Eldholm, 1994), or can be caused by small-scale convection induced by rifting, particularly at the edges of Archean cratons (Anderson, 1994a, b; King and Anderson, 1994). Thermal and magmatic events generated by mantle plumes and/or rifting can modify the diamond-bearing roots of continental lithosphere (Hoffman 1990a; Helmstaedt, 1993, 1995; Helmstaedt and Gurney, 1994); the ages of diabase dyke swarms precisely date the magmatic events. The global database for dyke swarms is improving rapidly (Ernst et al., 1996); applications of recent dyke studies include locating mantle plumes (LeCheminant and Heaman, 1989; Heaman et al., 1992; Zhao et al., 1994; Park et al., 1995; Ernst et al., 1995a, b) and establishing accurate time markers for regional correlation and paleocontinental reconstructions (Heaman and LeCheminant, 1993; Buchan et al., 1993, 1994).

Mapping, U-Pb geochronology and paleomagnetism indicate there are at least five major Proterozoic diabase dyke swarms in the central Slave Province near Lac de Gras (Fahrig and West, 1986; LeCheminant, 1994). The dykes are not deformed and range in age from 1.27 to 2.23 Ga (Fig. 1, 2). Other dyke swarms (not yet dated by U-Pb) intrude southwestern parts of the Slave Province (Fig. 1), and two younger dyke swarms penetrate the northern and western margins of the craton at 0.78 and 0.72 Ga (Heaman et al., 1992; LeCheminant and Heaman, 1994). The dykes are a surface record of Proterozoic thermal events, and their age, source, and distribution provide evidence about possible destruction or selective preservation of mantle roots beneath the Slave Province. In the Phanerozoic, kimberlites carried samples from deep crustal and mantle sources to surface, possibly providing a direct means of studying the post-Archean history of the craton and determining the effects of Proterozoic magmatism and rifting. Diabase dyke swarms in the central Slave Province are much older than the Eocene diamond-bearing kimberlites at Lac de Gras (Kjarsgaard, 1996; Davis et al., 1996). However, fracture sets along dyke contacts provided local zones of weakness for kimberlite emplacement (LeCheminant, 1994). Ages of newly discovered kimberlites in three other areas of the Slave craton (Jennings and Barker, 1995) have not been announced.

## *PALEOPROTEROZOIC DYKE SWARMS NEAR LAC DE GRAS: BREAK-UP OF AN ARCHEAN CRATON?*

The three oldest swarms in the central Slave Province, the Malley, MacKay, and Lac de Gras swarms, record events which may mark the progressive break-up of a large Archean craton (LeCheminant and van Breemen, 1994). The NE-striking Malley diabase dykes intrude Archean basement, but do not cut 1.90 to 1.97 Ga sedimentary rocks of the Goulburn Supergroup, Kilohigok basin (Fig. 1). U-Pb dates indicate Malley dykes were emplaced at 2.23 Ga, and are about 20 million years older than the MacKay diabase, a widely-spaced east-striking dyke swarm injected at 2.21 Ga. These two ages may indicate the timing of specific events related to rifting and break-up along the eastern and southern margins of the Slave Province. Interestingly, the 2.23 Ga Malley dykes are coeval with dyke swarms on other small Archean cratons, such as the 2235 Ma Kikkertavak dykes in the Hopedale block, Labrador (Cadman et al., 1993) and 2.24 Ga dykes in the Vestfold Hills, an Archean cratonic block within the East Antarctic Shield (Lanyon et al., 1993). The younger 2.21 Ga MacKay dykes are similar in age to 2.21 to 2.22 Ga Nipissing sills and Senneterre dykes of the southern Superior Province (Corfu and Andrews, 1986; Noble and Lightfoot, 1992; Buchan et al., 1993).

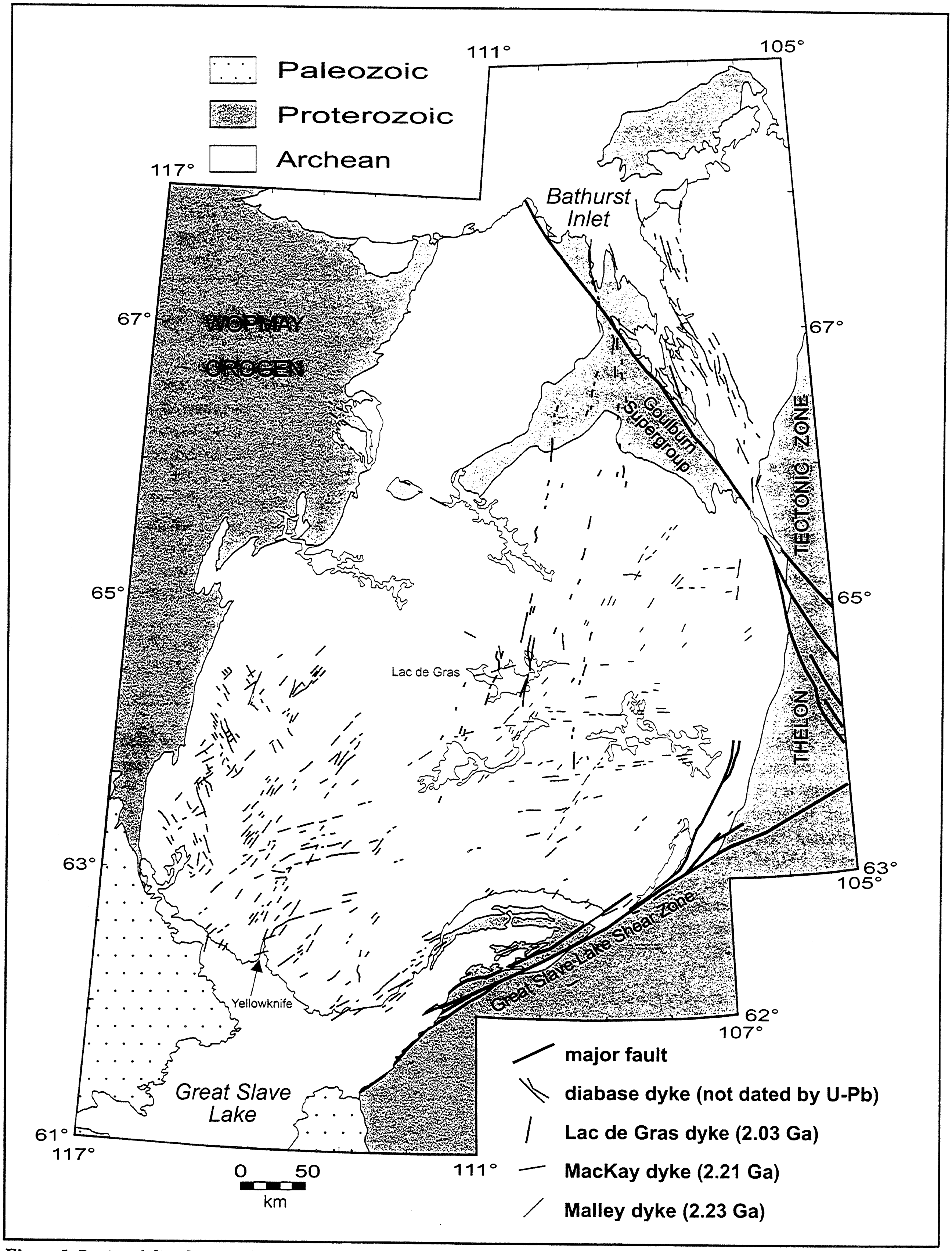


***Figure 1.*** *Regional distribution of Paleoproterozoic diabase dyke swarms in the Slave Province (modified after Fahrig and West, 1986).*

The 010°-striking Lac de Gras swarm consists of >10 dykes spaced over a width of up to 100 km (LeCheminant, 1994). The dykes converge to the north and extend more than 300 km from Lac de Gras to a focus beneath Kilohigok basin. U-Pb ages for two dykes of 2023 and 2030 Ma (LeCheminant and van Breemen, 1994) are slightly younger than the 2038 Ma age of a NE-striking Hearne dyke, East Arm, Great Slave Lake (Pehrsson et al., 1993). The Lac de Gras and Hearne dykes were emplaced shortly before collisional events along the eastern margin of the Slave Province that formed the 2.02 to 1.91 Ga Thelon tectonic zone (Thompson, 1989; Henderson et al., 1990), and the 1.98 to 1.92 Ga Great Slave Lake shear zone (Hanmer et al., 1992). However, the orientation, focus, and age of the Lac de Gras swarm suggest dyke emplacement is not related to the collisional events. Rather, magma was injected laterally, from the north, into the central Slave Province, from a source region beneath Kilohigok basin and near the coeval 2023 Ma Booth River intrusive suite (Roscoe et al., 1987). A pyroclastic unit, interpreted as erupted during the rift to drift transition on the western margin of the Slave Province, has a similar U-Pb age of about 2.02 Ga (S.A. Bowring, pers. comm., 1993). Therefore, emplacement of the Booth River suite and injection of the Lac de Gras dykes occurred at the same time as rifting along the western Slave margin. From about 1.97 to 1.90 Ga, the Kilohigok basin developed on thermally weakened lithosphere above the source region for the Lac de Gras dykes, synchronous with formation of the west-facing Coronation margin (Bowring and Grotzinger, 1992), and with collisional events in the Thelon tectonic zone.

## MESO- AND NEOPROTEROZOIC IGNEOUS EVENTS: MANTLE PLUMES AND RIFTING

The 1.27 Ga Mackenzie dyke swarm is the only large-scale igneous event to affect the entire Slave Province subsequent to the 2.0 to 1.8 Ga collisions that welded the Rae Province and Hottah terrane onto the Slave craton. The Mackenzie swarm is the largest known radiating dyke swarm on Earth (Ernst et al., 1995a, b); dykes are exposed for more than 2400 km along strike and form a radial array with >100° of arc (Fig. 2). They were emplaced laterally from a mantle plume-generated focal point (LeCheminant and Heaman, 1989; Ernst and Baragar, 1992) more than 700 km northwest of the Lac de Gras kimberlite field. U-Pb studies have determined the precise timing and short duration of the Mackenzie events (LeCheminant and Heaman, 1989; Heaman and LeCheminant, 1993). Ages are remarkably consistent, with dykes, thick sills, and the Muskox intrusion emplaced between 1267 and 1270 Ma, indicating an exceptionally short duration for the emplacement of large volumes of magma (>230 000 $km^3$).

Major gravity anomalies are linked to the Mackenzie events. North of the Coppermine basalts, five large 80 to 100 mGal anomalies form a U-shaped array along the outer edge of a regional gravity high enclosing the focal point of the dyke swarm (Fig. 3A). The anomalies suggest that large mafic intrusions, coeval with the Muskox intrusion, or thick flood basalt sequences underlie the younger cover to the north. The line enclosing the gravity anomalies marks the steepest gravity gradients and likely defines the region where the crust was extensively intruded and underplated by mafic igneous rocks at 1.27 Ga.

Two additional magmatic flareups, the 0.78 Ga Gunbarrel and the 0.72 Ga Franklin events, generated extensive diabase dykes, sills, and sheets along the northern and western margins of the Canadian Shield, but these intrusions only penetrated the outer edges of the Slave Province. LeCheminant and Heaman (1994) investigated the timing of widespread mafic igneous activity in the northwestern Canadian Shield, the northern Cordillera, and the Wyoming craton. U-Pb ages determined for gabbro sheets from south of Great Bear Lake, for a sill in the Mackenzie Mountains, and for dykes from British Columbia and Wyoming, are synchronous at 0.78 Ga. This short-lived magmatic flareup, named the Gunbarrel igneous event, is an excellent time-marker extending >2400 km from the western edge of the Slave craton west to the Mackenzie Mountains and south to the Wyoming craton. The magmatism signals a key event during formation of the western rift margin of North America and provides a precise age to test proposed Neoproterozoic connections between North America and East Antarctica-Australia (Moores, 1991; Hoffman, 1991). Park et al. (1995) suggested rifting was induced by a mantle plume and proposed that the same plume gave rise to ca 0.8 Ga magmatism in south-central Australia (Zhao et al., 1994).

The 0.78 Ga Gunbarrel event along the western margin of North America occurred about 60 million years before the abrupt onset of voluminous Franklin magmatism in the northwestern Canadian Shield. U-Pb ages for gabbro sills from Victoria Island, Coronation Gulf, and Bathurst Inlet, and for a diabase dyke from Baffin Island suggest that most of the igneous activity occurred at 723+4/-2 Ma (Heaman et al., 1992). The short-duration, large-volume, and mafic composition of the dykes and sills and their

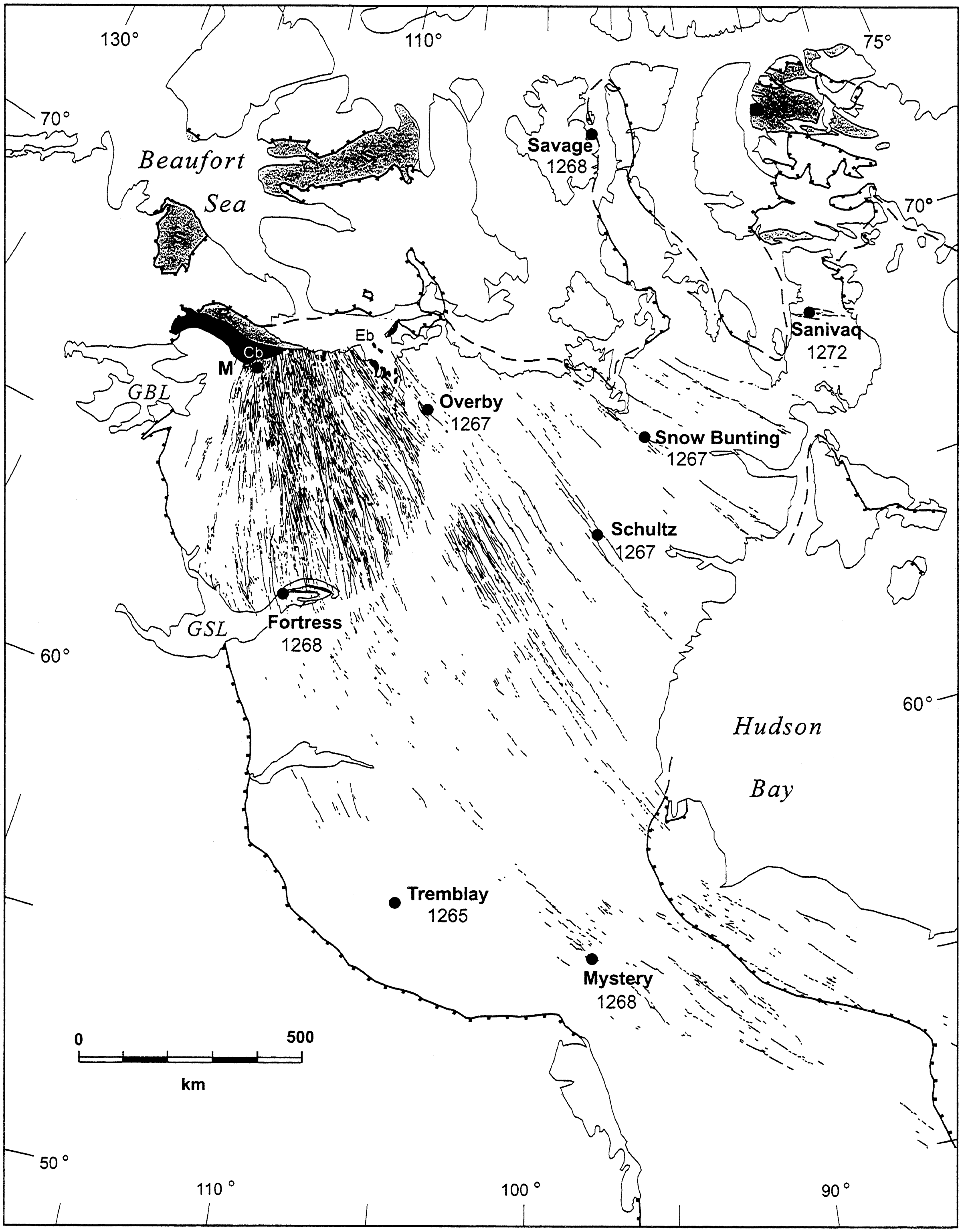


*Figure 2. Distribution of Mackenzie diabase dykes, gabbro sills and related flood basalts in northwestern Canada (after Fahrig and West, 1986). M = Muskox intrusion; Cb = Coppermine River basalts; Eb = Ekululia basalts; Nb = Nauyat basalts; R = Rae Group; S = Shaler Supergroup; GBL = Great Bear Lake; GSL = Great Slave Lake. Ages shown are average $^{207}Pb/^{206}Pb$ ages (Ma) for fractions that are less than 1% discordant (from Heaman and LeCheminant, 1993).*

association with the Natkusiak flood basalts suggest the magmatism occurred above a mantle plume. Stratigraphic relationships on Victoria Island indicate that differential uplift, interpreted as due to doming of the lithosphere above the upwelling plume, occurred just before the outbreak of igneous activity (Rainbird, 1993).

## DISCUSSION

Constraints on the origin of the cratonic mantle beneath the Canadian Shield were discussed by Hoffman (1990a), who concluded, based primarily on isotopic and seismic evidence, that mantle roots formed at about the same time as the overlying crust. Depleted Archean mantle roots survive by being more refractory and less dense than the convecting asthenosphere (Durrheim and Mooney, 1994; Polet and Anderson, 1995). Recent tomographic studies provide better resolution of the variation in mantle velocity beneath shields and support models suggesting that many old continental regions have roots extending to depths of 250 to 450 km (Grand, 1994; Polet and Anderson, 1995; Zhang and Tanimoto, 1993). Mafic magmatism triggered by mantle plumes, rifting, and continental break-up causes local destruction of the lithosphere by thermomechanical erosion (Davies, 1994), and/or by convective upwelling induced by pull-apart of the lithosphere (Anderson, 1994b).

The diabase dyke record in the Slave Province suggests it is a fragment of a larger Archean craton that survived repeated thermal and tectonic disturbances during the Proterozoic. The Malley and MacKay dykes likely were injected laterally into stable central areas of the Slave craton from rifted and thinned source regions that were subsequently destroyed by the collisional events that formed the Thelon tectonic zone. In the eastern Slave Province, both dyke swarms are metamorphosed, with metamorphic grade progressively increasing eastward towards the Thelon tectonic zone (Thompson, 1989; Henderson et al., 1990). The dykes provide a sensitive measure of the extent and metamorphic grade of the Paleoproterozoic reworking of the eastern edge of the Slave craton.

The orientation and focus of the Lac de Gras dykes suggest that magmas penetrated the Slave craton at 2.02 Ga from a source beneath Kilohigok basin. Flexural analysis of Kilohigok basin based on sedimentary sequences deposited between 1.97 and 1.92 Ga produced the unexpected result that the elastic strength of the underlying Slave craton was very low (Grotzinger and Royden, 1990), leading them to suggest that a thick continental root beneath this part of the Canadian Shield had not yet formed. This interpretation is, apparently, at odds with the view that thick cold roots existed beneath Archean cratons early in their history (Hoffman, 1990b). However, the low elastic strength is consistent with development of the basin above lithosphere thinned and thermally weakened by magmas feeding into the Booth River intrusive suite and the Lac de Gras dykes.

Collisional events between 2.0 and1.8 Ga along the margins of the Slave craton created the flanking Thelon tectonic zone and Wopmay orogen (St Onge and Lucas, 1996). Subsequently, the 1.27 Ga Mackenzie events may have substantially modified the mantle root beneath the northern Slave Province. A large starting plume (Campbell and Griffiths, 1990; Hill, 1991) explains both the sudden onset and short duration of the large-scale Mackenzie magmatism. Figure 3A shows a model for initiation of Mackenzie magmatism by an upwelling mantle plume, followed by rifting across the plume head. The radial pattern of the swarm suggests that the lithospheric tension necessary for dyke emplacement was caused by the plume head itself. Ernst and Baragar (1992) determined magma flow trajectories in Mackenzie dykes and showed that flow is mainly vertical near the focal point and becomes horizontal about 500 to 600 km from the centre (Fig. 3B). Hoffman (1990a) pointed out that mantle roots are thermally eroded by plumes, an idea effectively developed by Helmstaedt (1993, 1995) to distinguish "mantle-root-destructive" structures unfavourable for diamond preservation (mantle plumes, rifts, collision zones) from favourable "mantle-root-friendly" structures (dyke swarms, thin-skinned thrust belts). Helmstaedt (1993) suggested that the Mackenzie plume eroded the mantle root up to 500 km from the plume centre, but left the root intact elsewhere in the Slave Province (Fig. 3B).

The radial distribution of Mackenzie dykes and localization of flood basalts near the focal point suggests that magma generation at 1.27 Ga occurred directly over the axis of an underlying mantle plume. This may not be a general case for plume-generated magmatism beneath continents, provided the lithosphere is inhomogeneous and/or has been previously thinned nearby (Thompson and Gibson, 1991). For example, crustal thinning and intrusion of mafic magmas during the Mackenzie events created an anomalous crustal region, outlined by the U-shaped array of gravity anomalies, that may have controlled later Shaler Supergroup sedimentation and determined the surface location of the 0.72 Ga Franklin magmatic events. Franklin dykes, coeval sills, and flood basalts are spatially related to the anomalous region near

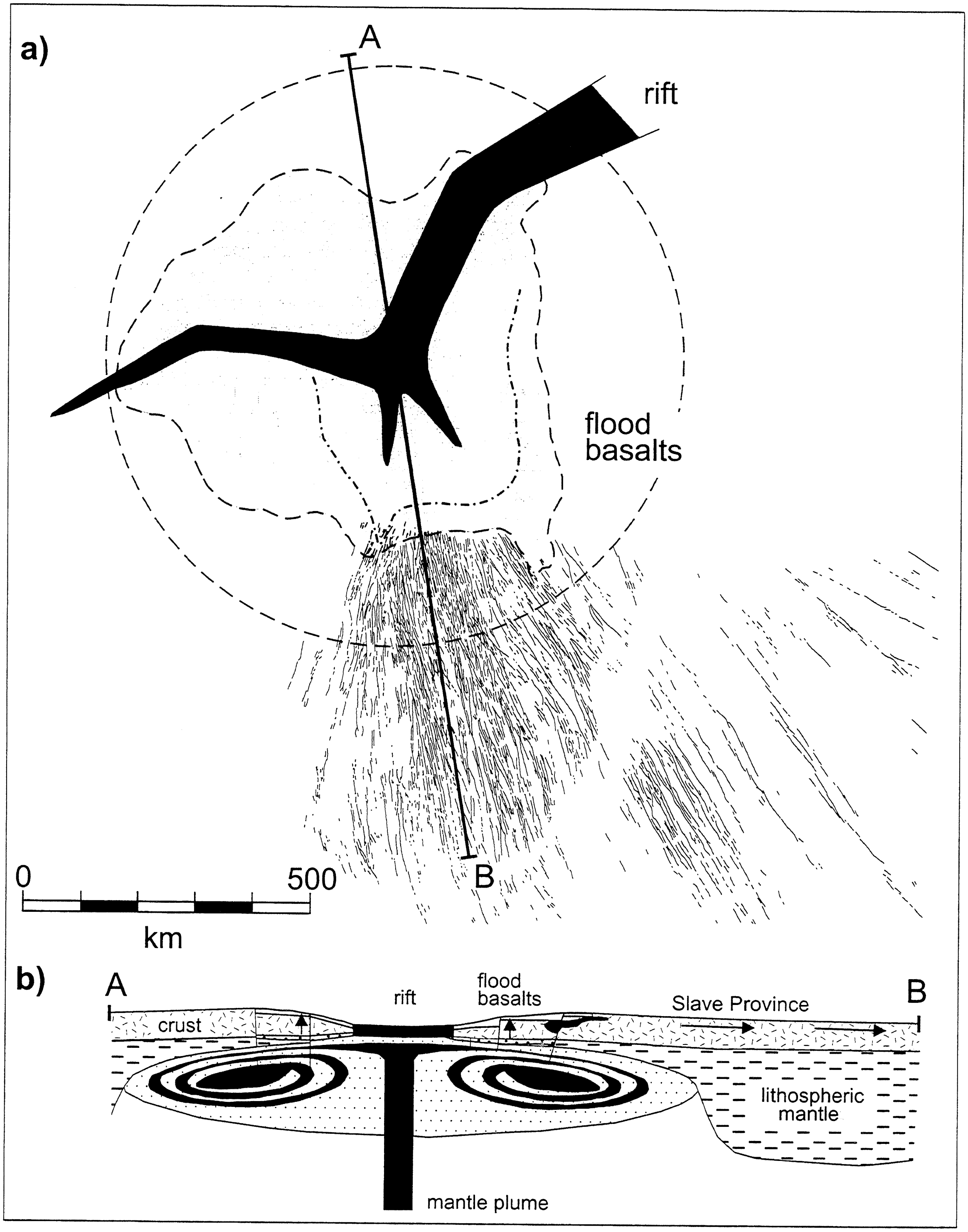


*Figure 3. a) Model for initiation of 1.27 Ga Mackenzie magmatism due to rift propagation across a domal mantle plume (LeCheminant and Heaman, 1989). The dash-dot line defines the outer limits of an U-shaped array of large gravity anomalies. The circle, 500 km from the focal point of the dyke swarm, is the proposed outer boundary of the mantle plume (Ernst and Baragar ,1992). Line A-B is the section line. b) Schematic cross-section through the Mackenzie plume, indicating erosion of the mantle root beneath the Slave craton (modified after Helmstaedt, 1993). The mantle plume is shown as a "starting plume" (Campbell and Griffiths, 1990; Hill, 1991) consisting of hot source material (black) and entrained mantle (dot pattern). Near-surface spreading of the rising plume has resulted in thermal and mechanical erosion of the lithospheric mantle and crust. Arrows indicate magma flow directions in the Mackenzie dyke swarm (Ernst and Baragar, 1992). A layered intrusion (black) is shown in the crust beneath overlying flood basalts, and coeval mafic rocks (heavy dots) are interpreted to underplate the crust.*

the focus of the Mackenzie events and may be displaced hundreds of kilometres from the axis of the Franklin plume (Rainbird, 1993). In any case, both the 0.78 Ga Gunbarrel and the 0.72 Ga Franklin events are peripheral to the Slave craton and, therefore, associated thermal and rifting events are likely to have had little effect on the Slave lithosphere.

The discovery of numerous diamond-bearing Eocene kimberlites in the central Slave Province indicates that the craton is anchored by a cool, deep-seated mantle root of uncertain age. This root may have stabilized in the Archean, shortly after crust formation (Hoffman, 1990a), or may not have formed until after 1.9 Ga (Grotzinger and Royden, 1990; Hoffman, 1990b). Thompson et al. (1996) suggested that diamond-bearing lithosphere was not present before ca. 1.5 to 2.0 Ga. The emplacement of dyke swarms into the Slave craton between 2.23 and 1.27 Ga probably had little effect on the lithosphere except near magma source regions. Importantly, mantle and crustal xenoliths transported from depth by the Lac de Gras kimberlites will provide a direct means of determining the age of the lithosphere and the extent of disturbance caused by Proterozoic mafic magmatism. Probable kimberlite discoveries near Takiyuiak Lake (Jennings and Barker, 1995) and expected discoveries in central Victoria Island (Northern Miner, 1995) are closer to source regions for the Mackenzie and Franklin magmatism and could provide important samples for comparison to mantle and deep crustal xenoliths carried by Lac de Gras kimberlites.

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# Diamonds associated with ultramafic complexes and derived placers

**A.N. LeCheminant and J. H. Bédard**

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## INTRODUCTION

Diamonds in orogenic belts, apparently unrelated to kimberlites or lamproites, have been reported for many years. Although many of the occurrences are poorly documented, numerous accounts link diamonds with platinum group minerals (PGM) derived from ophiolites and zoned ultramafic complexes (Janse, 1994). Tectonized ultramafic massifs within alpine orogenic belts and zoned Alaskan-type complexes can contain chromitite lenses with disseminated PGM (Cabri, 1981; Nixon et al., 1990), and rare diamonds and other high-pressure minerals have been recovered from PGM placers derived by erosion of ultramafic rocks (Janse, 1994; Bai et al., 1993). The peridotites, dunites and pyroxenites that make up the complexes typically occur in Phanerozoic fold belts far removed from Archean cratons and, in most cases, kimberlite or lamproite sources have not been identified for diamonds in the associated placers. Although verified in situ occurrences of diamonds in ultramafic rocks are rare (Kaminskiy and Vaganov, 1977; Janse, 1994), abundant graphite pseudomorphs after diamond have been recognized in pyroxenite layers from the Ronda and Beni Bousera peridotite massifs in Spain and Morocco (Nixon et al., 1991; Davies et al., 1993; Pearson et al., 1993). The diamondiferous Pamali breccia in Indonesia is interpreted as a proximal conglomerate derived from the adjacent Bobaris ophiolite, a likely primary source for the diamonds (Bergman et al., 1987).

The origin and emplacement history of many sheared ultramafic complexes is difficult to interpret because the relationship of the tectonized ultramafic slices to host rocks can be obscured by faulting, thrusting, metamorphism and hydrothermal alteration, leading to uncertainty about their age, setting, and pressure-temperature conditions of formation. Potentially diamond-bearing ultramafic rocks could be fragments of oceanic lithosphere that were subducted into the diamond stability field, or could form by high pressure crystallization of melts derived from subducted oceanic rocks (Pearson et al., 1993), or could be tectonically uplifted slivers of continental lithospheric mantle (Van der Wal and Vissers, 1993; Davies and von Blanckenburg, 1995).

## CORDILLERAN ULTRAMAFIC COMPLEXES

In western North America, fault-bounded ultramafic complexes containing PGM-bearing podiform chromitites are exposed along the Cordillera from Alaska to California. Unusual, non-gem quality diamonds weighing up to 33 carats, and characterized by a coarsely crystallized core encrusted by a fine-grained diamond overgrowth, have been recovered from placers in the Klamath Mountains of northern California, where the diamonds are associated with PGM and chromite derived from the Trinity peridotite (Kopf et al., 1990; Hausel, 1994). In adjacent areas of Oregon, diamonds were reported in PGM placers (Woehler, 1869) derived from the Josephine peridotite, a large dismembered ophiolite comprising mainly tectonized harzburgite, and containing minor podiform chromitite bands in serpentized dunite. PGMs have been found in chromitites from both the Josephine massif and the nearby Onion Mountain ophiolite (Stockman and Hlava, 1984). Diamonds are also associated with PGM-bearing ultramafic rocks from Goodnews Bay, Alaska, where two microdiamonds were recovered in the insoluble residue from dissolved platinum alloy nuggets (Mertie, 1976), and a single tiny diamond was discovered in bottom sediments in Goodnews Bay (Hoare and Cobb, 1977).

In Canada, chromitites in the dunite core of the Tulameen complex, British Columbia contain in situ PGM and are the source for PGM placers in the Tulameen-Similkameen river systems (Nixon et al., 1990). Microdiamonds were reportedly recovered from chromitite lenses in Tulameen dunite by a process involving fusion in platinum crucibles (Camsell, 1911a; b; 1913). The tiny diamonds, reported as "the first recorded discovery of diamonds in Canada either in the solid rock or in placer" (Camsell, 1911a), generated a staking rush of short duration (Camsell, 1954). X-ray work in 1949 indicated the Tulameen "microdiamonds" were actually synthetic periclase formed by heating the rock samples (Lang, 1956), an identification now confirmed by re-examination of the original crystals on the scanning electron microscope (Fig. 1).

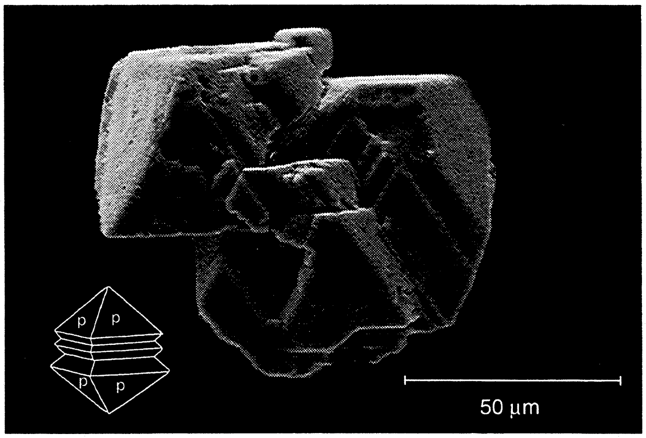


***Figure 1.*** *Backscattered electron image of twinned octahedral crystals identifed as "diamonds, British Columbia" (GSC catalogue no. 65566C). These crystals are periclase, produced during fusion of chromitite lenses in dunite from the Tulameen district (Camsell, 1911b). Similar tiny "diamond" crystals were reported from Thetford Mines chromitite. The white outline sketch, reproduced from Poitevin and Graham (1918), shows the original crystals, which have not been preserved. The twinned octahedral crystals, about 35 μm in height, look identical to the synthetic Tulameen periclase, and were produced by the same fusion process (Dresser, 1913).*

## APPALACHIAN OPHIOLITES

Shortly after the Tulameen "discovery", microdiamonds were identified in serpentinized chromite ore from the Montreal pit of the Dominion Chrome Company, Thetford Mines, Quebec (Johnston, 1912; Dresser, 1913). The chromitite is part of the Thetford Mines ophiolite, a fault-bounded, mafic-ultramafic complex in the southern Quebec Appalachians (Laurent and Hébert, 1989; Pinet and Tremblay, 1995). Samples that contained the tiny, euhedral "diamonds" reported by Johnston (1912) were processed by the same fusion method used on the Tulameen chromitite, and we suggest that the octahedral crystals were also synthetic periclase (Fig. 1). Despite these identification errors, however, both the Thetford Mines and Tulameen ultramafic rocks have similarities to potentially diamond-bearing ultramafic rocks, leading to continued speculation about the early reports of "microdiamonds" (Helmstaedt, 1993; Janse, 1994).

Diamonds have been reported in the central and southern Appalachians from New York to Alabama (Hausel, 1994) and, in several localities, alluvial diamonds were recovered during historical placer gold mining from streams draining ultramafic complexes (Janse, 1994; Hausel, 1994). Recent detailed mapping in the Blue Ridge belt of North Carolina has revealed small blocks of eclogite and altered ultramafic rocks in a 1 km thick shear zone, with mineral assemblages recording minimum pressures up to 17 kbar (Willard and Adams, 1994). The high pressure metamorphic rocks comprise a subduction-related accretionary mélange along the Paleozoic suture between North American rocks (Grenville orogen) and the accreted Piedmont terrane. Pyrope, chromite, ilmenite, and a few diamonds have been recovered in North Carolina southeast of the proposed suture (Kunz, 1887; Hausel, 1994).

Well-preserved mafic-ultramafic complexes, exposed in the Newfoundland and Quebec Appalachians, have been interpreted as dismembered fragments of oceanic lithosphere emplaced by a variable mix of obduction, subduction and collisional processes. The well-known Bay of Islands ophiolite complex in western Newfoundland, originally considered to be a fragment of ocean floor generated at a mid-ocean ridge, is now thought to have been generated in a supra-subduction zone environment (Bédard, 1991; Jenner at al., 1991; Elthon et al., 1991). Ophiolite generation is linked to opening of small ocean basins during early phases of arc-continent collision along the irregular margin of North America, followed shortly afterwards by emplacement of the young ophiolites onto reentrants along the margin (Cawood and Suhr, 1992). Ophiolites in southern Quebec, including the Thetford Mines complex, are interpreted to have formed in an intraoceanic arc environment (Laurent and Hébert, 1989) and been emplaced during arc-continent collision. Again, the irregular shape of the continental margin appears to control the location of well-preserved ophiolites, with re-entrants providing favourable settings for large-scale obduction (Pinet and Tremblay, 1995). In the Appalachains, highly deformed ultramafic remnants in accretionary mélanges near promontories on the continental margin are more likely to have undergone high-pressure, subduction-related metamorphism than are larger intact ophiolite complexes associated with re-entrants.

## PETROGENESIS

Ultramafic complexes have complicated histories involving processes of melt extraction, fractional crystallization, melt/rock reaction, metamorphism, deformation and thrusting (Keleman et al., 1992; Bédard, 1991; 1993). Some, such as the Tulameen complex, are interpreted as differentiated magmatic bodies that underwent high temperature deformation (Findlay, 1969; Nixon et al., 1990). Harzburgite and dunite layers in other tectonically-emplaced complexes are interpreted as 'depleted' mantle peridotites, that crystallized as restites after removal of a melt fraction in the mantle (Dick and Bullen, 1984). An alternative model for the formation of such refractory rocks has been proposed by Keleman et al. (1992), who suggested that dunite and harzburgite in the Trinity peridotite, a possible source of placer diamonds in northern California (Kopf et al., 1990), formed in the upper mantle by extensive interaction between rising basaltic melts and the surrounding mantle peridotite. Many ultramafic complexes preserve mineralogical evidence for formation at high temperatures (Nixon et al., 1990; Suhr, 1993), but pressures are more difficult to constrain. In some cases, pressures can be estimated from reaction curves in P-T space (Van der Wal and Vissers, 1993), or calculated from mineral assemblages in associated metamorphic rocks (Berman, 1996).

Detailed studies of inclusions in chromites from ultramafic complexes suggest a wide variety of PGM and mafic silicate inclusions are trapped as discrete phases at magmatic temperatures (Bird and Bassett, 1980; Stockman and Hlava, 1984; Talkington et al., 1984; Nixon et al., 1990). Dunite and harzburgite in the Trinity peridotite contain trains of Cr-rich spinel where Cr-rich pyroxenes were dissolved in melt/rock reaction zones (Keleman et al., 1992). Unusually high $Cr_2O_3$ contents in chromites requires strongly reducing conditions, and $Cr^{++}$-bearing chromites could contain diamond or graphite inclusions. Probable accessory minerals in peridotites and podiform chromitites in Tibetan ophiolites include diamond, graphite, moissanite (SiC), native metals (Cr, Ni-Fe alloys), and $Cr^{++}$-bearing chromite (Bai et al., 1993), suggesting formation in a high-pressure reducing environment. Diamond, graphite, SiC, Cr, Ni-Fe and PGM alloys may form within chromitites produced by melt/rock reactions that occur under reducing conditions, perhaps from melts originating in metasomatized mantle altered by carbon-rich, reduced fluids from a subducted slab. An interesting analogy is the use of metals to induce growth of synthetic diamonds. The key to successful synthesis is to reach the diamond stability field at P-T conditions where the metal (eg. Fe, Ni, Cr) is in a liquid state - the metal acts as a solvent for carbon and a catalyst for diamond crystallization (Hazen, 1993).

No high pressure minerals have been reported from PGE-bearing chromitites that occur in dismembered ophiolites in the Quebec Appalachians (Corrivaux and LaFlamme, 1990; Gauthier et al., 1990). The Hall deposit, south of Thetford Mines, is composed of dunitic rocks hosting chromitite (Morin et al., 1992a, b). Chromitite occurs stratigraphically above hydrothermally-altered gabbros that are interfolded with ultramafic cumulates (Bédard et al., 1992). PGE-rich chromitites, containing inclusion-bearing Cr-rich spinels, fill breccia zones in a discordant dunitic pipe that crosscuts the ultramafic cumulates (Morin et al., 1992; Tanguay et al., 1990).

Western Newfoundland ophiolites, such as the Bay of Islands complex, preserve up to 12 km of oceanic lithosphere, including a 6 km section of deformed mantle rocks (Suhr, 1992, 1993). The upper mantle tectonites and the base of the crustal section were intruded by dunite and pyroxenite intrusions that crystallized from boninitic magmas that reacted with the surrounding mantle rocks (Varfalvy et al., 1994, in press; Bédard et al., 1994). Chromitites deep within the mantle section are interpreted to represent intra-channel chromite cumulates (Edwards, 1990), but their pressure of formation is poorly constrained. Chromitites in the lower crustal section may have a shallow reaction origin (Bédard, 1994) and so are not good targets for diamond exploration. However, northeast of the Bay of Islands complex along the western edge of the Dunnage zone are major shear zones, where ophiolite suites, volcanic cover sequences and various intrusions are structurally juxtaposed, and amphibolites with eclogitic cores have been reported (Hibbard, 1983).

Diamond-bearing pyroxenite layers within the Beni Bousera peridotite massif crystallized at high pressures by crystal segregation in magma conduits intruding the peridotites (Pearson et al., 1993). The magmas crystallized clinopyroxene, orthopyroxene and garnet at pressures constrained to be above 45 kbar by the presence of abundant graphitized diamonds. Transport of high pressure ultramafic rocks into the crust from within the diamond stability field can follow various uplift paths (Nixon et al., 1991). In the Ronda and Beni Bousera massifs, the uplift history results in complete transformation of diamond to graphite (Davies, et al., 1993; Nixon et al., 1991; Van der Wal and Vissers, 1993). Other uplift paths permit metastable preservation of diamond, such as in the Luobusa and Donqiao ophiolites in Tibet (Bai et al., 1993). Partial tectonic exhumation of subducted slabs can bring diamond-bearing ultramafic rocks to relatively shallow levels, where they can be sampled by shallowly derived magmas, such as nephelinites and alkaline basalts in New South Wales, Australia (Barron et al., 1994).

## SUMMARY

Ultramafic rocks can be emplaced tectonically into the crust from the diamond stability field, suggesting that the pressure-temperature conditions of formation for fault-bounded ultramafic massifs need to be carefully investigated. Graphite pseudomorphs after diamond comprise up to 15% of four garnet pyroxenite layers in the Beni Bousera massif (Nixon et al., 1991; Slodkevich, 1983). Therefore, the geologically interesting search for in situ diamonds preserved in high-pressure ultramafic complexes could be rewarded by the discovery of rich diamond source rocks.

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# Diamonds in ultrahigh-pressure metamorphic rocks

**R.G. Berman**

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## INTRODUCTION

"Ultrahigh-pressure" (UHP) metamorphic rocks are a recently recognized metamorphic facies, defined by the occurrence of coesite and/or diamond, the high pressure polymorphs of quartz and graphite, respectively. Coesite of non-impact origin was initially discovered in the southern Dora Maira Massif of the western Alps (Chopin, 1984) in whiteschists, rocks consisting of talc, phengite, pyrope, kyanite, quartz, and rare jadeitic pyroxene. Coesite (or quartz pseudomorphs after coesite) has since been reported from eclogitic rocks of the Caledonides in Norway (Smith, 1984) and the Dabie Mountains in central China (Okay et al., 1989; Wang et al., 1989). Diamonds formed in situ in crustal rocks were first discovered in Proterozoic eclogitic rocks of the Kokchetav Massif, northern Kazakhstan, (Sobolev and Shatsky, 1990), and were subsequently also found in eclogites of Shandong (Enami and Zang, 1990) and Dabie Shan (Shutong et al., 1992) in south China. The eclogites generally occur as tectonic blocks within quartzofeldspathic gneisses and marbles. Diamonds reported to date from these rocks have an average grain size of about 15 μm.

Experimental phase equilibrium studies indicate that formation of coesite and diamond require pressures in excess of approximately 17 and 30 kbar, respectively, at geologically feasible temperatures. Figure 1, computed with the thermodynamic database of Berman and Aranovich (1993; in press) shows these stability fields along with those of various silicates + magnesite, the carbonate with the highest pressure stability field. Magnesite has been found in coesite- and diamond-bearing eclogites and peridotites (Yang et al., 1993; Zhang and Liou, 1994), and represents a likely carbon reservoir in the lower crust and upper mantle. Thermodynamic calculations show that diamond is stable under approximately 1 log $fo_2$ more oxidizing conditions for eclogitic rocks containing clinopyroxene compared to peridotites containing enstatite + forsterite (Luth, 1993; Ogasawara et al., 1994). This contrast indicates that eclogites can contain diamonds, while diamond-free peridotites contain magnesite at the same P, T, and $fo_2$, consistent with observations of diamond-bearing xenoliths recovered from kimberlites (Luth, 1993).

That diamonds found in metamorphic rocks formed in situ is supported by their occurrence exclusively in eclogite facies assemblages (mafic eclogites, quartzofeldspathic gneisses, marbles) that bear other indications of formation at very high pressure. These include high jadeite and high $K_2O$ content (~1 wt %) in clinopyroxene, high $SiO_2$ content of muscovite, high pyrope + grossular content of garnet, and high $Al_2O_3$ (~10 wt %) content of titanite. Eclogites as well as host gneisses display evidence of clockwise P-T-t paths with final stages of strong isothermal decompression (Wang et al., 1992; Ernst et al., 1994). The preservation of diamond (and coesite) is achieved by their occurrence as micro-inclusions in garnet, kyanite, and zircon, high strength minerals that act as natural "pressure vessels", shielding their inclusions from ambient physical conditions and from fluids. Preservation was probably also aided by the strong ductility contrast between the eclogite blocks and host marbles and gneisses, with strong strain partitioning into the host rocks.

The intimate association of graphite with diamond in most inclusions indicates that special tectonic processes are required not only in reaching the ultra-high pressures of diamond formation, but also in returning these metamorphic rocks to the Earth's surface with sufficient rapidity to preserve diamonds. A model that fits most observations from the Kokchetav Massif (Shatsky et al., 1993) and southern China (Li et al., 1993; Ernst et al., 1994), as well as from the western Alps (Chopin, 1987), is that UHP terranes form through continent-continent collision after closure of an oceanic basin. The negative buoyancy of eclogites and garnet peridotites formed from oceanic lithosphere acts as an "anchor" causing subduction of attached thin slivers of continental crust to extreme depths. Strong depression of ambient lower crust - mantle isotherms by as much as 1000 °C in a cold, thick subducting slab (e.g. Wiens et al., 1993) lowers the pressures required to cross into the coesite and diamond stability field relative to those on a continental geotherm (Fig. 1). Return to the surface of folded and

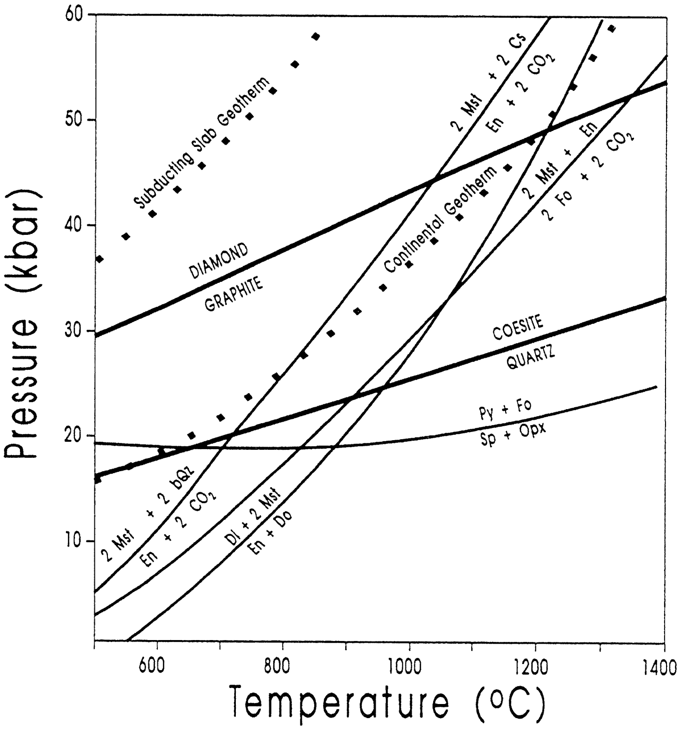


*Figure 1. Pressure - temperature diagram computed with thermodynamic data of Berman and Aranovich (in press) showing diamond and coesite stability fields relative to geotherm for stable continental shield (44 mW/m$^2$; Gurney and Harte, 1980) and approximated for subducting oceanic slab (see text). All equilibria calculated with the TWQ software (Berman, 1991) in the system $MgO$-$CaO$-$SiO_2$-$CO_2$-$H_2O$, with $X_{CO2} = 0.10$. Mineral abbreviations: Mst = magnesite; Cs = coesite; En = enstatite; Fo = forsterite; Py = pyrope; Sp = spinel; Opx = Al-bearing orthopyroxene; Di = diopside; Do = dolomite; bQz = beta-quartz.*

faulted blocks is accomplished by decoupling of the low density continental material from the eclogitized oceanic crust (Ernst et al., 1994). The essential difference between ultra-high pressure and high pressure subduction-related eclogites (and blueschists) may be the greater structural integrity and density of the cratonic crust relative to trench sediments, a difference that allows subduction to greater depths before the onset of decoupling and ascent (Ernst et al., 1994). Slab breakoff (Davies, J.H. and von Blackenburg, F., 1995) may also provide a means for rapid exhumation. This tectonic model also provides a genetic link between eclogites and high pressure ultramafic rocks derived from subduction of oceanic lithosphere (LeCheminant and Bédard, 1996). Barron et al. (1994) proposed an alternate model for returning diamonds to the surface in nephelenitic magmas which may sample partly exhumed eclogitized oceanic crust.

## CANADIAN ECLOGITES

Recent metamorphic map compilations (Read et al., 1991; Sinha et al., 1993; Berman et al., 1993) show areas with eclogite facies metamorphic rocks in Canada that represent the best possibilities to explore for diamonds in crustal rocks. Overall, eclogite is extremely rare, with occurrences in Phanerozoic rocks of the Cordillera (A in Fig. 2) and Newfoundland (B in Fig. 2), in the Proterozoic Grenville Province (C, D, E in Fig. 2), and in Archean rocks of northern Saskatchewan (F in Fig. 2). Only the Cordilleran eclogites appear to have formed directly through subduction of oceanic rocks.

Cordilleran eclogites are distributed over a strike length of several hundred kilometres in the hanging wall of a fault that separates overthrust Paleozoic - Mesozoic arc rocks of the Yukon-Tanana terrane from underlying rocks of ancestral North America (Erdmer, 1987, 1992). Eclogites occur as lenses up to 100 m thick within blueschists and mafic-ultramafic rocks and record P-T conditions of about 500°C - 10 kbar to 750°C - 15 kbar. Farther south, similar eclogites occur as tectonic blocks in ultramafic rocks near Pinchi Lake, British Columbia, and record P-T conditions of 350 °C - 10 kbar to 565 °C - 13.1 kbar (Ghent et al., 1993). These eclogites are interpreted to represent trench and oceanic rocks that have been obducted onto foreland rocks of the North American craton, and then offset by extensive right lateral movement along the Tintina fault (Erdmer, 1992).

Minor eclogite occurs in the core of amphibolite bodies within semipelitic schists and metaconglomerates of the East Pond Metamorphic Suite in Newfoundland (Hibbard, 1983). Pyroxene and amphibole chemistry (De Wit and Strong, 1975) suggest eclogite formation at relatively low pressures.

Eclogites have been recognized recently in various parts of the Grenville Province (Davidson, 1991, 1993; Indares, 1992, 1993). All appear to have been derived from basic anorthosite-gabbro igneous suites emplaced into, and metamorphosed and deformed along with adjacent quartzofeldspathic gneisses. In the western Grenville (C and D in Fig. 2), retrogressed eclogites occur as small tectonic lenses and pods strung out along a > 20 km zone. The eclogites contain plagioclase and augite with up to 10 mol % jadeite near Rosseau Lake, Ontario (Davidson, 1991) and between 10 to 25 mol %

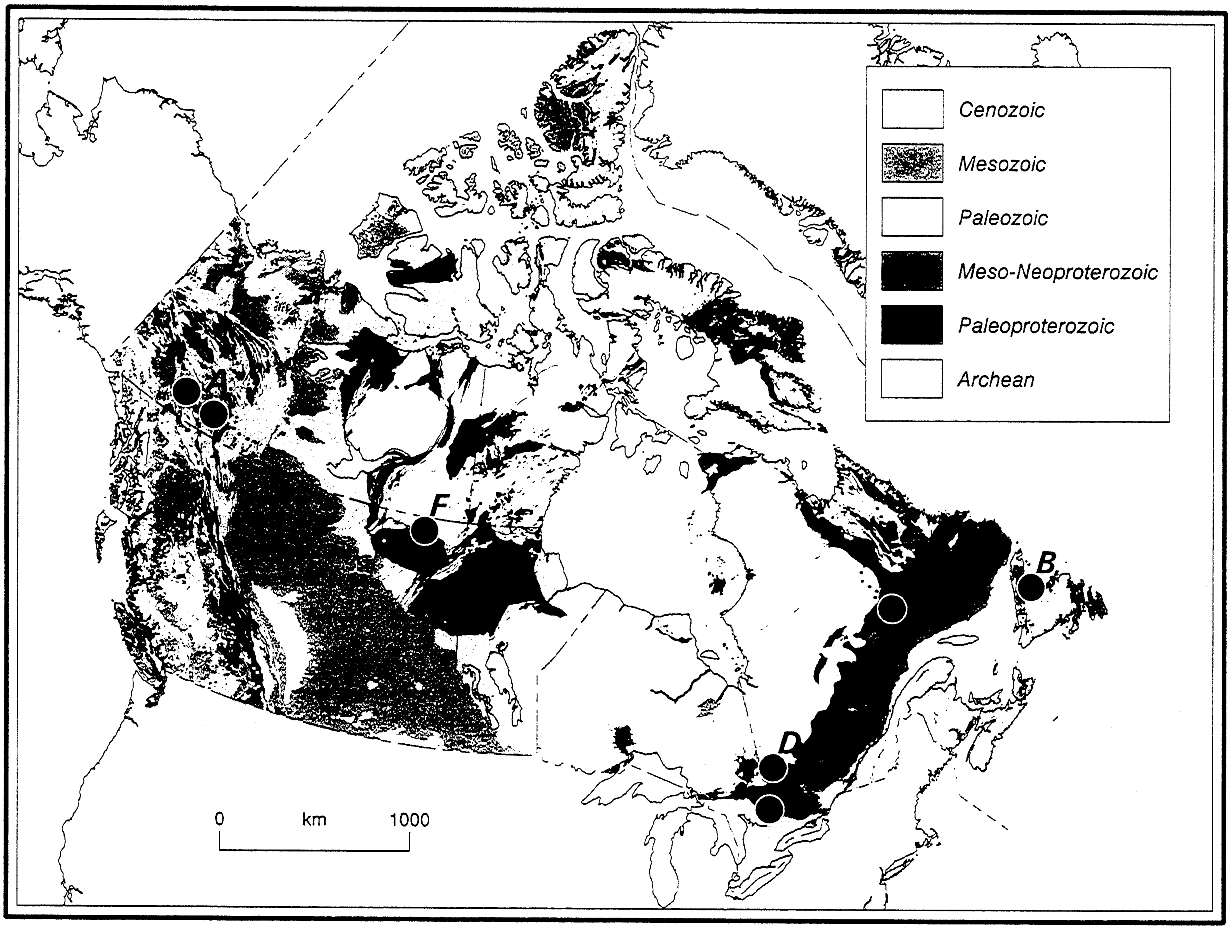


*Figure 2. Simplified geological map derived from the Geological map of Canada (Wheeler et al., in press), showing reported eclogite locations (see text).*

in the Timiskaming region (Indares, 1992). Grant (1988) estimated P-T conditions for one metagabbro of 15 kbar and 825 °C. Symplectitic intergrowths of clinopyroxene-plagioclase, pargasite-plagioclase, and corundum-spinel-sapphirine-plagioclase are interpreted as retrograde products of the breakdown of omphacitic pyroxene, reaction between omphacite and garnet, and reactions involving kyanite, clinozoisite, garnet, possibly with omphacite, respectively (Davidson, 1991).

The most extensive, as well as the best preserved and highest P-T eclogites in Canada have been described by Indares (1993, 1994) in the eastern Grenville Province (E in Fig. 2). In the Gagnon and Molson Lake terranes of the parautochthonous belt, continental tholeiitic gabbros of the 1.43 Ga Shabogamo intrusive suite are eclogitized into elongate bodies up to several hundred metres thick, with massive cores of omphacite (up to 39 mol % jadeite)-garnet and foliated margins of hydrated mafic gneisses. P-T conditions of 16 kbar and 700 to 800 °C were attained, followed by decompression along steep P-T paths (Indares, 1993). The largest eclogite body occurs within the Lelukuau domain of the Manicouagan shear belt which forms the southern boundary of the Gagnon terrane (Indares, 1994).

Estimated temperatures for the Grenville eclogites fall within the range of medium-temperature eclogites (Carswell, 1990) considered to form in tectonically overthickened continental crust. This origin is also supported by the occurrence of eclogites and meta-eclogites in the lowest structural domains of the Grenville Province, and by the clockwise P-T-time paths implied by equilibration of surrounding rocks at higher T and lower P (Davidson, 1991).

High-temperature eclogites also occur in the western Canadian Shield as layers up to 15 metres thick within ~10 kbar granulites of northern Saskatchewan interpreted to have formed in an intracontinental environment (Hanmer et al.,1994). The eclogites are unusual in that they represent the oldest eclogites, ~ 2.6 Ga, yet reported worldwide. Host quartzofeldspathic gneisses record P-T conditions in excess of 15 kbar and 1000 °C (Snoeyenbos et al., 1995). Farther northeast along the Snowbird zone, granulites of the Kramanituar Complex indicate peak P-T conditions of 14 kbar and 850 °C (Sanborn-Barrie, 1994).

## SUMMARY

UHP metamorphic rocks appear to form during continent-continent collisions during which oceanic or continental crust is subducted to depths within the diamond stability field. Known eclogite occurrences in Canada are limited in number, but have a wide geographic scope. Paleopressures recorded in these eclogites fall short of the diamond stability field.

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# Impact diamonds

**R.A.F. Grieve and V.L. Masaitis**

*Grieve, R.A.F. and Masaitis, V.L., 1996: Impact diamonds; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 183-186.*

## INTRODUCTION

Impact diamonds, like other high pressure polymorphs of various silicate minerals, occur as the result of the passage of a shock wave. The first indication of impact diamonds was the discovery in the 1960s of diamond with lonsdaleite, another (hexagonal) high pressure polymorph of carbon, in placer deposits, e.g., in the Ukraine, but their source was unknown. In the 1970s, diamond with lonsdaleite was discovered in the impact lithologies at the Popigai impact structure, Russia. Since then, they have been discovered at a number of other impact structures, e.g., Kara, Puchezh-Katunki, and Ternovka in Russia, Ries in Germany, and Zapadnaya in Ukraine (Fig. 1). Impact diamonds differ from kimberlite diamonds in a number of ways and details of their morphology, physical properties, and isotopic composition can be found in Masaitis (1993) and references therein. The study of impact diamonds has been almost exclusively carried out by workers from the former USSR and the number of studies are relatively limited.

## ORIGIN

Impact diamonds originate due to phase transitions that occur when their precursor carbonaceous lithologies (graphite or coal) are subjected to shock pressures $\geq 30$ GPa (300 kbar). Impact diamonds from graphite in crystalline targets usually occur as paramorphs, with inherited crystallographic features (Masaitis et al., 1990; Val'ter et al., 1992). They occur generally as microcrystalline aggregates, which can reach 1 cm in size, consisting of cubic diamond and lonsdaleite, with individual microcrystals of $10^{-4}$ cm. Impact diamonds from coal, or other carbon in sediments, are generally porous, white, black or brown, and may have a palimpsest biogenic texture, i.e., they have some of the structural properties of the original biological matter.

Diamonds occur up to about ten carats per tonne and in a variety of allochthonous lithologies at impact craters. They are most common as inclusions in impact melt rocks, in impact melt clasts in suevite breccias, and in pseudotachylyte dykes. For example, at the 3.8 km in diameter Zapadnaya structure, Ukraine, they occur in impact melt dykes in the central uplift and in suevite breccias in the peripheral trough. Zapadnaya is formed in Proterozoic granite containing graphite (Gurov et al., 1985). At the 100 km Popigai structure, the allochthonous breccia filling the peripheral trough is capped by diamond-bearing suevites and coherent bodies of impact melt rocks. The largest of these melt rock bodies can be traced for 10 to 15 km along strike and is 500 m thick (Masaitis et al., 1980). In the case of Popigai, the original source of the carbon is Archean gneisses containing graphite, and diamond are also occasionally recovered from shocked in situ gneisses (Fig. 2). The diamonds at the 65 km diameter Kara structure, Russia, also occur in impact melt rocks, with the source of carbon being Permian terrigenous sediments containing coal (Ezkerskii, 1982).

The distribution of diamonds within an individual impact structure and within individual impact melt and suevite bodies can be irregular. They are found in rays or zones emanating from areas where the original carbon-bearing lithologies are most abundant (Masaitis, 1993). They also tend to be radially restricted. Closer to the centre of the structure, high post-shock temperatures cause more rapid oxidation of the diamonds. On the other hand, at some critical distance from the centre, the shock pressures are too low to result in the phase transformation to diamond. Chilling of the impact melt aids preservation. When the melt matrix has cooled relatively slowly, the diamonds show evidence of reversion to graphite and oxidation. Although diamonds associated with known impact structures are not currently exploited commercially, those produced by shock transformation of graphite have some technical advantages over kimberlitic diamonds in that they tend to be harder and more resistant to breaking, because of their polycrystalline nature as a combination of cubic and hexagonal phases.

No searches have been made for impact diamonds at any of the 26 currently known impact structures in Canada (Fig. 1). A likely candidate structure is the 13 km diameter Deep Bay structure, Saskatchewan,

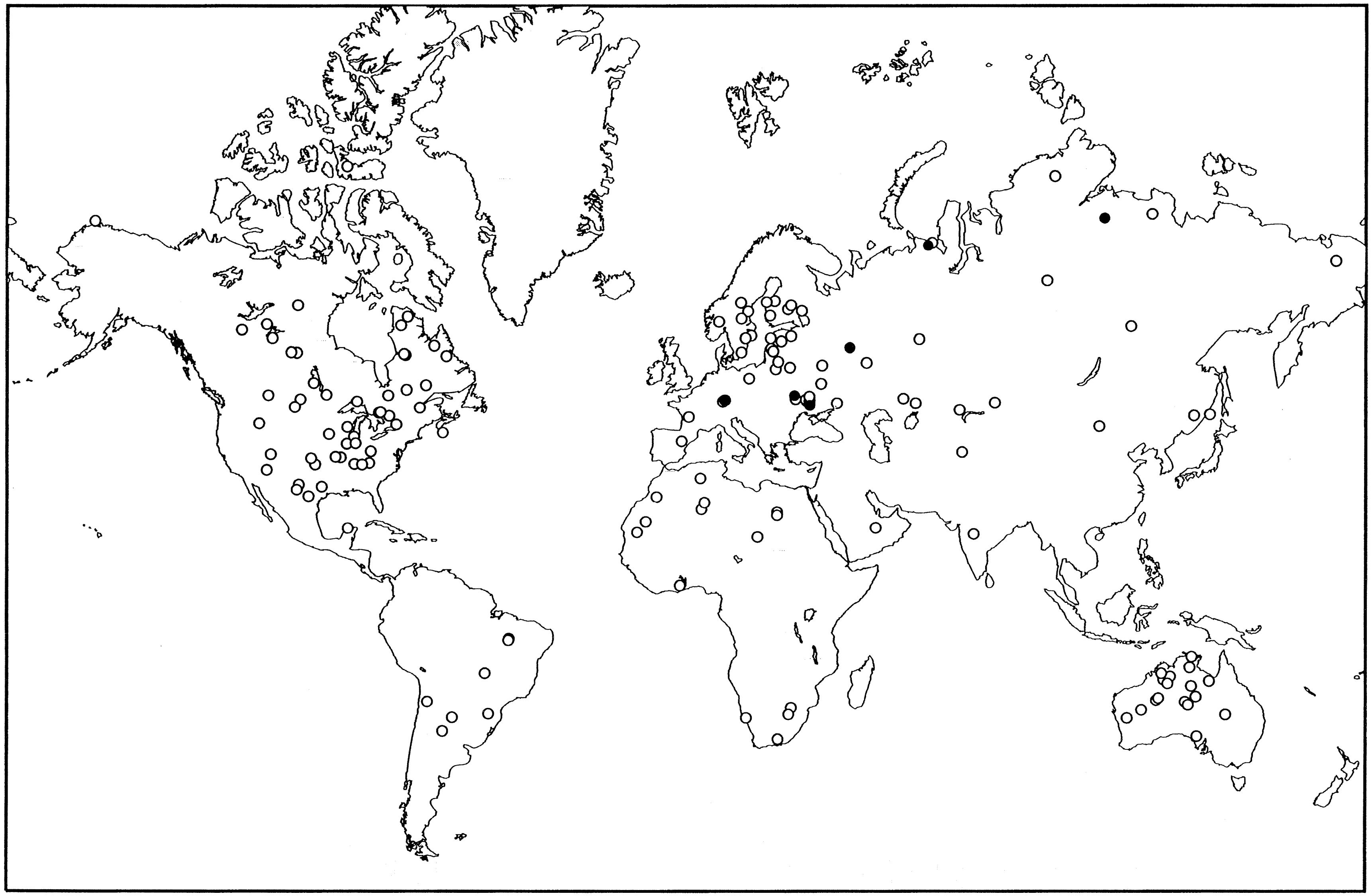

*Figure 1.* Location of currently known terrestrial impact structures (open circles). Structures known to contain impact diamonds are indicated by black circles.

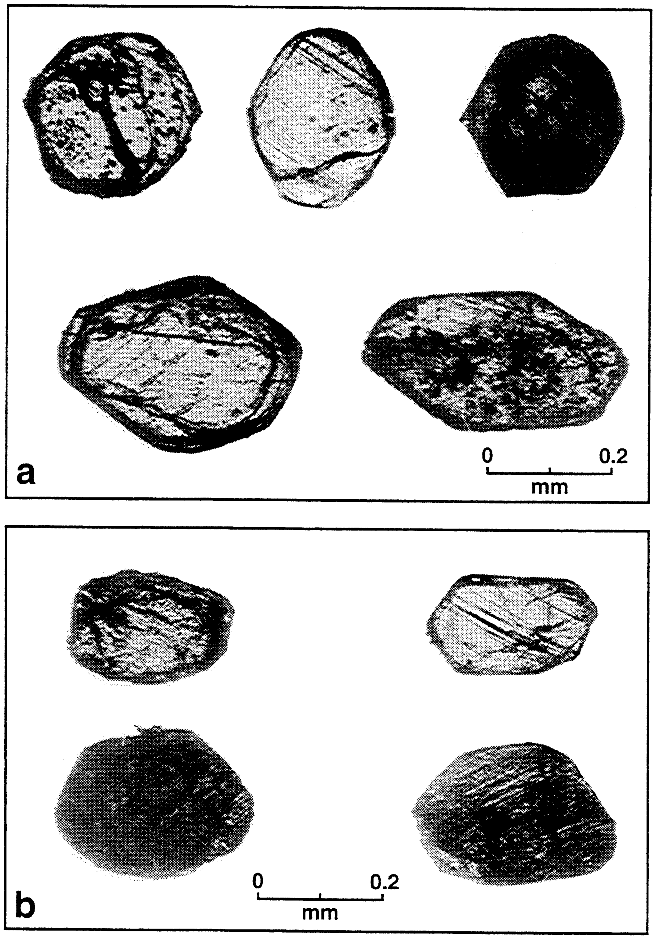


*Figure 2. Photomicrographs of impact diamonds recovered from the: a) Popigai impact structure and b) Ries impact structure.*

where the target rocks, in part, consist of graphite-bearing gneisses, with lenses estimated to contain 25% graphite (Innes et al., 1964). Another possible candidate is the originally ~ 250 km diameter Sudbury structure, Ontario. Here, the 1.8 km thick Onaping Formation, which is largely a suevite breccia contains ~ 0.5 to 1.0% carbon. Recently, a few parts per million of fullerenes ($C_{60}$ and $C_{70}$) have been identified within the Onaping (Becker et al., 1994). Fullerenes form under intense temperatures and pressures and their occurrence in the Onaping is related to the Sudbury impact event 1.85 Ga ago. The source of the carbon is unknown, and may have been the impacting body (Becker et al., 1994). Fullerenes have also been detected in Cretaceous/Tertiary (K/T) boundary deposits (Heymann et al., 1994), and diamonds have been recovered from K/T boundary deposits in Alberta (Carlisle and Braman, 1991). They are < $5 \times 10^{-7}$ cm in size and represent the only currently known diamonds in Canada linked to an impact event.

## CARBONADOS

Carbonados are irregular polycrystalline diamond aggregates that occur in placers and in low grade metamorphic rocks. They occur in Australia, Brazil, Russia, South Africa, Ukraine, Venezuela, Central African Republic, and Zaire. They are used for industrial diamond applications (Trueb and de Wys, 1971; Kaminsky et al., 1978; Kaminsky, 1994). Carbonados are associated with crustal parageneses and are not related to kimberlites. Some of the first impact diamonds, known as yakutite and found in placers in Northern Yakutia, Russia, were originally classified as a form of carbonado (Orlov and Kaminsky, 1981) and this has led to some confusion. The origin of carbonados is debatable, but the isotopically light character, noble gas contents, rare earth abundance patterns, and associated parageneses indicate a crustal source (Shibata et al., 1993). The wide variety of parageneses between carbonado deposits suggests a variety of progenitors. This range of sources and the apparent presence of lonsdaleite led Smith and Dawson (1985) to suggest that carbonados were formed by Precambrian-aged impacts into carbon-bearing lithologies. The apparent association with the high pressure polymorph lonsdaleite, however, is related to the original discovery of yakutite (Kaminsky, 1994). There is no evidence of shock metamorphism in associated minerals in carbonados. At the present time, impact must be considered as a relatively unlikely hypotheses for the genesis of carbonados. The hypothesis with the most consensus for the genesis of carbonados appears to be irradiation of organic carbon by high energy particles in a uranium-rich environment (Kaminsky, 1994; Shibata et al., 1993).

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# PART 2: DIAMOND EXPLORATION IN GLACIATED TERRAIN

## Introduction

**R.N.W. DiLabio**

*DiLabio, R.N.W., 1996: Part 2: Diamond exploration in glaciated terrain - Introduction; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 187-189.*

Diamond exploration in Canada has been hampered by the lack of exploration methods that are directly applicable in glaciated terrain. Most methods that rely on the search for indicator minerals or the geochemical signature of kimberlite were developed in nonglaciated regions of the world. They can be ineffective or misleading when applied to glaciated regions, where the provenance of surficial sediments is controlled by glacial dispersal, which complicates sampling design and interpretation. Although there are many research papers on exploration for gold and base metals using glacial drift as the sampling medium (e.g., Coker and DiLabio, 1989; DiLabio and Coker, 1989), until recently, only a few papers had been published that are pertinent to the search for diamonds in Canada (Brummer et al., 1992a,b). Attempts at the end of the last century to find the source of the mid-continent diamonds found in glacial drift recognized the complications involved and invoked glacial transport to explain the distribution of diamond discoveries (Hobbs, 1899). Much later, Lee (1965, 1968) showed that pyrope recovered from esker sediments near Kirkland Lake, Ontario could be traced up-ice to kimberlite bedrock. Recently, the Kirkland Lake area again became the focus for studies aimed at developing and refining exploration methods (McClenaghan, 1996). In new areas where diamond exploration has been active, research into drift prospecting methods continues (Ward et al., 1996; Garrett and Thorleifson, 1996), and biogeochemical exploration methods for finding kimberlite have been tested at several sites (Dunn and McClenaghan, 1996).

The single most important feature of glaciated terrain that complicates exploration using drift mineralogy or geochemistry is the exotic provenance of glacial drift. Bedrock is glacially eroded and dispersed away from its source along glacial flowlines, which can trend up regional slopes, cross drainage divides, shift through the course of a glacial cycle, and deviate markedly from post-glacial drainage patterns. For example, Veillette and McClenaghan (1995) have shown that ice flow during the last glaciation in the Abitibi-Timiskaming region of Ontario and Quebec, which contains several kimberlites, was opposite to the present flow patterns of rivers in the region. In addition, the paths followed by the ice shifted through 180° during the glaciation. Therefore, it is essential to know the history of ice flow in any region under exploration, so that a mapped pattern of the distribution of indicator minerals can be traced up-ice to its source in the bedrock. A generalized map of ice flow features from the last glaciation is shown in Figure 1. It is not meant to define the ice flow sequence at a detailed scale; shifts in ice flow cannot be displayed at the scale of this figure. Detailed mapping of glacial flow sequences is now available for many parts of Canada, particularly those areas where diamonds (and other commodities) may be found (Veillette and McClenagan,1995; Ward et al., 1994).

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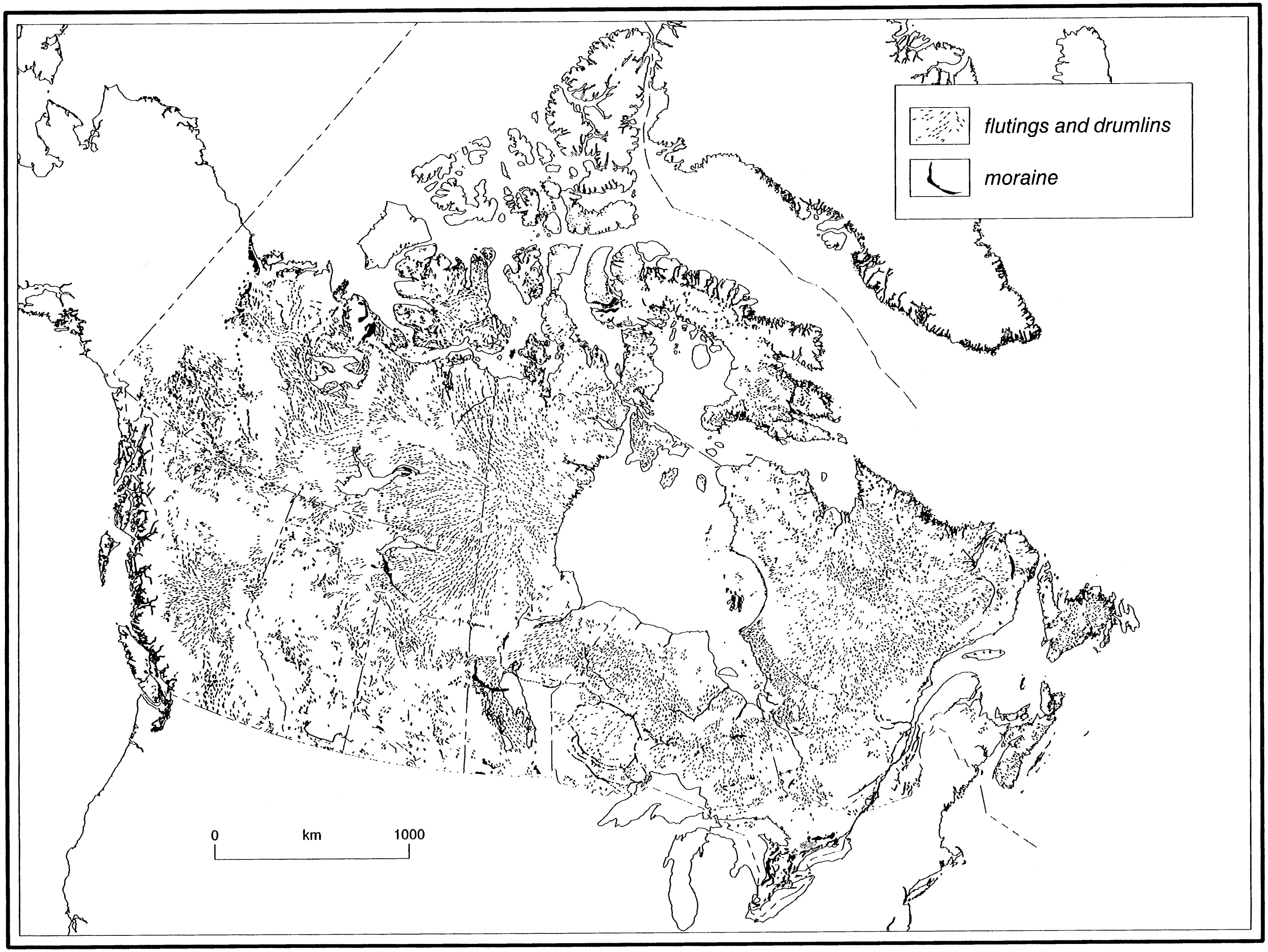


*Figure 1. Ice flow features in Canada (after Fulton, 1995).*

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# Kimberlite indicator minerals in glacial deposits, Lac de Gras area, N.W.T.

**B.C. Ward, L.A. Dredge, D.E. Kerr, and I.M. Kjarsgaard**

*Ward, B.C., Dredge, L.A., Kerr, D.E., and Kjarsgaard, I.M., 1996: Kimberlite indicator minerals in glacial deposits, Lac de Gras area, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 191-195.*

## INTRODUCTION

The Terrain Sciences Division component of the Slave Province National Mapping Programme (NATMAP) is centred in an area where numerous diamondiferous kimberlites have recently been discovered. The purpose of the surficial geology mapping and related research is to provide a regional framework for geological interpretation that will aid in land use planning and geotechnical studies, environmental assessment, and drift prospecting, in both the long and the short terms. Those aspects relevant to diamond exploration include determining the nature and distribution of surficial materials, interpreting glacial ice flow directions and dispersal patterns, analyzing kimberlite indicator minerals, and measuring concentrations of trace elements characteristic of kimberlite in glacial sediments. From these studies, regional patterns and background concentrations of kimberlite indicator minerals have been established (Dredge et al., 1995; Kerr et al., 1995; Ward et al., 1995). This work provides reference data for diamond exploration companies working in the Slave Province, as well as in other glaciated regions. Surficial geology mapping and sampling for 1:250 000 NTS map areas 86A, 86H, 86I, 76C, 76D, and 76E S1/2 have been completed. Maps and till geochemistry reports are available for fieldwork completed in 1992 and 1993 (Dredge et al., 1994a, b, c, and 1996b; Kerr et al., 1994a, b and in press; Rampton and Thomas, 1993; Thomas et al., 1993; Ward et al., 1994a, b, c, and in press). Preliminary results of 1995 investigations completed in the Kikerk Lake and Coppermine map areas (NTS 86P and 86O E1/2) were summarized in Kerr et al. (1996), and two GSC open file maps will be released in mid 1996. The following results pertain specifically to the Lac de Gras region (76C, D, and 86A), the current focus of diamond exploration activity.

A total of 194 10 kg bulk till samples were collected from shallow pits for mineral grain analysis, and 500 additional smaller samples were analyzed for trace element geochemistry. Heavy mineral concentrates from bulk samples were picked to recover kimberlite indicator minerals (pyrope, eclogitic garnet, Cr-diopside, Mg-ilmenite, chromite, and corundum) in the 0.25 to 0.50 and 0.50 to 1.00 mm size fractions. All potential indicator grains were analyzed using the electron microprobe (Stirling and Pringle, 1996). Grain morphology of pyropes was examined using the GSC scanning electron microprobe (Dredge et al., 1996a).

## INDICATOR MINERAL CONCENTRATIONS

### *Regional Patterns*

The regional distribution of kimberlite indicator minerals displays wide variability, and background values range from 0 in the Aylmer Lake area, to 7 to 10 in the Winter Lake area, to >15-20 in the northern portion of the Lac de Gras area (Fig. 1). This distribution can be explained by the ice flow history and the principal areas of known kimberlites for the area. Samples with the highest concentrations of indicator minerals occur either adjacent to, or down ice from, known clusters of kimberlite pipes in the northern half of the Lac de Gras (76D) map area. The area with the lowest concentrations of indicator minerals is the Aylmer Lake map area, which is up-ice from most of the known kimberlites. The abundance of samples with low concentrations of indicator minerals in the Winter Lake (86A) map area can be explained by glacial dispersal from kimberlites to the east. The ice flow history determined for the area indicates an earlier phase of southwestward and westward flow which could result in this diffuse dispersion from the Lac de Gras kimberlite field. This makes identification of anomalies difficult, since most samples would be expected to contain a few indicator minerals. However, a sample to the southwest of Winter Lake containing almost 100 indicator minerals is clearly anomalous.

The area with the highest concentrations of both the total number of indicator minerals and pyropes was used to informally define the Lac de Gras dispersal plume (Fig. 1). This plume reflects the combined signature of all

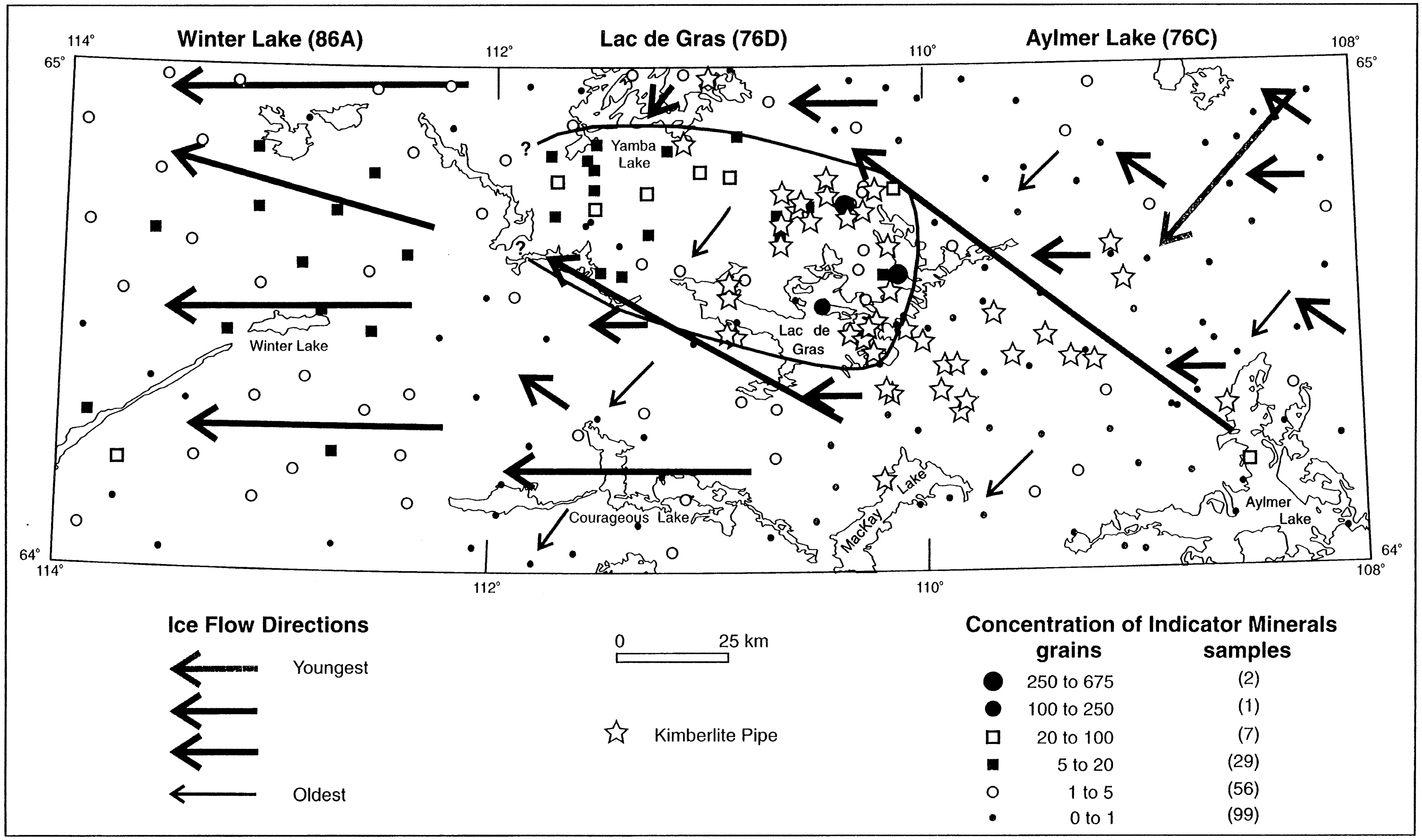


***Figure 1:*** *Distribution of all kimberlite indicator minerals for the 0.25 to 0.5 mm size fraction. The Lac de Gras dispersal plume is outlined by the bold solid line. Glacial flow directions are modified after Ward et al. (1994b) with relative ages determined by crosscutting relationships between striae. The longest arrows represent the dominant flow for a region, responsible for most of the transport of glacial debris. Locations of kimberlite pipes after Pell (1995).*

the pipes in the region, and indicates that till can be used as a regional exploration tool for clusters of kimberlite pipes. Its elongate nature to the northwest corresponds to the dominant direction of glacial transport. Sample density was too low to resolve individual dispersal trains.

There is some variation in distribution between the different indicator minerals. Almost no pyropes were found in the Aylmer Lake (76C) map area, but several samples contained Cr-diopsides (>1% $Cr_2O_3$). Since some of these are high-Cr-diopsides (>1.4% $Cr_2O_3$) they likely indicate the presence of kimberlites and the lack of pyropes implies that there are regional differences in kimberlite heavy mineral suites. This premise is supported by the sample with the highest concentration of indicator minerals in the Winter Lake (86A) map area since it contains mainly pyropes and no Cr-diopsides.

### *Grain Size Factors*

Total concentrations of indicator minerals ranged from 0 to >1000 grains per 10 kg sample. The majority of the indicator minerals were found in the 0.25 to 0.5 mm (35 - 60 mesh) size fraction (Fig. 2). In this fraction, 95 of the 194 samples contained indicator minerals confirmed by electron microprobe analysis. In the 0.5 to 1.0 mm (18 - 35 mesh) size fraction only 22 samples contained indicator minerals. These data suggest that subtle anomalies will be missed when only the coarser grain size is picked. Use of the 0.25 - 0.5 mm size fraction will become increasingly important as exploration proceeds and all the more obvious dispersal trains have been discovered.

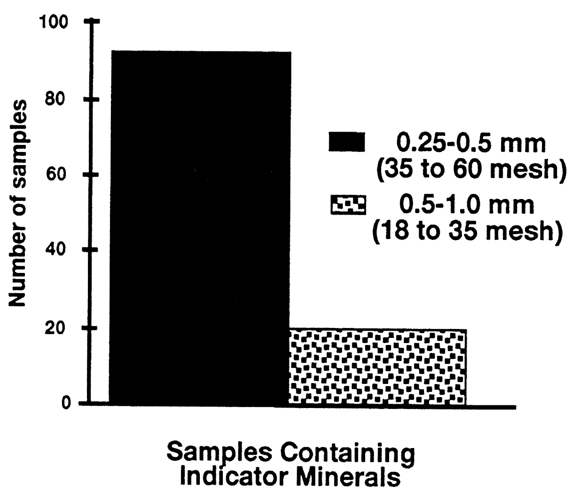


*Figure 2: Size fraction of samples containing indicator minerals.*

### *Mineral Types*

The relative proportion of indicator minerals for the entire (76C, D, and 86A) data set is ~73% pyropes, ~24% Cr-diopsides, ~2% Mg-ilmenites, ~1% chromites, and <<1% eclogitic garnets. There are differences between the different size fractions in the proportions of the minerals (Fig. 3); however, since indicator minerals were only found in 22 samples of the 0.5 to 1.0 mm fraction, this could be non-representative. The results to date show a higher proportion of pyrope garnets in the 0.25 to 0.5 mm size fraction. We suggest that this is a product of primary pyrope garnet size and fracturing in source kimberlites rather than an effect of glacial comminution.

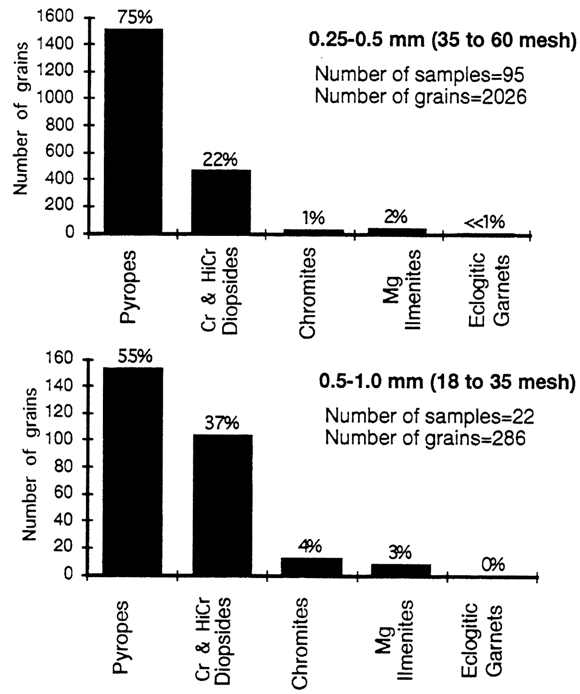


*Figure 3: Relative proportion of indicator minerals in the two size fractions examined.*

## DIAMOND POTENTIAL

Almost every sample with a sizeable number of pyrope garnets (>20 grains) contains "G10" or sub-calcic garnets (Fig. 4). Approximately 3% of all the pyropes analyzed are sub-calcic. These data indicate that most of the kimberlites in the area have incorporated xenoliths and xenocrysts from potentially diamondiferous harzburgite mantle. Conversely, only two eclogitic garnets were identified from more than a hundred potential grains that were analyzed. The virtual absence of eclogitic garnets

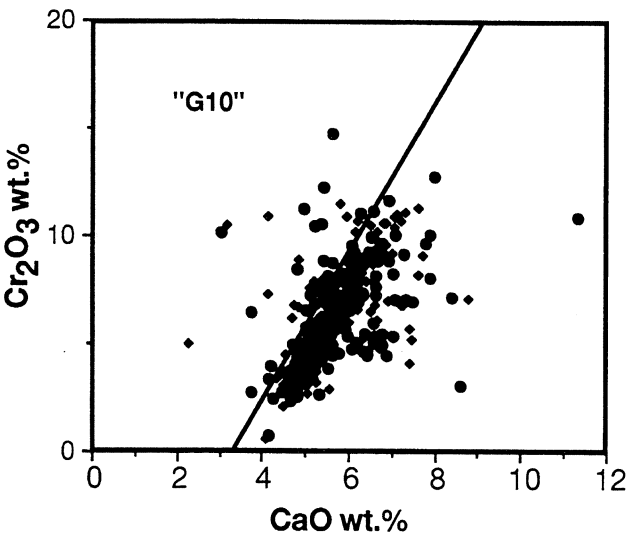


*Figure 4. A plot of $Cr_2O_3$ versus CaO for a representative selection of pyrope garnets from the data set. The subcalcic or "G10" garnets field is defined by the 85% line of Gurney (1984).*

may reflect difficulty in identifying the grains from background crustal garnets and/or a limited contribution from eclogitic mantle.

Chromites have a wide range of chromium contents and can come from non-kimberlitic sources. Chromites are the only indicator minerals present in several samples in the Aylmer Lake (76C) map area, increasing the probability of non-kimberlitic sources, such as mafic volcanics and intrusions, and pyroxenites. Three chromites with >61% $Cr_2O_3$ plot within the diamond inclusion field (Fipke, 1989); this composition increases the likelihood of a kimberlitic source.

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* Contribution to the Geological Survey of Canada's Slave Province National Geoscience Mapping (NATMAP) Program.

# Morphology and kelyphite preservation on glacially transported pyrope grains

**L. A. Dredge, B.C. Ward, and D.E. Kerr**

*Dredge, L.A., Ward, B.C. and Kerr, D.E., 1996: Morphology and kelyphite preservation on glacially transported pyrope grains; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 197-203.*

## INTRODUCTION

Pyrope garnets, brought up in xenoliths from the mantle, are abundant in many diamond-bearing kimberlites (Schulze, 1996). Reactions in the source regions of the xenoliths and interaction of the garnets with kimberlite magma affect their shape and composition, and complex rims (kelyphite) typically develop around individual pyrope grains (Garvie and Robinson, 1984). This report documents the surface morphology of pyrope grains recovered from glacial sediments; particular attention is focussed on kelyphite because its presence has been used previously to suggest the proximity of kimberlite pipes (e.g. Garvie and Robinson, 1984; Averill and McClenaghan, 1994). To date, most studies of pyrope morphology have been carried out in tropical areas on in situ kimberlite, or on alluvium. The results reported here, in contrast, are among the first from glacially transported material.

The Lac de Gras area, N.W.T., is an area of glaciated terrain; pyropes and other indicator minerals have been glacially transported and dispersed from kimberlite sources, and deposited in till and esker sediments. Regional kimberlite indicator patterns and analytic methods have been discussed by Ward et al. (1996) and in other publications (Ward et al., 1995; Dredge et al., 1995; Kerr et al., 1995). For the grain morphology study, four samples from glacial sediments within the Lac de Gras kimberlite field were examined (Fig. 1). They reflect a range of glacial transport conditions and distances. Three samples are from till, glacially transported <1 km (Fig. 2), and >25 km (Fig. 4) from kimberlite sources; the other is from an esker sample (Fig. 3), in which the garnet grains have been carried about 20 km by glaciofluvial processes. Although individual source pipes cannot be identified for the garnets on Figure 4, the transport distances and the part of the kimberlite field the pyropes are from are known. The transport distance for the esker sample was determined by comparing the indicator chemistry of esker and till samples (Ward et al., 1995).

## FRESH GARNET SURFACES

SEM images show that most pyrope grains in the 0.25 to 0.5 mm fraction display fresh, smooth, glassy, conchoidal fracture surfaces having sharp, angular edges. Primary crystal faces are absent. Irregular secondary fractures, developed on the conchoidal surfaces and seen within some pyrope grains, are very clean and fresh (Fig. 3E). The lack of weathering along fracture surfaces suggests that grains broke during glacial transport, rather than during kimberlite emplacement, as is common in pyropes from kimberlites in Africa (Garvie and Robinson, 1984). Hummocky surfaces textures that can develop by resorption during kimberlite groundmass crystallization (Garvie and Robinson, 1984) were not observed on the Lac de Gras grains.

All the grains that have undergone glacial transport, for up to 30 km, remain angular and retain their conchoidal fracture habit (Figs. 2, 4). There is minimal rounding of grain edges. Grains that were transported 20 to 30 km, however, have more conchoidal fracture faces (e.g. Fig. 4B,C) than those much closer to source pipes, and this may be a useful indication of transport distance. Grains from the esker, which have undergone at least 20 km of glaciofluvial transport, also remain angular (Fig. 3). This finding differs from that of Dummett et al. (1987) who observed that sand-sized, stream-transported, pyrope grains are round to subround.

## KELYPHITE

"Kelyphite" derives from the Greek word κελυφοσ, meaning shell or sheath; it refers to reaction coronas surrounding garnet (Garvie and Robinson, 1984). Reid and Dawson (1972) reported that it forms by reaction between garnet and surrounding olivine or orthopyroxene caused by pressure release associated with the ascent of peridotite xenoliths within kimberlite magma. Volatiles may also play a role (Nixon et al., 1963), particularly if the garnets are eclogitic rather than peridotitic. Kelyphite typically consists mainly of pyroxene, with spinel, phlogopite, or serpentine (Garvie and Robinson, 1984),

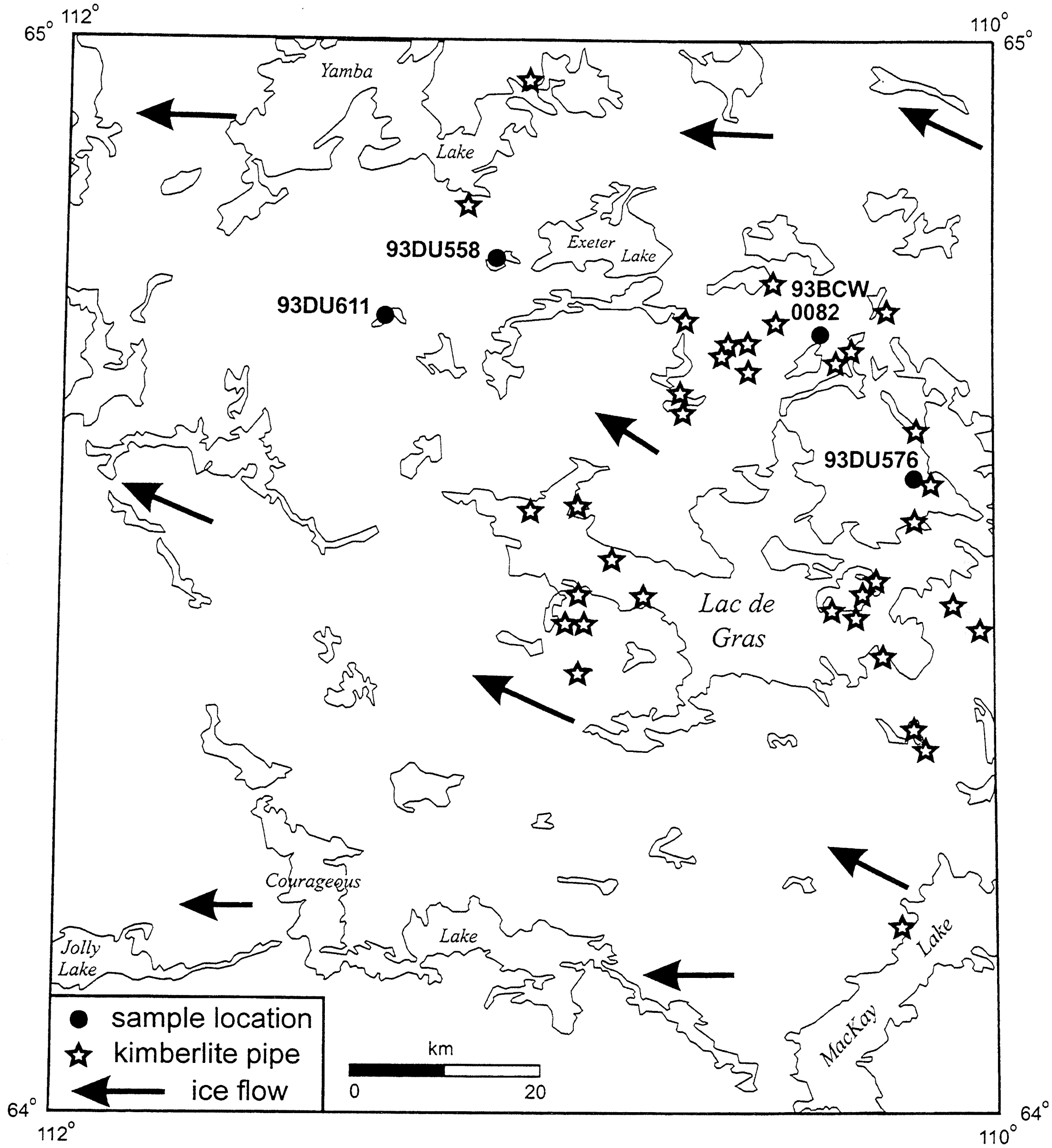



*Figure 1. The Lac de Gras area, and location of samples described in this paper.*

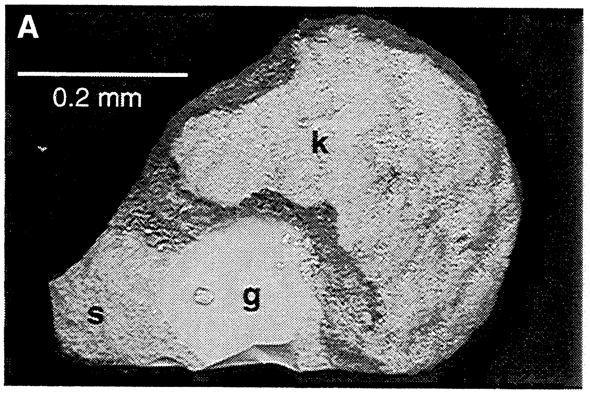
but the thickness and sequence of mineral zones in kelyphite varies from one kimberlite to another.

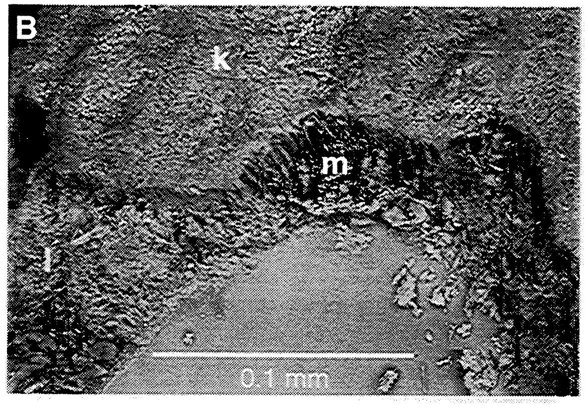
Garvie and Robinson (1984) reported that most garnets from peridotite nodules in kimberlite show evidence of kelyphite development. Since kelyphite layers break off readily from garnets during abrasion (Garvie, 1981 reported in Garvie and Robinson, 1984; McCandless, 1990), the presence of kelyphite on garnet grains should imply a lack of grain wear or transport, and hence the proximity of kimberlite pipes. Kelyphite rims have been found in the Lac de Gras area on grains which have been glacially transported for distances of at least 30 km, and in the esker sample, in grains that have travelled about 20 km. Thus, the general statement that kelyphite weathers and wears off easily, and that the presence of kelyphite implies a nearby pipe is unsupported by our observations.

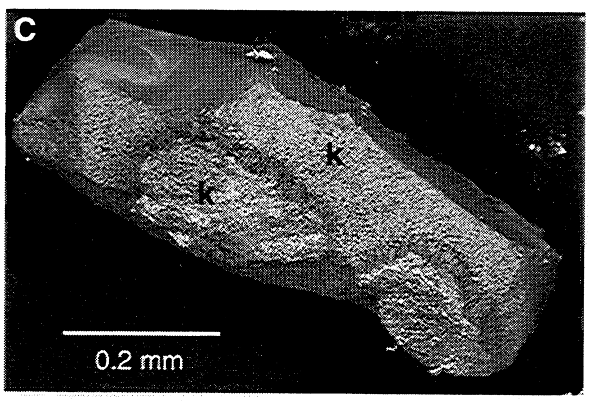
The percentage of pyrope grains with kelyphite rims in the four samples from Lac de Gras ranges from 4% to 23%. It varies without obvious relationship to mode or distance of travel (Table 1). In contrast, both the surface area and the thickness of kelyphite decrease as the distance over which the grains are glacially transported increases (Table 1). On sample 576, which lies nearest its source pipe, kelyphite covers more than half the grains' surfaces and is at least 40 μm thick (Fig. 2A-E), whereas farther away, the area covered declines, and the preserved rim becomes very thin, in places <5 μm (Fig. 4B-E).

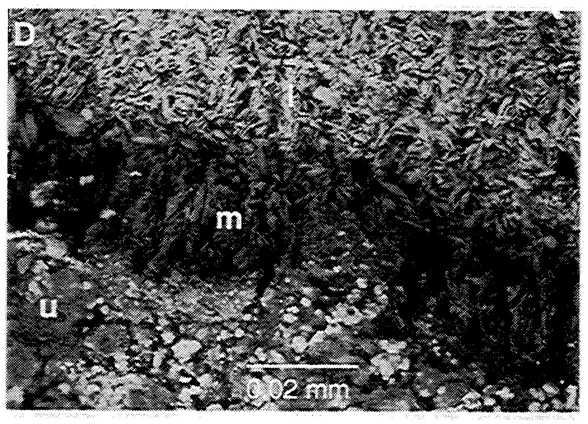
**Table 1.** Pyrope data, Lac de Gras region

| Sample | Material | Estimated distance from pipe (km) | Number of pyrope grains | Number of pyrope grains with kelyphite | Percent pyrope grains with kelyphite | Percent surface covered with kelyphite | Maximum kelyphite thickness (μm) |
|---|---|---|---|---|---|---|---|
| 93DU576 | till | < 1 | > 675 | 52 | ~ 8 | 50 | 25-40 |
| 93BCW0082 | esker | 20 | 267 | 27 | 10 | 50 | 10-35 |
| 93DU558 | till | 25 | 70 | 3 | 4 | < 25 | 2 |
| 93DU611 | till | 30 | 60 | 14 | 23 | 25 | 15 |



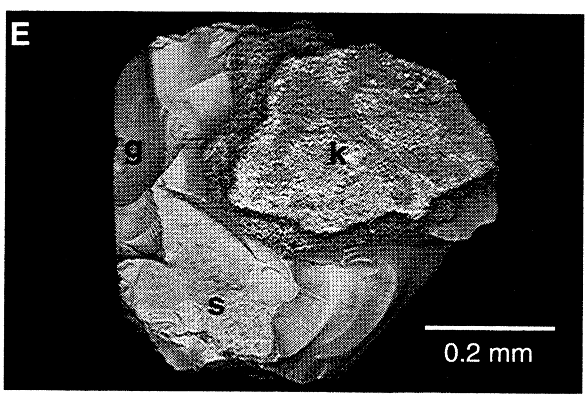




***Figure 2.*** *Pyrope grains from till transported < 1 km from the source pipe (sample 93DU576)*

- A *Kelyphite (k), and glassy (g) and matte (s) sub-kelyphite surfaces.*
- B *Close-up of A showing textural details of the kelyphite, particularly the sharp contact with unaltered garnet, a lower microcrystalline zone (l), a middle columnar zone (m), and the irregular upper surface (k).*
- C *Angular pyrope grain with conchoidal surfaces on the sides, and kelyphite (k).*
- D *Close-up of C showing lower (l), middle (m), and upper (u) kelyphite zones.*
- E *Fractured kelyphite surface (k), and underlying columnar crystals, Garnet surface (g) shows conchoidal fracture. Sub-kelyphite texture(s) is also present.*

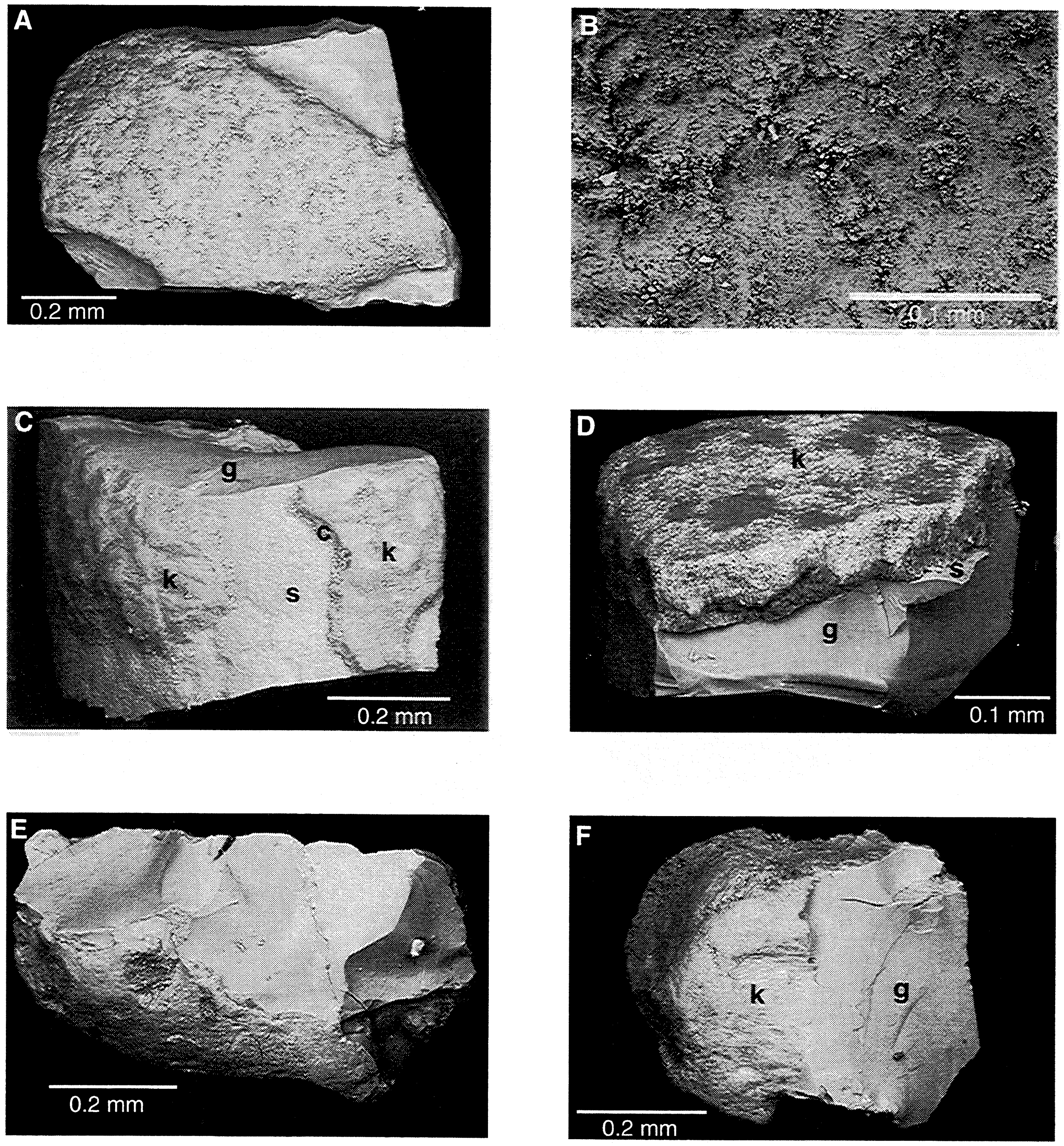



***Figure 3.*** *Pyrope grains from an esker (sample 93BCW0082), transported about 20 km from the kimberlite source.*

*A Irregular, gently rounded kelyphite surface, and underlying garnet.*

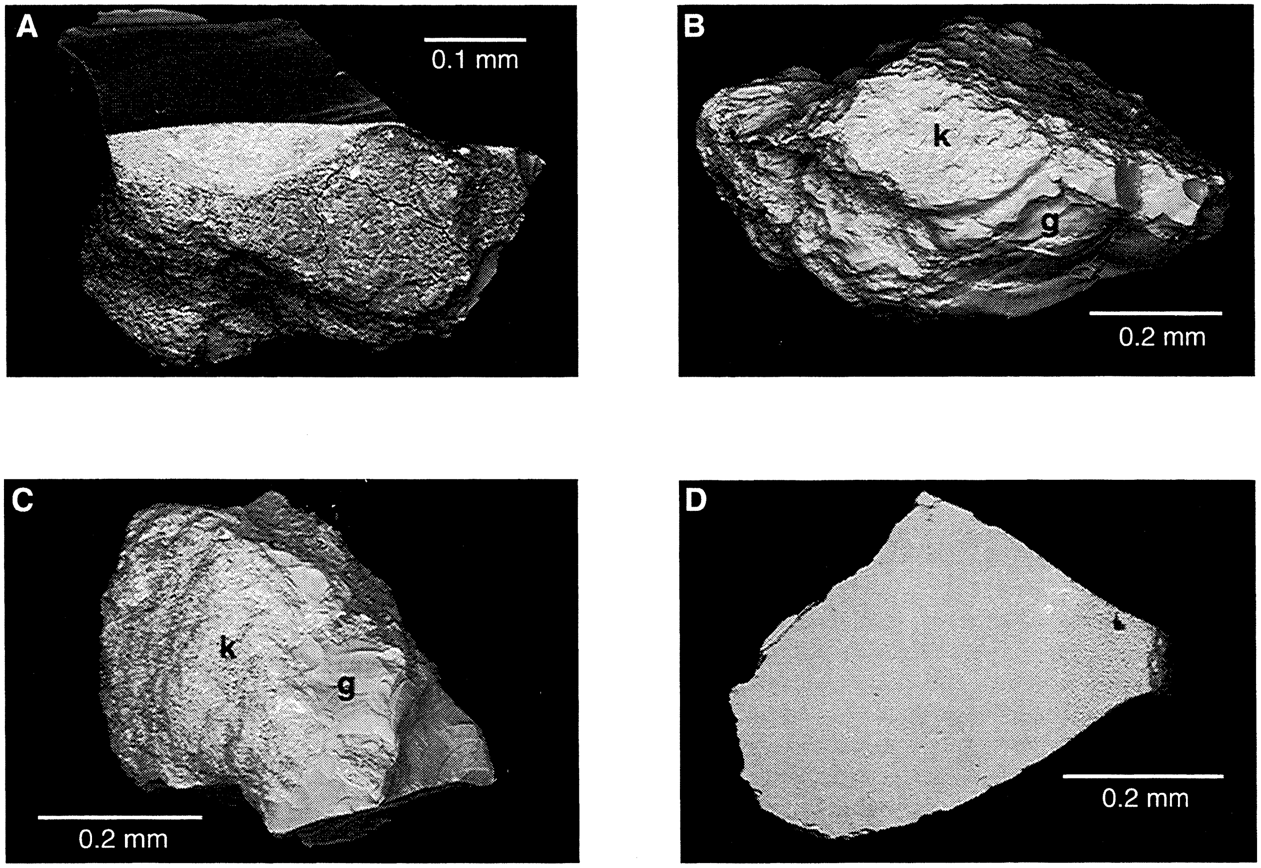
*B Close-up of A showing details of the surface.*

*C Fractured upper kelyphite (k), columnar intermediate zone (c), stippled sub-kelyphite surface (s), and fresh garnet (g).*

*D Thick kelyphite layer (k) with a sharp lower boundary. Smooth pyrope (g). Matte sub-kelyphite texture is visible on one corner (s)*

*E Clean, fresh fractures on kelyphite and garnet.*

*F Conchoidally fractured garnet grain (g) with angular edges, and rounded kelyphite surface (k) striated during glacial transport.*



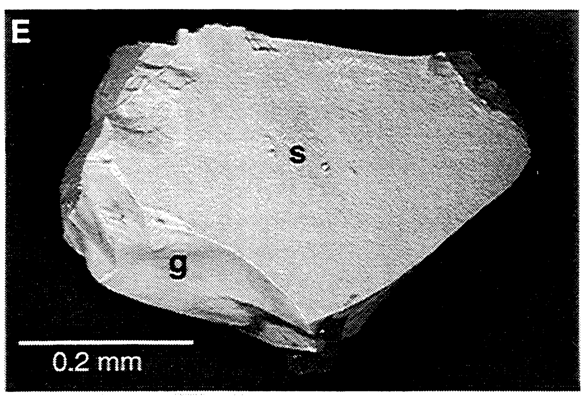




***Figure 4.** Pyrope grains glacially transported 25-30 km (Samples 93DU558 and 93DU611)*

*A Angular pyrope surface and fractured kelyphite. Some equant and rhombic crystals are visible on the kelyphite surface.*

*B Thin kelyphite (k), sub-kelyphite surface (s), and multifaceted pyrope (g)*

*C Thin, irregular kelyphite (k) and underlying pyrope (g) showing multiple conchoidal surfaces.*

*D Finely stippled sub-kelyphite surface on a pyrope crystal.*

*E Matte-to-stippled sub-kelyphite texture (s) and smooth pyrope surface (g).*

Although there are differences in amount and thickness of kelyphite with increasing glacial transport distance from source pipes, the basic appearance of the kelyphite (and fresh garnet) is remarkably similar in all the samples examined. Kelyphite surfaces appear irregular and gently rounded (Fig. 2A, 3A,B). Under high magnification, small crystals are visible within a more prevalent amorphous groundmass. Commonly, the kelyphite surface is fractured irregularly (Fig. 2E, 3E). Fractures on the kelyphite appear fresh - there are no infillings of secondary minerals that would suggest that the fracturing occurred during ascent in the kimberlite magma rather than during glacial transport. There is little difference between glacially transported grains and those that have undergone additional glaciofluvial transport. McCandless (1990) observed minor frosting, pitting and rough abrasion surfaces on grains that had undergone fluvial transport. These textures were not observed in the samples examined here. One reason may be that the fine sand grains in the esker sample tend to travel in water slurries (Dredge et al., 1994), in which grain-to grain impact is minimized. McCandless (1990) noted that grain abrasion was more apparent in gravel sizes rather than in the sand fraction.

Primary textural variation results from the mineralogy and environment of formation of the coronas. Kelyphite commonly displays a series of zones that are concentric to the pyrope surface. At Lac de Gras, the zone nearest the pyrope commonly has a microcrystalline texture or is composed of abundant small phlogopite plates (Fig. 2B,D). An intermediate zone of columns and plates, identified as pyroxene intergrown with phlogopite, forms a mesh oriented normal to the garnet surface. The outermost zone, forming the kelyphite surface of many grains, appears as coarse, irregular pyroxene prisms, with equant crystals of spinel/ chromite/titanium magnetite, and various unidentified minerals. (Fig. 3B, 4A).

## SUB-KELYPHITIC SURFACES

The contact between kelyphite and garnet is sharp. During glacial transport, some kelyphite has broken off the pyrope surfaces. Rarely, the newly exposed surface appears as glassy and flat (Fig. 2A). Much more commonly, the underlying pyrope surface has a grainy matte, stippled, or slightly pitted orange peel texture (Fig. 2E, 4D,E). Garvie and Robinson (1984) noted that this texture is formed of minute triangular or rhombic pits which are impressions of radially arranged (pyroxene) crystals. Deeper channelled or furrowed kelyphite impressions reported by Garvie and Robinson (1984), are also present on the Lac de Gras grains (not shown). Matte to pitted textures are characteristic of peridotite garnets, while rougher undulatory or flaky surfaces are generally developed under kelyphite coronas in eclogitic garnets as an impression of phlogopite grains (Garvie and Robinson, 1984). The textures seen on the Lac de Gras pyropes are similar to those for peridotitic, rather than eclogitic, garnets.

## CONCLUSIONS

This study describes textures of pyrope garnet from glacially transported material near Lac de Gras, an area of diamond exploration. This is a pilot study; although more than 1200 pyrope grains were examined, the results reported here are based on a set of only four samples. Nevertheless, it is one of the first examinations of grain shapes in material that has been glacially transported over considerable distances.

In the 0.25 to 0.5 mm size fraction, there is little change in garnet shape and surface texture with glacial transport. Kelyphite surfaces are rounded even near pipes; garnet grains are angular near pipes, and remain angular as distance of transport increases, but far-travelled glacial grains have more conchoidal surfaces than grains near pipes.

Not all kelyphite is removed during glacial transport, although the surface area and thickness of the kelyphite layer decreases as distance of transport increases. The presence of kelyphite does not imply proximity to a kimberlite source.

Kelyphite textures and zoning are similar to those reported for in situ kimberlite and stream-transported grains. In the Lac de Gras area, an inner microcrystalline layer consists of phlogopite, an intermediate columnar zone is pyroxene intergrown with phlogopite, and an outer zone is composed of coarse laths of pyroxene and equant grains of chromite/ magnetite/spinel.

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* Contribution to the Geological Survey of Canada's Slave Province NATMAP (National Geoscience Mapping) Program.

# Kimberlite indicator mineral and soil geochemical reconnaissance of the Canadian Prairie region

R.G. Garrett and L.H. Thorleifson

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## INTRODUCTION

In 1988, kimberlite discoveries in central Saskatchewan were announced. The resulting private exploration was accompanied by the initiation of government surveys designed to provide systematic data that would permit industry to place its local, more detailed, surveys into the broader regional context. The survey summarized here was proposed in 1990 in order to map regional trends in indicator mineral frequency and chemistry, to demonstrate and compare various mineralogical and geochemical exploration methods, to map drift composition as an indicator of its transport history, to test for the presence of metallic mineral deposits, and to map regional soil geochemistry in order to aid both exploration and environmental applications. Work was initiated in 1991 under the 1990-1995 Canada-Saskatchewan Partnership Agreement on Mineral Development (PAMD). Subsequently in 1992, the survey was extended to southern Alberta under the 1992-1995 Canada-Alberta Agreement on Mineral Development, and across southern Manitoba as a co-operative effort with Manitoba Energy and Mines.

The survey was limited to areas underlain by Phanerozoic sedimentary rocks, between the Rocky Mountains and the Canadian Shield (Fig. 1). The area is underlain by Cretaceous and Tertiary shale, sandstone, and gravel west of the Manitoba Escarpment. East of this feature is subcrop of a Paleozoic carbonate sequence. Throughout most of the region, these pre-Quaternary deposits are deeply buried by glacial sediments, primarily multiple till sequences commonly exceeding 100 m in thickness. At surface, and to a lesser extent in the subsurface, other Quaternary deposits such as glaciolacustrine clay and glaciofluvial sand and gravel also occur.

Till was chosen as the indicator mineral sampling medium, rather than fluvial or glaciofluvial sediments, due to its simpler transport history, more uniform composition, and greater usefulness in the study of drift provenance. Soil geochemistry was mapped using A and C horizon samples collected on a random basis from all parent materials within the extent of contiguous farmland only. Soil sampling was extended 100 km south into the U.S.A. with the co-operation of the United States Geological Survey (USGS) to facilitate trans-border geochemical mapping.

In 1991, a test set of till and soil samples was collected at 40 to 50 km intervals along transects from both Edmonton and Calgary to Winnipeg. Processing of these samples led to refinement of laboratory procedures and indicated that well defined regional trends could be mapped by low density soil and till sampling (Garrett and Thorleifson, 1993).

## SURVEY DESIGN

The project was initially designed as a low density soil geochemical reconnaissance meant to define broad regional trends in geochemical baselines. Similar surveys based on 100 x 100 km grids have been carried out by the USGS (Severson and Tidball, 1979; Severson and Wilson, 1990). Principles of low density survey design were discussed by Garrett (1983).

An 80 x 80 km grid was selected for the prairie survey, in conformance with IGCP 259 recommendations on low-density geochemical surveys (Darnley et al., 1995). These were subdivided into 40 x 40 km, 20 x 20 km and 10 x 10 km subcells and 1 x 1 km target cells were selected using a randomizing process for sampling (Garrett, 1994). This sampling design permits the spatial variability of the data to be quantified using Analysis of Variance. Rules specified procedures for selection of an alternative 1 x 1 km cell if the target could not be occupied. In each 80 x 80 km or 1600 km$^2$ cell, two sites were sampled for A horizon, C horizon, and till, if present at a nearby exposure such as a roadcut. Additional soil samples were selectively taken in order to characterize the spatial variability between more closely spaced samples.

## FIELD AND LABORATORY METHODS

Sampling was completed in the summer and fall of 1992, by staff from the Alberta Research Council, the Saskatchewan Research Council (SRC), Manitoba Energy and Mines, and the GSC. At 734 sites, till samples were obtained at depths of 1 to 2 m below surface, in order to minimize the effects of surface weathering, from the area within farmland. An additional 82 till samples were collected along road traverses which extended to the Phanerozoic limit in Saskatchewan and Manitoba. At 1273 sites in Canada and the United States, A (averaging 3-18 cm below surface) and C (averaging 40-66 cm below surface) horizon soil samples were obtained. Details of the sampling procedures are described by Thorleifson and Garrett (1993). The field data provide information on the general sampling environment and observations on the colour, texture, moisture content, and composition of the soils and till.

Geochemical sample preparation and initial recovery of <2 mm heavy mineral concentrates from the 25 litre bulk till samples was completed at the SRC. To conform with agricultural and environmental protocols the <2 mm fraction of the soils were recovered, and the <63 $\mu$m fraction of the tills were recovered for geochemical studies. Heavy mineral separations were completed by Overburden Drilling Management Ltd. of Nepean, Ontario, using methylene iodide diluted with acetone to a specific gravity of 3.2. The concentrates were then returned to the SRC and were examined for potential indicator minerals under a stereoscopic microscope. The 0.5 to 2.0 mm non-ferromagnetic concentrates were picked without further processing. The 0.25 to 0.5 mm non-ferromagnetic fraction was sorted by magnetic susceptibility using a Frantz isodynamic separator into strongly, moderately, and weakly paramagnetic fractions. Selected grains were subsequently examined by the mineralogical staff of Consorminex Inc. of Gatineau, Quebec, and minerals such as staurolite and hornblende were removed prior to grain mounting and microprobe analysis. Full details of the procedures are presented in Thorleifson and Garrett (1993).

Electron microprobe data were used to classify the minerals. Garnets were classified using a simplified version of the Dawson and Stephens (1975, 1976) and Gurney (1984) classifications. Diopsides with >0.50% $Cr_2O_3$ (Deer et al., 1982; Fipke, 1989) were regarded as chrome-diopsides. Magnesian ilmenites in every case contained well in excess of 6% MgO. Cr-spinels exceeding 60% $Cr_2O_3$ and 12% MgO were regarded as compositions comparable to those reported for diamond inclusions (Gurney and Moore, 1993). Details of the classification were discussed by Thorleifson et al. (1994).

Subsequent to the initial microprobe analyses, a number of additional studies were undertaken. Eclogitic and near-eclogitic garnets were re-analyzed at reduced detection limits. These analyses were carried out to obtain acceptable sodium data, for diamond grade prediction (Gurney and Moore, 1993), and to enhance titanium data which were being used for definition of eclogitic garnets. The G7, G9, G10, and G11 pyropes and all Cr-spinels were analyzed by proton microprobe at the University of Guelph for Ni and several other elements in order to utilize classification methods developed by Griffin and Ryan (1993). Nickel temperatures were calculated for peridotitic garnets using the equation presented by Griffin et al. (1989).

The 8 to 16 mm fraction of till samples was classified with respect to lithology using a classification based on that of Shetsen (1984). Bulk mineralogy of the 63 to 250 $\mu$m-non-ferromagnetic heavy mineral fraction was determined on the basis of identification under a stereoscopic microscope of 300 grains in an araldite mount.

Following grinding of the <2 mm fraction of the soils to approximately <150 $\mu$m, duplicates and both internal GSC and international standards were inserted in the A and C horizon soil and till sample sequences. The three sequences were randomized and renumbered prior to analysis to destroy any relationship between analysis order and spatial context. All samples were submitted for a suite of trace element analyses using Instrumental Neutron Activation Analysis (INAA) and Atomic Absorption Spectrophotometry (AAS) procedures after a total acid decomposition. In addition, the till and C horizon soil samples were analyzed for major and minor elements by X-Ray Fluorescence (XRF), and carbonate determinations were performed on the tills by the Chittick procedure. Full details are provided in Thorleifson and Garrett (1993).

## RESULTS

A total of 1253 kimberlite indicator minerals, 174 in the 0.5-2 mm fraction and 1079 in the 0.25 to 0.5 mm fraction, were reported by Garrett and Thorleifson (1993). The most abundant class consisted of 776 Cr-diopsides. A total of 206 Cr-pyropes (G7, G9, G10) including 12 G10 subcalcic Cr-pyropes, 136 titanian Cr-pyropes (G1, G2, G11), and 25 eclogitic garnets (titanian, calcic, magnesian almandines; G3, G4, G6) were initially identified. In excess of 76% of each of these mineral groups occurred in the finer 0.25 to 0.5 mm

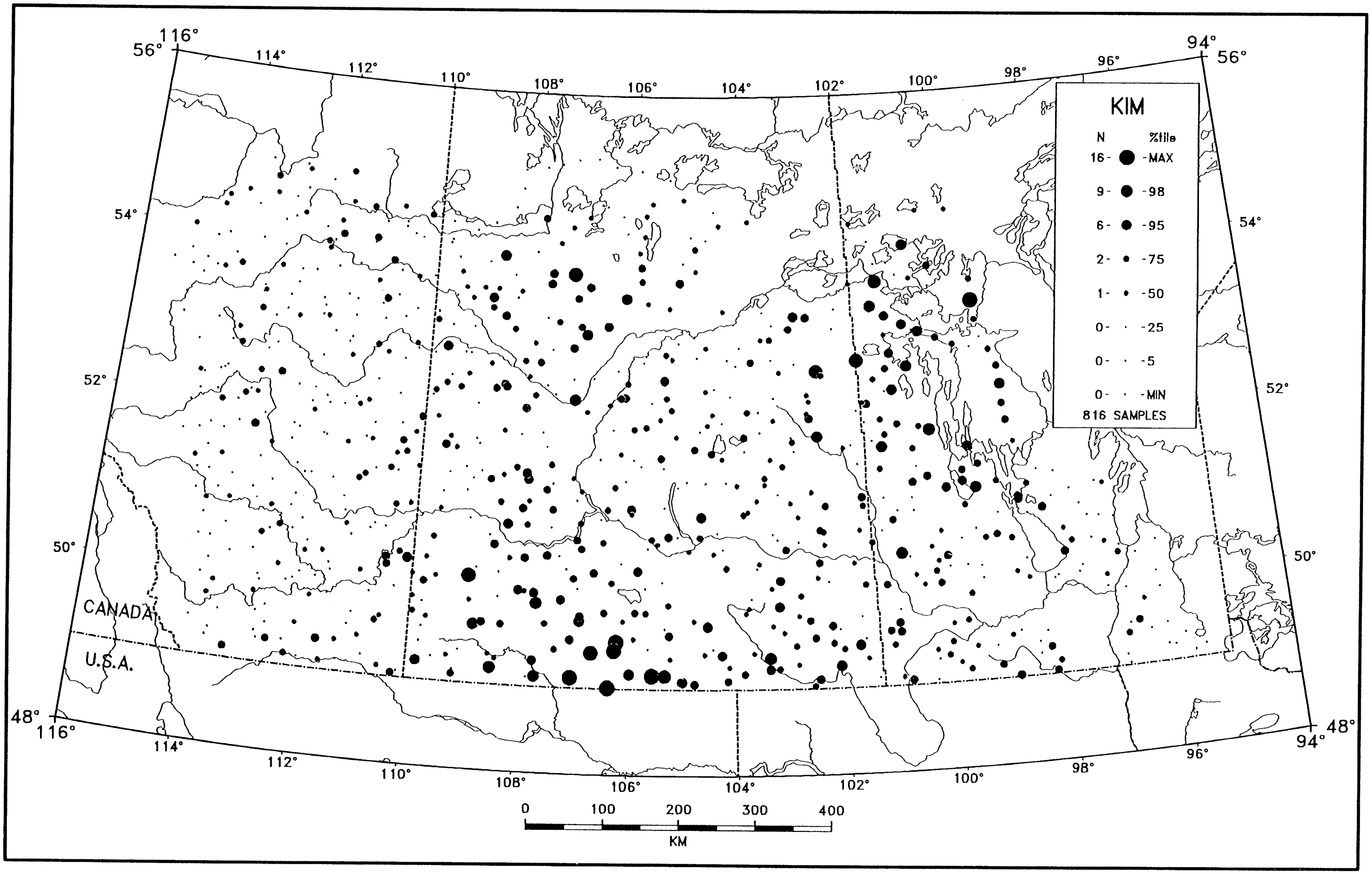


*Figure 1.* *Abundance of 0.25 to 2.0 mm kimberlite indicator minerals in 25 kg till samples.*

fraction. After the follow-up re-analysis, 76 garnets were re-classified as eclogitic using a lowered minimum $TiO_2$ value (>0.2%; Thorleifson et al., 1994). Magnesian ilmenite showed the greatest tendency to occur among the coarse grains, of a total of 110 Mg-ilmenites only 52% were obtained from the 0.25 to 0.5 mm fraction.

The indicator minerals are concentrated in southwestern Saskatchewan, west-central Manitoba, and central Saskatchewan (Fig. 1). Five of the G10 garnets, the best known predictor of diamond, were obtained in southern Manitoba. Additional occurrences of G10 garnets are scattered across southern Saskatchewan to the Alberta border. Indicator minerals in southwestern Saskatchewan show a spatial relationship to the outcrop of the Miocene Wood Mountain Formation. Therefore, the possibility of at least one stage of preglacial fluvial transport has to be considered in determining the ultimate provenance of these minerals.

After re-analysis of the 156 eclogitic and near-eclogitic garnet grains, 4 were found to contain marginally anomalous Na concentrations, >0.06% $Na_2O$. Three are from western Manitoba and adjacent Saskatchewan, and the remaining grain is from southwestern Saskatchewan.

The Ni determinations by proton microprobe on the kimberlitic garnets were used in a geothermometry study. Using the assumption of a 40 $mW/m^2$ geotherm (Griffin et al., 1989) and assuming acceptable calibration to other instruments, 13% of the garnets, mostly G11s, report temperatures above the diamond stability field (>1250 C°, >75 ppm Ni), 31% are in the diamond stability field (32-75 ppm Ni), and 56% are cooler (<950 C°, <32 ppm Ni). Of the grains in the Griffin diamond window (950-1250 C°, 32-75 ppm Ni), 80% contain <50 ppm Zr, a positive sign regarding diamond grade (Griffin and Ryan, 1993). These grains occur in Alberta, Saskatchewan, and Manitoba, with clusters around Diefenbaker Lake and Prince Albert.

Indicator minerals obtained from the till samples were transported by the continental ice sheet during the Pleistocene. Hence they were carried to the sampling sites by at least the final ice flow in the region. The dominant ice flow was towards the west and southwest, as indicated by drift composition in the region (e.g. Shetsen, 1984). This was overprinted in many areas by southeastward flow in late glacial time (Prest et al., 1968; Dyke and Prest, 1987). In most cases, the grains are likely to have undergone a complex transport history involving repeated glacial transport, and possibly interglacial and/or preglacial fluvial transport.

Useful evidence for interpreting glacial transport history was provided by lithological classification of the 8-16 mm fraction and the bulk mineralogy of the 63 to 250 $\mu$m non-ferromagnetic heavy mineral fraction. These patterns include abundant brown carbonate pebbles in southern Manitoba and southeastern Saskatchewan, a high concentration of shale clasts in southwestern Manitoba, the highest concentration of Precambrian Shield pebbles in Saskatchewan north of the Saskatchewan River, and Cordilleran-derived pebbles in the western half of southern Alberta. Heavy mineral counts indicate abundant garnet and hornblende in central Saskatchewan, as well as relatively more epidote and titanite near Winnipeg.

Geochemical data from till reveal broad-scale patterns that can be related to provenance. In addition, a number of elements exhibit single element patterns that, even though they are of low geochemical contrast, may be related to mineral occurrences.

The most notable feature in the data is a high carbonate, low As (Fig. 2), zone in the vicinity of southern Manitoba Paleozoic carbonate occurrences. However, of greater interest is the rapid dilution of the high carbonate till with Cretaceous shale material immediately to the west and southwest on the Manitoba Escarpment, and the persistence of an elevated carbonate pattern as far west as the Missouri Coteau, a northwest-trending escarpment lying west of Estevan-Regina-Saskatoon. In addition, the western carbonate domain of Shetsen (1984) is clearly visible in the data in the form of elevated carbonate values in the Calgary region. The majority of the data for the remaining elements show an inverse response, i.e., the carbonate terrain of eastern and central Manitoba are reflected by trace-element lows.

Tills down ice flow from Cretaceous shales of the Manitoba Escarpment are characterized by higher levels of Fe, Mn, V, Mo, As (Fig. 2), Sb and Zn, and to a lesser extent Cu and Ni, than the tills to the east or west. The Cd pattern is slightly different, whilst following the same general pattern, the highest values are concentrated in the northern tills in the Duck Mountain and Tisdale areas.

The line of the Missouri Coteau west of Estevan-Regina-Saskatoon appears to mark the boundary between tills of different regional geochemical compositions. The transition is most clearly seen in the data for Total Carbonate, Fe, V, As (Fig. 2), Rb, Ba and Br, and to a lesser extent with Pb, Zn and Cr. The Br data are of interest as the area east of this lineament is characterized by a zone of elevated levels.

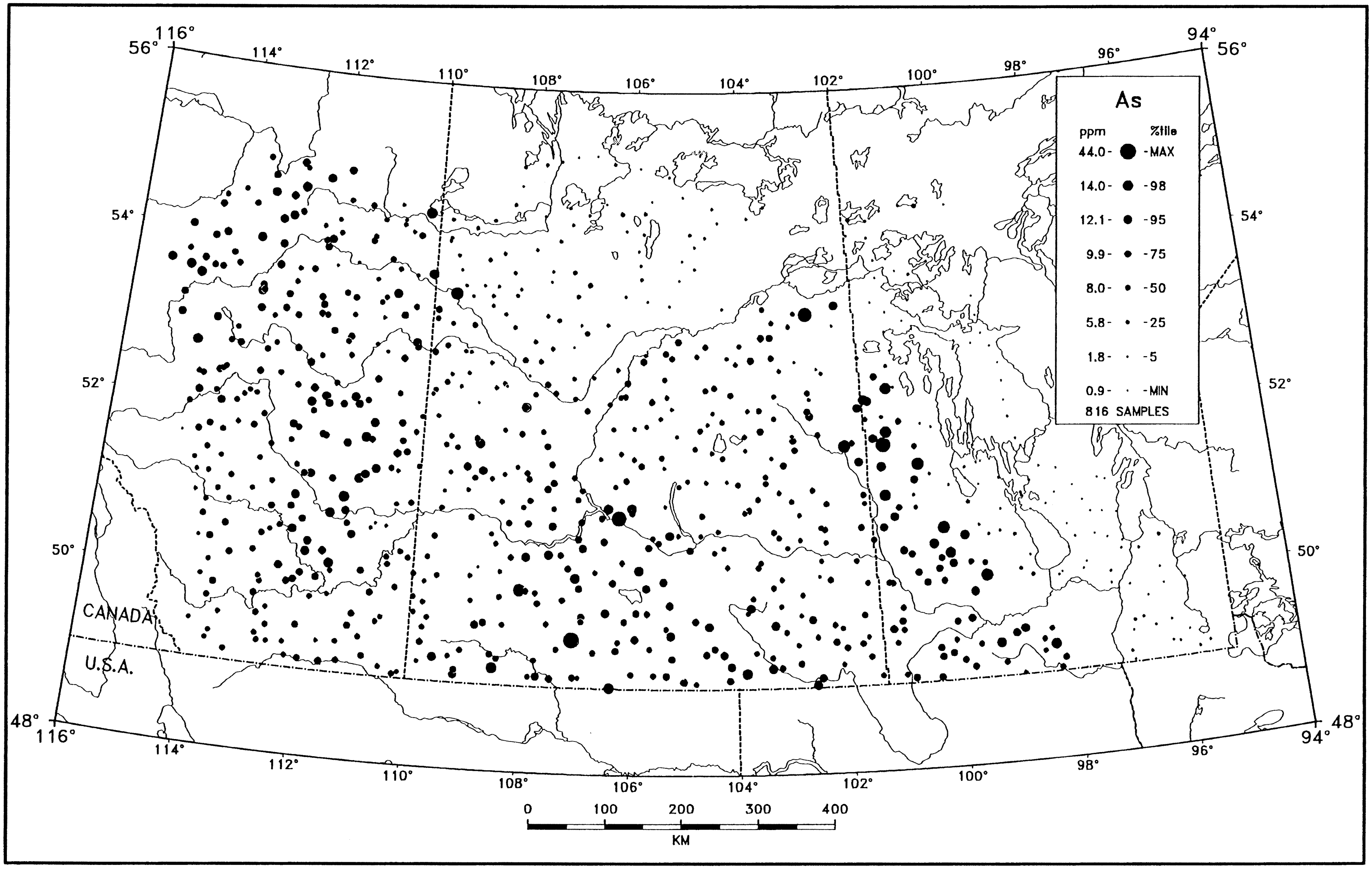


*Figure 2.* *As (ppm) in the < 63 μm fraction of till samples.*

Till north of the Saskatchewan River is characteristically low in a number of trace-elements in comparison with areas immediately to the south and/or west, e.g., V, Mo, U, Br, As (Fig. 2) and Zn. In contrast, Na and Th show the opposite, and are locally enriched; it is postulated that this may be due to the tills of the area containing a large proportion of Cretaceous Manville sediments or Precambrian Shield derived material. It is noteworthy that this area is also characterized by the highest amphibole contents in the heavy mineral separates.

A number of elements increase in level to the northwest in Alberta relative to the area to the south, e.g., Fe, V, Cu, Ni, Cr, Th, Sc, Rb and As (Fig. 2). This is likely a lithological control being exerted on the till composition. Shales are more widespread to the north and northwest, in comparison to the likely trace element poorer Tertiary sandstones and Foothills carbonates in the west.

The major task ahead is the interpretation of the chemical, mineralogical and lithological data for the tills. These data provide evidence for multiple ice advances from the northeast and east, and later advances from the north. There is also a suggestion that many of the surface tills east of the Missouri Coteau, except along the Manitoba Escarpment and in parts of the Interlakes region, have been transported long distances prior to deposition. In contrast, west of the Missouri Coteau, and particularly in the southern part of the survey area the surface tills may be of more local provenance.

The geochemical patterns in the C horizon soil data closely follow those observed for the tills. However, as soils were collected at all sites some patterns relate to the fluvial derivatives of the tills, e.g., coarser grained quartz-rich fluvial and eolian sands, etc., and clay-sized material of glaciolacustrine sediments. These two groups are characterized by generally lower trace-element levels in the coarser, quartz-rich, sediments and higher values for many elements in fine-grained parent materials. The A horizon samples have been influenced by pedological processes that have led to a modification of the C horizon patterns through concentration and depletion for different elements in different areas and soil regimes. These data are of particular interest to agricultural and environmental scientists, e.g., Cd (Garrett, 1994) and Hg (Garrett and Thorleifson, 1994).

## CONCLUSIONS

The prairie-wide kimberlite mineral reconnaissance project has provided the first systematic data on the distribution of kimberlitic and lamproitic indicator minerals, and till and soil geochemistry in the region. These demonstrate:

1) That indicator minerals are widespread across the prairie region. Local concentrations in the Prince Albert area of north-central Saskatchewan and southwestern Saskatchewan were previously known. This project brought to light the abundance of favourable indicator minerals in north-central and southern Manitoba as well as several areas of southern Alberta.

2) The till matrix geochemical, 8 to 6 mm fraction lithological classification, and bulk 63 to 250 $\mu$m non-ferromagnetic heavy mineral fraction geochemical and mineralogical data vary systematically and can be related to the bedrock provenance of the tills. These data will assist in identifying glacial flow history for the surface tills sampled in this project, the most important factor in determining the bedrock source areas of the kimberlitic indicator minerals identified in the survey.

3) Soil geochemical data from the A horizon have proven useful to baseline studies for at least Cd and Hg, and will be of future use in a variety of environmental and agricultural investigations. Availability of C horizon data paired to samples of overlying A horizon permits regional mapping of apparent surface enrichment. Further work will quantify carbonate loss from hand-augured samples, but it is apparent that many patterns indicated by the till samples collected below 1 m depth are also revealed in C horizon samples more easily collected from depths of approximately 0.5 m, which would also be suitable for refractory heavy mineral separation.

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# Geochemistry and indicator mineralogy of drift over kimberlite, Kirkland Lake, Ontario


**M. B. McClenaghan**



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## INTRODUCTION

Most bedrock in the Kirkland Lake area is covered by glacial sediments, from a few metres to 100 m thick. By applying a combination of indicator mineral and geophysical methods, several kimberlite pipes and dykes (Fig. 1) have been discovered in the Kirkland Lake area within the last 10 years (Brummer et al., 1992a; 1992b; McClenaghan, 1993). Indicator mineral tracing, however, is hindered by a lack of published information on glacial dispersal of kimberlitic debris. In 1992, the Geological Survey of Canada (GSC) began a 4 year project in the Kirkland Lake area to address this problem, to refine or develop new techniques for kimberlite exploration, and to evaluate the potential for the discovery of additional kimberlites near Kirkland Lake. Part of this project included study of the geochemical and mineralogical character of known kimberlitic intrusions and associated geochemical signatures in various sample media that are routinely used in mineral exploration (e.g. till, humus, soil, and vegetation).

## METHODS

Thirty-one overburden holes were drilled by the GSC up-ice (north), over and down-ice (south) from the C14, B30 (Nickila Lake), A4, and Diamond Lake kimberlite pipes in the winter of 1993 (Fig. 1). These pipes were chosen to examine glacial dispersal patterns in tills (C14, B30, A4) and in esker sediments (Diamond Lake). Sonic drilling was used to core the glacial sediments and bedrock because it provides a continuous, 9 cm diameter core from surface into bedrock. Kimberlite, being relatively soft, was preferentially eroded by preglacial weathering and glaciers such that the kimberlite pipe subcrop tends to be 10 to 35 m deeper than the

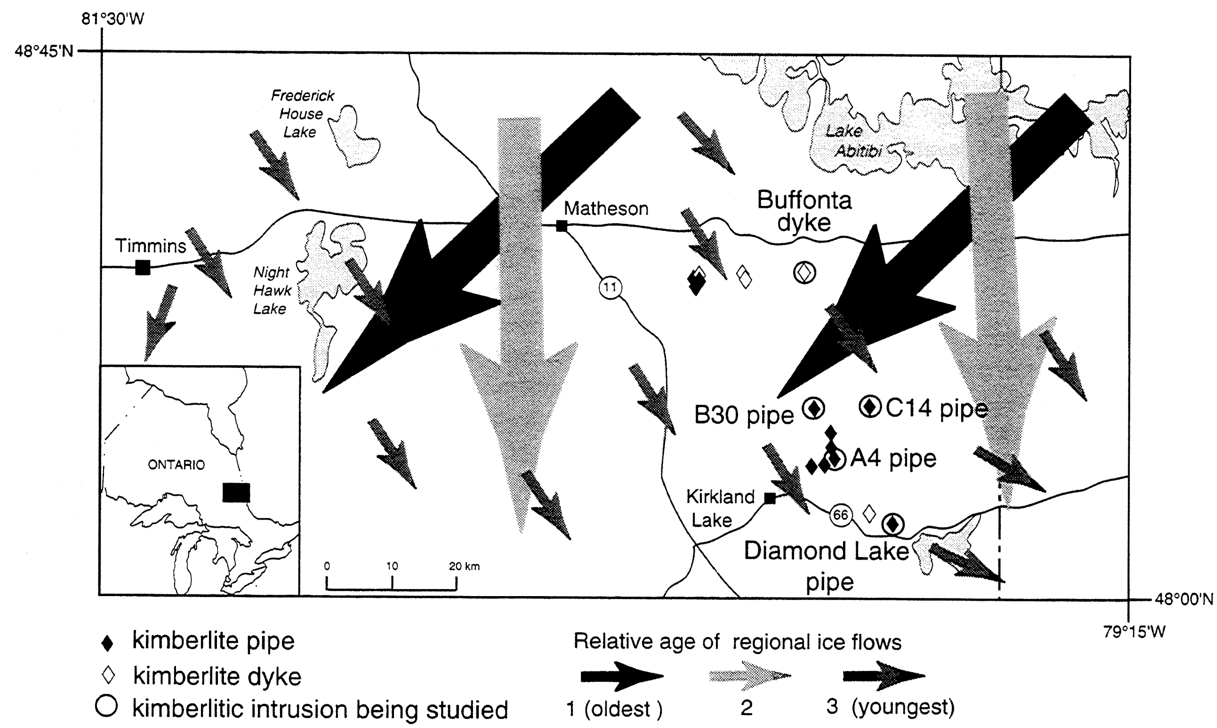


*Figure 1. Location of the C14, B30, A4, Diamond Lake, and Buffonta kimberlitic intrusions in the Kirkland Lake kimberlite field, northeastern Ontario (location of kimberlites from Zalnieriunas and Sage, 1995).*

surrounding bedrock surface. Because of this deep erosion, kimberlite pipes are covered by thick sequences of glacial drift, generally between 40 and 100 m, which contain some of the thickest deposits of Matheson Till in the region. These till sequences were sampled in detail.

Drill holes over the kimberlite pipes extended 10 m into bedrock to collect sufficient material for petrological, geochemical and mineralogical analyses in addition to U-Pb dating and Sr-Nd-Pb isotopic studies. Detailed studies of the kimberlite bedrock from the drill holes are being completed by Moser and Heaman (1994), Schulze and Anderson (1994), Vickers and Schulze (1994), Davis et al. (1996), and Katsube and Kjarsgaard (1996). Down-hole logging was performed in the winter of 1993 to document the geophysical character of kimberlites and the glacial sediments overlying them (Douma, 1995; Mwenifumbo et al., 1996). Geophysical logging was done in previously drilled and cased exploration diamond drill holes in addition to the three cased overburden holes drilled by the GSC.

Till samples were also collected over and down-ice from a small subcropping kimberlite dyke covered by 3 m of overburden on the Buffonta property in Garrison Township. Samples were collected from two backhoe trenches dug north to south across the northwestward-striking dyke to document the nature of down-ice dispersal from this small, kimberlitic bedrock source.

Soil (humus, B and C horizon) and vegetation samples were collected over and down-ice from the C14 and Diamond Lake pipes and the Buffonta dyke to determine the geochemical signatures of these media over kimberlite bedrock and to evaluate the potential use of these media in kimberlite exploration. Vegetation analyses and interpretations are reported in McClenaghan and Dunn (1995) and Dunn and McClenaghan (1996).

Bulk till, sand and kimberlite samples, weighing approximately 10 to 15 kg, collected from the sonic drill holes and the Buffonta property were processed by Overburden Drilling Management Ltd., Nepean, Ontario (ODM) using a combination of a double tabling procedure, developed by ODM and the GSC for kimberlite exploration, and heavy liquid separation to produce a <2.0 mm non-ferromagnetic heavy mineral (>3.2 SG) fraction. Gold and sulphide grains were also recovered from the heavy mineral fraction and counted because the area has potential to host gold and base metal deposits.

The non-ferromagnetic heavy mineral fraction was examined and picked by Lakefield Research, Lakefield, Ontario to recover kimberlite indicator minerals: pyrope and eclogitic garnet, Cr diopside (>1.0 wt.% $Cr_2O_3$), Mg ilmenite (>6 wt.% MgO) and diamond inclusion (DI) chromite (>62 wt.% and 12 - 17 wt.% MgO). Picked grains were analyzed at the GSC and Ontario Geological Survey (OGS) electron microprobe labs under the supervision of J. Stirling (Stirling and Pringle, 1996) and D. Crabtree (OGS).

Overburden samples were also analyzed geochemically to evaluate the potential use of till geochemistry as a less expensive exploration tool for kimberlite. Prior to indicator mineral picking, the <1.0 mm non-magnetic heavy mineral fraction was analyzed for 34 elements (INAA) at Activation Laboratories, Ancaster, Ontario. The <0.063 mm fraction of till and kimberlite samples was analyzed for Au (fire assay-ICP-AFS) plus 32 elements (ICP-ES) at Chemex Labs, Vancouver, B.C., and for Nb, Sr and Zr (XRF) at X-Ray Assay Laboratories, Toronto, Ontario. The matrix carbonate content and grain size distribution of till samples were determined at the GSC Sedimentology Lab.

## RESULTS

### *Ice flow patterns*

It is important to understand ice flow patterns for Kirkland Lake region when using indicator minerals and till geochemistry to explore for kimberlites. The oldest ice flows that crossed the region were likely towards the west-southwest and southeast (Veillette and McClenaghan, 1996). Till fabrics and distribution of Paleozoic carbonate from the Hudson Bay Lowlands are the main evidence for these older ice flows as they are not preserved in the striation record (see B30 pipe example below).

Three major ice flows (Fig. 1) are preserved in the striation record (McClenaghan et al., 1995a) and represent a counterclockwise shift in flow from northwest to southeast. The oldest of these ice flows crossed the region towards the west-southwest (220-240°) and is associated with the main phase of the Laurentide Ice Sheet. Towards the end of glaciation, ice flow shifted to the south (160-180°). As the Laurentide Ice Sheet thinned in the final stage of glaciation, ice flow shifted to the southeast (120-160°) towards the Harricana Moraine (Veillette, 1989). All three phases of ice flow are likely associated with the deposition of Matheson Till, the youngest and most ubiquitous till sheet (Veillette and McClenaghan, 1996).

### *Regional reconnaissance*

The potential for discovering additional kimberlite pipes north of the B30 and C14 pipes has been evaluated by examining archived heavy mineral concentrates from the OGS regional drift sampling program in the Black River-Matheson (BRiM) area (McClenaghan, 1991). The 0.25 to 2.0 mm non-ferromagnetic heavy mineral (>3.3 SG) fraction of 450 samples was examined to determine the number and composition of the kimberlitic indicator minerals present. The location, chemistry and quantity of kimberlite indicator minerals present in the OGS samples have been published (McClenaghan et al., 1993). Several samples contain elevated concentrations of Cr pyrope, Cr diopside, and Mg ilmenite that warrant further investigation.

### *Indicator mineral chemistry and abundance*

Electron microprobe analysis of indicator minerals from the kimberlite pipes and dyke and tills collected around them has been completed. Results will be used to characterize the indicator mineral assemblages in each pipe and the dispersal trains associated with them. A method for discriminating between regional and kimberlitic Cr diopside is being developed using several hundred microprobe analyses of grains from kimberlite and till from Kirkland Lake and Lac de Gras, NWT.

The four kimberlite pipes contain abundant indicator minerals (thousands of grains in a 10 kg sample) that have been glacially dispersed down-ice from the pipes. Till overlying the pipes contains hundreds to thousands of indicator minerals. Abundances in till decrease dramatically within hundreds of metres down-ice from the pipes, to tens of indicator minerals (Averill and McClenaghan, 1993). Background indicator mineral abundance varies, from a couple of grains north of the B30 (Fig. 2) and C14 pipes (north of the main cluster of pipes) to tens of grains north of the A4 pipe (Fig. 3), within the main pipe cluster. The relative abundance of each indicator mineral type varies significantly between the pipes. For example, the C14 pipe contains very few ilmenites whereas in the Diamond Lake pipe, ilmenites are the most abundant indicator mineral (Averill and McClenaghan, 1993). These differences in relative abundance are reflected in the indicator mineral concentrations in drift down-ice from the pipes.

Indicator minerals provide valuable information on ice flow patterns associated with till deposition. For example, the upper grey till contain tens of indicator minerals transported 800 m southwest from the B30 pipe (Fig. 2). The lowermost, brown till contains only background amounts of indicator minerals, indicating that

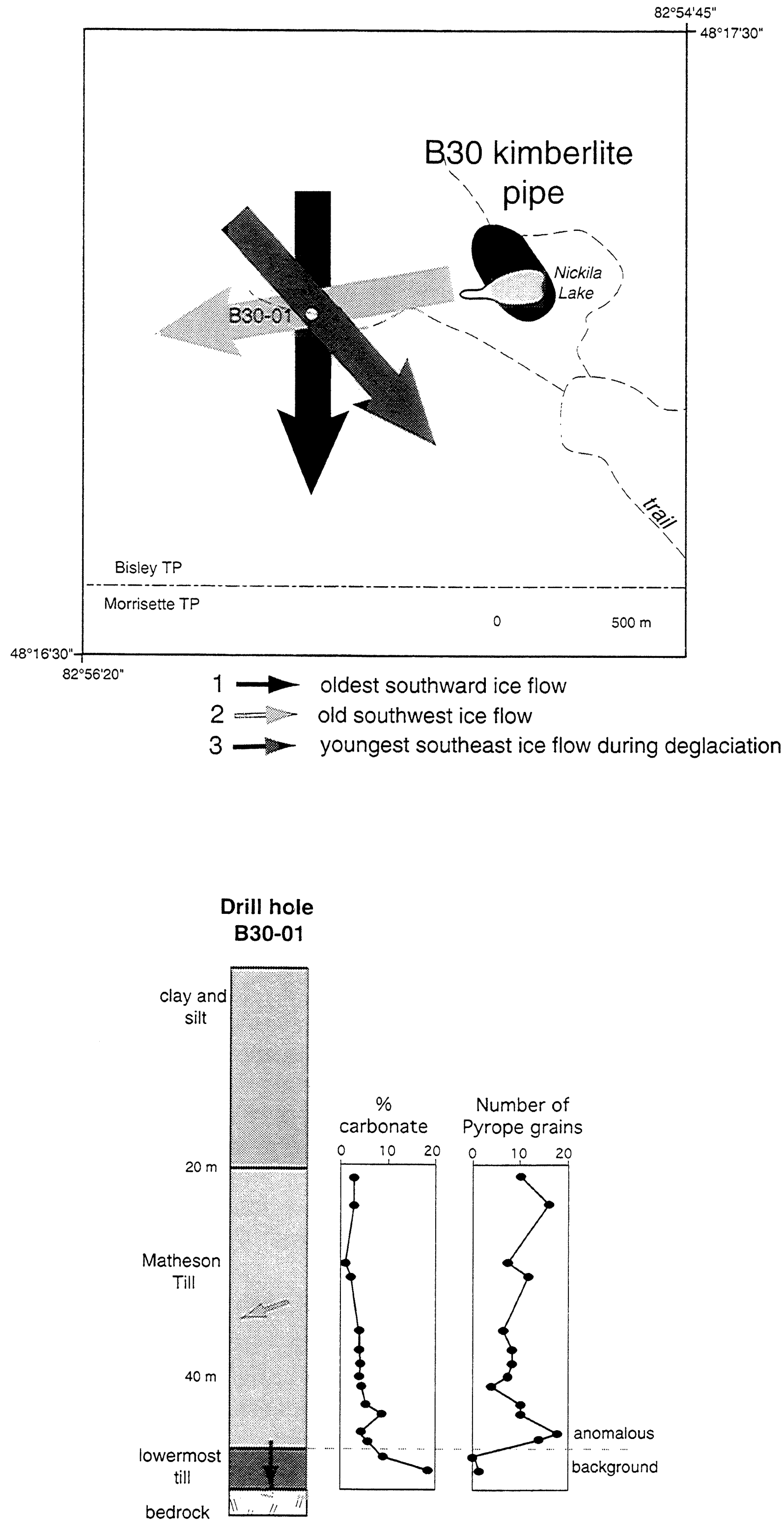


*Figure 2. Regional ice flow patterns and pyrope and matrix carbonate abundance for drill hole B30-01, 800m southwest of the B30 kimberlite pipe.*

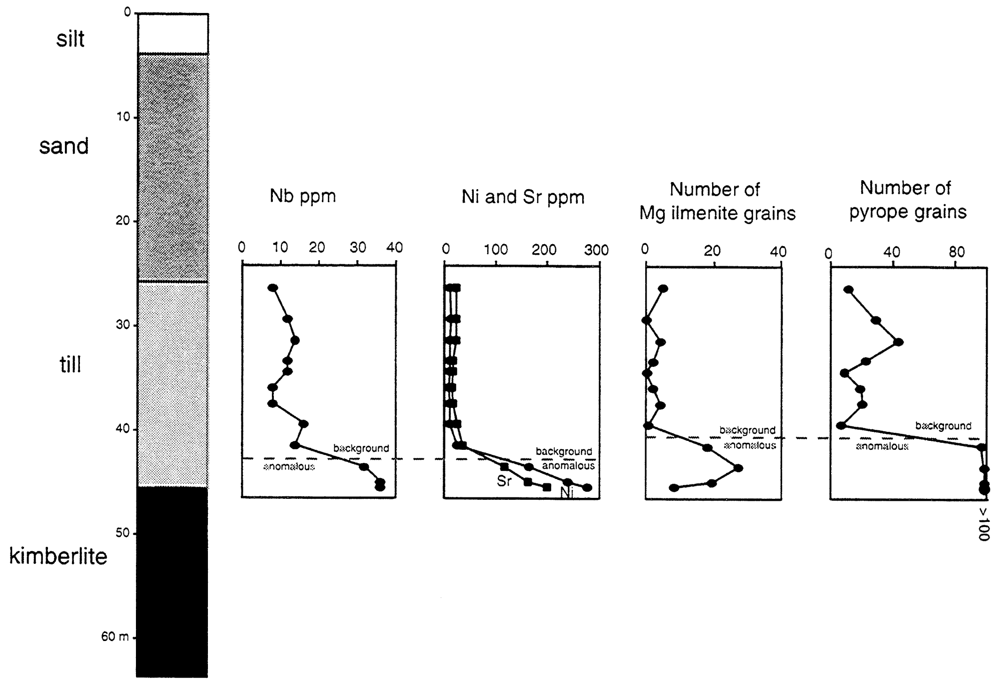


***Figure 3.*** *Nb, Ni, Sr, Mg ilmenite and pyrope abundances for drill hole A4-02, overlying the A4 kimberlite pipe.*

ice that deposited this till was not flowing southwestward. The high carbonate content of the lower till (Fig. 2) suggests ice flow was southward, bringing carbonate from the Hudson Bay Lowlands.

Results for the Buffonta dyke are summarized in McClenaghan et al. (1995b). The 1.0 m wide Buffonta kimberlite dyke contains four indicator minerals: Cr pyrope, Cr diopside, Mg ilmenite and diamond inclusion chromite. Relative to the much larger kimberlite pipes in the Kirkland Lake area, the dyke contains very few indicator minerals (tens of grains in a 10 kg sample). Because of its small size and low indicator mineral abundance, glacial dispersal from the dyke is minimal and would not be detected in a regional till sampling program.

### *Indicator mineral size and shape*

Preliminary results indicate that most Cr pyropes are found in the 0.25 to 0.5 mm size fraction of till because they break along pre-existing fractures when they are picked up by the overriding glacier. Subsequent glacial transport of garnets over several kilometres does not reduce their size or change their shape. Therefore, garnet shape and angularity are not conclusive indicators of glacial transport distance. Preliminary results have been published, along with colour photographs of garnets from kimberlite and glacial sediments down-ice, by Averill and McClenaghan (1993).

### *Till geochemistry*

Preliminary results indicate that tills with extremely anomalous indicator mineral concentrations overlying and down ice from kimberlites also contain elevated concentrations of Sr, Nb, Ba, Cr, Ni, Mg, Ca, Al, K, Ta, Th, and some REE. For example, elevated concentrations of Nb, Ni, and Sr in drill hole A4-02 (Fig. 3) correspond to anomalous levels of Mg ilmenite and pyrope. Elevated levels of trace and minor elements, however, cannot be detected as far down-ice as indicator mineral anomalies. Similar combinations of these elements should be looked for in till geochemistry data sets from past, as well as ongoing, gold and base metal till geochemistry programs in the region, as they may reflect, as yet undiscovered, kimberlites. Geochemical data for kimberlite and drift from the overburden drill holes will be published in 1996. Geochemical data for the Buffonta dyke and overlying till has been published (McClenaghan et al., 1995b).

### *Vegetation geochemistry*

Preliminary results indicate that balsam fir and black spruce trees growing over kimberlitic bedrock contain elevated concentrations of: Sr, Rb, Be, Mo, (Mn depletion); Cr, Na, Ni, Co, Cu, Ba, Cs, REE and Zn

(McClenaghan and Dunn, 1995). Results and interpretation are summarized in Dunn and McClenaghan (1996).

## ACKNOWLEDGMENTS

This project would not have been possible without the co-operation and assistance provided by current and former property holders in the Kirkland Lake area: Sudbury Contact Mines Ltd, W.A. Hubacheck Consultants Ltd., Regal Goldfields Ltd., J.E. Tilsley and Associates, Gwen Resources Ltd., and Monopros Ltd. The Ontario Geological Survey loaned archived heavy mineral concentrates from the KLIP and BRIM regional till sampling programs and provided electron microprobe analyses. Microprobe and SEM analyses were completed under the supervision of J. Stirling, G. Pringle, A. Tsai (GSC), I. Kjarsgaard, and D. Crabtree (OGS). L.H. Thorleifson is thanked for suggested improvements to the manuscript. J.J. Brummer, H.A. Lee, and C.F. Gleeson provided useful advice and information.

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# Biogeochemical studies of kimberlites

C.E. Dunn and M.B. McClenaghan

*Dunn, C.E. and McClenaghan, M.B., 1996: Biogeochemical studies of kimberlites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 219-223.*

## INTRODUCTION

Small-scale biogeochemical studies over proven kimberlite occurrences in Saskatchewan and Ontario have been undertaken to seek distribution patterns among trees (geobotany) and distinctive chemistry of tree and shrub tissues (biogeochemistry). Studies recording positive relationships between kimberlite pipes and vegetation type, abundance, and chemistry have been undertaken in Siberia (Buks, 1965), Africa (Cole, 1980), and India (Mathur and Alexander, 1983; Alexander and Shrivastava, 1984), but no previous studies of this nature are reported for Canada. The rationale for undertaking these studies is that vegetation can absorb chemical elements released from kimberlites and concentrate them in sufficient quantity to provide a distinctive biogeochemical signature. Given this knowledge, the recognition of such signatures might assist in characterizing and delineating kimberlites concealed by surficial materials.

## SAMPLE COLLECTION

### *Saskatchewan*

At the kimberlite occurrence near Sturgeon Lake in central Saskatchewan, the three most common plants are trembling aspen (*Populus tremuloides*), red-osier dogwood (*Cornus stolonifera*), and beaked hazelnut (*Corylus cornuta*). Samples of these species were collected along three profiles transecting, and to the south of, the kimberlite. Twigs (the most recent five years of growth) of each species were sampled, and the top 3 to 5 years growth of aspen stem was obtained by bending the tall spindly trees over by hand, and using pruning snips to cut 2 cm diameter stems. A total of 114 samples were collected from 32 sites. For additional details refer to Dunn (1989, 1993).

### *Northern Ontario*

Tree tissue samples were collected in the Kirkland Lake area along transects over two kimberlite pipes (Diamond Lake and C14) and one kimberlite dyke (Buffonta). The composition of forest cover is different over each kimberlite, permitting the evaluation of the geochemical response to kimberlite of three species: twigs of balsam fir (*Abies balsamea*) at Diamond Lake; twigs and outer bark of black spruce (*Picea mariana*) at the C14 pipe; and twigs of both balsam fir and pin cherry (*Prunus pennsylvanica*) at Buffonta. A total of 55 samples were collected from 34 sites. Details are given in McClenaghan and Dunn (1995).

## SAMPLE PREPARATION AND ANALYSIS

After air-drying, tissues were reduced to ash by controlled ignition at 470°C and analyzed for 35 elements by INA (instrumental neutron activation), and 30 elements by ICP-ES (inductively-coupled plasma emission spectrometry) following an aqua regia digestion. Details are given in Dunn (1993), and McClenaghan and Dunn (1995).

## RESULTS

### *Saskatchewan*

Studies in India (Alexander and Shrivastava, 1984) and Siberia (Buks, 1965) have recorded positive relationships between kimberlite pipes and vegetation type and abundance. Relatively high P levels derived from the kimberlite have promoted more luxuriant tree growth over the pipes. No similar geobotanical expression of the Sturgeon Lake kimberlite was evident, but tissue samples from the three dominant species showed a biogeochemical response with a spatial relationship to the kimberlite of enrichment in Ni, Rb, Sr, Cr, Nb, Mg, and P, and depletion of Mn and Ba (Dunn, 1993).

The Ni content of dogwood twigs was substantially higher than in twigs of the other species near the kimberlite, which is typically enriched in this element. All three species were enriched in Rb and Sr relative to sites on the surrounding Cretaceous and Quaternary deposits (Fig. 1). The Sr is probably from the carbonate, and the Rb may be derived from phlogopite within the kimberlite and mobilized as a highly soluble carbonate complex during weathering.

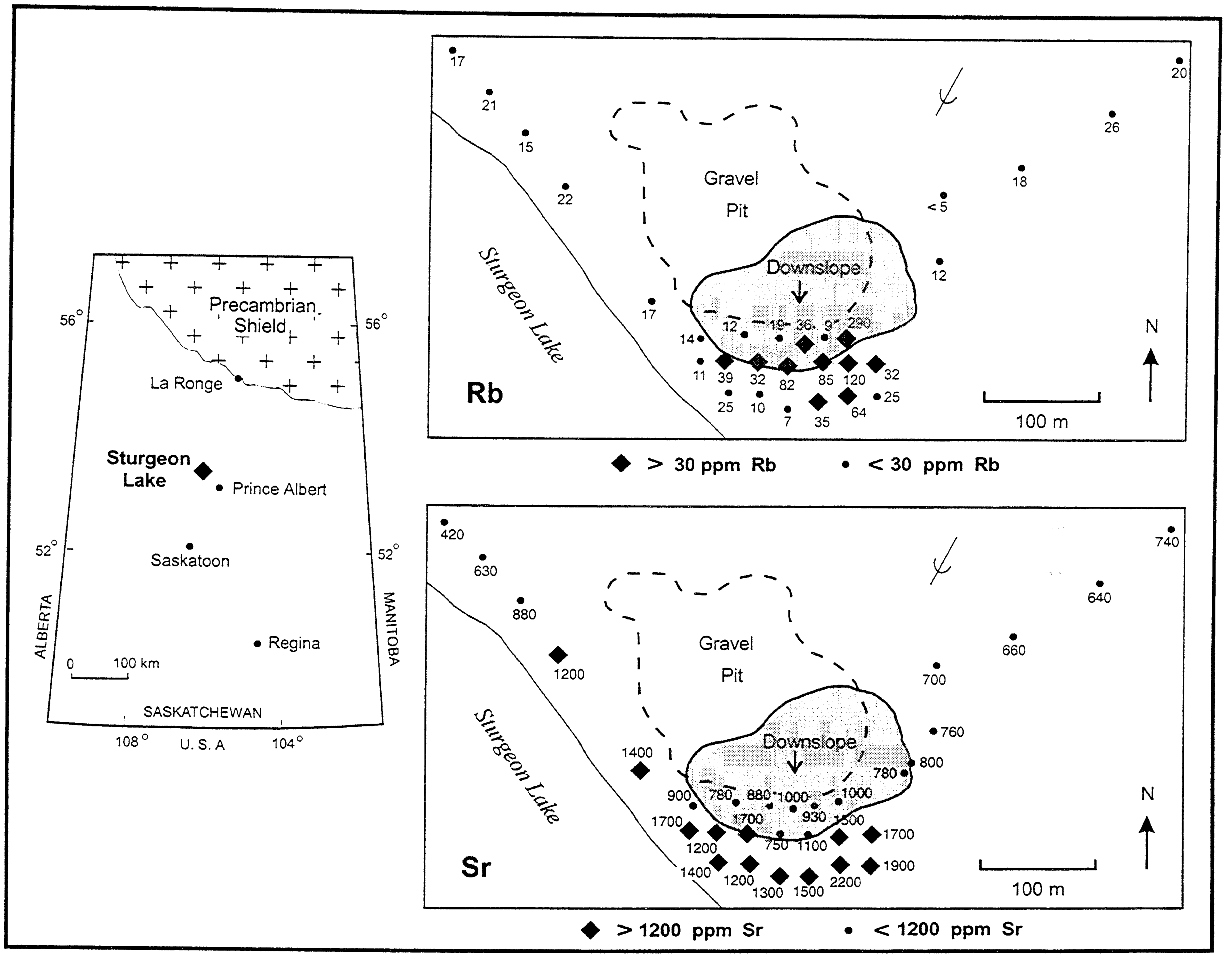


*Figure 1: Sturgeon Lake kimberlite. Distribution of rubidium in aspen tops, and strontium in aspen twigs (ppm in ash). Shaded area at south end of gravel pit indicates kimberlite.*

### *Northern Ontario*

There is no distinct geobotanical expression of the kimberlites. Biogeochemical transects across three kimberlite bodies showed that of the species and tissues collected, with some exceptions, their relative ability to accumulate base and heavy metals was black spruce twigs > black spruce bark > balsam fir twigs > pin cherry twigs (McClenaghan and Dunn, 1995).

As at Sturgeon Lake, Sr and Rb were enriched in most tissues from sites over, and adjacent to, the kimberlites (Fig. 2). Figure 3 shows expanded profiles of Sr, Rb, and Ba in tissues from spruce sampled across the C 14 pipe, demonstrating the difference in response of the two sample media and the relationships of element concentrations to the underlying lithologies.

Manganese depletion was noted in the three areas in all tissues except the spruce bark. Most other elements in the plant tissues did not show sufficient geochemical contrast to provide consistent help in locating the kimberlite bodies. The element enrichments/depletions over each kimberlite were:

a) **Diamond Lake:** *Balsam fir* - Sr, Rb, Be, Mo, (Mn depletion);

b) **Buffonta:** *Balsam fir* - Sr, Rb, Be, Mo, (Mn depletion), Au, Cr, Na, Cd, Ni, Al; *Pin Cherry* -Sr, Rb, Au;

c) **C14:** *Spruce twigs* -Sr, Rb, (Mn depletion), Au, Cr, Na, Cd, Co, Cu, Ba, Cs, REE, Zn; *Spruce bark* -Sr, Ba, Zn.

## CONCLUSIONS

Biogeochemical studies to examine the distribution patterns of a wide range of elements may help in locating kimberlite bodies at shallow depth and can be

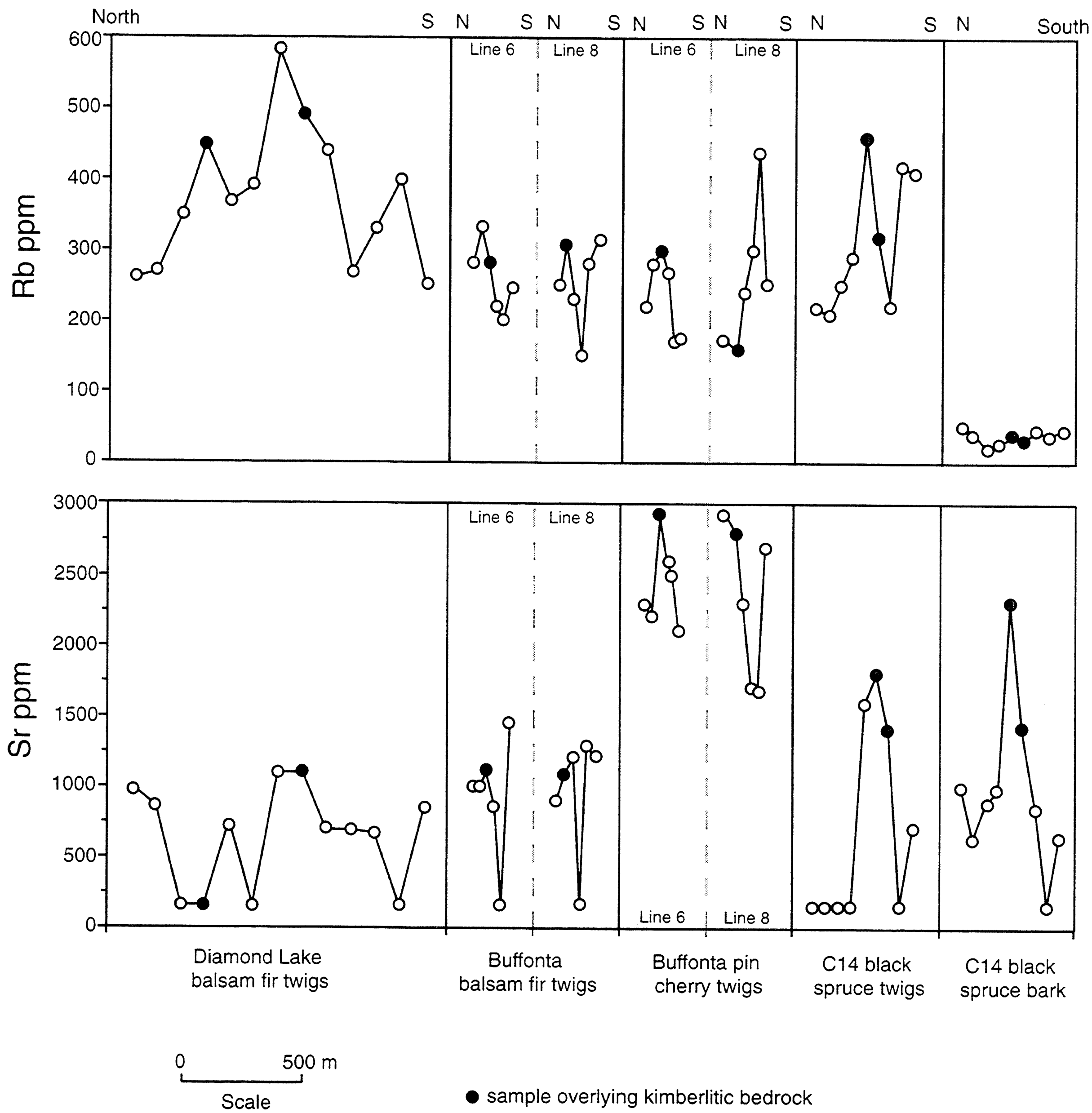


***Figure 2:*** *Kirkland Lake. Distribution of rubidium and strontium (ppm in ash) in trees from three kimberlite occurrences (Diamond Lake, C14 and Buffonta). Refer to Figure 1 in McClenaghan (1996) for kimberlite locations.*

useful in delineating their extent. Whereas the technique may be superfluous for delineating bodies that have a strong geophysical signature, it may be of value in locating those with indistinct geophysical responses, especially in a regime with upward movement of groundwaters capable of bringing kimberlite-related elements to the root systems of the plants. Of note in the surveys conducted so far is the consistent relative enrichment of Sr and Rb, thought to be derived respectively from carbonate and phlogopite, in trees and shrubs growing over and close to kimberlites. Recently, samples of dwarf birch (*Betula glandulosa*), willow (*Salix sp.*), and marsh tea (*Ledum palustre*) were collected from kimberlites in the Lac de Gras area of the Northwest Territories (B. Ward, Terrain Sciences Division). Data are pending.

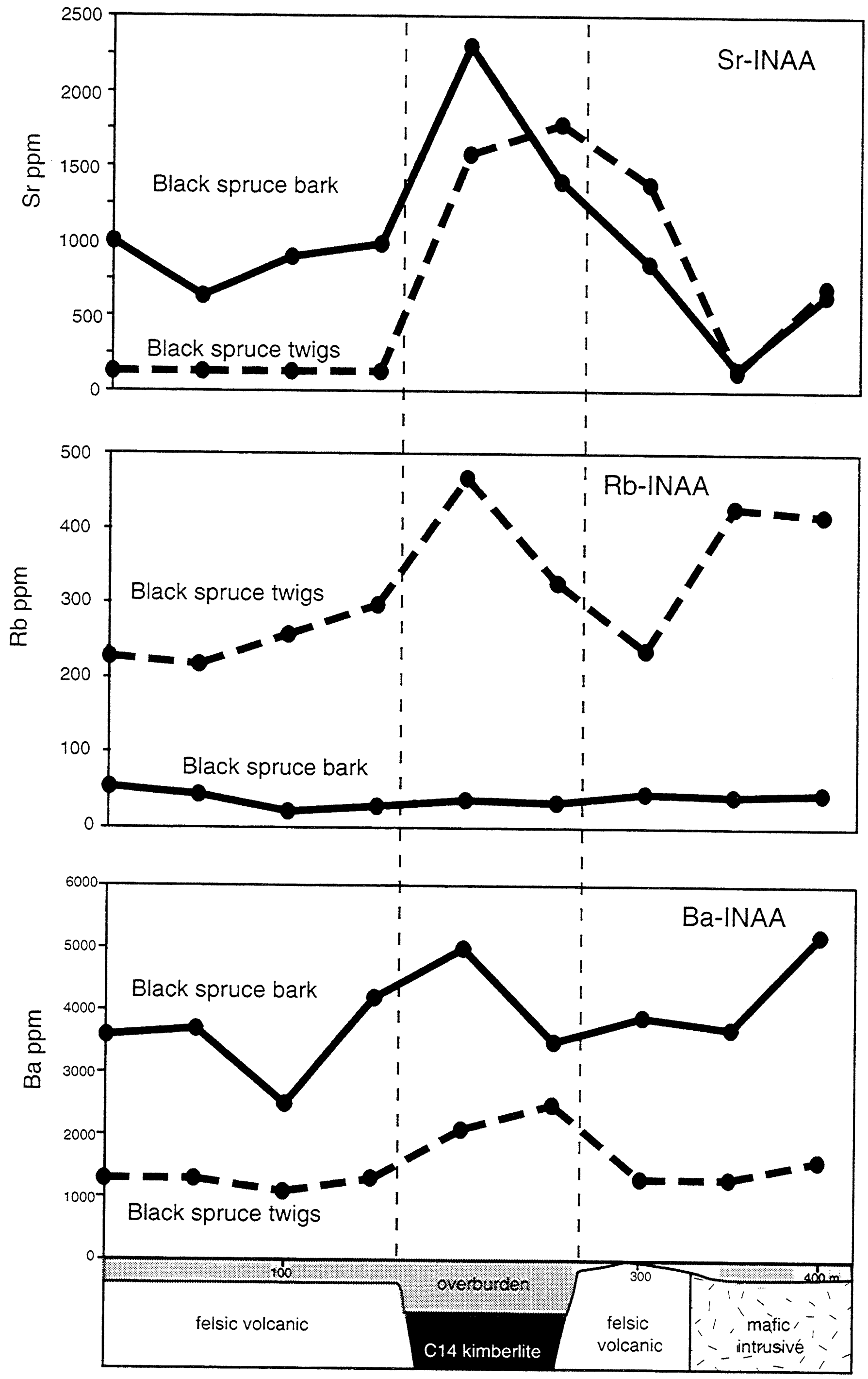


***Figure 3:*** *Kirkland Lake. Distribution of Sr, Rb, and Ba in the ash of black spruce bark and twigs.*

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# PART 3: GEOPHYSICAL EXPLORATION AND GEOGRAPHIC INFORMATION SYSTEM (GIS) APPLICATIONS

## Introduction

**K. A. Richardson**

*Richardson, K.A., 1996: Part 3: Geophysical exploration and Geographic Information System (GIS) applications - Introduction; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 225-228.*

The published record of using geophysics in the search for diamonds goes back 65 years to Stearn's (1930) report on a geomagnetic survey to locate diamond-bearing intrusions in Arkansas, U.S.A. Although ground and airborne magnetic surveying continue to play a fundamental role in diamond exploration, many other geophysical methods are now being used to define broad prospective terranes, and to detect and delineate diamond-bearing bodies.

Magnetic surveys were used extensively in Yakutia, Russia, in the 1950s and 1960s (Atkinson, 1989) and the use of aeromagnetic surveying, in conjunction with colour air photography in prospecting for "diamond deposits *in situ*" on the Siberian platform was reported by Barygin (1962). Other geophysical methods included in Gerryts' (1970) review of geophysical practice in diamond prospecting worldwide are gravity surveys (USSR and Africa), and resistivity surveys (Africa), in addition to magnetometer surveys (USA, USSR, and Africa). Gerryts concluded that gravity and resistivity methods were useful only on a local scale, and magnetics offered the only airborne method that produced positive results.

In 1970, Burley and Greenwood (1972) carried out a number of geophysical investigations (magnetic, resistivity, gravity, electromagnetic and seismic) to determine the most suitable exploration methods for diamond bearing kimberlite in Lesotho, Africa. They found that magnetic and resistivity methods were most useful for ground follow-up; gravity and electromagnetics could be useful in defining weathered kimberlites; gravity surveys produced small anomalies that could be used to locate buried pipes; and seismic refraction was unlikely to detect buried kimberlite bodies within the basalt of the study area.

By the end of the 1970s, when Macnae (1979) reviewed geophysical techniques for kimberlite prospecting, airborne electromagnetics (AEM) had proved effective to detect pipes in South Africa. He presented several examples from ground magnetic and resistivity surveys, and airborne EM and magnetic surveys, over two areas in South Africa, and concluded that AEM is a particularly important complement to aeromagnetics in areas of deep weathering. Smith (1985) discussed the use of ground and airborne EM at Ellendale, Western Australia, where AEM responses upgraded very weak magnetic responses.

A decade later, Atkinson (1989) reviewed the state of diamond exploration, and although many ground geophysical techniques had proved effective to delineate the boundaries of kimberlite pipes, aeromagnetic surveys still provided the only rapid, inexpensive reconnaissance exploration tool. The combination of radiometrics with magnetics, however, was the most widely used airborne geophysical method of diamond exploration in Australia.

There is little published on the results of radioactivity surveys applied to diamond exploration. Macnae (1979) referred to work by Paterson et al. (1977) that found ground radiometric techniques to be of limited use in Lesotho, but anomalies were too small for airborne reconnaissance detection. Although Drew (1986) described an aeromagnetic/radiometric survey of the area of the AK1 lamproite pipe in Western Australia, in which the radiometric data failed to locate a recognisable response over the pipe, Jenke and Cowan (1994) described a successful application of airborne radiometrics in mapping black soil cover in the Ellendale lamproite province, where a number of diatremes were detected. Small airborne gamma-ray spectrometer surveys were flown by the Geological Survey of Canada in the Sturgeon Lake area, Saskatchewan (Hetu, 1989) and at Lac de Gras, Northwest Territories (Shives and Holman,1995). Neither of these test surveys produced detectable gamma radiation responses from known kimberlites, but the gamma ray results, with accompanying magnetometer and VLF-EM data, did provide useful information on the bedrock and surficial geology. Macnae (1995) provided a recent and comprehensive review of the role of exploration geophysics in locating primary diamond deposits, with emphasis on integrated airborne EM and magnetic surveys. He referred to reports of anomalous radioactive responses for kimberlites in India and Siberia, and also gave an example of a seismic reflection profile over a Yakutian kimberlite.

Brummer's (1978) review, "Diamonds in Canada", mentioned the use of regional geophysical anomalies as broad indicators of areas favourable for kimberlite occurrences, but diamond exploration in Canada up to that time was based largely on glacial drift and stream sediment sampling for indicator minerals, rather than on geophysical surveying. When a later version of his paper was published (Brummer, 1984) high-resolution airborne magnetic and electromagnetic methods were increasingly applied to the search for diamonds in Canada, and the "Great Canadian Diamond Rush" of the 1990s, has seen extensive use of these airborne geophysical techniques, as well as ground geophysical methods for follow-up and location of drill targets.

The following 8 papers in this volume outline work by the Geological Survey of Canada involving geophysics and petrophysics, remote sensing, and the application of geographic information systems in searching for diamonds in Canada. Keating et al. (1996), presented the status of the National Aeromagnetic Data Base. Systematic aeromagnetic survey data, acquired since 1947, cover much of Canada, and provide a database that continues to be instrumental in the discovery of kimberlite fields. They also illustrated the extensive gravity coverage that can contribute on a regional scale to assess the potential of an area for diamond genesis and preservation (Morgan 1995). In a second paper, Keating (1996) presented an automated technique for identifying aeromagnetic anomalies that could be caused by kimberlite pipes, and showed how the combination of AEM data with magnetic data improves the technique.

Katsube and Kjarsgaard (1996) and Mwenifumbo et al. (1996) discussed physical property measurements of kimberlites, as measured in the petrophysics laboratory and in situ using borehole geophysical methods. The laboratory data show differences between hypabyssal, diatreme, and crater facies kimberlites. Borehole measurements of electrical conductivity and gamma radioactivity at Fort à la Corne enabled classification of kimberlite in one drill hole into five phases or separate eruptions.

Morgan (1995) discussed an exploration strategy based on the search for lithospheric conditions favourable to diamond genesis and preservation. Key characteristics for target areas are low surface heat flow over 300 to 400 km diameter regions, lithospheric thickness of >150 km, relatively deep lithospheric electrical conductors, slow seismic velocities, and great depths to the seismic low velocity zone. Jones et al. (1996) gave an overview of experiments from Canada's LITHOPROBE program (i.e., a new generation of teleseismic and deep-probing EM techniques), which contribute to the study of the continental lithosphere, and provide information about the age, thickness, and internal geometry of the upper mantle that optimizes strategies for regional kimberlite exploration.

On a more local scale, Gendzwill and Matieshin (1996) completed a reflection seismic investigation over a known kimberlite body at Fort à la Corne, Saskatchewan, as a test to define the geometry and structure of the body. Whereas refraction seismic has been used to map the edges of a kimberlite in South Africa (da Costa, 1989), only recently have applications of seismic surveying to kimberlite mapping (Erkhov et al., 1993) been reported ( Macnae, 1995). Gendzwill and Matieshin's work on a Fort à la Corne area kimberlite amplified the geological interpretation from drilling data, and indicated the shape, extent, and stratigraphic relationship of the body.

Rencz et al. (1996) presented an interesting application of thermal imagery from LANDSAT to locate kimberlite pipes. Lakes in the vicinity of Lac de Gras that have lower water temperatures relative to other lakes of similar size were identified from the satellite imagery. The lower temperatures indicate greater water depths, and this is attributed to deep glacial excavation of easily eroded kimberlites.

In addition to the variety of geophysical measurements from airborne and ground surveys, the database that is used in a diamond exploration program may include bedrock and surficial geological observations, geochemical survey data, and remotely sensed airborne and satellite imagery. The management, integration, analysis and interpretation of such a complex array of information, obtained from different sources and in different forms, presents a challenge to the user. In the final paper of this volume, Bowie et al. (1996) describe one technique, a Geographic Information System, that provides an efficient means of handling multidisciplinary data. The use of this technology in the Lac de Gras region demonstrates how GIS provides new ways to process large arrays of information, and how it might be applied in the search for diamonds.

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# The National Aeromagnetic Data Base

**P. Keating, J. Tod, and R. Dumont**

*Keating, P., Tod, J. and Dumont, R., 1996: The National Aeromagnetic Data Base; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 229-232.*

Much of the Canadian landmass is now covered by aeromagnetic surveys. Since the first GSC survey, in 1947, surveys have been typically flown at a mean terrain clearance of 305 m and a line spacing of 800 m. Flight lines were originally oriented north-south or east-west, however in more recent surveys they have been flown perpendicular to the main geological trend in the contract areas. Detailed surveys, mainly magnetic vertical gradiometer, have also been flown in specific areas as part of Federal-Provincial Mineral Development Agreements. In most cases the line spacing is 300 m or less and the flight height 150 m or less. Detailed gradiometer surveys have been flown in areas such as the Maritimes and the Flin Flon belt; elsewhere such as in the Kirkland Lake region, combined magnetic and electromagnetic surveys are available. A catalogue of products and sevices can be obtained from the Geophysical Data Centre.

The first 1:5 000 000 scale magnetic anomaly map of Canada was published in 1967. To help regional studies, the compilation of 1:1 000 000 scale maps was later initiated. Although modern surveys are digitally acquired and processed, older surveys had to be digitized from the original contour maps. The appropriate Definitive Geomagnetic Reference Field or International Geomagnetic Reference Field was subtracted and the data were gridded with a simple inverse distance formula. Discontinuities at survey boundaries were adjusted.

In 1987, the Ontario Geological Survey initiated a project to recompile the magnetic data so that maps could be plotted at a scale of 1:50 000. Small digitizing errors and survey adjustment problems that could not be detected at the larger compilation scales were corrected in order to produce a homogeneous 200 x 200m magnetic grid of Ontario. The project has now been expanded to cover the whole country. Surveys are merged and levelled to a common datum, data digitized from contour maps are edited and corrected. The project is well advanced and all of Eastern Canada, Manitoba, Saskatchewan, and the Northwest Territories now form one homogeneous data set. The British Columbia and Alberta data will be incorporated during the next year. Surveys flown at constant barometric altitude over these regions will have to be computationally draped to simulate mean terrain clearance so that adjacent surveys can be matched.

All the levelled data are stored in the National Aeromagnetic Data Base (NADB) as profile data. Derived products, such as contour maps and/or colour maps are easily produced on demand. The data base also archives magnetic and electromagnetic data from detailed surveys flown for the GSC and other government agencies.

Aeromagnetic data have been instrumental in the discovery of major kimberlite fields in Canada. Although the Point Lake kimberlite pipe in the Lac de Gras area of the Northwest Territories has only a small negative magnetic signature, it can still be detected by regional aeromagnetic surveys flown at a line spacing of 800 m and height of 300 m. The Fort à la Corne kimberlite field in Saskatchewan is easily identified on 1:50 000 aeromagnetic contour maps; moreover, Strnad (1991) noted that only five days of interpretation were necessary to outline this entirely new kimberlite field from the published aeromagnetic maps. Kimberlite pipes are also identified from detailed surveys; an example is the Kirkland Lake magnetic/electromagnetic survey where kimberlite pipes have both magnetic and conductivity responses (Keating, 1996).

Figure 1 shows the regional aeromagnetic coverage of Canada available from the Geophysical Data Centre. Data are available as profile data: latitude, longitude and magnetic values measured along the flight lines and/or gridded data, normally using 200 m pixels. Figure 2 shows the horizontal gradient of the gravity anomalies of Canada and illustrates the regional gravity coverage of the country. These maps helped delienate the major tectonics elements in Canada (Gibb and Thomas, 1976; Hoffman, 1989).

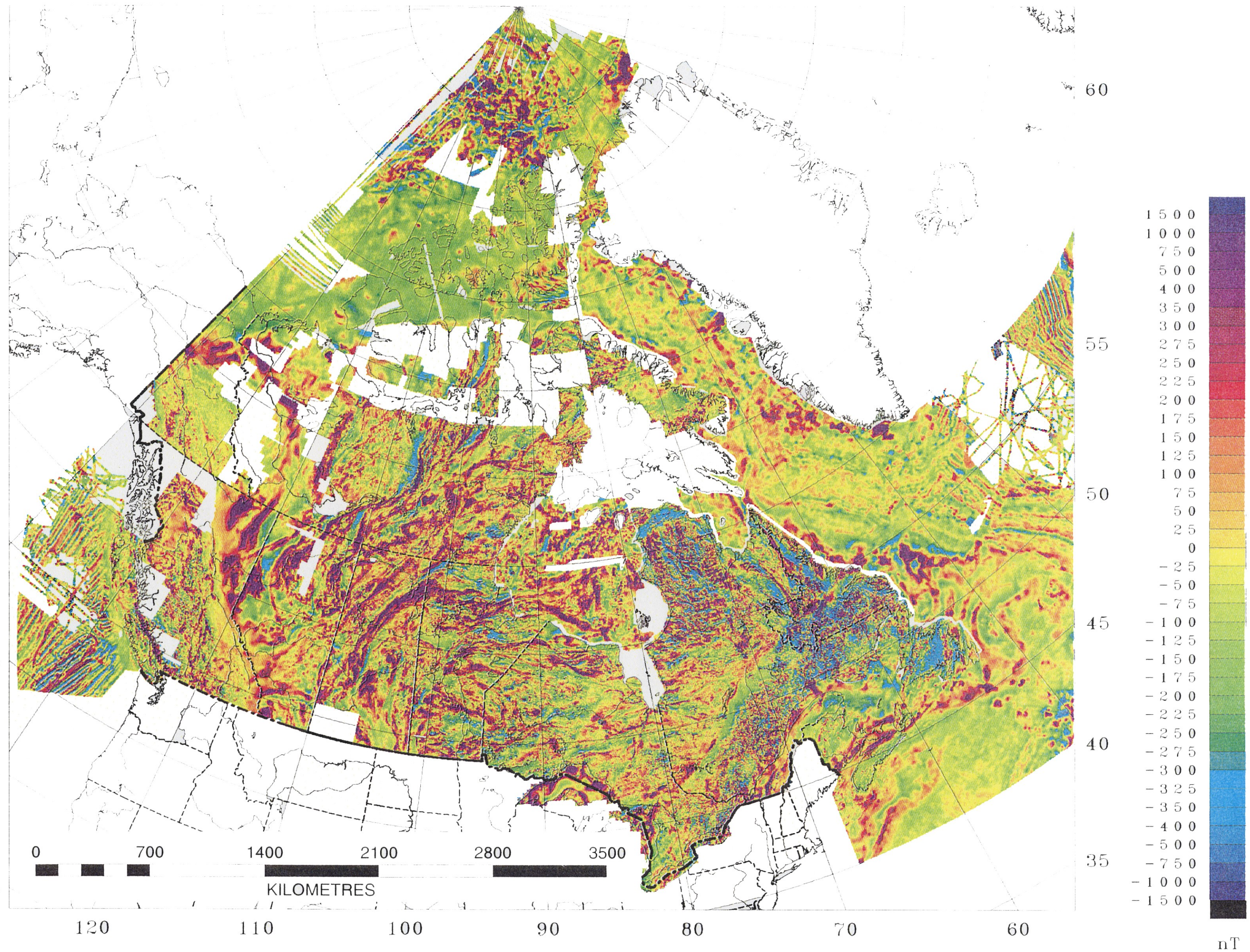


*Figure 1. Magnetic anomaly map of Canada.*

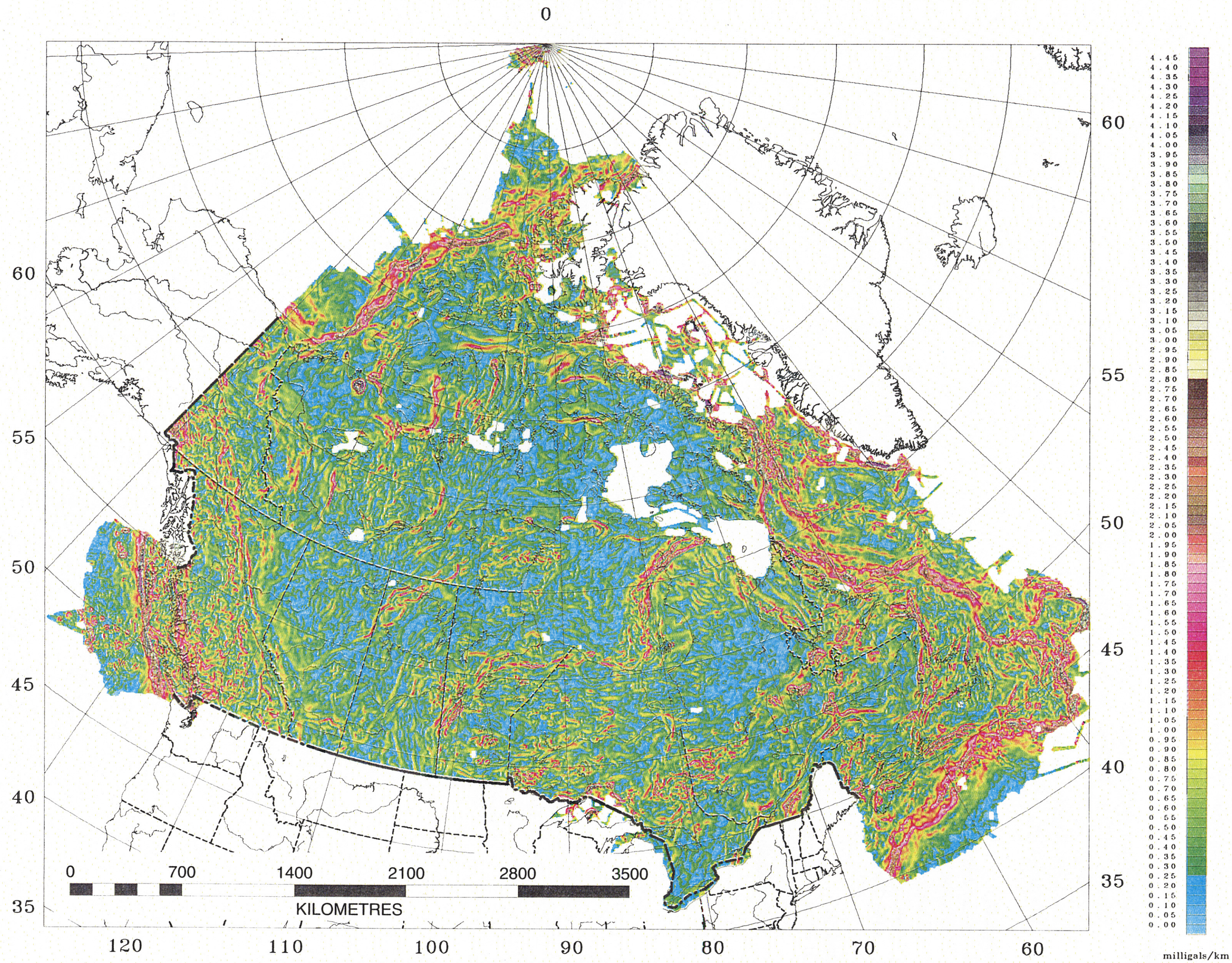


*Figure 2. Horizontal gradient of the gravity anomaly map of Canada.*

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# Kimberlites and aeromagnetics

**P. Keating**

*Keating, P., 1996: Kimberlites and aeromagnetics; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 233-236.*

## INTRODUCTION

Most of the world's known kimberlites have distinctive geophysical signatures, generally magnetic and/or electromagnetic. In the Canadian Shield, their aeromagnetic signature is often a circular anomaly. Target selection from aeromagnetic data is, therefore, based on the identification of roughly circular magnetic anomalies. A simple technique based on a linear regression between the theoretical anomaly of a vertical cylinder and the magnetic data within a moving window is an effective tool to identify potential kimberlites. Target selection can further be refined since most kimberlites are conductive.

## GEOPHYSICAL SIGNATURE OF KIMBERLITES

Most of the world's known kimberlites, a major source of diamonds, have distinctive geophysical signatures, generally magnetic and/or electromagnetic (EM). Macnae (1979) reviewed the various geophysical characteristics of kimberlites in South Africa and presented simple interpretation models. The suggested identification criteria are: 1) a bullseye appearance on the contoured magnetic and/or EM data; 2) the EM anomaly, if present, should appear to have a shallower source than the magnetic anomaly; 3) either the magnetic or the EM anomaly should be of large amplitude; 4) pipes tend to be clustered within a few kilometres and there should a coincident visible feature on areal photographs. Many kimberlite pipes are approximately circular in surface exposure. The diatreme zone is carrot-shaped in cross-section; its size diminishes with increasing depth and grades downward into a root zone with feeder dykes (Mitchell, 1991; Brummer et al., 1992).

In the Canadian Shield, the aeromagnetic signature of most kimberlites is a circular anomaly. However, at ground level, the anomaly is often more complex; it can be slightly elongated and even have internal highs. Typical examples are presented by Macnae (1979) and Brummer et al. (1992). Magnetic susceptibility can be variable, even within a single pipe. Remanence may or may not be present; for instance, many of the kimberlites from the NWT have a reversed magnetization (Reed, 1993a), while some do not have significant magnetic responses. Magnetic susceptibility of the Kirkland Lake area (Ontario) kimberlites range from 0.2 to 2.5 x $10^{-3}$ cgs units (Reed, 1993a). Reed (1993b) estimated, from model studies, that in the Northwest Territories their magnetic susceptibility is no higher than 0.6 x $10^{-3}$ cgs units. Macnae (1979) reported that in South Africa the apparent source dimension of the magnetic anomalies is in the range of 100 to 1600 m. Reed (1993b) indicated that in the James Bay Lowland diameters range from 50 m to 1.5 km. The median pipe diameter in the Fort à la Corne (Saskatchewan) kimberlite field is about 350 m and the maximum diameter < 1 km (Strnad, 1993). In the Kirkland Lake area, the diameter of the known kimberlite pipes is a few hundred metres (Brummer et al., 1992). The magnetic anomaly amplitude is proportional to the magnetic susceptibility and the areal size is proportional to the pipe diameter. Target selection from aeromagnetic data is, therefore, based on the identification of roughly circular anomalies.

Brummer et al. (1992) showed how aeromagnetic surveys have been able to successfully identify previously unknown kimberlites in the Kirkland Lake region in Northern Ontario. Their approach was to look for small, isolated circular anomalies. Paterson et al. (1991) have successfully used an automated magnetic interpretation technique, Euler deconvolution (Reid et al., 1990), to select circular magnetic targets for kimberlite exploration. Such automated techniques are most useful to process data from large regions; moreover, all the Canadian aeromagnetic data are available digitally.

## AUTOMATED TECHNIQUES

A simple regression (Davis, 1973) between the theoretical magnetic anomaly due to a vertical cylinder (Singh, 1978) and the observed magnetic field within a moving window has been found to be an effective tool to identify circular magnetic anomalies caused by kimberlite pipes (Keating, 1995). The diameter of the cylinder determines the size of the pipes being searched for. The absolute value of the correlation coefficient and the

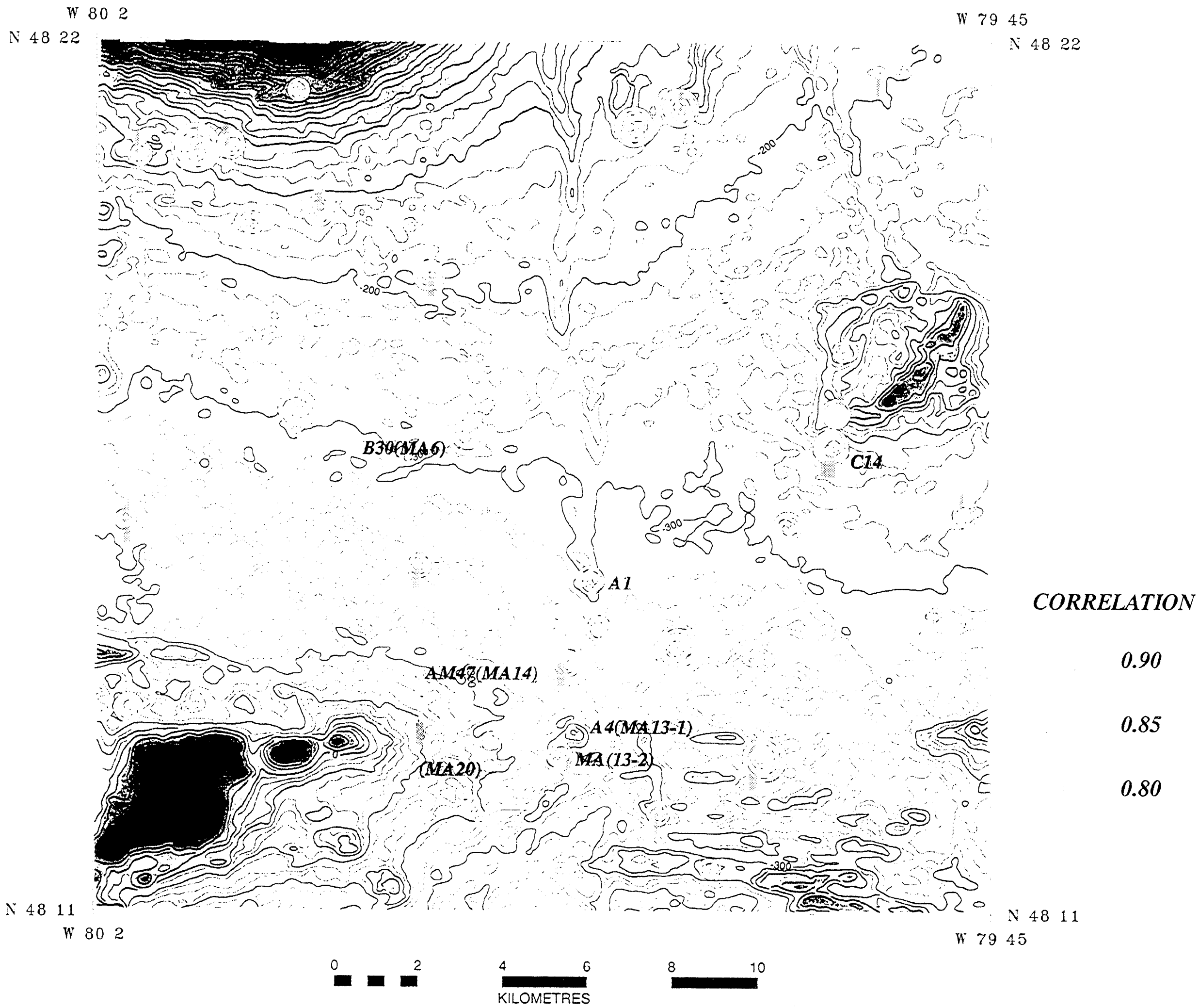


***Figure 1.*** *Magnetic anomaly map of the Kirkland Lake area, northern Ontario. Location and numbering of the kimberlite pipes is from Figure 2 of Brummer and al. (1992). Circle diameter is proportional to the absolute value of the correlation coefficient obtained from the linear regression between the theoretical anomaly of a vertical cylinder and the observed magnetic data within a moving window. Contour interval is 25 nT (nanoteslas).*

standard error of fit of the regression are used to screen the possible targets. Plotting the solutions on a conductivity map of the studied area further improves the technique, since most kimberlites have a positive conductivity response. The methodology was tested in the Kirkland Lake region, since known kimberlites there are well documented. The data are from a combined magnetic and electromagnetic survey flown as part of the 1991-1995 Canada-Ontario Subsidiary Agreement on Northern Ontario Development [NODA] (Geological Survey of Canada, 1993). Flight lines are oriented north-south and the magnetometer sensor located at an average height of 70 m above ground. The electromagnetic data (time-domain EM) were also used to produce a conductivity map. A total of 7 kimberlites (Fig. 1) are known in this area (Brummer et al., 1992). Solutions with an absolute value of the correlation coefficient greater than 0.75 and a standard error of regression less than 75 nT are shown on Figure 1. In the computations, the top of the pipes is assumed to be about 120 m under the magnetometer (depth of 50 m) and the radius to be 75 m.

Strong absolute correlations (>0.85) are found for most known kimberlites, except for MA 20 and MA 13-1 which have weaker absolute correlations, only slightly more than 0.75. This is because they are elongated ellipsoidal-shaped anomalies, departing from the circular model used. Plotting the solutions on the conductivity map (Fig. 2) improves target selection, since most kimberlites have a positive conductivity response. The increase in conductivity is due to the presence of near-surface conductive clays in the weathered kimberlite (Palacky, 1987). In the Kirkland Lake region, kimberlites documented by Brummer et al. (1992) are also associated with an increase of the overburden thickness, likely

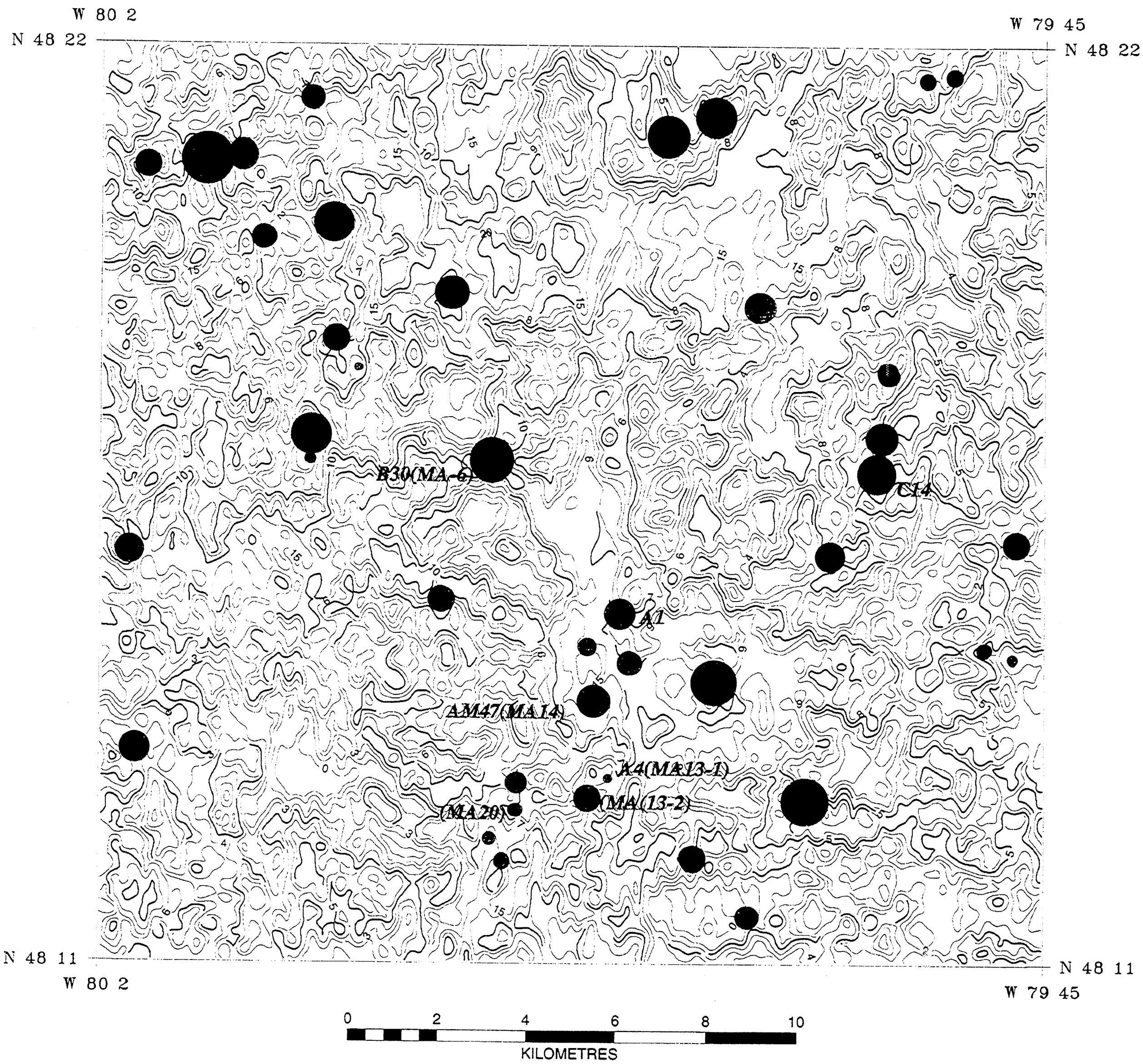


***Figure 2.*** *Apparent conductivity map of the Kirkland Lake area, northern Ontario. Location and numbering of the kimberlite pipes is from Figure 2 of Brummer and al. (1992). Shaded circle diameters are proportional to the absolute correlations identified in Figure 1. Contoured in mS/m (millisiemens/metre).*

caused by recessive erosion of the weathered top of the pipe prior to the last glaciation. This results in a higher apparent conductivity, a criterion that can be used to reduce the number of possible targets. All the known kimberlites in the Kirkland Lake region have a positive conductivity response.

## CONCLUSIONS

Most kimberlite pipes have distinct geophysical signatures, detectable by airborne surveys. In the Canadian Shield, their magnetic anomaly is roughly circular. A simple regression between the theoretical magnetic anomaly due to a vertical cylinder and the observed magnetic field within a moving window is an effective tool to identify magnetic anomalies caused by kimberlite pipes. The absolute value of the correlation coefficient and the standard error of fit of the regression are used to screen the possible targets. Plotting the solutions on a conductivity map of the studied area further improves the technique, since most kimberlites have a positive conductivity response. Interpretation is further improved by the use of geological criteria.

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* Contribution to the Canada-Ontario Subsidiary Agreement on Northern Ontario Development (1991-1995), completed under the Canada-Ontario Economic and Regional Development Agreement.

# Geophysical characteristics of Canadian kimberlites

**C.J. Mwenifumbo, J.A.M. Hunter, and P.G. Killeen**

*Mwenifumbo, C.J., Hunter, J.A.M., and Killeen, P.G., 1996: Geophysical characteristics of Canadian kimberlites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 237-240.*

## INTRODUCTION

In the past few years, a wide range of ground and airborne geophysical surveys have been conducted in the search for kimberlite pipes in Canada. Geophysical surveys have been mainly magnetic and electromagnetic (Keating, 1996), and to a lesser extent radiometric, seismic and gravity. Although some physical property data on Canadian kimberlites exist (e.g. Katsube and Scromeda, 1994) there is a lack of in situ data to aid in quantitative interpretation of geophysical data from airborne or ground surveys. These data would provide fundamental information for planning, interpreting and understanding a multitude of geophysical survey results.

The Geological Survey of Canada (GSC) is currently establishing an inventory of in situ physical property data, using borehole geophysics, for a variety of deposit types in Canada including kimberlites. The main objectives of the kimberlite studies are:

1) to compile information on the physical properties of the kimberlite pipes and their host rocks;
2) to develop a database of geophysical signatures of the different kimberlites; and
3) to determine the responses of different logging tools that may be used in delineation and exploration for kimberlites.

A compilation of kimberlite geophysical signatures will provide standards for comparing geophysical data and evaluating new borehole exploration technology aimed at finding new kimberlites in Canada. The in situ physical property data may also support the design of new generations of geophysical exploration equipment and the development of novel airborne, ground, and borehole methods.

## FORT À LA CORNE KIMBERLITE, SASKATCHEWAN

A 242 m borehole was drilled into a kimberlite near Smeaton, Saskatchewan, specifically for geological and borehole geophysical investigations by the Geological Survey of Canada. The hole intersected 100 m of Quaternary sediments, 40 m of Cretaceous shales and siltstones interbedded with kimberlite, 80 m of kimberlite with minor shale and sandstone layers, and 20 m of Mannville sandstone. Borehole geophysical data were acquired to determine the in situ geophysical characteristics of the kimberlite and its host rocks. Because the HQ hole showed signs of incompetence, it was cased with 2-inch ID PVC "plastic" pipe which restricted the use of galvanic electrical methods. However, seven different logging tools were run in the hole inside the PVC pipe. Nuclear, electromagnetic, and magnetic measurements can all "see" through the plastic pipe. The geophysical logs acquired included natural gamma ray spectrometry (Total Count gamma, K, U, and Th), magnetic susceptibility, inductive conductivity, spectral gamma gamma (SGG - density and heavy element indicator), temperature, three-component magnetometer, and sonic P-wave velocity. The last parameter was measured using a surface energy source and downhole recording with an array of twelve hydrophones.

Results indicate that, although there is a high degree of variability within the kimberlite, most of the geophysical measurements in kimberlite show distinct differences from the surrounding sediments. Density, magnetic susceptibility, and seismic P-wave velocity logs show distinctly higher values in kimberlite than in the host sediments. The gamma-ray signature, however, varies considerably and is not significantly different from that observed within Quaternary sediments or Cretaceous shales, siltstones and sandstones. The gamma-ray data alone, therefore, do not characterise the kimberlite. Figure 1 shows five of the geophysical logs: total count gamma, magnetic susceptibility, conductivity, density, and P-wave velocity. Both gamma-ray and electrical conductivity data show bimodal distributions within the kimberlite suggesting two distinct populations. However, when these two data sets are crossplotted, five distinct clusters emerge: 1) low gamma, low conductivity;

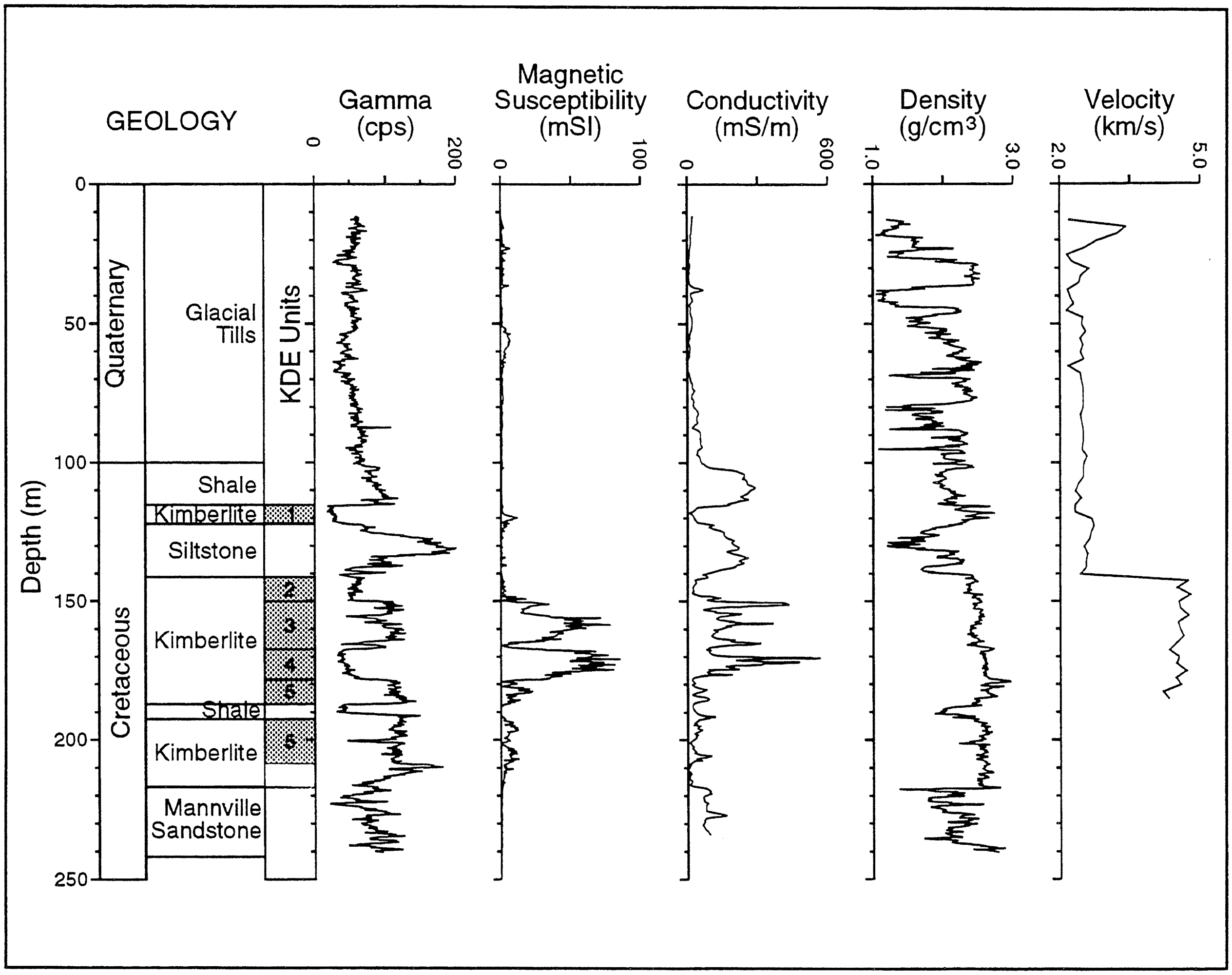


*Figure 1: Total count gamma, magnetic susceptibility, inductive conductivity, density, and P-wave velocity logs recorded in the GSC borehole at a Fort à la Corne kimberlite pipe. The numbers in the column labelled KDE Units refer to the distinct clusters identified in Figure 2.*

2) medium gamma, low conductivity; 3) high gamma, high conductivity; 4) low gamma, high conductivity; and 5) high gamma, low conductivity, which likely represent different phases of the kimberlite eruptions. These five clusters are identified in Figure 1 and in Figure 2, which shows a crossplot, contour map and a perspective view of the kernel density estimate (Mwenifumbo, 1993) of gamma-ray versus conductivity data within the kimberlite. Crossplots of other geophysical logging measurements were also made to investigate relationships between parameters. There is a strong linear correlation between magnetic susceptibility and conductivity (r=0.74).

## KIMBERLITE PIPES IN THE KIRKLAND LAKE AREA, ONTARIO

Four kimberlite pipes were investigated in the Kirkland Lake area, Ontario, during 1993. Borehole geophysical logging was carried out with the GSC R&D logging system in several company boreholes and three shallow rotosonic boreholes drilled by the GSC. The GSC boreholes were drilled for indicator mineral investigations through Quaternary glacial deposits overlying the kimberlite (McClenaghan, 1996). The geophysical variables measured included magnetic susceptibility, density, induced polarization, resistivity, self potential (SP), inductive conductivity, natural gamma ray spectrometry (Total Count gamma, K, U, Th), spectral gamma gamma (SGG - density and heavy element indicator), and temperature. Borehole 3-component magnetometer surveys were also conducted in three boreholes at two pipes. Results show that the four kimberlite pipes investigated have different geophysical characteristics. Two of the pipes are magnetic, highly resistive and have higher densities than the host rocks.

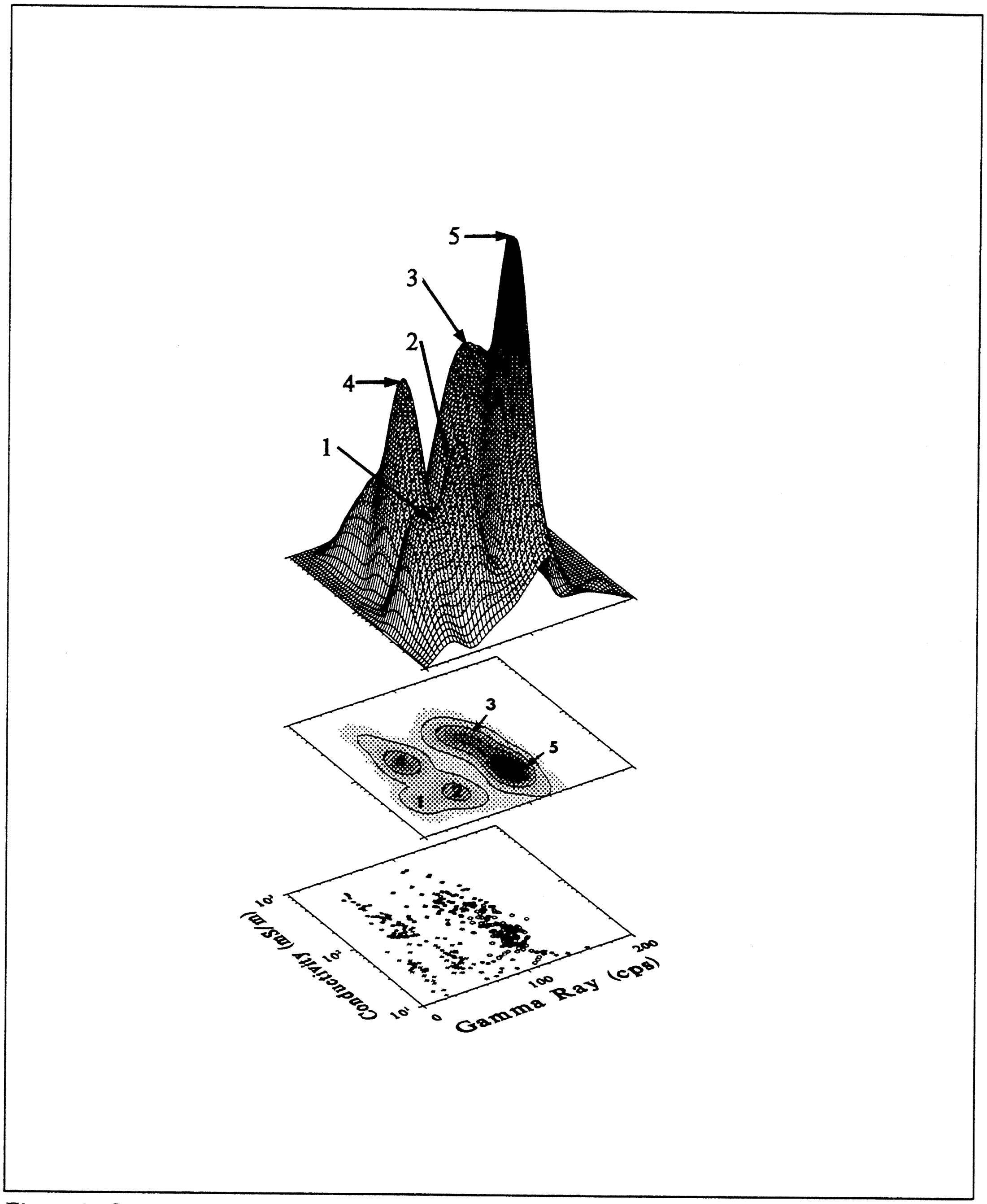


*Figure 2: Crossplot, contour map and a perspective view of the kernel density estimate of gamma ray versus conductivity data from the Fort à la Corne borehole. The perspective view and contour map of the kernal density estimate show five distinct clusters that may represent different kimberlite eruptive phases, which display: 1 = low gamma, low conductivity; 2 = medium gamma, low conductivity; 3 = high gamma, high conductivity; 4 = low gamma, high conductivity; 5=high gamma, low conductivity.*

The other two are nonmagnetic. There are also distinct differences in measured physical properties between epiclastic and intrusive kimberlite.

## CONCLUSION

The geophysical data from the kimberlites investigated indicate that the physical properties are variable within a kimberlite pipe and also between pipes. Although there is a high degree of variability of the physical properties within the kimberlite, most geophysical measurements show anomalous values which are characteristic of the kimberlites compared to the surrounding sediments. Density, magnetic susceptibility, and sonic P-wave velocity logs, for example, show distinctly higher values within kimberlites. The geophysical data may also be useful for classifying the different facies and source material of kimberlites. Five different phases of kimberlites were readily identified, based on borehole geophysical measurements at one Fort à la Corne kimberlite pipe.

## ACKNOWLEDGMENTS

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# Physical Characteristics of Canadian Kimberlites

**T.J. Katsube and B.A. Kjarsgaard**

*Katsube, T.J. and Kjarsgaard, B.A., 1996: Physical Characteristics of Canadian Kimberlites; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 241-242.*

## INTRODUCTION

Physical properties (e.g., density, porosity, resistivity, magnetic susceptibility) of 41 kimberlite samples from the Northwest Territories, Saskatchewan (Fort à la Corne and Sturgeon Lake), and Ontario have been measured (Katsube et al., 1992; Katsube and Scromeda, 1994; Scromeda et al., 1994) to obtain information on physical processes involved in an intruding kimberlite. This information is necessary to understand geophysical signatures to be expected from kimberlite pipes, and for development of geophysical methods with improved detecting capabilities below overburden, a requirement for geophysical exploration of kimberlite-hosted diamond pipes in central and northern Canada. While some kimberlite physical property data exist (e.g., da Costa, 1989), they are insufficient. This study includes, probably for the first time, physical properties related to the kimberlite facies: hypabyssal (HB), diatreme (DT), and crater (CR) facies kimberlites. This paper discusses the topics and progress being made in the GSC kimberlite physical property studies.

## METHODS OF MEASUREMENT

Laboratory measurements performed on kimberlite samples on a routine basis are bulk density ($\delta$), magnetic susceptibility (MS), porosity ($\phi$), electrical resistivity ($\rho_r$), and formation resistivity factor (F). Pore-size-distribution measurements are planned for selected samples. Methods and procedures used in these measurements are described in several GSC publications (e.g., Katsube et al., 1992; Katsube and Scromeda, 1994).

## PROGRESS

Physical property measurements have been completed on 41 kimberlite samples, with the results (more or less in the form of raw data) published in three GSC Current Research papers (Katsube et al., 1992; Katsube and Scromeda, 1994; Scromeda et al., 1994). Another paper is in preparation for a GSC publication on a kimberlite at Smeaton, Saskatchewan (Fort à la Corne field). Data analysis and interpretation have been performed for most of these samples. To date, mainly hypabyssal (HB) and crater (CR) facies kimberlites have been studied. Preparations are currently underway to study diatreme (DT) facies kimberlites from Guigge Township (Quebec), and Kirkland Lake (Ontario). Future studies will emphasize measurements on diatreme facies kimberlites.

## RESULTS

The results indicate that the physical properties are closely related to the kimberlite facies classification: hypabyssal (HB) and crater (CR) facies. They show that some kimberlites have electrical resistivity ($\rho_r$) values of 1000 to 60000 $\Omega$m, indicating probably for the first time, that there are kimberlites with $\rho_r$ above 500 $\Omega$m. The pore structure differences result in the two groups of kimberlites (HB and CR) showing distinctly different physical properties. HB-kimberlites are characterized by high bulk density ($\delta$), electrical resistivity ($\rho_r$), and formation factor (F) values, and low effective porosity ($\phi_E$) values, resulting from poor pore inter-connectivity. CR-kimberlites are characterized by low $\delta$, $\rho_r$ and F values and high $\phi_E$ values, resulting from good pore inter-connectivity. Examples of these differences for $\delta$ and $\rho_r$ are displayed in figures 1 and 2. Magnetic susceptibility (MS) is unrelated to pore structure, and thus shows no relation to the kimberlite facies. However, within each of the HB and CR kimberlite facies the relatively good relationship that exists between $\rho_r$ and MS varies with the kimberlite facies, as shown in figure 2.

These studies may provide some guidance to geophysical signatures of importance for kimberlite identification and economic significance. Although, kimberlites display both high and low $\rho_r$ and MS (Fig. 2), there may be cases where high $\rho_r$ and MS are less likely to be related to diamonds.

Results of these studies also indicate that at the hypabyssal-diatreme/crater facies boundary in the kimberlite pipe, there is a drastic change in effective

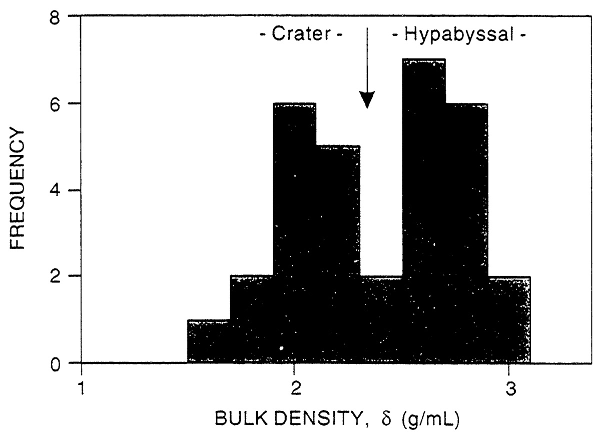


***Figure 1:*** *Frequency histograms of bulk density ($\delta$).*

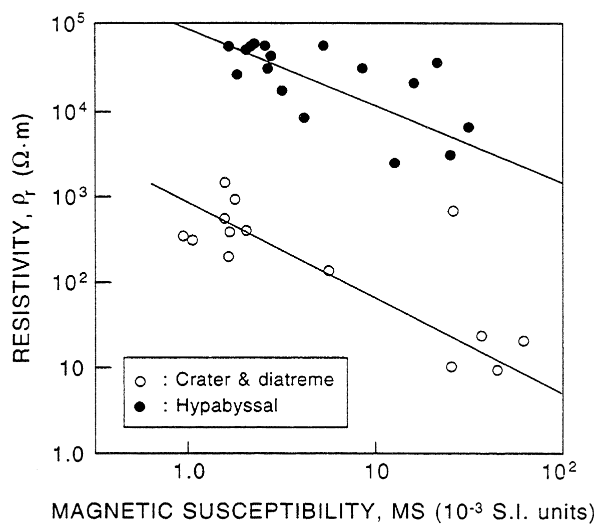


***Figure 2:*** *Bulk electrical resistivity ($\rho_r$) as a function of magnetic susceptibility (MS), showing two distinct groups: one for the hypabyssal (HB) kimberlites and the other for the diatreme/crater (DT/CR) kimberlites.*

porosity ($\phi_E$), total porosity ($\phi_T$) and pore interconnectivity ($\alpha=\phi_E/\phi_T$). These values change from 1.8-5.5 to 8.9-27 % for $\phi_E$, 5-16 to 12-28 % for $\phi_T$ and 0.21-0.36 to 0.66-1.0 for $\alpha$, as the facies changes from HB to CR-kimberlites. The marked increase in $\phi_E$ and $\alpha$ for CR-kimberlites is probably due to an explosion-like event that took place at the facies boundary.

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# Geophysical measurements for lithospheric parameters

**A.G. Jones, D.W. Eaton, D. White, M. Bostock, M. Mareschal, and J.F. Cassidy**

*Jones, A.G., Eaton, D.W., White, D., Bostock, M., Mareschal, M., and Cassidy, J.F., 1996: Geophysical measurements for lithospheric parameters; in Searching for Diamonds in Canada, A.N. LeCheminant, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 243-250.*

## INTRODUCTION

With the exception of mantle xenoliths and limited exposures of mantle rocks in collisional orogenic belts, the continental lithospheric mantle is generally inaccessible to direct observation. Knowledge of parameters such as the age, thickness, and internal geometry of the upper mantle, all of which could be used to optimize exploration strategies for kimberlites or lamproites likely to have originated in the diamond stability field, may only be available indirectly through geophysical techniques. A new generation of teleseismic and deep-probing electromagnetic methods has emerged recently, that provides passive and cost-effective means to infer this information and to map the lithospheric mantle in its entirely, from the base of the crust to the asthenosphere.

This paper provides an overview of these techniques, and describes ongoing experiments that comprise part of LITHOPROBE, Canada's national collaborative geoscience program involving the GSC and NSERC (Fig. 1). Other non-LITHOPROBE experiments have been proposed as well, and if funded will contribute further to our knowledge of the Canadian lithosphere. An outline of the relevant techniques is given here; more detailed discussions of the various methods are in the references cited.

## TELESEISMIC STUDIES

Conventional seismic reflection and refraction techniques are generally inadequate to resolve upper-mantle structure, as insufficient energy penetrates to subcrustal depths. Teleseismic studies, which involve the detection and analysis of waves generated by distant earthquakes, overcome this limitation. A number of complementary techniques are available to examine different portions of the recorded teleseismic wavefield, in order to extract information on various aspects of the upper mantle. In particular, four techniques, are well suited for the characterization of upper mantle structure, including: 1) S-wave splitting; 2) receiver-function analysis, 3) surface-wave inversion; and, 4) body wave traveltime tomography. These provide a range of complementary information on upper mantle structure and are summarized below.

### *Teleseismic S-wave splitting studies for upper-mantle deformation*

In general, a shear wave that passes through an anisotropic region splits into fast and slow modes of propagation. The two diagnostic parameters of this phenomenon are the delay time for the slow arrival, which is roughly proportional to both the thickness of the anisotropic layer and the magnitude of the anisotropy, and the polarization directions, which record the orientation of the elastic symmetry system (Crampin, 1981). Observations of these parameters for split "SKS" waves (shear-wave arrivals that have passed through the Earth's outer core as compressional waves) provide strong evidence that the upper mantle beneath the point of observation is anisotropic. Strain-induced alignment of constituent minerals (olivine and orthopyroxene) is now widely accepted as the principal cause of upper-mantle anisotropy (Kern, 1993; Babuska et al., 1993; Mainprice and Silver, 1993). This lattice-preferred orientation is usually attributed to flow regimes in the mantle, either from present-day plate motions or fossil strain due to the last major tectonic event (e.g. Vinnik et al., 1984; Silver and Chan, 1991; Silver and Kaneshima, 1993).

Both delay time and polarization for SKS arrivals are readily measured using a broadband three component seismograph. Application of this technique requires a reliable source distribution at epicentral distances of 85° to 140° from the location of interest. Based on this criterion, seismically active regions from the western Pacific island arcs and to a lesser extent South America provide suitable sources for the entire Canadian landmass (Fig. 2). Figure 3 illustrates preliminary results of splitting parameters measured on the Canadian National Seismograph Network [CNSN] (Bostock and Cassidy, 1995a), providing a large-scale framework for more detailed regional investigations.

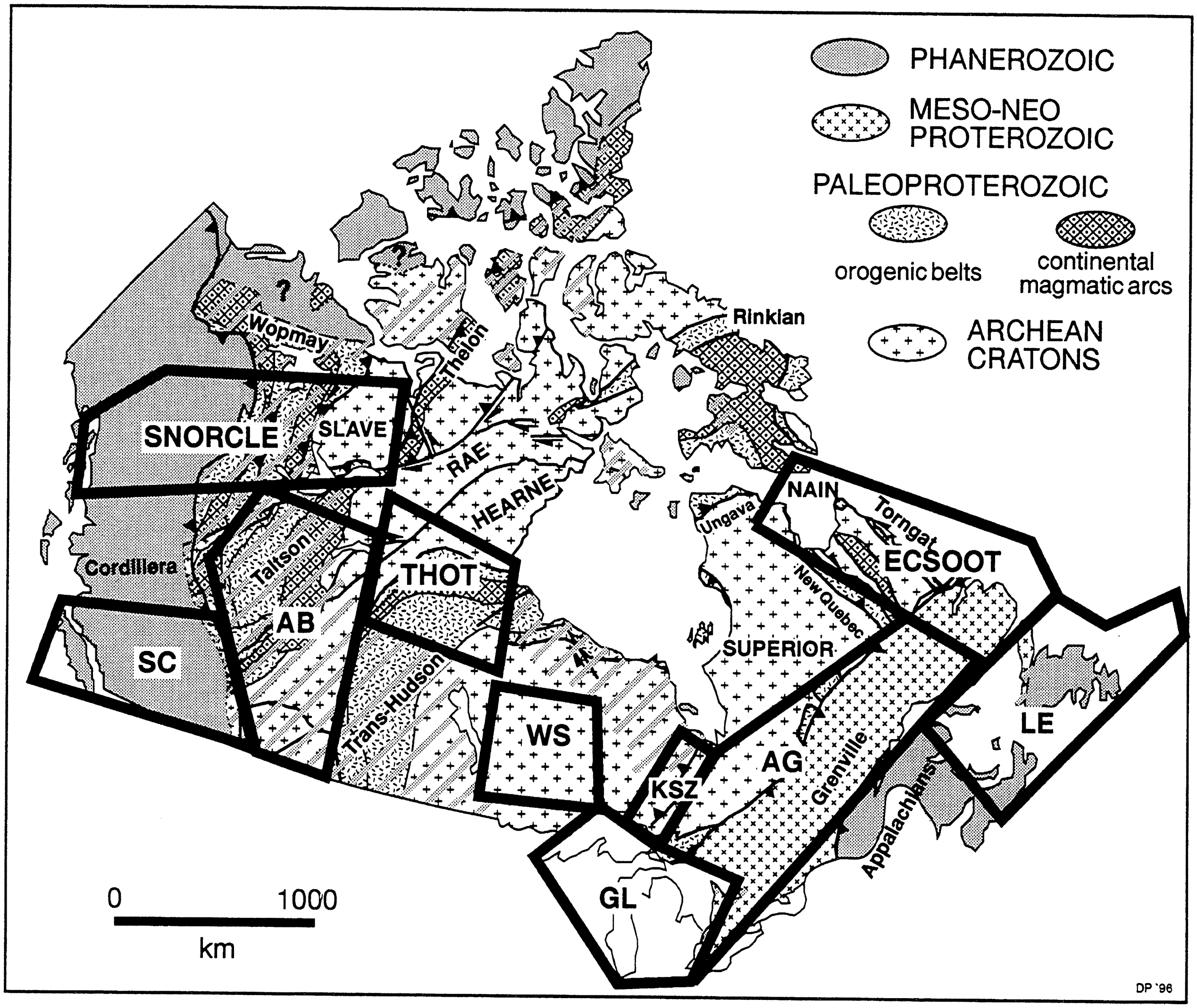



*Figure 1. Location of LITHOPROBE transects on a map showing the ages of major tectonic elements in Canada. The transects are: SC - Southern Cordillera, AB -Alberta Basement, SNORCLE - Slave-Northern Cordillera Lithospheric Evolution, THOT - Trans Hudson orogen, WS - Western Superior, KSZ - Kapuskasing Structural Zone, GL - Great Lakes International Program on Crustal Evolution, AG - Abitibi-Grenville, LE -Lithoprobe East, and ECSOOT - Eastern Canadian Shield Onshore-Offshore. Diagonal stripes represent areas where the Precambrian basement is covered by Phanerozoic platformal cover.*

### *Receiver function analyses of upper mantle discontinuities*

Teleseismic receiver-function analysis is used to derive a model of the S-wave velocity structure beneath a broadband seismograph. This is achieved by deconvolving the vertical-component signal (dominated by P-wave energy) from the radial and transverse components, to eliminate unwanted source and path effects and isolate local P-S conversions. The method has traditionally been applied to crustal problems (e.g. Mangino et al., 1993; Cassidy, 1995). However, it has recently been shown that, by application of techniques similar to beam forming used in seismic array studies, it is possible to determine velocity-depth products for the 410 and 670 km discontinuities (Vinnik, 1977; Bostock and Cassidy, 1995b). As these discontinuities are due to phase changes, their depths have been shown to be sensitive indicators of the ambient temperature regime and/or volatiles within the upper mantle (Vidale and Benz, 1992; Wood, 1995) with important implications for the question of whole-mantle versus layered-mantle convection.

The Cordillera is an ideal setting to address this question as the smooth velocity structure is known accurately (Bostock and VanDecar, 1995). As one example of the type of study proposed, we will use the velocity model of Bostock and VanDecar in conjunction

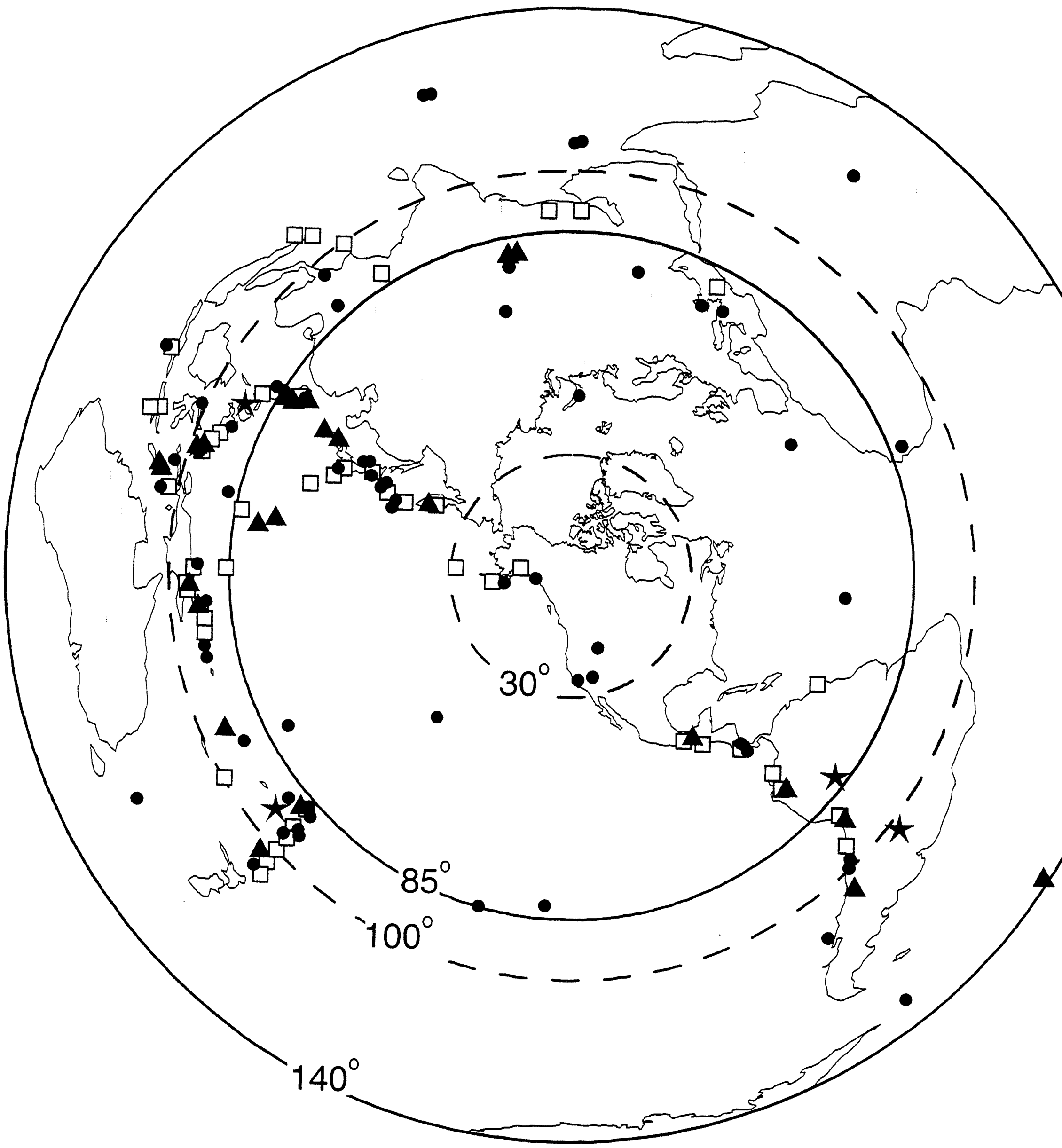
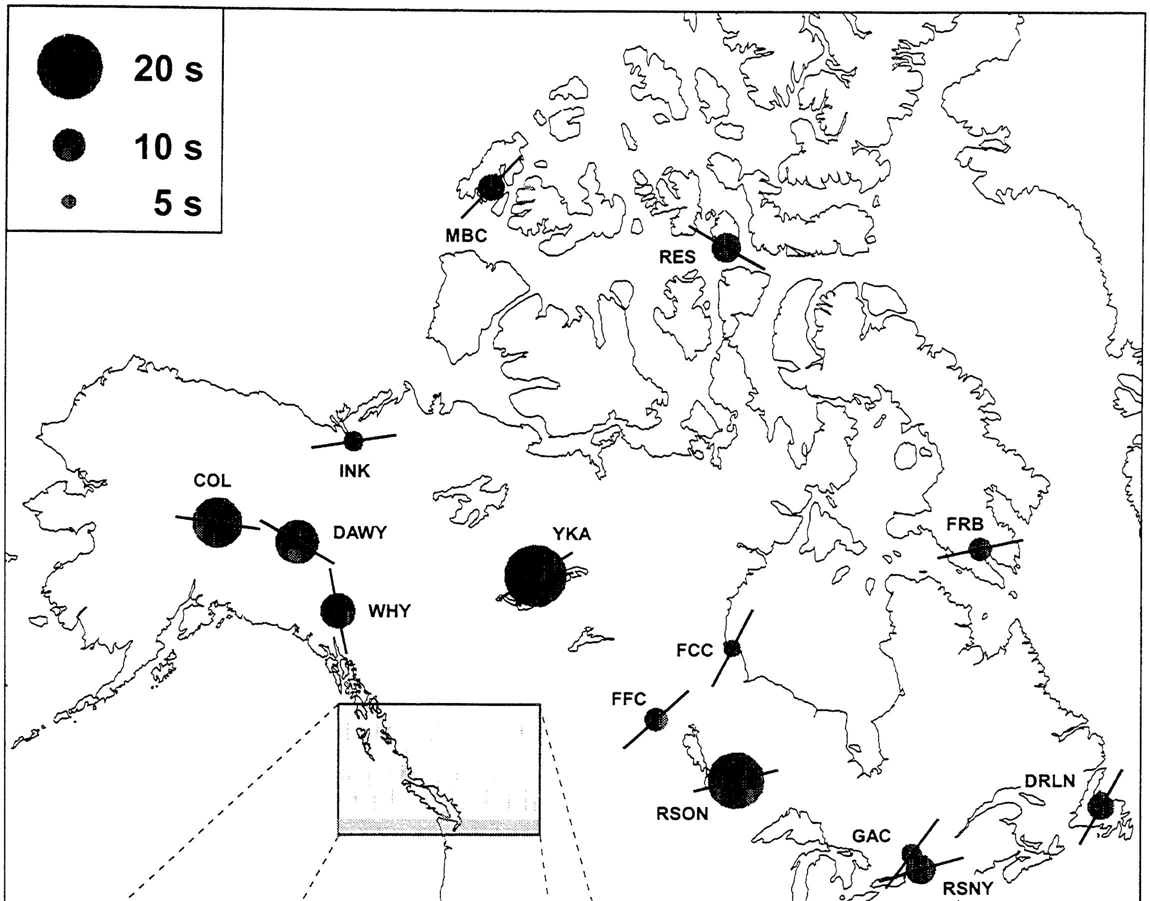


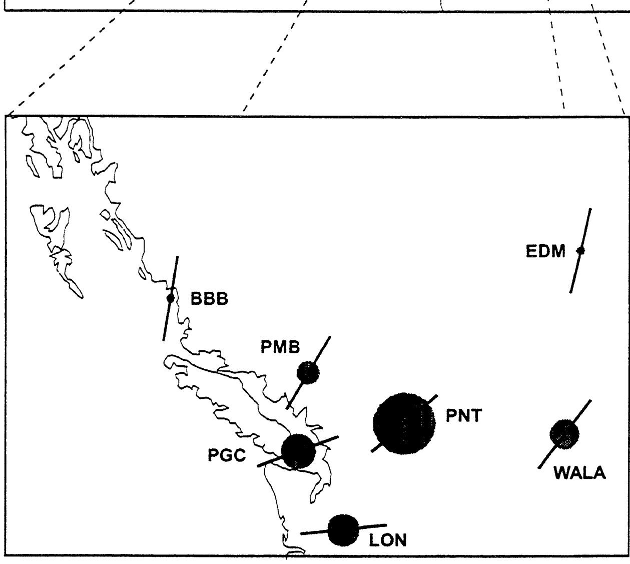

*Figure 2. Distribution of 1993 earthquakes with magnitude > 5.7. Symbols indicate focal depth as follows: circles < 35 km; squares, 35 to 70 km; triangles 75 to 300 km; stars > 300 km. Concentric circles show range (in degrees) for a study area centred on the Slave craton. Ideal range for shear-wave splitting studies is 85 to 140 ° (solid lines); for receiver-function analysis, 30 to 100 ° (dashed lines).*

with receiver function analyses of data from CNSN and temporary broadband stations in southern British Columbia, to place constraints on depths to the 410 km and 670 km discontinuities. Structure on the 670 km discontinuity will aid in discrimination between slab penetration into the lower mantle and accumulation of lithospheric material in the transition zone (Silver et al., 1988). Receiver-function analyses require earthquake sources at distances between 30 and 100°. As is evident in Figure 2, most of the Canadian landmass is well situated with respect to seismicity in the western Pacific and South American subduction zones.

### *Surface wave studies for gross upper mantle structure*

Surface waves propagate along great circle paths parallel to the Earth's surface. The velocity of a surface wave and its sensitivity to structure are dependent upon the frequencies of investigation. At lower frequencies the phase and group velocities increase, since the wave is sensitive to velocities at greater depths in the upper mantle. By analyzing the frequency dependence of velocity (i.e. dispersion) it is possible to infer the S-wave velocity structure of the upper mantle. In addition, whole-




| Abbreviations and locations of seismograph stations | |
|---|---|
| BBB | Bella Bella, British Columbia |
| COL | Fairbanks, Alaska, U.S.A. |
| DAWY | Dawson, Yukon Territory |
| DRLN | Deer Lake, Newfoundlnad |
| EDM | Edmonton, Alberta |
| FCC | Churchill, Manitoba |
| FFC | Flin Flon, Manitoba |
| FRB | Frobisher Bay, Northest Territories |
| GAC | Glen Almond, Quebec |
| INK | Inuvik, Northwest Territories |
| LON | Longmire, Washington State, U.S.A. |
| MBC | Mould Bay, Northwest Territories |
| PGC | Sidney, (Pacific Geoscience Centre) British Columbia |
| PMB | Pemberton, British Columbia |
| PNT | Penticton, British Columbia |
| RES | Resolute Bay, Northwest Territories |
| RSNY | Adirondack, New York State, U.S.A. |
| RSON | Red Lake, Ontario |
| YKA | Yellowknife, Northwest Territories |
| WALA | Waterton Lake, Alberta |
| WHY | Whitehorse, Yukon Territory |

***Figure 3.*** *Shear-wave splitting results obtained by Bostock and Cassidy (1995a, unpublished data). Diameter of circle symbol is proportional to the delay time between fast and slow arrivals; line shows the polarization of the fast arrival. See text for explanation of these parameters.*

waveform inversions where higher-order overtones are included are now feasible and dramatically improve resolution of the upper mantle (Nolet, 1990). The application of these techniques requires at least two broadband seismograph stations in line with earthquake sources and oriented perpendicular to the strike of features of interest. The derived information about velocity structure defines average properties of the upper mantle between two broadband seismic stations. This complements other teleseismic studies, which have excellent lateral resolution but provide less information on the depth extent of the heterogeneity. Since potential earthquake sources, along the western Pacific island arcs and to a lesser extent central and South America, span a large range of azimuths, surface wave studies are feasible in most Canadian regions of interest for diamond exploration (Fig. 2).

### *Teleseismic travetime tomography*

Teleseismic tomography provides a means of investigating smooth, three-dimensional variations in seismic P- and S-velocities in the upper mantle. It requires that the traveltimes of waves generated by distant earthquakes be measured at an array of seismic stations on the Earth's surface. The raypaths of the waves intersect in the upper mantle underlying the array, and when combined for many events, allow the construction of detailed subsurface velocity maps and cross sections. Recent studies (VanDecar et al, 1995, see also Bostock and VanDecar, 1995) have demonstrated the feasibility of traveltime tomography in portable 3-component array experiments to image upper mantle structure and identify contributions of both thermal and compositional origin. This information is essential to the characterization of the mantle reservoir whence diamondiferous kimberlites originate. The efficacy of the technique is dictated by the availability of a large number of receiver stations (>10) and global seismicity coverage. Vertical, short period instrumentation is sufficient for P-wave studies, however incorporation of S-waves will generally require broad-band seismometers. A comprehensive coverage of events in azimuth and at distances between 30 and 100 degrees is required for effective resolution, and is afforded by global seismicity for most locations in Canada (Fig. 2).

## ELECTROMAGNETIC STUDIES

Passive electromagnetic sounding for upper mantle studies can be carried out by simultaneous recording of: 1) time variations of the three components of the magnetic field (Hx, Hy, Hz) at a number of locations; or, 2) two electric (Ex and Ey) and three magnetic (Hx, Hy, Hz) components at a single location. The former technique is termed "Geomagnetic Depth Sounding" (GDS), and from the horizontal-field gradient the "Horizontal Spatial Gradient" (HSG) method, whereas the latter is called the "MagnetoTelluric" (MT) technique. The measurements include contributions from two parts: the external "source" field (i.e., ionospheric and magnetospheric electromagnetic waves caused mainly by the interaction between the Earth's magnetosphere and the Sun's ejected plasma), and the "induced" field (i.e., the secondary fields generated by currents induced within electrically conductive zones in the Earth).

The depth of penetration of the source field depends on the source frequency as well as on the electrical resistivity of the Earth material from the surface to that depth. For example, as the upper crust of the Canadian Shield is very resistive, the conventional period range of high-quality magnetotelluric data acquisition (0.02 - 500 s) resolves electrical structures from a few kilometres to about 50 km depth. To measure mantle responses over a geographical area covering a wide range of upper-crustal electrical responses, it is necessary to extend the conventional period range to 30 000 s, and, at selected sites, to several days (e.g. Shultz et al., 1993). The GSC has recently designed instruments called LIMS (Long period Intelligent Magnetotelluric System) that operate over this frequency range.

Deep-probing HSG and MT studies in Scandinavia in the early-1980s (Jones, 1984; Jones et al., 1983) illustrated that lateral variations in depth to the top of the "electrical asthenosphere" could be mapped with the appropriate EM methods. These depths were shown to be consistent with compressional lithospheric "lid" thicknesses determined from surface wave studies (Calcagnile, 1991; Calcagnile and Panza, 1987).

While resolution of subtle electrical features in the mantle was not possible only a few years ago (Jones, 1992), recent developments in magnetotelluric data collection and processing now allow the recognition of such features. The first of these was in the upper mantle beneath the eastern Canadian Shield, where a clear azimuthal electrical anisotropy was discovered from the interpretation of 140 MT soundings recorded as part of the Abitibi-Grenville and Kapuskasing transects (Mareschal et al., 1995). The ratios of horizontal resistivities can be as high as 1:15 and the anisotropic zone is found between approximately 50 and 150 km depth. The azimuth of enhanced electrical conductivities varies over horizontal distances in the order of a few hundred kilometres, but unlike seismic anisotropy, it does not show any clear relationship to geological boundaries. Although the physical mechanisms of electrical anisotropy are not clear, it unlikely that lattice-preferred orientation of olivine is the cause. Indeed, the conductivities of the major rock-forming minerals in the upper mantle are too small to explain the overall observed conductivity. Electrical conduction in the upper mantle is primarily through minor constituents such as graphite or sulphides or saline fluids, except where the mantle is hot enough to contain partial melt (Jones, 1992); thus electrical anisotropy must arise from the geometry of the interconnection between the conductive phases. In the Canadian Shield, the conductive azimuth correlates reasonably well with the trends of two crustal-scale shear zones which extend across the southern Superior Province. These shear zones are thought to have provided conduits for the migration of Au and $CO_2$-rich fluids as well as alkaline magmas from the upper mantle during the late Archean. Therefore, Mareschal et al. (1995) suggest that the upper mantle anisotropy beneath eastern Superior Province is of Archean age and is due to graphite-filled veins or microfractures within the mantle beneath crustal shear zones.

A proposed test of this hypothesis will be to examine the mantle electrical response on the western section of the Superior Province (through which the crustal shear zones continue) using 150 MT soundings which will be collected in 1998 as part of the Western Superior LITHOPROBE transect (Fig. 1). The transition from the Slave craton to progressively younger accreted terranes farther west (SNORCLE transect; Fig. 1) will provide another crucial test of the sources and significance of the upper mantle electrical response. The interpretation of these data, however, may be constrained by the highly conductive crust of the Cordillera, which obscured the resolution of mantle features in the MT survey across the southern Cordillera LITHOPROBE transect (Jones et al., 1992).

## PLANNED ACTIVITIES PRIOR TO THE YEAR 2000

### *Abitibi-Grenville Transect*

In the summer of 1994, a 3 month test teleseismic experiment using a maximum of 10 broadband 3-component seismometers (loaned by LITHOSCOPE, the French teleseismic program) was undertaken in the Abitibi belt. The data are of excellent quality, and shear-wave splitting analyses show strong splitting with the fast axis approximately east-west, consistent with the EM results (Senechal et al., 1995). In 1996, a full teleseismic experiment (30 stations) will take place in Abitibi, roughly along a N-S line from the Opatica to the Grenville. In addition to determining shear-wave splitting parameters, another goal of this experiment will be to resolve structural variations within the upper mantle along the main N-S Abitibi-Grenville transect for which we have reflection and refraction data. The experiment will include 6 to 10 months of continuous recording, using earthquakes from a southerly azimuth, largely from South America and the Caribbean. P-wave residuals (Buchbinder and Poupinet, 1977) will be supplemented with travel time delays of the S-waves (PKS phase), allowing evaluation of average Poisson's ratio. This experiment will run concurrently with a teleseismic array just south of our study area which is designed to cross the Appalachians into the Grenville Province, yielding a total transect length of 1000 km.

### *Trans-Hudson Orogen Transect*

The discovery of diamondiferous kimberlites in the late 1980s in north-central Saskatchewan, coupled with new constraints on the internal structure of the Trans Hudson orogen (St-Onge and Lucas, 1996), have motivated a series of teleseismic experiments to elucidate the lithospheric structure in the economically targeted part of the orogen (Ellis and Hajnal, 1993). A teleseismic feasibility study has been conducted using an 8 station array with individual seismographs operating for 4 to 6 months (Ellis et al., in press). Receiver function analysis for crustal structure have provided new evidence for large variations (~7 km) in crustal thickness in the southwestern area of Saskatchewan. SKS analysis, which shows rapid variations in, anisotropy within the orogen, also adds to the analysis. A more extensive program with seventeen 3-component stations is now underway in central Saskatchewan-Manitoba that will be used to map the 3-D lithospheric structure of the region, in conjunction with LITHOPROBE seismic reflection and refraction studies.

### *SNORCLE Transect*

A broadband seismic survey across the SNORCLE transect began in the summer of 1994, and will collect teleseismic data for a two-year period using permanent stations at Whitehorse, Yellowknife, and Churchill, and temporary stations at Watson Lake, Fort Simpson, and Snowdrift. S-wave splitting determinations over the Canadian Cordillera, which included broadband seismograph stations in eastern Alaska and Yellowknife (Silver and Chan, 1991), have already established the existence of shear-wave splitting anomalies and seismic anisotropy, probably reflecting deformation associated with the Cordilleran orogen. In contrast, shear-wave splitting in the cratonic mantle under the adjacent Canadian Shield, particularly in the region of the Western Superior LITHOPROBE transect, is dominated by fossil anisotropy possibly related to Archean orogenesis (Silver and Kaneshima, 1993). A broadband seismic array between Whitehorse and Churchill and a magnetotelluric survey will be integral parts of the LITHOPROBE SNORCLE transect, providing a unique opportunity to determine the variations in, and correlations between, seismic and electrical anisotropy in the upper mantle from the Archean craton to the Phanerozoic accreted terranes of the Canadian Cordillera.

Magnetotelluric experiments for deep structure are planned to commence in late 1995 with the installation of a small number of sites close to Yellowknife for a period of 4 to 6 months. One of these sites will take advantage of disused power lines to provide long electrode lines (many kilometres). This will reduce local distortion effects, and may permit mantle conductivity resolution as precise as that recently defined below the Superior craton by Schultz et al. (1993). MT data acquisition along the SNORCLE transect, under the auspices of LITHOPROBE, is scheduled for 1997 and 1999, with sites from the Slave craton to the west in 1997, and in the Cordilleran segment in 1999. Because of its proximity to the Lac de Gras kimberlite field, the Slave segment of the SNORCLE transect has been identified as a high priority corridor for additional lithospheric studies, both seismic and electromagnetic.

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# Seismic reflection survey of a kimberlite intrusion in the Fort à la Corne district, Saskatchewan

**D.J. Gendzwill and S.D. Matieshin**

*Gendzwill, D.J. and Matieshin, S.D., 1996: Seismic reflection survey of a kimberlite intrusion in the Fort à la Corne district, Saskatchewan; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 251-253.*

## INTRODUCTION

A seismic reflection survey in the Fort à la Corne district of central Saskatchewan has proven to be effective for evaluation of the geometry and structure of a kimberlite body. The kimberlite, located on property held by Uranerz Exploration and Mining Limited, in joint partnership with Cameco Corporation, Monopros Limited and Kensington Resources Limited, near Smeaton, Saskatchewan, was discovered by follow-up of an aeromagnetic anomaly. The site is identified as anomaly 169, and anomaly 269 (a smaller eastern extension of anomaly 169)(Fig. 1).

Several drill holes tested anomaly 169, which is about 1 km in diameter, and kimberlite was found in the drill core below about 100 m of glacial till. The holes drilled by the mining company penetrated as much as 100 m of the kimberlite and entered siltstone and sandstone of the Lower Cretaceous Manville Group (Cantuar

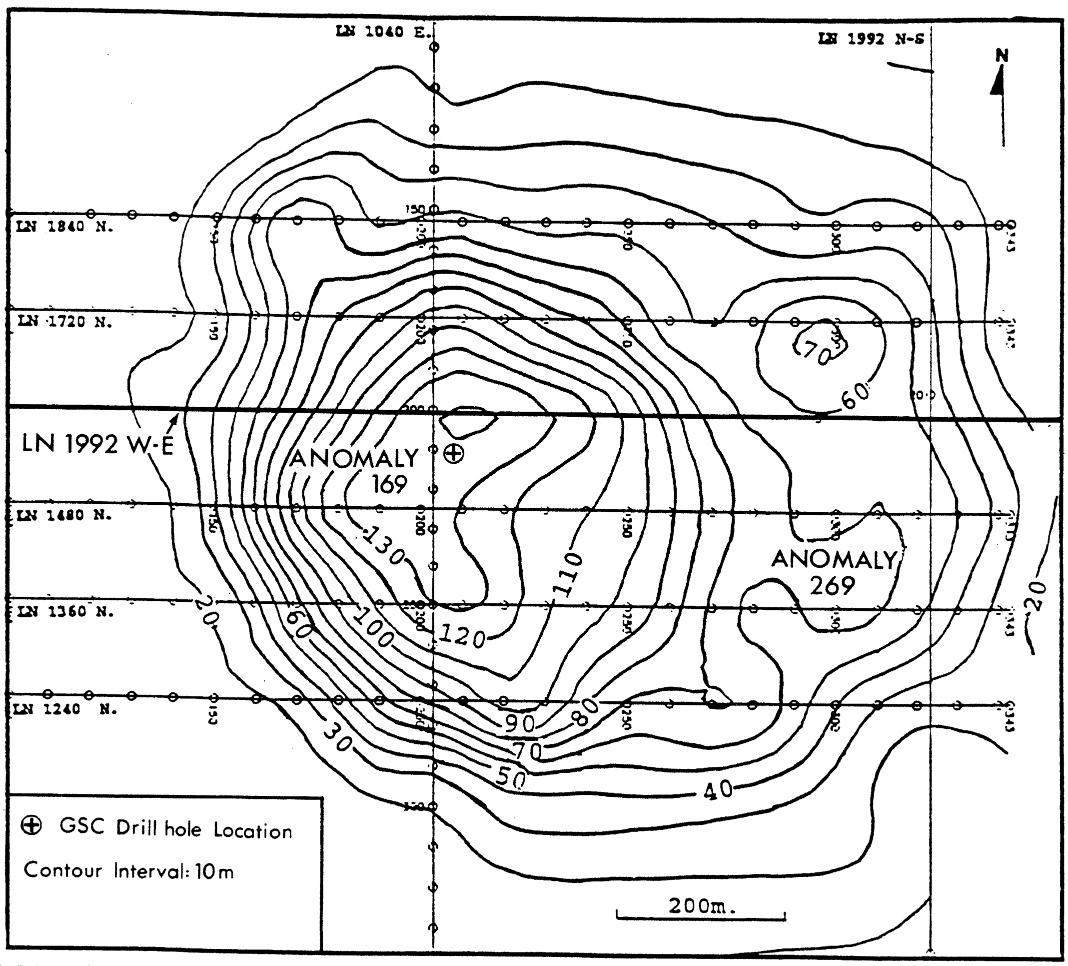


***Figure 1.** Map of the Fort à la Corne seismic survey area (Anomaly 169/269) showing the location of: the GSC drill hole; 1992 and 1993 seismic lines, including the trace of LN 1992 W-E shown in Figure 2; and kimberlite isopachs, based on seismic data.*

Formation) beneath it. One hole, drilled and logged by the Geological Survey of Canada, intersected approximately 20 m of siltstone and sandstone below the kimberlite, and then stopped in about 2 m of kimberlite at a depth of 242 m. Physical property data for the GSC Smeaton hole are provided in Mwenifumbo et al. (1996) and Richardson et al. (1996).

## METHODOLOGY

Vertical seismic reflection profiles totalling 13 km in length were collected in 1992 and 1993 over anomalies 169 and 269 with the objective of testing seismic reflection surveys as a method to determine the geometry and structure of kimberlite bodies. Conditions for this survey were excellent because the area was an open field with no obstructions, and the flat sedimentary rocks were expected to produce good seismic reflections. The 1992 survey investigated the kimberlite and surrounding area with two lines crossing and extending 1.5 km beyond the edge of the kimberlite, with geophones placed 20 m apart. The 1993 survey obtained more detailed information inside the kimberlite on six lines with geophones placed 5 m apart (Fig. 1). Dynamite was used for both surveys. Data processing used computer programs available at the University of Saskatchewan. Normal procedures of gain control, filtering, elevation, and weathering corrections, normal move-out, and stacking (12 fold) were applied; particular care was taken to recover the maximum possible signal frequency in the data. Signals with dominant frequency up to 200 Hz are present in the first 200 milliseconds of the final records.

For this report, only the principal 1992 east-west seismic line is shown (Fig. 2). The kimberlite appears as a domal structure in the middle of the line. The west side of the line contains a fold structure which may be controlled by several small faults. Undisturbed reflections from the tills, Upper Cretaceous shales, and the top of the Devonian limestones are shown on the east side of the line. Reflections from the deeper Paleozoic rocks and Precambrian surface are weak on this display. In the 1993 data, the reflection from the base of the kimberlite is clear near the periphery of the body but it is difficult to interpret near the centre, due to structural complexity. Under the kimberlite, reflections are distorted due to the high velocity of seismic waves in the kimberlite compared to the lower velocity in the adjacent sedimentary rocks.

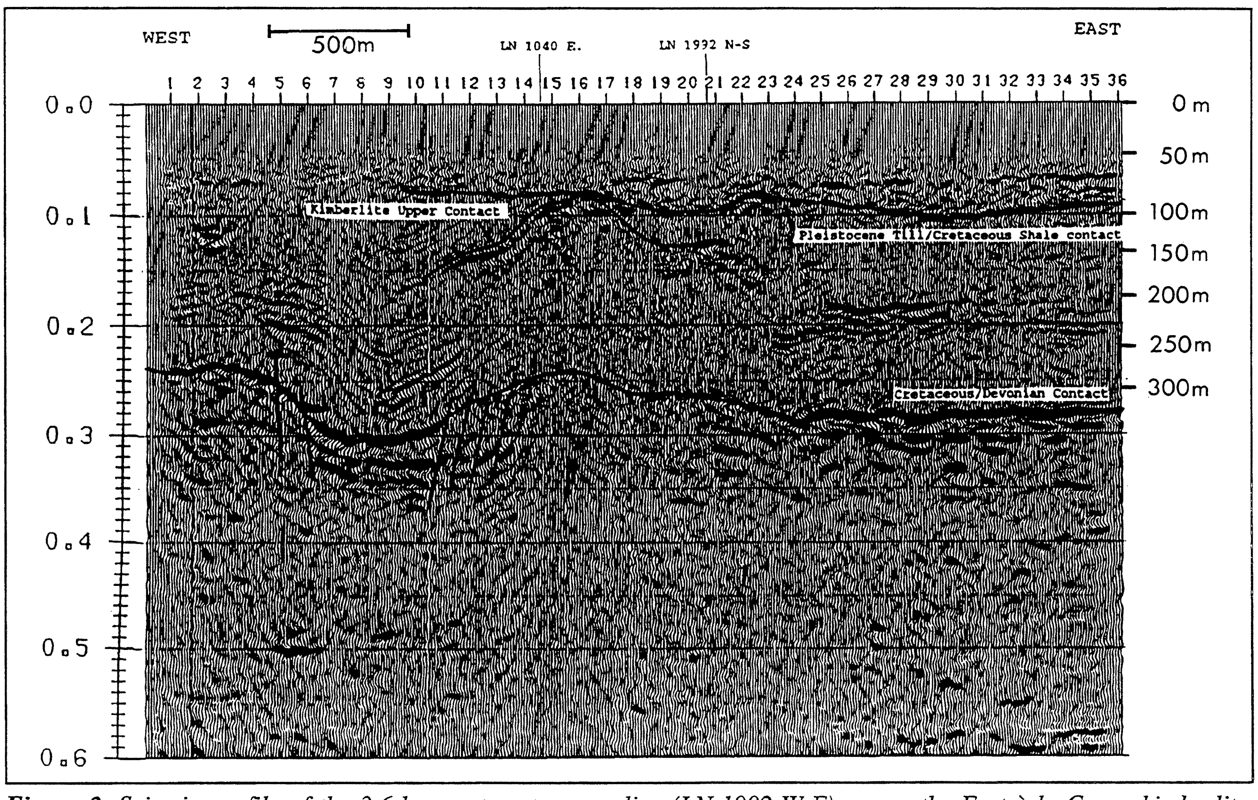


***Figure 2.*** *Seismic profile of the 3.6 km west-east survey line (LN 1992 W-E) across the Fort à la Corne kimberlite. Vertical scales are reflection time in seconds (left) and approximate depth in metres (right) assuming a velocity of 2300 m/s.*

## RESULTS

Interpretation of the seismic data (Fig. 2) indicates that the kimberlite is covered by 40 m of shale near its edges, diminishing to a thin layer near the centre of the body. Figure 1 is the isopach map of the kimberlite, based on seismic interpretation. In the centre of anomaly 169, the body is about 130 m thick, including shale, siltstone and sandstone interbeds. Anomaly 269 is represented by an extension and slight thickening of the kimberlite on the east side. The thickness and lens-like shape of the kimberlite body, the covering of stratified shale, and the existence of several laterally continuous reflectors within the body suggest that it is a crater facies, emplaced by several volcanic explosions with continuing sedimentation between explosions (see Kjarsgaard, 1996). The unique environment resulted in preservation of most of the crater facies breccia. The crater facies is rare because it tends to be quickly removed by erosion after subaerial emplacement.

There is no clear seismic evidence for a feeder dyke under anomaly 169. However, a small dyke with steep sides is a very difficult seismic target and it could well exist under the central thick portion of the 169 body where the seismic data quality is poor. Faulting is inferred from the seismic data in the adjacent area to the west of the kimberlite. A fault also could exist under the kimberlite, but there is no direct evidence for it. Seismic reflections are smooth and continuous under anomaly 269, showing that there is no feeder dyke under it. Anomaly 269 is interpreted as a lateral extension of anomaly 169, emplaced at the same time.

The seismic data have extended the geological interpretation from drilling data, showing the shape and extent of the body, its stratigraphic relationship, a nearby structure in the sediments, and evidence for multi-stage vulcanism. These interpretations are useful but they are restricted by the geometry of the linear profile seismic acquisition process. A more complete and accurate (and more expensive) interpretation with correct subsurface geometry would be possible with a true three-dimensional seismic survey over the kimberlite.

## ACKNOWLEDGEMENT

Work on the Smeaton kimberlite in Saskatchewan was done with the co-operation of Uranerz Exploration and Mining Limited on behalf of the Fort à la Corne Joint Venture, consisting of Uranerz (operator), Cameco Corporation, Monopros Limited, and Kensington Resources Limited.

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* Contribution to the Canada-Saskatchewan Partnership Agreement on Mineral Development (1990-1995), a subsidiary agreement under the Canada-Saskatchewan Economic and Regional Development Agreement.

# Application of thermal imagery from LANDSAT data to locate kimberlites, Lac de Gras area, District of Mackenzie, N.W.T.

**A.N. Rencz, C. Bowie, and B.C. Ward**

*Rencz, A.N., Bowie, C., and Ward, B.C., 1996: Application of thermal imagery from LANDSAT data to locate kimberlites, Lac de Gras area, District of Mackenzie, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 255-257.*

## INTRODUCTION

In mineral exploration, indirect methods are often used to locate potential zones of mineralization. In these methods, the presence of mineralization is inferred from observations on specific phenomena in the environment. We describe an indirect method, provide the background theory, and present an example of successful application to locate kimberlite pipes in the Lac de Gras area.

The method is based on the observation that kimberlite pipes may be associated with deep, cold lakes, and consequently the identification of an unusually cold lake may signal the presence of kimberlite. Information on the location of lakes and water temperatures was derived from LANDSAT TM data.

## GEOLOGICAL SETTING

Geoscientific data on bedrock and surficial geology were made available through the Slave NATMAP Project (Bowie et al, 1996).

Kimberlite is commonly more susceptible to weathering than its surrounding country rock. During glaciation, kimberlites in the Lac de Gras area were preferentially eroded, resulting in deep depressions and, wherever these depressions were not subsequently filled with sediment, deep lakes formed (DiLabio et al., 1992). The deeper lakes will contain a larger volume of water relative to other lakes of similar surface area. These deeper lakes should have measurable thermal differences that could be used to classify them. A characteristic of the deeper lakes would be cooler water temperatures during the summer and early fall, as a result of slower warming of the larger volume of water. The deeper lakes would also tend to be the last to freeze in winter and the last to thaw in spring.

Detection of these differences using thermal data acquired by the LANDSAT Thematic Mapper is the basis of this exploration method.

## LANDSAT TM DATA

The LANDSAT TM sensor acquires data in seven different, contiguous wavelength intervals along the electromagnetic (EM) spectrum (Fig. 1). Six of the seven bands measure reflected EM energy from pixels of 30 X 30 m (Siegal and Gillespie, 1980). The seventh band measures emitted EM energy in the 10.5 to 12.5 um region with a pixel size of 120 X 120 m. This band has been referred to as the "thermal band". LANDSAT TM data are acquired at 10:30 am (local time) so the temperature difference across the image may not be optimal, however differences between lake temperatures should not show significant diurnal fluctuation.

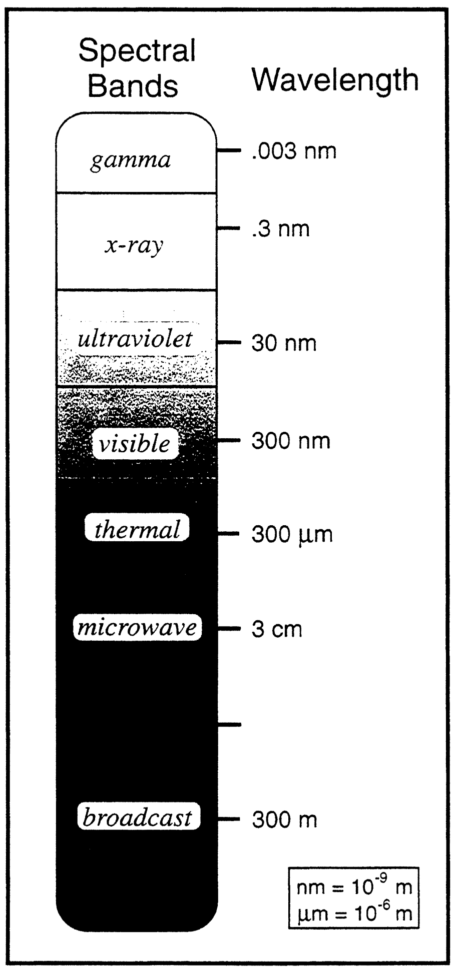


***Figure 1.** The electromagnetic spectrum.*

The classification of LANDSAT data into a map of relative lake temperatures is a two step process. First, lakes are identified on LANDSAT imagery using a threshold on one of the bands. Typically TM band 7 is used, because it is the least affected by atmospheric haze and it has the least penetration into the water. The thresholding procedure produces a binary image which identifies each pixel as either "water" or "not water". This image serves as a mask for the thermal band data. The values in the thermal band, if classified as a water pixel, are colour coded or classified so that each digital number renders a unique relative temperature code. In this code the higher numbers represent higher temperatures. It should be noted that the temperatures are only relative, as there is no ground calibration. Caution should also be exercised because differences in temperature can be due to factors other than lake depth, such as altitude and latitude, geological setting and lake size.

## EXAMPLE

The current example is from the Lac de Gras area in the Northwest Territories. Image processing was conducted using *EASI/PACE* software running on an IBM RISC 6000 (Anonymous, 1994).

The image in Figure 2 was created following the procedure outlined above, using a LANDSAT TM image acquired on September 12, 1984. In practice the optimal image to view temperature differences would be one that showed different colours for each temperature level at a scale of about 1:50 000. This was not possible for this article, so there is a considerable loss of information in the figure provided. The digital values, corresponding to water in the image (ranging from 107 through 117) were grouped into three different grey scale levels on Figure 2. The image illustrates some temperature differences. Obviously the larger lakes, notably Lac de Gras, are the coldest and the smaller lakes are typically warmer.

The search for kimberlite pipes was restricted to "round" lakes with a diameter of about 400 m, a size that reflects the diameter of typical kimberlite pipes. Several lakes in the area are significantly colder than others of the same size. In particular, Point Lake, the site of the first kimberlite discovery in the Lac de Gras area, is considerably colder than other lakes of similar size. Point Lake is the most westerly of the three lakes within the middle circle of inset map B (Fig. 2). To the authors'knowledge the other two lakes are not underlain by kimberlite. This method also correctly indicates a

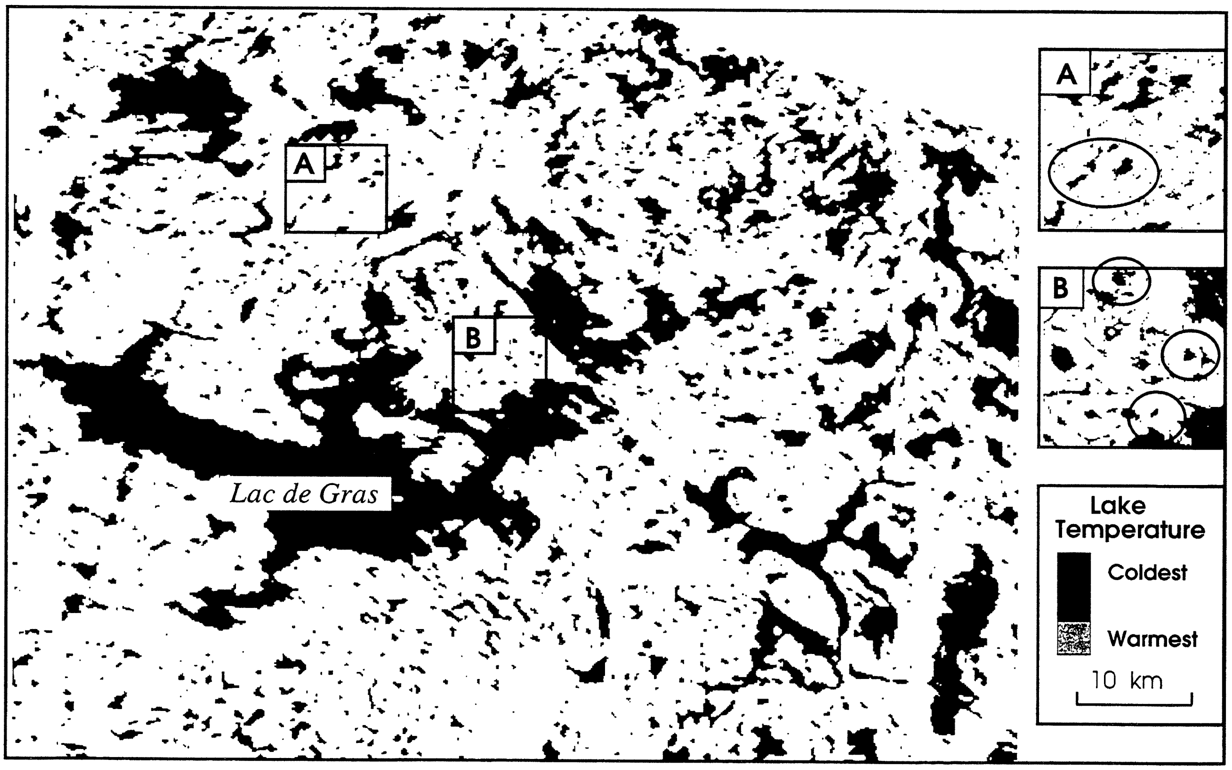


*Figure 2. Relative lake water temperature derived from LANDSAT imagery. Refer to text for discussion of inset maps A and B.*

lake over a kimberlite pipe circled at the northern edge of inset B. However, the lake circled in the southern part of inset B is known to cover a kimberlite, but the lake does not show an obvious temperature difference. Similarly, three lakes circled in inset A (Fig. 2) are considered to be underlain by kimberlite, but only two of the three display a temperature anomaly.

## CONCLUSION

In the Lac de Gras environment, LANDSAT thermal imagery can be used to identify locations of certain kimberlite pipes. The method is dependent on a significant temperature difference between lakes of similar size, being associated with kimberlite pipes. The technique is applicable in this particular geological setting, however this phenomenon may not extend to other geological settings.

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* Contribution to the Geological Survey of Canada's Slave Province NATMAP (National Geoscience Mapping) Program.

# GIS activities related to diamond research and exploration, Lac de Gras area, District of Mackenzie, N.W.T.

**C. Bowie, B.A. Kjarsgaard, H.J. Broome, and A.N. Rencz**

*Bowie, C., Kjarsgaard, B.A., Broome, H.J., and Rencz, A.N., 1996: GIS activities related to diamond research and exploration, Lac de Gras area, District of Mackenzie, N.W.T.; in Searching for Diamonds in Canada, A.N. LeCheminant, D.G. Richardson, R.N.W. DiLabio, and K.A. Richardson (ed.); Geological Survey of Canada, Open File 3228, p. 259-263.*

## INTRODUCTION

Successful exploration for kimberlites in Canada requires a multi-disciplinary approach. Bedrock geologists, surficial geologists, geophysicists, and geochemists must work together as an integrated unit to identify and evaluate prospective diamondiferous terranes. Effective utilization of the information base generated by each of these disciplines is enhanced by technologies that provide tools tailored for management and analysis of geoscience data. One such technology is a Geographic Information System (GIS).

This paper describes how GIS technology is being applied to support kimberlite research and mapping projects in the Lac de Gras region of the Slave Province. Discussion focuses on how GIS aided this research in aspects of data management, data distribution, and data analysis.

## BACKGROUND

The GIS methodology described in this paper was developed as part of the National Mapping Program (NATMAP), which was conceived in 1988 by the Geological Survey of Canada as a new initiative aimed at increasing the level of geoscientific mapping in Canada through multi-institutional and multi-disciplinary projects. One of the inaugural NATMAP projects, the Slave Province Project, is a co-operative effort led by the Continental Geoscience Division of the GSC in collaboration with other GSC Divisions (Mineral Resources Division, Terrain Sciences Division), the Canada-NWT Mineral Initiatives Office (MIO), and the Geology Office of the Department of Indian Affairs and Northern Development (DIAND). The goals of this project are to expand the geoscience database for the Slave Province through increased bedrock and surficial mapping and associated multidisciplinary studies including geochronology, and geochemical and geophysical investigations. For a concise description of how the Slave NATMAP project utilizes a Geographic Information System (GIS) and a linked relational database for management, distribution and analysis of the geoscience data sets, the reader is referred to Broome, et al. (1993).

With the discovery of diamonds in the Lac de Gras region of the Slave Province, efforts were made to expand the amount of up-to-date geoscientific information for this region. This included detailed regional bedrock and surficial mapping, till sampling for geochemistry and kimberlite indicator minerals, and high resolution airborne magnetic and gamma ray surveys. In addition, ancillary geoscientific information such as satellite imagery and digital 1:50 000 topographic maps were obtained.

## WHAT IS GIS ?

Geographic Information System technology has grown exponentially in its diversity of uses and applications. Hence, numerous definitions of what a GIS is have evolved, and there is no single, simple all-encompassing definition of a GIS. In light of this, the following definition taken from Environmental Systems Research Institute , ESRI (1991), is offered as a preliminary reference point:

> *Geographic Information System - An organized collection of computer hardware, software, geographic data, and personnel designed to efficiently capture, store, update, manipulate, analyze, and display, all forms of geographically referenced information.*

For a more concise definition related to geological applications, the reader is referred to Bonham-Carter (1994). The main point to remember when trying to define a GIS is that its ultimate purpose is to provide support for making decisions based on spatial data.

## DATA MANAGEMENT

A GIS was chosen for the Lac de Gras geoscience database because of its intrinsic ability to administer spatial information. Unlike traditional Relational

Database Management Systems (RDBMS), a GIS stores information using two major attribute types; descriptive (what is the measurement of the phenomenon) and spatial (what is the location and extent of the phenomenon being measured). The incorporation of the spatial attribute in its database management capabilities allows the GIS to compare and combine data at varying resolutions, scales and projections.

The GIS also provide a robust set of tools for populating the database with datasets originally available in a range of varying file formats, data structures, and map projections. These tools allow data from each of the contributing disciplines to be easily incorporated into the central GIS database file format with a common map projection. For example, structural measurements from field observations may be transferred from a field geologist as point data structures in a DXF file format, remotely sensed imagery and geophysical data may be transferred from a geophysicist as raster data structures in a variety of file formats, and hydrological data may be purchased from government mapping agencies as line data structures in an ISIF file format.

Table 1 lists the datasets which reside in the Lac de Gras database. In addition to the raw datasets listed, derivative data and interpretations from the raw data are also stored in the database. The main benefit of including the interpretive results along with the raw data, is that scientists from all disciplines may quickly and easily compare and share their interpretations with those of others.

**Table 1.** Lac de Gras digital Database

| MAPPING PRODUCTS | | | |
|---|---|---|---|
| **Data Type** | **Data Scale/Resolution** | **Data Format** | **Data Source** |
| Bedrock Geology | 1:50 000<br>1:250 000<br>1:1 000 000 | point/line/polygon<br>point/line/polygon<br>point/line/polygon | GSC Open Files 2739*, 2966*, 2967*<br>GSC Open File 2740*<br>Hoffman and Hall (1993) |
| Surficial Geology | 1:250 000 | point/line/polygon | GSC Open File 2798* |
| Ice Flow Indicators | 1:250 000 | point/line | GSC Open File 2808* |
| Proterozoic Dyke Swarms | 1:250 000 | line | GSC Open File 2975* |
| MINERAL INVENTORIES, GEOCHEMISTRY AND GEOCHRONOLOGY | | | |
| **Data Type** | **Data Scale/Resolution** | **Data Format** | **Data Source** |
| Kimberlite Pipes and Grades | variable | point | Pell (1995) and other sources |
| Lake Sediment Geochemistry | 2 mile sample grid | point | Allen et al. (1973),<br>Kjarsgaard et al. (1992) |
| Till Geochemistry | 10 samples/1:50 000 NTS sheet | point | GSC Open Files 2867*, 2868*, 2908* |
| Kimberlite Indicator Mineral Inventories | 3 samples/1:50 000 NTS sheet | point | Dredge et al. (1995); Kerr et al. (1995); Ward et al. (1995) |
| Geochronology | variable | point | GSC Open File 2972* |
| GEOPHYSICAL AND REMOTELY SENSED IMAGERY | | | |
| **Data Type** | **Data Scale/Resolution** | **Data Format** | **Data Source** |
| Airborne Total Field Magnetics (levelled) | 800m, 200 | raster grid | Geophysical Data Centre, GSC |
| Airborne Total Field Magnetics (un-levelled) | 250m line spacing, 50 m grid | raster grid | GSC Airborne Geophysics Section |
| Airborne Radiometrics | 250m line spacing, 50m grid | raster grid | GSC Airborne Geophysics Section |
| Bouguer Gravity | 8km sample spacing | raster grid | Geophysical Data Centre, GSC |
| Rock Density | 8km sample spacing | raster grid | Geophysical Data Centre, GSC |
| LANDSAT Thematic Mapper Imagery | 30m (7 bands) | raster grid | RADARSAT International |
| ERS-1 Radar Imagery | 30m (1 band) | raster grid | RADARSAT International |
| Scanned Airphotos | variable | raster grid | National Airphoto Library |
| TOPOGRAPHIC DATA | | | |
| **Data Type** | **Data Scale/Resolution** | **Data Format** | **Data Source** |
| NTS Hydrology | 1:250 000, 1:50 000 | point/line/polygon | Geomatics Canada |
| NTS Contours | 1:250 000, 1:50 000 | point/line/polygon | Geomatics Canada |
| * Detailed GSC Open File reference provided in Table 2. | | | |

## DATA DISTRIBUTION

In order for the database to be of maximum value to project participants from all disciplines, the data must be readily available. GIS provides the flexibility required to generate and distribute standardized and customized products. Data may be distributed as hardcopy maps, tabular reports, or digital files with relative ease. For example, during the 1994 field season, regional bedrock mapping in the Lac de Gras area (Kjarsgaard et al., 1994) was supported by both hardcopy maps and digital products derived from the database. Hardcopy products included LANDSAT TM imagery enhanced to discriminate between rock and vegetation, which was overlain by linework representing geological contacts and dykes derived from previous mapping projects. These maps were carried by each of the mapping parties on traverses to aid in mapping contacts and outcrop. Digital products used in the mapping project included a compilation of all of the geoscientific datasets archived for the region stored in CD-ROM format. The data on the CD-ROM were provided in a format compatible with ARCVIEW, a Windows 3.1 display and query software package. Relationships between the various datasets could be quickly viewed on laptop computers in the field camp as an aid to the mapping process.

Data release to the public is also facilitated by use of the GIS. Customized hardcopy maps, such as the overlay of 1:1 000 000 geology on shaded magnetics (Bowie, 1994), are easily created using GIS integration techniques. The digital nature of these maps allows them to be "plotted on demand", where a copy of the map is created from the GIS upon receiving a request for the map. This reduces costs from the more traditional map production techniques, which required estimating the number of maps necessary to meet demands, and then subsequently printing all of these maps at once.

To date, two CD-ROM compilations of data have been released to the public. Both releases use two common interchange file formats (DXF and ARC/INFO ASCII interchange) to insure maximum compatibility of the data with a wide range of GIS and CAD systems. The first CD-ROM released under this project as GSC Open File 2559, contained a 1:1 000 000 scale geological compilation of the Slave craton and environs (Hoffman and Hall, 1993), and a 1:1 000 000 scale hydrological base dataset derived from the Digital Chart of the World dataset. The second CD-ROM release was more ambitious, in that it included various maps and supporting data sets from a wider range of disciplines (Bowie, 1995). Table 2 lists the contents of this CD-ROM. In addition, a GSC-developed software package, SurView (Grant, 1993), was also included on the disc to allow for display, query, and plotting of the various datasets.

**Table 2.** Contents of GSC Open File 2974: Selected geoscience data for the Slave Province NATMAP Project, District of Mackenzie, Northwest Territories (Bowie, 1995).

| | |
|---|---|
| **BEDROCK GEOLOGY OPEN FILES** | |
| Kjarsgaard, B.A. (1993)<br>Geology of the Paul Lake area (NTS 76D/9), Lac de Gras, NWT; Geological Survey of Canada Open File 2739 (1:50 000 scale map). | |
| Thompson, P.H. and Kerswill, J.A. (1993)<br>Preliminary geology of the Winter Lake-Lac de Gras area, District of Mackenzie, N.W.T; Geological Survey of Canda Open File 2740 (1:250 000 scale map). | |
| Henderson, J.R., Henderson, M.N., and Kerswill, J.A. (1994)<br>Preliminary geological map of north central High Lake Greenstone Belt, NWT (NTS 76M/6E half; 76M/7): Geological Survey of Canada Open File 2782 (1:50 000 scale map). | |
| Kjarsgaard, B.A., Sparks, R.N., and Jakop, Z.J. (1994)<br>Preliminary geology, Koala, District of Mackenzie, NWT (NTS 76D/10); Geological Survey of Canada Open file 2966 (1:50 000 scale map). | |
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| **SURFICIAL GEOLOGY OPEN FILES** | |
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## ANALYSIS AND DECISION SUPPORT

A major strength of a GIS is its ability to represent an abstraction or simplification of the real world (in the form of data), and via user-guided manipulation of its relational database and software tools, derive conclusions on the relationships of these datasets to a specific problem. In kimberlite and diamond exploration, the main purpose is to identify prospective areas favourable for deposits of economic grade. In Canada, this process relies on a top down approach, whereby regions are ranked on their suitability to host economic kimberlites or lamproites, high ranking regions are investigated further by sampling for kimberlite indicator minerals and testing for geophysical anomalies. Targets are prioritized and drilled. This approach lends itself quite well to a GIS analysis paradigm.

Mimmi (1993), developed a GIS-based regional analysis system for identifying prospective regions of kimberlite and lamproite emplacement for the north-central continental United States, called the Diamond Exploration Geoscientific Information System (DEGIS). This system uses publicly available geoscience data and three different raster-based data models (empirical, staked, and proximity) to delimit terrane permissive for economic primary diamond deposits. The empirical model uses two critical factors in its prediction; age of Precambrian basement rocks and lithospheric thickness. The stacked model is the additive overlay of known and inferred point and linear features considered to be important to the emplacement and discovery of kimberlites and lamproites. The proximity model assesses the distance of each cell to the linear and point features in the stacked model in addition to the distance to "hinges" or centre lines of upwarps in the crust. The results of all three models are integrated to identify prospective zones. The efficiency of this modelling process is yet to be determined.

Kaminsky et al. (1995), proposed a similar system based on regional geology, structures, and tectonic environments to derive potential sites in Russia and Australia. The efficiency of this model has been proven in practice, and over the last five years four new kimberlite fields were predicted and subsequently discovered in northern Siberia, the Arkhangelsk region, Ukraine, and Belorussia. Similar analysis techniques could potentially be applied to the Slave Province, and Canada as a whole.

In the Slave Province, GIS will be applied to aid the identification of prospective kimberlite targets and to gain a better understanding of their environmental controls. For example, studies are currently underway to evaluate the spatial relationships of known kimberlites in the Lac de Gras region to local structural controls, such as dykes, faults, and fracture patterns derived from bedrock mapping and interpreted from airphotos and satellite imagery. Using the GIS, the locations of each kimberlite pipe are related to each of these features and a statistical measure of the probability of the occurrence is compared to the probabilities of random occurrence in the area. This statistical measure is then used to determine if any of these features have an influence on the placement of the pipes. Knowledge derived from this analysis will be combined with existing knowledge of normalized kimberlite indicator mineral counts (Ward et al., 1996) from surficial studies and geophysical anomalies to identify and rank prospective areas. Development of models based on small regions with abundant data will then be applied to larger regions with lower data densities.

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* Contribution to the Geological Survey of Canada's Slave Province NATMAP (National Geoscience Mapping) Program.

# APPENDIX A: Contributors Names and Addresses

**Y. Amelin (Yuri)**
Department of Geology
Royal Ontario Museum
100 Queen's Park
Toronto, Ontario
Tel: 416-586-5811
Fax: 416-586-5814
E-mail: yuria@rom.on.ca

**A.E. Armitage (Al)**
Department of Earth Sciences
University of Western Ontario
London, Ontario, N6A 5B7
Tel: 519-661-3178
E-mail: aarmitag@julian.uwo.ca

**A.D.Armstrong (Derek)**
Ontario Geological Survey
933 Ramsey Lake Road
Sudbury, Ontario, P3E 6B5
Tel: 705-670-5913
Fax:705-670-5905
E-mail:armstrong_d@torv05.ndm.gov.on.ca

**W.R.A. Baragar (Bob)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4864
Fax: 613-995-9273
E-mail: baragar@gsc.emr.ca

**J.H. Bédard (Jean)**
Centre géoscientifique de Québec,
Commission géologique du
Canada-Québec
2535 Laurier Blvd.
Sainte-Foy, Québec, G1V 4C7
Tel: 418-654-2671
Fax: 418-654-2615
E-mail: bedard@gsc.emr.ca

**R.G. Berman (Rob)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-6368
Fax: 613-995-9273
E-mail: berman@gsc.emr.ca

**M. Bostock (Micheal)**
Department of Geophysics and
Astronomy
University of British Columbia
2075 Wesbrook Place
Vancouver, British Columbia,
V6T 1Z4
Tel: 604-822-2082
Fax: 604-822-9014
E-mail: bostock@geop.ubc.ca

**C. Bowie (Cameron)**
INCO Exploration
Highway 17 West
Copper Cliffs, Ontario
Tel: 705 -682-8491
E-mail:

**H.J. Broome (John)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario. K1A 0E8
Tel: 613-995-6914
Fax: 613-947-1819
E-mail: broome@gsc.emr.ca

**K.L. Buchan (Ken)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4386
Fax: 613-995-7997
E-mail: kbuchan@gsc.emr.ca

**J.F. Cassidy (John)**
Geological Survey of Canada-
Pacific
9860 West Saanich Road
P.O. Box 6000
Sidney, British Columbia,
V8L 4B2
Tel: 604-363-6382
Fax: 604-363-6565
E-mail: cassidy@pgc.emr.ca

**L. Corriveau (Louise)**
Centre géoscientifique de Québec,
Commission géologique du
Canada-Québec
2535 Laurier Blvd.
Sainte-Foy, Québec, G1V 4C7
Tel: 418-654-2672
Fax: 418-654-2615
E-mail: corriveau@gsc.emr.ca

**K.L. Currie (Ken)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-1208
Fax: 613-995-7997
E-mail: currie@gsc.emr.ca

**A. Davidson (Tony)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4793
Fax: 613-995-9273
E-mail: tdavidson@gsc.emr.ca

**W.J. Davis (Bill)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-943-8780
Fax: 613-995-7997
E-mail: bdavis@gsc.emr.ca

**R.N.W. DiLabio (Ron)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-1380
Fax: 613-992-2468
E-mail: dilabio@gsc.emr.ca

**L.A. Dredge (Linda)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-5770
Fax: 613-992-2468
E-mail: ldredge@gsc.emr.ca

**R. Dumont (Regis)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-0735
Fax: 613-992-2787
E-mail: dumont@agg.emr.ca

**C.E. Dunn (Colin)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-996-2373
Fax: 613-996-3726
E-mail: dunn@gsc.emr.ca

**D.W. Eaton (Dave)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-947-2783
Fax: 613-992-8836
E-mail: eaton@cg.emr.ca

**R.E. Ernst (Richard)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4386
Fax: 613-995-7997
E-mail: rernst@gsc.emr.ca

**R.G. Garrett (Bob)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4517
Fax: 613-996-3726
E-mail: garrett@gsc.emr.ca

**D.J. Gendzwill (Don)**
Department of Geological
Sciences
University of Saskatchewan
114 Science Place
Saskatoon, Saskatchewan,
S7N 5E2
Tel: 306-966-5693
Fax:306-966-8593
E-mail: don.gendzwill@sask.usask.ca

**R.A.F. Grieve (Richard)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-5372
Fax: 613-952-8987
E-mail: impact@gsc.emr.ca

**L.M. Heaman (Larry)**
Department of Geology
University of Alberta, 1-26 Earth
Sciences Building
Edmonton, Alberta, T6G 2E3
Tel: 403-492-3265
Fax: 403-492-2030
E-mail:lheaman@gpu.srv.ualberta.ca.

**D.D. Hogarth (Don)**
Department of Geology
University of Ottawa
Ottawa, Ontario, K1N 6N5
Tel: 613-562-5800 (loc 6207)
Fax: 613-562-5192

**J.A.M. Hunter (Jim)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-2650
Fax: 613-992-2468
E-mail: jhunter@gsc.emr.ca

**O.J. Ijewliw (Olga)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-943-8060
Fax: 613-995-9273
E-mail: ijewliw@gsc.emr.ca

**A.G. Jones (Alan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-4968
Fax: 613-992-8836
E-mail: jones@cg.emr.ca

**A.L. Jones (Adrienne)**
Department of Earth Sciences
University of Western Ontario
London, Ontario, N6A 5B7
Tel: 519-661-3178
E-mail: ajones@julian.uwo.ca

**A.S. Judge (Alan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-996-9323
Fax: 613-992-2468
E-mail:ajudge@gsc.emr.ca

**T.J. Katsube (Jon)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-5239
Fax: 613-992-2468
E-mail: jkatsube@gsc.emr.ca

**P. Keating (Pierre)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-996-9318
Fax: 613-952-8987
E-mail: keating@gsc.emr.ca

**D.E. Kerr (Dan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4523
Fax:613-992-2468
E-mail:

**P.G. Killeen (Pat)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-996-2312
Fax: 613-996-3726
E-mail: pkilleen@gsc.emr.ca

**B.A. Kjarsgaard (Bruce)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-5705
Fax: 613-995-7997

**I.M. Kjarsgaard (Ingrid)**
Private Consultant
15 Scotia Place
Ottawa, Ontario
Tel: 613-730-0136

**A.N. LeCheminant (Tony)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4850
Fax: 613-995-7997

**T.J. Lewis (Trevor)**
Geological Survey of Canada-Pacific
9860 West Saanich Road
P.O. Box 6000
Sidney, British Columbia,
V8L 4B2
Tel: 604-363-6438
Fax:604-363-6739
E-mail:lewis@slocan.pgc-emr.ca

**S.B. Lucas (Steve)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4534
Fax: 613-995-9273
E-mail: slucas@gsc.emr.ca

**N.D. MacRae (Neil)**
Department of Earth Sciences
University of Western Ontario
London, Ontario, N6A 5B7
Tel: 519-661-3178
Fax: 519-661-3198
E-mail: nmacrae@julian.uwo.ca

**V.L. Masaitis (Victor)**
Karpinsky Geological Institute
Sredny prospekt 74
St. Petersburg 199026
Russia
Tel: 7-812-218-92-06
Fax: 7-812-213-57-38
E-mail: vsg@sovamsu.sovusa.com

**S.D. Matieshin (Scott)**
Department of Geological Sciences
University of Saskatchewan
114 Science Place
Saskatoon, Saskatchewan,
S7N 5E2
Tel: 306-966-5700
Fax:306-966-8593

**M.B. McClenaghan (Beth)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-7805
Fax: 613-992-2468
E-mail: bmcclenaghan@gsc.emr.ca

**A.D. McCracken (Sandy)**
Geological Survey of Canada-Calgary
3303-33rd Street N.W.
Calgary, Alberta, T2L 2A7
Tel: 403-292-7089
Fax: 403-292-6014
E-mail: mccracken@gsc.emr.ca

**D.C. McGregor (Colin)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel:613-995-4680
Fax:
E-mail: mcgregor@gsc.emr.ca

**D.J. McIntyre (Dave)**
Geological Survey of Canada-Calgary
3303-33rd Street N.W.
Calgary, Alberta, T2L 2A7

**A.R. Miller (Al)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4106
Fax: 613-996-9820
E-mail: amiller@gsc.emr.ca

**D. Morin (David)**
Centre géoscientifique de Québec,
Institut national de la Recherche scientifique, Géoressources
2535 boul Laurier,
P.O. Box 7500
Sainte-Foy, Quebec, G1V 4C7
Tel: 418-654-2604
Fax: 418-654-2672
E-mail: dmorin@gsc.emr.ca

**D. Moser (Desmond)**
Department of Geology
Royal Ontario Museum
Toronto, Ontario, M5S 2C6
Tel: 416-586-5811
Fax: 416-586-5814
E-mail: desmondm@rom.on.ca

**M.P. Mudry (Phillip)**
Comaplex Minerals Corp.
901-1015 4th Street S.W.
Calgary, Alberta, T2R 1J4
Tel: 403-265-2846
Fax: 403-265-7488
E-mail: mudryp@cadvision.com

**C.J. Mwenifumbo (Jonathan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-6520
Fax: 613-996-3726
E-mail: jarako@gsc.emr.ca

**W.W. Nassichuk (Walter)**
Geological Survey of Canada-Calgary
3303-33rd Street N.W.
Calgary, Alberta, T2L 2A7
Tel: 403-292-7150
Fax: 403-292-5377
E-mail: wnassichuk@gsc.emr.ca

**R.R. Parrish (Randy)**
Department of Geology
Leicester University
University Road
Leicester, LE1 7RH, U.K.
Fax: 011 44 115 936 3302
E-mail: rrp@le.ac.uk

**J. Pell (Jennifer)**
Northwest Territories Geological Mapping Division
Department of Indian Affairs and Northern Development
Box 1500
Yellowknife, NWT, X1A 2R3
Tel: 403-669-2640
Fax: 403-669-2709
E-mail:Jennifer.Pell@INAC.X400.gc.ca

**J.A. Percival (John)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4723
Fax: 613-995-9273
E-mail: jpercival@gsc.emr.ca

**T.D. Peterson (Tony)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-3573
Fax: 613-995-7997
E-mail: peterson@gsc.emr.ca

**G.J. Pringle (Gord)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-5006
Fax: 613-943-1286
E-mail: gpringle@gsc.emr.ca

**A.N. Rencz (Andy)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8
Tel: 613-995-4786
Fax: 613-996-3726
E-mail: arencz@gsc.emr.ca

**D.G. Richardson (Dan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8
Tel: 613-996-9151
Fax: 613-992-5051
E-mail: drichardson@gsc.emr.ca

**K.A. Richardson (Keith)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8
Tel: 613-992-4078
Fax: 613-996-3726
E-mail: krichardson@gsc.emr.ca

**B. Ryan (Bruce)**
Newfoundland Geological Survey Branch
95 Bonaventure Ave.
P.O. Box 8700
St. John's, Newfoundland, A1B 4J6
Tel: 709-729-2111
Fax: 709-729-3493
E-mail:
abr@zeppo.geosurv.gov.nf.ca

**D.J. Schulze (Dan)**
Department of Geology
University of Toronto
Erindale College
Mississauga, Ontario
L5L 1C6
Tel: 905-828-3970
Fax: 905-828-3717
E-mail:dschulze@credit.erin.utoronto.ca

**R.A. Stern (Richard)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8
Tel: 613-995-8935
Fax: 613-995-7997

**L.D. Stasiuk (Vern)**
Geological Survey of Canada-
Calgary
3303-33rd Street, N.W.
Calgary, Alberta, T2L 2A7
Tel: 403-292-7199
Fax:403-292-5377
E-mail:lstasiuk@gsc.emr.ca

**J.A.R. Stirling (John)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-943-0270
Fax: 613-943-1286
E-mail: stirling@gsc.emr.ca

**M.R. St-Onge (Marc)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-4935
Fax: 613-995-9273
E-mail: mstonge@gsc.emr.ca

**M. Tellier (Maxime)**
Centre géoscientifique de Québec,
Institut national de la Recherche
scientifique, Géoressources
2535 boul Laurier,P.O. Box 7500
Sainte-Foy, Quebec, G1V 4C7
Tel: 418-654-2604
Fax: 418-654-2672
E-mail: tellier@gsc.emr.ca

**P.H. Thompson (Peter)**
Geological Surrvey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-943-8950
Fax:613-995-7997
E-mail:pthompson@gsc.nrcan.ga.ca

**L.H. Thorleifson (Harvey)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-3643
Fax: 613-992-2468
E-mail: thorleifson@gsc.emr.ca

**J. Tod (Joan)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-6438
Fax: 613-992-2787
E-mail: jtod@gsc.emr.ca

**O. van Breemen (Otto)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-995-0810
Fax: 613-995-7997
E-mail: ovanbreemen@gsc.emr.ca

**B.C. Ward (Brent)**
Prince George Forest Region
5th Floor-1011 4th Avenue,
Prince George, B.C., V2L 3H9
Tel: 604-565-6174

**D. White (Don)**
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario, K1A 0E8
Tel: 613-992-0758
Fax: 613-992-8836
E-mail: white@cd.emr.ca