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GEOLOGY OF THE WIJINNEDI LAKE AREA, NORTHWEST TERRITORIES

John B. Henderson

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Cover illustration

Due to the juxtaposition of the various lithotectonic domains that make up the Wijinnedi map area, the same Archean Yellowknife Supergroup greywacke-mudstone turbidite protolith varies in metamorphic grade from chlorite-zone pelite to orthopyroxene-bearing migmatite. In these photomicrographs, crenulated chlorite-zone pelite is superimposed by a potassium-feldspar-zone migmatite in which a perthite porphyroblast is partially retrogressed to muscovite.

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GEOLOGY OF THE WIJINNEDI LAKE AREA, NORTHWEST TERRITORIES

Abstract

The geology of the Wijinnedi Lake area, presented as a 1:50 000-scale map sheet (Map 2023A) in the southwestern Archean Slave Province, is divided into four lithotectonic domains. The three western domains are separated by Archean ductile shear zones. In the northwest, the structurally highest Wijinnedi domain consists of a 2670 Ma Yellowknife Supergroup felsic volcanic centre, partly mantled by more mafic volcanic rocks that are overlain by a tightly to isoclinally folded, pelite-dominated siltstone and fine-grained greywacke turbidite sequence. Metamorphic grade in the lower pressure facies series ranges from lower greenschist to middle amphibolite, but reaches migmatite grade locally at the east end of the domain. To the south, the Hinscliffe domain consists mainly of a polyphase, largely recrystallized, ca. 2.65 Ga trondhjemite complex intruded by mafic amphibolite synplutonic dykes. To the southeast, the Ghost domain, representing mid-crustal structural levels, is dominated by plutonic bodies, with the areally most extensive tonalite and related tonalite gneiss intruded about 2.64–2.63 Ga and a group of mainly granite intrusions emplaced at about 2.60 Ga. The less abundant Yellowknife Supergroup pelitic metasedimentary rocks and intermediate volcanic rocks are represented by migmatite. Most of the domain is at granulite grade with most of the plutonic bodies crystallizing under granulite-grade conditions. The Dauphinee domain in the eastern part of the area is separated from the high-grade rocks to the west by a Paleoproterozoic cataclastic shear zone; it consists mainly of quartz dioritic to granitic rocks and minor metasedimentary migmatite that are not at granulite grade.

Résumé

La région du lac Wijinnedi, dans le sud-ouest de la province archéenne des Esclaves, se subdivise en quatre domaines lithotectoniques dont la géologie est présentée à l'échelle de 1/50 000 sur la carte 2023A. Les trois domaines à l'ouest sont séparés par des zones de cisaillement ductile archéennes. Dans le nord-ouest, le domaine de Wijinnedi, structuralement le plus élevé, est constitué d'un centre volcanique felsique de 2670 Ma qui fait partie du Supergroupe de Yellowknife et qui est partiellement recouvert de roches volcaniques davantage mafiques, elles-mêmes recouvertes par une séquence turbiditique composée de grauwacke à grain fin et de siltstone à prédominance de pélite et déformée en plis serrés à isoclinaux. Dans les séries métamorphiques de pression moins élevée, le degré de métamorphisme va du sous-faciès inférieur des schistes verts au sous-faciès intermédiaire des amphibolites, mais il atteint par endroits le faciès des migmatites à l'extrémité est du domaine. Au sud se situe le domaine de Hinscliffe, formé principalement d'un complexe polyphasé de trondhjémite de 2,65 Ga qui est largement recristallisé et que recoupent des dykes synplutoniques d'amphibolite mafique. Au sud-est se trouve le domaine de Ghost, qui représente des niveaux structuraux intermédiaires de la croûte et où prédominent des massifs plutoniques; la tonalite et le gneiss tonalitique, massifs les plus répandus, ont été mis en place vers 2,64 à 2,63 Ga alors qu'un groupe d'intrusions principalement granitiques a été mis en place vers 2,60 Ga. Les roches métasédimentaires pélitiques et les roches volcaniques intermédiaires moins abondantes du Supergroupe de Yellowknife sont représentées par de la migmatite. La plus grande partie du domaine a été métamorphisée dans le faciès des granulites, et la plupart des massifs plutoniques ont cristallisé dans des conditions associées au faciès des granulites. Le domaine de Dauphinee est situé dans l'est de la région; il est séparé des roches de fort métamorphisme à l'ouest par une zone de cisaillement cataclastique paléoprotérozoïque et comporte principalement des diorites quartziques passant à des roches granitiques et de faibles quantités de migmatite métasédimentaire qui n'appartiennent pas au faciès des granulites.

SUMMARY

Introduction

The Wijinnedi Lake map area, which consists of the Ghost Lake (NTS 85-O/14) and the western 10 minutes of the Dauphinee Lake (NTS 85-O/15) 1:50 000-scale map areas, is situated near the southwestern margin of the Slave Province, at the southern end of the Indin Lake supracrustal terrane.

SOMMAIRE

Introduction

La région cartographique du lac Wijinnedi englobe la région cartographique du lac Ghost (SNRC 85-O/14) et la portion ouest (10 minutes) de la région cartographique du lac Dauphinee (SNRC 85-O/15), les deux régions ayant été cartographiées à l'échelle de 1/50 000. Elle se trouve près de la marge sud-ouest de la Province des Esclaves, à l'extrémité sud du terrane supracrustal du lac Indin.

The area is underlain almost entirely by Archean rocks consisting of metasedimentary and metavolcanic rocks of the Yellowknife Supergroup that range in grade from lower greenschist to granulite, and a complex assemblage of granitoid rocks, foliated granitoid rocks, and granitoid gneiss, many of which are at granulite grade. The only known Proterozoic rocks include the extensively developed Paleoproterozoic Indin diabase dykes and at least one example of a Mesoproterozoic Mackenzie diabase dyke.

The area can be divided into four distinct, structurally bound domains. In the northwest, the Wijinnedi domain consists mainly of Yellowknife Supergroup supracrustal rocks that are for the most part at relatively low metamorphic grade and commonly contain well preserved primary features. To the south, sharply separated from the Wijinnedi domain by a high-grade shear zone, is the Hinscliffe domain consisting of variably deformed granitoid rocks with mainly mafic inclusions and few, if any, supracrustal rocks. The Ghost domain dominates the southern part of the area and extends through to the northern border; it contains both high-grade supracrustal rocks and a complex assemblage of granitoid intrusions and gneiss and is mostly at granulite grade. The domain boundary is a sharply defined, high-grade shear zone. Each of these three western domains is dominated by easterly to northeasterly trending, anticlinal or antiformal structures. The eastern part of the area consists of the Dauphinee domain. It is bounded by a low-grade, cataclastic shear zone and consists mainly of massive to weakly foliated, granitoid rocks with less abundant metasedimentary migmatite.

Wijinnedi domain

The dominant structure in the Wijinnedi domain is an east-trending anticline with an elongate dome of dominantly coarse-grained, intermediate to felsic volcaniclastic rocks in its core that closes to the west about 4 km beyond the map area. Its trend is at a high angle to the more northerly trends typical of the supracrustal rocks of the Indin Lake region. Zircons from these volcanic rocks have been dated at about 2675 Ma , which is toward the young end of the Yellowknife Supergroup age spectrum, but somewhat older than several other felsic volcanic centres in the southwestern Slave Province. The felsic dome is partially mantled by mafic to intermediate, locally pillowed volcanic rocks. Minor amounts of similar, more mafic volcanic rocks also occur within the felsic centre itself. To the east, primary structures in the mafic rocks have been lost due to the increasing degree of deformation and metamorphic grade.

La région se compose presque entièrement de roches archéennes, soit de roches métasédimentaires et métavolcaniques du Supergroupe de Yellowknife, dont le degré de métamorphisme va du sous-faciès inférieur des schistes verts au sous-faciès des granulites, et d'un assemblage complexe de roches granitoïdes, de roches granitoïdes foliées et de gneiss granitoïdes dont bon nombre appartiennent au faciès des granulites. Parmi les seules roches protérozoïques connues sont les dykes de diabase d'Indin très répandus du Paléoprotérozoïque et au moins un dyke de diabase de Mackenzie du Mésoprotérozoïque.

La région peut être subdivisée en quatre domaines structuralement délimités et distincts. Dans le nord-ouest, on trouve le domaine de Wijinnedi, qui est constitué principalement de roches supracrustales du Supergroupe de Yellowknife, dont la plupart sont de métamorphisme relativement faible et présentent généralement des caractéristiques primaires bien conservées. Au sud se situe le domaine de Hinscliffe, qui est nettement séparé du domaine de Wijinnedi par une zone de cisaillement fortement métamorphisée. Il est composé de roches granitoïdes qui ont été déformées à divers degrés et qui contiennent des inclusions principalement mafiques, ainsi que de quelques roches supracrustales, sinon aucune. Le domaine de Ghost couvre la plus grande partie du sud de la région et se prolonge jusqu'à la limite nord. Il renferme des roches supracrustales de fort métamorphisme et un assemblage complexe d'intrusions granitoïdes et de gneiss; ces roches appartiennent en majeure partie au faciès des granulites. Le domaine de Ghost est délimité par une zone de cisaillement fortement métamorphisée et nettement définie. Ces trois domaines occidentaux sont tous dominés par des antiformes ou des anticlinaux à direction généralement est à nord-est. Le secteur oriental de la région est constitué du domaine de Dauphinee. Ce dernier est délimité par une zone de cisaillement cataclastique faiblement métamorphisée et renferme principalement des roches granitoïdes massives à légèrement foliées ainsi que des quantités moindres de migmatite métasédimentaire.

Domaine de Wijinnedi

La principale structure du domaine de Wijinnedi est un anticlinal à direction est qui se ferme à l'ouest, à environ 4 km au-delà de la région cartographique. Cet anticlinal comporte un dôme allongé dont le noyau est composé en majeure partie de roches volcanoclastiques intermédiaires à felsiques à grain grossier. Sa direction est à angle aigu par rapport aux directions davantage vers le nord typiques des roches supracrustales de la région du lac Indin. La datation de zircons provenant de ces roches volcaniques donne un âge d'environ 2675 Ma, ce qui correspond à l'extrémité récente de la plage d'âges du Supergroupe de Yellowknife, mais est quelque peu supérieur à l'âge de plusieurs autres centres volcaniques felsiques du sud-ouest de la Province des Esclaves. Le dôme felsique est partiellement recouvert de roches volcaniques mafiques à intermédiaires qui forment, par endroits, des coussins. On trouve également de petites quantités de roches volcaniques similaires mais davantage mafiques dans le centre volcanique felsique lui-même. À l'est, les structures primaires autrefois présentes dans les roches mafiques ont disparu en raison du degré croissant de déformation et de métamorphisme.

Mudstone-dominated metasedimentary rocks of the Yellowknife Supergroup conformably overlie the volcanic rocks. At the contact, the metasedimentary rocks are commonly graphitic and locally gossanous along the southwest shore of Wijinnedi Lake due to the presence of pyrite, pyrrhotite, and rare chalcopyrite. In most cases, thin mafic intrusive sills are also present in these gossanous rocks. The minor sulphide occurrences have attracted economic interest from time to time from the late 1940s to the present. Rare carbonate lenses, usually less than 1 m thick, occur locally at this contact. The bulk of the metasedimentary rocks is a mudstone-dominant turbidite sequence in which the siliciclastic layers are thin and fine grained. Metre-scale, layered units of silicate iron-formation occur east of Wijinnedi Lake as well as north and south of 'Colson Lake'. Where these units contain sulphide minerals, they show local gold anomalies. They are more or less on strike with iron-formation units at Damoti Lake, 20 km to the north, where a major iron-formation-hosted gold prospect occurs. No iron-formation was recognized elsewhere in the metasedimentary rocks of the domain. At least one occurrence of oxide-dominated iron-formation occurs in similar, but very high-grade metasedimentary rocks of the Ghost domain.

Intruded into the core of the volcanic centre is a small gabbro to leucogabbro plutonic complex that has been metamorphosed in context with the rocks it intrudes. Dykes and sills related to this intrusion occur in varied abundance throughout the volcanic complex as well as locally within the metasedimentary rocks. They are particularly abundant southeast and east of the mafic plutons.

The lowest grade rocks occur west of Wijinnedi Lake where biotite is absent in the chlorite-bearing metasedimentary rocks. These lowest grade rocks may represent a retrograde assemblage. The northerly trending metamorphic pattern increases in grade both westward and eastward and is at a high angle to the dominant structural trend of the domain. To the east, the metamorphic zonation includes biotite, cordierite-andalusite, sillimanite, and quartz-plagioclase leucosome migmatite zones. The cordierite-andalusite- and sillimanite-zone boundaries north of the volcanic complex and the migmatite-zone boundary southeast of the complex represent prograde metamorphic isograds. The boundary between the sillimanite and migmatite zones in the north-central part of the area is a fault. Metamorphic grade within the volcanic complex increases southward to the bounding shear zone as well as eastward.

Hinscliffe domain

The Hinscliffe domain is lithologically distinct in that it consists almost entirely of metamorphosed igneous rocks. It is dominated by what was originally grey to pinkish-grey, leucocratic trondhjemite to granodiorite with minor pegmatite. These rocks have been extensively recrystallized and now have a fine-grained, sugary texture,

Des roches métasédimentaires à prédominance de mudstone du Supergroupe de Yellowknife reposent en concordance sur les roches volcaniques. Au contact entre les deux, les roches métasédimentaires sont généralement graphitiques et localement ferrugineuses sur la rive sud-ouest du lac Wijinnedi, ce qui est attribuable à la présence de pyrite, de pyrrhotite et de quantités rares de chalcopyrite. Dans la plupart des cas, on trouve également de minces filons-couches intrusifs mafiques dans ces roches ferrugineuses. Depuis la fin des années 1940, les indices mineurs de sulfures suscitent, par moments, l'intérêt des sociétés. De rares lentilles carbonatées mesurant généralement moins de 1 m d'épaisseur se rencontrent par endroits au contact. Une séquence turbiditique à prédominance de mudstone constitue la plus grande partie des roches métasédimentaires; les couches silicoclastiques y sont minces et à grain fin. À l'est du lac Wijinnedi et au nord et au sud du «lac Colson», on trouve des unités stratifiées d'échelle métrique de formation de fer à faciès silicaté. Là où elles renferment des sulfures, ces unités contiennent des quantités anomales d'or. Elles sont plus ou moins parallèles à la direction des unités de formation de fer au lac Damoti, à 20 km au nord, à l'endroit où se situe un prospect aurifère majeur logé dans une formation de fer. Aucune autre formation de fer n'a été reconnue ailleurs dans les roches métasédimentaires du domaine. Le domaine de Ghost renferme au moins une formation de fer à faciès principalement oxydé dans des roches métasédimentaires similaires, mais très fortement métamorphisées.

Un petit complexe plutonique gabbroïque à leuco-gabbroïque recoupe le noyau du centre volcanique; ce complexe a été métamorphisé dans les mêmes conditions que les roches qu'il recoupe. Un nombre plus ou moins élevé de dykes et de filons-couches apparentés à cette intrusion recoupent le complexe volcanique et, par endroits, les roches métasédimentaires. Ils sont particulièrement abondants au sud-est et à l'est des plutons mafiques.

Les roches le plus faiblement métamorphisées se trouvent à l'ouest du lac Wijinnedi, là où les roches métasédimentaires chloriteuses sont dépourvues de biotite. Ces roches pourraient représenter un assemblage rétromorphosé. Le gradient métamorphique, à direction nord, augmente vers l'ouest et vers l'est et forme un angle aigu par rapport à la direction struturale principale du domaine. À l'est, la zonalité comprend des zones de roches à biotite, à cordiérite-andalousite et à sillimanite et des zones de migmatite à leucosome de quartz-plagioclase. Les limites des zones de roches à cordiérite-andalousite et à sillimanite au nord du complexe volcanique, ainsi que la limite de la zone de migmatite à leucosome au sud-est du complexe, représentent des isogrades de métamorphisme prograde. Dans le centre nord de la région, la limite séparant les zones de roches à sillimanite et de migmatite est une faille. Le degré de métamorphisme au sein du complexe volcanique augmente vers le sud jusqu'à la zone de cisaillement limitrophe, ainsi que vers l'est.

Domaine de Hinscliffe

Le domaine de Hinscliffe est lithologiquement distinct, car il se compose presque entièrement de roches ignées métamorphisées. Il renferme surtout ce qui était, à l'origine, de la trondhjémite leucocrate grise à gris rosâtre passant à de la granodiorite avec de petites quantités de pegmatite. Ces roches ont été profondément recristallisées et présentent maintenant une texture saccharoïde à although in places the original coarser texture of the rocks, particularly the pegmatite, is still apparent. In most places in the complex, these rocks have a gneissic aspect with layering defined by minor variations in composition and texture. Pegmatite, where present, occurs as layer-parallel, commonly disaggregated, partially recrystallized, coarse crystals. The gneissic layering becomes more pronounced toward the bounding shear zones where the rocks become finely laminated. The best preserved rocks occur in the central part of the complex at the east end and south of 'Hinscliffe Lake', where several intrusive phases with evident crosscutting relationships as well as granitoid inclusions can be seen. Zircons from one of the oldest phases have a U-Pb age of about 2655 Ma, whereas titanite from the same rock has an age of about 2610 Ma. The zircon age is somewhat younger than the dominant mode in the age spectrum of Yellowknife Supergroup volcanic rocks in the Slave Province (2.73–2.66 Ga), about 20 Ma younger than the age of the metadacite from the Wijinnedi domain.

The meta-igneous rocks contain amphibolite inclusions in proportions ranging from zero to as much as 60% of the outcrop area. These inclusions are typically uniform in texture and composition, variably deformed, but commonly angular, and crosscut by multiple generations of variably deformed leucogranitoid veins. In rare cases, long, narrow zones of texturally distinct inclusions occur, suggesting that the mafic inclusions may represent the remnants of a suite of synplutonic mafic dykes.

A second suite of mafic inclusions commonly occurs along the margins of the complex. They differ from those previously described in that they are strongly compositionally layered on a scale of less than 10 cm and have a more intermediate average composition than the mafic amphibolite inclusions in the main part of the complex. These inclusions may have had a supracrustal origin, which, together with their marginal distribution, leads to the tentative suggestion that the Hinscliffe domain occurs in an antiformal structure.

Intruded into the granitoid rocks is a minor suite of small pyroxenite bodies to small lenses that are most abundant close to the northern domain contact and also occur in the volcanic rocks to the north. They now consist mainly of actinolitic amphibole. The largest of these bodies occur in central 'Hinscliffe Lake'. Also present are rare, late pegmatite and small, granite veins and small stocks that are considered to be related to the nearby, lithologically similar granite of the Ghost domain.

Ghost domain

The Ghost domain consists largely of igneous rocks and their metamorphic equivalents, and also contains a significant proportion of Yellowknife Supergroup supracrustal rocks. All the rocks of the domain are at

grain fin bien que, par endroits, la texture originelle plus grossière des roches, notamment de la pegmatite, demeure apparente. Dans la majeure partie du complexe, ces roches ont un aspect gneissique dont la foliation est définie par des variations mineures de composition et de texture. Là où elle est présente, la pegmatite se rencontre en couches parallèles de gros cristaux partiellement recristallisés et généralement désagrégés. La foliation gneissique devient plus prononcée en direction des zones de cisaillement limitrophes où les roches deviennent finement laminaires. Les roches les mieux conservées se trouvent dans la partie centrale du complexe, à l'extrémité est et au sud du «lac Hinscliffe», là où l'on peut voir plusieurs phases intrusives qui se recoupent entre elles de manière évidente, ainsi que des inclusions granitoïdes. Des zircons provenant d'une des plus vieilles phases ont un âge U-Pb d'environ 2655 Ma, alors que la titanite provenant de la même roche date d'environ 2610 Ma. Les zircons donnent un âge un peu plus récent que l'âge prédominant des roches volcaniques du Supergroupe de Yellowknife de la Province des Esclaves (de 2,73 à 2,66 Ga) et plus récent d'environ 20 Ma que l'âge de la métadacite du domaine de Wijinnedi.

Les roches méta-ignées renferment des inclusions d'amphibolite dont les proportions représentent de 0 % jusqu'à 60 % de la zone d'affleurement. Ces inclusions, qui ont habituellement une texture et une composition uniformes, ont été déformées à divers degrés mais sont généralement anguleuses, et elles sont recoupées par de multiples générations de filons leucogranitoïdes, qui ont eux aussi subi une déformation variée. Dans de rares cas, on trouve de longues et étroites zones d'inclusions distinctes par leur texture, ce qui laisse supposer que les inclusions mafiques pourraient représenter les restes d'une suite de dykes mafiques synplutoniques.

Une deuxième suite d'inclusions mafiques se rencontre souvent le long des marges du complexe. Ces inclusions se distinguent des inclusions décrites précédemment, car elles affichent une différenciation pétrographique marquée à une échelle de moins de 10 cm et leur composition moyenne est plus intermédiaire que celle des inclusions d'amphibolite mafique présentes dans la partie principale du complexe. Ces inclusions pourraient être d'origine supracrustale, ce qui pourrait mener, compte tenu de leur répartition marginale, à l'hypothèse provisoire que le domaine de Hinscliffe se trouve dans une structure antiforme.

Les roches granitoïdes sont recoupées par une suite mineure de petits massifs à de petites lentilles de pyroxénite qui sont le plus abondants à proximité du contact nord du domaine et qu'on trouve également dans les roches volcaniques au nord. Ces massifs et lentilles se composent maintenant principalement d'actinote. Le plus gros d'entre eux est situé dans la partie centrale du «lac Hinscliffe». On trouve aussi de rares filons de pegmatite tardive et des stocks et petits filons de granite qui seraient associés au granite avoisinant du domaine de Ghost, dont la lithologie est similaire.

Domaine de Ghost

Le domaine de Ghost est formé en grande partie de roches ignées et de leurs équivalents métamorphiques, ainsi que d'une quantité considérable de roches supracrustales du Supergroupe de Yellowknife. Toutes les roches du domaine ont été fortement high metamorphic grade or were emplaced under high metamorphic grade conditions. Most of the domain is at granulite grade. The domain has a pronounced northeasterly trend that contrasts with that of the Wijinnedi domain and, to a lesser extent, the Hinscliffe domain. This trend is the dominant structural pattern between Ghost Lake and Yellowknife, 160 km to the south.

The Ghost domain is dominated by a northeasterly trending and gently to moderately plunging antiformal structure, which contains at its core a large intrusive body of massive, coarsely megacrystic syenogranite. Zircons from this body have an age of about 2600 Ma, whereas monazite from the same rock has an age of about 2590 Ma. A similar body occurs north of the east end of Ghost Lake and may reflect the continuation of the antiformal structure to the north.

The large syenogranite pluton is thought to be flanked by major shear zones. The southeastern shear zone is not well exposed, but its existence is apparent from the structural discordance of the adjacent units. The shear zone itself is marked by a prominent positive aeromagnetic anomaly. The presence of the northwestern shear zone, located mostly under Ghost Lake, is largely hypothetical and is based on the presence of a similar, although less intense, magnetic anomaly and the topographic low expressed by the southwest part of the lake itself. This anomaly is symmetrical with the anomaly associated with the southeast shear zone about the antiformal axis and does not follow compositional layering around the antiformal structure.

The dominant rock types in the domain are orthopyroxene- and clinopyroxene-bearing tonalite and much less abundant granodiorite along with the closely related tonalitic gneiss. Uranium-lead dating of zircon suggests a crystallization age of between 2640 Ma and 2630 Ma; monazite from the same sample has an age of about 2595 Ma. The tonalite forms large, massive to moderately foliated bodies in which igneous textures are commonly well preserved in the least deformed rocks. The tonalitic gneiss consists largely of a more deformed and commonly layered equivalent of the tonalite intercalated with metasedimentary and amphibolitic layers at various scales and proportions. An orthopyroxene-free equivalent of the tonalite occurs northwest of the main part of Ghost Lake, although minor clinopyroxene and possibly relict orthopyroxene are found at its southwest margin. The tonalitic gneiss also occurs in an orthopyroxene-free form in the southwestern part of the area and is considered to be a retrograde expression of the granulite-grade gneiss.

The northwest limb of the antiform contains a more diverse assemblage of granitoid bodies. Most of the units as mapped are complex with, in addition to the main rock types of the particular unit, varied proportions of inclusions and intrusions of other granitoid phases as well as supracrustal rocks. Similarly, the supracrustal rocks in which the granitoid bodies are

métamorphisées ou mises en place dans des conditions de métamorphisme intense. La majeure partie du domaine se trouve dans le faciès des granulites. Le domaine a une direction nord-est prononcée, ce qui contraste avec la direction du domaine de Wijinnedi et, dans une moindre mesure, avec celle du domaine de Hinscliffe. Cette direction représente la configuration structurale dominante entre le lac Ghost et Yellowknife, qui est située à 160 km au sud.

Une structure antiforme à direction nord-est et à plongement faible à modéré domine dans le domaine de Ghost; son noyau est un gros corps intrusif de syénogranite massif et mégacristallin. Des zircons provenant du syénogranite remontent à environ 2600 Ma, alors que de la monzanite provenant de la même roche date d'environ 2590 Ma. Un massif similaire se trouve au nord de l'extrémité est du lac Ghost; il pourrait constituer le prolongement de la structure antiforme située au nord.

On suppose que le gros pluton de syénogranite est bordé par des zones de cisaillement majeures. La zone de cisaillement sud-est est mal exposée, mais sa présence est dévoilée par la discordance structurale des unités adjacentes. La zone de cisaillement est marquée par une anomalie aéromagnétique positive de forte intensité. L'existence de la zone de cisaillement nord-ouest, qui se trouve en grande partie sous le lac Ghost, est largement hypothétique et basée sur la présence d'une anomalie magnétique similaire de moins forte intensité et sur la présence de la dépression topographique que représente la partie sud-ouest du lac Ghost. Cette anomalie, de même que l'anomalie qui est associée à la zone de cisaillement sud-est, est symétrique par rapport à l'axe antiforme, mais elle ne suit pas la différenciation pétrographique autour de la structure antiforme.

De la tonalite à orthopyroxène et à clinopyroxène, de la granodiorite beaucoup moins abondante et du gneiss tonalitique étroitement apparenté sont les principaux types de roches trouvés dans le domaine. La datation U-Pb d'échantillons de zircon donne un âge de cristallisation qui se situe entre 2640 Ma et 2630 Ma; de la monazite provenant du même échantillon remonte à environ 2595 Ma. La tonalite forme de gros corps massifs à modérément foliés dans lesquels les textures ignées sont généralement bien conservées dans les roches les moins déformées. Le gneiss tonalitique est constitué en grande partie d'un équivalent plus déformé et généralement stratiforme de la tonalite qui est intercalé à diverses échelles et proportions dans des couches métasédimentaires et amphibolitiques. Un équivalent de la tonalite qui ne renferme pas d'orthopyroxène se rencontre au nord-ouest de la partie principale du lac Ghost, bien que l'on trouve des quantités mineures de clinopyroxène et possiblement des reliques d'orthopyroxène à la marge sud-ouest du lac. Du gneiss tonalitique sans orthopyroxène se rencontre aussi dans la partie sud-ouest de la région; on le considère comme étant une forme rétromorphosée du gneiss appartenant au faciès des granulites.

Le flanc nord-ouest de l'antiforme contient un assemblage plus varié de massifs granitoïdes. La plupart des unités cartographiées sont complexes et renferment, outre les principaux types de roche d'une unité en particulier, diverses proportions d'inclusions et d'intrusions d'autres phases granitoïdes, de même que des roches supracrustales. Pareillement, les roches supracrustales dans lesquelles sont mises en place les massifs granitoïdes renferment emplaced also commonly contain a variety of intrusive bodies too small to resolve at the present mapping scale. In addition to the previously mentioned tonalite, a long, narrow granitoid unit less than 1 km northwest of the main part of Ghost Lake varies gradationally from tonalite similar to that to the east to coarsely megacrystic granite similar to the main megacrystic syenogranite (discussed below). Several units of metamorphosed quartz diorite to tonalite occur near the west margin of the map area. Zircons from this unit at Ghost Lake have a U-Pb age of about 2605 Ma. A large unit of similar quartz diorite–tonalite occurs in the northern Dauphinee domain to the east. A large unit of granite to granodiorite with some pegmatite that is massive or significantly less foliated than most other granitoid units of the domain, occurs north of the main part of Ghost Lake. Small bodies of similar granite and pegmatite, too small to be mapped, occur southeast of the main body. Zircons from one such small granite and from the orthopyroxene-bearing granite at the southeast corner of the domain have similar ages between 2595 Ma and 2590 Ma. A large unit consisting mostly of various undivided granitoid phases occurs in the central part of the northwestern limb of the antiform. It includes many of the above-described granitoid units of the domain, as well as lesser amounts of other granitoid phases and supracrustal inclusions.

Amphibolite layers, ranging in width from a few centimetres to tens of metres, occur locally in varied proportions throughout the domain with only the largest units shown on Map 2023A. Most are considered to represent mafic intrusions, some of which may be related to the metagabbro intrusions of the Wijinnedi domain.

High-grade metasedimentary rocks occur throughout the domain, but are dominant along its northwestern margin and also in a distinct zone near the southeast corner of the map area. They are everywhere migmatitic to diatexitic and have been divided into four metamorphic zones: quartz-plagioclase-leucosome-bearing, K-feldspar-bearing, garnet-bearing, and orthopyroxene-bearing. Psammitic layers are rare, suggesting that the protolith was likely similar to the metasedimentary rocks of the Wijinnedi domain. Cordierite occurs locally in coarse crystals of almost gem quality.

Paragneiss is found in the northern part of the domain between its north border and Ghost Lake and as thin units associated with metasedimentary migmatite in the southeast corner of the area. The finely layered, mafic to intermediate migmatite consists of coarser grained leucosome and finer grained melanosome, both containing orthopyroxene- and clinopyroxene-bearing assemblages. These high-grade rocks are thought to represent a protolith similar to the intermediate volcanic rocks of the Wijinnedi domain.

généralement une gamme de massifs intrusifs trop petits pour être cartographiés à l'échelle de la carte. Outre la tonalite susmentionnée, on trouve une longue et étroite unité granitoïde à moins de 1 km au nord-ouest de la partie principale du lac Ghost; la composition de cette unité passe progressivement de la tonalite (semblable à celle qui se rencontre à l'est) à du granite mégacristallin (semblable au principal syénogranite mégacristallin, qui est décrit ci-après). Plusieurs unités de diorite quartzique métamorphisée allant à de la tonalite métamorphisée se rencontrent près de la marge ouest de la région cartographique. La datation U-Pb de zircons provenant de cette unité, au lac Ghost, donne un âge d'environ 2605 Ma. Une grosse unité de diorite quartzique-tonalite similaire se trouve à l'est, dans la partie nord du domaine de Dauphinee. Au nord de la partie principale du lac Ghost, on trouve une grosse unité de granite passant à de la granodiorite avec une certaine quantité de pegmatite, laquelle est massive ou considérablement moins foliée que la plupart des autres unités granitoïdes du domaine. Au sud-est du massif principal, on rencontre des massifs de granite et de pegmatite similaires qui sont trop petits pour être cartographiés. Des zircons provenant d'un de ces petits massifs de granite et du granite à orthopyroxène dans le coin sud-est du domaine ont des âges similaires qui se situent entre 2595 Ma et 2590 Ma. Une grosse unité constituée principalement de diverses phases granitoïdes non subdivisées se rencontre dans la partie centrale du flanc nord-ouest de l'antiforme. Elle contient nombre des unités granitoïdes du domaine qui sont décrites ci-dessus, ainsi que des quantités moindres d'autres phases granitoïdes et inclusions supracrustales.

Des proportions variées de couches d'amphibolite allant de quelques centimètres à des dizaines de mètres de largeur sont observées par endroits dans le domaine, mais seules les plus grosses unités figurent sur la carte 2023A. La plupart de ces couches sont considérées comme représentant des intrusions mafiques, et certaines d'entre elles pourraient être apparentées aux intrusions de métagabbro du domaine de Wijinnedi.

Des roches sédimentaires fortement métamorphisées se rencontrent partout dans le domaine, mais elles abondent particulièrement le long de la marge nord-ouest du domaine et dans une zone distincte près du coin sud-est de la région cartographique. Elles sont partout migmatitiques à diatexitiques, et elles ont été subdivisées en quatre zones métamorphiques, avec des roches à leucosome de quartz-plagioclase, à feldspath potassique, à grenat et à orthopyroxène. Les couches psammitiques sont rares, ce qui laisse supposer que le protolite était probablement similaire aux roches métasédimentaires du domaine de Wijinnedi. On trouve par endroits de la cordiérite en gros cristaux presque d'une qualité gemme.

On trouve du paragneiss dans le secteur nord du domaine, entre la limite nord de ce dernier et le lac Ghost, ainsi qu'en minces unités associées à de la migmatite métasédimentaire dans le coin sud-est de la région. La migmatite, mafique à intermédiaire et à lits fins, comporte un leucosome à grain plus grossier et un mélanosome à grain plus fin, qui contiennent tous les deux des assemblages à orthopyroxène et à clinopyroxène. Ces roches de métamorphisme intense représenteraient un protolite similaire aux roches volcaniques intermédiaires du domaine de Wijinnedi.

All the supracrustal rocks of the domain and most of the intrusive rocks have undergone high-grade metamorphism or were emplaced under relatively high-temperature and -pressure conditions. The eastern and southeastern two-thirds of the domain is at granulite grade, as indicated by the presence of orthopyroxene in rocks of appropriate composition and their characteristic yellow to greenish-grey to greenish-brown colour. Pressure and temperature estimates on a series of rocks from throughout the domain range from 6 kbar to 7 kbar and from 825°C to 900°C and show no regular regional variation. The orthopyroxene-zone boundary as shown on Map 2023A is regarded as a minimum estimate of the original extent of granulite-grade metamorphism, as relict altered pyroxene is locally recognized northwest of the zone boundary. Retrogressed zones also occur within the orthopyroxene zone. Most of the large, megacrystic syenogranite at the core of the domain antiform does not contain orthopyroxene. It has been suggested this is due to autoretrogression resulting from a progressive increase in the water content of the magma as it crystallized. Orthopyroxene has a local patchy distribution elsewhere in the domain as well, particularly in the gneiss along the antiformal axis northeast of the megacrystic granite. There the retrogression is thought to be related to fluids associated with the autoretrogression of the megacrystic syenogranite that presumably occurs below it.

Dauphinee domain

The Dauphinee domain, on the eastern edge of the map area, is separated from the Ghost domain by a throughgoing, low-grade, cataclastic shear zone. The dominant unit consists of massive to weakly foliated, undivided, leucocratic granite to granodiorite. The older, large, foliated tonalite to quartz diorite body in the northern part of the domain has been metamorphosed and intruded by several generations of variably deformed, leucogranitoid dykes and veins. Similar rocks of intermediate composition also occur in the Ghost domain. The thin unit of metasedimentary migmatite contained by the tonalite–quartz diorite unit expands to the north beyond the map area. In the south, another unit of metasedimentary migmatite occurs adjacent to the large southeastern metasedimentary zone of the Ghost domain, a relationship that would suggest minimal transcurrent transport on the bounding cataclastic shear zone if the two units are correlative. In contrast to the immediately adjacent rocks of the Ghost domain, no evidence has been found that would indicate that rocks of the Dauphinee domain have ever been metamorphosed to granulite grade.

Toutes les roches supracrustales du domaine et la plupart des roches intrusives ont subi un métamorphisme de forte intensité ou ont été mises en place à des températures et pressions relativement élevées. Les roches des deux tiers est et sud-est du domaine appartiennent au faciès des granulites, comme en témoigne la présence d'orthopyroxène dans les roches de composition appropriée et la couleur caractéristique de ces roches, qui va du jaune au gris verdâtre au brun verdâtre. Les estimations de la pression et de la température relatives à une série de roches provenant de partout dans le domaine vont de 6 kbar à 7 kbar et de 825 °C à 900 °C et elles n'affichent aucune variation régionale régulière. La limite de la zone à orthopyroxène telle que représentée sur la carte 2023A est considérée comme une estimation minimale de l'étendue initiale du métamorphisme dans le faciès des granulites, car du pyroxène altéré relique est présent par endroits au nord-ouest de la limite de la zone. On trouve aussi des zones rétromorphosées dans la zone à orthopyroxène. La plupart des gros massifs de syénogranite mégacristallin dans le noyau de l'antiforme du domaine ne contiennent pas d'orthopyroxène. Ce fait serait attribuable à une autorétromorphose résultant d'une augmentation progressive de la teneur en eau du magma pendant sa cristallisation. L'orthopyroxène est également réparti de façon inégale ailleurs dans le domaine, notamment dans le gneiss le long de l'axe antiforme au nord-est du granite mégacristallin. À cet endroit, la rétromorphose serait liée à des fluides associés à l'autorétromorphose du syénogranite mégacristallin qui se trouverait en-dessous.

Domaine de Dauphinee

Le domaine de Dauphinee, qui est situé sur la limite est de la région cartographique, est séparé du domaine de Ghost par une zone de cisaillement cataclastique faiblement métamorphisée. L'unité principale est formée de granite passant à de la granodiorite leucocrate, massif à légèrement folié et non subdivisé. Le gros massif plus ancien de tonalite foliée passant à de la diorite quartzique qui se trouve dans la partie nord du domaine a été métamorphisé et recoupé par plusieurs générations de dykes et de filons leucogranitoïdes plus ou moins déformés. Des roches similaires de composition intermédiaire se rencontrent également dans le domaine de Ghost. La mince unité de migmatite métasédimentaire que contient l'unité de tonalite-diorite quartzique s'élargit vers le nord, au-delà de la région cartographique. Dans le sud, une autre unité de migmatite métasédimentaire est contiguë à la vaste zone métasédimentaire sud-est du domaine de Ghost, ce qui laisse présumer, s'il y a corrélation entre les deux unités, qu'un coulissement minime est survenu le long de la zone de cisaillement cataclastique limitrophe. Contrairement aux roches adjacentes du domaine de Ghost, on n'a trouvé aucun indice montrant que les roches du domaine de Dauphinee ont déjà été métamorphisées dans le faciès des granulites.

Other geological features

Diabase dykes

Paleoproterozoic Indin diabase dykes are particularly abundant in the western three domains of the map area; they occur in north-northwesterly and northeasterly trending sets and have an average easterly dip of about 82°. In the very limited number of cases observed in which dykes from both sets intersect, the northeasterly trending dyke is the younger. In most cases, the original mafic mineral assemblage has been altered, with the northeasterly trending dykes generally showing less alteration. The highest grade alteration assemblage contains actinolitic amphibole, indicating a pervasive low-grade metamorphic event of postdating the Paleoproterozoic Indin dyke swarm.

At least one Mesoproterozoic Mackenzie diabase dyke, distinguished by its prominent aeromagnetic expression, occurs in the area.

Shear zones and faults

The three western domains are bounded by, and in most cases contain within them, major, through-going, ductile shear zones to mylonite zones. These major Archean structures are up to several hundred metres wide, and the deformed rocks within them are at a metamorphic grade similar to that of rocks outside the shear zone.

Also present throughout the area are numerous Paleoproterozoic brittle faults, the largest and most important of which is the cataclastic shear zone that marks the western margin of the Dauphinee domain. This structure is marked by a prominent, low aeromagnetic expression, is up to 800 m wide, and has no through-going fabric. The rocks within it are extensively fractured, chloritized, and hematized, with epidote-filled fractures and carbonate-cemented breccia present locally, all suggesting deformation under low-grade metamorphic conditions. In places, nearby minor shears and fractures are steep, suggesting a similar orientation for the main zone. The shear zone is asymmetric in that orthopyroxene is present in Ghost domain rocks within 50 m of the most intensely deformed part of the shear zone, whereas chlorite alteration associated with the shear zone extends over several hundred metres on the east side. Little evidence exists for a significant horizontal component of motion on the shear zone, although the contrast in metamorphic grade across the structure would indicate an important vertical component.

The main part of the map area is cut by a series of brittle faults considered to be related to the main cataclastic shear zone. In the south, they have a northerly trend that becomes more northwesterly in the northern part of the area. Like the main zone, the dominant component of movement appears to be vertical with west-side-up displacement in many, but not all

Autres entités géologiques

Dykes de diabase

Les dykes de diabase paléoprotérozoïques d'Indin sont particulièrement abondants dans les trois domaines occidentaux de la région cartographique; ils se présentent en essaims à direction nord-nord-ouest et nord-est et leur pendage moyen est d'environ 82° E. Dans les très rares cas observés où des dykes de ces deux essaims s'entrecroisent, c'est le dyke à direction nord-est qui est le plus jeune. Dans la plupart des cas, l'association originelle de minéraux mafiques a subi une altération, cette altération étant généralement moindre dans les dykes à direction nord-est. L'association de minéraux d'altération du plus fort métamorphisme contient de l'actinote, ce qui indique qu'un faible métamorphisme pénétratif est survenu après la formation de l'essaim de dykes paléoprotérozoïques d'Indin.

La région compte au moins un dyke de diabase mésoprotérozoïque de Mackenzie, celui-ci se distinguant par une expression aéromagnétique marquée.

Zones de cisaillement et failles

Les trois domaines occidentaux sont limités par d'importantes zones de cisaillement ductile passant à des zones de mylonite et, dans la plupart des cas, ils renferment ces zones. Ces structures archéennes majeures mesurent jusqu'à plusieurs centaines de mètres de largeur, et les roches déformées qu'elles contiennent sont métamorphisées à un degré similaire à celui subi par les roches situées à l'extérieur de la zone de cisaillement.

La région comporte également de nombreuses failles cassantes paléoprotérozoïques, dont la plus grosse et la plus importante est la zone de cisaillement cataclastique qui marque la marge ouest du domaine de Dauphinee. Cette structure, qui se caractérise par une expression aéromagnétique marquée de faible intensité, a jusqu'à 800 m de largeur et ne présente aucune fabrique pénétrative. Les roches qu'elle contient ont été profondément fracturées, chloritisées et hématisées, et elles présentent localement des fractures remplies d'épidote et des brèches à ciment carbonaté, ce qui laisse supposer une déformation dans des conditions de faible métamorphisme. Par endroits, les fractures et les cisaillements avoisinants sont abrupts, ce qui laisse penser que la zone principale aurait une orientation similaire. La zone de cisaillement est asymétrique, du fait de la présence d'orthopyroxène dans les roches du domaine de Ghost à moins de 50 m de la partie la plus fortement déformée de la zone de cisaillement, alors que la chloritisation associée à la zone de cisaillement s'étend sur plusieurs centaines de mètres du côté est. Il y a peu d'indices témoignant de l'existence d'une composante horizontale significative du mouvement dans la zone de cisaillement, bien que la différence d'intensité métamorphique dans l'ensemble de la structure indiquerait l'existence d'une importante composante verticale.

Une série de failles cassantes recoupe la partie principale de la région cartographique; ces failles seraient reliées à la principale zone de cisaillement cataclastique. Leur direction est vers le nord dans le sud de la région et davantage vers le nord-ouest dans le nord de la région. Comme dans la zone principale, le mouvement semble avoir été principalement vertical, avec soulèvement du compartiment de gauche dans nombre de cas, comme l'indiquent la cases, as shown by the map pattern and contrasts in metamorphism across two of the faults in the eastern Wijinnedi domain.

Diatremes

In the Hinscliffe domain, narrow linear diatremes a few metres in size occur at two localities. They are dominated by rounded clasts of either the local trondhjemite or pink, quartz-rich granite of a type not recognized in the area. The plagioclase in the clasts is weakly altered, whereas K-feldspar, where present, is completely altered to a fine-grained, feathery aggregate. The dark green matrix is an assemblage of fine-grained feldspar, chlorite, actinolite, and carbonate. Two more diatremes are associated with the volcanic rocks of the Wijinnedi domain. In both these examples, the clasts are what were originally euhedral to anhedral garnet crystals up to 10 cm that have been replaced by a low-grade mineral assemblage. Coarse-grained garnets are out of context with the low-grade volcanic rocks into which the diatremes were emplaced. The dark greenish-grey matrix is weakly foliated and consists mainly of feldspar, quartz, and chlorite. Because the matrix material is deformed, the diatremes are thought to be Archean.

Discussion

The Wijinnedi Lake area is at the north end of a region underlain by high-grade, granitoid gneiss; foliated and massive orthopyroxene-bearing, granitoid plutons; and migmatitic metasedimentary rocks. These high-grade rocks are prominently outlined by their aeromagnetic pattern.

The Wijinnedi Lake map area is situated at the northern margin of these high-grade rocks. Within the map area, the three western domains represent sequentially deeper crustal levels from north to south at the existing erosion level. The juxtaposition of these domains took place during the Archean, primarily along the domain-bounding shear zones while the rocks were under elevated metamorphic-grade conditions. This would indicate major uplift of parts of the region during the Archean. The 2675 Ma zircon age of the metadacite from the Wijinnedi domain and the 2655 Ma zircon age of trondhjemite from the Hinscliffe domain suggest that the trondhjemitic to granodioritic intrusive complex of the Hinscliffe domain may represent an originally subvolcanic, and perhaps somewhat more prolonged, expression of the surface magmatic event in the Wijinnedi domain. If so, the originally deeper level, Hinscliffe domain rocks have been translated laterally and vertically, presumably along the bounding and associated shear zones, with respect to the Wijinnedi domain to their present position adjacent to the intermediate-felsic volcanic centre.

configuration sur la carte et les différences de métamorphisme observées dans deux des failles dans la partie est du domaine de Wijinnedi.

Diatrèmes

Des diatrèmes linéaires étroits de quelques mètres se rencontrent à deux endroits dans le domaine de Hinscliffe. Ils contiennent en prédominance des clastes arrondis qui sont composés soit de trondhjémite locale, soit de granite quartzique rose d'un type inconnu dans la région. Le plagioclase présent dans les clastes est faiblement altéré, alors que le feldspath potassique, là où il existe, est entièrement altéré en agrégat plumeux à grain fin. La matrice vert foncé comporte une association de chlorite, d'actinote, de carbonates et de feldspath à grain fin. Deux autres diatrèmes sont associés aux roches volcaniques du domaine de Wijinnedi; les clastes qu'ils contiennent étaient à l'origine des cristaux de grenat automorphes à xénomorphes mesurant jusqu'à 10 cm qui ont été remplacés par une association de minéraux de faible métamorphisme. Des gros cristaux de grenat n'auraient pas pu cristalliser dans les roches volcaniques faiblement métamorphisées dans lesquelles les diatrèmes ont été mis en place. La matrice gris verdâtre foncé est légèrement foliée et se compose principalement de feldspath, de quartz et de chlorite. Étant donnée que la matrice est déformée, les diatrèmes pourraient remonter à l'Archéen.

Discussion

La région du lac Wijinnedi est située à l'extrémité nord d'une région composée de gneiss granitoïde fortement métamorphisé, de plutons granitoïdes foliés et massifs à orthopyroxène et de roches métasédimentaires migmatitiques. Ces roches fortement métamorphisées sont clairement délimitées de par leur expression aéromagnétique.

La région cartographique du lac Wijinnedi se trouve sur la marge nord de ces roches fortement métamorphisées. Dans la région cartographique, les trois domaines occidentaux représentent des niveaux crustaux dont la profondeur augmente du nord au sud, au niveau d'érosion actuel. La juxtaposition de ces domaines a eu lieu pendant l'Archéen, principalement le long des zones de cisaillement limitrophes, pendant que les roches se trouvaient dans des conditions propres à un métamorphisme de forte intensité, ce qui indiquerait un soulèvement majeur de certaines parties de la région au cours de l'Archéen. Compte tenu de l'âge de 2675 Ma fourni par des zircons provenant de la métadacite du domaine de Wijinnedi et de l'âge de 2655 Ma fourni par des zircons provenant de la trondhjémite du domaine de Hinscliffe, le complexe intrusif trondhjémitique à granodioritique du domaine de Hinscliffe pourrait représenter une expression au départ hypovolcanique et possiblement un peu plus prolongée de l'épisode magmatique superficiel qui est survenu dans le domaine de Wijinnedi. Si c'est le cas, les roches du domaine de Hinscliffe, qui proviennent d'un niveau plus profond, ont été déplacées latéralement et verticalement, vraisemblablement le long des zones de cisaillement limitrophes et associées, par rapport au domaine de Wijinnedi, jusqu'à leur position actuelle, qui est adjacente au centre volcanique intermédiaire-felsique.

On the basis of both geothermobarometric and textural evidence, it was suggested that rocks from the southeastern part of the Ghost domain may have undergone relatively quick decompression from granulite conditions. This is supported by the geochronological data in which four zircon U-Pb ages from granitoid intrusions of the Ghost domain range from about 2605 Ma to about 2590 Ma. They suggest that the ca. 2590 Ma monazite from the ca. 2600 Ma megacrystic syenogranite, whether interpreted as a down-temperature growth of secondary monazite or as a cooling age, may reflect this decompression event. The uplift and relative positioning of the domains may be, at least in part, a result of gravitational instability following formation and emplacement of the large, relatively low-density, megacrystic syenogranite bodies among the higher density, tonalite and tonalitic gneiss bodies at lower crustal levels and the even denser supracrustal rocks at higher crustal levels. This instability could be accommodated, in part, by the formation of, and extensional motion along, the domain-bounding and associated internal shear zones. These shear zones are symmetrical with respect to the large syenogranite bodies and could well be a consequence of the buoyant uplift of the lighter intrusions.

In addition to the adjustment of crustal levels during the Archean, major crustal uplift occurred during the Paleoproterozoic that involved much of the terrane within the map area as well as large areas to the north and south. Most of the uplift took place along the cataclastic shear zone between the Ghost and Dauphinee domains within the Wijinnedi Lake area and its southward and northward extension, as indicated by the abrupt metamorphic change across the shear zone. This is further supported by the presence on an over 25 mgal Bouguer gravity gradient across the approximately 35 km wide uplifted block, one of the greatest in the southwestern Slave Province. The steep gravity gradient may reflect the exposure of deeper crustal rocks on the west side of the main cataclastic shear zone. The Indin dykes within the map area dip on average 82°E. If this dip is due to rotation of the Archean block during uplift, it would suggest that the east side rose on the order of 3 km or 4 km higher than the west side. The total uplift of the block could have been significantly greater. Since the dykes are deformed in the cataclastic shear zone, the maximum age of uplift is considered to be that of the Indin dykes, about 2.1 Ga. This uplift may be a consequence of the indentation of much of the Slave Province into the western Churchill Province along the Bathurst and McDonald faults between 1.84 Ga and 1.74 Ga.

En se basant sur des indices géothermobarométriques et texturaux, on a émis l'hypothèse selon laquelle des roches de la partie sud-est du domaine de Ghost auraient subi une décompression relativement rapide à partir de conditions propres au faciès des granulites. Cette hypothèse est appuyée par des données géochronologiques qui situent entre 2605 Ma et 2590 Ma environ quatre âges U-Pb de zircons provenant d'intrusions granitoïdes dans le domaine de Ghost. Selon ces auteurs, qu'il soit interprété comme une croissance due à une chute de température de la monazite secondaire ou comme un âge de refroidissement, l'âge d'environ 2590 Ma de la monazite que contient le syénogranite mégacristallin d'environ 2600 Ma pourrait refléter cet événement de décompression. Le soulèvement et la position relative des domaines pourraient, du moins en partie, résulter d'une instabilité gravitationnelle postérieure à la formation et à la mise en place des gros massifs de syénogranite mégacristallin relativement peu denses parmi des massifs de tonalite et de gneiss tonalitique plus denses à des niveaux inférieurs de la croûte ainsi que dans les roches supracrustales encore plus denses à des niveaux supérieurs. Cette instabilité pourrait être compensée en partie par la formation des zones de cisaillement en bordure des domaines et des zones de cisaillement internes associées et par un mouvement d'extension le long de ces zones de cisaillement. Ces zones de cisaillement sont symétriques par rapport aux gros massifs de syénogranite et pourraient très bien être le produit du soulèvement des intrusions moins lourdes.

Outre l'ajustement des niveaux crustaux pendant l'Archéen, un soulèvement majeur de la croûte s'est produit au Paléoprotérozoïque, celui-ci touchant un grande partie du terrane dans la région cartographique ainsi que de vastes étendues au nord et au sud. La plus grande partie du soulèvement s'est concentrée le long de la zone de cisaillement cataclastique entre les domaines de Ghost et de Dauphinee, dans la région du lac Wijinnedi et dans le prolongement nord et sud de cette région, comme l'indique le brusque changement métamorphique qui s'observe à travers la zone de cisaillement. Celà est également corroboré par la présence d'un gradient de pesanteur de Bouguer de plus de 25 mgal dans le bloc soulevé d'environ 35 km de largeur (l'un des plus gros du sud-ouest de la Province des Esclaves). Ce fort gradient de pesanteur pourrait traduire la mise à découvert de roches de niveau crustal plus profond sur le côté ouest de la principale zone de cisaillement cataclastique. Les dykes d'Indin de la région cartographique ont un pendage moyen de 82° E. Si ce pendage est attribuable à une rotation du bloc archéen pendant le soulèvement, on peut présumer que le soulèvement du côté est s'est avéré de 3 à 4 km supérieur à celui du côté ouest. Le soulèvement total du bloc aurait pu être beaucoup plus important (Henderson et Chacko, 1995). Puisque les dykes de la zone de cisaillement cataclastique sont déformés, on considère que l'âge maximal du soulèvement équivaut à celui des dykes d'Indin, qui est d'environ 2,1 Ga (Henderson et Chacko, 1995). Ce soulèvement pourrait avoir été provoqué par le poinçonnement de la Province de Churchill occidentale par une grande partie de la Province des Esclaves, le long des failles de Bathurst et de McDonald, il y a entre 1,84 Ga et 1,74 Ga.

INTRODUCTION

The Wijinnedi Lake area contains within its boundaries the transition from well preserved, lower greenschist-grade, Archean Yellowknife Supergroup pelitic metasedimentary rocks to their granulite-grade, mid-crustal equivalents. This contrast in environments, as now exposed at the surface is the result of a combination of uplift of the mid-crustal rocks along a series of ductile shear zones during the Archean, together with the pop up of a major crustal block during the Paleoproterozoic.

The map area (Map 2023A, included with bulletin; Fig. 1, 2) which consists of the Ghost Lake (NTS 85-O/14) and the western 10 minutes of the Dauphinee Lake (NTS 85-O/15) 1:50 000-scale map areas, is located near the southwestern margin of the Slave Province, at the southern end of the Indin Lake supracrustal terrane (Fig. 3). Several of the geographical names that appear on Map 2023A are not included in the federal government's list of formal geographic names; however, these informal names are considered important local landmarks and are included on the map and referred to in the report, but are offset by single quotation marks.

Much of the surrounding region was originally mapped by Lord (1942) and Yardley (1949), and most of the area covered in this report was mapped by Wright (1950a, b, 1954). The first rocks reflecting granulite-grade metamorphic conditions to be recognized anywhere in the Slave Province were identified by Robertson and Folinsbee (1974) in the southeast corner of the map area. The few localities of economic interest in the area include gold showings associated with iron-formation in Yellowknife Supergroup metasedimentary rocks east of Wijinnedi Lake and gossans with largely pyrite and/or pyrrhotite mineralization near the volcanic-sedimentary contact in the metasedimentary rocks southwest of Wijinnedi Lake.

With the exception of its easternmost part, the area has a very high proportion of superb rock exposure. Much of the area has been exposed to periodic forest fires that, in many cases, have resulted in large, lichen-free rock outcrop surfaces. The area is within the high boreal–low subarctic forest at the boundary of the approximately 130 km wide transition from forest tundra to low arctic tundra (Timoney et al., 1992). In 1992 and 1993, the larger lakes in the area were accessible to float-equipped aircraft during the second week of June. Relief over much of the area is relatively low with local variations being less than a few tens of metres. Exceptions occur at the contact between Yellowknife Supergroup metavolcanic and metasedimentary rocks south of Wijinnedi Lake and southwest of 'Glazebrook Lake' where the local relief is as much as 90 m. Relief is up to 70 m along parts of the shoreline of Ghost Lake, particularly where Paleoproterozoic faults are present, and locally up to 100 m along a major Paleoproterozoic cataclastic shear zone at the east end of Ghost Lake. Ghost Lake is the main catchment basin for most of the area. It drains into Wijinnedi Lake via the Ghost River and then out of the area through 'West Wijinnedi Lake' into the Snare River and eventually to Great Slave Lake.

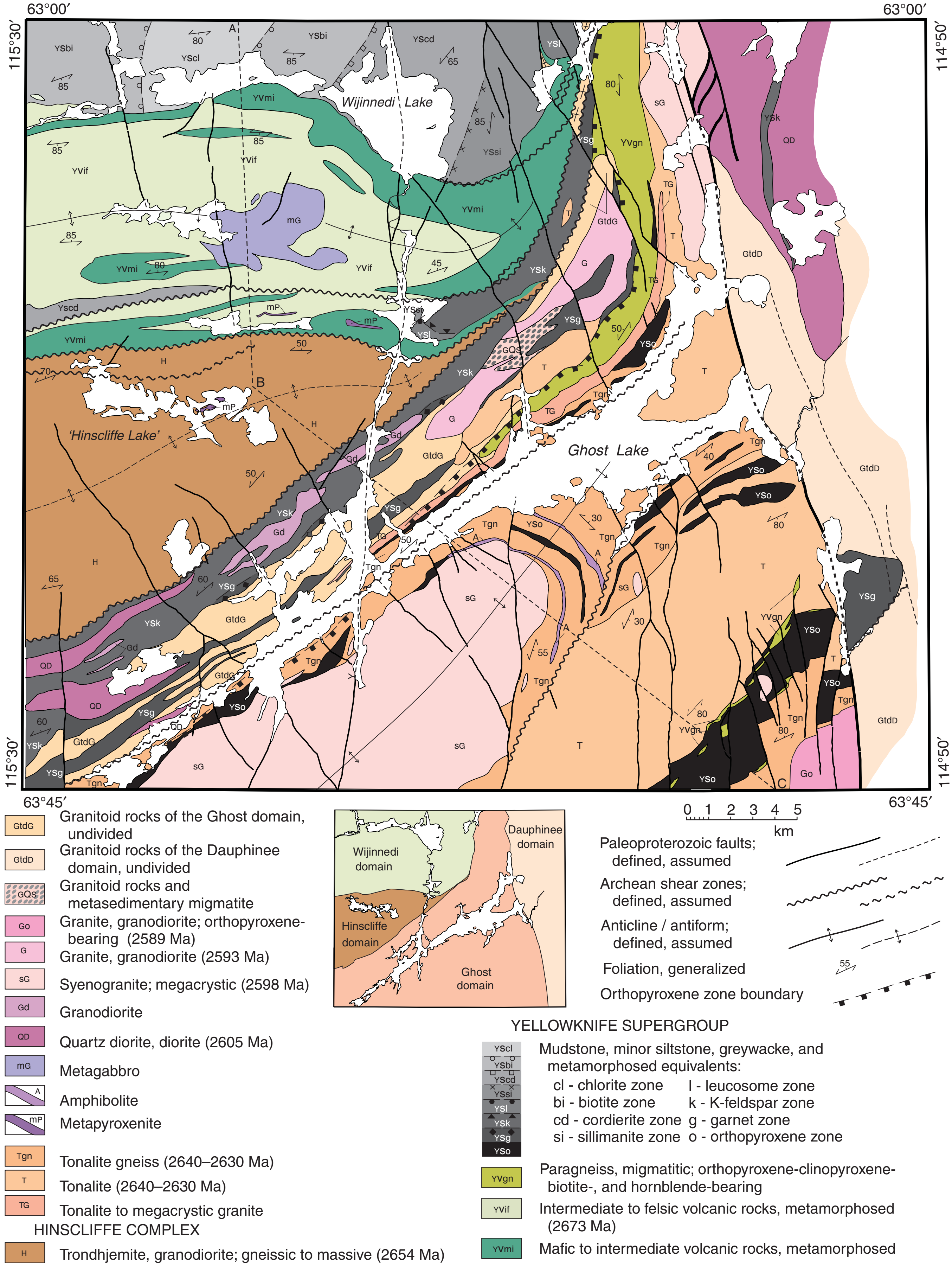
The Wijinnedi Lake area, which includes the area between latitude 63°45′N and 63°00′N and longitude 114°50′W and 115°30′W, was mapped in 1992 and 1993 (Henderson and Schaan, 1993; Henderson, 1994, 1998b). A reconnaissance visit was made to the high metamorphic grade area south of the map area in 1994 (Henderson and Chacko, 1995). The region west and south of the area was mapped by the Department of Indian and Northern Affairs (Jackson, 2003).

Acknowledgments

The fieldwork was greatly aided by the enthusiastic and capable assistance of Susan E. Schaan, senior assistant, and student assistants Kellie A. Emon, Robert E. Johnson, and Mark D. Smith in the 1992 field season, and Steven Bauke in 1993. The hospitality and good company of Robert Curtis, resident at Wijinnedi Lake in 1992, is appreciated. The use of his road network and vehicle greatly facilitated access to part of the area. The analysis and synthesis of field data for the preparation of the map and report on the area were greatly facilitated by the use of FieldLog, a data management program (Brodaric et al., 2001). The support and guidance of Boyan Brodaric and Deborah Lemkow in this regard are greatly appreciated. The geochronological data used in this report are from material collected during the course of fieldwork and were provided by M.E. Villeneuve and the staff of the Geological Survey of Canada Geochronology Lab (Villeneuve and Henderson, 1998; this report). The aeromagnetic bases for several figures in the report were provided and refined from the National Aeromagnetic Data Base by Warner Miles of the Geological Survey of Canada Geophysical Data Centre. This report has benefited from careful reviews by Sally J. Pehrsson and Simon Hanmer.

GENERAL GEOLOGY

The Wijinnedi Lake area in the southwestern Slave Province (Map 2023A; Fig. 1, 2) is situated at the abrupt transition between the relatively low-grade rocks of the Indin Lake supracrustal terrain to the north and west, the central part of which is at sub-biotite-greenschist metamorphic grade (Frith, 1993; Pehrsson and Kerswill, 1997a, b), and a much higher grade, largely granitoid-dominated terrane at granulite grade to the south (Fig. 4). The Archean supracrustal rocks of the Indin Lake terrane are considered part of the Yellowknife Supergroup, although they are physically separated from Yellowknife Supergroup rocks to the east by an extensive granitoid-rock-dominated terrane. As elsewhere in the Slave Province, the Yellowknife Supergroup rocks are dominated by metagreywacke-mudstone turbidite deposits. Mafic volcanic rocks are significantly more abundant than intermediate and felsic volcanic rocks, and limited geochronological data suggest an age bracket similar to that of most Yellowknife Supergroup volcanic rocks elsewhere in the province (Villeneuve and Henderson, 1998; Pehrsson and Villeneuve, 1999).


63°00′
63°00′
115°30′
114°50′
115°30′
114°50′
63°45′
63°45′
Wijinnedi Lake
'Hinscliffe Lake'
Ghost Lake
A
B
C
YSbi
YScl
YScd
YSsi
YSl
YSk
YSg
YSo
YVmi
YVif
YVgn
mG
mP
H
GQS
G
Gd
GtdG
GtdD
QD
sG
Go
T
TG
Tgn
Wijinnedi domain
Dauphinee domain
Hinscliffe domain
Ghost domain
0 1 2 3 4 5
km
GtdG Granitoid rocks of the Ghost domain, undivided
GtdD Granitoid rocks of the Dauphinee domain, undivided
GQS Granitoid rocks and metasedimentary migmatite
Go Granite, granodiorite; orthopyroxene-bearing (2589 Ma)
G Granite, granodiorite (2593 Ma)
sG Syenogranite; megacrystic (2598 Ma)
Gd Granodiorite
QD Quartz diorite, diorite (2605 Ma)
mG Metagabbro
A Amphibolite
mP Metapyroxenite
Tgn Tonalite gneiss (2640–2630 Ma)
T Tonalite (2640–2630 Ma)
TG Tonalite to megacrystic granite
HINSCLIFFE COMPLEX
H Trondhjemite, granodiorite; gneissic to massive (2654 Ma)
Paleoproterozoic faults; defined, assumed
Archean shear zones; defined, assumed
Anticline / antiform; defined, assumed
Foliation, generalized
55
Orthopyroxene zone boundary
YELLOWKNIFE SUPERGROUP
YScl YSbi YScd YSsi YSl YSk YSg YSo Mudstone, minor siltstone, greywacke, and metamorphosed equivalents:
cl - chlorite zone
bi - biotite zone
cd - cordierite zone
si - sillimanite zone
l - leucosome zone
k - K-feldspar zone
g - garnet zone
o - orthopyroxene zone
YVgn Paragneiss, migmatitic; orthopyroxene-clinopyroxene-biotite-, and hornblende-bearing
YVif Intermediate to felsic volcanic rocks, metamorphosed (2673 Ma)
YVmi Mafic to intermediate volcanic rocks, metamorphosed

The high-grade rocks within the map area are at the northern end of a largely structurally bound, triangular terrane 80 km long by 35 km wide, composed of dominantly Archean granitoid rocks that are in large part at granulite grade (Fig. 4, 5). At the time of mapping, the geology of this high-grade terrane was known largely from a series of reconnaissance landings (locations shown in Fig. 5) (Henderson and Chacko, 1995); a detailed mapping, geochronology, and geochemical study of part of the terrane has since been completed (Perks, 1997), and much of it has been systematically mapped by Indian and Northern Affairs Canada (Bennett and Dunning, 1998; Jackson, 1998, 2000, 2003; Bennett et al., 2000). The terrane consists mainly of weakly to moderately foliated, mainly granodioritic to tonalitic gneiss with layering defined by subtle compositional and textural variations; this gneiss commonly contains pegmatitic granite pods, layers, and veins, and amphibolitic layers and inclusions. Migmatitic cordierite-, garnet-, and K-feldspar-bearing metasedimentary rocks ranging from layered metatexite to homogeneous diatexite are less abundant and are believed to be derived from Yellowknife Supergroup metasedimentary rocks similar to those seen to the north. In addition, Jackson (2000) reported the occurrence of volcanic protoliths and minor carbonate. Massive to weakly foliated granitoid rocks ranging from coarsely megacrystic granite to medium-grained, equigranular, more compositionally heterogeneous intrusions are also present. Megascopic orthopyroxene in granitoid or granitoid gneiss phases, indicating granulite metamorphic conditions is most common in the central and eastern parts of the terrain (Fig. 5). It is not clear whether the paucity of orthopyroxene in the west indicates that these rocks never reached granulite conditions or were retrogressed. The region east of the high-grade terrane is underlain mainly by undivided granitoid rocks (Yardley, 1949); an extensive area of granulite-grade rocks occurs northeast of the area (Pehrsson et al., 2000).

The high-grade terrane is characterized by high-relief magnetic topography that contrasts strongly with the low-amplitude–long-wavelength pattern to the north and northwest (Fig. 5). Within the Wijinnedi Lake area, a prominent magnetic low corresponds to a major zone of migmatitic metasedimentary rocks, as may also be the case elsewhere in the domain to the south. The eastern side of the high-grade terrane is marked by a prominent magnetic low that corresponds, within the Wijinnedi Lake area, to a major cataclastic shear zone. This structure is the eastern margin of a Paleoproterozoic uplift that can be traced to the north and northwest as far as Indin Lake (Henderson and Chacko, 1995). There, in the lower grade rocks, the narrow greenschist-grade zone southwest of the fault zone suggests greater uplift than in the terrane to the northeast, where the adjacent greenschist-grade zone is much wider (Fig. 6; Henderson and Chacko, 1995). The southwestern margin of the terrane as defined by the 60 400 nT magnetic contour corresponds to a relatively straight, north-northwesterly trending lineament considered to possibly represent a fault with a largely vertical sense of displacement (Henderson and Chacko, 1995). Although no abrupt lithological change occurs across this lineament, there may be a metamorphic jump (Jackson, 1998, 2000). One of the steepest and longest gravity gradients in the southeastern Slave Province occurs across this high-grade block (Fig. 5).

***Figure 1** (previous page). General geology of the Wijinnedi Lake map area. The map area can be divided into four domains bounded by high metamorphic grade shear zones as outlined in the inset figure. The Wijinnedi domain is dominated by relatively low metamorphic grade Yellowknife Supergroup supracrustal rocks; the Hinscliffe domain, by a metamorphosed trondhjemitic to granodioritic igneous complex; and the Ghost domain, by foliated granitoid rocks, granitoid intrusions, and high metamorphic grade Yellowknife supracrustal rocks with most of the domain at granulite grade. The Dauphinee domain consists largely of massive granitoid rocks with minor metasedimentary migmatite. It is separated from the Ghost domain by a cataclastic shear zone. Line ABC is the position of the cross-section shown in Figure 2.*

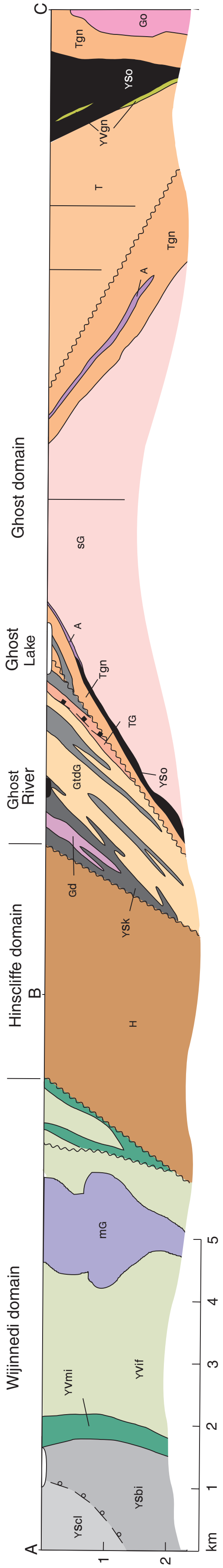


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***Figure 2.** Geological cross-section across the Wijinnedi Lake map area along line ABC indicated in Figure 1.* See *legend for Figure 1.*

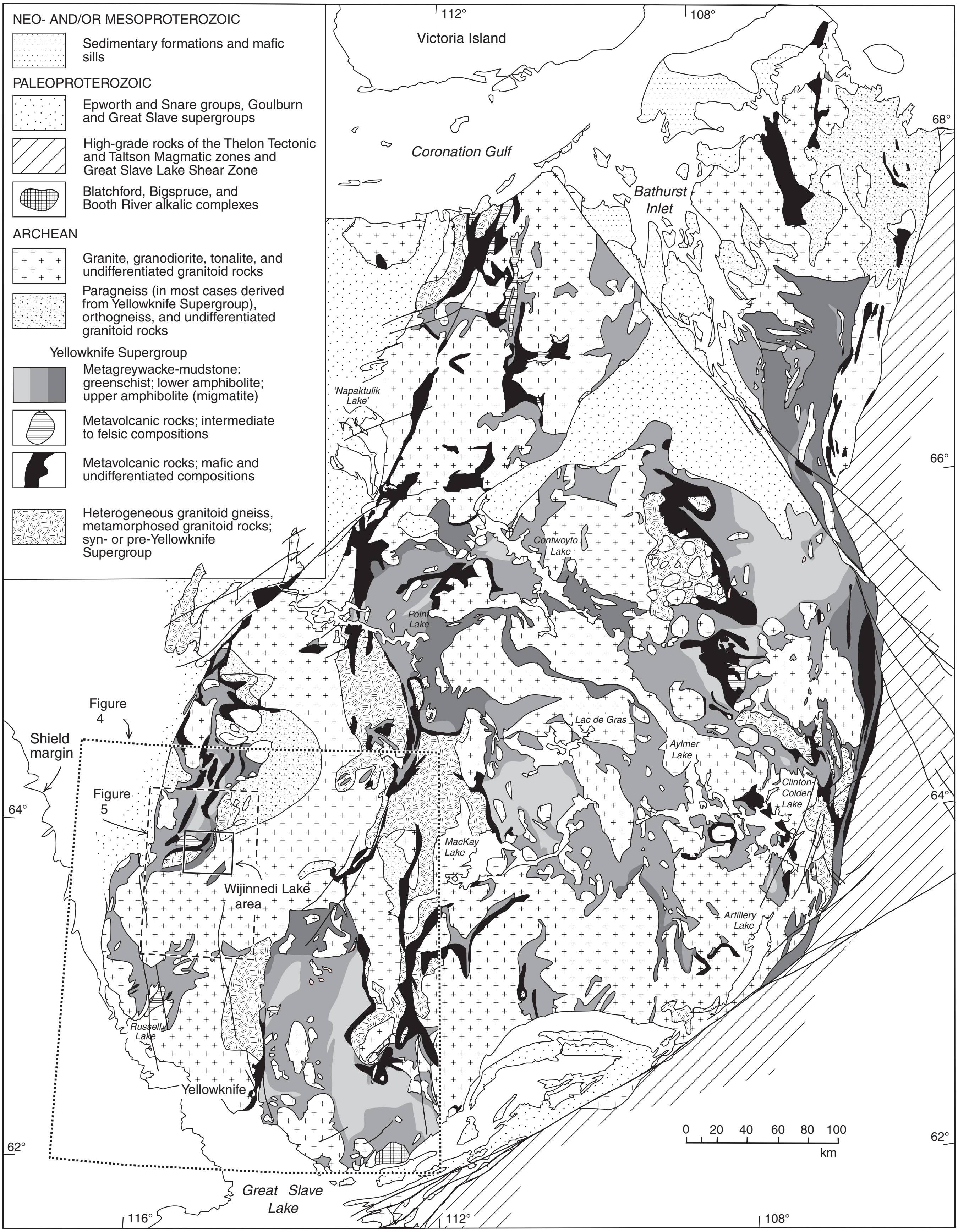


Figure 3. General geology of the Slave structural province. The location of the Wijinnedi Lake area in the southwest part of the province is indicated. Other areas outlined show the locations of figures in the report. The geological map is primarily modified from *McGlynn (1977) and Hoffman and Hall (1993), but also other published sources.*

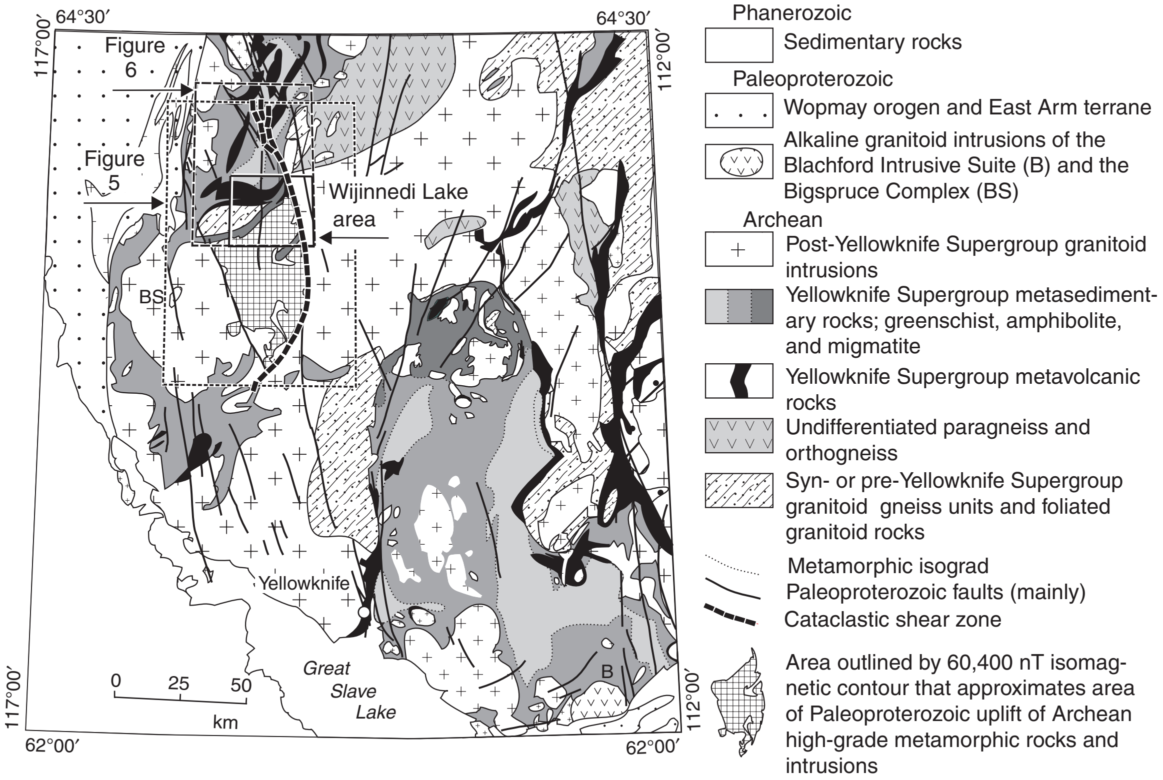


Figure 4. *Geology of the southwestern Slave Province showing the location of the Paleoproterozoic uplift as expressed by the aeromagnetic anomaly pattern. The location of the Wijinnedi Lake map area and the areas covered by Figures 5 and 6 are indicated. Geology* modified from *McGlynn (1977) and Hoffman and Hall (1993).*

Wijinnedi Lake map area

The geology of the Wijinnedi Lake map area can be thought of in terms of four lithotectonic domains representing distinct geological environments (Fig. 1; Map 2023A). Three of these domains are dominated by easterly to northeasterly trending anticlinal or antiformal structures and are separated by major Archean ductile shear zones formed at elevated metamorphic grade. In contrast, the fourth domain on the east side of the map area is separated from the others by a northerly trending, Paleoproterozoic, low-grade, cataclastic shear zone. Both within and north and south of the area, this cataclastic shear zone appears to be the easterly limit of the Paleoproterozoic component of the uplift of high-grade rocks discussed above (Fig. 6).

The Wijinnedi domain, in the northwest quadrant of the area, is dominated by moderately to well preserved metavolcanic and metasedimentary rocks of the Yellowknife Supergroup whose metamorphic grade varies from lower greenschist to middle amphibolite facies, reaching upper amphibolite facies in the easternmost part of the domain. The domain is dominated by a coarsely volcaniclastic, intermediate to felsic volcanic centre that is partially mantled by mafic, locally pillowed, volcanic rocks that in turn are stratigraphically overlain by a pelite-dominated, turbidite sequence. The easterly trend of the volcanic centre contrasts strongly with the more north-northeasterly trends typical of the Indin Lake supracrustal terrane as a whole (Fig. 6).

The Hinscliffe domain occurs south of the Wijinnedi domain. It consists of a moderately to strongly deformed granitoid complex. No Yellowknife Supergroup supracrustal rocks have been recognized within it. Much of the complex is strongly deformed, particularly toward its margins, resulting in a gneissic aspect, with layering defined by small compositional and textural variations. The complex is better preserved in its centre where intrusive relationships between various phases are evident, suggesting that the complex was never a simple foliated intrusion, but more likely formed through the multiple injection of numerous granitoid bodies. Locally present are deformed angular blocks of amphibolite of uniform composition that are invariably intruded by several generations of thin granitoid veins and dykes.

The Ghost domain is the largest domain in the area and occurs east and southeast of the Wijinnedi and Hinscliffe domains. Its boundary with the Hinscliffe domain is clear because of the strong lithological contrast between the two domains. The boundary with the lithologically more similar Wijinnedi domain to the northeast is less obvious. For purposes of definition, it is taken as the northeasterly to ultimately northerly continuation of the amalgamated Archean shear zones that bound the Hinscliffe domain to the southeast. This shear-zone complex is offset to the northwest by a Paleoproterozoic fault before extending into Daran Lake. The domain contains the high-grade migmatitic equivalents of both the lower grade metasedimentary and metavolcanic

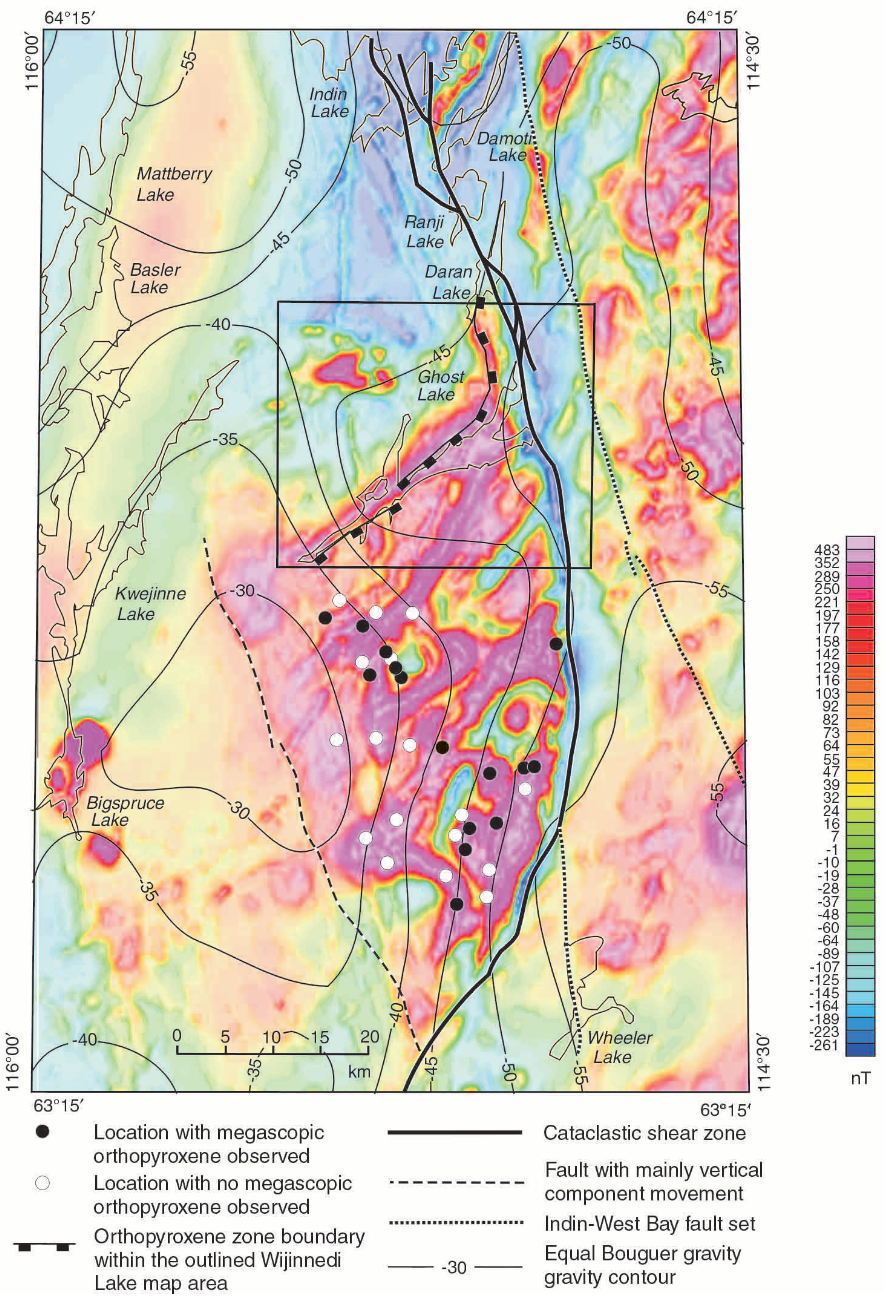


Figure 5.

Residual total-field magnetic map of the Ghost Lake region with the location of the Wijinnedi Lake area outlined. The roughly triangular, slightly elevated, high-relief magnetic anomaly that outlines the high-grade part of the Paleoproterozoic fault-bounded uplift of Archean rocks occurs at the eastern margin of a region characterized by a low-amplitude, long-wavelength magnetic pattern. The shaded relief map is presented with directly overhead illumination that results in the flatter tops and bottoms of anomalies having less saturated colours, whereas the anomaly margins with steeper gradients have more highly saturated colours. Superimposed on the magnetic map are contours of equal Bouguer gravity anomalies (Department of Energy, Mines and Resources, 1969). One of the longest and steepest gravity gradients in the southeastern Slave Province occurs across the triangular uplift terrain. Magnetic map provided by the Geological Survey of Canada Geophysical Data Centre from the National Aeromagnetic Data Base. Modified from *Henderson and Chacko (1995).*

rocks of the Wijinnedi domain. It is dominated, however, by a series of lenticular sheets of granitoid rocks ranging in composition from granite to quartz diorite. Most are fairly homogeneous within the individual outlined map units, although some are highly heterogeneous to a degree impossible to resolve at the present scale of mapping. Except for a zone along the northwestern margin of the domain, most Archean rocks were metamorphosed at, or were emplaced under, granulite-grade metamorphic conditions.

The Dauphinee domain occurs on the eastern side of the map area and is dominated by granitoid rocks varying from foliated quartz diorite to massive granite. Migmatitic metasedimentary rocks are also present. One large body of metasedimentary migmatite occurs on strike from a major belt of metasedimentary migmatite in the Ghost domain, which, if not coincidental, might suggest little, if any, horizontal transport along the cataclastic shear zone separating the two domains.

Paleoproterozoic, northeasterly and northwesterly trending, Indin diabase dykes are abundant in all four domains. One strongly magnetic, north-northwesterly trending, Mesoproterozoic dyke at the east end of ‘Hinscliffe Lake’ is considered to be part of the 1267 Ma (LeCheminant and Heaman, 1989) Mackenzie diabase dyke swarm (Fig. 7).

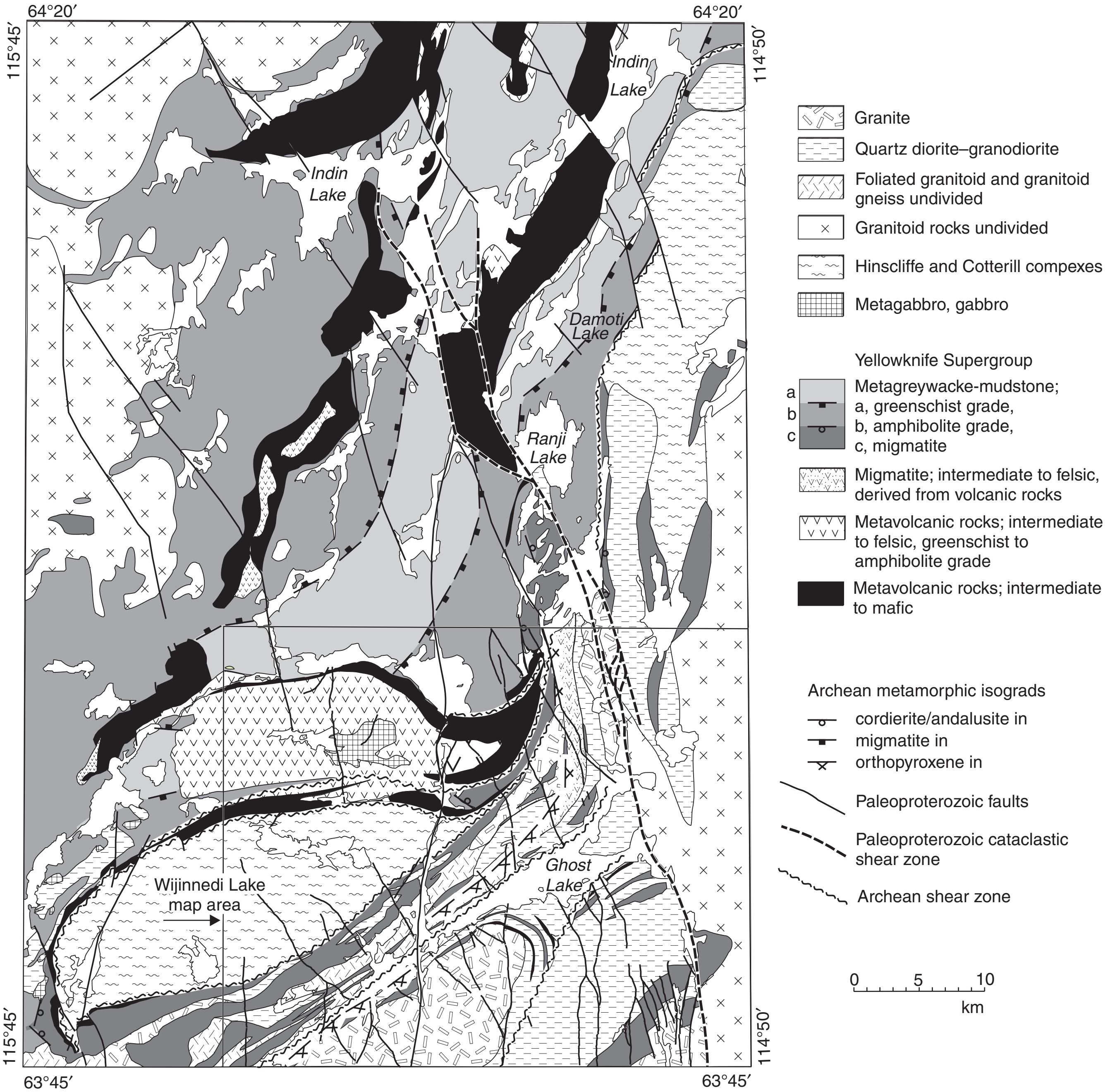


Figure 6. *General geology of the Indin Lake–Ghost Lake region. The Wijinnedi Lake map area is outlined. The pronounced northeasterly trends of the high-grade, deeper level rocks of the Wijinnedi Lake area are also present to some extent in the generally lower grade rocks of the Indin Lake region to the north. The easterly trend of the supracrustal rocks within the Wijinnedi Lake map area represents a major departure from this trend. The low-grade Paleoproterozoic cataclastic shear zone, along which the granulite-grade rocks southeast of Ghost Lake have been uplifted against the relatively lower grade rocks to the east, appears to continue north through to Indin Lake where it appears to splay out. There, a similar sense of west-side-up is indicated by the contrast in width of the greenschist zone in the metasedimentary rocks on either side of the shear zone; the greenschist zone is significantly narrower in the uplifted western side.* Modified from *Lord (1942), Frith (1993), Pehrsson and Kerswill (1997a, b), and Jackson (2003).*

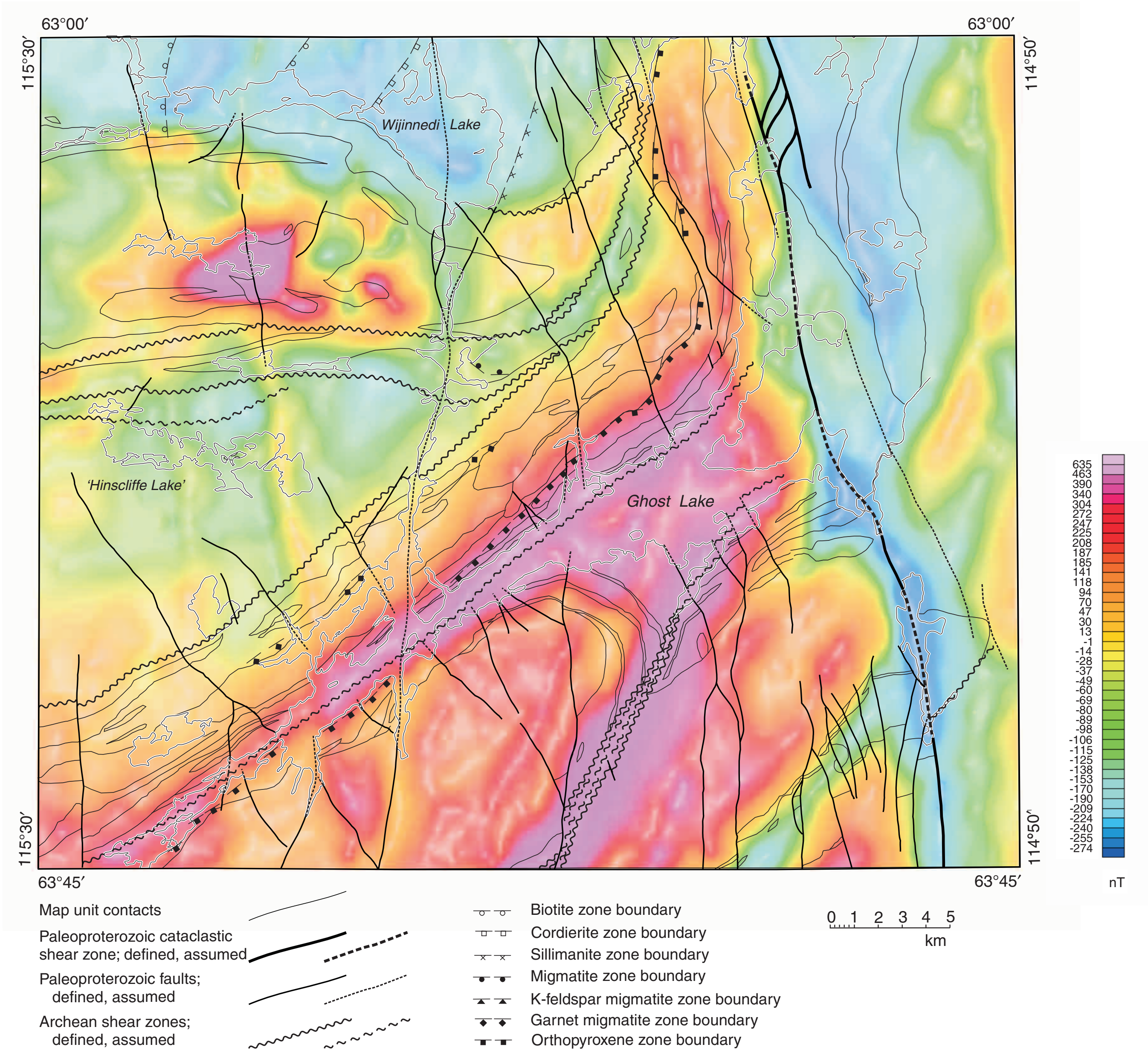


Figure 7. *Residual total-field magnetic map of the Wijinnedi Lake area with superimposed geological contacts, faults, shear zones, and metamorphic isograds. Magnetic map provided by the Geological Survey of Canada Geophysical Data Centre from the National Aeromagnetic Data Base.*

The domain-bounding shear zones are Archean as their metamorphic grade is similar to that of the rocks they displace. Other shear zones of this vintage are common, particularly within the Ghost domain. Collectively, they are largely responsible for the juxtaposition, during the Archean, of the various crustal levels represented by the three western domains to their present positions, and so are ultimately responsible for the short lateral transition from greenschist- to granulite-grade rocks within the map area. Northerly to northwesterly trending, Paleoproterozoic faults occur throughout the map area and are thought to be related at least in part to the cataclastic shear zone that separates the Dauphinee and Ghost domains. As mentioned previously, the cataclastic shear zone marks the eastern limit of uplift of the western three domains during the Paleoproterozoic.

YELLOWKNIFE SUPERGROUP

The Yellowknife Supergroup, as currently defined and used, includes all Archean rocks of supracrustal origin within the Slave Province (Henderson, 1970; Padgham and Fyson, 1992). It can be divided into three lithologically and chronologically distinct components reflecting differing tectonic

Table 1. Table of formations.

Eon	Era	Period/ Epoch	Group/ Suite	Formation/ Lithodeme	Map symbol	Lithology and radiometric age where known
	Cenozoic	Quaternary				Unconsolidated till, gravel, sand and silt
Unconformity						
Proterozoic	Mesoproterozoic			Mackenzie	PM	Diabase, gabbro dykes, northerly tend
	Paleoproterozoic			Indin	PI	Diabase, gabbro dykes, northeasterly and northwesterly trend
Archean						Diatremes
	Intrusive contact with Hinscliffe complex and Yellowknife Supergroup					
					AG, AG.o	Undivided granitoid rocks of the Ghost domain
					ADg	Undivided granitoid rocks of the Dauphinee domain
					AYs-g-t	Mixed granitoid rocks and Yellowknife Supergroup metasedimentary migmatite
					Ag, Ag.o	Granite, granodiorite; orthopyroxene granite; U-Pb (zircon) ages 2593+6/-4 Ma , 2589+1/-2 Ma
					Agm.o	Megacrystic syenogranite; U-Pb (zircon) age 2598 ± 2 Ma, U-Pb (monazite) age 2589+1/-2 Ma
					Agd	Granodiorite, tonalite
					Aqd	Quartz diorite, diorite; U-Pb (zircon) age 2605 ± 3 Ma
					Amg	Metagabbro, leucogabbro
					Aam	Amphibolite
					Aum	Metapyroxenite
					At, At.o	Tonalite, granodiorite
					Atgn, Atgn.o	Tonalite gneiss; U-Pb (zircon) age 2630–2640 Ma
					At-gm, At-gm.o	Tonalite, granodiorite to megacrystic granite
			Intrusive contact with Yellowknife Supergroup; tectonic contact with Hinscliffe complex			
			Hinscliffe complex		AH	Trondhjemite, minor granodiorite and tonalite, rare pegmatite, and foliated to gneissic equivalents, inclusions of metadiorite to metagabbro; U-Pb (zircon) age 2654+4 Ma, U-Pb (titanite) age 2610 ± 4 Ma
			Tectonic contact with Yellowknife Supergroup			
			Yellowknife Supergroup			
					AYs.cl, AYs.bi, AYs.cd AYs.si, AYs.m, AYs.mk, AYs.mg, AYs.mo	Metamudstone, siltstone and greywacke, minor silicate iron-formation, metamorphosed from lower greenschist to granulite facies
					AYvmi	Metabasalt, meta-andesite
					AYvif, AYvif.o	Metadacite, minor meta-andesite and metarhyolite volcanic breccia, volcaniclastic sandstones; U-Pb (zircon) age 2673.3 ± 1.4 Ma

environments. The growing catalogue of geochronological data on these rocks suggests that these components together may represent a period of time longer than the Phanerozoic, indicating this far too all-encompassing terminology has outlived its usefulness.

By far the largest component of the Yellowknife Supergroup consists of extensive outcrop areas of metagreywacke-mudstone turbidite. They are locally bounded by or, less commonly, contain within them much smaller areas of mafic or mafic to felsic volcanic sequences (Fig. 3). Together, they make up over 99% of the outcrop area of Yellowknife Supergroup supracrustal rocks. In most cases, the volcanic rocks range from about 2.72 Ga for mafic volcanic rocks near Yellowknife (Isachsen and Bowring, 1997), to about 2.66 Ga for felsic volcanic centres and the thin, felsic tuff units that occur within the turbidite sequence in the Yellowknife region (Mortensen et al., 1992; Bleeker and Villeneuve, 1995). A few exceptions are known, however, including a unit of felsic volcanic rocks in the northernmost Slave Province that has been dated at about 2616 Ma (Henderson et al., 2000), as well as a 2647 ± 2 Ma rhyolite breccia within a thick-bedded, greywacke-dominant turbidite unit 30 km north of the area (Pehrsson and Villeneuve, 1999). Pehrsson and Villeneuve (1999) also reported zircon ages from another turbidite unit from the same region; the zircons are interpreted as detrital grains that provide ages for source-terrane components as young as about 2630 Ma. Isachsen and Bowring (1994) reported a 2612 Ma age for a tuff within turbidite about 50 km south of the area at Wheeler Lake (Fig. 5).

In the western half of the Slave Province, an older and a younger suite of supracrustal rocks occur locally below or above the main sequence. The older sequence consists mainly of quartz-rich sandstone with minor iron-formation and mafic to felsic volcanic rocks (Bleeker et al., 1999). North of Yellowknife, the felsic volcaniclastic rocks have a minimum age of about 2.84 Ga (Isachsen and Bowring, 1997). Unconformably overlying the mafic to felsic volcanic rocks of the main sequence are local occurrences of conglomerate and shallow-water sandstone (Henderson, 1975b, 1998a; Hrabi et al., 1995; Corcoran et al., 1998), which have been dated at about 2.60 Ga (Isachsen, 1992; Isachsen et al., 1993).

In the Wijinnedi Lake area (Map 2023A), only supracrustal rocks of the main component of the Yellowknife Supergroup are known. They are everywhere metamorphosed and range from sub-biotite greenschist to granulite grade. In the interests of simplicity, therefore, the prefix 'meta' is omitted in the following discussion of these metamorphosed rocks.

Intermediate to felsic metavolcanic rocks

Low- and medium-grade metavolcanic rocks (unit ĀYvif)

A major unit (ĀYvif) of dominantly intermediate to felsic volcanic rocks of the Yellowknife Supergroup occurs in the central and southern parts of the Wijinnedi domain. It forms the core of an easterly trending, ellipsoidal, dome-like structure that extends about 4 km to the west beyond the area (Fig. 6). Stratigraphically, these rocks are the oldest in the area. Their trend is at a high angle to the more northerly to northeasterly trends typical of the Indin Lake region supracrustal rocks as seen to the north (Fig. 6).

Two other units of what were originally intermediate to felsic volcanic rocks occur to the east and southeast in the much higher grade Ghost domain. These rocks are everywhere migmatitic and are at granulite grade.

Structural setting

The intermediate to felsic volcanic rocks occur in a doubly plunging, easterly trending anticline. Facing indicators within the sequence are rare and, in many cases, of questionable validity, but the mafic to intermediate volcanic rocks that surround these rocks are locally pillowed, suggesting that the more mafic volcanic rocks face away from the felsic to intermediate volcanic unit. Similarly, the lower grade sedimentary rocks close to the contact with the mafic volcanic rocks tend to face away from them. The volcanic rocks are moderately to strongly foliated, and a moderate to steep stretching lineation is commonly evident. At the outcrop scale, primary compositional or textural layering is commonly parallel to the easterly trending principal foliation. On airphotos, however, trend lines possibly representing volcanic layering are evident locally and suggest the presence of smaller order folds within the anticline. Foliations in the southern limb of the structure dip moderately to steeply north. They are roughly concordant with the foliations in the northern Hinscliffe domain to the south and steepen northward. Dips in the northern limb are steeply north or south.

The southern part of the structure is transected by an Archean ductile shear zone, a smaller version of the shear zones that separate three of the four domains within the map area. In general, there is commonly an abrupt lithological change and an apparent increase in metamorphic grade across the shear zone.

Metamorphic grade increases from lower greenschist grade immediately south of the west end of 'West Wijinnedi Lake'. It increases southward and eastward such that most volcanic rocks are at amphibolite grade. Zones of retrogression occur locally within the higher grade rocks and are presumably related to late movement along shear zones.

Contacts

No evidence has been found to indicate the type of material on which the intermediate to felsic volcanic rocks were deposited, nor can any useful estimate be made of their original stratigraphic thickness. In most places, these rocks are conformably overlain by more mafic volcanic rocks. The contact zone, where seen, is in most places gradational with varied proportions of both mafic and dacitic units over several tens of metres. In the southwestern part of the volcanic complex, the intermediate volcanic rocks are assumed to be conformably overlain by amphibolite-grade metasedimentary rocks. The contact has not been observed, in part because of the high relief between the volcanic and sedimentary rocks.

The volcanic rocks are intruded by a metagabbro–anorthositic gabbro plutonic complex. A swarm of related metagabbro sills occurs throughout the volcanic dome; the sills are present in varied proportions, but are particularly abundant south and east of the intrusion where, locally, they exceed the volume of volcanic rocks present. Volumetrically much less important intrusive phases include minor, older Archean mafic dykes; the much more abundant Paleoproterozoic Indin dykes; coarse-grained, garnet-bearing diatremes; and a fine-grained, intermediate to felsic intrusive phase that may well be related to the volcanic complex. These felsic intrusive rocks also occur locally in the overlying mafic volcanic rocks, particularly in the south.

Lithology

In decreasing order of abundance, the intermediate to felsic volcanic unit consists of pale yellow-grey- to buff-weathering, light to medium grey, volcanic breccia; finer grained, volcaniclastic rocks; and extrusive flows or shallow intrusive rocks. Most of the complex is dacitic to rhyolitic in composition, but more mafic units occur locally on both the north and south limbs of the structure, the most prominent being a mafic dacitic to andesitic sequence south of ‘Glazebrook Lake’. They are described in the section entitled ‘Mafic to intermediate metavolcanic rocks (unit ĀYvmi)’.

Dacitic volcanic breccia is by far the most common rock type and is most abundant in the central, northern, and eastern parts of the complex (Fig. 8, 9). The clasts vary from a few centimetres to about 25 cm; deposits with clasts in the range of 4–6 cm are the most common. In places, bedding-unit contacts can be defined on the basis of abrupt changes in clast size. Clasts are everywhere tectonically elongated with a typical ratio of 4:1 or 5:1 on horizontal surfaces and even more in the third dimension (Fig. 10). In general, the degree of elongation increases southward to the extent that breccia units become difficult to recognize as such. Clasts vary in composition from rhyolitic to andesitic and are most commonly fairly uniform within a given unit, although heterogeneous units are not uncommon. In one very unusual case, two medium-grained, granitoid clasts occur with the normal dacite clasts (Fig. 11). In some cases, the clasts are sparsely vesicular, with quartz fillings up to 1 cm most common. Carbonate vesicle fillings are also present in the lower grade rocks. Matrix composition varies from similar to that of the clasts it contains to more mafic, commonly quite biotitic. This is commonly, but not exclusively, the case for the volcanic breccia units in the vicinity of the more mafic zone south of ‘Glazebrook Lake’.

The volcanic rocks that form the breccia units are typically very fine grained, with sparse phenocrysts of plagioclase or quartz (Fig. 12). At the lowest metamorphic grade, the volcanic rocks tend to vary in grain size with diffusely bound, patchy zones of coarser, more quartzofeldspathic material. With increasing metamorphic grade, grain size increases and becomes more homogeneous. The compositional variation of these rocks is expressed petrographically with the varied proportion of mafic minerals, chlorite at lowest grade through actinolitic amphibole to hornblende that can form poikilitic porphyroblasts in the higher grade rocks.

In addition to breccia, extensive, fine-grained, generally featureless volcaniclastic units occur throughout the complex; they are particularly abundant in the west and southwest. They are similar in colour and presumably in compositional range to the volcanic breccia units. They form layered units that are defined by both compositional and textural variations and range in thickness from a few centimetres to several metres. Where they are best preserved, a faint clastic texture (Fig. 13) is suggested with clasts up to 5 mm.

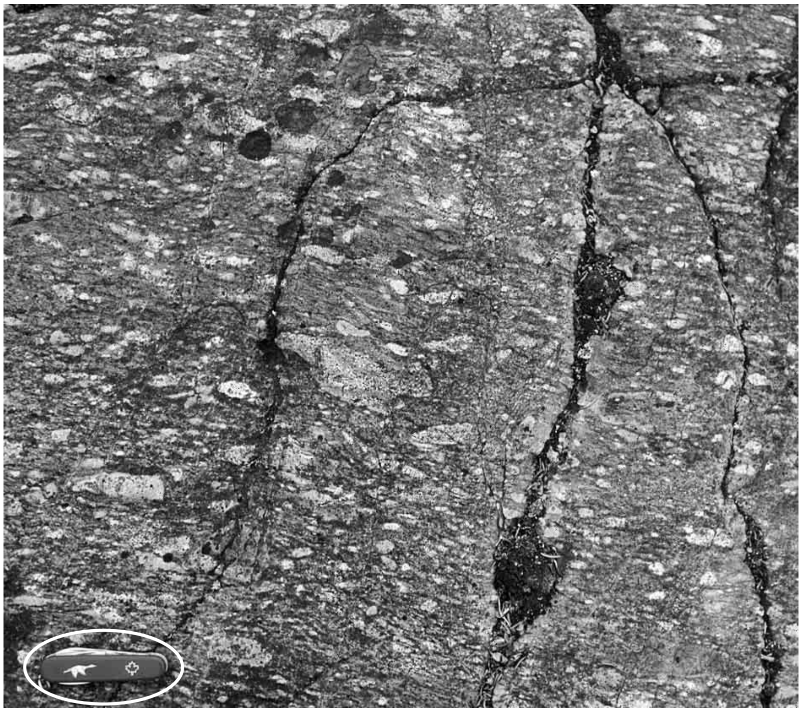


Figure 8.

Heterolithic volcanic breccia with a wide range of compositional and textural variation. The matrix is more mafic than most of the clasts. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416A

No other primary sedimentary features have been recognized, although any that may have been present may have been obscured by the relatively strong foliation in these rocks.

A third facies consists of homogeneous, massive units that locally form topographic highs. The thickness varies from a few metres to more commonly a few tens of metres, and in rarer cases exceeds 100 m. They have a much weaker fabric relative to the surrounding units and commonly break with a conchoidal fracture. In contrast with the other volcanic facies, they are typically more densely porphyritic with subhedral plagioclase and less commonly anhedral to rounded quartz (Fig. 14). In areas where this facies is dominant, individual units are separated by volcanic breccia. They are thought to represent either extrusive flows to domes or possibly shallow sills.

A minor suite of concordant to low-angle felsic sheets occurs locally throughout the volcanic complex. These sheets are most common, or at least apparent, north of the metasedimentary unit in the southwestern part of the complex. They occur in pinkish-grey, metre-scale units that are locally up to 20 m thick. Where abundant, they can form as much as half the outcrop.

One carbonate unit was noted within the complex. This lenticular unit is up to 1.5 m thick and consists of 5–8 cm thick layers of ferruginous dolomite separated by darker green, siliceous mafic material.

A metamorphosed alteration zone occurs within the intermediate to felsic volcanic rocks over an extent of at least 1.5 km southeast of Gale Lake. The rocks contain assemblages of

Figure 9.

Compositionally uniform volcanic breccia with very coarse, poorly sorted clasts. This rock may be an autobreccia, the product of subaqueous, synextrusive fracturing due to quenching and the subsequent alteration along the fractures of the original homogeneous felsic flow or synvolcanic intrusion. Some of the breccia units within the volcanic complex may be the product of such a process. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416B

Figure 10.

Prominent, steep tectonic lineation expressed by the elongation of volcanic breccia clasts. This vertical section is of the same volcanic breccia as shown in Figure 8. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416C quartz, plagioclase, biotite, cordierite, cummingtonite, and anthophyllite (Fig. 15). The rare granitoid clasts found in otherwise dacitic volcanic breccia, mentioned previously, occur within this zone and also contain cummingtonite as the main mafic mineral. These assemblages suggest a significant enhancement in magnesium compared to most of the volcanic unit. Such alteration is common in areas associated with base-metal mineralization (Franklin et al., 1981), although no sulphide mineralization was noted within this region. It is not known whether this alteration is related to volcanogenic processes or to emplacement of the nearby, possibly synvolcanic, metagabbro intrusions.

Immediately north of the volcanic complex, within the dominantly fine-grained, greywacke-mudstone turbidite, a 10 m unit consists of a series of distinctive, thin-bedded, fine-grained, graded, felsic tuff beds that, because of their proximity to the volcanic complex, may be related to it. If so, their presence is an argument for there being no major time

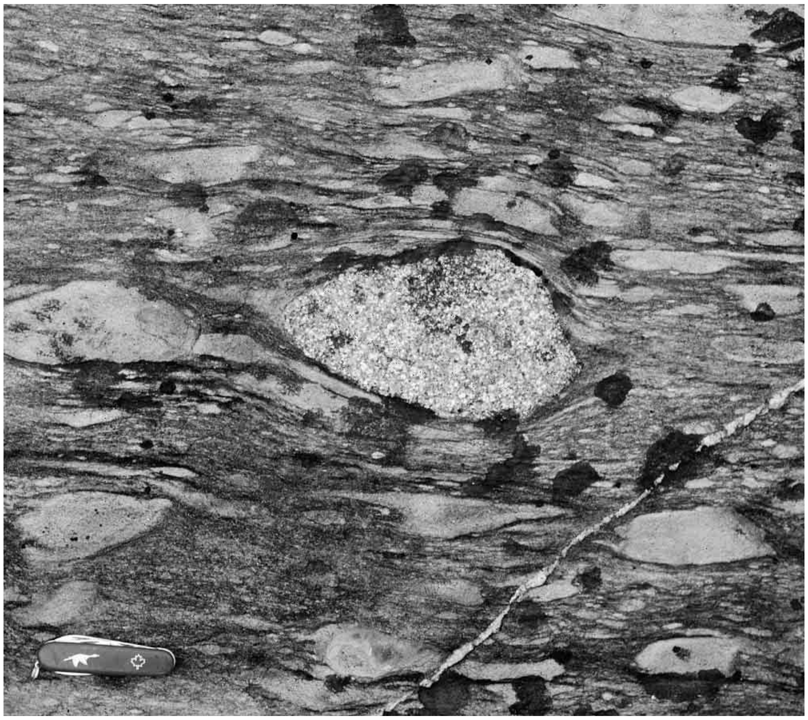


Figure 11.

Rare example of an isolated, medium-grained, granitoid clast within a volcanic breccia. Black spots are lichen. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416D

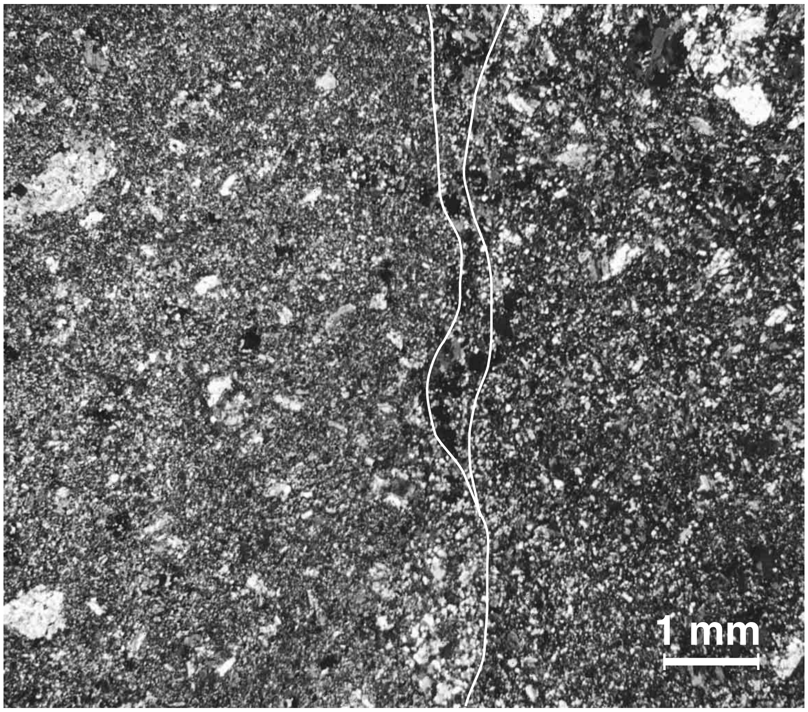


Figure 12.

Photomicrograph of low-grade, felsic volcanic breccia. Note slight variation in texture between clasts demarcated by white lines. Matrix between clasts in central part of figure contains abundant biotite. Low phenocryst abundance is fairly representative and in this case consists almost entirely of plagioclase. Plane-polarized light.

break between volcanism and subsequent turbidite sedimentation. They are discussed more fully in the section "Volcanogenic sedimentary rocks and iron-formation".

Geochronology

A metadacite sample from the greenschist-grade, north margin of the volcanic complex at 'West Wijinnedi Lake' has been dated. Three zircon fractions have concordant, overlapping ages with an average $^{207}Pb/^{206}Pb$ age of 2673 ± 1.4 Ma while a fourth fraction, just over 1% discordant, was interpreted as a 2791 ± 35 Ma xenocryst (Villeneuve and Henderson, 1998; Appendix 1). The 2673 ± 1.4 Ma age is essentially identical to ages determined on two felsic bodies within the volcanic sequences in the vicinity of Indin Lake, about 30 km to the north of the area (2670 ± 2 Ma and 2671 +8/-7 Ma (Pehrsson and Villeneuve, 1999)), although they also reported a 2647 ± 2 Ma age of a felsic volcanic breccia within the metasedimentary

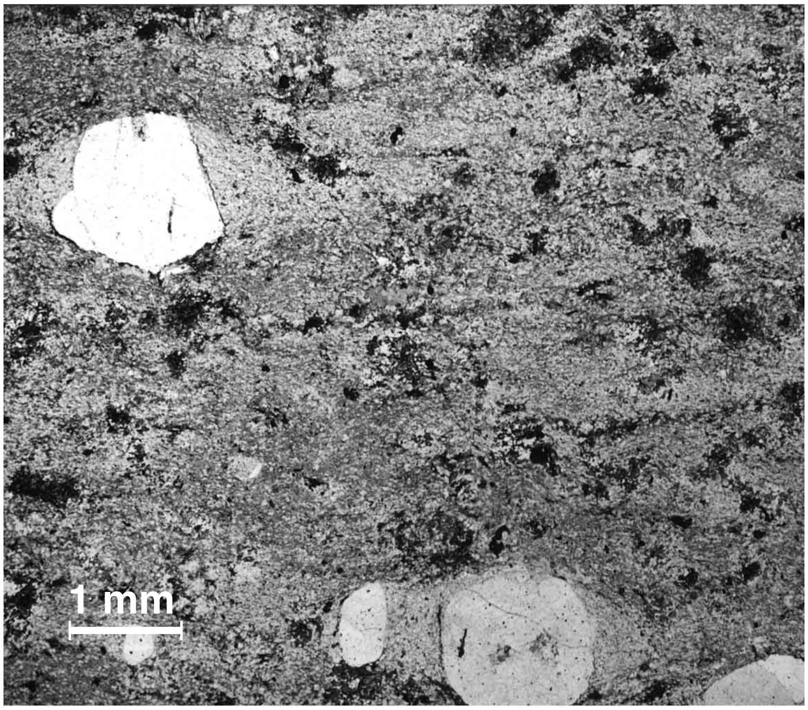


Figure 13.

Photomicrograph of felsic volcaniclastic sedimentary rock in which faint clastic texture is suggested by slightly texturally varied domains in part outlined by mafic mineral concentrations, here mainly chlorite and epidote. This rock is anomalous in the relatively large number of quartz phenocrysts present and lack of plagioclase phenocrysts. Plane-polarized light.

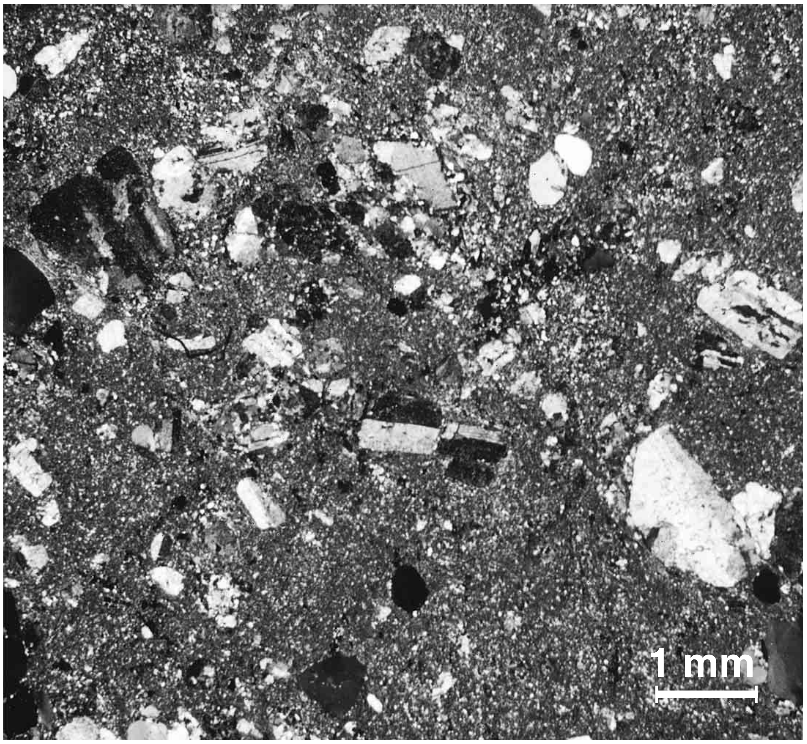


Figure 14.

Photomicrograph of synvolcanic intrusion that characteristically has more plagioclase and quartz phenocrysts, but is similar in composition to the associated volcanic rocks. Plane-polarized light.

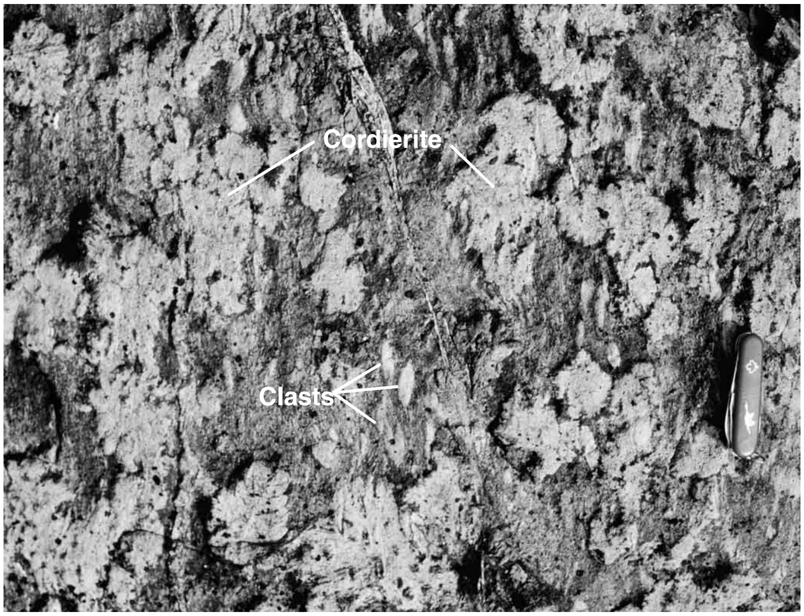


Figure 15.

Altered felsic volcanic rocks with prominent coarse, light-coloured cordierite masses with associated dark patches of biotite, and radiating aggregates of anthophyllite locally at the margins of the cordierite. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416E

rocks. The age of the volcanic rocks at 'West Wijinnedi Lake' is toward the young end of the dominant 2720–2660 Ma Yellowknife Supergroup volcanic age spectrum. It is, however, somewhat older than several other intermediate to felsic volcanic centres in the southwestern Slave Province such as the 2658 +1.2/-0.8 Ma and 2661 +1.2/-1.1 Ma felsic volcanic centres at Russell Lake and Clan Lake, 80 km south and 95 km south-southeast of the area, respectively (Mortensen et al., 1992). The ca. 2.79 Ga age of the presumed xenocryst in the metadacite is at the young end of the age spectrum of basement rocks predating Yellowknife Supergroup in the Slave Province (Bleeker and Davis, 1999).

High-grade intermediate to felsic metavolcanic rocks (unit ĀYvif.o***)***

In the adjacent Ghost Lake domain to the east and southeast of the main volcanic complex are two areas of migmatitic rocks that are considered to be the higher grade equivalents of the intermediate to felsic volcanic complex. The largest of these occurs in a narrow belt between Daran and Ghost lakes. It extends beyond the map area for about 1 km, but is then covered by Daran Lake and is not seen again to the north (Pehrsson and Kerswill, 1997b). These rocks were originally mapped and interpreted as volcanic rocks by both Lord (1942) and Wright (1950b).

Due to the high metamorphic grade and degree of deformation, no primary features other than possible compositional layering have been preserved. The metavolcanic rocks are bounded on each side, in part, by metasedimentary rocks. This would appear to have been a primary relationship as minor units of metasedimentary migmatite occur locally within the metavolcanic unit near its contacts, even where the metasedimentary rocks have been replaced by intrusions of tonalite. It therefore seems reasonable to suggest that the volcanic unit may occur in an anticlinal structure as do the better preserved volcanic rocks in the Wijinnedi domain to the west. The migmatite is best preserved in the northern part of the unit (Fig. 16); to the south the rocks are more deformed with a stronger foliation and straighter layering (Fig. 17).

The unit is a migmatitic gneiss with interlayered melanosome and leucosome that is everywhere at granulite metamorphic grade. It is locally intruded, particularly in the north, by small dykes and veins to outcrop-scale bodies of tonalite, pegmatite, and granite related to nearby major intrusions. These can be distinguished from the leucosome phases of the gneiss without difficulty. The brownish-black- to light-brownish- to rusty-grey-weathering melanosome occurs in 1–10 cm laminated layers. In contrast to the coarser leucosome, it is fine to medium-fine grained, even grained, and dark to light grey to olive green. Mineral assemblages consist of varied proportions of some or all of the following: plagioclase, quartz, biotite, hornblende, clinopyroxene, orthopyroxene, and, in rare cases, garnet. Where best preserved in the north, the leucosome occurs as lower proportions of anastomosing veinlets and lenses to diffuse patches of pale brown- to rusty-weathering, light olive-green, medium-grained tonalite consisting mainly of quartz and plagioclase, but with local coarse-grained hornblende or pyroxene. No equivalents of the metagabbro dykes and sills that are so prominent in the main sequence of lower grade felsic to intermediate volcanic rocks of the Wijinnedi domain were recognized.

A sample of orthopyroxene-bearing leucosome from these migmatitic volcanic rocks has been geochronologically analyzed by Pehrsson et al. (2000) from a small island immediately north of the map area. They suggested the younger, dominant zircon population at about 2588 Ma crystallized under granulite-grade metamorphic conditions while an older population with an age of about 2624 Ma likely represented an inherited component from the original volcanic protolith. As it is now known that the tonalite (unit Āt.o) that intrudes the metavolanic rocks within the area is 2640–2630 Ma, the dated leucosome zircon populations probably both represent metamorphic ages (*see* 'Appendix' for further discussion).

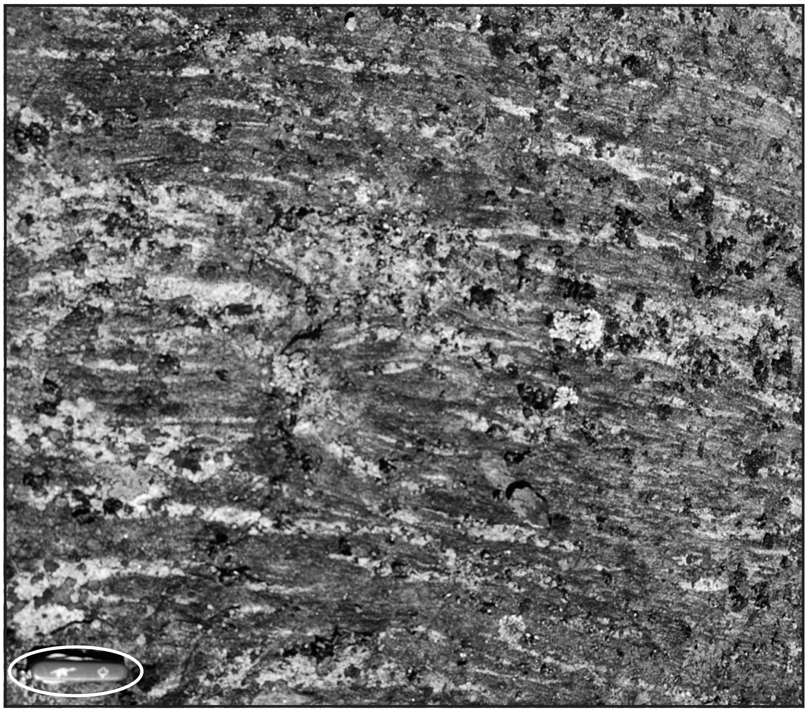


Figure 16.

Less deformed, metavolcanic migmatite at granulite grade from the northern part of the high-grade belt. Orthopyroxene occurs both in the sparse wispy coarse-grained leucosome lenses and patches and in the fine-grained melanosome. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416F

Figure 17.

Finely layered, straight, deformed granulite-grade metavolcanic migmatite from the central part of the belt. Note the large variation in composition of the layers expressed by their colour. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416G

The second group of high-grade migmatitic rocks, presumed to be of intermediate volcanic origin, occurs in the southeastern part of the map area, locally at or near the north margin of the major metasedimentary migmatite unit. In several places, they appear to be almost gradational with the adjacent metasedimentary migmatite due to the intimate intermixing of pyroxene-biotite and garnet-cordierite-biotite layers. Although a much smaller scale unit, the bulk of these migmatitic supracrustal rocks are similar in many respects to the unit north of Ghost Lake. These rocks were presumably more extensive, as zones of inclusions of similar rocks occur locally within the bounding tonalite both north and south of the metasedimentary migmatite belt. Elsewhere in the area, at lower metamorphic grades, the felsic to intermediate volcanic rocks are older than the associated metasedimentary rocks. If these gneiss bodies represent high-grade metavolcanic units and are therefore stratigraphic units, their occurrence on both the northwest and, to a lesser extent, on southeast side of the zone of the metasedimentary migmatite units might suggest that the belt of supracrustal rocks in the southeast corner of the map area occurs in a regionally horizontal synclinal structure.

These two zones of granulite-grade migmatitic gneiss of probable intermediate to felsic volcanic origin are separated by an extensive terrain of granulite-grade tonalite and tonalite gneiss. Supracrustal inclusions, mainly metasedimentary migmatite to diatexite, are common locally. Pyroxene- and amphibole-bearing gneiss inclusions, lithologically similar to the contiguous volcanic gneiss bodies to the northwest and southeast, are abundant only on the large bulbous peninsula at the east end of Ghost Lake and, to a lesser extent, in the tonalitic gneiss east of the small river flowing north into Ghost Lake. These inclusions may represent a link between the two.

Discussion

The felsic-intermediate volcanic centre of the Wijinnedi domain is similar in many respects to other volcanic centres in the Slave Province. These include the volcanic centre east of Russell Lake (Henderson, 1985; Jackson, 1990) and the volcanic centre at Clan Lake (Hurdle, 1983), about 110 km to the south-southwest and south-southeast of the map area, respectively. A somewhat smaller example, occurs just north of Point Lake (Henderson, 1998a), but the largest and best understood example is the Back River volcanic complex in the eastern Slave Province (Lambert, in press, and references cited therein). Each of these are more or less equidimensional centres, surrounded by greywacke-mudstone turbidite units, in which mafic volcanic rocks, where present, are a relatively minor component.

An unusual feature of this volcanic centre is the very high proportion of dacitic volcanic breccia. Similar breccia facies occur in the other centres, but on the whole are a relatively minor feature. Such breccia units commonly represent quench-fragmented lava flows, crypto domes, shallow intrusions, lava dome and accompanying flow-front talus deposits, and epiclastic redeposition and mass wastage (Cas and Wright, 1987). Given the rather high degree of deformation of most of this unit as reflected in the generally large aspect ratios of the clasts, it may be difficult to define particular volcanic environments more precisely. The overlying formations, pillowed mafic to intermediate volcanic rocks which in turn are overlain by fine-grained, greywacke-mudstone turbidite units, are subaqueous. This may suggest the abundant breccia units are more likely a product of autobrecciation in the coarse of extrusion of the lava than due to subaerial epiclastic redeposition and mass wastage.

The major felsic to intermediate volcanic centre of the Wijinnedi domain and the two smaller units to the east and southeast, although differing significantly in scale, metamorphic grade, and degree of deformation, may well have been originally stratigraphically equivalent. While this would be difficult to prove, it remains an attractive, if only potential, stratigraphic link between the domains.

Mafic to intermediate metavolcanic rocks (unit AYvmi*)*

Mafic to intermediate volcanic rocks conformably overlie the intermediate to felsic volcanic rocks on the northern and eastern sides of the dome in the Wijinnedi domain within the map area. The greatest thickness of these more mafic rocks are at the eastern end of the dome. They may continue to the west beyond the area and join up with a volcanic centre that occurs immediately to the northwest of the intermediate to felsic volcanic dome (Fig. 6) (Lord, 1942). The composition of this centre to the west is mainly intermediate with lesser amounts of mafic volcanic rocks (Jackson, 2003). Mafic volcanic rocks also occur along the southern margin of the dome, although the mafic rocks and the intermediate to felsic volcanic rocks that they are associated with are separated from the main part of the dome by a shear zone of unknown displacement. Metagreywacke-mudstone turbidite deposits, not mafic volcanic rocks, overlie the intermediate to felsic volcanic rocks north of this shear zone, a relationship that continues to the west beyond the map area (Fig. 6) (Jackson, 2003). Thus the relationship of the mafic volcanic rocks south of the shear zone to those to the east and north of the dome proper is not directly apparent. Two zones of more mafic volcanic rocks occur within the intermediate to felsic volcanic dome. Their roughly symmetrical distribution north and south of the axial trace of the structure suggests that they may be correlative.

Contacts

The contact between the mafic to intermediate volcanic rocks and the underlying intermediate to felsic volcanic rocks is conformable and gradational. In the contact area, more mafic and more felsic compositions are interlayered, commonly over tens to hundreds of metres. On the north side of the dome, the contact between the more mafic volcanic rocks and the overlying metasedimentary rocks is also conformable, but sharp on the scale of a metre or so. On the southeast side of the dome, where the rocks are also at much higher metamorphic grade, the contact between the mafic volcanic rocks and metasedimentary migmatite is a shear zone, although the local occurrence of marble at this contact, similar to that found locally along the northern contact, suggests the present contact is likely a sheared primary contact. The contact between the southernmost, mainly mafic volcanic rocks that marks the boundary between the Wijinnedi and Hinscliffe domains is a knife-sharp break within a kilometre-scale mylonite zone.

Like the underlying, more felsic volcanic rocks, the mafic volcanic rocks are locally intruded by varied proportions of metagabbro sills and dykes presumably related to the metagabbro intrusive complex in the central part of the volcanic complex. In the more deformed, high-grade rocks in the south it is commonly difficult to differentiate between the intrusive and extrusive mafic rocks. Thin granitoid dykes and sills are a very minor intrusive phase in these higher grade rocks. In the southeast, these minor intrusions are granite and pegmatite, similar in some respects to the large granite to the southeast in the Ghost domain (unit Ag). The thin, annealed mylonitic granitoid sills north of the Hinscliffe complex are of unknown affinity.

Lithology

The mafic to intermediate volcanic unit consists of a rather heterogeneous assemblage of volcanic rocks at a varied range of metamorphism and degree of deformation. Best preserved primary structures and textures occur in the north. The degree

Figure 18.

Irregular, pillowed mafic volcanic rocks. Brunton compass is 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417A

of deformation increases toward the boundary of the Wijinnedi domain, with the mafic volcanic rocks adjacent to the Hinscliffe domain being essentially mylonite. Metamorphic grade increases from lower greenschist, chlorite-dominant assemblages through to hornblende-, plagioclase-, garnet-, and titanite-bearing assemblages toward the domain boundary. Clinopyroxene was noted locally close to the boundary, but no orthopyroxene was seen.

The consistently most mafic material occurs at the top of the unit over the apparent thickest part of the sequence southeast and east of Wijinnedi Lake. This would also appear to be the pattern on the south side of the volcanic complex as well, although the mafic to intermediate sequence there is severely telescoped due to subsequent deformation. The most mafic mylonitic amphibolite occurs at the contact with the Hinscliffe complex, but grades in the northern part of the unit into a more heterogeneous volcanic assemblage. Where best preserved, the unit consists mainly of pillowed and massive flows with lesser amounts of mafic volcaniclastic material (Fig. 18). The flows are locally amygdaloidal with quartz- or carbonate-filled vesicles. The concentration of these vesicles toward the top of individual pillows commonly provides a more useful top indicator than the all-to-commonly deformed shape of the pillows themselves. To the east and south as the mafic volcanic rocks become increasingly deformed and the metamorphic grade increases, the rocks assume a more layered aspect defined by both compositional and textural variations. With increasing deformation, the layering becomes both thinner and more pronounced. In the northwestern most part of the unit, west of 115°22′W longitude, the entire unit consists of interlayered, more andesitic volcanic units and layers to lenses of dacite and dacitic volcanic breccia.

Within the dacite-dominated volcanic centre are several zones of mafic to intermediate volcanic rocks that are similar in many respects to the heterogeneous volcanic sequence. A rather distinct mafic zone dominated by mafic dacite to perhaps felsic andesite extends for almost 10 km in the volcanic sequence south of ‘Glazebrook Lake’. At least one unit of probable basaltic composition locally contains pillow-like structures. On the other hand, amphibolite sills related to the central metagabbro intrusion are common in this region and the two can be difficult to differentiate. The sills elsewhere contain early, altered, premetamorphic fractures that, due to further deformation and metamorphism, look like pillow structures (Fig. 19).

This relatively thin unit of mafic-dominated, heterogeneous volcanic rocks differs from the major, thick, homogeneous mafic volcanic sequences that dominate the Slave Province, particularly, but not exclusively, in the west. These would include the Kam Group at Yellowknife (Henderson and Brown, 1966), the Cameron River and Sunset Lake basalt units to the northeast of Yellowknife (Lambert, 1988), and the Point Lake Formation at Point Lake (Henderson, 1998a), among others. Given the close association of the mafic volcanic rocks in the Wijinnedi Lake area with the felsic volcanic centre and their heterogeneity, they may have a closer affinity, in character if not scale, to the mafic rocks of the Back River volcanic complex (Lambert et al., 1992) or parts of the Hackett River Group to the north (Frith, 1987).

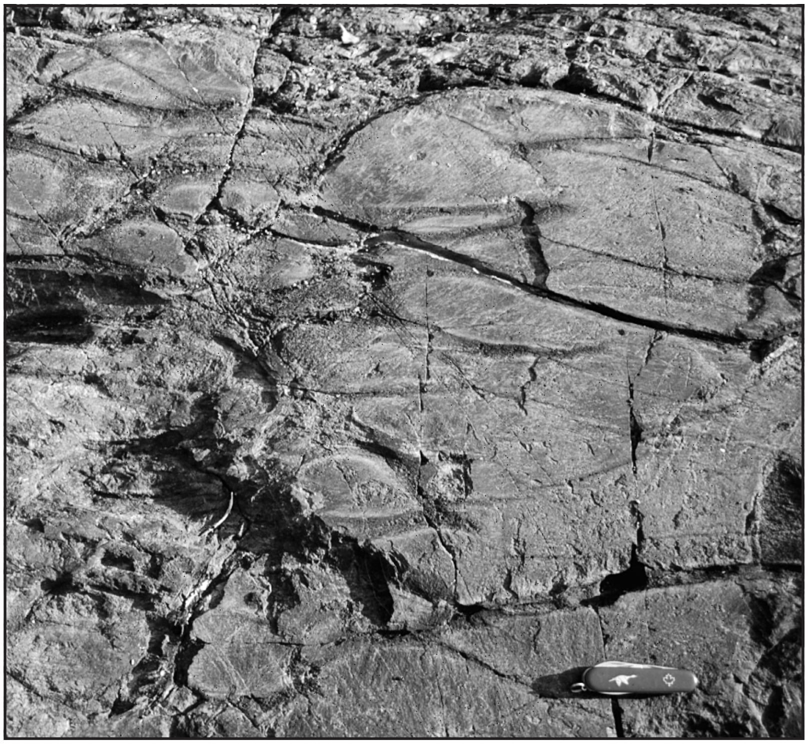


Figure 19.

Pseudo-pillow structures in otherwise massive mafic unit are due to alteration along premetamorphic fractures that were subsequently deformed. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416H

Associated carbonate rocks

At several localities at the top of the mafic to intermediate volcanic sequence are metre-scale occurrences of interlayered marble and calc-silicate. Examples were seen at the contact between the mafic volcanic rocks and the overlying metasedimentary rocks south of 'Colson Lake' and 2 km east of 'Gehl Lake'. The material is discontinuous, although in the southern example was traced over 100 m. Lord (1942) reported the occurrence of another example near the west end of 'West Wijinnedi Lake' as well as an unusually thick, 27 m unit from the same stratigraphic position about 2 km or 3 km west of the map area. The occurrence of carbonate units at the transition between volcanic rocks, particularly intermediate to felsic volcanic units and overlying greywacke-mudstone turbidite units is not unusual elsewhere in the Slave Province (Henderson, 1998a, p. 58). Where deformation and metamorphism is minimal, such units can be stromatolitic (Henderson, 1975a; Lambert, 1998).

Metasedimentary rocks

Variably metamorphosed greywacke-mudstone turbidite of the Yellowknife Supergroup occurs throughout the area. At lower metamorphic grades, primary sedimentary features are commonly preserved in metasedimentary rocks of the Wijinnedi domain; however, in the Ghost, Dauphinee, and even the easternmost Wijinnedi domains the metasedimentary rocks are present as their high-grade migmatitic equivalents. In the following sections, the primary sedimentary aspects of these rocks are discussed based mainly on those of the Wijinnedi domain followed by a discussion of the high-grade metamorphic rocks, primarily of the Ghost domain, in which few, if any, primary features are preserved.

Low-grade metasedimentary rocks of the Wijinnedi domain (units AYs.cl, AYs.bi, AYs.cd, AYs.si)

The metasedimentary rocks consist almost entirely of greywacke-siltstone-mudstone turbidite as is the case for most of the sedimentary rocks of the Slave Province. These rocks, however, are distinctive due to the very high proportion of pelitic material throughout the area. The lowest grade rocks, the chlorite zone of the greenschist facies, occur north of 'West Wijinnedi Lake'. The metamorphic gradient increases both to the west and east where it rises through the biotite, cordierite, and sillimanite zones of the amphibolite facies and ultimately to migmatite-zone rocks north and east of the volcanic rocks of the domain. Primary sedimentary structures and textures are moderately to well preserved in the greenschist-grade rocks, the main variable being the degree of development of cleavage in the rocks (Fig. 20). Most primary textures are lost due to recrystallization in the amphibolite-grade rocks, although bedding commonly is evident and facing directions can be determined by grain-size gradation in the coarsest grained sandstone even at elevated metamorphic grades (Fig. 21).

The sedimentary rocks conformably overlie the volcanic complex that has a U-Pb (zircon) age of 2673.3 ± 1.4 Ma (Villeneuve and Henderson, 1998; Appendix 1). No younger stratigraphic units are known to overlie them. The sedimentary rocks within the area have not been dated directly, but the turbidite units continue to the north beyond Indin Lake (Fig. 6). There, zircons within felsic volcanic breccia units that occur within a relatively coarser grained turbidite facies have been dated at 2647 ± 2 Ma (Pehrsson and Villeneuve, 1999). Zircons in a finer grained facies at Damoti Lake (Fig. 6) have been dated at 2629 ± 2 Ma (Pehrsson and Villeneuve, 1999). These have been interpreted as detrital grains, and if their age represents the age of a component of the source terrain, this would suggest a maximum age of sedimentation of 2629 ± 2 Ma. If these sedimentary rocks are correlative with those within the Wijinnedi Lake area, a time break of as long as 42 million years across the conformable contact between volcanic and sedimentary rocks would be implied. The existence of such a long period of time in an active volcanic environment during which no structural discordances are recognized that formed prior to sedimentation taking place is somewhat paradoxical, particularly given the emplacement of two major plutonic suites in the region during that time period (*see* sections 'Hinscliffe complex (unit ĀH)' and 'Tonalite (units Āt, Āt.o, Ātgn, Ātgn.o, Āt-gm, Āt-gm.o)').

The sedimentary rocks are intruded locally by Archean mafic dykes, sills, and irregular bodies that are metamorphosed in context with the metasedimentary rocks they intrude. These mafic intrusions are thought to be related to the

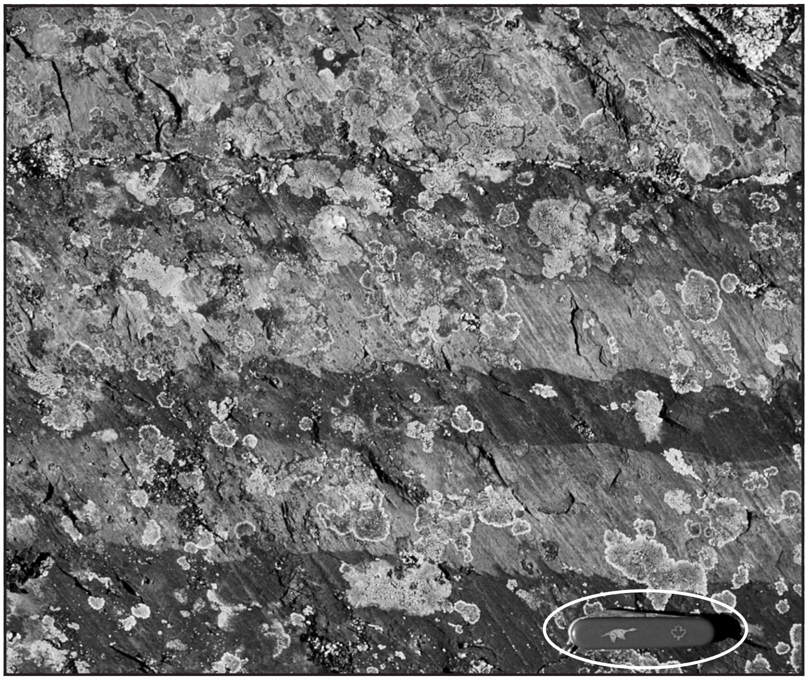


Figure 20.

Light-coloured, thin-bedded, fine-grained, biotite-zone greywacke to siltstone turbidite layers grade into darker coloured pelite. Grain-size gradation, ripples, flame structures, and included shale chips can be seen between lichen patches. The moderately developed cleavage resulting in the asymmetry of the flame structures at the base of the beds is approximately parallel to the more prominent glacial striae. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416I

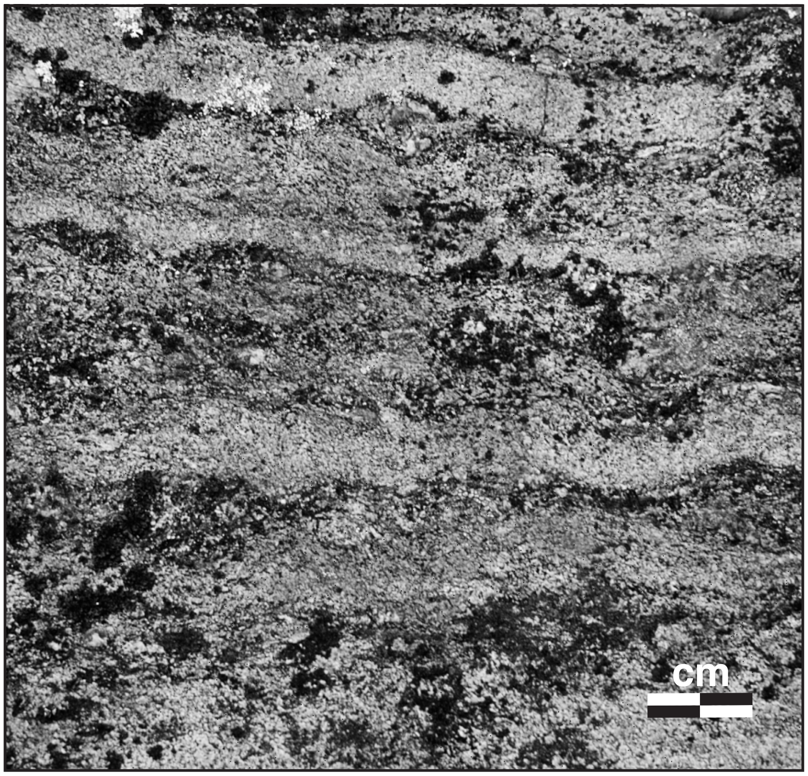


Figure 21.

Thinner bedded, finer grained, and perhaps more typical turbidite layers than those seen in Figure 20, but in the sillimanite zone. The porphyroblasts most evident in the now coarser grained metapelite are cordierite. Photograph by S.E. Schaan. GSC 2002-417B metagabbro plutons and associated dykes and sills that intrude the core of the volcanic complex to the south. The sillimanite-zone metasedimentary rocks west of 'Colson Lake' and the cordierite-zone rocks southwest of 'Glazebrook Lake' contain a few small granite plugs.

The metasedimentary rocks have a similar east-northeasterly trend with steep dips mainly to the north in the main part of the domain. This trend swings to the northeast and north with steep southeast dips dominant in the higher grade metasedimentary rocks to the east. Facing directions of beds is either to the north or south and can reverse over short distances indicating the sedimentary sequence is tightly folded. The everywhere present, but variably developed principal foliation is in most cases subparallel to bedding.

The contact with the underlying volcanic rocks is sharp. As is commonly, but not invariably the case elsewhere in the Slave Province, there is no interlayering of volcanic and sedimentary rocks after sedimentation began. The sedimentary rocks at the contact are locally carbon rich, sometimes with finely disseminated sulphide minerals. Patchy, as opposed to bedding-controlled gossans on the scale of a few metres to ten or so metres are developed on these rocks at several places along the contact. These zones are invariably associated with the metagabbro sills that are common throughout the region, there presumably being a cause-and-effect relationship. Similar gossans are not known to be associated with these rocks away from the contact area.

Greywacke-mudstone turbidite

The bulk of the formation consists of a pelite-dominated turbidite sequence with at least three-quarters of the unit being pelitic. Bedding couplets consist of light grey to brownish-grey basal siltstone to fine-grained greywacke typically a few centimetres thick grading up into dark grey to black pelite that is commonly in excess of 10 cm thick (Fig. 20). Locally the siltstone to fine-grained greywacke is thicker, with groups of beds up to several tens of centimetres in thickness, whereas the pelite component can be proportionately less. Invariably the maximum grain size is 0.5 mm or less and the pelite component dominates. Sedimentary structures include graded bedding, climbing ripples, flame and load structures at the bases of beds, silt and fine-grained sand intrusions, mud-chip breccia, and rare carbonate concretions in thicker greywacke units (Fig. 20).

Randomly superimposed on this background of thin-bedded, pelite-dominated sediments are thick units of yellow-brown-weathering, fine-grained greywacke up to 4 m thick, but more commonly on the order of 1 m or less. On an outcrop scale at least, no evidence of major channelling associated with these unusually thick units was noted. Elsewhere in the Slave Province where unusually thick greywacke beds occur, there is usually evidence for amalgamation, that is, the thick bed represents two or more depositional events with the finer grained upper parts of the earlier deposits eroded by the current that brought in subsequent deposits (Henderson, 1975b). This may well be the case here as well, but due to the fine grain size of these sandstone units, erosional scours within the bed are difficult to recognize. Although some of the more quartz-poor examples were previously interpreted as possible epiclastic deposits derived from nearby felsic volcanic centres (Henderson, 1994, 1998a), petrographic examination indicates these sediments are much the same as Yellowknife Supergroup greywacke elsewhere in the Slave Province. These unusually thick units are thought to represent large turbidity currents that were ponded in the area of deposition due to topographic variations on the depositional floor. In addition, reflection of the depositing current from topographic highs may have resulted in the passage of the same current over the same point several times resulting in the anomalous thickness in the manner suggested by Hiscott and Pickering (1984).

Petrographically, the coarser greywacke consists of varied proportions of lithic clasts, quartz, and feldspar that grade from a maximum grain size of about 0.5 mm to matrix-sized particles (Fig. 22). Diffuse to sharply bounded lithic clasts are the most common component. They consist of fine-grained aggregates of quartzofeldspathic material with varied amounts of fine-grained chlorite, clinozoisite, carbonate, and opaque minerals. Some are texturally similar to felsic volcanic rocks, whereas others appear to represent the end-member of progressively altered feldspar. Most are of indeterminate origin. Invariably present, but rare, perhaps due to the fine-grained nature of the sandstone units, are quartz-feldspar aggregates suggestive of granitoid provenance (Fig. 22). Anhedral, broken to subrounded quartz varies from about 10% to 40% of the sandstone. The coarser grains are multidomainal with irregular sutures reminiscent of plutonic quartz (Fig. 22). The typical squarish outlines of volcanic quartz, where present, are rare. Feldspar varies from between 5% and 15% of the rock and is almost entirely plagioclase. It is variably altered, with the most altered grains similar to many of the lithic clasts. Rare examples of microcline, perthite, and myrmekite are present in some cases. Detrital flakes of muscovite and biotite are also a minor component.

Volcanogenic sedimentary rocks and iron-formation

Although the vast majority of Yellowknife Supergroup metasedimentary rocks in the map area consist of mudstone-dominant, greywacke-mudstone turbidite sequences, two other volumetrically minor facies also occur within the unit. On the Snare River about 600 m north of the volcanic contact is a 10 m thick unit of volcanogenic sedimentary rock that extends over at least 1 km along strike. It consists of distinct yellow-brown-weathering, very fine-graded siltstone units that contain green, millimetre-scale, metacryst-like structures (Fig. 23). In thin section, these structures consist of a core composed of quartz and minor biotite surrounded by a decussate mass of chlorite, the outer margin of which defines an euhedral crystal form, possibly that of an amphibole (Fig. 24). The chlorite zone is followed by a diffuse region that is slightly coarser grained and more biotite rich than the surrounding sediment. These sediments are similar to the somewhat more abundant and thicker bedded sedimentary rocks that have been interpreted as volcanogenic and are associated with a small felsic volcanic centre in the Contwoyto Formation at Point Lake (Henderson, 1998a). These are considered to be tuffs; as such they are indicative of contemporaneous volcanism during sedimentation.

Iron-formation occurs within the greywacke-mudstone turbidite sequences east of Wijinnedi Lake and in particular north, west, and south of 'Colson Lake'. Individual iron-formation units can be traced over several hundred metres and the northerly trending iron-formation-bearing zone can be traced for several kilometres within the map area. To the north, the same zone occurs more or less continuously to the north end of Indin Lake (Fig. 6) (Pehrsson and Kerswill, 1997a, b). Significant gold prospects associated with these iron-formation units are known at Damoti Lake and east of Fishhook Lake (Fig. 6) (Pehrsson and Kerswill, 1997b).

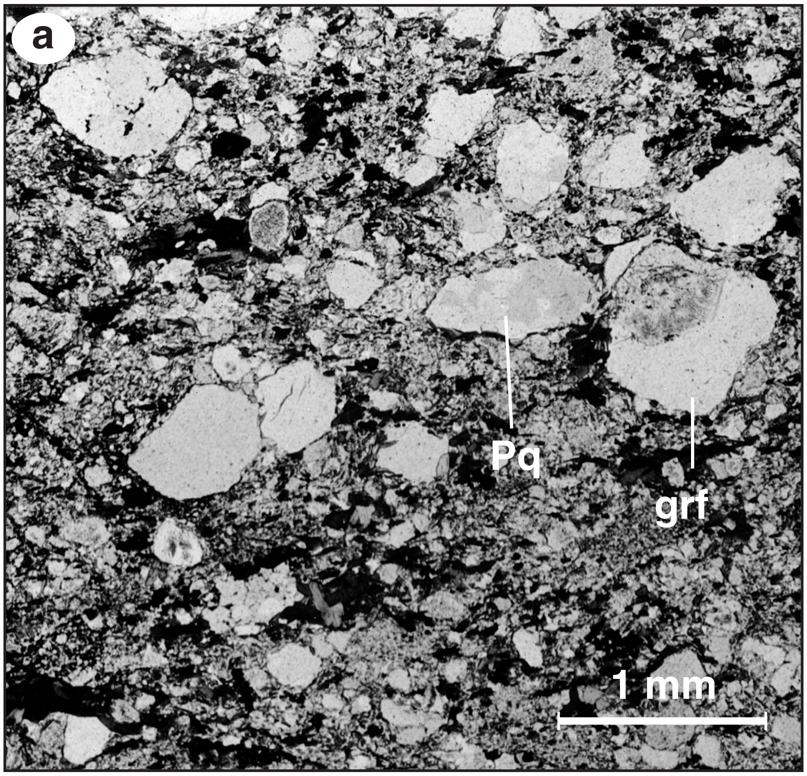


Figure 22.

Photomicrograph of unusual coarser grained greywacke consisting primarily of angular quartz mainly of pluton origin, very minor plagioclase, but abundant fine-grained, quartzofeldspathic, lithic clasts thought to represent in large part weathered feldspar grains. Note occasional quartz-feldspar aggregate suggesting at least in part a granitoid provenance; Pq, plutonic quartz; grf, granitoid rock fragment. ***a****) Plane-polarized light.*

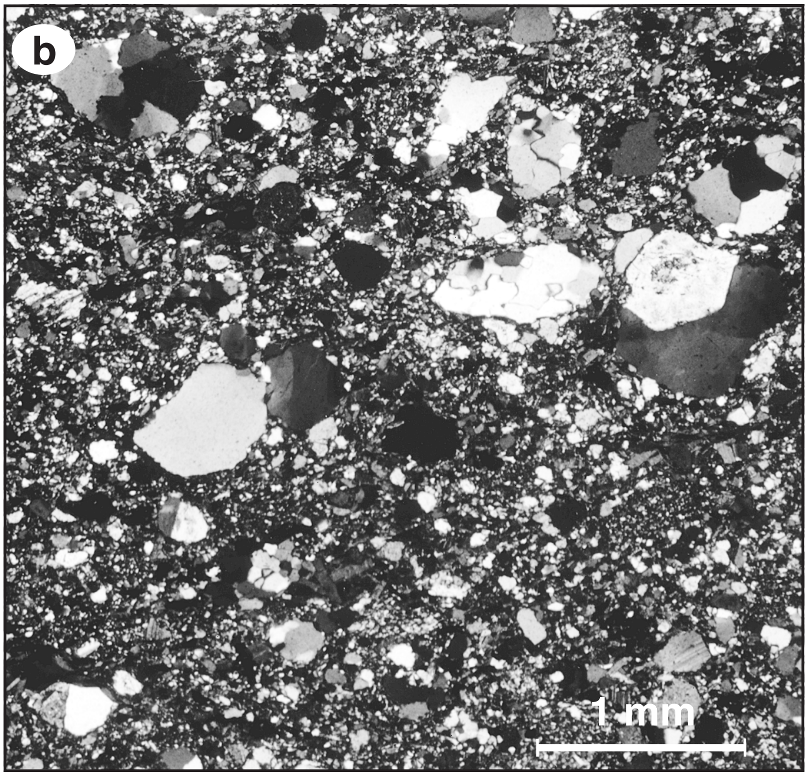


Figure 22.

b*) Cross-polarized light.*

Within the map area individual iron-formation units are up to 10 m thick and are separated by a few metres to several hundred metres of the greywacke-mudstone turbidite sequences. It is not known if the repetition is due to the presence of more than one unit, folding, or both. Pelitic rocks near the iron-formation units are more iron rich than normal as shown by the presence of garnet, something not commonly seen in the pelitic metasedimentary rocks at this metamorphic grade elsewhere in the area. The iron-formation units are everywhere of the silicate type and are composed of assemblages of grunerite, garnet, hornblende, and quartz, with minor sulphide present locally (Fig. 25, 26, 27). Individual iron-formation units consist of 1 cm to 10 cm layers defined by the varied proportion of iron-silicate minerals present. Quartz is present

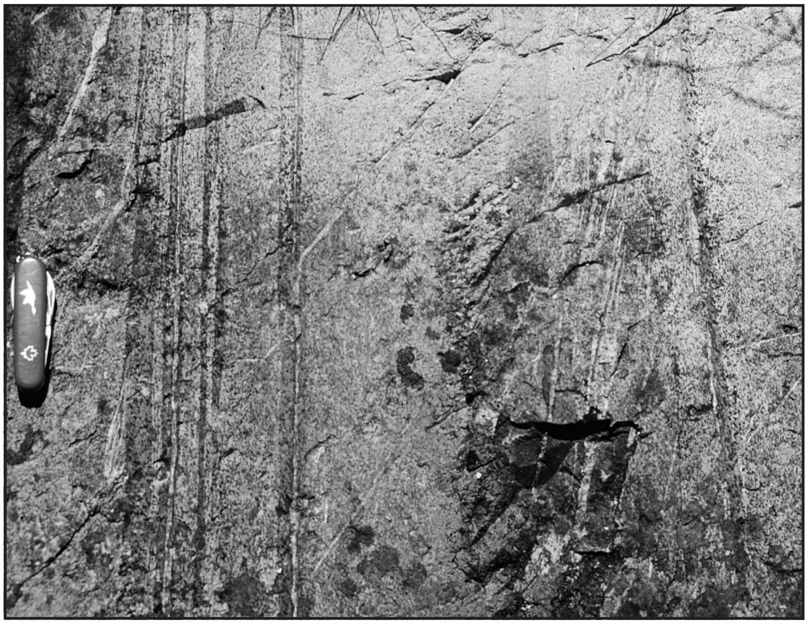


Figure 23.

Fine-grained, weakly graded, volcaniclastic beds separated by thin pelitic layers to laminae. These unusual yellow-weathering rocks with green, millimetre-scale, probable amphibole porphyroblasts, now retrograded, are quite distinct from the normal greywacke-mudstone turbidite sedimentary rocks. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416J

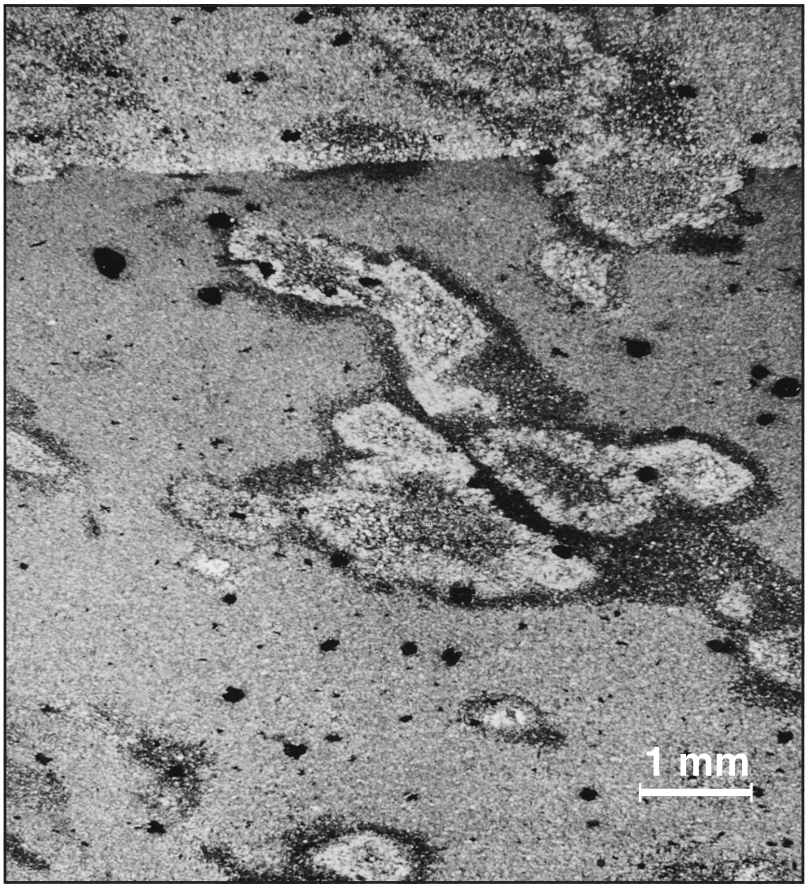


Figure 24.

Photomicrograph of graded volcaniclastic siltstone of Figure 23 with retrograded porphyroblasts of what may have been originally an amphibole. The original porphyroblast has been replaced by a light zone of decussate chlorite that outlines the original euhedral to subhedral crystal shape; a darker core zone of quartz and minor biotite; and a diffuse outer, coarser grained, darker zone in the matrix surrounding the pseudomorph that is enriched in biotite. Plane-polarized light.

Figure 25.

Silicate iron-formation in sillimanite zone southwest of 'Colson Lake'. Thick garnet-amphibole layers containing fine-grained quartz nodules that represent original disrupted cherty laminae are interlayered with laminated amphibole-rich layers. Fine-grained greywacke occurs within the sequence of layered iron-formation. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416K

in most layers and is dominant in some, generally thin, layers. It also occurs as several centimetre-scale nodules within some of the iron-silicate-rich layers. The southernmost of these iron-formation units are gold-bearing in association with pyrite and pyrrhotite where they are intersected by quartz veins. These showings were trenched and drilled during the late 1940s (Wright, 1950a, 1954). The iron-formation units of the area, along with the greywacke-mudstone turbidite sequences that they are contained in, are similar in many respects to those of the Contwoyto Formation between Point and Contwoyto lakes that are more completely discussed in Kerswill (1993, 1996) and Henderson (1998a) and references cited therein.

Metasedimentary migmatite and diatexite of the Ghost, Wijinnedi, and Dauphinee domains (units ĀYs.m, ĀYs.mk, ĀYs.mg, ĀYs.mo)

Most of the metasedimentary rocks within the map area are migmatite. They occur throughout the Ghost domain, in the easternmost Wijinnedi domain and locally within the Dauphinee domain. The metamorphic grade of the migmatite increases toward the southeast with four metamorphic zones outlined on the map. The lowest grade migmatite (unit ĀYs.m) occurs in the eastern Wijinnedi domain north of 'Colson Lake' where the migmatite, with well developed leucosome phases, does not contain K-feldspar or show any evidence of having done so. The other migmatite of the Wijinnedi domain, southeast of the volcanic rocks and the northwesternmost migmatite of the Wijinnedi domain (unit ĀYs.mk) are, or were, K-feldspar-bearing. Of the remaining migmatite bodies to the southeast, the bulk of the migmatite units northwest of Ghost Lake are in the garnet zone (unit ĀYs.mg). Garnet also occurs locally in the central part of the K-feldspar zone. It is not clear whether this is due to local, possibly stratigraphically controlled compositional anomalies, the occurrence of the granitoid plutons in the same area and hence local elevated metamorphic conditions, or unrecognized structural complications. The highest grade metasedimentary rocks occur at, and southeast of, Ghost Lake and contain orthopyroxene (unit ĀYs.mo). The mineralogy of these rocks is discussed more fully in the 'Metamorphic geology' section.

The metasedimentary migmatite units have lost almost all primary sedimentary features, the only exception being possible grading in some of the relatively rare psammitic rocks. The generally gradational contact between the metasedimentary migmatite and the high-grade, intermediate metavolcanic rocks between Ghost and Daran lakes is considered

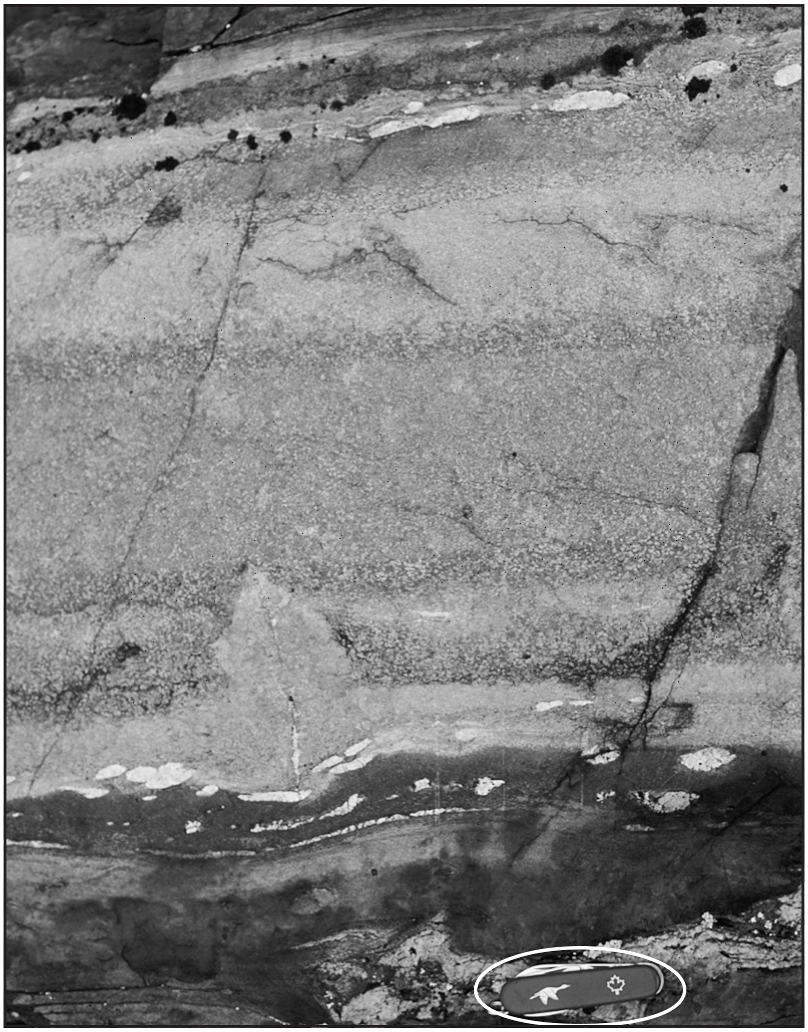


Figure 26.

Silicate iron-formation in sillimanite zone southwest of 'Colson Lake'. This unit is dominated by a thick, diffusely layered, garnet-rich zone. White quartz nodules presumably represent disrupted chert layers. The amphibole-rich layer at the base of the photograph contains disseminated sulphide. The small wedge-shaped structure near the base of the garnet layer is a synsedimentary intrusion of material from the basal garnetite layer. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416L

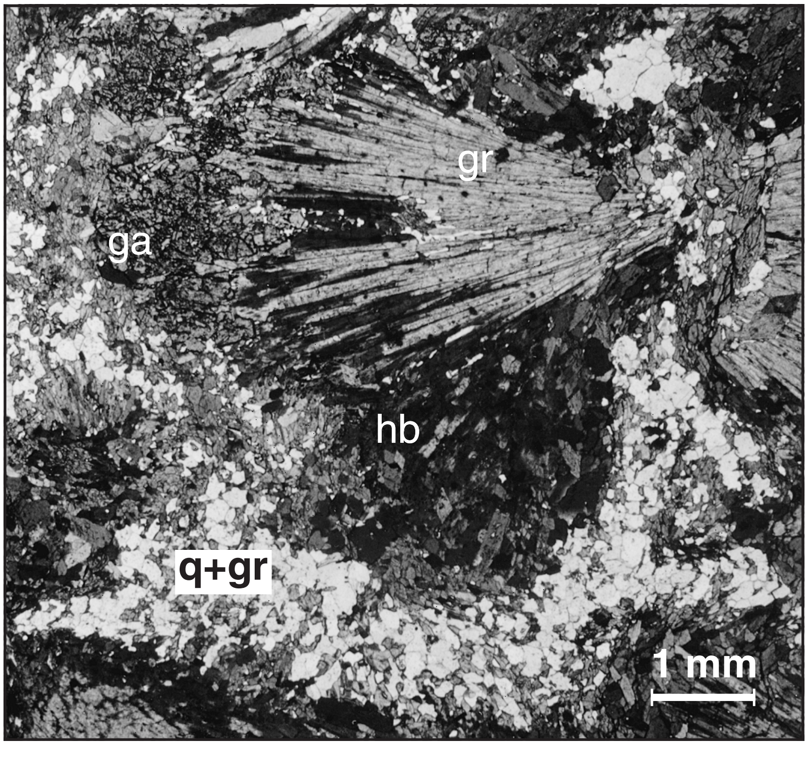


Figure 27.

Photomicrograph of sillimanite-zone, silicate iron-formation consisting of radiating sheaths of neutral-coloured grunerite (gr) overgrown by dark hornblende (hb). Anhedral garnet (ga) contains abundant grunerite inclusions. Fine-grained quartz (q) occurs interstitially to the coarse-grained, iron-silicate mineral aggregates. Plane-polarized light.

to be an original stratigraphic relationship. Similar high-grade, presumed metavolcanic rocks occur locally along the northern margin of the large metasedimentary migmatite unit in the southeastern corner of the map area. They have a similar gradational contact with intimately interlayered garnet-cordierite-orthopyroxene metasedimentary migmatite and leucocratic clinopyroxene-orthopyroxene metavolcanic migmatitic gneiss over a few metres, again suggesting a stratigraphic relationship. While these contacts may be stratigraphic, there remains no direct evidence of their facing direction. If, however, the metavolcanic migmatite units are correlative with the main volcanic centre in the Wijinnedi domain, perhaps not an unreasonable suggestion, then the metasedimentary migmatite units may well lie above the metavolcanic rocks in the Ghost domain as well.

The metasedimentary migmatite bodies, particularly in and above the garnet zone, are extensively intruded by granitoid rocks that are too small to outline at the present mapping scale. Similarly, the mapped intrusions can contain large amounts of metasedimentary inclusions, only the largest of which are mapped separately or shown as inclusion-rich zones on the map. The tonalite gneiss (units Atgn, Atgn.o) is gneissic due in part to the presence of deformed metasedimentary inclusions.

Metatexite

The migmatite units include both stromatic migmatite or metatexite consisting of interlayered to lenticular quartz-plagioclase-rich leucosome and biotite-rich melanosome, and relatively homogeneous diatexite in varied proportions in all metamorphic zones. In general, metatexite is more common at lower grades, but on the other hand, is more abundant than diatexite in the highest grade rocks in the southeastern part of the area. The typically white leucosome varies from foliation-parallel and possibly original bedding-parallel, fine-grained, millimetre-scale laminae to coarse-grained layers a few tens of centimetres thick. It also occurs as elongate lenses to irregular patchy bodies at a variety of scales (Fig. 28, 29, 30). The leucosome consists of quartz and plagioclase with cordierite, K-feldspar, and garnet variably present at increasingly higher grades. The leucosome cordierite, originally recognized by Folinsbee (1940a, b), can occur in spectacular blue to violet-blue subhedral crystals up to 10 cm in length and, although commonly fractured, can be of gem or near-gem quality. Thicker mobilized sills and less commonly discordant dykes of similar composition to the leucosome are also associated with the metatexite. The proportion of leucosome is highly varied, but usually less than about 60%. It does not appear to be related to the grade of metamorphism, but is more likely a function of the pelite content of the original sediment.

The brown- to sometimes rusty-weathering melanosome (used in the descriptive sense of Ashworth (1985)) consists of a more varied, medium-grained, inequigranular, generally biotite-rich sequence of layers. They are typically more abundant and occur in thicker layers than the associated leucosome. Quartz, plagioclase, and biotite are everywhere present and dominant. Potassium feldspar, muscovite, cordierite, sillimanite, garnet, and orthopyroxene have a more varied presence and amount depending on composition and degree of prograde and retrograde metamorphism. Locally garnetite layers are present with in excess of 50% garnet. Less commonly, more psammitic material is recognized as fine- to medium-grained layers composed of quartz and plagioclase with only minor biotite. In rare examples, 1–2 m long sequences of such beds consisting of a series of 5–10 cm beds,

Figure 28.

Migmatite-zone metatexite close to sillimanite-zone boundary with thin leucosome that occurs both as in situ laminae and thicker, possibly mobilized, leucosome sheets. Brunton compass 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417C

Figure 29.

Potassium-feldspar-zone metatexite with sparse, ptygmatically folded leucosome layers. Brunton compass 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417D

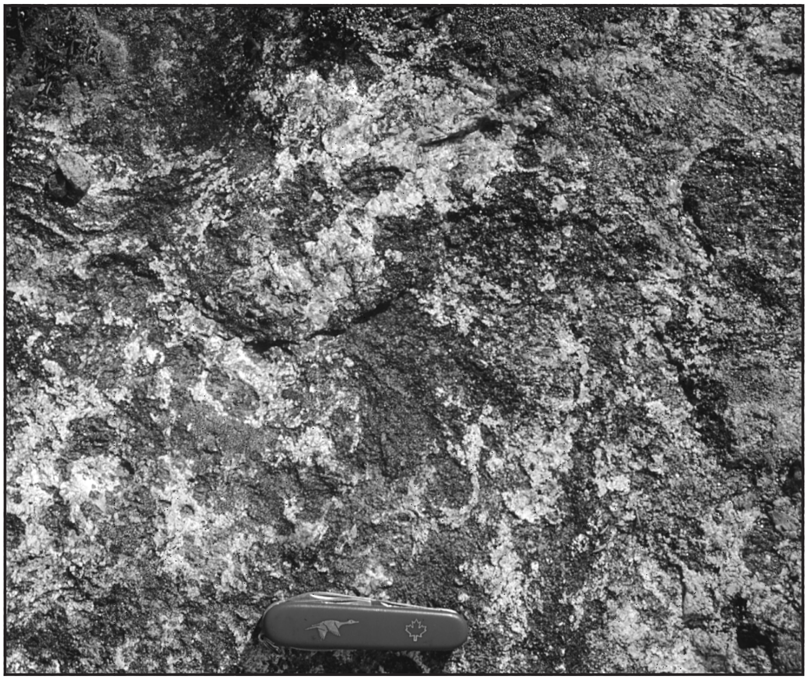


Figure 30.

Garnet-zone metatexite with coarse-grained garnet particularly in leucosome phases. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-417E

in some cases graded, separated by millimetre-scale, biotite-rich and leucosome laminae occur within the metatexite (Fig. 31). Amphibolite layers are a relatively rare occurrence.

Diatexite

At least two types of diatexite occur within the higher grade metasedimentary rocks. Commonly associated with the metatexite are a few to many metre-scale units of what appears to be a homogenized equivalent of the metatexite. The brownish rock is massive to moderately foliated, homogeneous to less commonly vaguely layered, and inequigranular, consisting primarily of about 5 mm subhedral laths of plagioclase in a somewhat finer grained, mainly quartz- and biotite-rich matrix with lesser amounts of the minerals found in the normal melanosome. Randomly and rather widely dispersed through the rock are coarse crystals, up to 10 cm, of mainly feldspar and cordierite, and less commonly sillimanite and garnet depending on metamorphic conditions (Fig. 32). No particular pattern of occurrence of these units were recognized nor were individual bodies of this material mapped out. They are most common in the K-feldspar zone, but also occur

Figure 31.

Sequence of psammitic beds in garnet zone with thin pelitic layers to partings and leucosome laminae. Primary grain-size gradation is preserved in some of the more psammitic layers. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416M

Figure 32.

Potassium-feldspar-zone melanocratic diatexite with coarse cordierite porphyroclast. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416N

in the other zones. In some cases, they appear to be concordant and alternate with metatexite units on the scale of a few tens of metres, possibly as thick sills within the metatexite sequence. In one instance the homogeneous migmatite appeared to be intrusive into a metatexite sequence with a high angle of discordance (Fig. 33). These units appear to represent the collapsed, homogenized, and mobilized equivalents of metatexite similar to that with which they are associated. They may form when the proportion of leucosome to melanosome exceeds some limit such that the metatexite no longer has any structural integrity and collapses. The mass then represents a mechanical mixture of the generally finer grained melanosome and coarser grained leucosome phases and could have behaved as a mobile crystal mush intruding the intact metatexite in thick sills or crosscutting bodies.

The second type of diatexite is a much more leucocratic rock. The commonly white- to buff-weathering rock is similarly medium grained and inequigranular, massive to weakly foliated, but sometimes with vague layering due to subtle textural and compositional variations (Fig. 34, 35). It has a distinctive texture consisting of blocky subhedral plagioclase surrounded by clotty to wispy aggregates of fine-grained biotite that is greatly reduced in amount over the previously described diatexite. Depending on grade, it can also contain cordierite, garnet, orthopyroxene, and K-feldspar that are commonly, but not always coarser grained. The presence of garnet and cordierite serves to distinguish these diatexitic rocks from texturally similar granodioritic intrusions. In several instances the two appear to be gradational. A good example of this occurs with the map-scale granodiorite (unit Agd) intrusions in the K-feldspar zone. In at least one transect, one granodiorite contact is distinguished only by the disappearance of cordierite from the marginal diatexite, whereas the other contact is sharply defined against metatexite. A similar relationship can be seen with the large megacrystic granite (unit Agm.o) in the southern part of the area. In several places, as the contact with the granite is approached, the size and number of K-feldspar porphyroblasts in the diatexite increase to the mapped contact where the now dominant K-feldspar is pink, cordierite is gone, and garnet is reduced to fine grains locally in the matrix. As in the other high-grade rocks, minor, fine-grained psammite inclusions occur locally

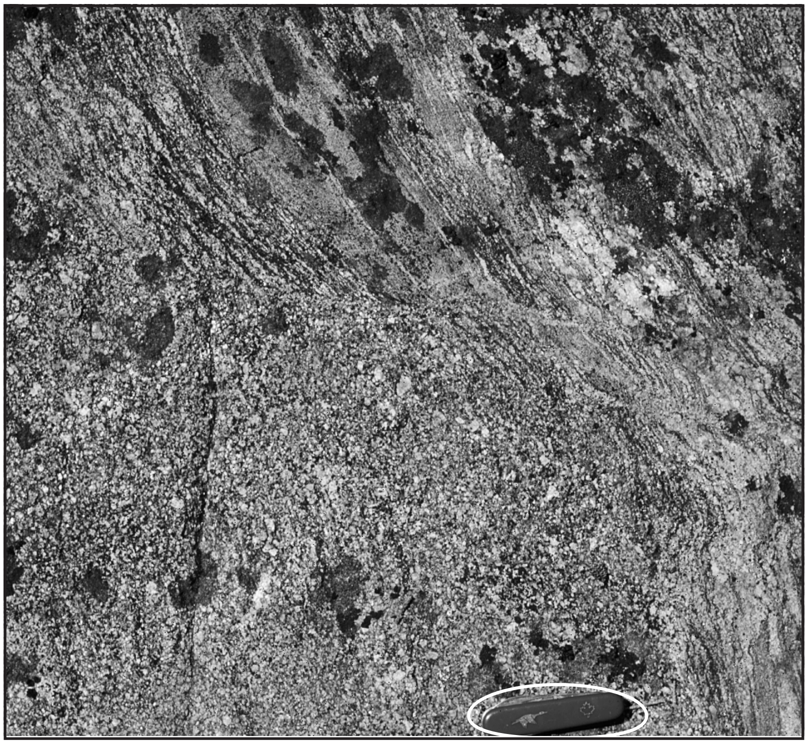


Figure 33.

Mobilized homogeneous melanocratic diatexite in intrusive contact with thinly layered metatexite. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416O

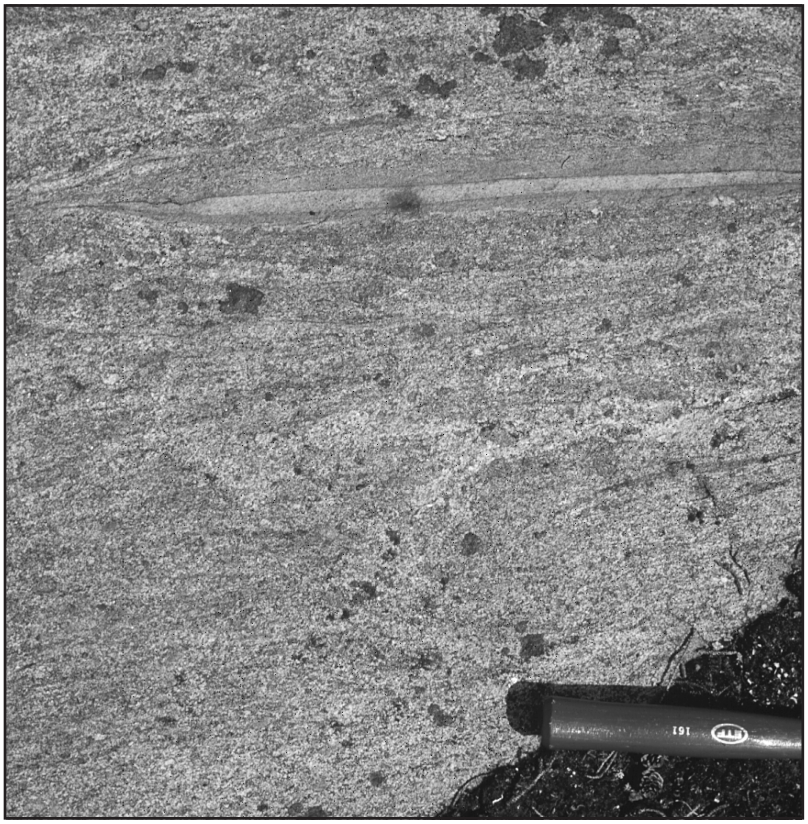


Figure 34.

Diffusely layered, leucocratic, garnet-zone diatexite in which coarsest minerals are garnet and cordierite. Elongate inclusion represents a refractory psammite layer. Hammer handle 3.5 cm wide. Photograph by J.B. Henderson. GSC 2002-416P

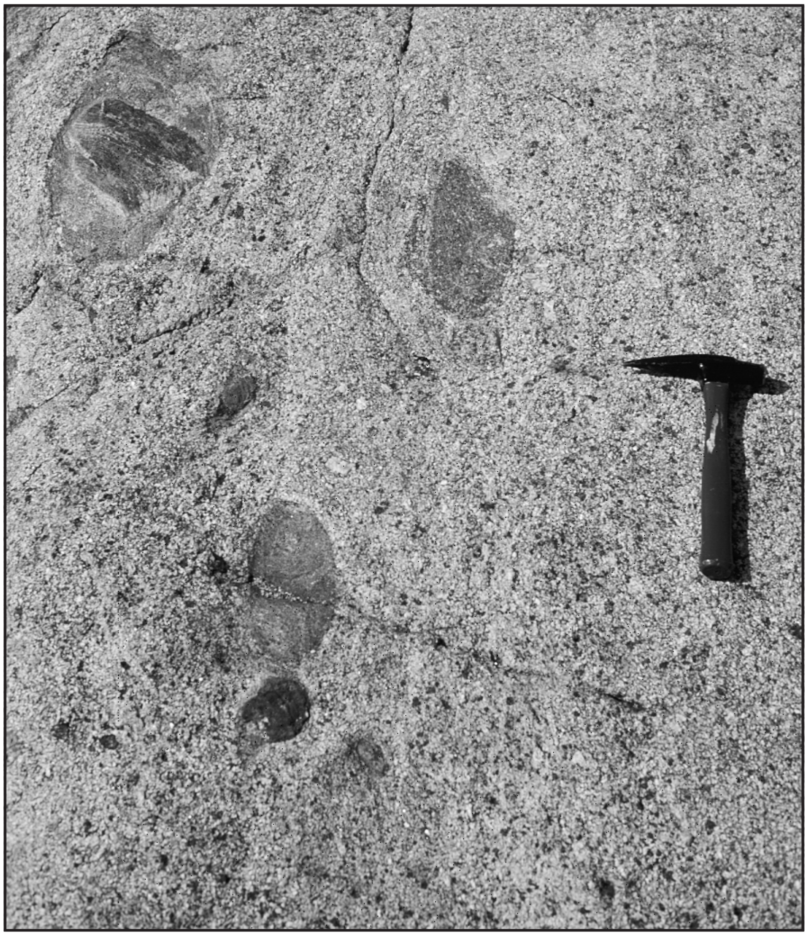


Figure 35.

Homogeneous garnet-cordierite diatexite with rounded inclusions of psammite. Photograph by J.B. Henderson. GSC 2002-416Q

and more rarely amphibolite inclusions are present. This leucocratic type of diatexite is most common in the garnet zone where the reduced biotite content suggests a greater degree of melting as would be expected in the higher grade rocks. They are also present in the two lower grade, migmatite zones (units ĀYs.m and ĀYs.mk) where they may represent mobilized concentrations of leucosome, particularly in the case of the lower grade zone.

The large area of high-grade metasedimentary rocks in the southern part of the Dauphinee domain are similar in many respects to the metasedimentary rocks to the west. Both metatexite and diatexite are present and mineral assemblages include K-feldspar and abundant garnet; one difference is the apparent lack of cordierite noted both in outcrop and the few thin sections examined. While megascopically similar to the rocks to the west, in thin section the garnet crystals in these rocks are highly fractured and partially altered to chlorite, and epidote is a common part of the mineral assemblage. This presumably is related to alteration associated with the domain-bounding, cataclastic shear zone.

The relative paucity of psammitic material in these high-grade metasedimentary rocks, occurring only locally as minor inclusions, suggests these rocks were originally similar to the lower grade, pelite-dominant, metasedimentary rocks in the northwestern part of the map area. The otherwise unexplained, apparently random distribution of metatexite versus diatexite perhaps may be in part explained by the more abundant occurrence of somewhat more refractory, but now recrystallized beyond recognition, siltstone in the protolith of what is now preserved as metatexite.

ARCHEAN INTRUSIVE ROCKS POSTDATING YELLOWKNIFE SUPERGROUP

Approximately half the map area is underlain by intrusive rocks that postdate the Yellowknife Supergroup. In the sections that follow, these rocks are discussed mainly in terms of three compositional groupings that is also to some extent reflected in their age. The oldest dated intrusive rock is a group, intermediate in composition, consisting of trondhjemite, tonalite, and quartz diorite with some granodiorite that occur both in the Ghost and Hinscliffe domains. They range from 2655 Ma to 2605 Ma. The youngest group consist of several suites of granite with ages younger than 2600 Ma. Also present is a group of largely mafic intrusions that have not yet been dated radiometrically, that predate the young granitoid suite, and either postdate or are synchronous with the various members of the intermediate composition suite. These include a major metagabbroic to anorthositic gabbro suite in the Wijinnedi domain, amphibolitic sheets in the Ghost domain, and small metapyroxenite intrusions in both the Wijinnedi and Hinscliffe domains. In addition to these groups, there are two units on the map that consist of undivided granitoid rocks of either the Ghost or Dauphinee domains. They consist of intrusions that include members of the previously outlined groups as well as other texturally and/or compositionally distinct bodies on a scale too fine to reliably outline at the present scale of mapping.

Intermediate intrusions

Major intermediate intrusive units occur in all but the Wijinnedi domain. The Hinscliffe domain consists almost entirely of trondhjemitic rocks, whereas tonalite and related tonalite gneiss units are the major component of the Ghost domain. The Ghost domain also contains granodiorite as a relatively minor phase, as well as a variety of other intermediate intrusions too small or complex to be outlined at the present scale of mapping that are among the components of an undivided unit. Large bodies of quartz diorite occur both in the Ghost domain and the Dauphinee domain.

Hinscliffe complex (unit ĀH)

The Hinscliffe complex consists entirely of igneous rocks and their deformed and metamorphosed equivalents. The complex forms the Hinscliffe domain, which is a wedge-shaped body between the Wijinnedi and Ghost domains that is separated from them by ductile shear zones (Fig. 1; Map 2023A). The domain marks the transition from the steeper dipping, easterly structural trends characteristic of the lower grade Wijinnedi domain to the north and the generally more moderately dipping and more varied northeasterly trends of the higher grade Ghost domain to the south.

The only geochronological data on the complex comes from the south shore of 'Hinscliffe Lake' in the central part of the domain where the rocks are better preserved and original intrusive relations can be seen. The rock dated, a trondhjemite, is the oldest of three phases recognized in the outcrop area. Zircon from this rock has an age of 2654 ± 4 Ma, based on a weighted average of ^{207}Pb/^{206}Pb ages of three near-concordant zircon fractions and was interpreted as the age of crystallization of the rock (Villeneuve and Henderson, 1998; Appendix). A fourth fraction has a near-concordant age of 2665 ± 12 Ma which is similar, within error, to the more precise 2673.3 ± 1.4 Ma age of a Yellowknife Supergroup metadacite from the Wijinnedi domain to the north (Villeneuve and Henderson, 1998; Appendix). A titanite fraction from the same rock has a ^{207}Pb/^{206}Pb age of 2610 ± 4 Ma which is similar to the U-Pb (zircon) ages of two of the dated intermediate intrusions in the area (2605 ± 3 Ma for a quartz diorite, unit Āqd; ca. 2605 Ma for a granodioritic phase in a tonalite gneiss, unit Ātgn). This date was interpreted by Villeneuve and Henderson (1998) (Appendix) as an estimate of the time of cooling of the rock through the approximately 600°C closure temperature of titanite (Heaman and Parrish, 1991).

Contacts

Both northern and southern boundaries of the domain are ductile shear zones. The northern shear zone is in excess of a kilometre wide at its western end and narrows somewhat toward the east. The contact itself, between the leucocratic mylonite of the Hinscliffe domain and the black, similarly finely laminated amphibolite, derived in large part from the mafic volcanic rocks of the Wijinnedi domain, is exposed locally and is knife-sharp (Fig. 36). The intensity of the foliation in the Hinscliffe mylonite units decreases to the south over a distance of several hundred metres to a second prominent shear zone, across which the granitoid rocks are significantly less deformed. The mafic minerals in the shear zone consist mainly of hornblende with few, if any phases suggesting retrogression. This would indicate the rocks were deformed under amphibolite-grade conditions.

Figure 36.

Exposed contact between mylonitic trondhjemite gneiss of the Hinscliffe domain to the south and finely foliated mylonitic amphibolite of the Yellowknife Supergroup of the Wijinnedi domain to the north. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416R

The southern-bounding shear zone is much narrower with the rocks of the Hinscliffe domain becoming increasingly foliated toward the southern boundary, but over a distance of only tens of metres. The actual contact was nowhere seen due to coverage by a 10 m or more wide peat bog between Hinscliffe domain rocks and the mainly migmatitic metasedimentary rocks of the Ghost domain to the south. Given that the actual contact zone between the Hinscliffe and Ghost domains is recessive and not seen, it is conceivable that the covered contact zone could represent reactivation of part of the shear zone under greenschist-grade conditions or by brittle deformation. In the bordering rocks that are exposed, however, no retrogressive gradient was recognized that might support this possibility.

Lithology

The bulk of the complex consists of trondhjemite, or more precisely, its deformed and metamorphosed equivalents. Where best preserved in the core of the complex, irregular primary intrusive relationships can be seen between the various phases of trondhjemite that differ slightly in composition and texture. These variations are commonly apparent on an outcrop scale with the presence of three or more phases not uncommon. The degree of deformation increases toward the margins of the complex with most parts of it at least foliated and much of it gneissic (Fig. 37). The structural transposition of the slightly compositionally varied, complexly interrelated, phases of the trondhjemite in the more deformed parts of the complex result in the typically subtle aspect of the resulting gneiss units that are layered on a scale of 1 cm to a few tens of centimetres. Deformation is more intense along the northern margin as the rocks change from the subtly layered gneiss through porphyroclastic gneiss to finely laminated, fine-grained mylonite.

The trondhjemite is a light-grey- to pinkish-grey- to less commonly buff-weathering rock and is massive to foliated due to the orientation of sparse biotite and quartz elongation. It is typically fine grained and most rocks have a sugary aspect due to recrystallization of the original igneous mineral phases. This recrystallization gives the rock in outcrop an almost translucent look (Fig. 38). Where better preserved, the originally coarser grained nature of the rock is evident through the distribution of the sparse biotite around the original coarser plagioclase grains.

The trondhjemite consists of anhedral, equidimensional, weakly to nontwinned plagioclase with, in some cases, minor K-feldspar exsolution, almost as much amoeboid, to elongate quartz, and scattered fine-grained biotite flakes that make up usually less than 10%, commonly less than 5% of the rock (Fig. 39). Minor, interstitial microcline is locally present and can be accompanied by myrmekite. With increasing microcline or biotite content, some phases of the complex grade into the granodiorite or tonalite fields, respectively (Fig. 40). The coarsest grain size is less than 2 mm, commonly less than 1.5 mm. Metamorphic epidote, sometimes overgrowing allanite, is everywhere present, commonly associated with biotite, which it sometimes almost approaches in abundance. Titanite is abundant in the trondhjemite sampled for geochronology (Villeneuve and Henderson, 1998; Appendix), but is not common elsewhere.

Intrusions

The trondhjemite bodies are commonly intruded by a series of decimetre-scale, discordant dykes of aplitic granite (Fig. 38). Many of these postdate the principal foliation, but are finely recrystallized in a similar style to that of the trondhjemite. In the northern domain-bounding shear zone, minor, strongly deformed, slightly discordant, felsite layers occur on both sides of the contact, suggesting that at least some granitoid activity in the complex was synchronous with

Figure 37.

Trondhjemitic gneiss in the north-central part of the Hinscliffe complex with strongly deformed amphibolite inclusions. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416S

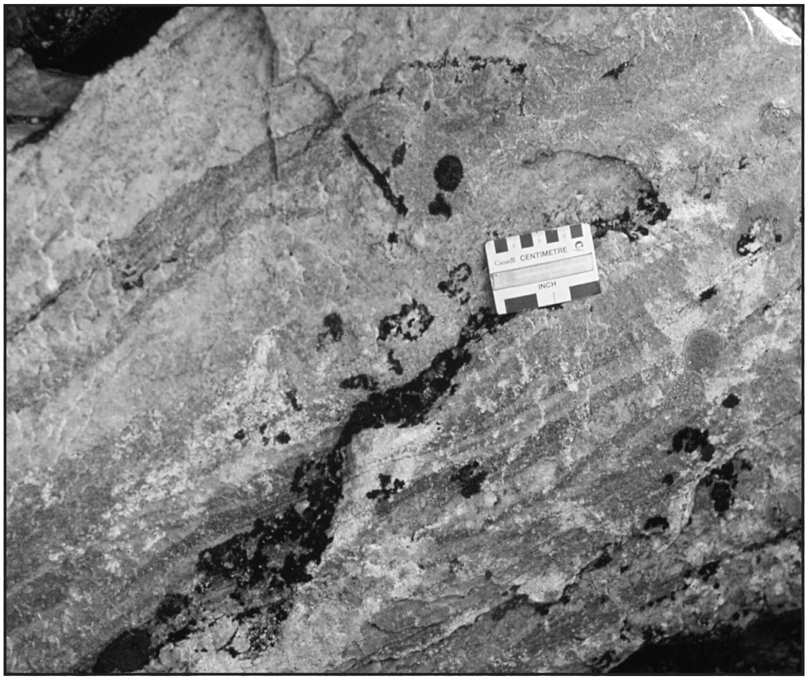


Figure 38.

Hinscliffe trondhjemite with aplitic dykes, veins, and pegmatite that are common, if not this abundant, throughout the complex. Both the trondhjemite and the aplitic phases that intrude it are strongly recrystallized to a fine grain size giving the rock a translucent aspect and making the contact relations seem slightly out of focus. Photograph by S.E. Schaan. GSC 2002-417F

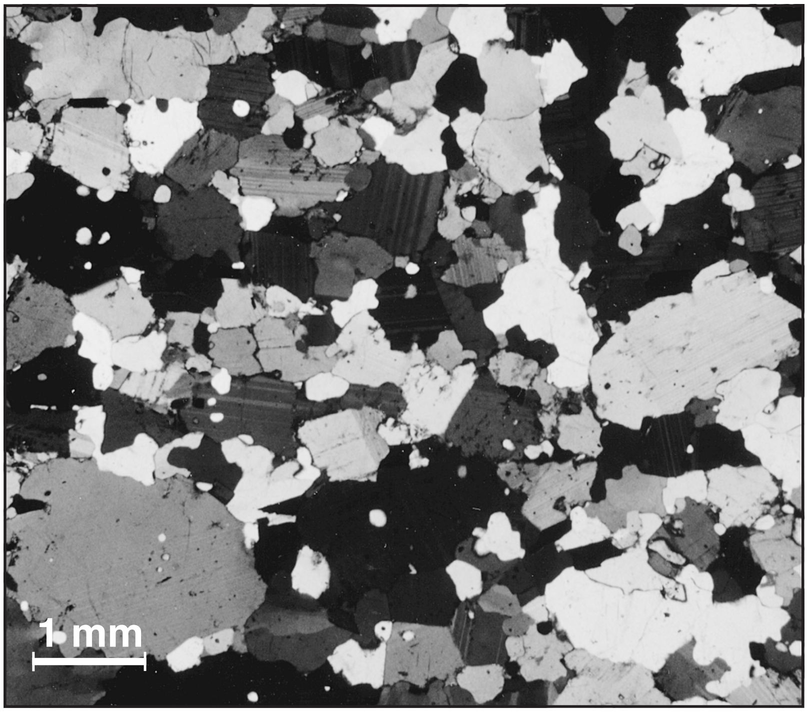


Figure 39.

Photomicrograph of representative Hinscliffe complex trondhjemite with varied grain size consisting of anhedral, more or less equidimensional, moderately to faintly twinned oligoclase with minor quartz inclusions and K-feldspar exsolution in some cases and with some of the larger grains weakly zoned. Quartz occurs in both large, multidomainal, amoeboid masses and fine spheres, and biotite in scattered isolated flakes. Anhedral to euhedral metamorphic epidote is commonly associated with the biotite.

the bounding shear zone. Pegmatite is also present, most commonly as disaggregated trains of coarse-grained, pink K-feldspar parallel to the foliation or gneissosity, but also as more intact, discordant bodies in the less deformed regions (Fig. 38, 41). They, like the trondhjemite and aplite are also recrystallized, but in many cases only the outer parts of the coarse K-feldspar crystals are recrystallized.

At the northern margin of the complex, in particular between the contact and the shear zone parallel to it to the south, is a series of deformed gabbro to diorite intrusions that are now foliation parallel (Fig. 42). The proportion of the mafic intrusions is highly varied and is locally up to 70%. The mafic intrusions are, on the whole, less deformed than the granitoid rocks in which they are emplaced. They are themselves intruded by younger granitoid dykes, all of which become increasingly transposed as the northern contact is approached. The resulting rock unit has a much more prominent compositional layering relative to the more subtle gneiss of the complex proper.

A somewhat similar heterogeneous assemblage of mafic to intermediate gneiss units occurs at the southern margin of the complex, although there they occur mainly as inclusions

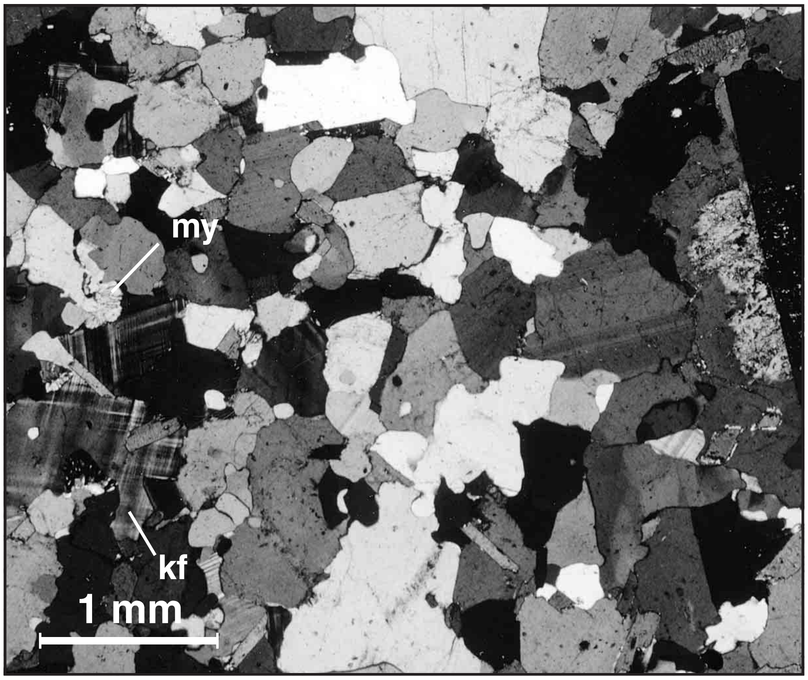


Figure 40.

Photomicrograph of Hinscliffe complex trondhjemite with some interstitial K-feldspar (kf) and associated myrmekite (my), generally considered to be diagnostic of deformation of the rock (Simpson and Wintsch, 1989). Some of coarser plagioclase grains contain more altered calcic cores.

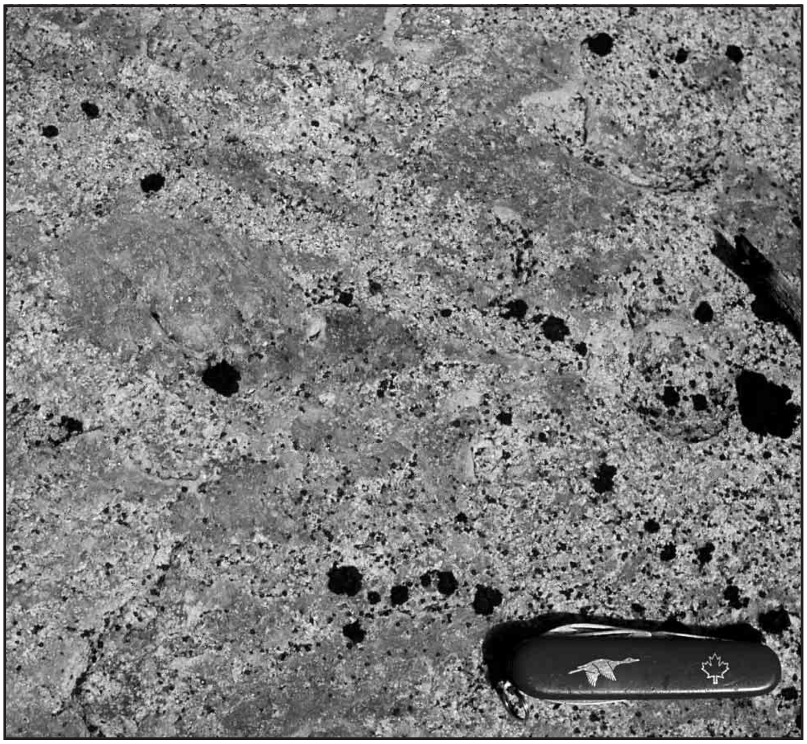


Figure 41.

Disaggregated pegmatite vein in Hinscliffe complex trondhjemite. As in most of the complex, the original textures are lost due to recrystallization, but the process affects only the margins of the coarse K-feldspar crystals of the pegmatite. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416T

within less deformed, recrystallized trondhjemitic intrusions. These vary from what were originally straight, recrystallized, mylonitic, amphibolitic gneiss near the boundary that are now folded (Fig. 43), to more diffusely layered, compositionally heterogeneous gneiss in the northern part of the boundary zone (Fig. 44). The similarity of these more mafic bodies with those to the north, together with their symmetric distribution at the margins of the complex, suggest the complex as a whole may occur in an antiformal structure.

At least two small metamorphosed diatremes intrude the complex (Map 2023A). They contain clasts of a type of granite not recognized in the complex. They are discussed more completely in the section ‘Diatremes’.

The complex also contains minor undeformed intrusions on the scale of a few tens of metres of massive, medium-grained, pink biotite granite and dykes of minor, locally magnetite-bearing, undeformed pegmatite (Fig. 45). These small intrusions are lithologically most similar to the large body of

Figure 42.

Mylonitic trondhjemite gneiss with less deformed, dioritic sills in north border shear zone of the Hinscliffe complex about 200 m south of the contact. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416U

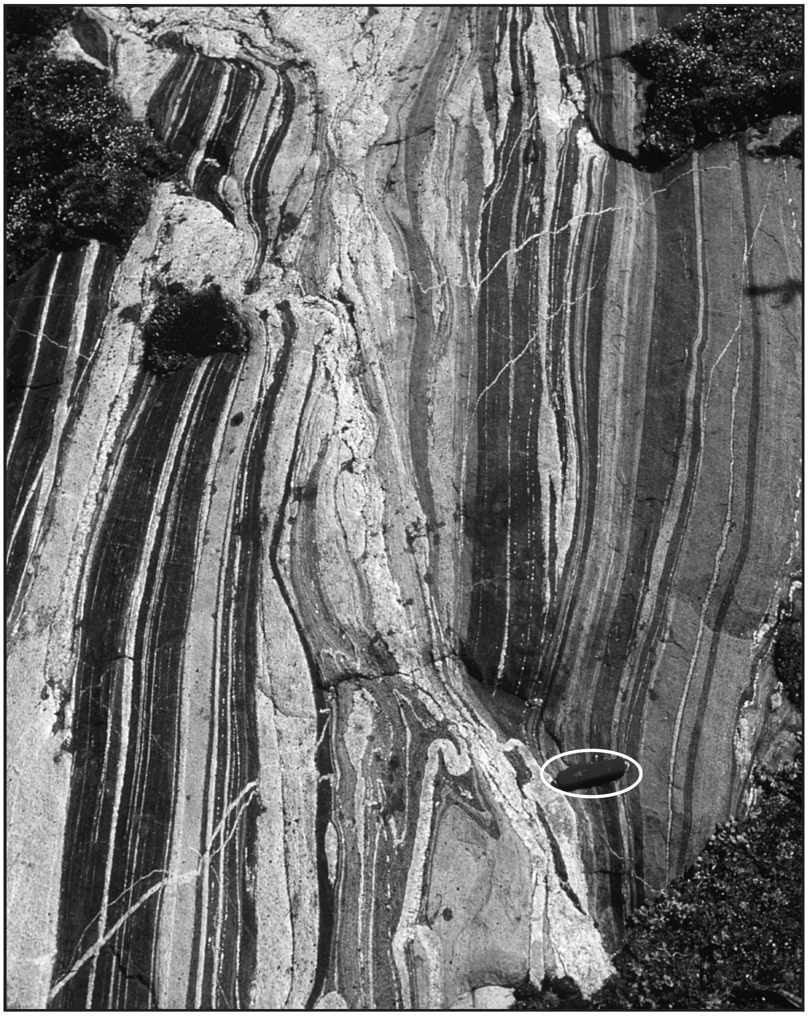


Figure 43.

Highly deformed inclusions of heterogeneous intermediate gneiss to amphibolite contained in less deformed trondhjemite within a few tens of metres of the southern margin of the Hinscliffe domain. Note the now bent, parallel-sided nature of the layering suggesting the gneiss was mylonitic. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416V

Figure 44.

Extended granodioritic to tonalitic gneiss inclusion in less deformed Hinscliffe complex trondhjemite in the southern border zone. Part of an amphibolite inclusion occurs at the top of the photograph. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416W

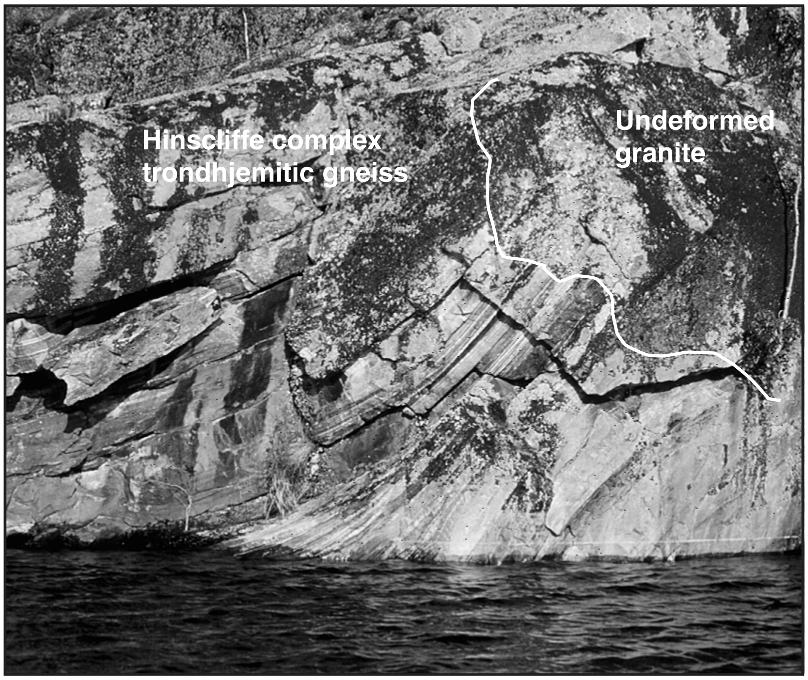


Figure 45.

Interlayered trondhjemite and mafic mylonitic gneiss near the north margin of the Hinscliffe complex on the Ghost River are intruded by small, undeformed intrusions of granite (contact outlined by white line). Photograph by J.B. Henderson. GSC 2002-416X

granite (unit Ag, *see* section 'Granite and pegmatite (units Ag, Ag.o)') in the Ghost domain east of the Hinscliffe complex and its associated small bodies of pegmatite and granite, one of which has been dated at ca. 2595 Ma, that occur locally nearby.

In addition to these more felsic intrusions, a few occurrences of ultramafic rocks are known with the complex. The metapyroxenite body on 'Hinscliffe Lake' (unit Aum, described below) is the largest of these.

Mafic inclusions

A common feature of the trondhjemite complex is the occurrence of black amphibolite inclusions. They are found throughout, but vary greatly in their proportion from none at all, to isolated enclaves, to concentrated zones in which the proportion of amphibolite exceeds the trondhjemite (Fig. 46). The zones of concentration are a few metres to tens of metres wide and are separated by inclusion-free zones of trondhjemite. Where best preserved they occur as angular blocks, a few tens of centimetres to several metres in size. They are intruded by several generations of variably deformed leucocratic granitoid dykes and veins similar to those within the trondhjemite (Fig. 47). The inclusions are on the whole quite uniform in composition and have a medium-grained, even-grained texture. An exception to this is a zone of inclusions a few metres wide that contained distinctive, relict, 1 cm phenocrysts of plagioclase. These were traced over a distance of 150 m and are considered strong evidence that these enclaves represent disrupted mafic dykes.

Figure 46.

Swarm of angular amphibolite inclusions from the central part of the Hinscliffe complex. In the minimally deformed parts of the complex such as this, linear trains of concentrated inclusions are separated by zones with few, if any inclusions. They have been interpreted as synplutonic dykes both intruded into, and intruded by, the still hot and mobile crystallizing trondhjemite. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416Y

Figure 47.

Amphibolite inclusion in Hinscliffe complex trondhjemite. Note the multiple generations of intrusive granitoid veins in the inclusion. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416Z

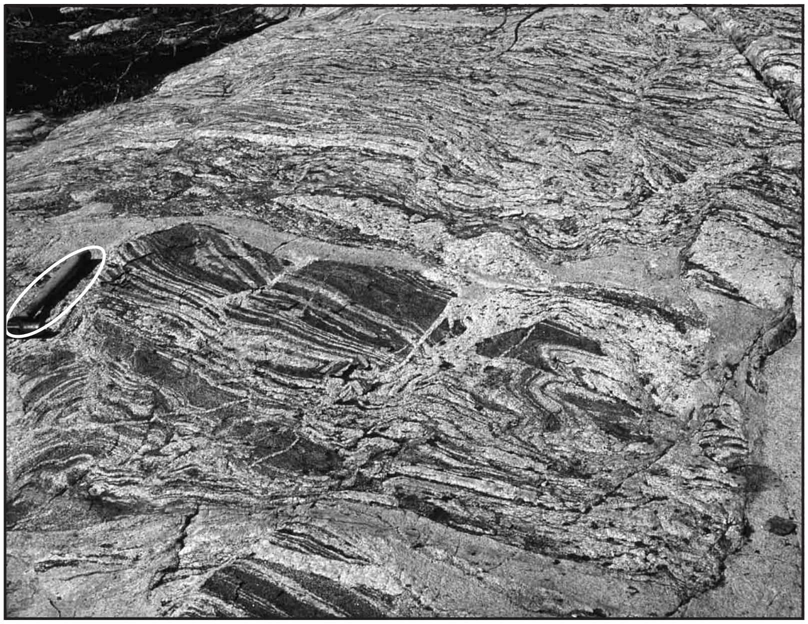


Figure 48.

Partially assimilated amphibolitic inclusion due to the injection of fine veinlets of trondhjemite introduced at a centimetre scale. At its most extreme, as in the top part of the photograph, the rock looks not unlike some of the pelitic metasedimentary migmatite bodies of the Ghost domain except that the dominant mafic mineral here is hornblende as opposed to biotite. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416AA

Most of the inclusions, like the trondhjemite they are contained in, are deformed to some degree, are elongate to elliptical in outline, and become foliated to gneissic with the transposition of the original crosscutting granitoid dykes and veins (Fig. 37). In the southwestern part of the complex some of the inclusions are closely injected with granitoid veins giving the rock a migmatitic aspect not unlike that seen in the migmatitic metasedimentary rocks of the Ghost domain to the south (Fig. 48). Rocks such as these, then, may represent another path to the mafic to intermediate gneiss units that occur along the southern boundary of the complex.

These mafic inclusion swarms are thought to represent synplutonic mafic dykes in the sense of Roddick and Armstrong (1959) and Pitcher (1991) among others, in that the best preserved examples have apparent conflicting age relations. While these arrays of inclusions have a dyke-like form, they are at the same time intruded by the trondhjemite in which they are emplaced. The simplest explanation is that the still hot, partially crystallized trondhjemite was capable of being fractured during a short-term deformational event, perhaps seismic shocks. This would allow the emplacement of mafic intrusions that would quickly chill and crystallize within the relatively cooler, still crystallizing trondhjemite. Subsequent deformation in the still-hot, viscous trondhjemite could result in fracturing of the dyke, now cooled several hundreds of degrees below its melting point, and flow of the still-mobile trondhjemite around the pieces (Pitcher, 1991). Their presence is clear evidence of the contemporaneity of felsic and mafic magmatism.

Discussion

The Hinscliffe complex is a multiple intrusive complex with most of its components falling within the trondhjemite compositional range. As such, it differs from most of the other plutonic bodies within the map area, which are, for the most part, simple and compositionally homogeneous, with the granitic phases at least being significantly younger. The intrusive complex sits structurally adjacent to the compositionally similar, volcanic dome of the Wijinnedi domain and it is perhaps of interest to speculate on a possible relationship. On the basis of the single U-Pb (zircon) date available from each, such a relationship might be considered if not proven (Villeneuve and Henderson, 1998). The Hinscliffe complex sample dated at 2654 ± 4 Ma is about 20 million years younger than the Yellowknife Supergroup sample which has an age of 2673.3 ± 1.4 Ma. The trondhjemite sample, however, does contain a zircon population with a ^{207}Pb/^{206}Pb age of 2665 ± 12 Ma which is within error of the dacite age. The Hinscliffe complex could have behaved as a long-lived, open magmatic system due to the periodic introduction of mafic magma and, most importantly, its accompanying heat into the unconsolidated silicic magma chamber, thereby extended its life possibly as much as the time implied by the geochronological data (Macdonald and Smith, 1988; Koyaguchi and Kaneko, 2001). That this may have happened is supported by the abundance of mafic synplutonic dykes in the Hinscliffe complex and an example of magma-mixing textures seen in a mafic dyke emplaced in a synvolcanic felsic intrusion at one locality in the volcanic complex. The trondhjemite sample dated is also almost within error of the 2647 ± 2 Ma age of a rhyolite breccia unit (that also contains 2.70–2.66 Ga xenocrystic zircon) that occurs within thick-bedded, coarse-grained turbidite sequences 30 km north of the area on Indin Lake (Pehrsson and Villeneuve, 1999).

Other potential links include the same mafic synplutonic dykes of the Hinscliffe complex and the mafic volcanic rocks of the Wijinnedi domain that, for the most part, occur stratigraphically above the felsic volcanic rocks and toward the east end of the dome (Fig. 1; Map 2023A). A perhaps more likely link, given the geochronological data, might be the suite of metagabbro plutons and sills that occur in particular abundance in the central and eastern part of the volcanic complex. These metagabbroic rocks are discussed in the

'Metagabbro to meta-anorthositic gabbro (unit Amg)' section where it is suggested that there may be a fairly close temporal relationship between metagabbro intrusions and the volcanic system.

The closest temporal link with other plutonic rocks would be with the tonalite–tonalite gneiss suite (units At.o, Atgn.o) that dominate the Ghost domain. These compositionally somewhat similar rocks that appear to have crystallized under granulite-grade conditions between 2.64 Ga and 2.63 Ga (M. Villeneuve, pers. comm., November 2000; Appendix) are on the order of 20 Ma younger than the Hinscliffe trondhjemite. The granite intrusions, on the other hand, are some 60 Ma younger than the Hinscliffe complex rocks.

Tonalite (units At, At.o, Atgn, Atgn.o, At-gm, At-gm.o)

The areally most important rock type of the Ghost domain is tonalite. On the map and in the discussion that follows the tonalitic rocks have been divided into several units: orthopyroxene-bearing tonalite thought to have crystallized under granulite-grade conditions (unit At.o); a closely related, but structurally and lithologically more complex orthopyroxene-bearing tonalitic gneiss (unit Atgn.o); and a unit that variably grades along strike from orthopyroxene-bearing tonalite to orthopyroxene-bearing megacrystic granite (unit At-gm.o). The equivalents of all three, but without orthopyroxene, occur in the northwestern part of the Ghost domain where they either crystallized under lower grade conditions or have been retrograded so that orthopyroxene is no longer present (units At, Atgn, At-gm).

Tonalite

Tonalite without orthopyroxene (unit At) occurs in two bodies, the larger northwest of Ghost Lake and a small body in the Wijinnedi domain, southeast of the main accumulation of mafic volcanic rocks. The larger body is white weathering, grey, medium to coarse grained, even grained, and massive to weakly foliated. It consists mainly of subhedral plagioclase, no K-feldspar, and a varied content of quartz and mafic minerals, with only a weakly modified igneous texture, and so varies from diorite to tonalite with the latter predominating. The mafic minerals include biotite in all cases, commonly hornblende, and usually rather altered clinopyroxene that in some cases is overgrown by hornblende. It contains inclusions to local zones of metasedimentary migmatite and granitoid gneiss. The tonalite is cut by a series of large, pink pegmatite bodies presumably related to the granite (unit Ag) to the north. The smaller, massive tonalite body to the north is similar in most respects.

Most of the orthopyroxene-bearing tonalite (unit At.o) occurs south, and to a lesser extent, north of the east end of Ghost Lake. It intrudes both metasedimentary and metavolcanic rocks of the Yellowknife Supergroup and is intruded by several major granite bodies and associated dykes, veins, and minor plugs. Its contacts with both older and younger units are sharp and generally concordant. High-angle dykes and veins of tonalite in the adjacent metasedimentary and metavolcanic rocks are not known, but in the vicinity of the contact, sills parallel to it are locally present. Supracrustal inclusions, mainly metasedimentary migmatite to diatexite, are locally common in the tonalite close to contacts with similar rocks, but are particularly abundant southeast of Ghost Lake where they occur as mappable bodies several tens of metres to a few kilometres wide. Rather small amphibolitic gneiss inclusions, lithologically similar to the contiguous volcanic gneiss bodies to the northwest and southeast, are locally abundant on the large peninsula at the east end of Ghost Lake (Fig. 49), in the tonalitic gneiss east of the small river flowing north into Ghost Lake, and southeast and north of the east end of the major belt of mainly metasedimentary rocks in the southeastern part of the area.

Figure 49.

Granulite-grade amphibolitic to amphibolitic gneiss inclusions, thought to be derived from Yellowknife Supergroup metavolcanic rocks, in tonalite. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416BB

The tonalite has the buff- to yellow- to rusty-weathering, greasy yellow to greenish grey colours that are characteristic of many granulite-grade rocks. Where best preserved the rock has igneous textures, is medium grained and equigranular, with subhedral laths of plagioclase, similarly sized, anastomosing quartz commonly consisting of one or two crystallographic domains, and smaller biotite flakes and anhedral orthopyroxene. The best preserved igneous textures (Fig. 50) tend to occur near the intrusive contacts of the tonalite with the supracrustal rocks. Most of the tonalite, particularly in its central parts, is recrystallized to varied degrees (Fig. 51) with the rock generally finer grained and more inequigranular, the plagioclase more anhedral, and the quartz, where more coarsely grained, is more polygonized, but is also present throughout the rock as fine grains. In the better preserved examples of the recrystallized tonalite, the relict, coarser igneous textures are still apparent in outcrop. The tonalite is variably deformed from massive to having a moderately developed foliation defined primarily by biotite concentrations. The strength of the foliation is commonly varied, even at outcrop scale.

Tonalite is by far the most common constituent of the unit, but with varied quartz, K-feldspar, and mafic mineral content, compositions range into the quartz diorite, trondhjemite,

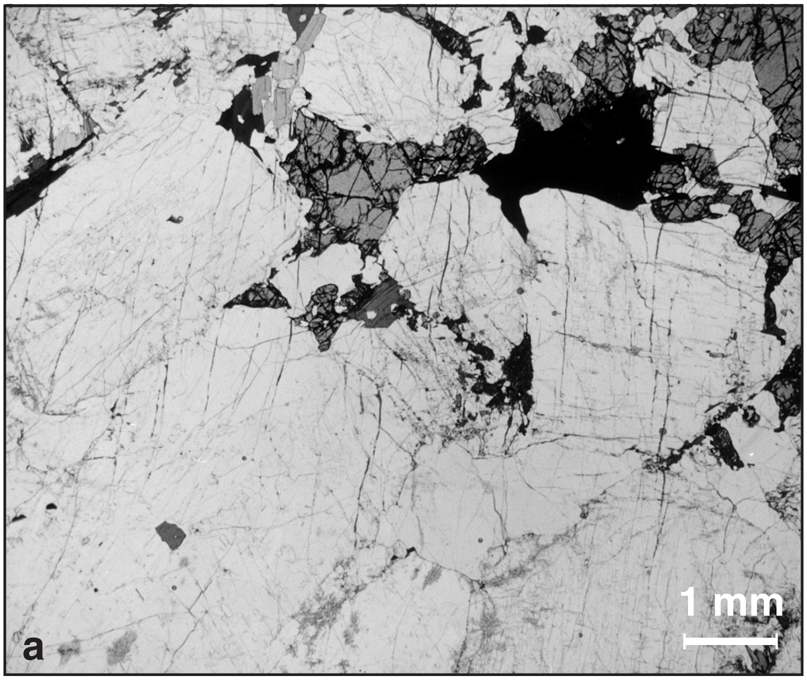


Figure 50.

Photomicrograph of tonalite with well preserved, igneous textures consisting of subhedral plagioclase with K-feldspar exsolution, anhedral lobate quartz masses, anhedral orthopyroxene, and biotite. The tonalite is considered to have crystallized under granulite-grade conditions. ***a****) Plane-polarized light.*

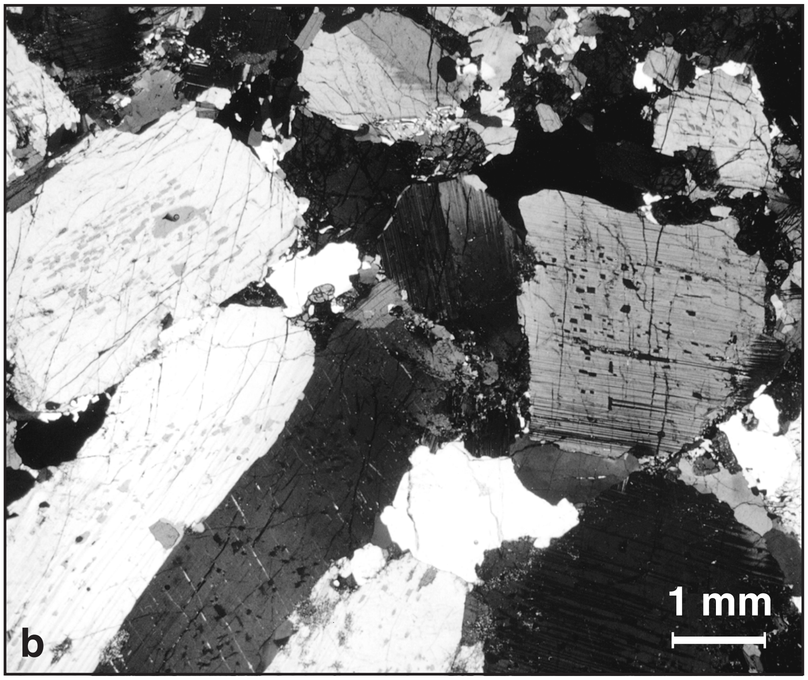


Figure 50. b) *Cross-polarized light.*

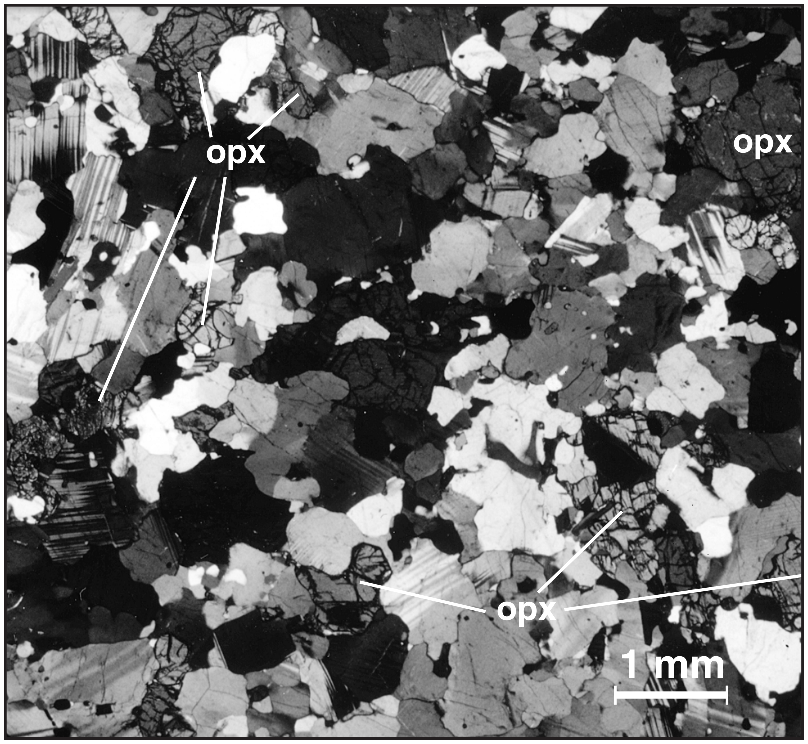


Figure 51.

Photomicrograph of recrystallized, but not particularly foliated, orthopyroxene-bearing tonalite. Cross-polarized light.

and granodiorite fields. Plagioclase is commonly antiperthitic to varied degrees, more so in the rocks with better preserved igneous textures, but also in some of the recrystallized tonalite bodies as well. Potassium feldspar as a separate phase occurs in about one-third of the rocks examined, mainly as a minor phase. The main mafic minerals include orthopyroxene — the dominant mafic phase, biotite, hornblende, and clinopyroxene. Clinopyroxene, a major mafic phase in the tonalite in lower grade rocks, was noted in only a few cases in the granulite-zone tonalite. Hornblende, like the clinopyroxene, was noted in only a few examples where it overgrows orthopyroxene. Biotite occurs in most of the tonalite where it occurs in relatively fine-grained, scattered flakes to aggregates, commonly associated with the orthopyroxene. The anhedral to subhedral, generally fractured orthopyroxene was present in almost all of the tonalite examined. Most are replaced to varied degrees by low-grade alteration products along its margins and fracture surfaces, but in some cases are completely replaced.

Small bodies of K-feldspar-rich rocks are associated with the orthopyroxene-bearing tonalite. They are distinct from the sharply bound, presumably intrusive pegmatite bodies and small granite pods that are a minor occurrence locally throughout much of the domain. The extent of these K-feldspar-rich rocks typically varies from a few tens of metres to a few hundred metres. The rock has a distinct pinkish cast in outcrop and its contact with the K-feldspar-poor to -absent, yellow-weathering tonalite is commonly gradational on a scale of a few metres. They occur throughout the tonalite outcrop area, but they are more abundant on the peninsula at the east end of Ghost Lake. It is speculated that these small K-feldspar-rich bodies in the tonalite, particularly where abundant, is largely a metasomatic addition. These rocks appear to occur structurally above the younger megacrystic granite which might be a good candidate as a source of metasomatic fluids.

Potassium-rich rocks associated with the tonalite are, however, most abundant along the northwesternmost part of the tonalitic outcrop area where a subunit has been outlined (unit Āt-gm.o). There the unit grades back and forth along its trend from tonalite to a megacrystic granite, with moderately abundant coarse megacrysts in excess of 4 cm (Fig. 52, 53), which is not unlike the major units of megacrystic granite at the north and south boundaries of the area.

Tonalite gneiss

Closely related to the above-described tonalite units is the tonalite gneiss subunit (unit Āt-gn). It differs from the tonalite in that it tends to be more strongly foliated and commonly has a gneissic aspect due to minor variations in texture and composition. This is enhanced by the local occurrence of other, more granitic phases commonly present as thin, concordant sheets. To the northwest, the tonalitic gneiss is essentially gradational with the undivided granitoid rock (unit ĀG) as the tonalite becomes less dominant. Amphibolite to amphibolite gneiss, with centimetre-scale internal layering defined by variations in plagioclase and mafic minerals, is also a common occurrence within the gneiss subunit. These amphibolitic layers are considered to be correlative with the map-scale amphibolite (unit Āam) that occurs within the tonalite gneiss. Similar, thin amphibolite sheets also occur within the undivided granitoid rocks (unit ĀG) on the north side of Ghost Lake. Where present within the tonalite gneiss they vary in proportion up to 50%, but typically constitute less than 10%, occurring in layers a few centimetres to tens of centimetres thick. Layers to irregular inclusions of metasedimentary migmatite (Fig. 54) are a locally present, minor component particularly in the vicinity of larger, map-scale units of metasedimentary migmatite.

Although it is clear that the rocks of the tonalite gneiss unit formed under or were exposed to granulite-grade metamorphic conditions, large parts of the unit are retrograded. This is particularly true along the south shore of Ghost Lake and in particular in the rocks over the culmination of the antiform cored by the large megacrystic granite body. Granulite-grade mineralogy is well preserved in the gneiss within a few hundred metres to 1 km of the intrusion. The retrogression is low grade with the orthopyroxene replaced with a yellow chloritic material. It is suspected that this retrogression is related to the cryptic shear zone that is believed to underlie the length of most of Ghost Lake. Similar alteration occurs in the vicinity of the less cryptic, but still poorly exposed shear zone within the gneiss units east of the megacrystic granite and is also

Figure 52.

Orthopyroxene tonalite containing sparse K-feldspar megacrysts. Brunton compass 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417G

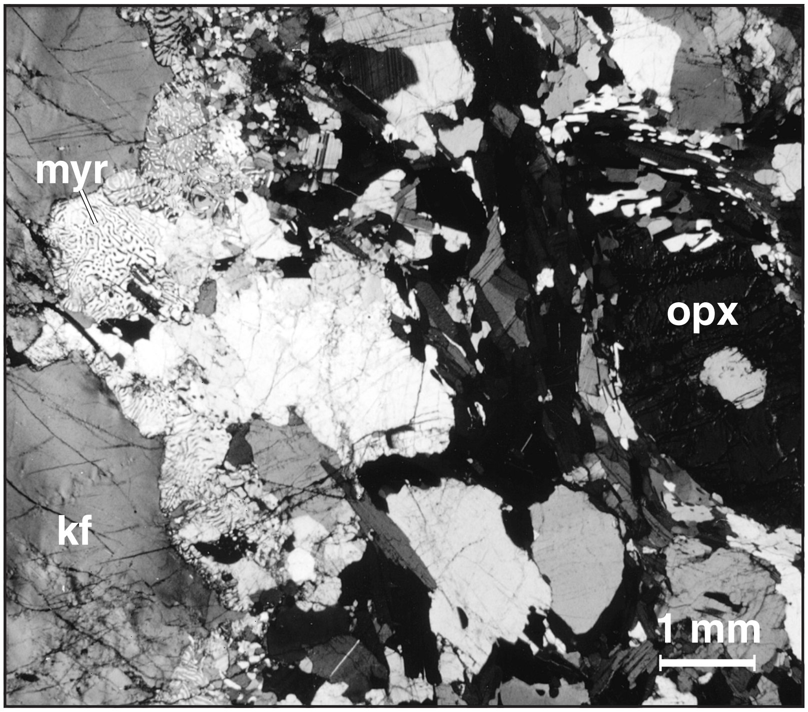


Figure 53.

Photomicrograph of same tonalite shown in Figure 52. Matrix of rock is typical of igneous-textured, orthopyroxene-bearing (opx) tonalite seen throughout the unit, but contains coarse microcline (kf) megacrysts, part of one of which appears on the left, here surrounded by myrmekite (myr). Cross-polarized light.

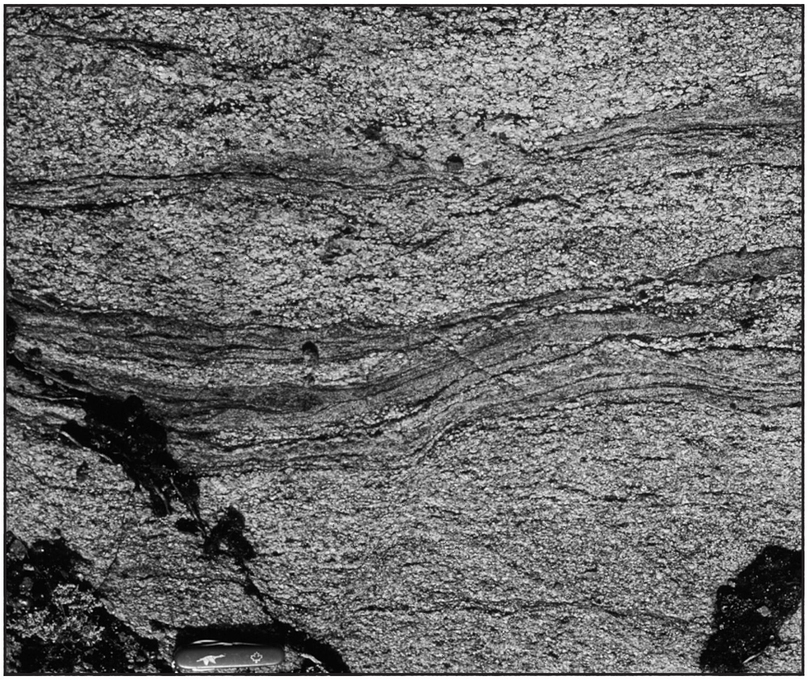


Figure 54.

Tonalite gneiss with garnet-K-feldspar-cordierite metasedimentary migmatite. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416CC

seen in the tonalite units in the immediate vicinity of the cataclastic shear zone that marks the eastern margin of the Ghost domain.

Pressure-temperature estimates on rocks associated with the tonalite units in this region suggest that they equilibrated between 5.5–7.4 kbar and 827–920°C (Chacko et al., 1995a; Farquhar and Chacko, 1996). This is discussed further in the 'Metamorphic geology' section.

Geochronology

Zircons from a tonalite layer within the tonalite gneiss (unit At-gn.o) on the north shore of Ghost Lake near its eastern end suggest a crystallization age of between 2640 Ma and 2630 Ma (M.E. Villeneuve, pers. comm., November 2000; Appendix). Perks (1997) tentatively suggested an age of crystallization of 2639 ± 6 Ma (U-Pb (zircon), 2 fractions only) for a tonalite sheet at 'Forked Lake' within Yellowknife Supergroup metasedimentary diatexite units, 13 km south-southwest of the contact between the tonalite and the megacrystic granite at the southern boundary of the map area. This is within the estimated age range of of the tonalite at Ghost Lake. Monazite from the same tonalite sample at Ghost Lake has an age of between 2590 Ma and 2580 Ma (M.E. Villeneuve, pers. comm., November 2000; Appendix), which is similar to zircon and monazite U-Pb ages for the three granite bodies in the area that have been dated. The age of the monazite, with a closure temperature of about 700°C , whether interpreted as due to retrogressive metamorphic growth of secondary monazite or as reflecting the time of closure of U-Pb system for primary igneous monazite, represents the end of the granulite-grade conditions under which the tonalite crystallized (*see also* 'Megacrystic granite (unit Agm.o)' section and Villeneuve and Henderson (1998)).

A granodioritic phase within the retrograde granulite tonalitic gneiss (unit At-gn) from an island in Ghost Lake south of the Ghost River has a U-Pb (zircon) age of about 2605 Ma (Villeneuve and Henderson, 1998; Appendix). This age is significantly younger than the tonalite, but is similar to that of the quartz diorite that is described in the following section. Two zircon fractions from the granodiorite that are presumably xenocrystic have similar $^{207}Pb/^{206}Pb$ ages to those in the tonalite (Appendix). The same granodioritic sample has an εNd_T value of 1.0 and a depleted mantle model age (T_{DM}) of 2.8 Ga (Yamashita et al., 1999). Two samples from the tonalite at 'Forked Lake' had εNd_T values at 2650 Ma of 1.5 and -0.9 and T_{DM} ages of 2.88 Ga and 3.03 Ga, respectively (Perks, 1997).

Quartz diorite (unit Aqd)

A quartz diorite unit that also includes some diorite and tonalite occurs in both the northern part of the Dauphinee domain and the southwestern part of the deeper level Ghost domain. In the Dauphinee domain, the quartz diorite occurs in a single elongate intrusion that continues 20 km to the north beyond the map area (Fig. 6; Frith, 1993; Pehrsson and Kerswill, 1997b). It intrudes migmatitic metasedimentary rocks of the Yellowknife Supergroup and is intruded by granitic phases of the Dauphinee domain. The intrusion is compositionally varied over its extent, but is quite homogeneous on an outcrop scale. Inclusions are rather rare and include metasedimentary rocks. More commonly they are rounded to elliptical clasts of a more mafic version of the quartz diorite, up to 50 cm or 60 cm, and can be quite abundant where present (Fig. 55).

The rock is light grey- to whitish-weathering, medium grey, medium grained and inequigranular, with distinctive blocky plagioclase crystals up to 0.5 cm in a finer grained, dark matrix. The rock is commonly weakly foliated and has a metamorphic aspect in outcrop. In thin section the relict igneous texture is dominated by subhedral plagioclase laths that are in part mantled by finer grained hornblende and commonly kinked and altered biotite that make up between 20% and 25% of the rock (Fig. 56). Titanite is commonly associated with the mafic minerals both as discrete grains and as fine grains within the biotite and along its cleavage surfaces. Epidote is also commonly associated with the mafic minerals.

A characteristic feature of the unit is the pink to white pegmatite dykes, typically a few centimetres, but locally up to a metre wide, that are variably deformed, but less deformed than the quartz diorite (Fig. 57). In the western contact zone of the pluton, the quartz diorite is extensively intruded by various granite bodies of the Dauphinee domain and can exceed half the outcrop. No intrusions of the distinctive coarsely and densely megacrystic granite of the Ghost domain that occurs immediately to the west across the bounding shear zone were noted within the quartz diorite. From within the quartz diorite toward the cataclastic shear zone the intrusion becomes increasingly altered and brittly fractured with all mafic minerals going to chlorite and the feldspar minerals completely replaced by alteration products over a distance of several hundred metres to 1 km.

The quartz diorite intrusion approximately corresponds to a magnetic low (Fig. 7). It is not clear if this is a feature of the intrusion itself or if the magnetic low is part of the regional magnetic low that underlies the Hinscliffe and Wijinnedi domains to the west and continues to the north beyond the map area (Fig. 5).

Figure 55.

Quartz diorite intrusion from the northern Dauphinee domain here with abundant elongate more mafic enclaves. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416DD

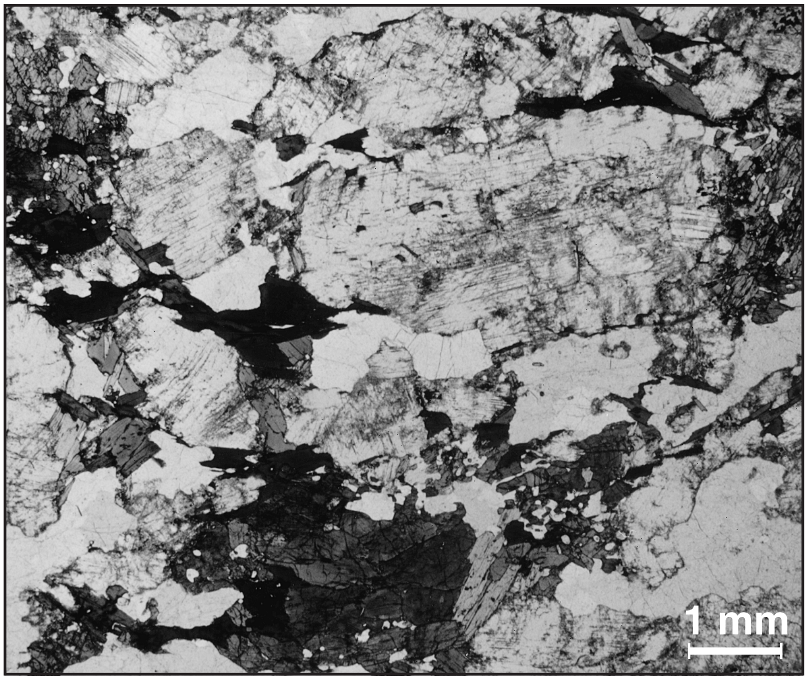


Figure 56.

Photomicrograph of moderately deformed Dauphinee domain hornblende-biotite quartz diorite in which relict igneous texture is quite apparent. Plane-polarized light.

Similar quartz diorite intrusions also occur in the Ghost domain at the western margin of the area. Although in general more deformed and metamorphosed, these rocks are similar in composition and, where best preserved, have the same texture, same suite of more mafic inclusions, and are crosscut by similar granite-pegmatite bodies (Fig. 58), strongly suggesting that they are closely related to the Dauphinee domain intrusion. In the Ghost domain these rocks have been mapped as three lens-shaped bodies. In addition, occurrences of the meta-quartz diorite, too small to resolve at the present scale of mapping, also occur as small bodies to inclusions within the undivided granitoid unit and as thin sills within the migmatitic Yellowknife Supergroup metasedimentary unit in the southwestern part of the domain (Fig. 59). The meta-quartz diorite is typically more strongly foliated, recrystallized, and extensively intruded by a variety of granitoid phases that

Figure 57.

Quartz diorite from the Dauphinee domain with abundant crosscutting dykes of minimally deformed pegmatite and granite. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416EE

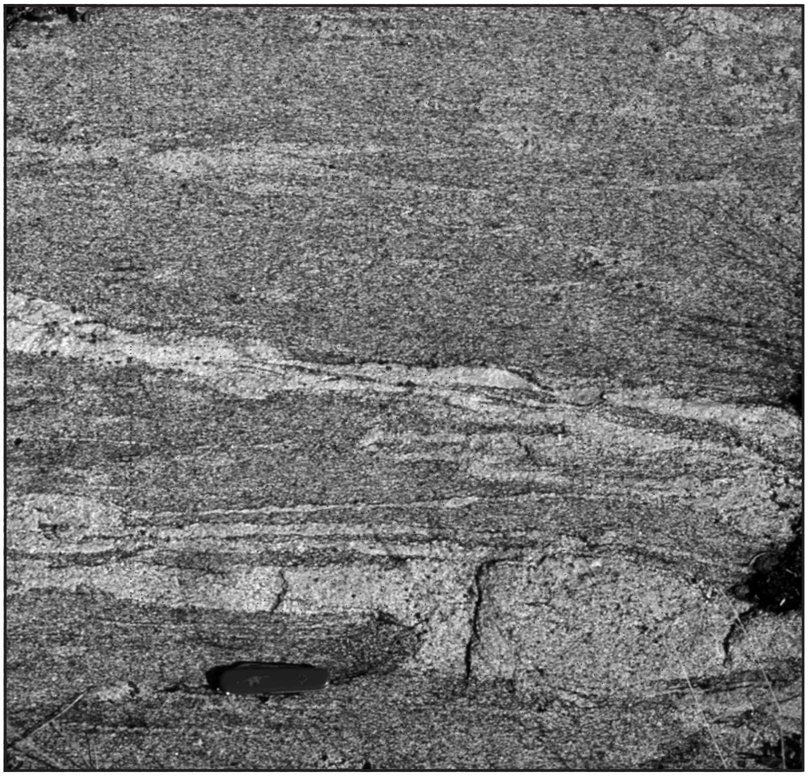


Figure 58.

Quartz diorite of the Ghost domain with a strongly developed foliation with pegmatite dykes transposed into the foliation. Compare with much less deformed equivalent from the Dauphinee domain (Fig. 57). Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416FF occur in this part of the domain. As in the Dauphinee domain the rock consists mainly of plagioclase, quartz, hornblende, and biotite, but somewhat altered clinopyroxene is usually also present (Fig. 60), and the small unit north of Ghost Lake contains remnant cores of orthopyroxene. This indicates that these rocks were at a significantly higher metamorphic grade than those to the northeast, locally into granulite grade, although they are now largely retrogressed.

The meta-quartz diorite body on Ghost Lake has a U-Pb (zircon) age of 2605 ± 3 Ma (Villeneuve and Henderson, 1998; Appendix). One of the zircon fractions analyzed consisted of a 2691 ± 1 Ma xenocryst which is within the age range of Yellowknife Supergroup volcanic rocks. It is also similar to the 2680 ± 3 Ma age of the main tonalitic component of the Cotterill gneiss complex that is intruded by the Dauphinee domain quartz diorite about 10 km north of the

Figure 59.

In addition to large mappable bodies, the quartz diorite commonly occurs in narrow sills and enclaves such as this quartz diorite inclusion within Yellowknife Supergroup metasedimentary migmatite. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416GG

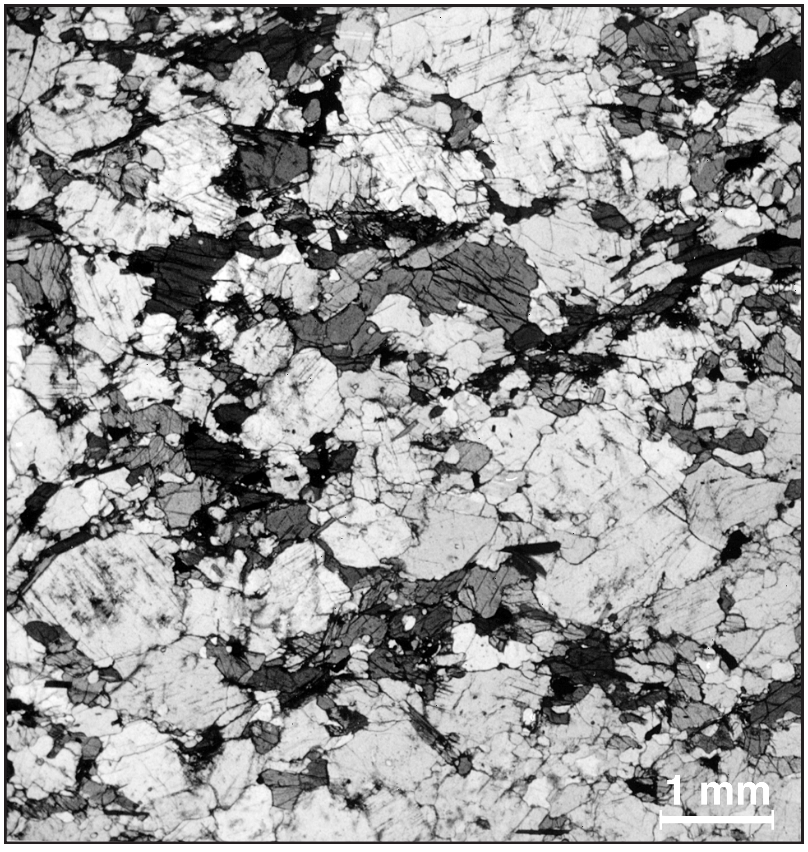


Figure 60.

Photomicrograph of metamorphosed and deformed Ghost domain, hornblende-biotite-clinopyroxene diorite which is largely recrystallized. Compare with similar, but less deformed rock from the Dauphinee domain (Fig. 56). Plane-polarized light.

area (Pehrsson and Villeneuve, 1999). The dated meta-quartz diorite sample on Ghost Lake has an εNd_{T} value of 1.1 and a depleted mantle model age (T_{DM}) of 2.9 Ga (Yamashita et al., 1999). The Strachan tonalite which occurs about 25 km north-northeast of the north end of the Dauphinee domain quartz diorite intrusion has a similar U-Pb (zircon) age of 2609 ± 1.5 Ma (Pehrsson and Villeneuve, 1999).

Granodiorite (unit Āgd)

A series of elongate bodies of granodiorite occur in the central part of the K-feldspar zone of the migmatitic Yellowknife Supergroup metasedimentary rocks. In places these rocks are in sharp contact with the migmatite that contains them. Elsewhere they grade from one to the other, particularly where the metasedimentary rocks are diatexitic. Inclusions of metasedimentary migmatite are common, although elongate rafts of dioritic to quartz dioritic gneiss possibly related to the previously described quartz diorite unit occur in the northeasternmost body. The unit is intruded by minor biotite-tourmaline pegmatite bodies.

The granodiorite is typically white weathering, grey, and weakly to moderately foliated to locally gneissic due to more deformed inclusions. The rocks are coarse grained and commonly megacrystic with subhedral feldspar crystals up to 4 cm in length, although 1–2 cm crystals are more typical. The megacrysts consist of both plagioclase and microcline and the proportion is highly varied such that the composition of the unit is actually quite heterogeneous ranging from tonalite to granite. Biotite varies in abundance on all scales and occurs as fine flakes in irregular aggregates through the rock. Fine-grained muscovite is a very minor, but everywhere present phase associated with the biotite, as is minor epidote. Myrmekite commonly occurs at the margins of the microcline megacrysts.

The close association of this unit with the migmatitic metasedimentary rocks, their gradational contacts, particularly with the diatexitic phases, suggests that there is a close relationship between the two.

Discussion

With a crystallization age older than 2630 Ma, the tonalite bodies are older than most other intrusions of the Slave Province, most of which range between 2630 Ma and 2580 Ma (Davis and Bleeker, 1999). Intrusions emplaced between 2660 Ma and about 2630 Ma, the young end of the age range of the bulk of Yellowknife Supergroup volcanism, are relatively rare. Some examples beyond this map area include phases of the Anton Complex foliated granodioritic and granitic intrusions at 2.64 Ga, 30 km north of Yellowknife (Dudás et al., 1989) and the Suse granite at 2641 ± 3.5 Ma, 100 km northeast of Yellowknife (James and Mortensen, 1992). The Providence granite at 2645.8 ± 1.6 Ma, 20 km south of the east end of Point Lake (Villeneuve et al., 1997) and the Olga suite tonalite units at 2650 ± 5 and 2649 ± 2 Ma, 50 km northeast of the east end of Point Lake (van Breemen et al., 1990) are two examples from the central Slave Province.

Younger intermediate intrusions ranging in composition from diorite to granodiorite and similar in age to the quartz dioritic rocks of the Wijinnedi Lake area occur throughout the Slave Province. These would include the Defeat Suite in the Yellowknife area (Henderson, 1985) (ca. 2625 Ma; Davis and Bleeker, 1999), the Tarantula granodiorite and Charloit tonalite in the easternmost Slave Province (Henderson et al., 1999) (ca. 2622–2616 Ma (van Breemen et al., 1987; van Breemen and Henderson, 1988)), and the Concession Suite in the Contwoyto Lake area (Davis et al., 1994) (ca. 2608 Ma; van Breemen et al., 1992). Davis and Bleeker (1999) suggested that these rocks were emplaced diachronously across the Slave Province.

Yamashita et al. (1999) noted a geochemical similarity between the quartz diorite of the map area and the somewhat older (ca. 2626 Ma; Davis and Bleeker, 1999) Defeat Suite intermediate intrusions at Yellowknife. On the basis of their Nd and Pb isotopic systematics, they suggest that the intermediate intrusions from these two areas, as well as similar intrusions about 12 km south of the Wijinnedi Lake area, were derived by partial melting of Yellowknife Supergroup mafic volcanic rocks that, in the course of their evolution, had been contaminated to some degree by older crust.

Mafic and ultramafic intrusions

Most of the mafic intrusions in the map area are part of the suite of metamorphosed gabbro to anorthositic gabbro plutons that has intruded the Yellowknife Supergroup in the Wijinnedi domain. Associated with them are numerous dykes and sills that occur in varied proportions throughout the volcanic dome and to a lesser extent in the metasedimentary rocks to the north. Other amphibolite sheets are present in varied amounts, some at mappable scale, in the higher grade Ghost domain. Their relationship to the metagabbro intrusions of the Wijinnedi domain is unknown. A few small bodies of metapyroxenite are also present within the three western domains. Their age relative to the other mafic intrusions is not known. A major suite of mafic synplutonic dykes are a characteristic feature of the Hinscliffe domain and have been described in the section ‘Hinscliffe complex (unit ĀH)’. The Paleoproterozoic Indin diabase dykes that are particularly abundant throughout the map area are described in the section ‘Paleoproterozoic diabase dykes (units PM, PI)’.

Metagabbro to meta-anorthositic gabbro (unit Āmg)

The metamorphic mafic intrusion between ‘Glazebrook Lake’ and Gale Lake is a complex suite of metagabbroic to meta-anorthositic phases. Closely related to it is the swarm of mafic sheets in the surrounding metavolcanic rocks that locally show a similar range of composition.

The contact of the intrusion with the surrounding volcanic rocks is quite sharp where seen on the eastern and northern margins of the complex, but to the south and west is, on the whole, rather gradational and is generalized on the map. There the marginal phases of the intrusion can carry a high proportion of volcanic inclusions, which together with the abundance of related mafic sills and perhaps small stocks in the country rock, together with the presence of more mafic volcanic units adjacent to parts of the intrusion, make it difficult to place the contact precisely. In many places the contact is marked by an elevation change with the intrusive rocks topographically higher.

The mafic rocks form a compositionally and texturally heterogeneous complex ranging from hornblendite to anorthosite, although the end members are not particularly common. In general the rocks in the western part of the complex are more hornblende rich, consisting of dark greenish-grey to green, originally fine- to medium-grained rocks. The most common rock is an originally coarser grained, more leucocratic metagabbro with somewhat greater than 50% plagioclase. Zones of originally coarser anorthositic metagabbro (Fig. 61), with plagioclase crystals up to 8 cm, occur locally, particularly in the eastern part of the complex, but do not appear to be areally extensive. Compositional and textural variations can be gradational or very sharp (Fig. 62) on an outcrop scale. The best preserved part of the intrusion is in the northeast where very coarse-grained, anorthositic gabbro occurs (Fig. 61). Coarse orbicular structures are locally present within these rocks (Fig. 63). Inclusions of metadacite are not common, but do occur in local concentrations, suggesting the complex is made up of several coalesced plutonic lobes that could be defined by more detailed mapping.

Figure 61.

Anorthositic metagabbro with very coarse-grained plagioclase crystals preserved within mainly recrystallized rock. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416HH

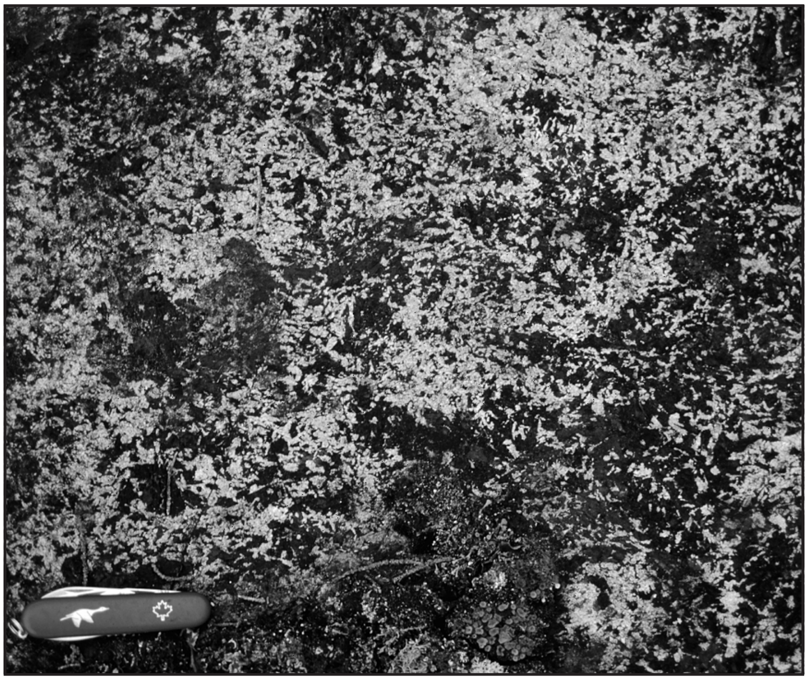


Figure 62.

Leucogabbro to anorthositic gabbro showing compositional and textural variation over small area. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416II

The complex has been metamorphosed to amphibolite grade, although in places parts of the southern portion of the complex are at upper greenschist grade. As the regional metamorphic gradient is thought to increase to the south, these low-grade rocks are considered to be the result of later retrogression. The original plagioclase-pyroxene igneous mineralogy is not preserved, but in many cases relics of the igneous textures are apparent. This texture can be seen both in hand specimen and in thin section where the mafic minerals occur as coarse-grained hornblende fringed by fine-grained hornblende. Plagioclase varies from coarse crystals to aggregates of recrystallized plagioclase (Fig. 64, 65). The intrusion is deformed, but the fabric is rather weakly developed, with much of the deformation being partitioned into discrete shear zones a few metres to several tens of metres wide (Fig. 66).

Figure 63.

Coarse orbicular structures over 20 cm in diameter in a more leucocratic phase from an unusually well preserved part of the complex in the northeastern part of the intrusion. Brunton compass 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417H

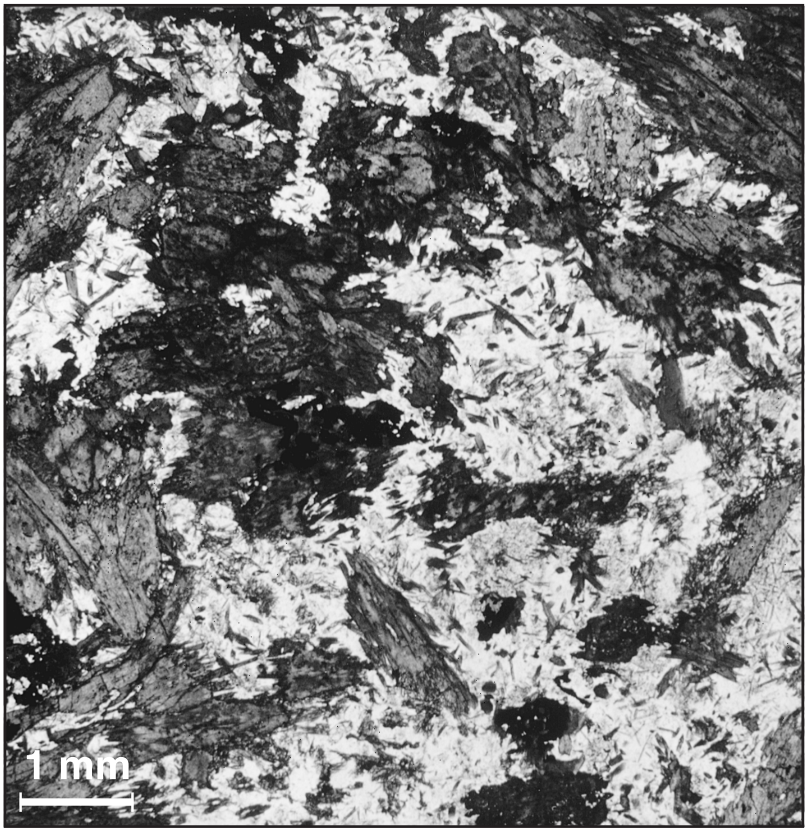


Figure 64.

Photomicrograph of leucogabbro in which coarse-grained hornblende crystals fringed by fine-grained hornblende is presumed to be pseudomorphous after the original igneous pyroxene. Plagioclase is partially recrystallized and contains fine-grained hornblende and a colourless amphibole. Plane-polarized light.

Although the intrusion as a whole does not have a magnetic expression, certain phases within it appear to be quite magnetic. Against the rather low, flat magnetic field that underlies both the Wijinnedi and Hinscliffe domains, there is a prominent, roughly triangular anomaly situated over the western margin of the complex (Fig. 7; Geological Survey of Canada, 1969). One apex extends about 5 km to the west, south of 'Glazebrook Lake' and another 3 km to the northeast, mainly within the mafic complex, toward Lac Avril. The less intense, third apex extends about 3 km to the southwest. Although the magnetic source(s) was not identified, it is thought that it is probably dykes both within and extending beyond the mafic complex, but related to it. A smaller magnetic anomaly occurs over the end of the easternmost arm of the intrusion.

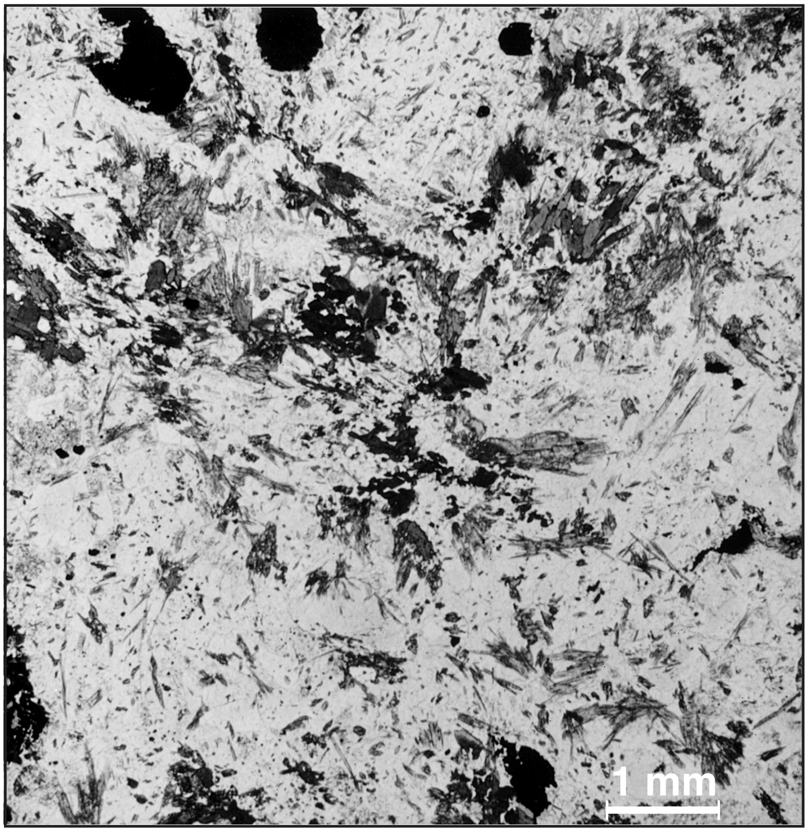


Figure 65.

Photomicrograph of anorthositic gabbro in which the plagioclase is largely recrystallized. Mafic minerals consist of both hornblende and a colourless amphibole. Plane-polarized light.

Figure 66.

Shear zone in gabbroic anorthosite. Although most of the rocks of the complex have a weak fabric, most of the strain is partitioned into several-metre-scale shear zones. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416JJ

Mafic sheets in the Wijinnedi domain

Mafic sheets occur throughout much of the Yellowknife Supergroup volcanic complex. They are particularly abundant east and southeast of the above-described metagabbro plutonic complex where locally they can form up to half the outcrop area. They are less common to the north and southwest of the complex and relatively rare west of ‘Glazebrook Lake’. They are more abundant in the vicinity of the contact between the Yellowknife Supergroup metavolcanic and metasedimentary rocks south of Wijinnedi Lake and ‘West Wijinnedi Lake’ and are a minor occurrence in the sedimentary rocks to the north.

The sheets are everywhere metamorphosed in context with the rocks they intrude. Most of the dykes at higher grades are dark green- to black-weathering, dark grey-green, medium- to fine-grained rocks, have a metamorphic texture, and consist mainly of hornblende, plagioclase, and titanite. In the lower grade metasedimentary rocks north of ‘West Wijinnedi Lake’ the intrusions are light grey-green and consist of assemblages of plagioclase, chlorite, actinolitic amphibole, and carbonate. Most of the dykes are metagabbro, but more leucocratic metagabbro to anorthositic metagabbro units occur locally, particularly east of the plutonic complex. The mafic sheets are up to a few tens of metres, but most commonly are much thinner and can be as thin as a few centimetres. In the metasedimentary rocks the metagabbro is mainly layer parallel (Fig. 67), but also occurs in crosscutting dykes and small pods to plugs. In the metavolcanic rocks, particularly at higher grades, the mafic sheets typically have a generally east trend, more or less parallel to the principle foliation in the rocks they intrude. In detail, there can be a low-angle discordance between the dykes and the foliation (Fig. 68). The dykes are deformed, commonly having schistose margins and a foliation that is typically less prominent than that in the rocks they intrude. In some cases they have early fracture sets, now somewhat curved and altered due to the subsequent deformation and metamorphism (Fig. 19) that in some cases produces structures not unlike deformed pillows. At higher grades, in particular in the more deformed southernmost part of the volcanic complex, they become difficult to differentiate from some of the more mafic volcanic units which are more abundant there.

Although most of the mafic sheets are considered to be comagmatic with the main plutonic complex (unit Amg) due to their lithological similarity, there are other, probably minor, mafic intrusions that are clearly older. One such dyke is shown in Figure 69, where the older dyke, folded in the principle foliation, is crosscut by one of the more common metagabbro dykes. The older mafic dyke which is garnet bearing (a mineral not known to occur in the crosscutting suite of metagabbro dykes at this metamorphic grade) could conceivably be related to the Yellowknife Supergroup mafic volcanic rocks that overlie the dacitic rocks and do locally contain metamorphic garnet. Another mafic dyke shows magma-mixing textures with a dacitic synvolcanic intrusion associated with the Yellowknife Supergroup felsic centre (Fig. 70). This would be further evidence of mafic intrusions contemporaneous with Yellowknife Supergroup volcanism.

A few mafic sheets occur that have a strongly developed internal layering up to 10 cm (Fig. 71). Each layer is strongly compositionally and texturally graded from medium-grained anorthosite or gabbro anorthosite to fine-grained hornblendite. This might be considered evidence that the sheet formed in a horizontal position and the layering is the product of gravitational settling and thus useful evidence for the timing of emplacement of the intrusions. On the other hand, a Paleoproterozoic Indin dyke from this area that was emplaced vertically and has since been tilted at most only a few degrees, shows somewhat similar if not so strongly developed features (Fig. 72). Similar structures have been described and discussed by McCall and Peers (1971) in the very large, vertical Binneringie Dyke of Western Australia.

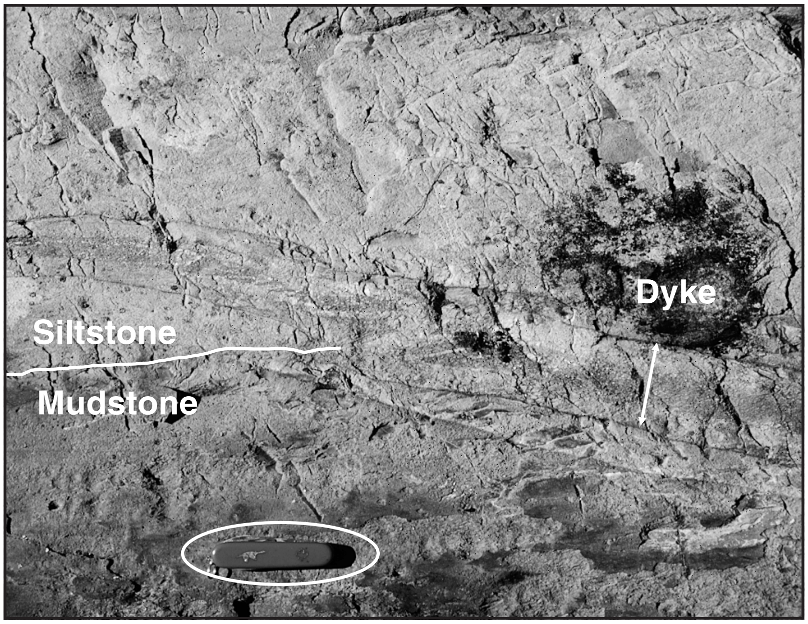


Figure 67.

Very thin, metagabbro dyke, slightly discordant to both bedding and cleavage, intruding biotite-zone metasiltstone (upper part of photo) and metamudstone. Note the contact metamorphic zone almost half the width of the dyke that is most apparent in the mudstone. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416KK

Mafic sheets in the Ghost domain (unit Aam)

Amphibolite sheets are a fairly common occurrence in the Ghost domain, although only a few of the largest bodies are shown on the map (Map 2023A). They are common as thin layers a few centimetres to, less commonly, a few metres in width in the tonalitic gneiss (unit Atgn.o) south and southeast of Ghost Lake; indeed their presence contributes to the gneissic aspect of this rock. They also occur in some of the other rocks of the domain, in particular in the undivided granitoid rocks (unit AG) northwest of Ghost Lake where they occur both as intrusions and inclusions depending on the relative age of the particular phase in the unit. They have not been recognized in the three main granite units of the domain (units Agm.o, Ag.o, Ag).

The mafic sheets are black to dark greenish-grey, weakly foliated, faintly layered, fine- to medium-grained, feldspathic amphibolite gneiss to amphibolite. Some contain sparse, centimetre-scale, coarse-grained, plagioclase-hornblende leucosome layers. In the granulite zone they contain assemblages of plagioclase, clinopyroxene, orthopyroxene, hornblende, an opaque phase, and less commonly biotite, titanite, and in one case, spinel. In the lower grade parts of the domain the pyroxene minerals are missing, but in all cases the textures are metamorphic and no igneous mineralogy has been preserved.

Metapyroxenite (unit Aum)

A series of small metapyroxenite bodies occur within the Wijinnedi, Hinscliffe, and Ghost domains. None are known from the Dauphinee domain, but that may be more a function

Figure 68.

Mafic dyke emplaced at low angle to principal foliation as indicated by orientation of dacitic breccia fragments. The dyke itself is extensively fractured with the fracture surfaces altered during metamorphism. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416LL

Figure 69.

Most of the mafic sheets associated with the metagabbro are roughly concordant with the easterly trending principal foliation as is the case here with the larger dyke. The dyke intrudes an older, lithologically distinct mafic dyke that is deformed in the principal foliation. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416MM of greater Quaternary cover and less close traversing of that domain than anything else. The largest known bodies occur on islands in ‘Hinscliffe Lake’, in metatrondhjemite of the Hinscliffe domain, and in the higher grade, metavolcanic rocks of the southern Wijinnedi domain immediately north and east of ‘Peaks Lake’ (Wright, 1950a, b). The other occurrences are much smaller; the bodies are on a scale of a metre to a few metres, and as such appear in exaggerated form on the map.

The metapyroxenite is a dark greenish-black to dark green, medium- to coarse-grained (locally up to 4 cm), massive, homogeneous rock composed almost entirely of actinolite. The Hinscliffe Lake body has abundant metatrondhjemite country rock inclusions to the extent that it is difficult to precisely define the contact, let alone its orientation. The metapyroxenite is intruded by veins of very fine-grained, pink-weathering, grey, almost glassy granite to massive to weakly foliated grey trondhjemite to coarse-grained pegmatite (Fig. 73). Although the metapyroxenite bodies are massive, the larger ones are up to 2 km long and 150 m wide and are concordant with the Archean structure. Although some of the smaller bodies are small plugs, others occur as zones of angular inclusions of metapyroxenite in a leucocratic granitic matrix and may represent disrupted metapyroxenite sheets (Fig. 74).

The metapyroxenite consists almost entirely of coarse, subhedral grains of pale green actinolite (Fig. 75). The actinolite occurs in coarse, subhedral to anhedral grains, as fine-grained granular aggregates, as a matrix to the coarser

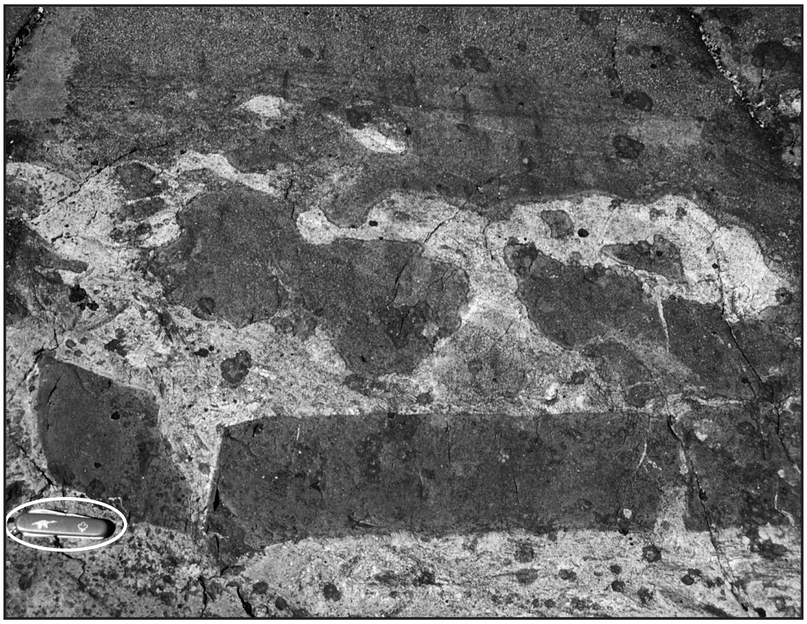


Figure 70.

Mafic dyke intruding synvolcanic felsic intrusion. Lobate structures suggest magma mixing structures, which would suggest the felsic and mafic intrusions were contemporaneous. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416NN

Figure 71.

Mafic sheet consisting of texturally and compositionally graded layers which might suggest the sheet was emplaced in a subhorizontal position; however, similar features also occur in vertical Paleoproterozoic Indin dykes (Fig. 72). Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416OO material, and as much finer elongate prismatic crystals. In some cases ragged cores of clinopyroxene are preserved within the coarser grains. Minor phases which may or may not be present include plagioclase, biotite, and quartz and are never more than a few per cent of the rock. They occur both as inclusions within the coarse actinolite and interstitially between actinolite laths. In some cases, the actinolite also contains fine-grained epidote group minerals, chlorite, and carbonate and are thought to be alteration products of the clinopyroxene core material.

The granitoid dykes and veins that intrude the metapyroxenite vary from fine-grained, almost glassy-appearing thin dykes to coarse pegmatite with crystals of dark green to black amphibole up to 10 cm long oriented perpendicular to the walls of the dyke. In the case of the intrusion at ‘Hinscliffe Lake’ which was examined in thin section, the dykes are similar to the trondhjemitic country rock in which the intrusion was emplaced and have the sugary, recrystallized texture that characterizes much of the trondhjemite of the Hinscliffe complex. The main difference between the two is that hornblende is a common mafic mineral in the dykes, whereas biotite only is present in the trondhjemite country rock.

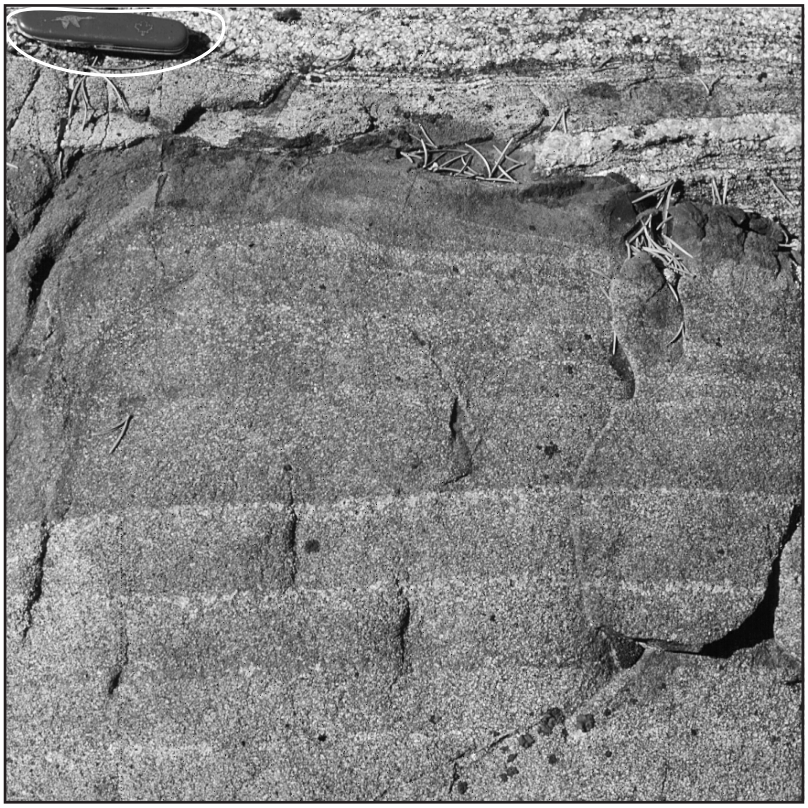


Figure 72.

Subvertical and nondeformed Paleoproterozoic Indin diabase dyke containing texturally and compositionally graded layers adjacent to and parallel to chilled margin of dyke. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416PP

Figure 73.

Metapyroxenite with crosscutting granitoid to pegmatitic dykes. Coarsest amphibole occurs in crosscutting pegmatitic dykes. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416QQ

Discussion

The metagabbro plutonic complex and associated mafic sheets in the volcanic rocks of the Wijinnedi domain have not been radiometrically dated, but are clearly premetamorphic. Since the trend of the mafic sheets is roughly parallel to the principal foliation and the intrusions are to some degree deformed, they would appear to have been emplaced during formation of the principal foliation. Mafic intrusions also occur in the Yellowknife Supergroup metasedimentary rocks, which are the youngest rocks in which they occur and as such, their sedimentation age represents a maximum age limit for the metagabbro sills.

Similar mafic intrusions occur elsewhere in the Slave Province and are everywhere closely associated with Yellowknife Supergroup volcanic rocks. These include the anorthositic gabbro body in the Chan mafic volcanic sequence north of Yellowknife (Padgham, 1987) and the gabbro-anorthositic gabbro at Camsell Lake south of MacKay Lake in the south-central Slave Province (Johnstone, 1992). The

Figure 74.

Angular blocks of metapyroxenite in a granitic matrix. Hammer is 33 cm long. Photograph by J.B. Henderson. GSC 2002-416RR

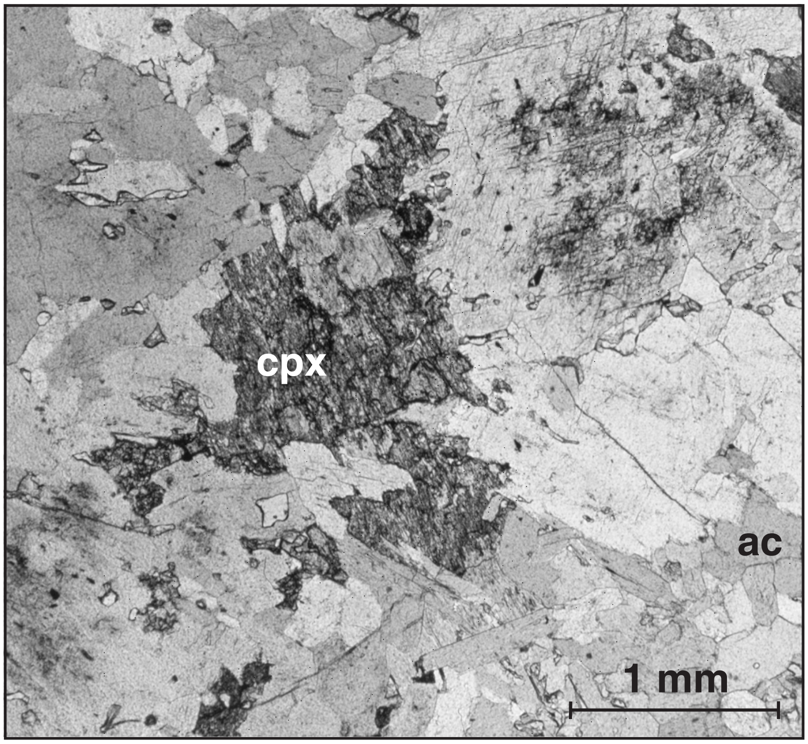


Figure 75.

Photomicrograph of metapyroxenite with relatively high proportion of relict clinopyroxene (cpx) cores contained in coarse-grained actinolite (ac) laths. Plane-polarized light.

best known example occurs at, and south of, Clinton-Colden Lake in the eastern Slave Province (Henderson et al., 1999), parts of which have been studied in detail by Macfie (1987, 1989). Like the plutonic complex in the Wijinnedi domain, parts of the intrusion are strongly magnetic. The Clinton-Colden intrusion, emplaced in the lower part of the dominantly mafic volcanic sequence, is synvolcanic with a U-Pb (zircon) age of 2686 ± 3 Ma (Macfie et al., 1990), whereas the U-Pb (zircon) age of a presumed stratigraphically higher rhyolite dome in the volcanic sequence is 2671 +2/-4 Ma (van Breemen and Henderson, 1988). Archean megacrystic anorthositic rocks from elsewhere in the world have been described and discussed by Ashwal (1993) and Ashwal and Myers (1994).

Any relationship between the mafic sills in the Ghost domain and the mafic intrusions of the Wijinnedi domain are difficult to establish with any confidence. The Ghost domain mafic sheets, although deformed, are assumed to be intrusive into the ca. 2640–2630 Ma tonalite mainly on the basis of their continuity over considerable distances. Since they have not been recognized within the granite units, they are presumably older than ca. 2600 Ma.

The metapyroxenite is regarded as an Archean intrusion that, given its massive nature, was probably emplaced late in the deformational history of the region. In the Wijinnedi and Ghost domains, the rocks in which they are emplaced are at middle to upper amphibolite grade. The amphibolite dykes in the Hinscliffe domain are at a similar grade. The fact that the metapyroxenite intrusions consist almost entirely of actinolite might suggest that they are metamorphically out of context with the rocks in which they occur. On the other hand, since the intrusions consist mainly of actinolite, there is very little available to react with that would result in a more stable, higher grade assemblage. Boyd (1959), in a hydrothermal experiment, showed that tremolite, the magnesian end-member of the tremolite-ferroactinolite series, could persist to about 870°C at 2 kbar before breaking down to enstatite, diopside, and quartz. Also, the crosscutting granitoid dykes, which are metamorphically recrystallized, do contain hornblende. Although all the rocks in the map area were metamorphosed during the Paleoproterozoic to greenschist grade as shown by the metamorphic assemblages in the Paleoproterozoic Indin dykes, the effect is not nearly as pervasive as seen in the metapyroxenite rocks since the clinopyroxene in the diabase dykes are only marginally replaced by fine-grained actinolitic fringes. The chlorite-, carbonate- and epidote-group-mineral assemblages seen in some metapyroxenite samples may be a response to this later metamorphism.

The relationship of the metapyroxenite intrusions to the other mafic intrusions in the area is not known. No ultramafic bodies are known to be associated with the more extensive metagabbro intrusions and they and the associated mafic sills tend on the whole to be more deformed than the metapyroxenite bodies. At the western end of the Hinscliffe complex, about 10 km west of the map area, several similar ultramafic intrusions occur as part of a gabbroic complex that ranges in composition from ultramafic to granodiorite (Jackson, 2003). V.A. Jackson considered these intrusions to be Archean due their participation in northerly trending Archean folds (V.A. Jackson, pers. comm., 2000).

Granite intrusions

Four major granite bodies occur within the map area. Two consist of orthopyroxene-bearing, megacrystic granite and the other two of more even-grained granite, only one of which crystallized under granulite-grade conditions. In general, they are less deformed than the intermediate and more mafic intrusions and are younger, having been emplaced between about 2600 Ma and 2590 Ma (Villeneuve and Henderson, 1998).

Megacrystic granite (unit Agm.o)

Several bodies of distinct, coarsely and densely megacrystic granite occur in the area (Fig. 76). The largest of these occurs in the core of the large antiformal structure south of Ghost Lake. A second, smaller body occurs north of the east end of Ghost Lake and extends about 3 km north of the map area (Pehrsson and Kerswill, 1997b). The intrusion south of Ghost Lake is more or less concordant with structure in the adjacent tonalitic gneiss allowing the possibility that the intrusion has a sheet-like form. Indeed the two megacrystic granite bodies may well be parts of the same intrusion with the northern body representing the northward extension of the antiformal structure south of Ghost Lake. Southeast of Ghost Lake are two small intrusions of similar megacrystic granite.

The contact of the granite is typically sharp, particularly with the tonalite. The contact with diatexitic metasedimentary rocks is commonly more gradational, with the size and proportion of K-feldspar and garnet megacrysts in the diatexite increasing and decreasing, respectively, toward the contact. Near the contact, minor small inclusions of either tonalite or metasedimentary diatexite are present in the otherwise rather homogeneous granite. In one case, small, fine-grained, mafic enclaves occur that also contain sparse, but otherwise similar K-feldspar megacrysts (Fig. 77). Minor lenticular bodies to sills a few tens of metres wide occur locally in the country rock in the vicinity of the contact.

The granite weathers pink to pinkish grey; the megacrysts are pink and their matrix more grey to white (Fig. 76). The granite is commonly deeply weathered. It is on the whole quite homogeneous, the main variations being the size and proportion of the megacrysts. These are usually several centimetres in length and can be up to 8 cm and are typically densely packed (Fig. 76). They are, for the most part, randomly oriented. Locally the megacrysts are aligned which can vary on an outcrop scale and is presumably due to local, preconsolidation turbulent flow. In some cases, the megacrysts are aligned parallel to the contact, close to the intrusion margin. In the northern body, the megacrysts tend to be parallel to the regional northerly structural trend.

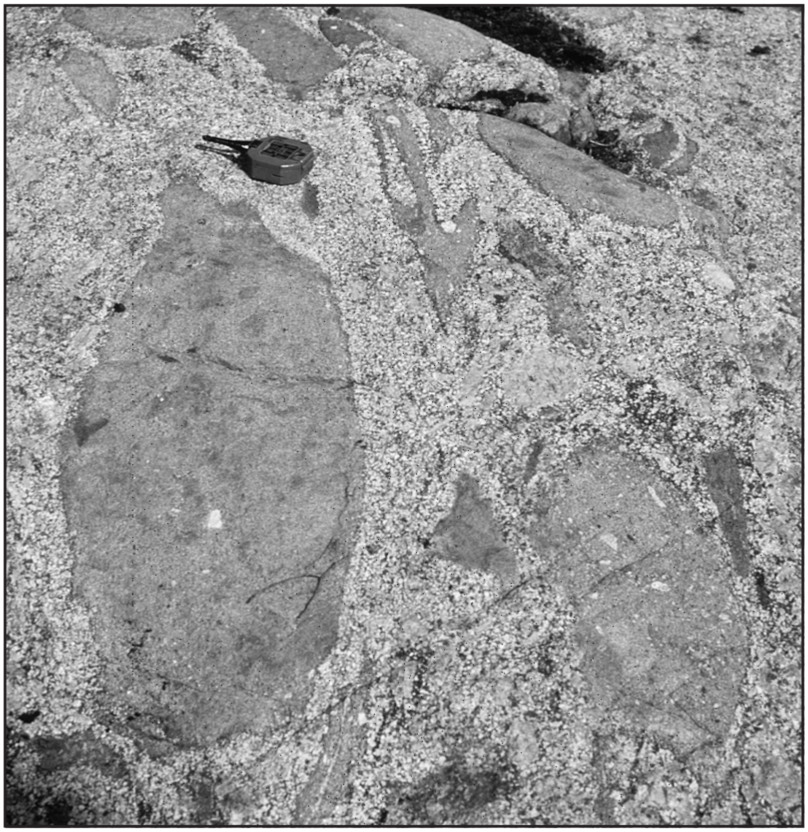
The granite has a coarse-grained matrix and in most cases is not foliated. The megacrysts are microcline with typically very finely developed microcline twinning (Fig. 78). They

Figure 76.

Megacrystic granite with densely packed, randomly oriented, very coarse-grained, pink, mainly K-feldspar megacrysts in a grey to white, coarse-grained quartz, plagioclase, and biotite matrix. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416SS

Figure 77.

*Fine-grained mafic enclaves containing sparse K-feldspar megacrysts identical to the K-feldspar crystals in the megacrystic granite that contains the enclaves. The presence of such megacrysts in the mafic enclaves has been cited as evidence for the metasomatic origin of the feldspar by some, whereas others suggest they are a result of commingling of more mafic magma with the granitic magma during the coarse of which some K-feldspar megacrysts of the granite were entrapped within the mafic globules (*see *Vernon (1986) for a review). Brunton compass 7 cm wide. Photograph by S.E. Schaan. GSC 2002-417I*



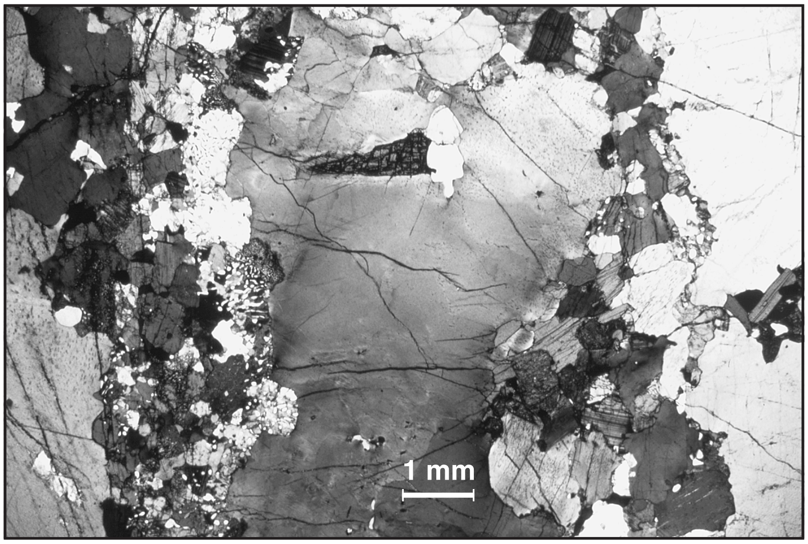


Figure 78.

Photomicrograph of megacrystic granite with examples of rarely preserved orthopyroxene. Coarse-grained microcline with orthopyroxene inclusion is in part mantled by myrmekite. This rock is moderately deformed, leaving the matrix of the rock with a finer grained, granulated aspect. If the microcline megacryst is a phenocryst, the occurrence of orthopyroxene within the microcline is evidence for the crystallization of the granite under granulite-grade conditions. Cross-polarized light.

contain minor inclusions of plagioclase most commonly, and quartz crystals, that in some cases, are subhedral to euhedral. Less commonly biotite and orthopyroxene inclusions are present (Fig. 78). Myrmekite commonly occurs at the margin of the microcline. Weakly to untwinned plagioclase occurs in 0.4–1 cm, subhedral grains, with minor K-feldspar exsolution in some cases, and is typically altered to some degree. Quartz forms irregularly shaped, polydomainal masses commonly in excess of 1 cm, with no particular orientation in most cases. The dominant mafic mineral is dark brown to olive-brown biotite, typically with fine-grained opaque exsolution that usually occurs in aggregates of finer grained flakes. Preserved orthopyroxene is rarely seen (Fig. 78) (Chacko et al., 1995a; Perks, 1997); more commonly pyroxene altered to low-grade chloritic minerals is present, particularly in the northern body. Secondary minerals include minor, fine-grained muscovite, chlorite, and epidote. Igneous textures are sometimes moderately well preserved, but where the rock is more deformed, the granite has a more recrystallized aspect with the K-feldspar megacrysts remaining much the same, but in a more granulated, finer grained matrix of plagioclase, quartz, and biotite (Fig. 78).

The pink colour of the megacrystic granite is in strong contrast to the characteristic greenish-brown to yellow colours of the nearby granulite-grade tonalite and granite. This, together with the lack of obvious pyroxene, at least in hand specimen, might suggest that the megacrystic granite was emplaced after granulite-grade metamorphism; however, the country rocks at the contact are not retrogressed, and although orthopyroxene does not occur in most examples examined, it or its retrograde equivalents are present locally which would indicate the granite has been exposed to granulite-grade conditions. In Figure 78 orthopyroxene is shown in a rather deformed example of the granite, but one grain occurs within a K-feldspar megacryst. If, as argued by Vernon (1986), megacrysts such as those that occur in this megacrystic granite are phenocrysts as opposed to porphyroblasts, the occurrence of orthopyroxene within the K-feldspar megacryst is strong evidence that the granite crystallized under granulite-grade conditions. That Vernon's (1986) interpretation is likely correct in this case is supported by the occasional presence of fine-grained quartz with square β-quartz formed as inclusions within the K-feldspar megacrysts. Chacko et al. (1995a) suggested that due to the progressive increase of water in the liquid phase of the crystallizing granite, most of the original orthopyroxene was retrograded to biotite.

In addition to the two main areas of megacrystic granite, the southern body has several 0.5 km satellite lenses of lithologically similar granite within a few hundred metres of the contact. Two other, more remote stocks of megacrystic granite occur to the east, one within the tonalite gneiss and the other, containing fine-grained garnet, within the zone of migmatitic metasedimentary rocks.

The megacrystic granite has a U-Pb (zircon) age of 2598 ± 2 Ma (Villeneuve and Henderson, 1998; Appendix), which is similar to, or somewhat older than the other two granite samples dated (*see* below) and is within the age range of most granite intrusions postdating Yellowknife Supergroup rocks (2605–2580 Ma; van Breemen et al., 1992; Villeneuve and van Breemen, 1994) of the Slave Province. The megacrystic granite also has a U-Pb (monazite) age of 2589 ± 2 Ma. This compares with the 2590–2580 Ma age range of monazite in the 2640–2630 Ma tonalite intruded by the megacrystic granite. As suggested in the 'Tonalite' section, the monazite ages (with a closure temperature of about 700°C) (Heaman and Parrish, 1991) for both the granite and the significantly older tonalite, whether interpreted as due to retrogressive metamorphic growth of secondary monazite or as reflecting the U-Pb system closure of primary igneous monazite, represents the end of the granulite-grade conditions under which the granite crystallized (Villeneuve and Henderson, 1998; Appendix).

Yamashita et al. (1999) determined $\mathrm{Nd_T}$ values of 0.6 for the megacrystic granite sample that was dated and a corresponding $\mathrm{T_{DM}}$ age of 2.9 Ga. They argued on the basis of these and other geochemical parameters that this granite and the other two analyzed from the map area (described below) were derived from a metasedimentary source that in turn had mixed juvenile, presumably synchronous with the Yellowknife Supergroup, and older components.

Granite and pegmatite (units Āg, Āg.o)

A large body of granite and associated pegmatitic bodies occurs in the sub-granulite-grade rocks of the northwestern Ghost domain, northwest of the main part of Ghost Lake (unit Āg). Lithologically similar and presumably related rocks also occur in small-scale, generally concordant sheets to small stocks in many places in the map area, but most abundantly along the northwest shore of Ghost Lake and in the zone dominated by undivided granitoid rocks and gneiss. A second major granite pluton (unit Āg.o) occurs in the southeast corner of the map area, where it is associated with granulite-grade rocks.

The largest body, northwest of Ghost Lake, is a locally homogeneous, but regionally somewhat compositionally heterogeneous intrusive complex that is not particularly well exposed relative to most of the area, due to glacial deposits. The rock is pink weathering, pink to pinkish grey, medium to medium-fine grained, and typically even grained, although locally it is sparsely megacrystic with 1–1.5 cm K-feldspar megacrysts. It is typically massive to locally weakly foliated and, as such, contrasts with most of the older granitoid rocks of the region. Contacts with the metasedimentary rocks, tonalite, and other undivided granitoid rocks were not seen, but these rock units contain generally concordant sills of the granite and also occur as locally abundant, sharply bounded, angular inclusions within the granite. On the whole, metasedimentary rocks are the most common inclusion type. An area has been outlined on the map in which metasedimentary material is particularly abundant (unit ĀYs-g-t). The other inclusion types dominate near contacts with similar rocks.

The composition of the unit varies from alkali-feldspar granite to granodiorite. On the whole, igneous textures are well preserved although the various phases have anhedral to less commonly subhedral crystal form (Fig. 79). Evidence of deformation, where present, is expressed by minor granulation along coarser grain margins and minor development of myrmekite. Biotite, which occurs in varied proportions, is the only mafic mineral and tends to show no particular orientation.

The second major granite pluton occurs in the southeast corner of the map area, well within the zone of orthopyroxene-bearing, granulite-grade rocks. It is more or less symmetrically disposed with respect to the larger granite body about the main antiformal structure of the domain. The contact with the surrounding tonalite to tonalitic gneiss is not well defined due to the abundance of granite dykes and veins within the tonalite bodies and large inclusions (tens of metres scale) of tonalite and tonalite gneiss within the granite body. The eastern boundary of the granite is the domain-bounding cataclastic shear zone, within which the granite is retrogressed over a distance of about 50 m. The rock is dark pinkish weathering, but on the fresh surface is typically the characteristic greenish brown of granulite-grade rocks. The granite is similar in most respects to the previously described pluton except for the presence of somewhat altered orthopyroxene and, less commonly hornblende, in addition to biotite as the mafic mineral component of the rock (Fig. 80). The granite is commonly coarse grained, but not megacrystic and pegmatitic phases were not noted. It is generally massive, but contains local shear zones, within which the granite is retrogressed.

In addition to the two main granite plutons, thin, generally concordant sheets and lenticular bodies of granite and less commonly pegmatite, for the most part on too small a scale to map separately, occur within the tonalitic gneiss and tonalite, the undivided granitoid unit, and metasedimentary migmatite. They vary greatly in size from a few centimetres

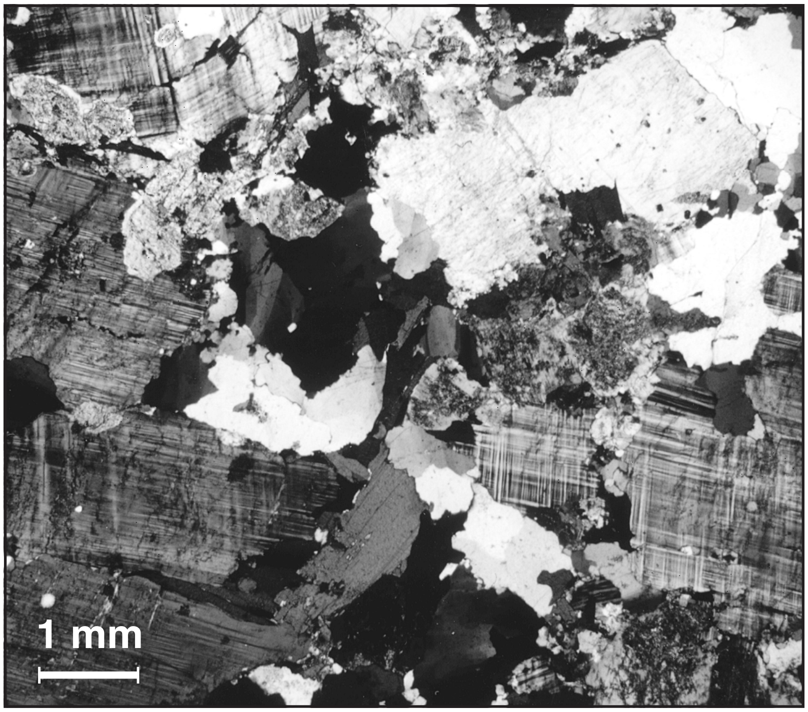


Figure 79.

Photomicrograph of granite consisting of mainly coarse-grained anhedral microcline, coarse-grained, irregular masses of quartz, minor plagioclase, and biotite from the large heterogeneous pluton from the north-central part of the area that crystallized under amphibolite-grade conditions. Compare with Figure 80 from the granite pluton from the southeast corner of the map area. Cross-polarized light.

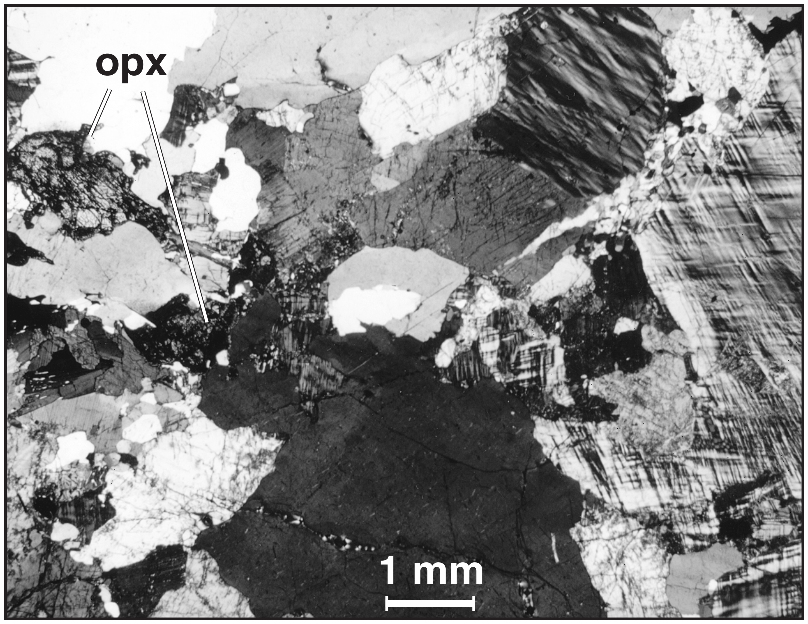


Figure 80.

Photomicrograph of granite from the southeast corner of the map area that crystallized under granulite-grade conditions. It is compositionally and texturally similar to the granite of Figure 79, but contains somewhat altered orthopyroxene (opx) and in some cases hornblende in addition to biotite as the main mafic minerals. Cross-polarized light.

to tens of metres and form less than 20% to 30% of the outcrop, typically less than 5% to 10% where present at all. A few small granite intrusions also occur within the Hinscliffe domain and it is assumed that they are related to the more abundant, lithologically similar, granite sheets in the rocks of the Ghost domain (Fig. 45). The massive nature of these granite bodies contrasts strongly with the generally strongly foliated rocks they intrude.

A U-Pb (zircon) age determination was made on one of the sills of pink, massive granite within the undivided granitoid intrusions and gneiss on an island south of the Ghost River. The sill has a U-Pb (zircon) age of 2593 +6/-4 Ma (Villeneuve and Henderson, 1998; Appendix). The orthopyroxene granite has a U-Pb (zircon) age of 2589 +1/-2 Ma (Villeneuve and Henderson, 1998; Appendix). This is within error of the age of the granite sill and somewhat younger than the U-Pb (zircon) age of the megacrystic granite (2598 ± 2 Ma). Like the megacrystic granite, these granite ages are within the age range of most of the granite bodies postdating Yellowknife Supergroup rocks in the Slave Province (2605–2580 Ma; van Breemen et al. (1992); Villeneuve and van Breemen (1994)).

Yamashita et al. (1999) determined Nd_T values of 0.6 for the granite sill and 1.1 for the orthopyroxene granite, and a corresponding T_{DM} ages of 2.9 Ga and 2.8 Ga for the same geochronological samples, respectively. They argued on the basis of these and other geochemical parameters that these granite samples, as well as the megacrystic granite described previously, were derived from a metasedimentary source that in turn had mixed juvenile, presumably synchronous with Yellowknife Supergroup, and older components.

Granitoid rocks of the Dauphinee domain (unit ADg)

The largest unit in the Dauphinee domain consists largely of granite with smaller amounts of granodiorite and less commonly tonalite. These rocks are less well exposed than the rocks of the domains to the west and were not traversed as closely. These granite bodies intrude both the quartz diorite–tonalite unit in the northeastern corner of the area and the Yellowknife metasedimentary rocks. The western boundary of the unit is the cataclastic shear zone that separates the Dauphinee domain from the Ghost domain. Although the granite is a major component within the area in the Dauphinee domain, no significant analogues are known to the west in the Ghost domain. It is similar in some respects to the Awry granite about 80 km south of the area, northwest of Yellowknife (Henderson, 1985).

The granite is pink to pale pink to white and generally massive, although locally it can have a weak foliation or is vaguely layered. It is medium to coarse grained, and usually sparsely megacrystic with subhedral microcline crystals usually less than 1 cm, rarely more than 2 cm. It is dominantly a leucocratic granite with minor granodiorite and rare tonalite phases that were not outlined. Pegmatitic phases are more common in the northern part of the unit. Inclusions are locally present, but are minor volumetrically. They vary from other, generally more mafic, granitoid to granitoid gneiss phases, to amphibolite and amphibolitic gneiss, to psammitic to pelitic metasedimentary rocks. One rather fine-grained and altered tonalitic inclusion near the domain boundary contained altered clinopyroxene and possibly orthopyroxene, the only assemblages seen in the domain reminiscent of the granulite-grade rocks a few hundred metres to the west in the Ghost domain. Most of the granite retains igneous textures. It is dominated by anhedral to subhedral plagioclase and microcline, with both, particularly the microcline, containing inclusions of biotite, the other feldspar, and quartz. The rock also contains abundant quartz, minor biotite, and usually small amounts of fine-grained muscovite.

Granite up to a distance of 800 m to 1 km from the cataclastic shear zone domain boundary is noticeably affected by the Paleoproterozoic deformation. This is in strong contrast to the Ghost domain rocks on the east side of the shear zone where the zone of alteration is only a few tens of metres wide. As the zone is approached from the east, the rocks are first reddened and then a dull greenish grey, the mafic minerals become chloritized, and fractures are more abundant and are commonly filled with epidote group and carbonate minerals (Fig. 81). The granite eventually becomes disaggregated consisting of feldspar porphyroclasts in a fine-grained, grey, quartzofeldspathic matrix typically with no consistent foliation.

Discussion

The rather abrupt arrival at about 2.59 Ga of the several major bodies of relatively low-density granite into intermediate-level crust dominated by higher density, ca. 2.64 Ga tonalite and older gneiss units of supracrustal origin as represented by the rest of the Ghost domain, may have had important tectonic implications as far as the tectonic evolution of the area is concerned (Henderson, 1998b). If the granite units, particularly the megacrystic bodies, represent extensive sheets of unknown thickness, their arrival at about 2.59 Ga may have had a significant buoyant affect on that crust. Prior to that time, the intermediate crustal level rocks appear to have been under granulite-grade conditions. Most of the granite intrusions upon their arrival also crystallized under granulite-grade conditions. The monazite data from both the granite and the older tonalite, however, indicate passage of this crustal level into lower temperature conditions (about 700°C — the blocking temperature of monazite; Heaman and Parrish (1991)) at or immediately following the intrusion of the granite (Appendix: Figure A-1, part 9). Much of this uplift may have initiated and taken place along the extensive shear-zone system in the area; some of these shear zones define the boundaries between several of the domains.

Undivided intrusions of the Ghost domain (unit ĀG)

Part of the Ghost domain, in particular north of the southwestern part of Ghost Lake, is underlain by a heterogeneous assemblage of granitoid and other rocks, which cannot realistically be separated at the present scale of mapping. These include smaller bodies of units already described, both granitoid and supracrustal rocks as well as other lithological units.

The boundaries of the unit are gradational and somewhat arbitrary in most cases in that they are established where the adjacent unit becomes dominant at more than a local level. In the case of contacts with the metasedimentary rocks this is fairly straightforward as the amount of intruded phases tends to be significantly less; however, for the granitoid units, such as the quartz diorite or tonalite gneiss, the proportion of intruded or included phases can be high and the break can be less obvious.

In general, the unit consists of two dominant, well foliated to gneissic lithologies with quartz dioritic to tonalitic rocks dominating in the northwestern part of the unit west of the Ghost River, and granodioritic to tonalitic rocks more common in the southeast. The quartz dioritic to tonalitic phases are similar in most respects to the quartz diorite–tonalite (unit Āqd). They are quite uniform at any given locality, but compositionally fairly heterogeneous on a regional scale. The granodioritic rocks do not have an analogous mapped unit, the closest perhaps being some of the gneiss of the tonalite gneiss (unit Ātgn). These rocks are pinkish grey to grey, leucocratic, and medium to fine grained. They occur as complex, wispy, very irregular gneiss; through well foliated, but weakly layered rocks; to more extensive, homogeneous, weakly foliated bodies.

Both lithologies are intruded by a variety of, in general, less deformed granitic phases that locally can be dominant. They vary from strongly deformed gneiss to weakly foliated rocks, in which the original igneous texture is still evident.

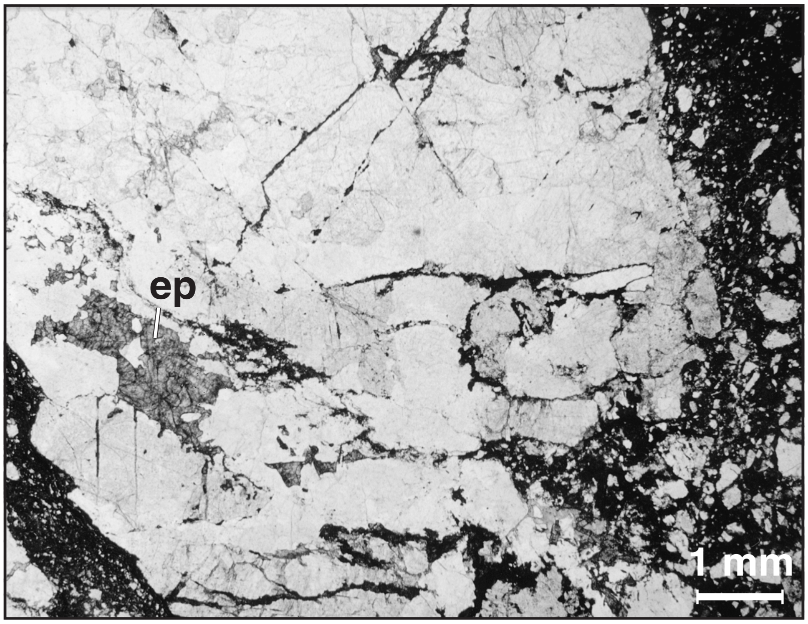


Figure 81.

Photomicrograph of Dauphinee domain granite about 100 m from the domain-bounding, cataclastic shear zone. Granite is broken up into angular clasts of highly varied grain size in a matrix of comminuted quartzofeldspathic material, chlorite, carbonate, and epidote group minerals. Plane-polarized light.

They have considerable compositional and textural variety, with perhaps the most common type similar to the still rather heterogeneous granite to granodiorite to pegmatite of unit Āg. Another fairly common phase is a sparsely megacrystic granite reminiscent of the Dauphinee domain granite suite (unit ĀDg). The granite typically occurs as concordant sheets a few centimetres to several tens of centimetres thick to less commonly several metres or more. Granitic rocks also occur locally as larger bodies of undefined extent and shape.

Also present in the unit, particularly in its southeastern parts, are thin, amphibolitic sills to trains of inclusions. They are typically on the order of 10 cm thick and form less than 10% of the rock where most abundant. They are assumed to be related to similar amphibolite sills (unit Āam) that are more common in the tonalite gneiss (units Ātgn and Ātgn.o) where they are locally of mappable extent. Thin units of metasedimentary migmatite along with inclusions are a common constituent of the unit. Rare metapyroxenite as angular inclusions in a leucocratic granite occur at several localities and may be related to the small units of metapyroxenite (unit Āum) at 'Hinscliffe Lake' and elsewhere.

A small body of similar, complex heterogeneous granitoid rocks and gneiss bodies occurs south of and within Daran Lake and is known to extend 2.5 km north of the area (Pehrsson and Kerswill, 1997b). It is possible these rocks are faulted remnants of either the undivided granitoid complex or the more leucocratic Hinscliffe complex; the greater heterogeneity and presence of some fairly mafic phases suggest the former may be the case.

DIATREMES

Two distinct types of diatreme, based on clast type, occur within the map area. An originally garnet-bearing type occurs within Yellowknife Supergroup metavolcanic rocks, one about 1 km south of west-central 'West Wijinnedi Lake' within dacitic metavolcanic rocks, and another 2 km west-southwest of 'West Wijinnedi Lake', just south of Snare River at the margin of the volcanic complex in meta-andesitic rocks (Map 2023A). Two examples of granite-clast-bearing diatremes occur in trondhjemitic rocks of the south-central Hinscliffe domain: 0.5 km southeast of 'Hinscliffe Lake' and the other near the west boundary of the map (Map 2023A).

The 'garnet-bearing' diatreme in the meta-andesitic rocks occurs in a body approximately 35 m long and 20 m across. It consists of a dark grey-green, altered matrix containing normally about 25%, but locally up to 50% clasts that were originally garnet, based on clast morphology, but now consist of a low-grade mineral assemblage. The clasts are pale pink to white, up to 8 cm in diameter, but are recrystallized to a very fine grain size and are not sorted within the diatreme. Although some retain the euhedral garnet crystal form, most are fairly rounded and equidimensional (Fig. 82, 83). None of the primary garnet has been preserved, but has been replaced by a fine-grained, heterogeneous assemblage of clinozoisite, white mica, plagioclase, carbonate, chlorite, and epidote in varied proportions. No other clast types were recognized in the diatreme. The dark matrix is also uniformly fine grained and consists of chlorite, clinozoisite, a fine-grained quartzofeldspathic aggregate, carbonate, and epidote. Some presumed original euhedral mineral outlines are evident suggesting a grain size of less than 1 mm. A mod-

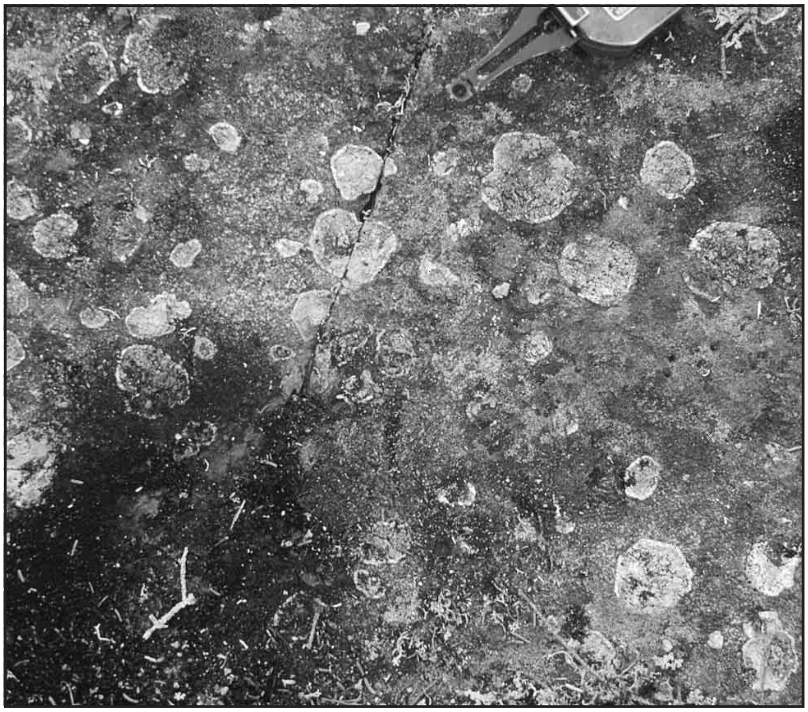


Figure 82.

Diatreme within dacitic, greenschist-grade Yellowknife Supergroup metavolcanic rocks containing coarse-grained, euhedral to anhedral, altered garnet. Brunton compass hinged sight 7 cm long. Photograph by S.E. Schaan. GSC 2002-417J fabric is evident wrapping around the coarse garnet crystals (Fig. 83). The second occurrence of 'garnet-bearing' diatreme in the dacitic metavolcanic rocks is of unknown extent, but is otherwise similar. The original garnet was out of context with the metamorphic grade of the matrix in which it presently occurs as well as that of the greenschist-grade volcanic rocks within which the diatremes are contained. This would suggest they formed at considerably greater depths under more extreme conditions.

The two diatremes in the Hinscliffe domain are quite distinct from those hosted by the volcanic rocks, but are themselves almost identical. They both occur in dyke-like bodies 2.5–3 m wide that have an easterly to somewhat north of easterly trend. The easternmost diatreme is associated with a prominent east-northeasterly trending linear. Both contain rounded to angular clasts of granite, mainly 10–20 cm, but as much as 80 cm that vary from being densely packed to sparsely distributed (Fig. 84). Two granite clast types are present with the most common being a pink, massive, coarse- to very coarse-grained, quartz-rich and mafic-poor granite of a type not recognized elsewhere in the area. Also present to a lesser extent, but more common close to the walls of the diatreme dyke are trondhjemite clasts similar to that of the Hinscliffe complex. The original igneous textures of the clasts are well preserved, but locally modified by the apparent

Figure 83.

Euhedral to anhedral, altered garnet in foliated mafic matrix. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416TT

Figure 84.

Granite-clast-bearing diatreme from the Hinscliffe complex. The angular to rounded clasts consist mainly of a coarse-grained, quartz-rich granite of a type not recognized in the region as well as some clasts of the local Hinscliffe complex trondhjemite. The mafic matrix contains a low-grade metamorphic mineral assemblage, but is not foliated as is the case in the garnet-bearing diatremes. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416UU

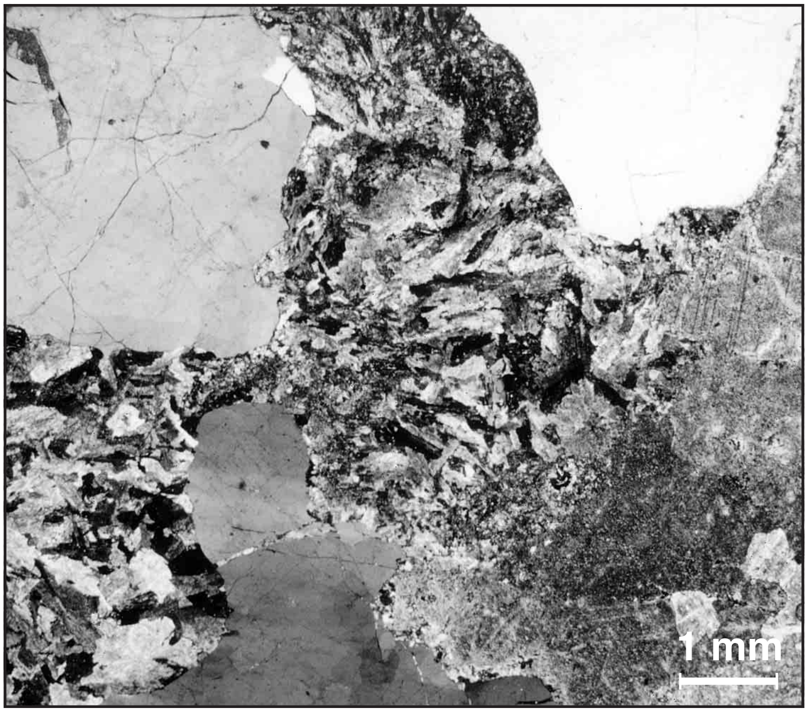


Figure 85.

Photomicrograph of altered granite clast in greenschist-grade mafic matrix. Note fine-grained, feathery intergrowth of the now recrystallized original K-feldspar of the granite. Cross-polarized light.

local penetration of thin seams of fine-grained quartz- and feldspar-rich matrix material along some of the igneous grain boundaries. The plagioclase is moderately altered, but the K-feldspar is recrystallized into fine, elongate to radiating to feathery aggregates of K-feldspar (Fig. 85). Any mafic minerals present are now chlorite. The dark green matrix is massive and fine grained, aside from coarser grains of feldspar and quartz that presumably represent minerals of disaggregated granite clasts. In the western occurrence, the matrix consists largely of altered clinopyroxene, little if any plagioclase, together with some anhedral to skeletal opaque phases which may represent primary minerals. Also present is abundant chlorite, actinolite, and carbonate. The eastern locality consists of a similar assemblage, but plagioclase greatly exceeds the altered clinopyroxene. Both diatremes are crosscut by Paleoproterozoic Indin dykes.

The two diatreme types are quite distinct and may well be unrelated. Both have largely lost their primary mineralogy which has in whole or in part been replaced by a low-grade metamorphic mineral assemblage. For the diatremes in the volcanic rocks this is compatible with the metamorphic mineral assemblage in the volcanic rocks they are contained in, which do not appear to have ever been at a higher metamorphic grade. Although the Paleoproterozoic Indin diabase dykes can also contain a similar low-grade metamorphic mineral assemblage (*see* 'Paleoproterozoic diabase dykes (units PM, PI)'), the fact that the matrix of the garnet-bearing diatreme is foliated may indicate the diatremes are Archean. The granite-clast-bearing diatremes in the Hinscliffe complex also contain a similar low-grade metamorphic assemblage, whereas the Hinscliffe complex rocks in which they are contained are at amphibolite grade, which would indicate their emplacement postdated Archean metamorphism. These diatremes, however, lack a tectonic fabric, which may suggest they could be Paleoproterozoic.

PALEOPROTEROZOIC DIABASE DYKES (UNITS PM, PI)

The Wijinnedi Lake area contains several of the Proterozoic dyke sets that commonly occur in the southwestern Slave Province. These include the Paleoproterozic Indin and possibly the Dogrib dykes as well as at least one example of a Mesoproterozoic Mackenzie dyke.

Indin diabase dykes

The most common dykes are the Indin dykes originally named by McGlynn and Irving (1975) and include the dykes of "Sets II" and "IV" of Burwash et al. (1963) and Leech (1966). These dykes are unusually abundant in the area, although only a few examples of the more prominent dykes are shown on Map 2023A. Many more are shown on Wright's map of the Ghost Lake area (Wright, 1954) and indeed, examples occur just about everywhere throughout the area. The Indin dykes occur in two sets, one with a north-northwesterly trend and the other with a northeasterly trend (Fig. 86). In the Wijinnedi Lake area the north-northwesterly trending dykes are more common (Fig. 86), although the reverse is the case in the Yellowknife area (Henderson, 1985). The dykes are abundant throughout most parts of the area. East of 'Hinscliffe Lake' where they are particularly abundant, Wright (1950a) suggested an minimum crustal extension of 10% over a distance of 1.5 km due to the emplacement of the dykes.

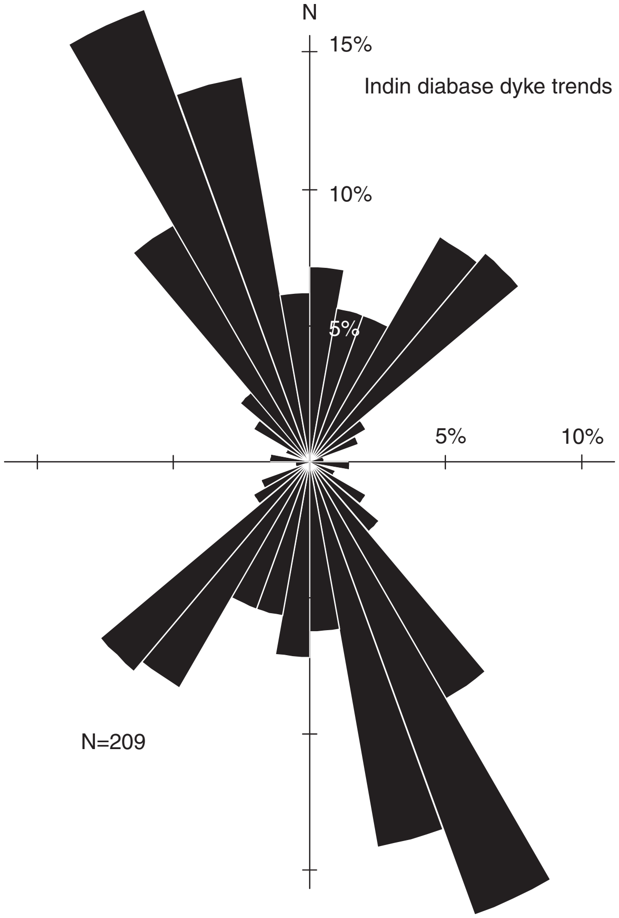


Figure 86. Rose diagram showing bimodal trends of 209 diabase dykes of the Indin set from throughout the map area.

Individual dykes are up to 90 m wide and can be traced over a distance of 20 km. Most commonly they are about 10 m thick. In contrast to most diabase dyke sets in the shield, the Indin dykes within the map area dip on average 82° to the east (Fig. 87). This is thought to be due to rotation of the Wijinnedi, Hinscliffe, and Ghost domain during Paleoproterozoic uplift of those domains relative to the Dauphinee domain, primarily along the major cataclastic shear zone that marks the eastern boundary of the Ghost domain. Only the widest dykes have a moderate aeromagnetic expression (Geological Survey of Canada, 1963). The dykes consist of black diabase that weathers a dull greyish green to reddish brown depending on the degree of alteration. They are most commonly medium to fine grained and even grained although some of the north-northwesterly trending dykes are variably porphyritic with plagioclase crystals up to 2 cm. The characteristic ophitic texture and chilled dyke margins are clearly evident in most examples. Almost all the dykes are massive, although a few of the north-northwesterly trending dykes have a weakly developed fracture cleavage filled with fine-grained alteration minerals in some cases. One or two examples of multiple dykes were noted. In one example, a dyke has a well developed, margin-parallel, compositional and textural layering at the margins of the dyke (Fig. 72). The graded, centimetre-scale layers become thinner closer to the dyke margin. Similar zonation was seen in a few of the Archean mafic sheets associated with the metagabbro bodies in the Yellowknife Supergroup felsic volcanic centre (Fig. 71). Inclusions in the dykes are rare, but in one dyke in the megacrystic granite south of Ghost Lake, a train of granite inclusions about 1 m from the margin of the dyke could be traced over a distance of about 800 m. Elsewhere, dacite inclusions up to 4 m occur. In one case, a dyke emplaced in low-grade Yellowknife Supergroup metasedimentary rocks, 3 km north of the sedimentary-volcanic contact, contained dacite inclusions.

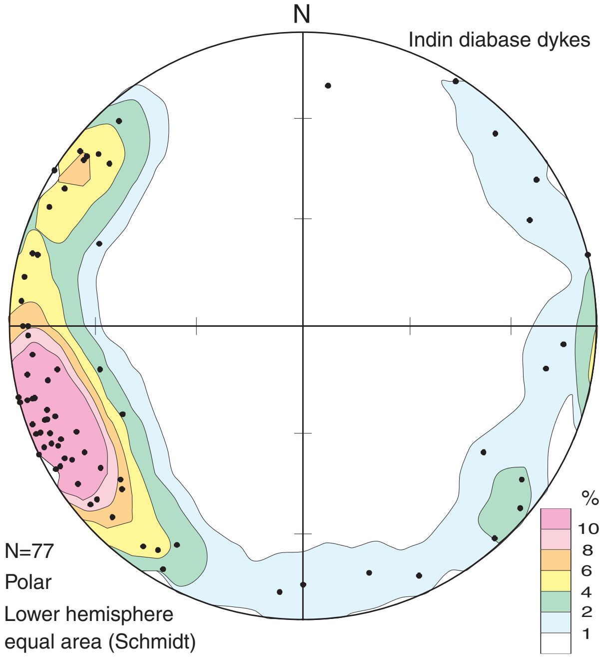


Figure 87. Stereogram of poles to Indin diabase dykes within the Wijinnedi Lake map area. The data suggests the dykes are steeply dipping in an easterly direction.

In most cases, the dykes consist of more or less equal parts plagioclase and clinopyroxene with opaque oxide minerals as a ubiquitous minor component. Olivine and minor biotite or their altered equivalents occur in most of the northeasterly trending dykes (Fig. 88) whereas some of the northwesterly trending dykes contain minor quartz. This would suggest the two trends are compositionally distinct. While most dykes show some degree of alteration, the northwesterly trending dykes, with few exceptions, are more altered than the northeasterly trending dykes. The mafic minerals are most affected, with the olivine altering to serpentine and opaque phases and then chlorite and more abundant opaque phases, whereas the pyroxene is replaced by actinolitic amphibole and chlorite (Fig. 89). The plagioclase is, on the whole, only minimally affected.

The northeasterly trending dykes are offset by the northwesterly trending Indin faults. The northwesterly trending dykes are parallel to the faults and indeed, in the Ranji Lake area to the north, they are coincident with the faults in several cases (Tremblay, 1948; Pehrsson and Kerswill, 1997b). This presumably explains the greater degree of alteration and local shearing in the northwesterly trending dykes. The dykes within the cataclastic shear zone seem to occur as large isolated blocks that, while fractured, are significantly less deformed than the cataclastic granitoid rocks in which the diabase blocks are contained, perhaps a reflection of the mechanical contrast of the two rock types under the particular deformational conditions. The only isotopic geochronological data available on the dykes comes from elsewhere in the Slave Province. They include K-Ar data suggesting an age of about 2100–2000 Ma (Leech, 1966) and Rb-Sr data suggesting a 2067 ± 45 Ma age for the northeast-trending set and 2174 ± 180 Ma for the northwest-trending set (Gates and Hurley, 1973).

Despite the abundance of the dykes, examples of intersecting relations are not commonly exposed. In four examples found in the course of mapping the area, the northeasterly set is apparently younger in three of the cases, a finding also noted by Frith (1993). Wright (1950a), on the other hand,

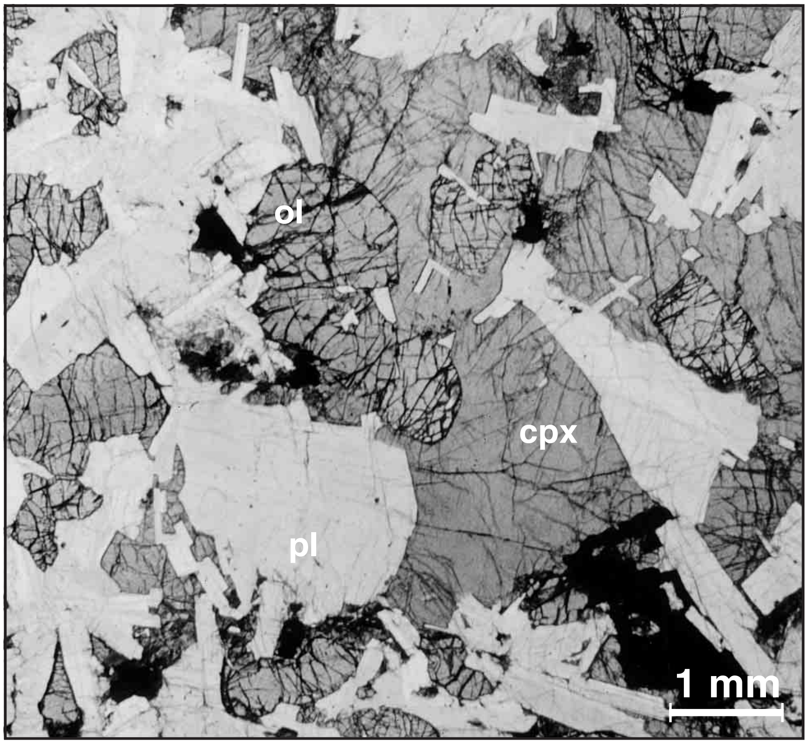


Figure 88.

Photomicrograph of northeasterly trending Indin dyke showing well developed, ophitic texture consisting of fresh plagioclase, clinopyroxene with lesser amounts of olivine, and minor biotite and opaque oxide minerals. Plane-polarized light.

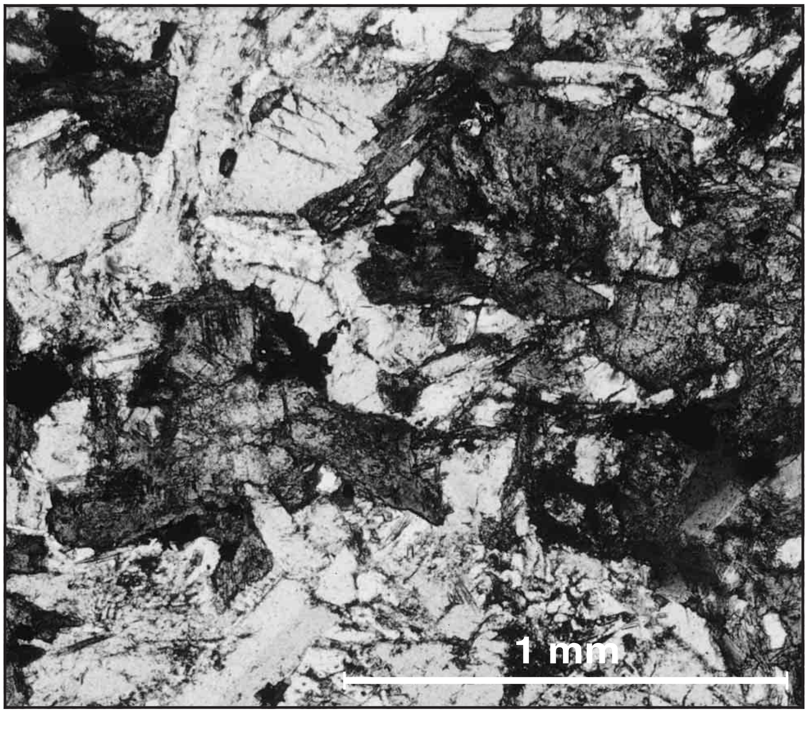


Figure 89.

Photomicrograph of representative north-northeasterly trending Indin dyke. These dykes are significantly more altered than the olivine-bearing northeasterly dykes as shown in Figure 88. There is no evidence of olivine in these dykes and clinopyroxene is in part altered to actinolitic amphibole and chlorite, whereas the plagioclase remains relatively clear of alteration products. Plane-polarized light.

illustrated the reverse ratio on his map and concluded that there is no consistent relationship. This is also the case at Yellowknife which led Henderson and Brown (1952, 1966) to suggest the two sets were likely contemporaneous. The similarity of their K-Ar ages led Leech (1966) to suggest the dykes of the two trends were conjugate, an interpretation also implied by Wright (1950a) on the basis of their pattern in the Ghost Lake–Ranji Lake area. The paleomagnetic directions for samples from the two sets are also not significantly different which led McGlynn and Irving (1975) to concur with this interpretation. More recent data led W.F. Fahrig and K.L. Buchan to suggest that the Indin dykes of both trends are retrograded and more than one swarm of pre-Mackenzie dykes may be present (W.F. Fahrig and K.L. Buchan, pers. comm. (1985) *in* Frith (1993)).

Other diabase dykes

In addition to the abundant northwesterly and northeasterly trending dykes of the Indin set, two other dyke sets are represented in the area. Wright (1954) showed a few east-northeast-trending dykes that are cut by both trends of the Indin set. These dykes have a similar trend to the Dogrib dykes of the Yellowknife area which are 2.19 Ga (LeCheminant et al., 1997). Also present in the area is a single Mackenzie dyke with a 350° trend and the very prominent magnetic expression characteristic of this dyke set. Mackenzie dykes have been dated elsewhere at 1267 Ma (LeCheminant and Heaman, 1989).

QUATERNARY GEOLOGY

The Wijinnedi Lake area was glaciated by the Laurentide Ice Sheet that reached its maximum extent in northwestern Canada from between 25 ka BP and 30 ka BP, but had retreated from the map area by about 10.0 ka BP (Lemmen et al., 1994). The area lies within the prominent 700 km long by 100 km wide zone along the eastern margin of the Canadian Shield south of Great Bear Lake that contains only minor Quaternary deposits (Fulton, 1995). This zone approximately corresponds to the slope west of the higher plateau

Figure 90. Airphoto of area east of 'Hinscliffe Lake' that is reasonably representative of the largely Quaternary deposit-free terrane in the central and western part of the map area. Several small areas underlain by low-relief sand plains may represent relict esker system after reworking by glacial Lake McConnell. Modified from *part of NAPL A8749-108*

that underlies most of the Slave Province. With the eastward retreat of the glacial ice through the region the short-lived, but extensive (up to about 215 000 km²) glacial Lake McConnell formed along the shield margin during the time the ice front passed through the map area (Lemmen et al., 1994).

The western two-thirds of the map area contains almost no glacial deposits, resulting in the superb bedrock exposure that characterizes much of the area (Fig. 90). The eastern (up-ice) part of the area is covered only by a thin, discontinuous, till blanket such that the local relief is basically that of the bedrock surface and through which bedrock faults, shear zones, and fracture patterns are for the most part still evident. Geomorphic forms such as drumlins, flutes, and roches moutonnées developed on the till are minor, if present at all. Sporadically collected glacial striae trends indicate a southwesterly transport sense (Fig. 91). This trend is at a low angle to the dominant structural trend northwest of Ghost Lake, resulting in the relatively high local relief in that area. The presence of more easily eroded rock associated with the Archean shear zone hypothesized to underlie the length of Ghost Lake may explain the presence of Ghost Lake itself.

Sandy glaciofluvial deposits occur locally, but are most abundant in the northeastern part of the area (Fig. 91). Most sand areas are rather flat and featureless, but in the eastern and

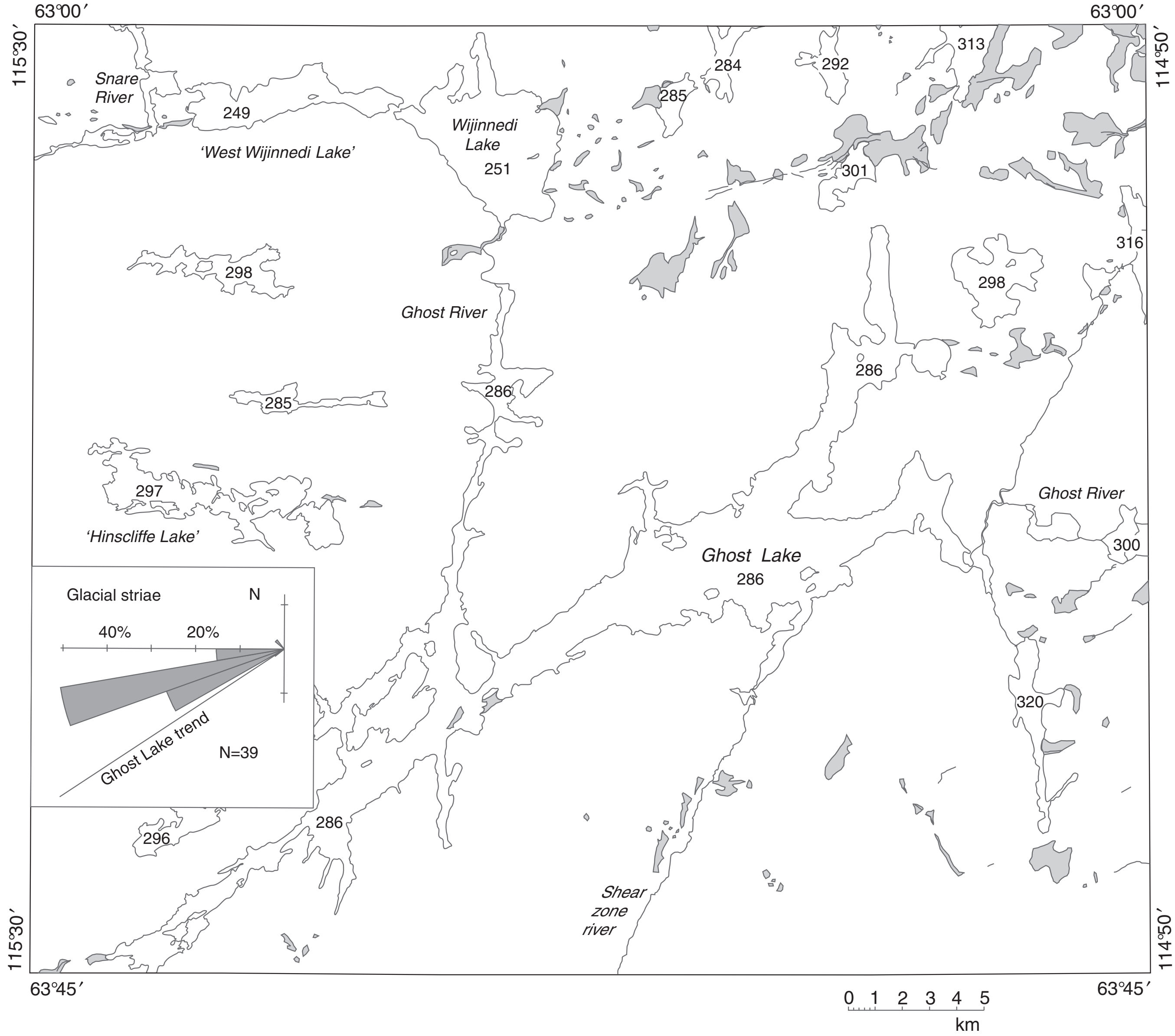


***Figure 91.** Glaciofluvial sand deposits in the Wijinnedi Lake area. Shaded areas represent low-relief sand plains while the black lines represent sharp-crested esker ridges that are oriented parallel to glacial striae in the area (inset) and occur on sand plains above 290 m elevation. Elevation of the major lakes indicated in metres. Text "Shear zone river" refers to an unnamed river in prominent valley following major shear zone.*

in particular the northeastern part of the area, sharp-crested esker-like ridges oriented parallel to the striae trend occur on the sand areas. The sharp-crested sand ridges occur at or above 290 m elevation which is Lemmen et al.'s (1994) estimate of the maximum elevation of glacial Lake McConnell. Thus, part of the eastern margin of the maximum extent of the lake passed through the map area. The shoreline itself must have been very irregular as much of the terrain in the western two-thirds of the area is above 290 m (e.g. Ghost Lake is at an elevation of 286 m). Any sharp-crested sand ridges that might have existed below that level were presumably reworked by the waters of the lake.

Parts of three fluvial systems appear to be represented by the remnant sand deposits. The most complete is the esker system that comes in through the northeast corner and drains out of the area through present-day Wijinnedi Lake and 'West Wijinnedi Lake' and the Snare River. A second system appears to follow the course of present-day Ghost Lake. The third system comes into the area south of the eastern Ghost River and is deflected to the south-southwest along the topographically low course of the valley associated with the Archean shear zone east of the large megacrystic granite (unit Āgm.o).

STRUCTURAL GEOLOGY

The Wijinnedi Lake map area has been divided into four lithotectonic domains, of which the three western domains are separated by major Archean ductile shear zones. The fourth, on the east side of the map area is separated from the others by a Paleoproterozoic cataclastic shear zone (Fig. 92). In general, the three western domains successively increase in metamorphic grade from northwest to southeast. Foliations vary from steep, dipping north or south in the northernmost domain, and become progressively shallower dipping in the domains to the southeast (Fig. 93). The rocks become steeply dipping again in the southwesternmost part of the Ghost domain. The major element in each of the three western domains is an anticline or antiformal structure. As only a small part of the Dauphinee domain, the easternmost domain of the map area, was mapped, little can be said of any regional structures within it other than the generally northern trend of map units, more evident in the northern part of the domain, that continues to the north (Fig. 6) and the presence of a varied, but generally northeasterly foliation trend that is steeply northwesterly dipping (Fig. 93).

Wijinnedi domain

The Wijinnedi domain mainly consists of Yellowknife Supergroup rocks, for the most part at relatively low metamorphic grade. The southern boundary of the domain, in the western part of the area, is clearly separated from the dominantly metagranitoid Hinscliffe domain by a concordant ductile shear zone up to 1.5 km wide that involves rocks of both domains. To the east, however, the metamorphic grade within the Wijinnedi domain increases and the granitoid Hinscliffe domain is pinched out between its bounding shear zones. This leaves little lithological or metamorphic contrast between the metasedimentary rocks of the Wijinnedi and adjacent Ghost domains. The domain boundary becomes somewhat more arbitrary and is taken as the north-northeasterly extension of the now-joined bounding shear zones of the Hinscliffe domain. These are joined to the north by a subsidiary, but presumably related shear zone from within the Wijinnedi domain and all continue to the north through Daran Lake and beyond the map area as a single, north-trending, amalgamated shear zone for at least 50 km (Fig. 6; Pehrsson and Kerswill, 1997a, b).

The dominant structural trend in the Wijinnedi domain is easterly, expressed in both the metavolcanic and metasedimentary rocks. This easterly trend contrasts with the more north-northeasterly to northeasterly trends seen in similar rocks north of the map area (Fig. 6; Frith, 1993; Pehrsson and Kerswill, 1997a, b).

In the eastern part of the domain where the rocks are increasingly more highly metamorphosed, the trends in both the metavolcanic and metasedimentary rocks swing to the north-northeast. This pattern also is seen in the adjacent rocks of the Ghost domain immediately to the east where the normally northeasterly structural trends of the Ghost domain become northerly trending east of the Wijinnedi domain. This flexure may be related to the transition of the trends of the several Archean shear zones of the area from easterly and northeasterly to northerly as they leave the map area, in large part through Daran Lake.

The volcanic rocks occur in an elongate antiformal dome-like structure consisting mainly of dacitic volcaniclastic rocks mantled by more mafic volcanic rocks that are most abundant at the eastern end of the dome. The dome closes to the west about 5 km beyond the map area (Fig. 6; Lord, 1942; Jackson, 2003). The anticlinal form is based on top determinations from local pillowed mafic volcanic flows on the north side of the structure in the vicinity of Wijinnedi Lake and on the overall symmetry of the structure, since the rocks on the south side are too deformed for preservation of primary structures. At outcrop scale, primary bedding units that are traceable over any significant distance in the felsic volcaniclastic rocks were rarely, if ever, identified. Possible large-scale bedding units, however, are locally visible on airphotos and would suggest that stratigraphic units within the overall domal structure are folded at a smaller scale (Fig. 94). The foliation in the main part of the dome is easterly trending and moderately to steeply dipping to the north on the southern side of the dome, whereas on the north side of the dome the dips are steeper with some southerly directed (Fig. 93). In most places, the foliation is defined by the flattening of breccia clasts which are also commonly elongated down-dip (Fig. 9, 10, 92).

Both stretching and mineral lineations in the main part of the volcanic complex are common (Fig. 92). They are moderately to steeply southerly plunging on the north limb of the complex and more moderately northerly plunging on the south limb. Where associated with the domain-bounding shear zones, the lineations are northeasterly directed with a more moderate plunge. Lineations abruptly become a much more prominent feature at, and east of, the fault along the Ghost River south of Wijinnedi Lake where they are strongly northerly directed with a moderate plunge. It is not clear if these lineations are related to the northerly bend in the volcanic rocks in this area or possibly the fault along the Ghost

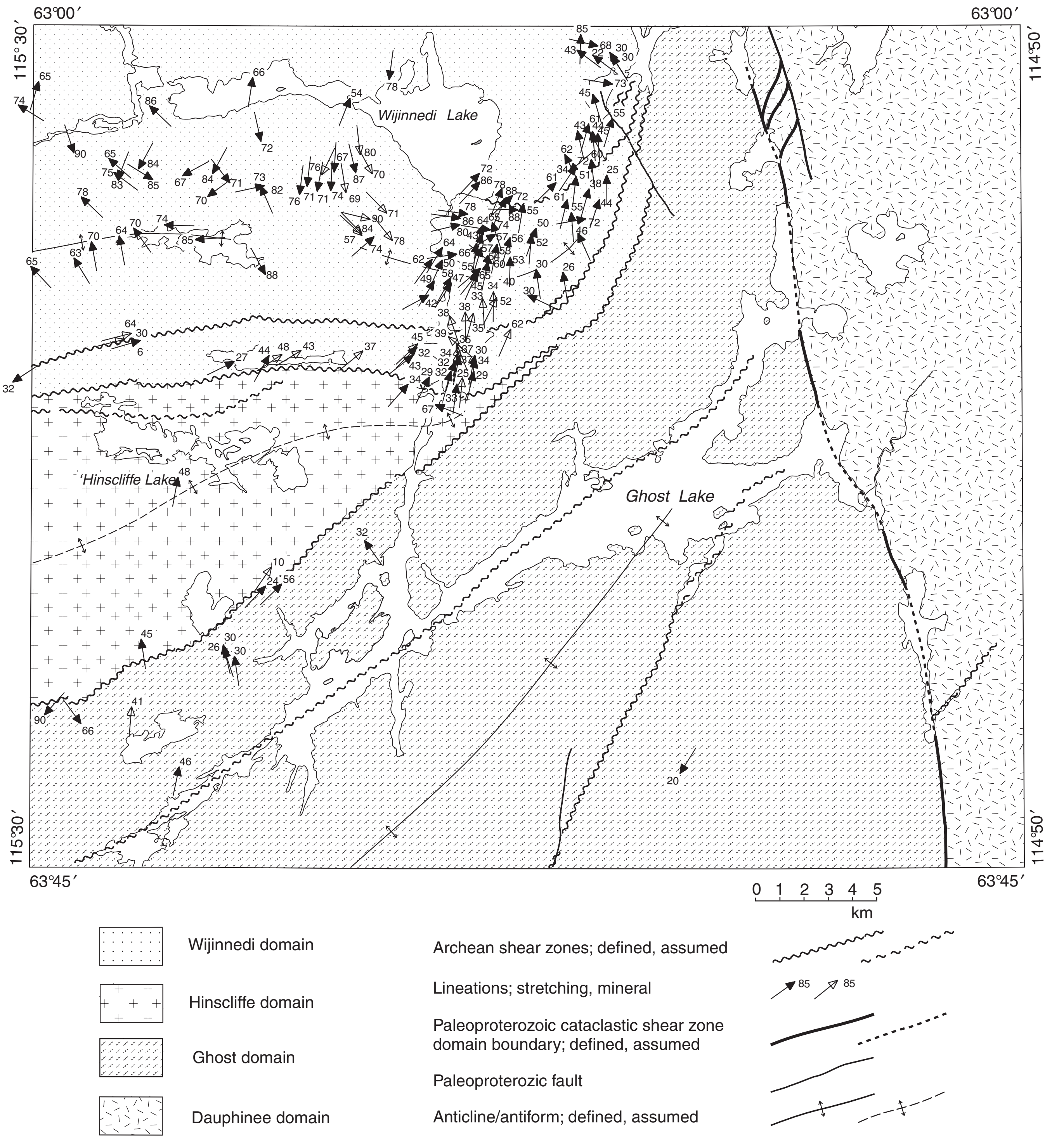


Figure 92. *Archean ductile shear zones separate the Wijinnedi, Hinscliffe, and Ghost domains that represent successively deeper structural levels. A major Paleoproterozoic cataclastic shear zone separates the Dauphinee domain from the Ghost domain. Moderately plunging stretching and mineral lineations are concentrated in the eastern Wijinnedi domain where the structural trends change from easterly to northerly.*

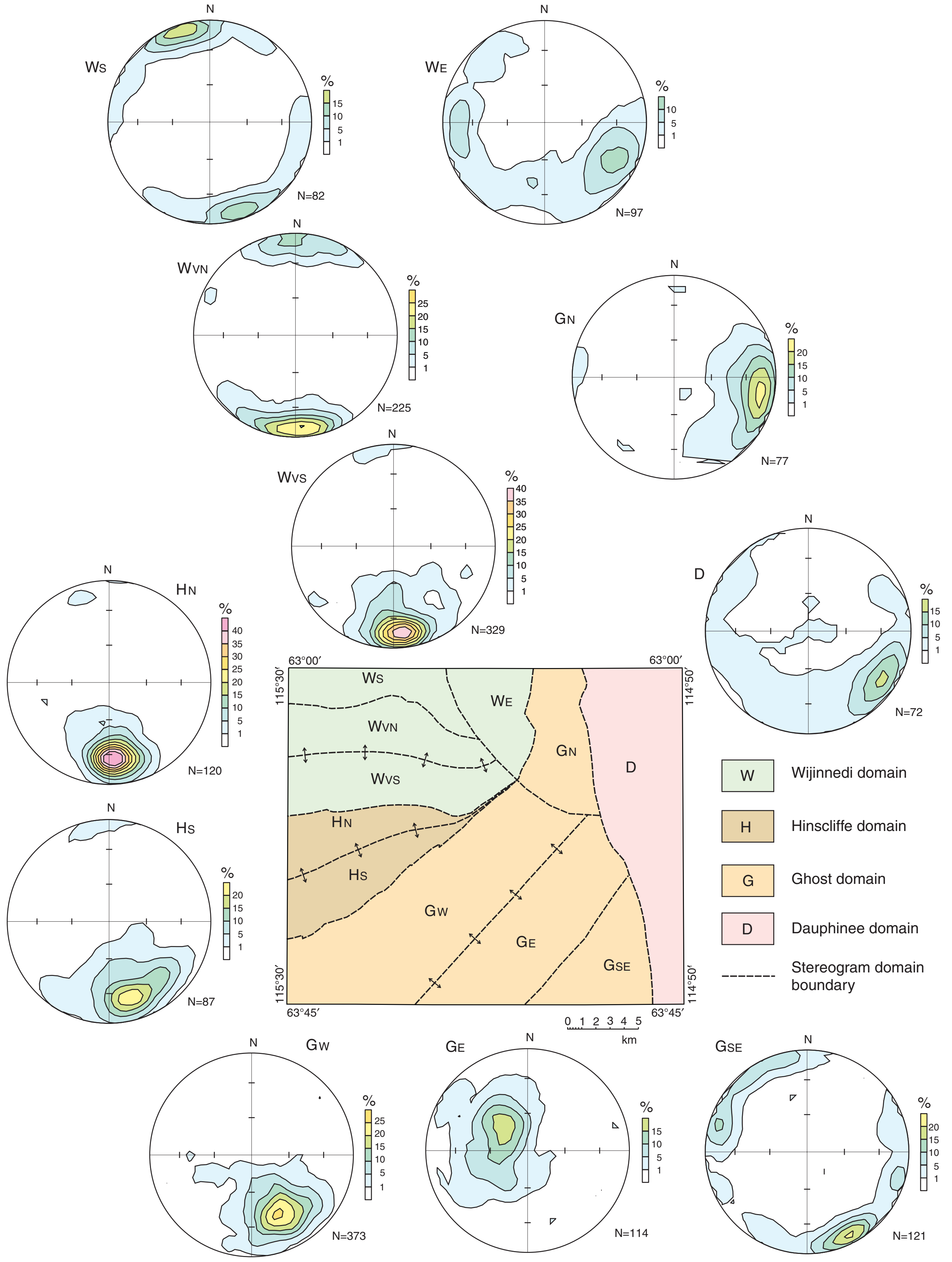


Figure 93. *Stereograms of principal foliation data for various domains and subdomains of the Wijinnedi map area. Contours are expressed as a percentage and were calculated by the Starkey fixed-circle counting method in Spheristat 2, a stereonet plotting program (Stesky, 1998).*

Figure 94. Airphoto of part of the intermediate to felsic dome southwest of 'Glazebrook Lake' showing probable smaller order fold closures within the volcanic dome (arrow). Fine white lines indicate contacts between mainly felsic metavolcanic rocks (AYv) and/or metasedimentary rocks (AYs) and/or metagabbro (Amg). The heavy white line represents a fault while the wiggly line represents a shear zone. Modified from *part of NAPL A8749-15*

River and other related faults that occur where the lineations are most dense. These faults are thought to be related to the domain-bounding cataclastic shear zone to the east, although no lineations associated with this particular structure were recognized. Lineations are by far much more abundant in the Wijinnedi domain than in the other domains.

In the sedimentary terrane to the north of the volcanic dome, the trend of bedding is east-northeasterly west of Wijinnedi Lake and bends around to northerly east of the lake. This a similar pattern to that seen in the volcanic rocks to the south, although the bedding trends in the sedimentary rocks have a somewhat more northerly component in their trend. The trend of the principal foliation is similar, on average, to the bedding in both situations, although at somewhat more than half the locations observed there is a greater than 10° discordance between the two. Layering in the metasedimentary rocks is steeply dipping to both the north and the south. As is commonly the case with the metagreywacke-mudstone turbidite units elsewhere in the Slave Province, the rocks are tightly to isoclinally folded as indicated by reversals in bed facing directions that can occur over a distance of a few tens of metres.

The easterly trending folds and the strongly developed parallel-trending principal foliation, presumably representing the axial-planar foliation to the folds, are the oldest structures recognized in the relatively small area of metasedimentary rocks mapped. The only other major fold structure recognized in these rocks is the bend east of Wijinnedi Lake that results in the change in trend of both the volcanic and sedimentary rocks. Both primary layering and the principal foliation are involved in this bend. This structure, however, appears to be unique and is not part of a regional fold pattern. Younger structures include minor cleavages and crenulations (Fig. 95, 96) that may be correlative with other deformational generations recognized in similar rocks elsewhere in the region and throughout the Slave Province (Jackson, 1989; Bleeker and Beaumont-Smith, 1995; Pehrsson and Kerswill, 1997b). Textural relations suggest that development of the principal foliation was synchronous with, and to some extent succeeded, the peak of metamorphism. Cordierite porphyroblasts in a matrix of finer grained quartz, feldspar, and aligned biotite and muscovite commonly contain randomly oriented biotite and muscovite. In some cases, however, the mica minerals within the porphyroblast are also aligned. Randomly oriented, coarse-grained biotite, ragged in outline, overprints the aligned matrix in many cases (Fig. 97). Inclusions or trains of inclusions within the biotite parallel the older matrix fabric.

Hinscliffe domain

The Hinscliffe domain, south of the Wijinnedi domain, is a distinct lithological entity consisting mainly of deformed and metamorphosed trondhjemitic to granodioritic intrusions

Figure 95.

Chevron cleavage pattern in biotite-grade, graded, fine-grained, metagreywacke-mudstone couplets. The more prominent cleavage is evident in more pelitic material in the lower half of the photograph while the more subtle cleavage that is parallel to the 8 mm wide mechanical pencil occurs in less pelitic material in the upper half of the photograph. Photograph by J.B. Henderson. GSC 2002-416VV

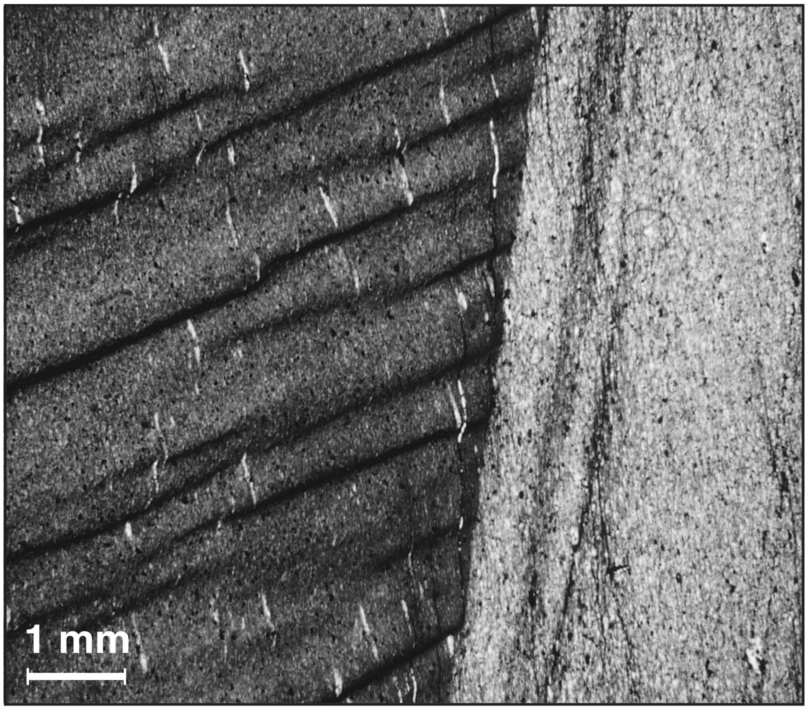


Figure 96.

Thinly bedded, chlorite-zone metasiltstone interlayered with metamudstone with earlier cleavage at a low angle to bedding and crenulations of that cleavage prominently developed in the mudstone. Colour version of photomicrograph appears on cover. Plane-polarized light.

with varied amounts of mafic synplutonic dykes, now represented by zones of inclusions. On the basis of the mineralogy of the mafic inclusions throughout the domain, the complex as a whole is at amphibolite grade and so is at a similar or higher metamorphic grade than the adjacent Wijinnedi domain. It is bounded both to the north and south by Archean ductile shear zones that join together at the east end of the domain, where they turn to the north and continue beyond the area.

The least deformed rocks occur in the core of the domain where intrusive contacts between igneous phases differing slightly in composition are evident. A relatively tightly focused easterly trending foliation with moderate to steep northerly dips is developed to the north (Fig. 93). The foliation is mainly defined by the orientation of biotite, and to a lesser extent, by the weak orientation of quartzofeldspathic phases. The foliation is parallel to compositional layering where the trondhjemite bodies are gneissic. To the south, the otherwise similar foliation has a wider ranging easterly to east-northeasterly trend with generally shallower, but more varied northerly directed dips (Fig. 93). The foliation becomes increasingly pronounced toward the margins of the domain as the rocks change from massive, through foliated, to gneissic granitoid rocks, to eventually mylonite at the domain margins where the foliation grades into the domain-bounding shear zones.

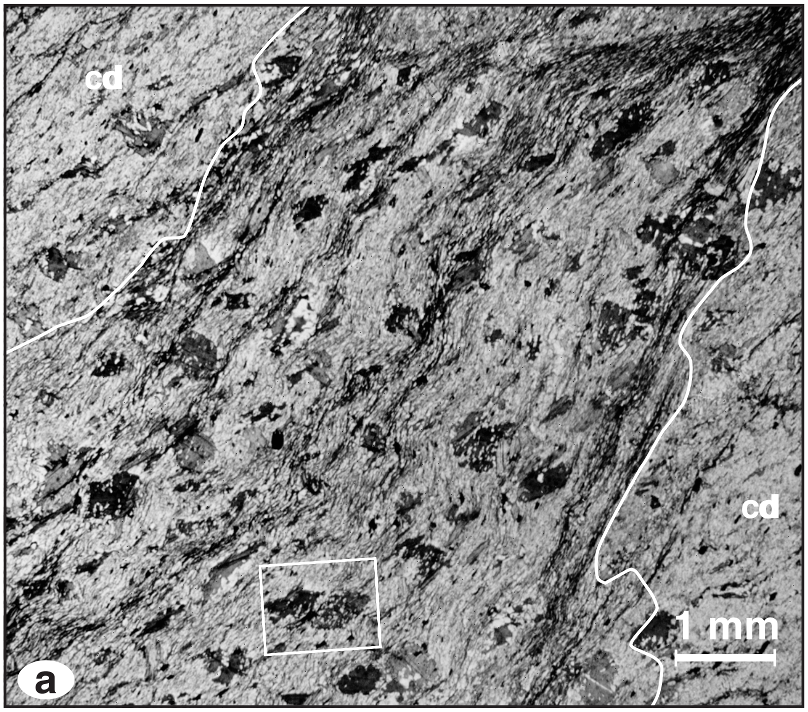


Figure 97.

***a**) Sillimanite-zone metapelite in which cordierite (cd) overgrows strongly oriented fine-grained biotite and muscovite. The strongly oriented biotite and muscovite is also overgrown by coarser grained biotite. Both the fine-grained biotite and muscovite and the younger, coarser grained biotite are deformed in a late crenulation. Plane-polarized light.*

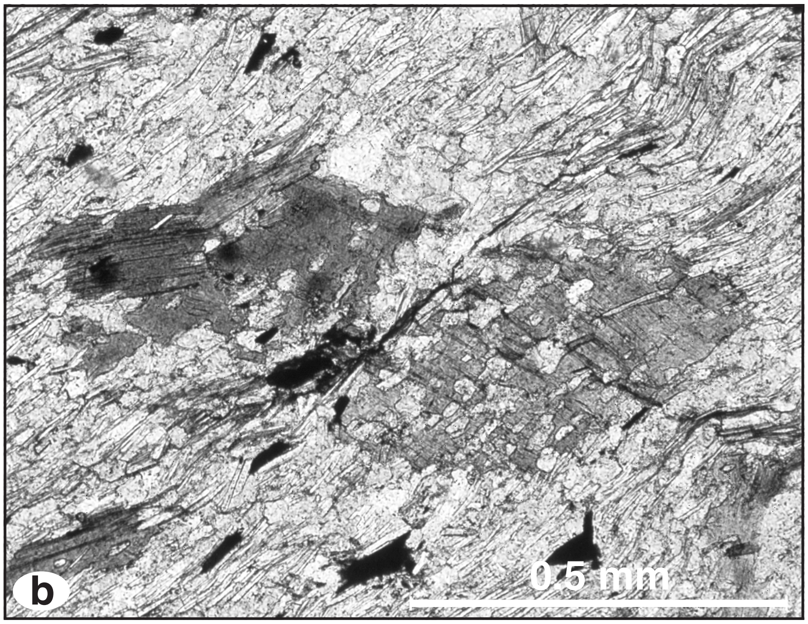


Figure 97.

***b**) Detail of area outlined in a. Plane-polarized light.*

No evidence of younger, crosscutting foliations was recognized. Small intrusions of granite within the Hinscliffe complex that are similar in many respects to granite unit Ag and considered to be possibly correlative with it, do not contain the principle foliation. The only evidence of deformation recognized that predated the principal foliation are the variably deformed granitoid phases that crosscut some of the mafic inclusions (Fig. 47). These earlier deformations would appear to be synchronous with at least some stages of trondhjemitic igneous activity, in contrast to the principal foliation which postdated all but the latest granite event.

Along the north margin of the domain are a series of layers ranging in composition from metagabbro to metadiorite. As with the trondhjemite, they become increasingly deformed to the north. Where best preserved they appear to be a suite of igneous sills that, in general, are somewhat less deformed than the granitoid rocks in which they occur. These rocks are discussed more completely in the section on the Hinscliffe complex. Along the southern margin of the complex, fine-grained, parallel-layered, mylonitic rocks of similar composition, commonly occur as inclusions within the less deformed trondhjemite (Fig. 43). These rocks are not found in the interior of the complex. Although the reason for the significance of these rocks is not understood, on the basis of the symmetry of their distribution at the margin of the domain, it is suggested that the domain itself may represent an overturned antiformal structure.

Ghost domain

The dominant trend in the Ghost domain and, to a somewhat lesser extent, in the adjacent Dauphinee domain, is northeasterly. This is a common trend seen in the western to southwestern Slave Province. It is particularly well developed in the western Point Lake area as evident in both aeromagnetic and geological maps (Geological Survey of Canada, 1980; Easton, 1981; Henderson, 1998a). A similar trend is also well developed in the rocks between Yellowknife and the western margin of the Slave Province (Lord, 1942; Henderson, 1985). Parallel trends are also apparent in deep seismic anisotropy measurements at Yellowknife (Silver and Chan, 1988) although Bank et al. (2000), on the basis of a teleseismic reconnaissance throughout the Slave Province, found similar trends even where surface rocks have regional northerly and northwesterly trends. Bank et al. (2000) suggested the seismic orientation is related more to present-day motion of the North American Plate than ancient structural trends.

The Ghost domain is the highest metamorphic grade domain, with all but a 4 km wide zone along the northwestern margin of the domain at granulite grade. Like the Wijinnedi domain, and possibly the Hinscliffe domain, the Ghost domain is dominated by a major antiformal structure that is clearly evident in the southern part of the domain, where it is cored by a large body of megacrystic granite (Fig. 2). A similar megacrystic granite occurs north of the east end of Ghost Lake suggesting the structure, which presumably has a near-horizontal fold axis, may plunge shallowly under Ghost Lake, re-emerging north of the lake. The overall symmetry of the domain is further expressed by the distribution of metasedimentary rocks which are dominant along both the northwesterly and southeasterly parts of the domain. As well, the major intermediate migmatitic metavolcanic unit northwest of Ghost Lake may be represented on the southeast limb by the thin, discontinuous felsic metavolcanic rocks adjacent to the large body of metasedimentary migmatite.

Foliations in the northwest limb of the fold, on the whole, dip moderately to the northwest, although they are quite varied (Fig. 93). Foliations in the southeast limb are similarly varied, but generally have a shallow to moderate dip (Fig. 93). In the southeastern part of the domain, dips are mainly steeply north or south and are more similar to the low-grade Wijinnedi domain (Fig. 93).

Smaller scale folds may also be present in the domain. The high metamorphic grade, felsic metavolcanic unit to the west of the north end of Ghost Lake is bounded in part on both sides by, and in some cases appears to grade into, metasedimentary migmatite bodies. If these migmatitic felsic volcanic rocks in the Ghost domain represent a deeper level equivalent of the felsic volcanic-dominated dome in the Wijinnedi domain that there underlie the metasedimentary rocks, the high-grade volcanic rocks may occur in the core of an anticline. By a similar argument, the several kilometre wide zone of metasedimentary migmatite in the southeast corner of the domain, which is in part bordered by thin discontinuous units of migmatitic metavolcanic rocks, may represent a synformal structure. On the south side of Ghost Lake, east of the large megacrystic granite body, the pattern of thin metasedimentary and amphibolitic units within the tonalite gneiss suggest a possible synformal structure with a moderately dipping axial plane, cored by a prominent shear zone. The symmetrically disposed cryptic shear zone along the northwest part of Ghost Lake (discussed in the next section) could conceivably occupy the sheared axial plane on another synclinal structure.

Archean shear zones

The three western domains are bounded by, and in most cases contain within them, major through-going ductile shear zones to mylonite zones (Fig. 92). These major structures are up to several hundred metres wide (up to 1 km for parts of the shear zone between Wijinnedi and Ghost domains). The rocks within a given shear zone were deformed under the similar metamorphic conditions as those represented by the rocks outside the shear zone. Because of this, together with the fact that the Paleoproterozoic diabase dykes are not offset by them (Wright, 1954), the shear zones are considered to be Archean. In some cases, a low-grade retrogression is associated with some of the shear zones. This is particularly evident in the shear zone within the granulite-grade rocks south of the main part of Ghost Lake, where orthopyroxene is pseudomorphed by chlorite minerals. This low-grade mineralogy is similar to that seen in the weakly altered Paleoproterozoic diabase dykes and, as a result, the retrogression in the shear zones is considered to be Paleoproterozoic.

The easterly to northeasterly trending shear zones within and bounding the Wijinnedi and Hinscliffe domains become more northerly trending. They join to become a single shear zone in Daran Lake where they are truncated by the

Paleoproterozoic cataclastic shear zone just beyond the north border of the map area. These shear zones continue over 50 km to the north where they separate the relatively low metamorphic grade, Yellowknife Supergroup supracrustal rocks to the west from the largely granitoid intrusions and high-grade granitoid gneiss that locally contain retrograded granulite mineral assemblages of the Cotterill complex and migmatite-grade Yellowknife Supergroup equivalents in the east (Fig. 6; Pehrsson and Kerswill, 1997a, b). Within the map area the shear zones separate domains that progressively increase in metamorphic grade from northwest to southeast, although no metamorphic jump was recognized across any given shear zone.

Of the two domain-bounding shear zones, the shear zone between the Wijinnedi and Hinscliffe domains is the most prominent. It is by far the widest zone, up to 1 km wide, and is usually well exposed with a knife-sharp contact between the finely laminated amphibolite of the Wijinnedi domain and almost as finely laminated mylonitic trondhjemite of the Hinscliffe domain (Fig. 36). Foliation attitudes at and near this shear zone are much more tightly focused than elsewhere (Fig. 93). The contact between the Hinscliffe and the Ghost domains, on the other hand, is typically represented by a 10 m covered zone, the recessive nature of the contact presumably being related to ductile shearing along it. The first metasedimentary migmatite units of the Ghost domain to the south are not more strongly foliated than normal. The Hinscliffe domain granitoid rocks, however, become increasingly foliated to gneissic toward the boundary over a distance of several tens of metres to 100 m, particularly in the central and eastern parts of the shear zone. In the west, a late intrusive phase, similar in composition to much of the Hinscliffe domain, is more prominent. These rocks, while foliated, are much less deformed than the Hinscliffe domain rocks to the east and contain within them deformed inclusions of parallel-layered, fine-grained, mylonitic, intermediate to amphibolitic gneiss (Fig. 43) that are similar to the sheared gneiss units at the boundary toward the eastern end of the domain.

Very sparse lineation data suggests moderate to shallow plunges in the vicinity of the shear zones and remote from the northerly bend. In the area of the bend, lineations are more abundant and more steeply plunging as is the case in the interior of the domains (Fig. 92).

A prominent Archean shear zone also occurs within the Ghost domain south of Ghost Lake and east of the large megacrystic granite intrusion. It is discordant to metasedimentary and amphibolitic layers in the adjacent tonalitic gneiss (Map 2023A). It would appear to disrupt the core of a steeply inclined, synformal structure defined by shallowly dipping, southeasterly to southerly trending foliations and metasedimentary migmatite and amphibolite on the west side, and more moderately dipping, southwesterly trending foliations and lithological units on the east. The shear zone is recessive relative to the adjacent tonalite and rocks within the zone are rarely exposed. Its trace is coincident with a prominent magnetic anomaly (Fig. 5, 7). Granulite-grade mineral assemblages are retrograded within the immediate vicinity of the shear zone relative to those in the surrounding tonalite. A somewhat smaller positive magnetic anomaly underlies the length of the main part of Ghost Lake and it, along with the magnetic anomaly east of the megacrystic granite, are symmetrically disposed about the main antiformal fold of the domain. The fact that the Quaternary glacial transport direction is subparallel to the anomaly (Fig. 91) and the feature is now covered by the long, narrow length of Ghost Lake might suggest the presence of a more easily eroded zone. It is hypothesized that this anomaly may represent another shear zone similar to the recessive structure to the southeast.

Cataclastic shear zone

The eastern margin of the largely granulite-grade Ghost Lake domain is defined by a major cataclastic shear zone. This structure is marked by a prominent, low aeromagnetic expression (Fig. 5, 7), is up to 800 m wide, and has no through-going fabric. Small-scale brittle faults with short offsets are common in the vicinity of the cataclastic shear zone (Fig. 98). Rocks within the main structure are extensively fractured, chloritized, and hematized with epidote-filled fractures and carbonate-cemented breccia present locally, all suggesting deformation under low-grade metamorphic conditions (Fig. 99). In places, nearby minor shears and fractures are steep, suggesting a similar orientation for the main zone. The shear zone is asymmetric in that orthopyroxene is present in Ghost domain rocks within 50 m of the most intensely deformed part of the shear zone, whereas on the east side, the alteration of mafic minerals to chlorite, associated with the shear zone, extends over several hundred metres.

The cataclastic shear zone locally contains blocks of diabase that are presumably part of the Paleoproterozoic Indin diabase dykes. The only isotopic age data currently available on the Indin dykes consists of K-Ar data suggesting an age between 2100 Ma and 2000 Ma and Rb-Sr data at 2067 ± 45 Ma (Leech, 1966) and 2174 ± 180 Ma (Gates and Hurley, 1973), and so may be part of the major, relatively brief episode of diabase dyke emplacement in the Slave Province ca. 2230–2210 (LeCheminant and van Breemen, 1994). A mid-Paleoproterozoic age, probably after 2.21 Ga, for the cataclastic shear zone is indicated.

The cataclastic shear zone continues south of the map area for about 55 km based on its magnetic expression (Fig. 5). Its trend gradually changes from southerly to southwesterly. To the north, it continues with a more northwesterly trend another 40 km through to Indin Lake (Fig. 6). There, a west-side-up sense of movement on the structure is suggested by the significantly wider zone of greenschist-grade metamorphism in the metasedimentary rocks northeast of the structure compared to the much narrower zone in similar rocks to the southwest. This, together with the contrast in metamorphic grade across the structure within the map area, the apparent continuation of the zone of steeply dipping, migmatitic metasedimentary rocks in the Ghost domain across into the Dauphinee domain, as well as the smooth change in trend of the structure throughout its length, suggest movement across the shear zone has been largely vertical with only minor, if any, horizontal component. The cataclastic shear zone, then, marks the eastern margin of an uplifted block, in part fault-bounded, the higher grade central and southern part of which is outlined by a prominent high-relief magnetic pattern. This uplifted block is coincident with one of the longer and steeper gravity gradients in the south-western Slave Province (Fig. 5).

Paleoproterozoic faults

Paleoproterozoic brittle faults occur throughout the area with most showing apparent displacements of a few tens of metres to a few hundreds of metres. The most common trend in the southern and southeastern part of the area is northerly (Fig. 100). The trend of many of the faults change to a north-northwesterly direction in the central and northern parts of the area. Although most of the faults in the area have an apparent sinistral offset, exceptions are not unusual. The amount of apparent offset can vary along the length of the fault and, in some cases, is even reversed, suggesting a scissors-type motion and a significant vertical component. For example, offset relations along the northerly trending fault at the end of the large megacrystic granite south of Ghost Lake suggests a significant west-side-up vertical component. On the other hand, in the vicinity of 'Colson Lake' a jump in the

Figure 98.

Small-scale, brittle faults in Archean migmatitic granitoid gneiss related to deformation associated with main cataclastic shear zone. Location is within a few tens of metres of the main structure. Knife is 9 cm long. Photograph by J.B. Henderson. GSC 2002-416WW

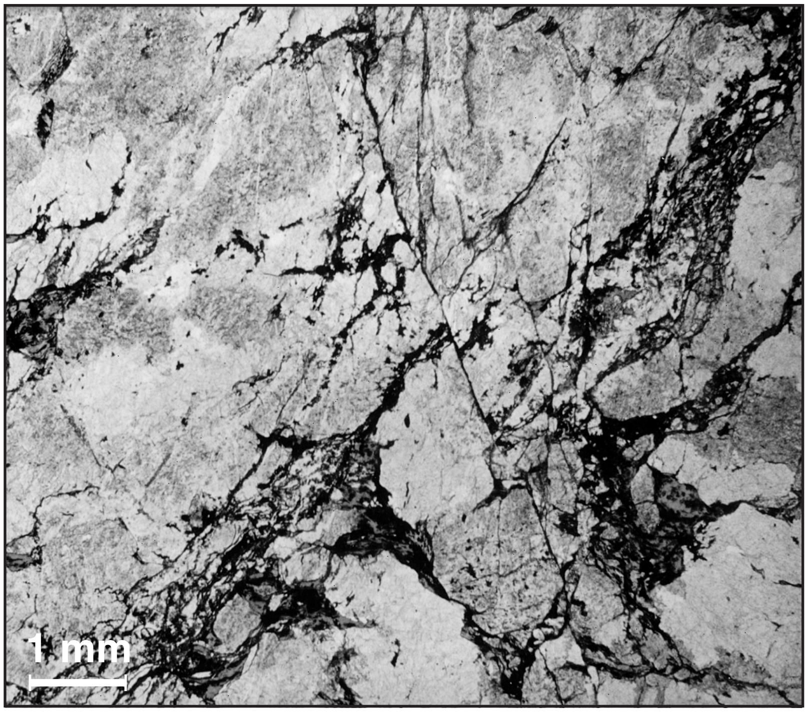


Figure 99.

Photomicrograph of cataclastically deformed, originally orthopyroxene-bearing granite in which all igneous minerals are altered to chlorite and brittly fractured quartzofeldspathic minerals. Plane-polarized light.

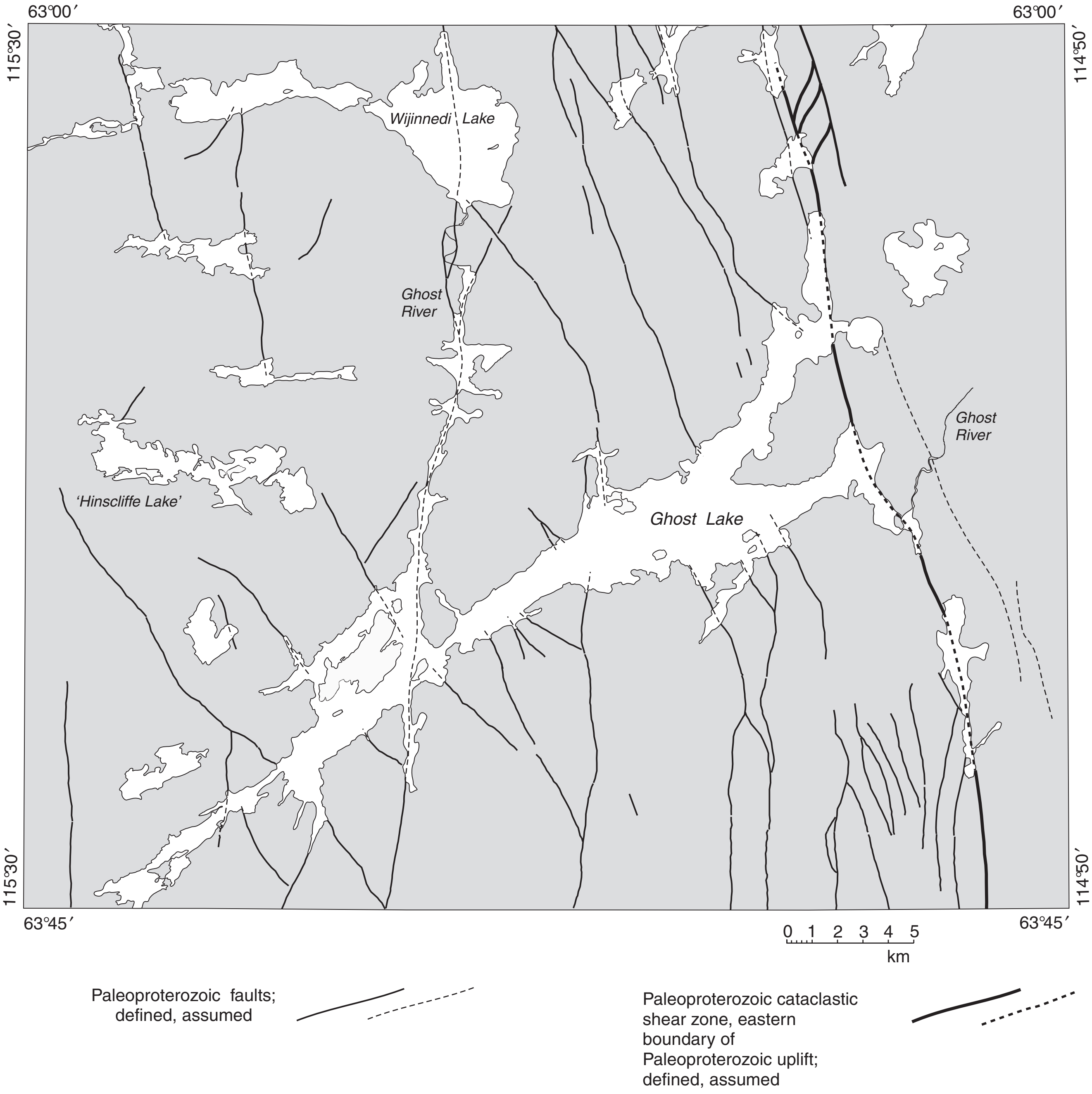


Figure 100. *Paleoproterozoic brittle faults and cataclastic shear zone within the map area in which the faults are northerly trending in the southern part of the area, but swing into a more northwesterly trend in the north.*

metamorphic grade across a fault suggests a significant east-side-up vertical component. Faults with the north-northwesterly trend are more abundant north of the area and have much greater sinistral offsets (Frith, 1993; Pehrsson and Kerswill, 1997a, b). Faults with similar trends and displacement sense are also common in the Yellowknife region (Brown, 1955; Henderson, 1985). The prominent fault that is followed by the Ghost River and passes through Wijinnedi Lake is an exception to the normal pattern in that its northerly trend continues throughout its length until it joins the splayed northern extension of the cataclastic shear zone 12 km north of the map area (Pehrsson and Kerswill, 1997b).

Discussion

The major phases of deformation in the Wijinnedi Lake area took place during two distinct periods, one during the Archean and the other in the Paleoproterozoic.

The Wijinnedi, Hinscliffe, and Ghost domains, as exposed at the present erosion level, represent sequentially deeper crustal levels from north to south. The juxtaposition of these domains took place during the Archean, primarily along the domain-bounding shear zones while the rocks were under elevated metamorphic conditions. As discussed earlier, the 2673.3 ± 1.4 Ma zircon age of the metadacite from the Wijinnedi domain and the 2654 ± 4 Ma zircon age of trondhjemite from the Hinscliffe domain led Villeneuve and Henderson (1998) to suggest that the trondhjemitic to granodioritic intrusive complex of the Hinscliffe domain may represent an originally somewhat longer lasting, subvolcanic magmatic event related to the mainly dacitic volcanism of the Wijinnedi domain. If so, the originally deeper level, Hinscliffe domain rocks may have been translated laterally and vertically, with respect to the Wijinnedi domain, along the domain-bounding and associated shear zones, to their present position adjacent to the intermediate-felsic volcanic centre. A similar linkage might be considered between the these rocks and the approximately 2640–2630 Ma tonalitic rocks that dominate the presumable deeper level, dominantly granulite-grade Ghost domain. Pressure estimates for a series of rocks from the Ghost domain average about 6.3 kbar (Chacko et al., 1995a), whereas a reasonable pressure estimate at the cordierite isograd such as occurs in the Wijinnedi domain is 3 kbar or less (Spear and Cheney, 1989). This would suggest a vertical component of displacement across the shear zones of more or less 10 km.

Farquhar et al. (1993) suggested rocks from the southeastern part of the Ghost domain may have undergone relatively quick decompression from granulite conditions on the basis of both geothermobarometric and textural evidence. Villeneuve and Henderson (1998) suggested this decompression may be reflected in the 2589 ± 2 Ma monazite age, whether interpreted as down-temperature growth of secondary monazite or as a cooling age that occurred in the 2598 ± 2 Ma megacrystic syenogranite. Gerya et al. (2001) argued that rocks under low- to medium-pressure granulite conditions would have both a lower density than rocks of similar composition under lower grade conditions and a lower viscosity, conditions that would result in a potentially mobile gravitational instability. The uplift and relative positioning of the domains in the Wijinnedi Lake area may be, at least in part, a response to such a gravitational instability imposed by deeper level conditions. These would have been exacerbated by the emplacement of the large, relatively low-density, megacrystic syenogranite bodies among the higher density, tonalite and tonalitic gneiss at lower crustal levels. This instability could, in part, be accommodated with the formation of, and extensional motion along, the domain-bounding and associated internal shear zones that are symmetrical with respect to the large syenogranite bodies and which could well be a consequence of the buoyant uplift of the lighter intrusions.

Paleoproterozoic deformation in the region is most prominently expressed as the curved, 125 km long, cataclastic shear zone, part of which defines the boundary between the Ghost and Dauphinee domains. The terrane to the west of this structure consists of the three domains that, to the southeast, represent successively deeper crustal samples that were juxtaposed during the Archean. During the Paleoproterozoic, the block represented in the area by the three domains has been further uplifted along the cataclastic shear zone. The western margin of this uplift is not precisely defined within or beyond the map area. Southwest of the map area is a prominent north-northwesterly trending linear across which is typically a significant break in slope in east-west aeromagnetic profiles that may represent a western limit of uplift (Fig. 5; Henderson and Chacko, 1995) (*see also* Jackson, 1998, 2003). The Indin dykes within this terrane dip on average 82°E (Fig. 87). This would suggest the block also rotated during uplift with the east side rising several kilometres more than the west side. The total uplift of the block may have been significantly more than this (Henderson and Chacko, 1995). Since the dykes are deformed in the cataclastic shear zone, the maximum age of uplift is considered to be that of the Indin dykes, about 2.1 Ga. It has been suggested (Henderson and Schaan, 1993) that this uplift may be a consequence of the indentation of much of the Slave Province into the western Churchill Province along the Bathurst and McDonald faults between 1.84 Ga and 1.74 Ga (Henderson et al., 1990; Henderson and van Breemen, 1991).

METAMORPHIC GEOLOGY

A remarkably large range of metamorphic conditions is represented by the rocks at the present erosion surface within the relatively small area of the Wijinnedi Lake map area (Fig. 101). These vary from sub-biotite greenschist grade in parts of the Wijinnedi domain to granulite grade in most of the Ghost domain as originally recognized by Robertson and Follinsbee (1974). As noted previously, the shear-zone-bounded domains expose successively deeper structural levels across the map area. The metamorphic mineral sequence in the Wijinnedi domain represents a low-pressure (andalusite-sillimanite) metamorphic pressure series at greenschist and amphibolite grades, as is the case throughout the Slave Province (Thompson, 1978), but the higher grade Ghost domain rocks approach a more intermediate metamorphic pressure series.

Wijinnedi domain

The Wijinnedi domain is metamorphically the most varied. The lowest grade, chlorite-zone rocks at 'West Wijinnedi Lake' represent the southern end of a metamorphic trough that extends some 60 km to the north-northeast in the central part of the Indin Lake supracrustal belt (Tremblay et al., 1953; Stanton et al., 1954; Frith, 1993; Pehrsson and Kerswill, 1997a, b) (Fig. 6). The metamorphic trough is closed off to the south as shown by the presence of cordierite-zone metasedimentary rocks south of 'Glazebrook Lake'. In the largely intermediate to felsic volcanic dome, metamorphic grade increases both to the east and to the south. The metamorphic grade rises to the migmatite zone over a distance of about 10 km to the east, whereas to the west, with a somewhat shallower gradient, the cordierite isograd occurs just at the northwestern corner of the map area (Tremblay et al., 1953; Frith, 1993; Pehrsson and Kerswill, 1997b). The eastern gradient is somewhat telescoped by the vertical movement component across the northerly to northwesterly trending faults. One such fault in part forms the boundary between the sillimanite and cordierite zones, whereas to the south-southeast there is an abrupt change in the amount of sillimanite present on either side of the fault, suggesting a metamorphic jump. Similarly, north of 'Colson Lake' the boundary between the sillimanite and migmatite zones is a fault with leucosome very abundant in the migmatite immediately adjacent to the fault. There may be a metamorphic jump across the Archean shear zone internal to the domain between the mafic volcanic rocks and the metasedimentary rocks to the east. The migmatite bodies east of the volcanic rocks in the southern

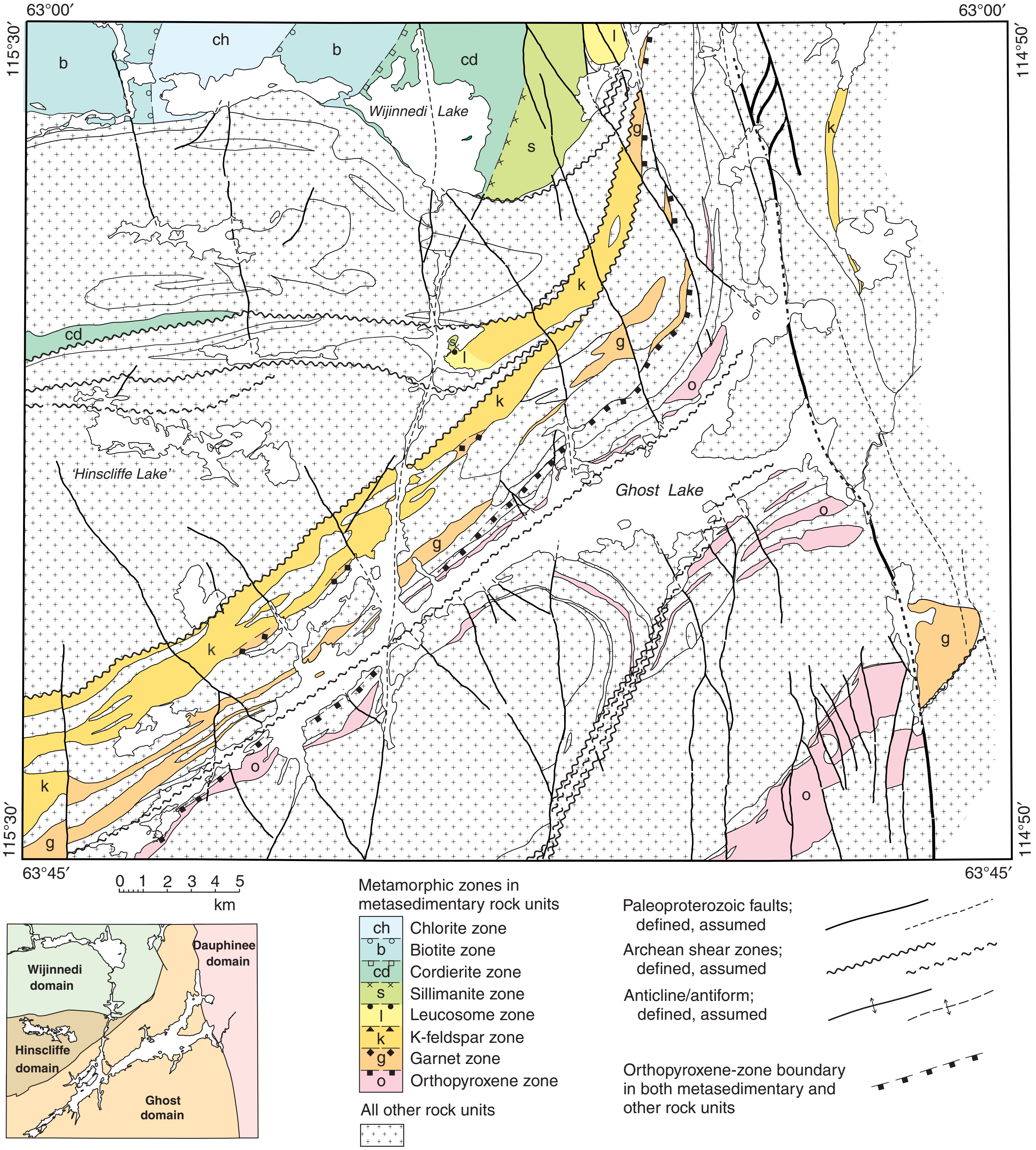


Figure 101. *Metamorphic zonation in Yellowknife Supergroup pelitic rocks in the Wijinnedi Lake map area.*

panel are K-feldspar-bearing whereas K-feldspar was not recognized in the migmatitic metasedimentary rocks north of 'Colson Lake' and west of the projected shear zone through Daran Lake.

The chlorite-zone metasedimentary rocks (unit ĀYs.cl) consist of fine-grained to very fine-grained assemblages of plagioclase-quartz-chlorite-white mica-opaque phases± epidote±tourmaline±carbonate (Fig. 96). In a few examples, very minor, small flakes of biotite are present well within the zone, which might suggest that the rocks of this domain were at biotite grade and were subsequently retrogressed, possibly during the same event that altered the igneous minerals of many of the Paleoproterozoic Indin dykes. The biotite isograd therefore is perhaps better regarded as a retrograde line. Biotite-zone metasedimentary rocks (unit ĀYs.bi) contain a similar mineral assemblage except for the common presence of biotite. All nondetrital minerals are coarser grained, of which biotite is commonly the coarsest. In some examples, biotite is partially to completely replaced by chlorite and in a few cases, coarse-grained chlorite overgrows the usually strongly developed fabric in the rock.

In the cordierite zone (unit ĀYs.cd), the sedimentary detrital textures are less well preserved as the rock becomes more coarsely recrystallized. Mineral assemblages include combinations of plagioclase-quartz-biotite-cordierite±andalusite±muscovite. No staurolite was recognized. The very coarse-grained cordierite is anhedral and highly poikilitic with inclusions of quartz, plagioclase, and minor biotite and muscovite that are significantly finer grained than that in the matrix of the rock. The mica inclusions are commonly, but not always, oriented and the cordierite is variably altered to pinnite. The less common andalusite is significantly less poikilitic with similar, but coarser grained minerals to those in the cordierite. Muscovite, typically as fine-grained flakes, is a minor component and, if present, is a major contributor to the fabric of the rock. Northeast of Wijinnedi Lake, a few layers within the normal metagrewacke-mudstone sequence, considered in the field to be probable felsic volcaniclastic beds, consist of a plagioclase-quartz-biotite-chlorite assemblage, but contains pseudomorphs possibly after amphibole (Fig. 24).

The appearance of sparse sillimanite needles defines the beginning of the sillimanite zone (unit ĀYs.si). It greatly increases in abundance into the zone forming large, irregular fibrolitic mats, which rarely have acicular sillimanite crystals associated with them. Muscovite on the whole is more abundant in the sillimanite zone, forming relatively coarse, somewhat poikilitic flakes, which in some cases appear to overgrow any biotite-defined fabric present. The muscovite is probably retrograde and although no K-feldspar was recognized within the zone, the muscovite may represent the breakdown of earlier formed, but no longer present K-feldspar. Silicate iron-formation occurs locally in the sillimanite-zone pelitic rocks and consists of assemblages of grunerite-garnet-hornblende-quartz-opaque minerals and sulphide is locally present. Pelitic metasedimentary rocks interlayered with, or in close association to, the iron-formation units can be garnet bearing, a mineral not normally present in the pelitic rocks other than in the highest grade metasedimentary migmatite zones of the Ghost domain.

To the east of the sillimanite zone, north of 'Colson Lake' the metasedimentary rocks are migmatitic (unit ĀYs.m) with the leucosome phase everywhere abundantly developed. The boundary between the two zones is a fault. The zone is quite homogeneous consisting mainly of leucosome and paleosome interlayered on a several millimetre scale to less than 2 cm scale, with the latter predominating. Homogenized zones in which layering is lost are present, but rare. Even less common are thicker accumulations of leucosome into layers or lenses tens of centimetres in scale. Mineralogy present includes assemblages of plagioclase-quartz-biotite-sillimanite±cordierite±muscovite. Cordierite occurs in largely inclusion-free, clear grains that at its most coarse is only slightly larger than the leucosome quartz and feldspar present. Muscovite is usually present, although the proportion is highly varied. In places it overgrows sillimanite, suggesting it is a retrograde phase (Fig. 102). Potassium feldspar was not recognized, but in one case a few grains of myrmekitic plagioclase were present, suggesting K-feldspar may have retrograded out. Similar mineral assemblages representative of this zone also occur in the southeastern part of the domain in the metasedimentary rocks just east of the Ghost River, south of 'Gehl Lake'.

The remaining metasedimentary migmatite bodies of the Wijinnedi domain east of the volcanic rocks are mineralogically similar to the preceding zone, but are K-feldspar bearing or at least have evidence that they were K-feldspar bearing at one time (unit ĀYs.mk). Aggregates of quartz and skeletal muscovite commonly with myrmekite adjacent to them are thought to represent retrograded K-feldspar porphyroblasts in the migmatite. These migmatite bodies are more heterogeneous than the migmatite bodies to the north, due in large part to the more varied proportion of psammitic-to-pelitic compositions. The rocks vary from metatexite with layering ranging from millimetres to tens of centimetres, to homogenized metatexite to diatexite. They are very similar to the more abundant K-feldspar-zone metasedimentary migmatite of the adjacent Ghost domain and are more completely described below in 'Ghost domain'.

The metamorphic pattern in the volcanic rocks of the domain is less precisely defined due to compositional constraints. The more quartzofeldspathic, felsic metavolcanic rocks are particularly insensitive. In the amphibole-bearing, intermediate to mafic volcanic rocks and the metagabbro plutons and sills related to them, the transition from actinolitic amphibole to hornblende indicates the metamorphic grade in the volcanic rocks increases to the south as well as to the east. This transition roughly corresponds to the change from greenschist to amphibolite facies as represented by the cordierite isograd in the pelitic rocks and crosses through the volcanic complex just north of 'Glazebrook Lake'.

A metamorphosed alteration zone occurs within the intermediate to felsic volcanic rocks extending over at least 1.5 km southeast of Gale Lake. The rocks contain assemblages of quartz, plagioclase, biotite, cordierite, cummingtonite, and

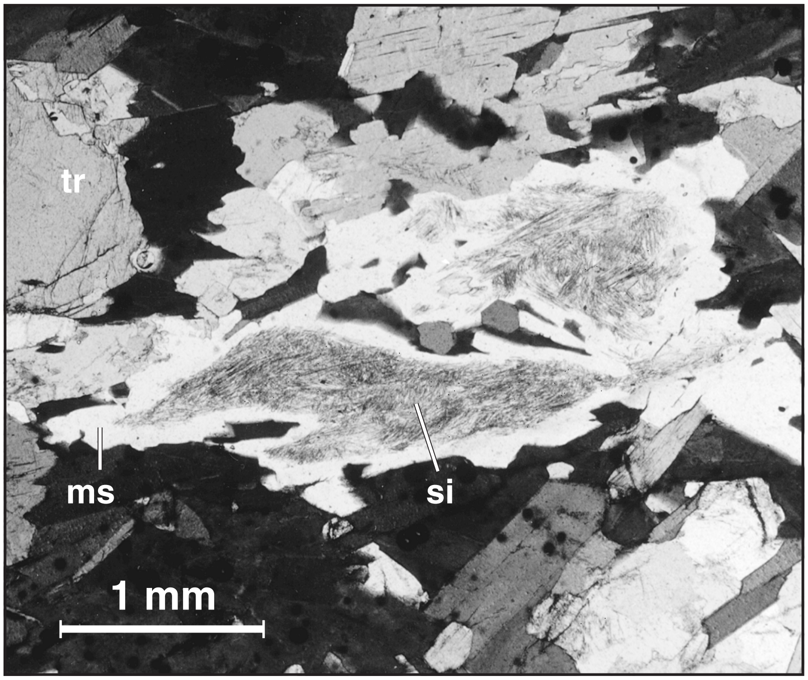


Figure 102.

Photomicrograph of retrogressive muscovite (ms) overgrowing sillimanite (si) which is abundant in the lowest grade, migmatite-zone rocks, but rare or absent in the higher grade zones. Coarse grains of tourmaline (tr) are also present. Plane-polarized light.

anthophyllite (Fig. 15). These assemblages suggest a significant enhancement of magnesium over that of most of the volcanic unit. It is not known if this alteration is related to volcanogenic processes or to hydrothermal effects related to the emplacement of the nearby metagabbro intrusions.

Ghost domain

Throughout the domain, the Yellowknife Supergroup metasedimentary rocks are everywhere migmatite. They have been divided into three zones based on the presence or appearance of K-feldspar, garnet, and orthopyroxene at successively higher grades. Other aspects of these rocks are described in the 'Metasedimentary migmatite and diatexite of the Ghost, Wijinnedi, and Dauphinee domains (units ĀYs.m, ĀYs.mk, ĀYs.mg, ĀYs.mo)' section.

The lower grade, K-feldspar-bearing, but garnet-absent zone (unit ĀYs.mk) occurs along the northwestern margin of the domain and is mineralogically similar to the K-feldspar-bearing migmatite bodies of the Wijinnedi domain, suggesting no significant metamorphic jump across the shear zones that separate the domains, at least at the east end. The migmatite bodies consist of assemblages of plagioclase, quartz, biotite, K-feldspar, cordierite, sillimanite, and muscovite. Potassium feldspar is commonly present as somewhat coarser grained porphyroblasts that are commonly mantled or overgrown by muscovite or an intergrowth of commonly skeletal muscovite and quartz due to retrogression (Fig. 103). Myrmekite is commonly associated with the K-feldspar, but also occurs in aggregates with muscovite, suggesting that K-feldspar may have been present during peak metamorphic conditions, but is now retrograded to muscovite and quartz. Sillimanite is typically a rather minor to rare constituent of the rock, occurring most commonly as fibrolite associated with biotite, but also as groups of fine needles within cordierite in some cases. Sillimanite can be present only within cordierite, or in other cases, only as extremely fine-grained fibrolite in fractures or grain boundaries with plagioclase and probably represents a retrograde product. Garnet occurs at a few localities in the central part of the zone in the vicinity of the granodiorite intrusions. This may be due to local compositional anomalies, locally elevated metamorphic conditions due to the proximity of the intrusions, or unrecognized structural complications.

The garnet-bearing migmatite zone (unit ĀYs.mg) occurs to the southeast along the north shore of Ghost Lake. Both diatexite and metatexite are present, although white-weathering diatexite is more common than in the other migmatitic zones. The migmatitic rocks of this domain consist of assemblages of plagioclase, quartz, biotite, garnet, K-feldspar, cordierite, and muscovite (Farquhar et al., 1993; Chacko et al., 1995b). The proportion of K-feldspar is significantly greater than in the K-feldspar-bearing zone, occurring in commonly coarser grains that, as before, vary from being unaltered to partially to totally retrogressed. In a few cases the plagioclase is antiperthitic. The proportion of biotite is typically reduced and is more commonly red than in the lower grade zones. Sillimanite is rarely, if ever, present and then only as very small amounts of fine-grained fibrolite in fractures and along grain margins. Subhedral to anhedral garnet is present in varied proportions, usually with fine-grained inclusions of quartz, feldspar, and biotite. Cordierite is common as clear, sometimes twinned grains or aggregates of grains. Muscovite, presumably retrogressive, is less common and, where present, occurs in relatively small amounts, suggesting the degree of retrogression of the rocks of this zone is less complete. It does not form a distinct fabric element, suggesting the retrogression of the K-feldspar takes place relatively early in the evolution of the migmatite.

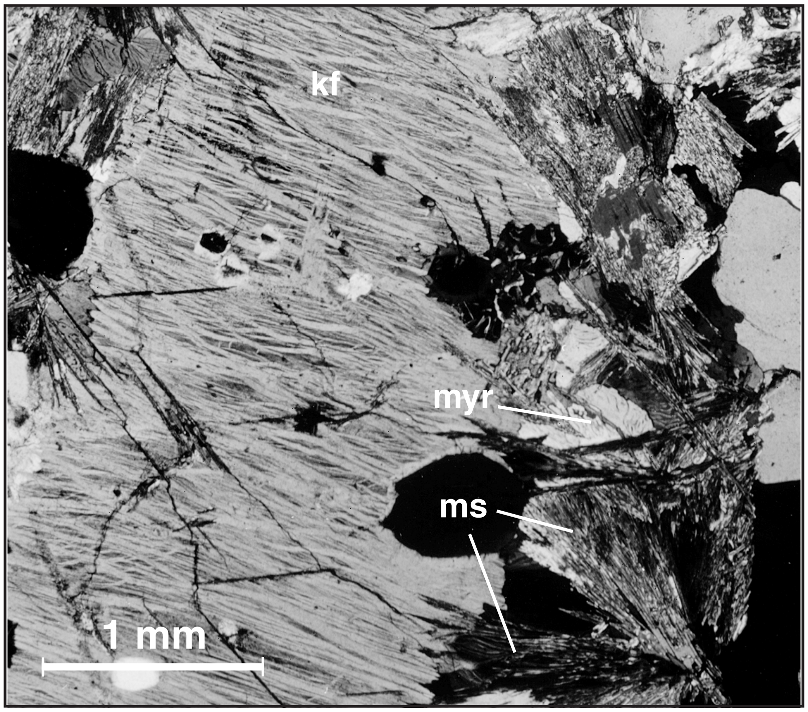


Figure 103.

Photomicrograph of Yellowknife Supergroup metasedimentary migmatite from the K-feldspar zone. Coarse perthitic K-feldspar porphyroblast (kf) is mantled in part by fine-grained myrmekite (myr) and is partially overgrown by radiating sprays of retrogressive muscovite (ms). Colour version of photomicrograph appears on cover. Cross-polarized light.

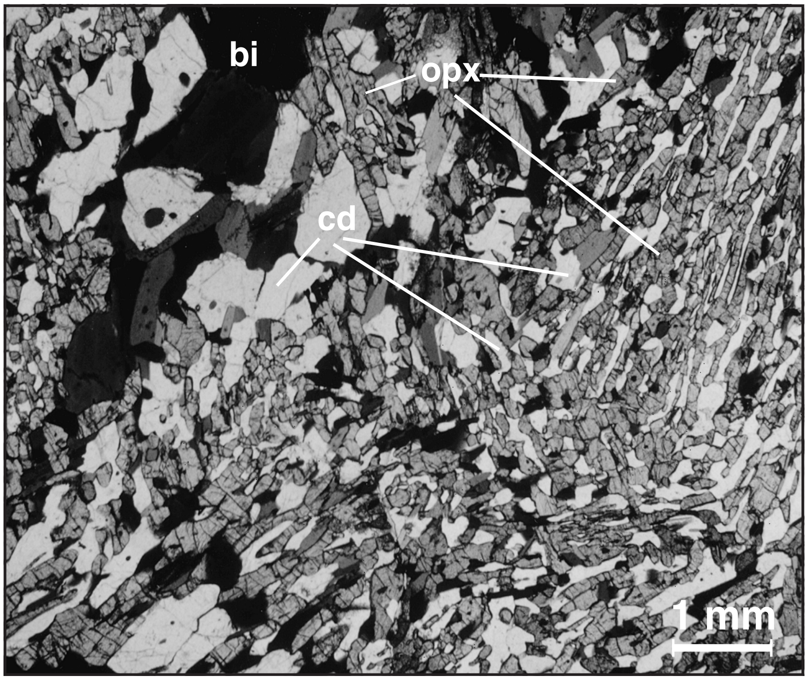


Figure 104.

Photomicrograph of Yellowknife Supergroup metasedimentary migmatite which consists in large part of a fine intergrowth of orthopyroxene (opx) and cordierite (cd) along with some biotite (bi). Plane-polarized light.

The highest grade zone in the metasedimentary rocks contains assemblages of plagioclase, quartz, K-feldspar, biotite, garnet, cordierite, orthopyroxene, and rare examples of spinel (Farquhar et al., 1993; Chacko et al., 1995b) (Fig. 104, 105). The highest grade metasedimentary rocks occur for the most part southeast of Ghost Lake, primarily as several kilometre-scale rafts to screens of inclusions within the orthopyroxene-bearing tonalite and tonalitic gneiss. A major metasedimentary migmatite unit up to 2.5 km wide occurs in the southeast corner of the area, but appears to continue south of the area as much as 10 km based on its distinctive, low-magnetic expression (Fig. 5). As with the other migmatitic zones, both metatexite and diatexite occur. The white-weathering diatexite is in strong contrast to the typical yellow to greenish-brown granulite colours of the various orthopyroxene-bearing intrusive bodies with which the diatexite is commonly intimately associated. Not all the migmatite bodies in this zone contain orthopyroxene, and where present, are commonly extensively altered. A few of the rocks from this domain also contain spinel, either within cordierite grains or cordierite grain aggregates or, in one case, within garnet.

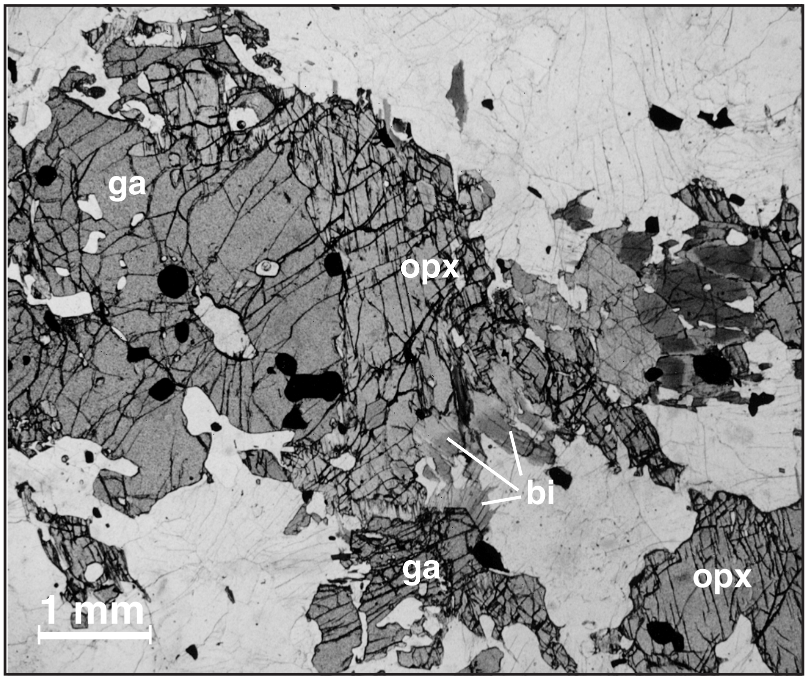


Figure 105.

Photomicrograph of orthopyroxene-zone metasedimentary migmatite consisting of quartz, plagioclase, cordierite, and garnet (ga), with orthopyroxene (opx) in part overgrowing the garnet. Biotite (bi) is a minor component. Plane-polarized light.

Pressure-temperature estimates for the highest grade rocks using the garnet-orthopyroxene-plagioclase-quartz barometer and the aluminum-in orthopyroxene thermometer (Fitzsimons and Harley, 1994; Pattison and Begin, 1994) have been made by Chacko et al. (1995a) and Farquhar and Chacko (1996) (Fig. 106). The data came from a series of samples above the orthopyroxene line throughout the Ghost domain and showed no regular variation in pressure or temperature (Chacko et al., 1995a). Pressure and temperature estimates ranged between 5.5 kbar and 6.8 kbar and 827°C and 899°C (Chacko et al., 1995a; Fig. 106). A refinement of these calculations on the same or similar samples resulted in a somewhat higher pressure and temperature range at 5.9 kbar to 7.4 kbar and 845°C to 920°C (Farquhar and Chacko, 1996; Fig. 106). The local occurrence of spinel-cordierite assemblages within this zone suggests a decompression event (Fig. 106) that perhaps is related to the proposed uplift resulting from the emplacement of the large volume of low-density granite bodies at about 2595 Ma (*see* 'Discussion' in 'Granite intrusions').

In addition to the mainly pelitic, migmatitic metasedimentary rocks, two units of intermediate to felsic metavolcanic rocks occur within the Ghost domain that have been suggested as possible correlatives with the main, largely intermediate volcanic centre in the Wijinnedi domain. These rocks are everywhere migmatitic and contain assemblages of plagioclase, quartz, biotite, orthopyroxene, and clinopyroxene, with retrograde amphibole, chlorite, and carbonate assemblages locally rimming to partially replacing the pyroxene.

The supracrustal rocks are intruded by a series of granitoid rocks that volumetrically dominate the domain. East and southeast of the orthopyroxene line all the intrusions are orthopyroxene bearing to a greater or lesser extent. In all cases, parts of these intrusive units are sufficiently well preserved to suggest that the orthopyroxene is a primary igneous

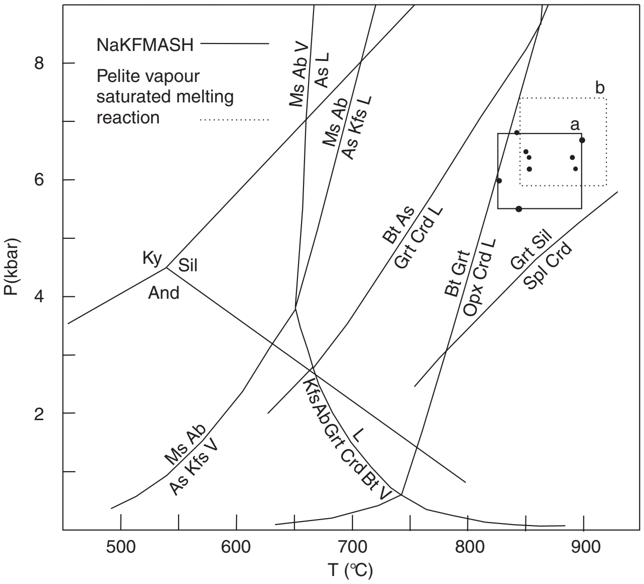


Figure 106. *Pressure-temperature diagram for high-grade pelitic rocks showing selected melting and dehydration reactions in the NaKFMASH system* modified from *Spear et al. (1999). Solid line box outlines pressure and temperature estimates from metasedimentary migmatite rocks from the highest grade rocks of the Ghost domain determined by Chacko et al. (1995a). Dashed lines outline a refined version of these estimates by Farquhar and Chacko (1996). Ab = albite, And = andalusite, As = aluminosilicate, Bt = biotite, Crd = cordierite, Grt = garnet, Kfs = K-feldspar, Ky = kyanite, L = liquid, Ms = muscovite, Opx = orthopyroxene, Sil = sillimanite, Spl = spinel, V = vapour.*

mineral and that the intrusions crystallized under granulite-grade conditions. The most voluminous of these intrusions, the tonalite and related tonalitic gneiss, are also the oldest. Based on one U-Pb (zircon) age determination from within the area (Appendix), they would appear to have crystallized between 2.64 Ga and 2.63 Ga. This is supported by a 2639 ± 6 Ma U-Pb (zircon) age for a presumably similar tonalite sheet about 10 km south of the area (Perks, 1997). As the highest grade, orthopyroxene-bearing, metasedimentary rocks occur immediately adjacent to or as inclusions within these tonalite bodies, these ages represent a point within the metamorphic age range. A younger point is provided by the U-Pb (zircon) ages of the megacrystic granite and the orthopyroxene granite, both of which also crystallized under granulite-grade conditions and have been dated at 2598 ± 2 Ma and 2589 +1/-2 Ma respectively (Appendix; Villeneuve and Henderson, 1998). This would suggest the rocks of most of the domain were exposed to granulite-grade metamorphic conditions for over 40 million years, there being no evidence recognized that more than one granulite event took place. Monazite fractions from both the tonalite and the megacrystic granite fall within the 2590–2580 Ma age range (Villeneuve and Henderson, 1998; Appendix) and whether representing the time of U-Pb system closure of igneous monazite at about 700°C (Heaman and Parrish, 1991) or the retrogressive metamorphic growth of monazite, their age is considered to mark the end of the period of granulite-grade conditions for these rocks (Appendix; Villeneuve and Henderson, 1998). As discussed previously in the ‘Discussion’ sections of ‘Structural geology’ and ‘Granite intrusions’, the end of granulite-grade conditions may correspond to the time of movement on the domain-bounding shear zones which are thought to be a consequence of the buoyant effect of the emplacement of large volumes of low-density granite.

There is ample evidence of retrogression from peak metamorphic grades in rocks of the Ghost domain. This has already been mentioned previously with the common occurrence of muscovite and the local breakdown of K-feldspar in the metasedimentary migmatite. Retrogressive effects are also seen in the granitoid rocks as well. In particular, the large megacrystic granite body only rarely contains orthopyroxene. Chacko et al. (1995a, b) and Henderson and Chacko (1995) suggested that this was a result of autoretrogession due to the progressive increase of water in the remaining magma as the granite crystallized. Water would then be present in sufficient quantities to retrograde most of the previously crystallized orthopyroxene as the body cooled. Presumably related to this is the common occurrence of retrograded tonalite gneiss northeast of the megacrystic granite. The megacrystic granite is thought to gently plunge below the tonalitic gneiss in this area and water from it may have contributed to the retrogression in the tonalitic gneiss. Orthopyroxene crystals in the tonalite and tonalitic gneiss also tend to be more commonly altered in the vicinity of the prominent shear zone south of Ghost Lake. What appear to be altered pyroxene crystals also occur locally northwest of the orthopyroxene isograd as mapped. This would suggest the original extent of granulite-grade conditions as defined by the presence of orthopyroxene may have been somewhat beyond the location of the mapped isograd.

Hinscliffe domain

The Hinscliffe complex consists mainly of trondhjemite or trondhjemite gneiss toward the margin of the complex where the rocks are more strongly deformed. Minor amounts of thin, aplite dykes and pegmatite are commonly present. None of these provide particularly useful mineral assemblages that would indicate the metamorphic grade; however, all the phases are variably recrystallized, commonly having a sugary aspect in outcrop or hand specimen. The marginal parts of the complex are intruded by deformed mafic to intermediate intrusive sheets that are hornblende-plagioclase-quartz±biotite±epidote±titanite±ilmenite bearing. Varied amounts of mafic inclusions occur throughout the complex which, where locally well enough preserved, appear to represent synplutonic mafic dykes. They consist of hornblende-plagioclase-biotite±titanite±epidote±chlorite assemblages. This would suggest the complex as a whole is at amphibolite grade.

Dauphinee domain

The Dauphinee domain, east of the major low-grade, cataclastic shear zone consists mainly of granitoid rocks with lesser amounts of metasedimentary migmatite. Granitic rocks dominate the domain and they are also the youngest rock unit. In many cases, igneous textures are well preserved. Retrograded pyroxene is present in an altered tonalitic inclusion near the boundary with the high-grade Ghost domain and is the only assemblage suggestive of the high-grade rocks that occur to the west. The quartz diorite in the northern part of the domain has been metamorphosed. A relict igneous texture is preserved and consists of subhedral plagioclase in a finer grained matrix of the metamorphic assemblage hornblende-biotite-quartz-plagioclase±epidote±titanite. The pelitic migmatite units have plagioclase-quartz-biotite-garnet-K-feldspar±muscovite assemblages similar to those seen in the Ghost domain. No cordierite- or orthopyroxene-bearing assemblages were recognized within the map area, but do occur immediately to the north (S. Pehrsson, pers. comm., 2001).

ECONOMIC GEOLOGY

Mineral exploration began in the Wijinnedi Lake area after interest was shown in the Indin Lake region to the north. Most prospects recognized to date were discovered in the 1940s with further exploration and follow-up in the 1970s and 1980s. To date, no significant economic discoveries have been made. The two main targets of primary interest have been the iron-formation in the Yellowknife Supergroup metasedimentary rocks east of Wijinnedi Lake and the contact area between the Yellowknife Supergroup metavolcanic and metasedimentary rocks south of the lake.

Silicate iron-formation that is locally sulphide bearing occurs in the Yellowknife Supergroup metagreywacke-mudstone turbidite units of the easternmost Wijinnedi domain. Locations of iron-formation encountered in the course of mapping are noted on the map and represent a minimum estimate of those present. These iron-formation units have been described elsewhere in this report. A quartz vein in a silicate iron-formation near the south end of 'Colson Lake' contained 3620 ppb gold, although most other veins assayed from that particular iron-formation unit and others elsewhere in the area contained significantly less than 100 ppb Au (the iron-formation units were not assayed) (Bryan, 1992). Similar iron-formation units occur more or less along strike to the north where at Damoti Lake (Fig. 6) greenschist-grade, but otherwise similar silicate iron-formation units carry as much as 26 800 ppb gold (Brophy, 1993). No iron-formation was recognized in the greenschist-grade rocks west of Wijinnedi Lake, although Pehrsson and Kerswill (1997b) noted the occurrence of iron-formations in the greenschist-grade metasedimentary rocks immediately north of the area. A sample of greenschist-grade, arsenopyrite-bearing, silicate iron-formation was found at an old camp site in 'West Wijinnedi Lake' that contained over 2500 ppb Au (John Brophy, DIAND, pers. comm., 1993) and may have been originally collected in the region.

The contact area between the Yellowknife Supergroup metavolcanic and metasedimentary rocks, mainly southwest of Wijinnedi Lake, but to some extent to the west and east as well, has attracted attention over the years due to the local development of prominent gossans at the contact. The contact zone typically consists of black, carbonaceous metamudstone conformably overlying the mafic to intermediate metavolcanic rocks to the southwest and more normal metapelite and associated thin-bedded metasiltstone to the northeast. Thin, metagabbro sills (10 cm to a few metres) are common in the zone and are more abundant here than in metasedimentary and metavolcanic rocks to the north and south, respectively. Patchy discontinuous gossans on the order of 10 m occur along this zone and involve both the carbonaceous metasedimentary rocks and the metagabbro. Mineralization is mainly pyrrhotite-pyrite with minor chalcopyrite locally. The mineralization in the metasedimentary rocks varies from finely disseminated to finely laminated to, in one case at the central showing noted on the map, a 2 m wide lens of massive pyrrhotite with lesser pyrite and chalcopyrite (Bryan, 1992). Assays, largely on quartz vein material associated with the gossans, indicate less than a few tens of parts per billion gold in most cases, but up to 226 ppb in two or three instances (Seaton et al., 1987, p. 239; Bryan, 1992). The margins of the metagabbro sills are also pyrrhotite bearing in the gossan zones, but not in the interior of the sills or in sills remote from gossanous areas. Most of the zone is not mineralized, which would suggest the mineralization is controlled by the intrusion of the sills into local sulphur-rich zones in the carbonaceous metasedimentary rocks.

A third potential economic target is the previously described zone of altered, largely felsic metavolcanic rocks east of Gale Lake in the felsic to intermediate volcanic dome. These rocks are anomalous in that they contain cordierite-cummingtonite-anthophyllite assemblages, suggesting the presumably normal, original volcanic rocks were altered due to hydrothermal activity related either to volcanogenic processes or perhaps related to the emplacement of the nearby metagabbro intrusive complex to the west and its associated sills, which occur throughout the volcanic rocks east of the intrusion. No mineralization associated with this alteration was recognized in the course of mapping the region, but a 0.6 m wide, quartz vein containing massive sulphides found in the course of an EM survey occurs at the small lake 1 km south of Gale Lake (Seaton et al., 1987, p. 239).

REFERENCES

Ashwal, L.D.
1993: Anorthosites; Springer-Verlag, Berlin, Germany, 422 p.

Ashwal, L.D. and Myers, J.S.
1994: Archean anorthosites; *in* Archean Crustal Evolution, (ed.) K.C. Condie; Elsevier, Amsterdam, Netherlands, p. 315–355.

Ashworth, J.R.
1985: Introduction; *in* Migmatites, (ed.) J.R. Ashworth; Blackie, Glasgow, p. 1–35.

Bank, C.-G., Bostock, M.G., Ellis, R.M., and Cassidy, J.F.
2000: A reconnaissance teleseismic study of the upper mantle and transition zone beneath the Archean Slave craton in NW Canada; Tectonophysics, v. 319, p. 151–166.

Bennett, V., Dunning, G., and Indares, A.
2000: Preliminary data from the Kwejinne Lake supracrustal belt – Ghost Lake granulite domain transect: impact of a steep thermal gradient on upper crustal rocks; *in* 28th Yellowknife Geoscience Forum Program and Abstracts of Talks and Posters, Northwest Territories Chamber of Mines, p. 10–11.

Bennett, V.R.C. and Dunning, G.R.
1998: Geological transect across the southern Indin Lake supracrustal belt to the central Ghost Lake granulite domain; *in* 26th Yellowknife Geoscience Forum; Indian and Northern Affairs Canada, Northwest Territories Chamber of Mines and Resources, Wildlife and Economic Development, Government of the Northwest Territories, Program and Abstracts of Talks and Posters, p. 14–16.

Bleeker, W. and Beaumont-Smith, C.
1995: Thematic structural studies in the Slave Structural Province: preliminary results and implications for the Yellowknife Domain; *in* Current Research 1995-C; Geological Survey of Canada, p. 87–96.

Bleeker, W. and Davis, W.J.
1999: The 1991–1996 NATMAP Slave Province Project: introduction; *in* NATMAP Slave Province Project, (ed.) W. Bleeker and W. Davis; Canadian Journal of Earth Sciences, v. 36, p. 1033–1042.

Bleeker, W. and Villeneuve, M.
1995: Structural studies along the Slave portion of the SNORCLE transect; *in* LITHOPROBE Slave/Northern Cordillera Lithospheric Evolution Workshop, LITHOPROBE Report 44, p. 8–13.

Bleeker, W., Ketchum, J.W.F., Jackson, V.A., and Villenueve, M.E.
1999: The Central Slave Basement Complex, Part I: its structural topology and autochthonous cover; *in* NATMAP Slave Province Project, (ed.) W. Bleeker and W. Davis; Canadian Journal of Earth Sciences, v. 36, p. 1083–1109.

Boyd, F.R.
1959: Hydrothermal investigations of amphiboles; *in* Researches in Geochemistry, (ed.) P.H. Abelson; Wiley, New York, New York, p. 377–396.

Brodaric, B., Harrap, R., and Lemkow, D.
2001: FieldLog v. 3.0: Users guide and reference; Geological Survey of Canada, Open File 3239, 170 p.

Brophy, J.A.
1993: BIF Island, Damoti Lake area: the first documented iron-formation-hosted gold showing in the Indin Lake supracrustal belt; *in* Exploration Overview 1992 Northwest Territories, Mining, Exploration and Geological Investigations; Indian and Northern Affairs Canada, Yellowknife, Northwest Territories, p. 18–19.

Brown, I.C.
1955: Late faults in the Yellowknife area; Geological Association of Canada, Proceedings, v. 7, p. 123–138.

Bryan, D.
1992: Wijinnedi Lake report; Indian Affairs and Northern Affairs Canada, Assessment Report 083072, 14 p.

Burwash, R.A., Baadsgaard, H., Campbell, F.A., Cumming, G.L., and Folinsbee, R.E.

1963: Potassium-argon dates of diabase dyke systems, District of Mackenzie, N.W.T.; Canadian Institute of Mining and Metallurgy Transactions, v. 66, p. 303–307.

Cas, R.A.F. and Wright, J.V.

1987: Volcanic successions modern and ancient; Chapman and Hall, London, United Kingdom, 528 p.

Chacko, T., Creaser, R.A., Farquhar, J., and Muehlenbachs, K.

1995a: The deep crust of the western Slave Province — initial petrologic and isotopic data from the high-grade rocks of the Ghost Domain; *in* LITHOPROBE Slave/Northern Cordillera Lithospheric Evolution Workshop, LITHOPROBE Report 44, p. 4–7.

Chacko, T., Farquhar, J., and Creaser, R.A.

1995b: A petrologic study of granulites and associated rocks from the Ghost Lake area, southwestern Slave Province; *in* Geological Association of Canada–Mineralogical Association of Canada Annual Meeting, Victoria 1995, Program with Abstracts, v. 20, p. A-15.

Corcoran, P.L., Mueller, W.U., and Chown, E.H.

1998: Climatic and tectonic influences on fan deltas and wave- to tide-controlled shoreface deposits: evidence from the Archaean Keskarrah Formation, Slave Province, Canada; Sedimentary Geology, v. 120, p. 125–152.

Davis, W.J. and Bleeker, W.

1999: Timing of plutonism, deformation and metamorphism in the Yellowknife Domain, Slave Province, Canada; *in* NATMAP Slave Province Project, (ed.) W. Bleeker and W. Davis; Canadian Journal of Earth Sciences, v. 36, p. 1169–1187.

Davis, W.J., Fryer, B.J., and King, J.E.

1994: Geochemistry and evolution of Late Archean plutonism and its significance to the tectonic development of the Slave craton; Precambrian Research, v. 67, p. 207–241.

Department of Energy, Mines and Resources

1969: Rae, Northwest Territories; Gravity Map Series; Observatories Branch, No. 90, scale 1:500 000.

Dudás, F.Ö., Henderson, J.B., and Mortensen, J.K.

1989: U-Pb ages of zircons from the Anton Complex, southern Slave Province, N.W.T.; *in* Radiogenic Age and Isotopic Studies: Report 3; Geological Survey of Canada, Paper 89-2, p. 39–44.

Easton, R.M.

1981: Geology of 86H/3, H/4, H/5, and H/6, District of Mackenzie; Department of Indian Affairs and Northern Development, Northern Affairs Program, Geological Division, Preliminary Geological Map, EGS-1981-5, scale 1:30 000.

Farquhar, J. and Chacko, T.

1996: Thermobarometry of high-temperature granites and granulites of the western Slave Province, District of Mackenzie, NWT, Canada; *in* 1996 Spring Meeting, EOS, American Geophysical Union, v. 77, no. 17, Supplement, p. 283–284.

Farquhar, J., Snavely, J.A., and Chacko, T.

1993: Granulite facies metamorphism near Ghost Lake, Slave Province, NWT; *in* Geological Association of Canada–Mineralogical Association of Canada Joint Annual Meeting, Edmonton 1993, Program and Abstracts, v. 18, p. A-28.

Fitzsimons, I.C.W. and Harley, S.L.

1994: The influence of retrograde cation exchange on granulite P-T estimates and a convergence technique for the recovery of peak metamorphic conditions; Journal of Petrology, v. 35, p. 543–576.

Folinsbee, R.E.

1940a: Gem cordierite from northern Canada; M.Sc. thesis, University of Minnesota, Minneapolis-St. Paul, Minnesota, 85 p.

1940b: Gem cordierite from the Great Slave Lake area, N.W.T., Canada; American Mineralogist, v. 25, p. 216.

Franklin, J.M., Lydon, J.W., and Sangster, D.F.

1981: Volcanic-associated massive sulphide deposits; *in* Economic Geology, Seventy-fifth Anniversary Volume, 1905–1980, (ed.) B.J. Skinner; Economic Geology Publishing Company, New Haven, Connecticut, p. 485–627.

Frith, R.A.

1987: Precambrian geology of the Hackett River area, District of Mackenzie, N.W.T.; Geological Survey of Canada, Memoir 417, 61 p.

1993: Precambrian geology of the Indin Lake map area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Memoir 424, 63 p.

Fulton, R.J.

1995: Surficial materials of Canada; Geological Survey of Canada, Map 1880A, scale 1:5 000 000.

Gates, T.M. and Hurley, P.M.

1973: Evaluation of Rb-Sr dating methods applied to the Matachewan, Abitibi, Mackenzie, and Sudbury dyke swarms in Canada; Canadian Journal of Earth Sciences, v. 10, p. 900–919.

Geological Survey of Canada

1963: Ghost Lake, District of Mackenzie, Northwest Territories; Aeromagnetic Series; Geological Survey of Canada, Geophysics Paper 2955, Map 2955 G, scale 1:50 000.

1969: Wecho River, District of Mackenzie, Northwest Territories; Aeromagnetic Series; Geological Survey of Canada, Geophysics Paper 7196, Map 7196G, scale 1:253 440.

1980: Point Lake, District of Mackenzie, Northwest Territories; Geophysical Series (Aeromagnetic); Geological Survey of Canada, Map 7874G, scale 1:250 000.

Gerya, T.V., Maresch, W.V., Willner, A.P., Van Reenen, D.D., and Smit, C.A.

2001: Inherent gravitational instability of thickened continental crust with regionally developed low- to medium-pressure granulite facies metamorphism; Earth and Planetary Science Letters, v. 190, p. 221–235.

Heaman, L. and Parrish, R.

1991: U-Pb geochronology of accessory minerals; *in* Applications of Radiogenic Isotopic Systems to Problems in Geology, (ed.) L. Heaman and J.N. Ludden; Mineralogical Association of Canada, v. 19, p. 59–102.

Henderson, J.B.

1970: Stratigraphy of the Yellowknife Supergroup, Yellowknife Bay–Prosperous Lake area, District of Mackenzie; Geological Survey of Canada, Paper 70-26, 12 p.

1975a: Archean stromatolites in the northern Slave Province, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 12, p. 1619–1630.

1975b: Sedimentology of the Archean Yellowknife Supergroup at Yellowknife, District of Mackenzie; Geological Survey of Canada, Bulletin 246, 62 p.

1985: Geology of the Yellowknife–Hearne Lake area, District of Mackenzie: a segment across an Archean basin; Geological Survey of Canada, Memoir 414, 135 p.

1994: Geology of the Wijinnedi Lake area — a Paleoproterozoic(?) asymmetric uplift of Archean rocks on the southwestern Slave Province, District of Mackenzie, Northwest Territories; *in* Current Research 1994-C; Geological Survey of Canada, p. 71–79.

1998a: Geology of the Keskarrah Bay area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Bulletin 527, 122 p.

1998b: Preliminary geology, Wijinnedi Lake area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 3609, scale 1:50 000.

Henderson, J.B. and Chacko, T.

1995: A reconnaissance of the high grade metamorphic terrane south of Ghost Lake, southwestern Slave Province, Northwest Territories; *in* Current Research 1995-C; Geological Survey of Canada, p. 77–85.

Henderson, J.B. and Schaan, S.E.

1993: Geology of the Wijinnedi Lake area: a transect into mid-crustal levels in the western Slave Province, District of Mackenzie, Northwest Territories; *in* Current Research, Part C; Geological Survey of Canada, Paper 93-1C, p. 83–91.

Henderson, J.B. and van Breemen, O.

1991: K-Ar (hornblende) data from the Healey Lake area, District of Mackenzie: a potential time constraint on the intracratonic indentation of the Slave Province into the Thelon Tectonic Zone; *in* Radiogenic Age and Isotopic Studies: Report 4; Geological Survey of Canada, Paper 90-2, p. 61–66.

Henderson, J.B., James, D.T., and Thompson, P.H.

1999: Geology, Healey Lake–Artillery Lake, Northwest Territories–Nunavut; Geological Survey of Canada, Open File 3819, scale 1:250 000.

Henderson, J.B., McGrath, P.H., Thériault, R.J., and van Breemen, O.

1990: Intracratonic indentation of the Archean Slave Province into the early Proterozoic Thelon Tectonic Zone of the Churchill Province, northwestern Canadian Shield; Canadian Journal of Earth Sciences, v. 27, p. 1699–1713.

Henderson, J.F. and Brown, I.C.

1952: The Yellowknife Greenstone Belt; Geological Survey of Canada, Paper 52-28, 41 p.

1966: Geology and structure of the Yellowknife Greenstone Belt, District of Mackenzie; Geological Survey of Canada, Bulletin 141, 87 p.

Henderson, J.R., Henderson, M.N., Kerswill, J.A., and Dehls, J.F.

2000: Geology, High Lake greenstone belt, Nunavut; Geological Survey of Canada, Map 1945A, scale 1:100 000.

Hiscott, R.N. and Pickering, K.T.

1984: Reflected turbidity currents on an Ordovician basin floor; Nature, v. 311, p. 143–145.

Hoffman, P. and Hall, L.

1993: Geology, Slave craton and environs, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 2559, scale 1:1 000 000.

Hrabi, R.B., Nelson, M.D., and Helmstaedt, H.

1995: Diverse metavolcanic sequences and late polymictic conglomerate-associated metasedimentary rocks in the Winter Lake supracrustal belt, Slave Province, Northwest Territories; *in* Current Research 1995-E; Geological Survey of Canada, p. 137–148.

Hurdle, E.

1983: Geology of a volcanic pile at Clan Lake, N.W.T; Indian and Northern Affairs, Canada, EGS 1983-5a, b, c, scale 1:10 000.

Isachsen, C.E.

1992: U-Pb zircon geochronology of the Yellowknife volcanic belt and subjacent rocks, N.W.T., Canada: constraints on the timing, duration, and mechanics of greenstone belt formation; Ph.D. thesis, Washington University, St. Louis, Missouri, 164 p.

Isachsen, C.E. and Bowring, S.A.

1994: Evolution of the Slave craton; Geology, v. 22, p. 917–920.

1997: The Bell Lake group and Anton Complex: a basement-cover sequence beneath the Archean Yellowknife greenstone belt revealed and implicated in greenstone belt formation; Canadian Journal of Earth Sciences, v. 34, p. 169–189.

Isachsen, C.E., Bowring, S.A., and Northrup, C.J.

1993: Geochronologic constraints on the structural evolution of the Point Lake greenstone belt; *in* Exploration Overview 1993 Northwest Territories, Mining, Exploration and Geological Investigations, Indian and Northern Affairs Canada, Northwest Territories Geology Division, p. 34.

Jackson, V.

1989: Metamorphic and structural evolution of Archean rocks in the Keskarrah Bay area, Point Lake, District of Mackenzie, N.W.T.; M.Sc. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 313 p.

Jackson, V.A.

1990: Geology of the Russell–Slemmon Lake area; part of 85 O/4; Indian and Northern Affairs Canada, EGS 1990-11, scale 1:30 000.

1998: The Snare River Project: objectives and initial results; *in* 26th Yellowknife Geoscience Forum; Indian and Northern Affairs Canada, Northwest Territories Chamber of Mines and Resources, Wildlife and Economic Development, Government of the Northwest Territories, Program and Abstracts of Talks and Posters, p. 61–63.

2000: The Snare River Project: results from 2000 mapping; *in* 28th Yellowknife Geoscience Forum Program and Abstracts of Talks and Posters, Northwest Territories Chamber of Mines, p. 33–34.

2003: Preliminary compilation of the geology of the Snare River (1998-2002 results), Wijinnedi Lake, Labrish Lake and Russell Lake area; parts of 85N and 85O; C.S. Lord Northern Geoscience Centre, Yellowknife, Northwest Territories, NWT Open Report 2003-002, scale 1:100 000.

James, D.T. and Mortensen, J.K.

1992: An Archean metamorphic core complex in the southern Slave Province: basement-cover relations between the Sleepy Dragon Complex and the Yellowknife Supergroup; *in* The Tectonic Evolution of the Superior and Slave Provinces of the Canadian Shield, (ed.) K.D. Card and J.E. King; Canadian Journal of Earth Sciences, v. 29, p. 2133–2145.

Johnstone, R.M.

1992: Preliminary geology of the Camsell Lake area, parts of NTS 75M/6,10,11; Department of Indian and Northern Affairs Canada, EGS-1992-2, scale 1:50 000.

Kerswill, J.A.

1993: Models for iron formation-hosted gold deposits; *in* Mineral Deposit Modeling, (ed.) R.V. Kirkham, W.D. Sinclair, R.I. Thorpe, and J.M. Duke; Geological Association of Canada, Special Paper 40, p. 171–199.

Kerswill, J.A. (cont.)

1996: Iron-formation-hosted stratabound gold; *in* Geology of Canadian Mineral Deposit Types; (ed.) O.R. Eckstrand, W.D. Sinclair, and R.I. Thorpe; Geological Survey of Canada, Geology of Canada, no. 8, p. 367–382 (*also* Geological Society of America, The Geology of North America, v. P-1).

Koyaguchi, T. and Kaneko, K.

2001: Thermal evolution of silicic magma chambers after basalt replenishments; Transactions of the Royal Society of Edinburgh: Earth Sciences, v. 91, p. 47–60.

Lambert, M.B.

1988: Cameron River and Beaulieu River volcanic belts of the Archean Yellowknife Supergroup, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Bulletin 382, 145 p.

1998: Stromatolites of the late Archean Back River stratovolcano, Slave Structural Province, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 35, p. 290–301.

in press: Back River volcanic complex — an Archean stratovolcano, Nunavut–Northwest Territories; Geological Survey of Canada, Bulletin 581.

Lambert, M.B., Beaumont-Smith, C., and Paul, D.

1992: Structure and stratigraphic succession of an Archean strato volcano, Slave Province, Northwest Territories; *in* Current Research, Part C; Geological Survey of Canada, Paper 92-1C, p. 189–200.

LeCheminant, A.N. and Heaman, L.M.

1989: Mackenzie igneous events, Canada; middle Proterozoic hotspot magmatism associated with ocean opening; Earth and Planetary Science Letters, v. 96, p. 38–48.

LeCheminant, A.N. and van Breemen, O.

1994: U-Pb ages of Proterozoic dyke swarms, Lac de Gras area, N.W.T.: evidence for progressive breakup of an Archean supercontinent; *in* Geological Association of Canada–Mineralogical Association of Canada, Program with Abstracts, v. 19, p. A62.

LeCheminant, A.N., Buchan, K.L., van Breemen, O., and Heaman, L.M.

1997: Paleoproterozoic continental breakup and reassembly: evidence from 2.19 Ga diabase dyke swarms in the Slave and western Churchill provinces, Canada; *in* Geological Association of Canada–Mineralogical Association of Canada Annual Meeting, v. 22, p. A86.

Leech, A.P.

1966: Potassium-argon dates of basic intrusive rocks of the District of Mackenzie, N.W.T; Canadian Journal of Earth Sciences, v. 3, p. 389–412.

Lemmen, D.S., Duk-Rodkin, A., and Bednarski, J.M.

1994: Late glacial drainage systems along the northwestern margin of the Laurentide Ice Sheet; Quaternary Science Reviews, v. 13, p. 805–828.

Lord, C.S.

1942: Snare River and Ingray Lake map-areas, Northwest Territories; Geological Survey of Canada, Memoir 235, 55 p.

Macdonald, R. and Smith, R.L.

1988: Relationship between silicic plutonism and volcanism: geochemical evidence; Transactions of the Royal Society of Edinburgh: Earth Sciences, v. 79, p. 257–263.

Macfie, R.I.

1987: The Clinton-Colden hornblende gabbro-anorthosite intrusion, Artillery Lake map area, District of Mackenzie; *in* Current Research, Part A; Geological Survey of Canada, Paper 87-1A, p. 681–687.

1989: The petrology and mineralogy of the Clinton-Colden gabbro-anorthosite intrusion, Slave Province, N.W.T.; M.Sc. thesis, University of Ottawa, Ottawa, Ontario, 126 p.

Macfie, R.I., van Breemen, O., and Loveridge, W.D.

1990: U-Pb zircon age of the Clinton-Colden gabbro-anorthosite intrusion, eastern Slave Province, N.W.T.; *in* Radiogenic Age and Isotopic Studies: Report 3; Geological Survey of Canada, Paper 90-2, p. 45–48.

McCall, G.J.H. and Peers, R.

1971: Geology of Binneringie Dyke, Western Australia; Geologische Rundschau, v. 60, p. 1174–1263.

McGlynn, J.C.

1977: Geology of the Bear-Slave Structural Provinces, District of Mackenzie; Geological Survey of Canada, Open File 445, scale 1:1 000 000.

McGlynn, J.C. and Irving, E.

1975: Paleomagnetism of early Aphebian diabase dykes from the Slave Structural Province, Canada; Tectonophysics, v. 26, p. 23–38.

Mortensen, J.K., Henderson, J.B., Jackson, V.A., and Padgham, W.A.

1992: U-Pb geochronology of the Yellowknife Supergroup felsic volcanic rocks in the Russell Lake and Clan Lake areas, southwestern Slave Province; *in* Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2, p. 1–7.

Padgham, W.A.

1987: Access to anorthosite and sheeted dykes in the Chan Formation; *in* Yellowknife Guidebook, A Guide to the Geology of the Yellowknife Volcanic Belt and its Bordering Rocks, (ed.) W.A. Padgham; Geological Association of Canada, Mineral Deposits Division, p. 41–42.

Padgham, W.A. and Fyson, W.K.

1992: The Slave Province: a distinct Archean craton; *in* The Tectonic Evolution of the Superior and Slave Provinces of the Canadian Shield, (ed.) K.D. Card and J.E. King; Canadian Journal of Earth Sciences, v. 29, p. 2072–2086.

Pattison, D.R.M. and Begin, N.J.

1994: Zoning patterns in orthopyroxene and garnet in granulites: implications for geothermobarometry; Journal of Metamorphic Geology, v. 12, p. 387–410.

Pehrsson, S.J. and Kerswill, J.A.

1997a: Geology, Chalco–Strachan lakes and parts of Origin–Truce lakes, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 3395, scale 1:50 000.

1997b: Geology, Ranji–Cotterill lakes, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 3396, scale 1:50 000.

Pehrsson, S.J. and Villeneuve, M.E.

1999: Deposition and imbrication of a 2670–2629 Ma supracrustal sequence in the Indin Lake area, southwestern Slave Province, Canada; *in* NAPMAP Slave Province Project, (ed.) W. Bleeker and W. Davis; Canadian Journal of Earth Sciences, v. 36, p. 1149–1168.

Pehrsson, S.J., Chacko, T., Pilkington, M., Villeneuve, M.E., and Bethune, K.

2000: The Anton terrane revisited: late Archean exhumation of a moderate pressure granulite terrane in the western Slave Province; Geology, v. 28, p. 1075–1078.

Perks, M.J.

1997: The mid-crust of the western Slave Province — geological mapping, geochemistry, and U-Pb geochronology of the Forked Lake area, southwestern Slave Province, N.W.T.; M.Sc. thesis, University of Alberta, Edmonton, Alberta, 95 p.

Pitcher, W.S.

1991: Synplutonic dykes and mafic enclaves; *in* Enclaves and Granite Petrology, (ed.) J. Didier and B. Barbaran; Elsevier, Amsterdam, Netherlands, p. 383–391.

Robertson, D.K. and Folinsbee, R.E.

1974: Lead isotope ratios and crustal evolution of the Slave craton at Ghost Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 11, p. 819–827.

Roddick, J.A. and Armstrong, J.E.

1959: Relict dykes in the coast mountains near Vancouver, British Columbia; Journal of Geology, v. 67, p. 603–613.

Seaton, J.B., Brophy, J.A., and Crux, J.C.

1987: Slave Structural Province; *in* Mineral Industry Report — 1984-85 Northwest Territories, (ed.) J.A. Brophy, J.C. Crux, W.A. Gibbins, P.J. Laporte, W.A. Padgham, and J.B. Seaton; Indian and Northern Affairs Canada, p. 161–270.

Silver, P.G. and Chan, W.W.

1988: Implications for continental structure and evolution from seismic anisotropy; Nature, v. 335, p. 34–39.

Simpson, C. and Wintsch, R.P.

1989: Evidence for deformation-induced K-feldspar replacement by myrmekite; Journal of Metamorphic Geology, v. 7, p. 261–275.

Spear, F.S. and Cheney, J.T.

1989: A petrogenetic grid for pelitic schists in the system SiO_2-Al_2O_3-FeO-MgO-K_2O-H_2O; Contributions to Mineralogy and Petrology, v. 101, p. 149–164.

Spear, F.S., Kohn, M.J., and Cheney, J.T.

1999: P-T paths from anatectic pelites; Contributions to Mineralogy and Petrology, v. 134, p. 17–32.

Stanton, M.S., Tremblay, L.P., and Yardley, D.H.

1954: Chalco Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1023A, scale 1:63 360.

Stesky, R.M.

1998: Spheristat 2 Users Manual; Pangaea Scientific, Brockville, Ontario, 248 p.

Thompson, P.H.

1978: Archean regional metamorphism in the Slave Structural Province: a new perspective on some old rocks; *in* Metamorphism in the Canadian Shield, (ed.) J.A. Fraser and W.W. Heywood; Geological Survey of Canada, Paper 78-10, p. 85–102.

Timoney, K.P., La Roi, G.H., Zoltai, S.C., and Robinson, A.L.

1992: The high subarctic forest-tundra of northwestern Canada: position, width and vegetation gradients in relation to climate; Arctic, v. 45, p. 1–9.

Tremblay, L.P.

1948: Ranji Lake map-area, Northwest Territories; Geological Survey of Canada, Paper 48-10, 7 p.

Tremblay, L.P., Wright, G.M., and Miller, M.L.

1953: Ranji Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1022A, scale 1:63 360.

van Breemen, O. and Henderson, J.B.

1988: U-Pb zircon and monazite ages from the eastern Slave Province and Thelon Tectonic Zone, Artillery Lake area, N.W.T.; *in* Radiogenic Age and Isotopic Studies: Report 2; Geological Survey of Canada, Paper 88-2, p. 73–83.

van Breemen, O., Davis, W.J., and King, J.E.

1992: Temporal distribution of granitoid plutonic rocks in the Archean Slave Province, northwest Canadian Shield; *in* The Tectonic Evolution of the Superior and Slave Provinces of the Canadian Shield, (ed.) K.D. Card and J.E. King; Canadian Journal of Earth Sciences, v. 29, p. 2186–2199.

van Breemen, O., Henderson, J.B., Sullivan, R.W., and Thompson, P.H.

1987: U-Pb zircon and monazite ages from the eastern Slave Province, Healey Lake area, N.W.T.; *in* Radiogenic Age and Isotopic Studies: Report 1; Geological Survey of Canada, Paper 87-2, p. 101–110.

van Breemen, O., King, J.E., and Davis, W.J.

1990: U-Pb zircon and monazite ages from plutonic rocks in the Contwoyto–Nose lakes map area, central Slave Province, District of Mackenzie, Northwest Territories; *in* Radiogenic Age and Isotopic Studies: Report 3; Geological Survey of Canada, Paper 89-2, p. 29–37.

Vernon, R.H.

1986: K-feldspar megacrysts in granites — phenocrysts, not porphyroblasts; Earth-Science Reviews, v. 23, p. 1–63.

Villeneuve, M. and Henderson, J.B.

1998: U-Pb geochronology of Wijinnedi Lake area, Slave Province, District of Mackenzie, Northwest Territories; *in* Current Research 1998-F; Geological Survey of Canada, p. 99–106.

Villeneuve, M.E. and van Breemen, O.

1994: A compilation of U-Pb age data from the Slave Province; Geological Survey of Canada, Open File 2972, 53 p.

Villeneuve, M.E., Henderson, J.R., Hrabi, R.B., Jackson, V.A., and Relf, C.

1997: 2.70–2.58 Ga plutonism and volcanism in the Slave Province, District of Mackenzie, Northwest Territories; *in* Radiogenic Age and Isotopic Studies: Report 10; Geological Survey of Canada, p. 37–60.

Wright, G.M.

1950a: Geology of the Ranji Lake and Ghost Lake areas, Northwest Territories, Canada; Ph.D. thesis, Yale University, New Haven, Connecticut, 117 p.

1950b: Ghost Lake map-area, Northwest Territories; Geological Survey of Canada, Paper 50-13, 10 p.

1954: Ghost Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1021A, scale 1 inch to 1 mile.

Yamashita, K., Creaser, R.A., Stemler, J.U., and Zimaro, T.W.

1999: Geochemical and Nd-Pb isotopic systematics of late Archean granitoids, southwestern Slave Province, Northwest Territories, Canada: constraints for granitoid origin and crustal isotopic structure; *in* NATMAP Slave Province Project, (ed.) W. Bleeker and W. Davis; Canadian Journal of Earth Sciences, v. 36, p. 1131–1147.

Yardley, D.H.

1949: Wecho River (east half), Northwest Territories; Geological Survey of Canada, Paper 49-14, scale 1 inch to 2 miles.

APPENDIX

GEOCHRONOLOGY

Eight U-Pb dates from various separate map units have been produced from the Wijinnedi Lake area, of which seven have been previously published (Villeneuve and Henderson, 1998). This publication should be consulted for a more complete description of the fractions analyzed. What follows is an abstract of the interpretation of each of the earlier seven dates as well as a more complete description and discussion of the eighth, more recent date provided by M.E. Villeneuve (pers. comm., November, 2000). In addition, a U-Pb data table for all zircon, monazite, and titanite fractions analyzed is provided (Table A-1) along with a concordia diagram for each of the eight samples (Fig. A-1a–h) and a summary concordia diagram showing all fractions from all eight samples (Fig. A-1i). Figure A-2 is a summary of interpreted ages for the eight samples analyzed. The location of each sample is noted on Map 2023A. All analyses were made by M.E. Villeneuve and staff of the Geochronology Laboratory of the Geological Survey of Canada, using techniques and standards current for the laboratory at the time of analysis (Villeneuve and Henderson, 1998). Additional geochronological data for rock units immediately north and south of the area is available in Pehrsson and Villeneuve (1999) and Perks (1997) respectively. A comment on an analysis of zircons from the migmatitic intermediate metavolcanic unit (unit ĀYvif.o) (Pehrsson and Villeneuve, 1999), which is relevant to this map area, is provided at the end.

Sample 1. Yellowknife Supergroup metadacite, Wijinnedi domain (unit ĀYvif*)*

The sample is a grey to greenish-grey, fine-grained, volcaniclastic, feldspar- and quartz-phyric metadacite with a small zircon yield. One fraction of a single, clear zircon and two multigrain frosted zircon fractions (fractions 1-A, E, and F, Table A-1) gave nearly identical concordant results and overlap at an average ^{207}Pb/^{206}Pb age of 2673.3 ± 1.4 Ma, which is taken as the age of crystallization. This is near the young end of the volcanic age spectrum for the Yellowknife Supergroup as a whole (Villeneuve and van Breemen, 1994). Fraction 1-B had a ^{207}Pb/^{206}Pb age of 2791 ± 35 Ma and was interpreted as an older xenocryst.

Sample 2. Trondhjemite, Hinscliffe domain (unit ĀH*)*

The trondhjemite sample analyzed represents the oldest of three granitoid phases present in the sample locality outcrop, based on crosscutting relationships, in the best preserved, least deformed, central part of the complex. Three single-grain, low-uranium fractions plot on concordia with a weighted average ^{207}Pb/^{206}Pb age of 2654 ± 4 Ma, which is interpreted as the crystallization age of the trondhjemite. This age is only about 20 Ma younger than that of the Yellowknife Supergroup metadacite and may conceivably represent a deeper, more prolonged expression of the surface volcanic event. A titanite fraction from this sample has a ^{207}Pb/^{206}Pb age of 2610 ± 4 Ma, which is interpreted as an estimate of resetting of the U-Pb system on cooling below about 600°C, the approximate closure temperature of titanite (Heaman and Parrish, 1991). This titanite age groups with the package of U-Pb zircon and monazite ages from the younger intrusions of the predominantly granulite-grade Ghost domain (Fig. A-1i).

Sample 3. Quartz diorite, Ghost domain (unit Āqd*)*

This sample of quartz diorite from the Ghost domain represents the only intrusive unit that occurs both in the Ghost domain and the Dauphinee domain. Three zircon fractions agree within error and overlap concordia with a weighted average of ^{207}Pb/^{206}Pb ages at 2605 ± 3 Ma. The large error of one of these is due to the low U content and small fraction size leading to a high common-to-radiogenic Pb ratio. A fourth near-concordant fraction at 2691 ± 1 Ma, suggests the incorporation of inclusions or partial melts of Yellowknife Supergroup supracrustal rocks in the source area of the intrusion.

Sample 4. Granodiorite gneiss, Ghost domain (unit Ātgn*)*

This sample is a grey, well layered and well foliated, fine-grained granodioritic phase of an interlayered granodiorite, granite, amphibolite gneiss sequence that is part of a gneissic unit that elsewhere is commonly dominated by tonalite (*see* sample 8). Visible cores are evident in many of the zircons. Although cores were not apparent in the fractions analyzed, the scatter of the data would suggest they are present. The near concordant fraction 4-B together with the similar ^{207}Pb/^{206}Pb age of the more discordant fraction 4-E suggests an age of about 2605 Ma as a reasonable estimate of the of the maximum age of one of the components of the gneiss (Table A-1; Fig. A-1d). Fraction 4-D, a distinct zircon different from those in the other fraction, plots slightly above concordia with a ^{207}Pb/^{206}Pb age of 2640 ± 1 Ma. This zircon plots close to those from the dominant tonalitic phase of the gneiss of sample 8 (Fig. A-1h, i). Fraction 4-F would appear to be on a similar lead-loss line as fractions of the tonalite (Fig. A-1i).

Sample 5. Megacrystic syenogranite, Ghost domain (unit Āgm.o*)*

The sample comes from the larger, originally granulite-grade, but now largely autoretrogressed (Chacko et al., 1995a), massive, homogeneous, coarsely and densely K-feldspar megacrystic granite body south of Ghost Lake. The four single-grain analyses have similar ^{207}Pb/^{206}Pb ages, although fraction 5-A may have a minor inherited Pb component. A regression line through all four fractions results in an upper intercept age

Table A-1. Geochronology U-Pb data and sample locations

							Radiogenic ratios (± 1σ, %)[f]				
Fraction[a]	Wt.[b] (µg)	U (ppm)	Pb[c] (ppm)	^{206}Pb[d] / ^{204}Pb	Pb[e] (pg)	^{208}Pb[f] / ^{206}Pb	^{207}Pb / ^{235}U	^{206}Pb / ^{238}U	^{207}Pb / ^{206}Pb	^{207}Pb / ^{206}Pb[g] age (Ma)	Disc. %[h]
1. Yellowknife Supergroup metadacite (unit ĀYvif), Wijinnedi domain (Z2915[i]; 63 58′30″, 115 22′00″)											
A-clear(Z)	8	121	71	512	42	0.149	12.890 ± 0.26	0.5133 ± 0.26	0.18214 ± 0.12	2673 ± 4	0.09
B-clear(Z)	2	106	68	203	27	0.196	14.461 ± 1.46	0.5358 ± 0.77	0.19576 ± 1.09	2791 ± 35	1.13
E-frost(Z)	6	113	69	847	25	0.197	12.943 ± 0.17	0.5151 ± 0.14	0.18222 ± 0.08	2673 ± 3	-0.24
F-frost(Z)	7	121	72	2469	11	0.172	12.879 ± 0.14	0.5125 ± 0.14	0.18226 ± 0.06	2674 ± 2	0.28
2. Trondhjemite (unit ĀH), Hinscliffe domain (Z2914[i]; 63 53′53″, 115 29′43″)											
B (Z)	16	51	29	928	1	0.123	12.677 ± 0.22	0.5109 ± 0.25	0.17997 ± 0.09	2653 ± 3	-0.36
C (Z)	6	131	72	471	23	0.083	12.592 ± 0.36	0.5072 ± 0.43	0.18007 ± 0.16	2654 ± 5	0.41
E (Z)	5	55	30	249	33	0.077	12.717 ± 0.69	0.5087 ± 0.75	0.18131 ± 0.36	2665 ± 12	0.63
Z (T)	1288	30	18	700	1743	0.207	12.009 ± 0.16	0.4965 ± 0.09	0.17542 ± 0.11	2610 ± 4	0.53
3. Quartz diorite (unit Āqd), Ghost domain (Z2991[i]; 63 46′18″, 115 24′22″)											
A (Z)	5	221	143	5331	7	0.269	13.119 ± 0.11	0.5165 ± 0.10	0.18421 ± 0.03	2691 ± 1	0.31
B (Z)	15	103	62	4283	11	0.229	11.989 ± 0.11	0.4969 ± 0.10	0.17498 ± 0.03	2606 ± 1	0.25
E (Z)	6	137	79	945	26	0.169	12.006 ± 0.15	0.4982 ± 0.12	0.17478 ± 0.07	2604 ± 2	-0.1
F (Z)	5	50	28	285	12	0.156	11.899 ± 0.64	0.4966 ± 0.52	0.17377 ± 0.40	2594 ± 13	-0.24
4. Granodiorite gneiss (unit Āt-gn.o), Ghost domain (Z2993[i]; 63 48′59″, 115 15′04″)											
B (Z)	4	808	413	10507	10	0.027	11.919 ± 0.10	0.4951 ± 0.09	0.17462 ± 0.03	2602 ± 1	0.46
D (Z)	4	177	108	5175	2	0.219	12.524 ± 0.11	0.5086 ± 0.10	0.17859 ± 0.03	2640 ± 1	-0.49
E (Z)	10	162	83	8129	6	0.077	11.522 ± 0.11	0.4777 ± 0.09	0.17492 ± 0.05	2605 ± 2	4.08
F (Z)	8	196	102	6066	8	0.058	11.943 ± 0.14	0.4884 ± 0.13	0.17736 ± 0.03	2628 ± 1	2.98
5. Megacrystic syenogranite (unit Āgm.o), Ghost domain (Z3220[i]; 63 57′50″, 115 15′02″)											
A (Z)	10	140	79	8574	4	0.179	11.709 ± 0.12	0.4875 ± 0.11	0.17419 ± 0.03	2598 ± 1	1.8
B (Z)	7	278	147	8384	7	0.084	11.738 ± 0.10	0.4893 ± 0.09	0.17400 ± 0.03	2597 ± 1	1.36
D (Z)	5	231	127	4718	7	0.12	11.879 ± 0.10	0.4946 ± 0.09	0.17418 ± 0.04	2598 ± 1	0.35
E (Z)	4	200	108	2382	11	0.098	11.904 ± 0.11	0.4956 ± 0.09	0.17422 ± 0.04	2599 ± 1	0.18
X (M)	18	1792	14525	33717	24	22.497	9.411 ± 0.11	0.3950 ± 0.10	0.17281 ± 0.03	2585 ± 1	19.94
Y (M)	5	1165	9442	19864	8	17.992	11.658 ± 0.10	0.4879 ± 0.09	0.17330 ± 0.03	2590 ± 1	1.32
Z (M)	4	659	5044	9225	9	16.718	11.779 ± 0.10	0.4935 ± 0.09	0.17309 ± 0.03	2588 ± 1	0.08
6. Granite (unit Āg) Ghost domain, (Z2992[i]; 63 57′05″, 115 15′31″)											
A (Z)	3	226	118	2223	9	0.098	11.414 ± 0.19	0.4789 ± 0.18	0.17285 ± 0.05	2585 ± 2	2.93
D (Z)	8	744	412	9611	19	0.156	11.535 ± 0.10	0.4843 ± 0.08	0.17275 ± 0.03	2584 ± 1	1.8
E (Z)	5	370	197	8054	7	0.083	11.822 ± 0.11	0.4939 ± 0.10	0.17361 ± 0.03	2593 ± 1	0.26
F (Z)	2	424	222	6683	4	0.066	11.715 ± 0.15	0.4909 ± 0.14	0.17310 ± 0.03	2588 ± 1	0.63
7. Orthopyroxene granite (unit Āg.o), Ghost domain (Z3371[i]; 63 45′59″, 114 54′15″)											
A (Z)	3	153	84	2579	5	0.118	11.832 ± 0.12	0.4951 ± 0.11	0.17331 ± 0.05	2590 ± 2	-0.14
C (Z)	5	109	59	1373	11	0.105	11.802 ± 0.13	0.4938 ± 0.12	0.17336 ± 0.06	2590 ± 2	0.16
E (Z)	6	129	70	6120	1	0.096	11.784 ± 0.11	0.4937 ± 0.10	0.17310 ± 0.05	2588 ± 2	0.04
8. Tonalite gneiss (unit Āt-gn.o), Ghost domain (Z5835[i]; 63 52′16″, 115 04′48″)											
A1 (Z)	13	209	117	1767	47	0.109	12.365 ± 0.13	0.5041 ± 0.09	0.17790 ± 0.07	2633 ± 2	0.09
A2 (Z)	19	117	66	18465	4	0.118	12.486 ± 0.12	0.5065 ± 0.10	0.17878 ± 0.04	2642 ± 2	-0.01
B1 (Z)	18	262	141	27591	5	0.076	12.335 ± 0.11	0.5009 ± 0.09	0.17862 ± 0.04	2640 ± 1	1.04
B2 (Z)	17	232	125	17191	7	0.076	12.285 ± 0.11	0.4998 ± 0.09	0.17826 ± 0.04	2637 ± 1	1.09
C1 (Z)	11	171	96	13435	4	0.107	12.567 ± 0.11	0.5092 ± 0.09	0.17901 ± 0.04	2644 ± 1	-0.43
C2 (Z)	15	191	106	20015	3	0.094	12.525 ± 0.11	0.5065 ± 0.09	0.17935 ± 0.04	2647 ± 1	0.24
M1 (M)	7	192	9144	13050	3	110.694	11.680 ± 0.12	0.4912 ± 0.09	0.17246 ± 0.05	2582 ± 2	0.27
M4 (M)	10	310	8936	27156	3	66.367	11.777 ± 0.12	0.4928 ± 0.10	0.17331 ± 0.04	2590 ± 2	0.33
M5 (M)	16	142	10582	4958	14	172.56	11.842 ± 0.12	0.4938 ± 0.10	0.17395 ± 0.05	2596 ± 2	0.42
M6 (M)	19	89	8242	1200	45	215.478	11.811 ± 0.16	0.4935 ± 0.14	0.17358 ± 0.11	2592 ± 4	0.31

[a] Fractions include zircon (Z), monazite (M), and titanite (T), all zircon fractions are abraded;

[b] Error on weight = ±0.001 mg;

[c] Radiogenic Pb;

[d] Measured ratio corrected for spike and Pb fractionation of 0.09 ± 0.03%/AMU;

[e] Total common Pb on analysis, corrected for fractionation and spike, of blank model Pb composition;

[f] Corrected for blank and spike Pb and U and common Pb (Stacey-Kramers model Pb equal to the ^{207}Pb/^{206}Pb age);

[g] Age error is ±2SE in Ma;

[h] Discordance along a discordia to origin;

[i] GSC geochronolgy lab number.

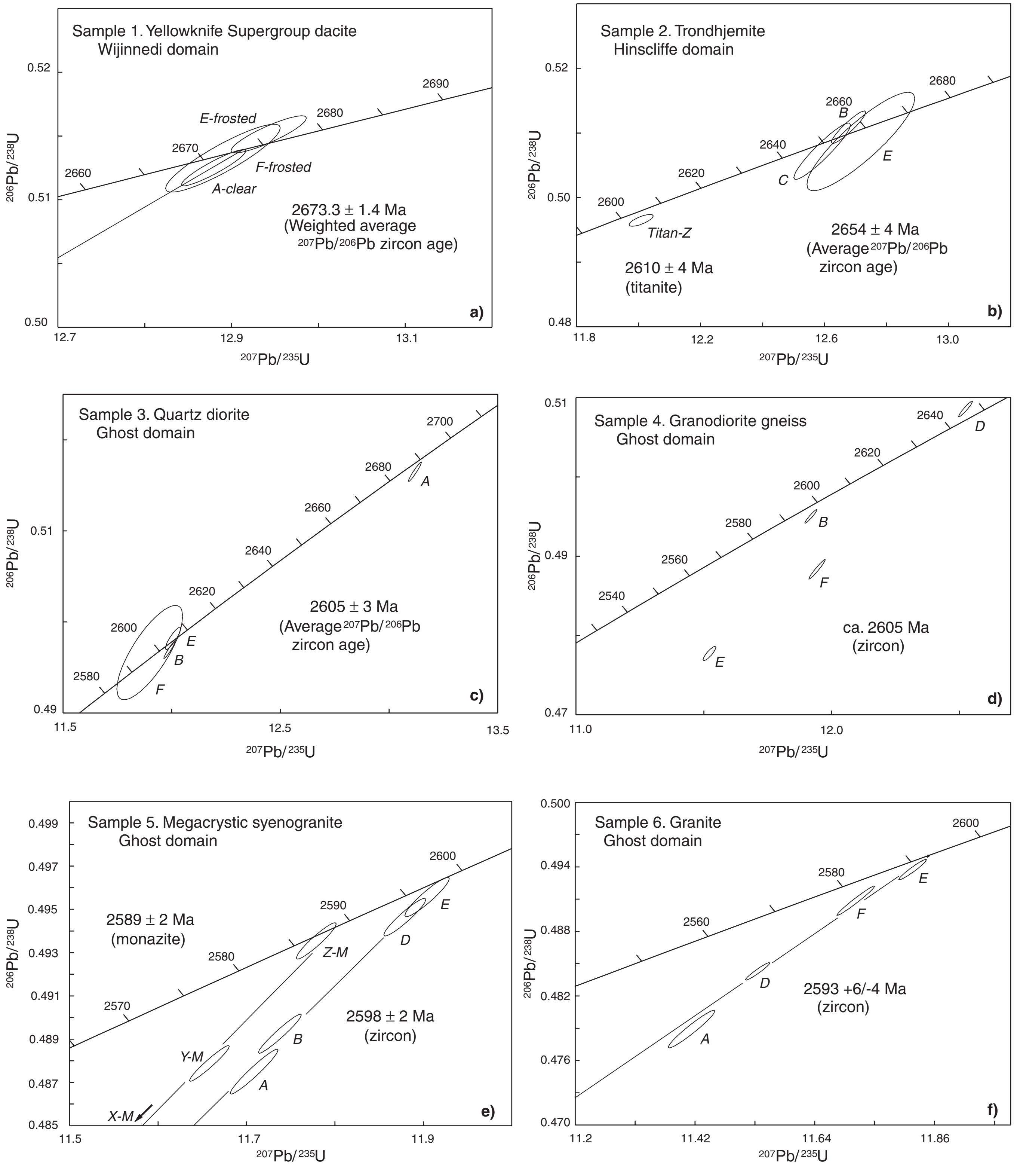


Figure A-1. *Concordia diagrams for samples analyzed from Wijinnedi Lake area. Error ellipses are shown at 2σ. Figure A-1a–g are from Villeneuve and Henderson (1998). Figure A-1h is bsed on new data. Figure A-1i represents an amalgamation of all data from all eight samples.*

of 2598 ± 2 Ma (lower intercept age is 127 ± 400 Ma, MSWD=4). Analyses of three, variably discordant, single-grain, monazite fractions from the same sample resulted in an upper intercept age of 2589 ± 2 Ma, with a lower intercept age of 40 ± 36 Ma (MSWD=9)(Fig. A-1e).

Sample 6. Granite, Ghost domain (unit Āg*)*

This sample comes from a small island in Ghost Lake south of the Ghost River. It is one of several small plugs to sills of granite that are particularly common along the north shore of Ghost Lake, east of the sample site. They are considered to be

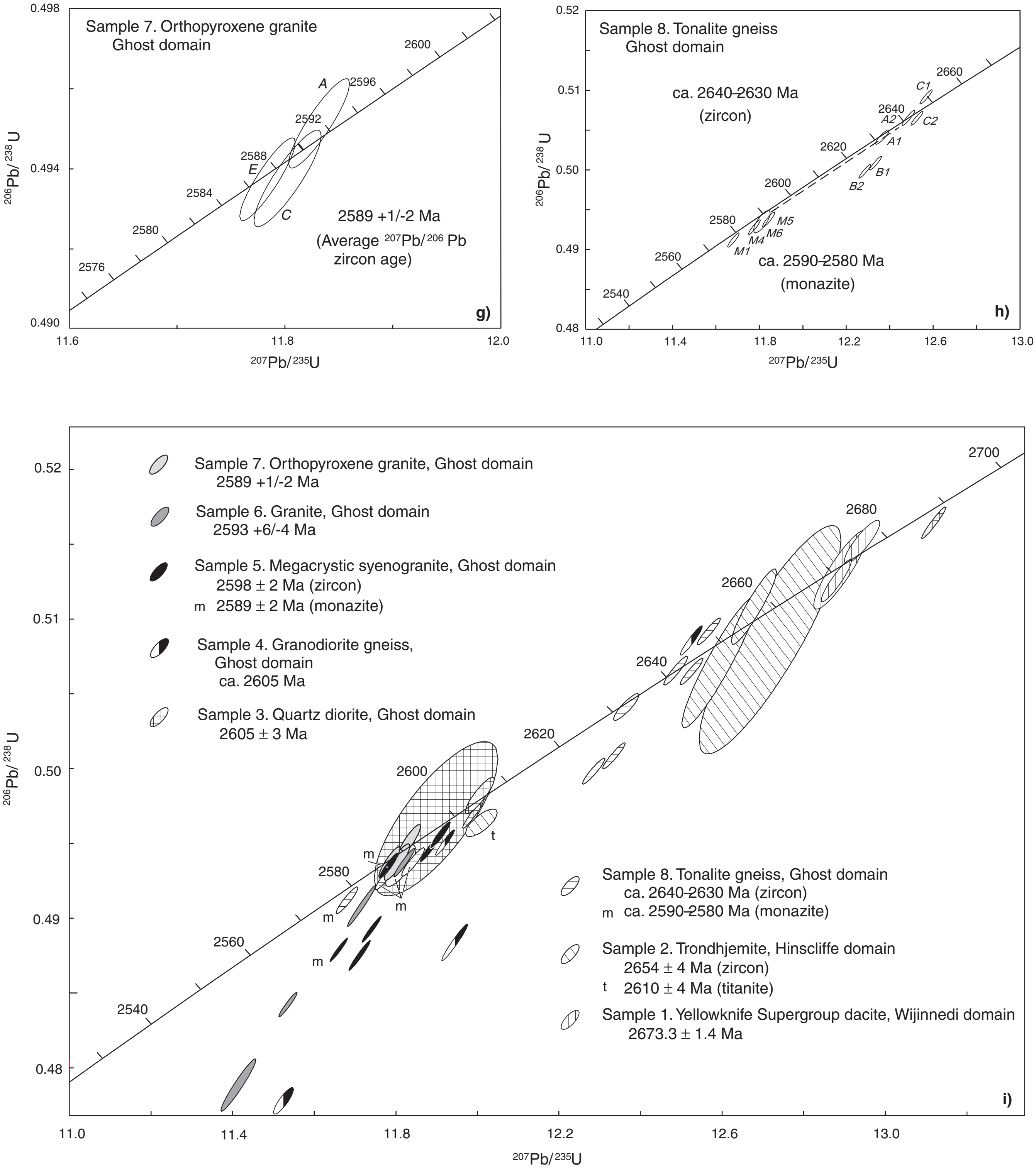


***Figure A-1** (cont.)*

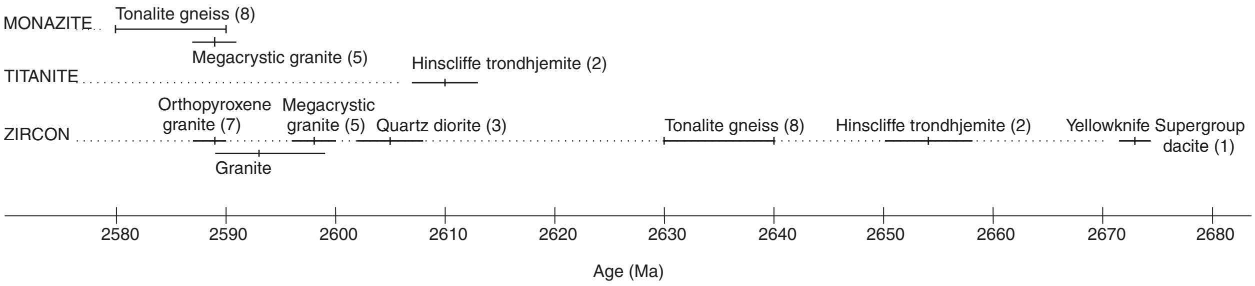


***Figure A-2**. Summary of age distributions with confidence levels for the eight samples analyzed for the zircon, titanite, and monazite mineral systems.*

small satellites related to the large granite to granodiorite body with associated pegmatite, 5 km northeast of the sample site. The zircons from the granite represent a surprisingly heterogeneous population with high-uranium cores present in many cases. Four single-grain fractions form the basis of a poorly defined discordia that includes one nearly concordant point with an upper intercept age of 2593 +6/-4 Ma and a lower intercept age of 741 ± 410 Ma (MSWD=15). The scatter of the analyses about the discordia line is due to inheritance. Since the ^{207}Pb/^{206}Pb ages for all fractions fall within 10 Ma of each other and two of the fractions are close to being concordant (Table A-1) the upper intercept age is considered to be a reasonable estimate of the age of intrusion.

Sample 7. Orthopyroxene granite, Ghost domain (unit Āg.o*)*

This granite from the southeast corner of the map area is petrographically quite similar to the granite from the northern part of the domain described above (sample 6) except for the presence of orthopyroxene, suggesting that it crystallized under granulite-grade conditions. The three zircon fractions analyzed overlap on concordia yielding an average ^{207}Pb/^{206}Pb age of 2589 +1/-2 Ma. This is interpreted as the age of crystallization of the granite.

Sample 8. Tonalite gneiss, Ghost domain (unit Ātgn.o*)*

The tonalite gneiss sample comes from the north shore of Ghost Lake, north of the widest part of the lake. The gneiss at that locality consists of primarily of texturally and compositionally varied tonalitic layers and less than 15% granite-pegmatite layers. The layered tonalite is similar and considered comagmatic with the main, more homogeneous tonalite of unit Āt. The rock is buff- to yellowish-weathering, medium grey, well laminated, medium-grained, moderately recrystallized orthopyroxene- and biotite-bearing tonalite.

Three zircon morphologies can be recognized in the rock. Fractions 8-A1 and 8-A2 consist of highly elongate zircons with slightly rounded or broken tips and long inclusions running the length of the crystal. These fractions give the most concordant results and would be the least likely to contain inheritance (Fig. A-1h). Fractions 8-B1 and 8-B2 represent more equant, slightly rounded grains that are light pink. Minor inclusions and rare fractures are present. These two fractions show clear evidence of Pb loss (Fig. A-1h). Fractions C1 and C2 are deep pink and have smooth irregular surfaces. Under transmitted light, possible cores are visible in some grains, suggesting a core-overgrowth relationship. This may be borne out by the older ages of ca. 2650 Ma arrived at for the two fractions.

Four fractions consisting of single monazite grains were analyzed. Fractions 8-M4, 8-M5, and 8-M6 all cluster around 2596–2590 Ma, suggesting growth during retrograde conditions or possible partial resetting of original magmatic monazite. Fraction 8-M1, the youngest grain with a ^{207}Pb/^{206}Pb age of 2582 ± 1.5 Ma (0.3% discordant), may point toward a younger thermal event responsible for scattering the other monazite ages. It is the smallest and thinnest grain and likely represents the outer portion of a larger grain. It would be the grain most susceptible to thermal effects.

In summary, zircon fractions 8-C1 and 8-C2 are interpreted to consist of older cores with younger overgrowths and give a maximum age of crystallization. Zircon fraction 8-A1 may have a component of Pb loss as is the case for zircon fractions 8-B1 and 8-B2. As such, it marks the minimum age of crystallization at 2633 ± 2.1 Ma. Concordant fraction 8-A2 may be a more realistic interpretation of the age of the rock at 2641.6 ± 1.5 Ma. The overall Pb-loss trajectory is consistent with resetting at ca. 2590–2580 Ma, as shown by the scattered monazite fractions.

A comment: metavolcanic migmatite leucosome (unit ĀYvif.o*)*

A sample of orthopyroxene-bearing leucosome from the migmatitic volcanic rock unit has been geochronologically analyzed by Pehrsson et al. (2000). The sample comes from a small island in Daran Lake, 1.2 km north of the map border, that represents the northernmost exposed extent of the unit (Pehrsson et al., 2000; S. Pehrsson, pers. comm., 2001). Three single-grain fractions of the dominant zircon population based on crystal morphology have an tightly grouped, close to concordant, average ^{207}Pb/^{206}Pb age of 2588 ± 1 Ma and were interpreted by Pehrsson et al. (2000) as having crystallized during granulite-facies metamorphism. This age is also within error of the age of a major granite suite within the area (units Ag, Ag.o (described above)). Two single zircon fractions of a second morphological population, slightly more discordant, have ^{207}Pb/^{206}Pb ages of 2623 ± 3 Ma and 2626 ± 6 Ma. Pehrsson et al. (2000) considered that this older population represented either zircons crystallized during an earlier, ca. 2624 Ma metamorphic event or zircons inherited from the original volcanic protolith. Given the morphological similarity of the ca. 2624 Ma zircons to the youngest of a suite of zircons in low metamorphic grade metagreywacke at Damoti Lake (Fig. 6) whose ages ranged between 3048 Ma and 2629 Ma, together with the lack of any evidence for an earlier metamorphic event, they opted for the latter interpretation with the older zircon population representing the age of crystallization of an original volcanic protolith; however, more recent data on the tonalite and tonalitic gneiss (units At.o and Atgn.o) that crystallized under granulite conditions and intrude the migmatitic metavolcanic rocks (*see* 'Tonalite' section) are reported in the above section to have crystallized between 2640 Ma and 2630 Ma. On the basis of this new data it is suggested that the other possible interpretation suggested by Pehrsson et al. (2000), that the older population of zircons in the migmatite leucosome may also be metamorphic in origin, is correct and is presumably related to the emplacement of the tonalite. The monazite grains analyzed from the tonalite gneiss sample have a similar age to that of the dominant zircon population in the metavolcanic leucosome (Fig. A-1i).