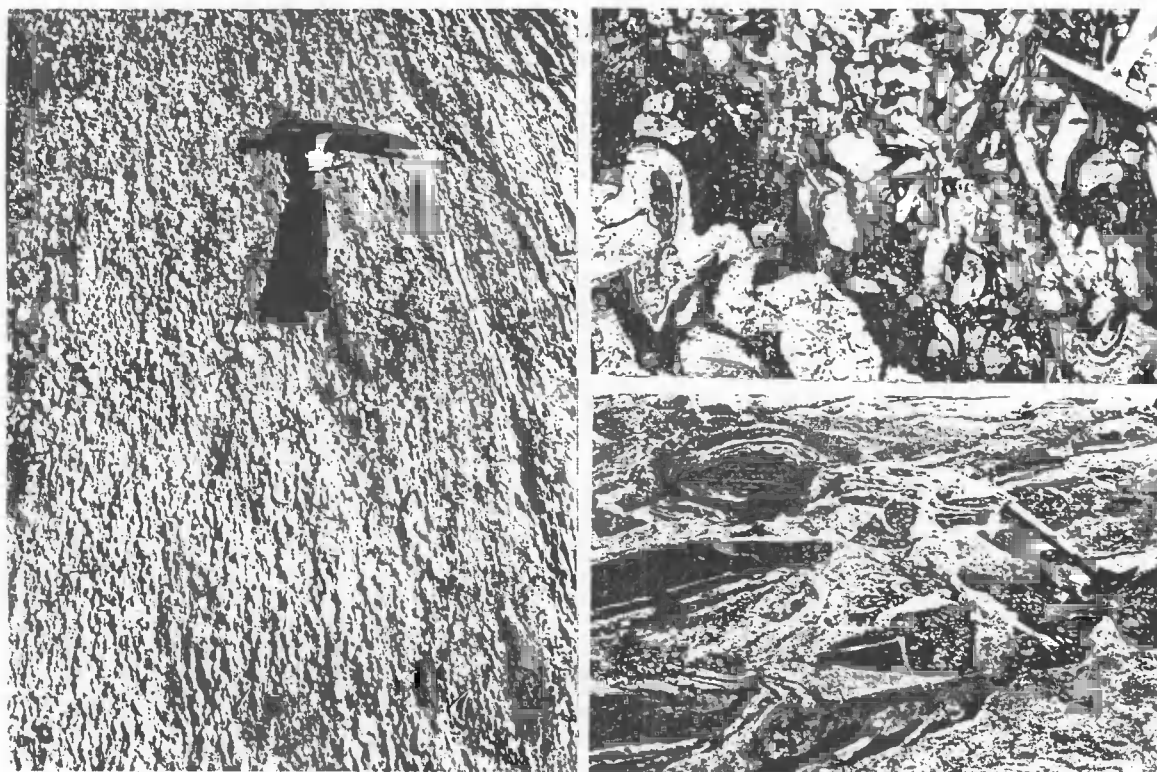
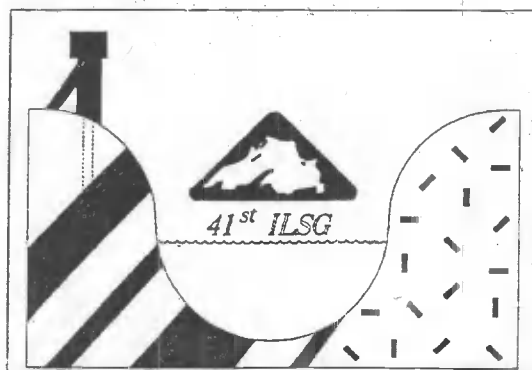

*Geology, Structure and Age Relationships of
the Manitouwadge Greenstone Belt and the
Wawa-Quetico Subprovince Boundary,
Northwestern Ontario*



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Field Trip Guidebook**



**Geology, structure and age relationships
of the Manitouwadge greenstone belt
and the Wawa-Quetico subprovince boundary,
northwestern Ontario¹
Field guidebook**

by

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Frontispiece: Sillimanite knots in metamorphosed altered felsic rocks near the Willroy mine (left). Folded and brecciated iron formation (upper right). Mafic to intermediate metavolcanic inclusions in tonalite along the folded attenuated extensions of the Manitouwadge greenstone belt (lower right).

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INTRODUCTION

Despite its relatively small size, the Manitouwadge greenstone belt is host to significant economic Cu-Zn deposits, the most productive being Noranda's Geco mine. It is among the largest of Canadian volcanogenic massive sulphide deposits, with total lifetime production projected at 55.9 million tonnes. The deposits of the belt have been known since the 1950's, and early theories of their origin and geological history are a microcosm of the major debates on massive sulphide petrogenesis. In more recent years, a volcanogenic origin has been generally accepted for the Manitouwadge deposits (Friesen et al., 1982; Bakker et al., 1985; Williams et al., 1990), although their stratigraphic, structural and tectonic setting, their relationships and detailed petrogenesis were largely unknown or controversial.

In recognition of the economic interest of the area and the impending closure of the Geco mine, the Manitouwadge project was conceived to further understanding of mineralization, alteration, metamorphism and deformation and their relationships in the Manitouwadge camp. Within this mandate, the project has focussed on 1) the evolution of the Manitouwadge belt; 2) the influences of primary depositional setting, deformation, and metamorphism on the belt and its mineral deposits; and 3) the relationship of the belt to the enclosing plutonic rocks and the adjacent Wawa-Quetico boundary.

The Manitouwadge project is still ongoing; although most of the field component has been completed, laboratory and follow-up investigations continue. The first part of this volume is intended as a comprehensive interim report of our results to date. Some of the material is of a preliminary nature although, we hope, presented with sufficient background and interpretation to stimulate discussion. The second part of this volume is a field guide to the Manitouwadge area, consisting of stops, outcrop descriptions and maps arranged as a series of day or part-day field trips. The organization of the field trips is mostly thematic, with stops selected to demonstrate critical or representative observations.

Methodology

Our approach in the Manitouwadge project has been primarily field-based, entailing mapping varying in scale from 1:5000 in the area of known economic Cu-Zn deposits, to 1:10000 and 1:20000 mostly along the extensions of the belt and in peripheral areas. In detailed mapping, we made use of existing grids, cut for mineral exploration purposes and varying greatly in age, condition and utility. The co-operation of private-sector companies, prospectors, and Ontario government geologists contributed to the success of the project. Unpublished data, mostly in the form of geological, geophysical and grid maps, drill logs, reports and geochemical analyses, were contributed by Noranda Inc., Granges Inc., Minnova Inc., and A. Turner (prospector). F. Breaks and H. Williams (Ontario Geological Survey) allowed access to their field notes, station location maps and thin sections, providing useful supplemental observations in areas of plutonic rock and near the Quetico boundary. In combination with the published geological maps of Pye (1957), Milne (1974) and Williams and Breaks (1990b), we used this information to direct mapping and, in some cases, to interpolate or extrapolate geological contacts. Our field observations were compiled in digital maps and data bases.

Aeromagnetic information, in combination with field observations, proved to be a powerful tool in resolving the complex structure of the Manitouwadge belt. The aeromagnetic data were collected during high resolution surveys, flown by Dighem Surveys and Processing Inc. for Noranda Exploration Ltd., and released by Noranda to the Geological Survey of Canada for compilation by the Geophysical Data Centre. Published (Geological Survey of Canada, 1993a, 1993b; Zaleski and Peterson, 1995) and unpublished aeromagnetic maps based on these surveys were invaluable for focussing our mapping, extrapolating contacts in areas of poor exposure, and interpreting the locations of faults and diabase dykes. The accompanying 1:25000 map, showing geology draped over shaded relief of total field magnetics, demonstrates the close correspondence between map-scale structure and aeromagnetic expression.

Field observations and interpretations were followed up by laboratory studies, including petrographic observations, in particular petrofabric studies of oriented samples and documentation of metamorphic mineral assemblages. In conjunction with metamorphic studies, electron-microprobe determination of mineral compositions has been initiated. Whole-rock geochemical analyses, including major, trace and rare-earth elements, were done on a sample suite representing apparently unaltered or 'least-altered' metavolcanic rocks and their altered equivalents, metasedimentary rocks from the Manitouwadge belt and Quetico subprovince, iron formation, and intrusive rocks. Geochronology samples were collected with the aim of dating primary volcanism, bracketing deformational events and metamorphism, bracketing sedimentation and determining provenance, and dating intrusive activity.

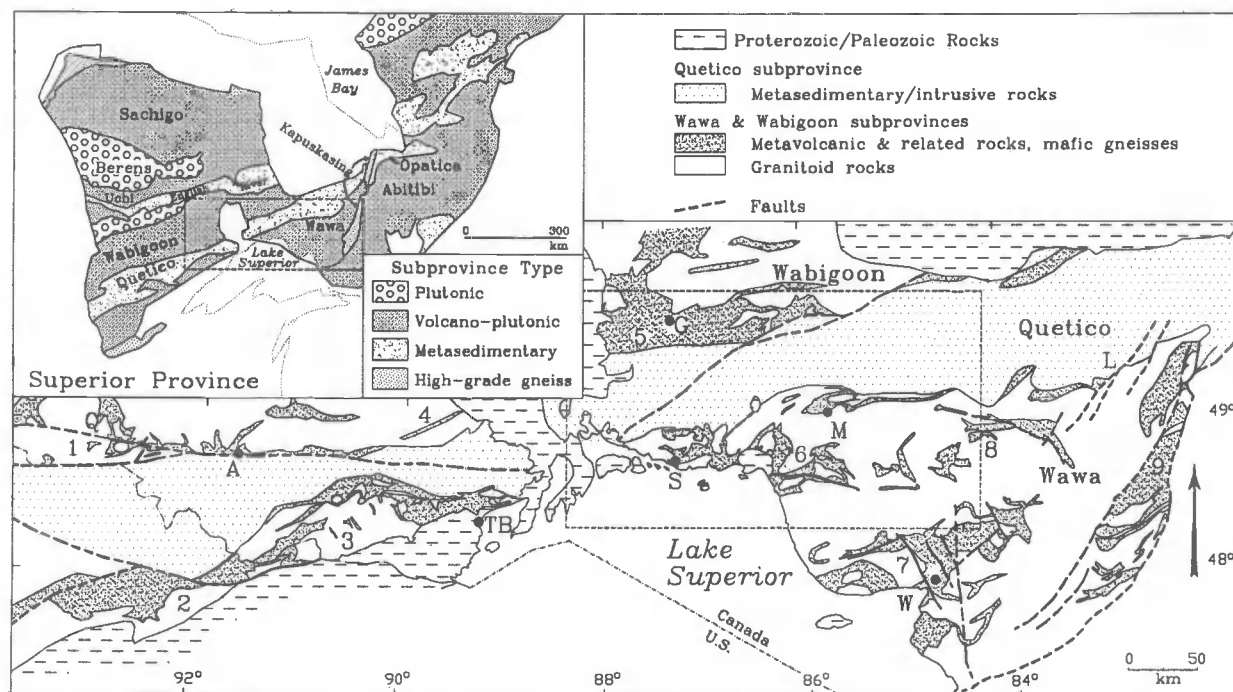


FIG. 1. Tectonic map of the south central Superior Province showing the Wawa, Quetico and Wabigoon subprovinces with the area of Figure 2 outlined. 1 = Rainy Lake area; 2 = Vermilion district, Minnesota; 3 = Shebandowan greenstone belt; 4 = Max Creek conglomerate; 5 = Beardmore-Geraldton belt; 6 = Schreiber-Hemlo greenstone belt; 7 = Michipicoten greenstone belt; 8 = Dayohessarah-Kabinagami greenstone belt; 9 = Kapuskasing structural zone; Q = Quetico fault; A = Atikokan; TB = Thunder Bay; S = Schreiber; G = Geraldton; M = Manitouwadge; W = Wawa; L = Lepage fault zone. Adapted from Williams et al. (1991) and Percival (1989). The inset shows the subprovinces of the southern Superior Province after Card and Ciesielski (1986), and the area detailed in the tectonic map.

REGIONAL SETTING

Relationships in the Wawa subprovince

The Manitouwadge greenstone belt lies in the volcano-plutonic Wawa subprovince of the Superior Province, near the boundary with the metasedimentary-migmatitic Quetico subprovince (Fig. 1). The belt is a remnant of volcanic and sedimentary rocks, highly deformed and metamorphosed to upper amphibolite facies. Williams et al. (1991) proposed correlations with the Dayohessarah-Kabinagami and Schreiber-Hemlo greenstone belts, and suggested that these are dismembered parts of an originally continuous greenstone terrane. Regional mapping established that supracrustal rocks (mostly mafic metavolcanic rocks) extend east of Manitouwadge to the Moskinabi belt, and southeast to the Faries Lake belt (Fig. 2), as a semicontinuous unit and as inclusions in foliated plutonic rocks (Williams and Breaks, 1989; 1990a, b; Williams et al., 1992). The Moshkinabi and Faries Lake belts are intruded by a layered complex comprising gabbro, leucogabbro, anorthosite and peridotite (Williams and Breaks, 1989; 1990b).

The Manitouwadge belt is bounded to the west and south by foliated multiphase plutonic rocks, collectively known as the Black Pic batholith (Fig. 2). To the south, foliations in the Black Pic batholith define a structural dome (Williams and Breaks, 1990a, b), and plutonic rocks extend to the Schreiber-Hemlo greenstone belt (Williams et al., 1991). To the west of the Manitouwadge greenstone belt, abundant inclusions of mafic to intermediate supracrustal rocks were likely derived from the belt. On aeromagnetic maps, the corresponding area shows a pronounced striping parallel to foliation trends (accompanying 1:25000 map), interpreted to reflect the presence of discontinuous supracrustal septa. Aeromagnetic lineaments that trend southwest toward the Schreiber-Hemlo greenstone belt support linkage of the greenstone terranes (Williams et al., 1991). In contrast, the Black Pic batholith south of Manitouwadge contains few supracrustal inclusions and the aeromagnetic signature is low and flat.

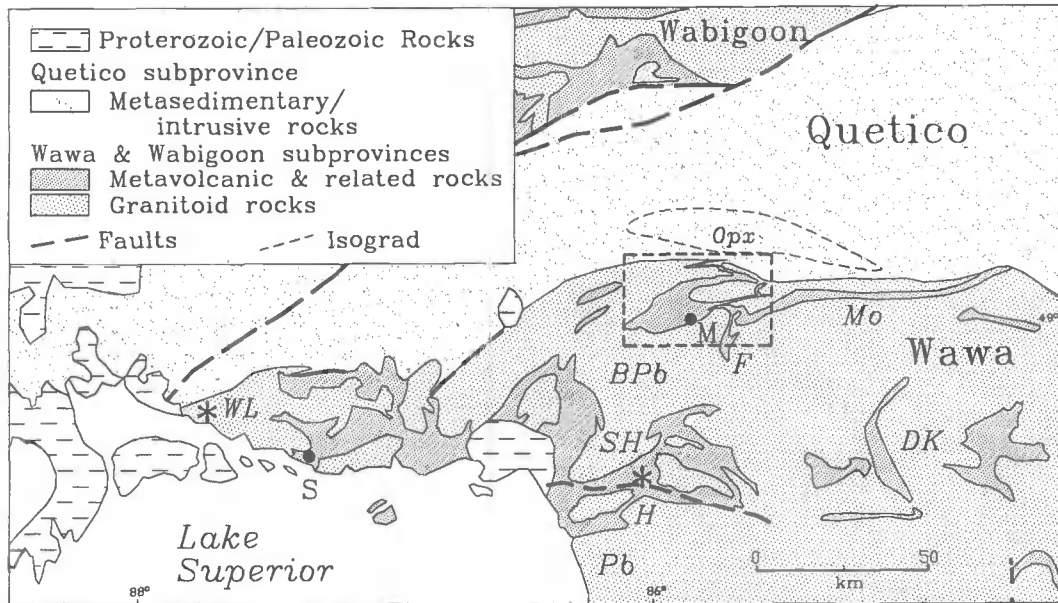


FIG. 2. The Wawa, Quetico and Wabigoon subprovinces with the Manitouwadge greenstone belt and the area of Figure 3 outlined. SH = Schreiber-Hemlo greenstone belt; F = Faries Lake belt; Mo = Moshkinabi greenstone belt; DK = Dayohessarah-Kabinakagami greenstone belt; BPb = Black Pic batholith; Pb = Pukaskwa batholith; Opx = orthopyroxene granulite isograd; S = Schreiber; M = Manitouwadge; WL = Winston Lake Zn-Cu mine; H = Hemlo Au camp. Adapted from Williams et al. (1991).

Structural and tectonic setting

Stockwell (1964) originally used structural trends and style to subdivide the Superior Province (Fig. 1), contrasting the linear east-west trends of the Quetico subprovince with the curvilinear trends of the Wawa and Wabigoon subprovinces, the latter defined by remnants of greenstone belts lying between domical batholiths. Subsequently, lithological features were also used to define subprovinces, by contrasting the dominantly volcanic origin of supracrustal rocks in the volcano-plutonic terranes with the sedimentary origin of the Quetico subprovince (Stockwell, 1970; Card and Ciesielski, 1986). In some cases, boundaries were defined to exclude mafic metavolcanic rocks from the Quetico subprovince (Williams, 1991). Supracrustal sequences in the volcano-plutonic subprovinces resemble arc-type deposits in modern orogenic belts (Card, 1990). Among the recent tectonic models for the Wabigoon, Quetico and Wawa subprovinces, an arc-accretionary model has been favoured by Percival (1989), Percival and Williams (1989) and Williams (1990). In this scenario, the Quetico subprovince represents an accretionary complex that accumulated southward from the margin of the Wabigoon arc, above a northerly dipping subduction zone. Collision of the Wawa arc from the south amalgamated the three subprovinces. The relatively high temperature-low pressure metamorphic assemblages preserved within the Quetico and northern Wawa subprovinces could be explained by ridge subduction or by post-subduction thermal relaxation (Percival, 1989).

The identification of structures produced by dextral transcurrent motion is a feature common to various structural studies of the Quetico subprovince and its boundaries. Both ductile and brittle structures have been identified, typically in shear zones or deformation zones near subprovince boundaries (Fig. 1). Examples include the Quetico fault (Borradaile et al., 1988), the Vermilion district (Hudleston et al., 1988), the Beardmore-Geraldton belt (Devaney and Williams, 1989), the Shebandowan belt (Corfu and Stott, 1986), the Shebandowan-Quetico boundary (Borradaile and Spark, 1991), and the Hemlo-Schreiber belt (Williams, 1989). Zones of distributed dextral motion have been noted in the western Quetico (Bauer et al., 1992), as have dextral shear zones well within the Wawa subprovince, for example in the Dayohessarah-Kabinakagami greenstone belt (Leclair, 1990). In the Shebandowan belt, the timing of dextral transcurrent motion (local D₂) is constrained to between 2689±3/-2 and 2684±6/-3 Ma (Corfu and Stott, 1986); however, temporal correlation with dextral motion elsewhere in the Wawa-Quetico-Wabigoon region is uncertain. Although the timing and style of dextral motion varies widely, in several areas it is associated with later phases of deformation

(Hudleston et al., 1988; Devaney and Williams, 1989; Bauer et al., 1992). In the western Quetico subprovince, dextral motion has been attributed to dextral transpression following oblique arc collision (Hudleston et al., 1988; Borradaile and Spark, 1991; Bauer et al., 1992).

Evidence of thrusting prior to dextral transcurrent motion has been observed in several areas of the Wawa-Quetico-Wabigoon region. In the Beardmore-Geraldton belt, stratigraphic packages are separated by dip-slip faults attributed to thrust stacking in an accretionary wedge (Devaney and Williams, 1989). Early (local D₁) thrusting and nappe formation in the Vermilion district of the western Wawa subprovince may have been associated with accretion (Jirsa et al., 1992). Along the Wawa-Quetico boundary near Schreiber, Williams (1989) identified structures possibly related to early dip-slip thrusting. Early (D₂ and earlier) thrust structures in the Michipicoten greenstone belt were described by Arias and Helmstaedt (1990) and McGill (1992). Detailed mapping in the Manitouwadge greenstone belt has also led to the recognition of ductile shear zones, interpreted as the loci of early thrusting (Zaleski and Peterson, 1993a).

Williams et al. (1992) interpreted the Manitouwadge, Moshkinabi and Faries Lake greenstone belts (Fig. 2) as a semicontinuous easterly trending supracrustal suite (Manitouwadge-Hornpayne assemblage), belonging to a discontinuous mafic metavolcanic layer along the northern margin of the Wawa subprovince, traceable to the east to the Lepage fault zone (Williams et al., 1991) (Fig. 1). In the area of the Manitouwadge, Moshkinabi and Faries Lake belts and the Quetico subprovince to the north, Williams and Breaks (1989; 1990a) described a 5-phase sequence of deformation in which D₂ was the main fabric-forming event. The easterly regional trend of the greenstone belts is modified by D₃ northeasterly trending Z-shaped folds, from west to east; Blackman Lake antiform, Manitouwadge synform, Banana Lake antiform and Moshkinabi Lake synform (Williams and Breaks, 1990b). The axial traces of the Z-folds curve, becoming parallel with the east-west regional trends of the Quetico subprovince to the north (Williams et al., 1992).

Metamorphism

There is a regional metamorphic gradient from amphibolite facies in the Schreiber-Hemlo greenstone belt (Corfu and Muir, 1989b), through upper amphibolite at Manitouwadge, to upper amphibolite and granulite facies in the Quetico subprovince immediately north of Manitouwadge (Williams and Breaks, 1989) (Fig. 2). Thermobarometric results are consistent with upper amphibolite facies (600–700°C, 3–6 kb) in the Manitouwadge belt (Petersen, 1984; Pan and Fleet, 1992), and granulite facies (680–770°C, 4–6 kb) in the Quetico subprovince to the north (Pan et al., 1994). In the Manitouwadge belt, metamorphic grade increases to the north, with muscovite-sillimanite-quartz schists near the Geco mine on the southern limb of the Manitouwadge synform, giving way to sillimanite-microcline-quartz schists (Fig. 3). Migmatites are ubiquitous north of the Blackman Lake antiform. Metagreywackes along the southern limb of the Manitouwadge synform contain garnet-sillimanite-biotite in pelitic layers, but are not migmatitic. Similar rocks near the Quetico boundary contain tonalitic segregations, in some cases with cordierite and garnet. In the Manitouwadge belt, dominant D₂ fabrics are defined by high grade minerals and deformed by D₃ folds; hence, peak metamorphism was broadly synchronous with D₂. Petrographic evidence of decompression reactions (e.g. cordierite mantling sillimanite and orthoamphibole) suggests that cooling was protracted and that high temperatures persisted during D₃ deformation.

Along most of its length, the Quetico subprovince is symmetrically zoned from low grade margins adjacent to the Wawa and Wabigoon subprovinces, to central high grade migmatites (Percival, 1989). The Manitouwadge area is unusual in two respects. Firstly, the Manitouwadge belt and the Wawa-Quetico boundary preserve upper amphibolite-facies assemblages, and secondly, amphibolite-facies assemblages are transitional to orthopyroxene-bearing granulites, the latter found in a lensoid area (about 80×10 km) near the southern margin of the Quetico subprovince (Williams and Breaks, 1990a, b; Pan et al., 1994) (Fig. 2). Moreover, sampling during this study shows that orthopyroxene-bearing rocks in the Quetico subprovince extend somewhat south of the Williams and Breaks (1990b) isograd, to within 1.5 km of the subprovince boundary.

Geochronological constraints

In the Manitouwadge belt, felsic metavolcanic rocks have primary volcanic ages of circa 2720 Ma (Zaleski et al., 1994; Davis et al., 1994), as does a subvolcanic trondhjemite north of the Willroy and Geco mines (Zaleski et al., 1994). In the Schreiber-Hemlo belt, felsic volcanism ranges in age from 2772±2 Ma (Hemlo-Black River assemblage, Williams et al., 1991) near the Hemlo gold camp to 2695±2 Ma for the Heron Bay volcanic complex (Corfu and Muir, 1989a). Felsic metavolcanic rocks (Schreiber assemblage, Williams et al., 1991) associated with the Winston Lake Zn-Cu mine are 2723±2 Ma (Schandl et al., 1991), within error of those at Manitouwadge. The ages suggest that 'Manitouwadge-Hornpayne' assemblage is related to the Schreiber assemblage, rather than the Hemlo-Black River assemblage as originally proposed by Williams et al. (1991). In the western Abitibi subprovince, east of the Kapuskasing structural zone (Fig. 1), felsic rocks associated with base-metal mineralization at Timmins, Abitibi greenstone belt, have been bracketed at 2710–2717 Ma (Bleeker and Parrish, pers. comm., 1994). Felsic tuff underlying the Shunsby base-metal occurrence in the Swayze greenstone belt is 2724±2 Ma (Heather et al., 1995). Evidently, circa 2710–22 Ma

felsic volcanism in the Wawa and western Abitibi subprovinces tended to be associated with hydrothermal activity and mineralization.

The U-Pb systematics of metavolcanic rocks from the Manitouwadge belt do not suggest any inheritance of older crustal components. A model age of 3000 Ma based on Pb isotopic analyses of galena from the Geco mine (Tilton and Steiger, 1969) reflects the inadequacies of early models of Pb isotopic evolution; recalculation gives model ages near or slightly younger (i.e. by circa 5–15 Ma) than the 2720 Ma age of the volcanic host rocks (R. Thorpe, written comm., 1994). The slight discrepancy could be the result of limited isotopic resetting during regional metamorphism (*ibid.*), an interpretation supported by the presence of Pb-bearing amazonite in pegmatites in contact with massive sulphide (Stevenson, 1985).

U-Pb zircon geochronology of various phases of the Black Pic batholith continues under the auspices of the Manitouwadge project. Two dioritic phases of the Black Pic batholith have ages of $2687 \pm 3/-2$ and 2677 ± 2 Ma (see Geochronology). Two pre- to syn-D₂ bodies of K-feldspar porphyry, the Loken Lake and Nama Creek plutons, near the margins of the Manitouwadge belt (Zaleski et al., 1995), have zircons that yield intrusive ages of $2687 \pm 2/-3$ Ma and 2680 ± 3 Ma, respectively. Titanites from the Nama Creek pluton and from the 2677 Ma Black Pic diorite have U-Pb ages of 2672 ± 3 and 2674 ± 2 Ma, respectively, tentatively interpreted as metamorphic cooling ages. In comparison, south of the Schreiber-Hemlo belt, marginal rocks of the Pukaskwa batholith were intruded at $2719 \pm 6/-4$ Ma (Corfu and Muir, 1989a), broadly coeval with felsic volcanism at Winston Lake and Manitouwadge (Fig. 2). Near Hemlo, several granodiorite plutons cluster closely around 2687–2688 Ma, whereas amphibolite-facies metamorphism dated by titanite peaked around 2676–2678 Ma, coeval with the late tectonic Gowan Lake pluton at 2678 ± 2 Ma (Corfu and Muir, 1989b). Synmetamorphic monazites have ages of 2675 ± 1 Ma at Geco and 2677 ± 1 Ma at Winston Lake (Schandl et al., 1991); younger monazite (2661 ± 1 Ma) in biotite schist at Geco was interpreted to date local K-metasomatism (Davis et al., 1994). In general, metamorphic titanite and monazite ages suggest that metamorphism was nearly synchronous, or diachronous over about 5 Ma, throughout this part of the Wawa subprovince. In the Quetico subprovince, Percival (1989) suggested that metamorphism was approximately coeval with granitic plutonism from 2670–2650 Ma.

Metagreywackes of the Quetico subprovince have been interpreted as an accretionary complex continuous with the Wawa subprovince at least since 2689–2684 Ma (Percival and Williams, 1989). Geochronological studies of Quetico metasedimentary rocks have been mainly confined to west and north of Thunder Bay (Fig. 1). Detrital zircons constrain the maximum age of sedimentation to 2702 ± 4 Ma for the southern Quetico near Thunder Bay (Percival and Sullivan, 1988), and to 2698 ± 3 Ma for the northern Quetico near Atikokan (Davis et al., 1990). Minimum depositional ages were constrained by intrusions of $2687 \pm 19/-13$ Ma and 2688 ± 4 Ma, respectively for each area. In both cases, zircons older than 3000 Ma indicated input from old sources (*ibid.*). In the Atikokan area, the zircons could have relatively proximal sources to the north, including the 3003 ± 5 Ma Marmion Lake batholith (Davis et al., 1990). In the case of zircons greater than 2800 Ma in the southern Quetico, no local potential sources are known (Percival and Sullivan, 1988). The Coutchiching metagreywackes in the Rainy Lake area of the Wabigoon subprovince contain detrital zircon populations similar to those of Quetico rocks (Davis et al., 1989). The maximum depositional age was established at 2704 ± 3 Ma and the minimum age at 2692 ± 2 Ma, the age of a cross-cutting intrusion. The presence of zircons as old as 3059 Ma indicated a Mesoproterozoic source.

In both the Wabigoon and Wawa subprovinces, conglomeratic sequences are comparable to Timiskaming rocks in facies associations and age. These sequences include (Fig. 1); the Max Creek conglomerate with a maximum depositional age of 2687 ± 3 Ma (Percival and Sullivan, 1988), the Seine group near Rainy Lake bracketed between $2696 \pm 5/-3$ and $2686 \pm 2/-1$ Ma (Davis et al., 1989), and the Shebandowan group (Borradaile and Brown, 1987) with a maximum depositional age of 2704 ± 2 Ma and in depositional contact with alkalic volcanic rocks of $2689 \pm 3/-2$ Ma (Corfu and Stott, 1986).

HISTORICAL BACKGROUND OF THE MANITOUWADGE BELT

History of deposits

In 1931, J.E. Thomson, Ontario Department of Mines, while engaged in mapping the north shore of Lake Superior, was told by the local Ojibway Indians of rusty rocks near Manitouwadge Lake to the north (Brown, 1963). Thomson and his guide, M. Fisher, completed a canoe traverse to the area and found sulphide mineralization and 'considerable magnetic disturbance' (Thomson, 1932) on the future sites of the Geco and Willroy mines (Fig. 3). Fisher returned in 1943 and staked the showing, but he was unable to find financial backing during the war years and allowed his claim to lapse (Pye, 1957). Ten years later, interest in base metal potential was heightened and the showings on Thomson's map were staked by 'weekend prospectors', Dawd, Barker and Forster (*ibid.*). Within the year, drilling intersected an orebody and Geco Mines Limited was incorporated. In the ensuing staking rush, the Willroy, Nama Creek and Willecho deposits were discovered, all of which were eventually mined. The Willroy and Geco mines went into production in 1957.

The Geco mine, scheduled to close in late 1995, is the only mine currently active in the area. Geco was by far the largest deposit; total production for its lifetime is projected to be 55.9 million tonnes with an overall grade of 1.85% Cu, 3.76% Zn and 46.9 gram/tonne Ag. Minor amounts of Au, Cd and Bi are recovered and, until 1988, an average of 0.30% Pb was extracted (Williams et al., 1990) as a by-product of purification of Cu concentrate. Combined production from the Willroy, Nama Creek and Willecho deposits, until the closure of the Willroy mine in 1977, was 8.0 million tonnes with a grade of 0.9% Cu, 4.9% Zn and 50 gram/tonne Ag. Exploration continues in the area, mainly focussed on identifying targets at depth and along the folded extensions of the belt.

Previous mapping and interpretations

The first systematic mapping of the Manitouwadge greenstone belt was undertaken by Pye (1957) for the Ontario Department of Mines, from 1954 to 1957. He outlined the general shape of the Manitouwadge synform and recognized mafic metavolcanic rocks along the inner and outer margins of the folded supracrustal sequence. At the time of Pye's mapping, most of the presently known economic deposits had already been discovered and, in many cases, his detailed maps of mine properties are the only published record of geological relationships that were obliterated by mining. Pye regarded all the mineralization as Algoman-age replacement that post-dated intrusive activity, with the possible exception of diabase dykes. He interpreted most supracrustal rocks as metasediments. From 1968 to 1970, Milne (1974) remapped the area of major economic interest for the Ontario Division of Mines, reinterpreting most of the felsic rocks as volcanic or volcanogenic.

The base-metal deposits of the Manitouwadge greenstone belt are a classic example of the problems encountered in petrogenetic studies of mineralization in complex polydeformed terranes. Early studies tended to focus on evidence of local replacement, remobilization and structural control (Pye, 1957; Timms and Marshall, 1959; Brown et al., 1960; Watson, 1970). Associated orthoamphibole-cordierite-garnet gneiss has been interpreted as metasediment (Pye, 1957; Brown et al., 1960; Watson, 1970), restite from the extraction of partial melt (Robinson, 1979), and metamorphosed altered volcanic or volcanoclastic deposits with possible pelitic interbeds (James et al., 1978; Friesen et al., 1982). Sillimanite-muscovite-quartz schist in proximity to orebodies has also been variously interpreted as metasediment, with the most muscovitic zone corresponding to a concordant fault (Milne, 1974); epigenetically altered metasediment (Pye, 1957; Brown et al., 1960); metamorphosed altered volcanic and volcanoclastic rock (Friesen et al., 1982; Bakker et al., 1985); and syn-metamorphically altered rock, possibly with a prehistory of synvolcanic alteration (Schandl et al., 1991; Davis et al., 1994). Suffel et al. (1971) were the first to propose a volcanogenic origin, interpreting textures and microstructures the result of superimposed metamorphism and remobilization. More recently, a syngenetic origin for the base-metal deposits and altered rocks has been increasingly accepted (James et al., 1978; Friesen et al., 1982; Bakker et al., 1985).

Historical nomenclature, the 'Mine Series'

Most previous detailed mapping has been done by private-sector geologists who have adopted the historical 'Mine Series' lithological units (Table 1). This confusing nomenclature is of wide use locally and has found its way into the literature. The 'Mine Series' consists of a four-fold subdivision of supracrustal rocks (Friesen et al., 1982; Williams et al., 1990) into; Granite Gneiss group, Sericite Schist group, Grey Gneiss group and Hornblende Schist group. The 'Granite Gneiss group' refers to orthoamphibole-cordierite-garnet gneiss, interpreted as metamorphosed synvolcanically altered rocks lying in the stratigraphic footwall (structural hanging wall) to the mineral deposits (*ibid.*). The misleading name dates back to the discovery drill holes of the Geco mine; by a quirk of nature and karma, the holes penetrated foliated granite before emerging into massive sulphide, and hence, all the rocks of the structural hanging wall were grouped as 'granite gneiss'. 'Sericite schist' was originally Pye's field term for muscovite-quartz±sillimanite schist, and (although corrected by Pye (1957) on the basis of petrography) the misnomer persists in the 'Mine Series'. The 'Sericite Schist group' has been viewed as the lateral equivalent of the 'lower Grey Gneiss group', representing either a sedimentary facies (Milne, 1974) or an alteration product (Friesen et al., 1982; Bakker et al., 1985). The 'Grey Gneiss group' corresponds to Pye's metasedimentary unit, later subdivided into lower (northerly) and upper (southerly) members. The 'lower Grey Gneiss group' is approximately equivalent to felsic to intermediate metavolcanic rocks, interlayered with iron formation, extending from Geco to north of the Willecho area (Fig. 3). The 'upper Grey Gneiss group' mainly includes metagreywackes and foliated tonalite occupying the centre of the southern limb of the Manitouwadge synform. The contact between the 'lower' and 'upper Grey Gneiss' has been considered as transitional, reflecting a change from a depositional regime dominated by volcanism and chemical sedimentation (iron formation and sulphide mineralization), to one dominated by clastic sedimentation (Friesen et al., 1982; Bakker et al., 1985). Our U-Pb dating of detrital zircons shows that the 'upper Grey Gneiss group' is at least 25 Ma younger than the 'lower Grey Gneiss group', and that the former is likely correlative with migmatized metagreywackes of the Quetico subprovince (Zaleski et al., 1995; see Geochronology). The 'Hornblende Schist group' corresponds to mafic metavolcanic rocks, now recognized to lie at the stratigraphic base of the entire sequence.

TABLE 1. Comparison of Map Units and Historical 'Mine Series'

Mine series	Map units, this work	depositional age
Upper Grey Gneiss	Metasedimentary rocks (and tonalite) (Unit 10)	<2693 Ma ¹
	 unconformity or fault	
Lower Grey Gneiss	Felsic metavolcanic rocks (and iron formation), quartzitic variant of sillimanite-muscovite-quartz schist (Units 1, 6, 7, 8, 9)	2720 Ma ¹
Sericite Schist	Sillimanite-muscovite-quartz schist, altered felsic volcanic rocks (Unit 1)	2720 Ma ²
Granite Gneiss	Orthoamphibole-cordierite-garnet gneiss, altered mixed mafic-felsic volcanic rocks (Unit 2)	2720 Ma
Hornblende Schist	Mafic metavolcanic rocks (Unit 3)	≥2720 Ma

¹Zaleski et al. (1995)²Davis et al. (1994)

Contradictions exist in the literature with respect to the stratigraphic succession and younging direction of the 'Mine Series'. The confusion is related to the presence of an early fold, originally proposed by Suffel et al. (1971) and Touborg (1973), that repeats the supracrustal succession across the southern limb of the Manitouwadge synform. Pye (1957) assumed northerly younging for the entire belt. Reinterpretation of the massive sulphide deposits as volcanogenic implied a southerly younging sequence for the Willroy-Geco area (Suffel et al., 1971) and, in some cases, southerly younging was extrapolated to the entire belt (Friesen et al., 1982).

DESCRIPTION OF UNITS

The Manitouwadge greenstone belt comprises a mafic-to-felsic succession, the felsic component of which is circa 2720 Ma. The map pattern is dominated by D₃ folds, and the greatest thickness of supracrustal rocks lies in the hinge region and along the southern limb of the D₃ Manitouwadge synform (Fig. 3). Away from the hinge region, supracrustal units are attenuated and invaded by plutonic rocks. On the southern limb, the volcanic succession is repeated to north and south by a D₂ syncline with a central zone of metagreywacke (see Structure). The northern sequence is here referred to as the 'inner' volcanic belt, and the southern as the 'outer' volcanic belt, with respect to the inner and outer regions of the Manitouwadge synform. All known Cu-Zn deposits, including the Geco, Willroy, Nama Creek and Willecho deposits, lie in the inner volcanic belt along the southern limb and inner hinge region of the Manitouwadge synform.

Our definition of map units partly reflects differences in scale of mapping, exposure and accessibility; for example, mafic, mixed mafic-felsic, and felsic to intermediate metavolcanic rocks (Units 3, 4 and 6) in the inner volcanic belt give way to undifferentiated intermediate to mafic metavolcanic rocks to the north (Unit 5). All supracrustal rocks are metamorphosed, although the prefix 'meta' is sometimes omitted depending on the context. 'Alteration' is used to mean metasomatic changes in whole-rock composition, inferred from metamorphic mineral assemblages and abundances. Unless stated, timing is not implied by 'alteration' and it could be synvolcanic (premetamorphic), synmetamorphic or post-peak metamorphic. The unit numbers, locations, geographic names and references to aeromagnetic trends in the following discussion refer to the accompanying 1:25000 map, generalized in figure 3. Stops and figures prefixed by a letter refer to field stops and maps found in the field-trip section of this volume.

Supracrustal rock units

Mafic metavolcanic rocks (Unit 3) occur in both the inner and outer volcanic belts. The outer belt is dominated by mafic rocks, comprising a steeply dipping section of up to 3 km of interleaved hornblende-rich mafic schists, laminated rocks and foliated gabbros. These are folded around the hinge of the Manitouwadge synform near Mills Lake, and continue to the northeast in a zone of high strain along the northwestern margin of the Manitouwadge belt until truncated(?) by the Nama Creek pluton (Unit 13a). To the east on the southern limb of the Manitouwadge synform, the unit thins and is increasingly engulfed by plutonic rocks.

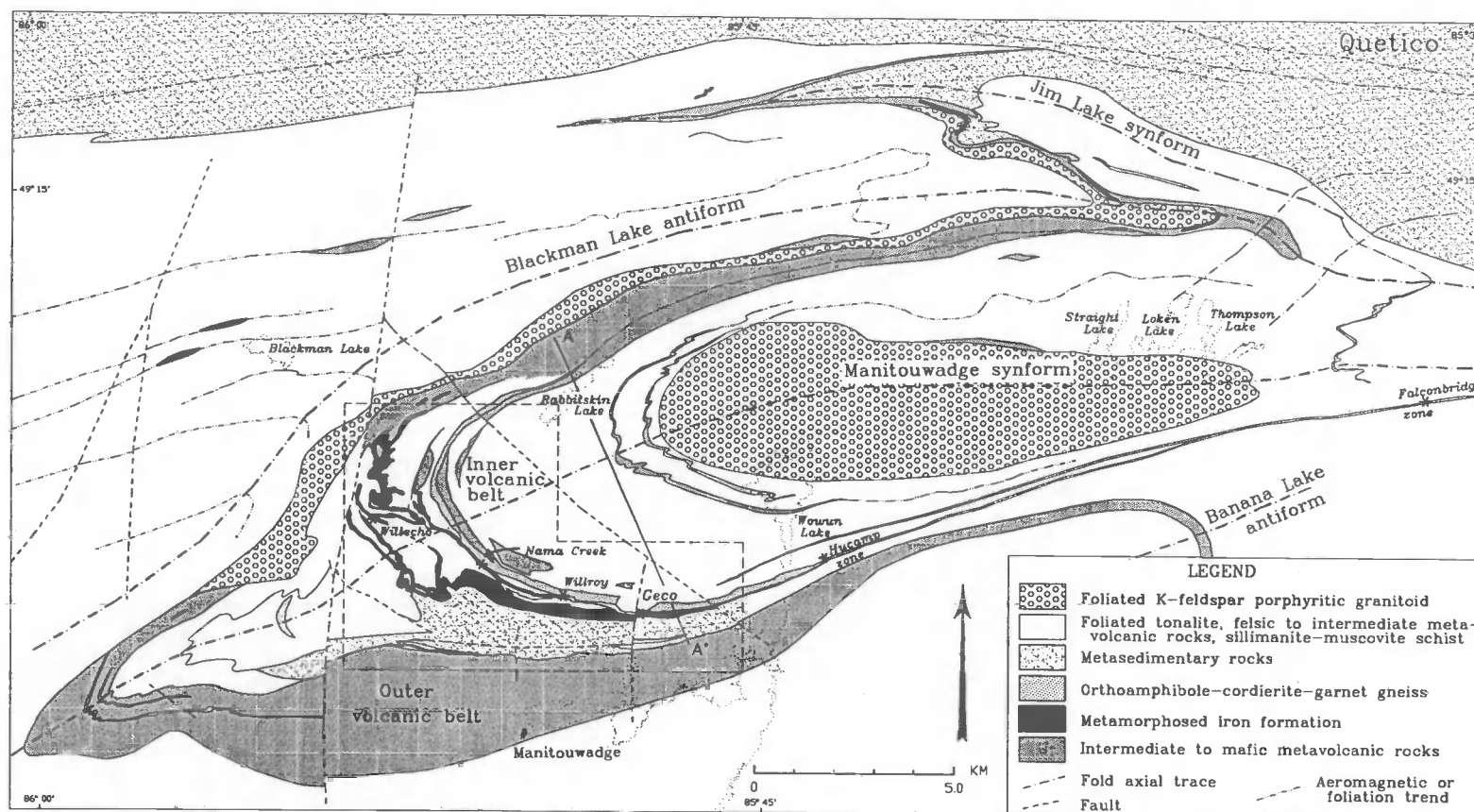


FIG. 3. Generalized geology of the Manitouwadge greenstone belt, omitting diabase dykes (see the accompanying 1:25000 map for details). The regional structure is dominated by major D_3 folds; the Manitouwadge synform, Blackman Lake antiform and Jim Lake synform. The inner and outer volcanic belts are correlative, repeated by a D_2 syncline, the axial trace of which lies in metasedimentary rocks on the southern limb of the Manitouwadge synform (Fig. 5). Asterisks mark economic and sub-economic base-metal deposits. The area detailed in Figure 4 is outlined. A-A' is the trace of the projection in Figure 7.

The hinge region of the Banana Lake antiform is poorly exposed, but foliation trends in the enclosing tonalite suggest a possible connection to the Faries Lake belt (Fig. 2). Aeromagnetic trends are more consistent with an easterly continuation of the mafic unit, although weak trends obscured by diabase dykes may be consistent with folding by the Banana Lake antiform. In the inner volcanic belt, mafic inclusions in synvolcanic trondjemite (Unit 12), in some cases, form near-continuous screens traceable in outcrop for distances of 3.5 km (from the Mose Lake fault to the road east of Wowun Lake). Mafic rocks of Unit 3 are transitional to the south to interleaved mafic-felsic metavolcanic rocks (Unit 4).

Mafic metavolcanic rocks are strongly foliated, fine grained, homogeneous to thinly layered schists, and medium to coarse grained metagabbro, typically interlayered. In the area of detailed mapping between Mills and Swill Lakes, layers of schist and metagabbro are 50–100 metres thick (Fig. B3). Near Gaug Lake, metagabbro forms a larger body in fine grained schist. The metagabbro is commonly a hornblende-augen schist in which lenticular augen (0.5–1 cm in size), locally with well developed asymmetric tails, lie in a fine grained schistose matrix. In some high strain zones, metagabbro or augen schist is transitional to fine grained homogeneous schist, suggesting that some homogeneous schist could have originated as gabbro, comminuted by deformation (Stop B15).

In both the inner and outer volcanic belts, mafic schists are locally pillowed but, with few exceptions, deformation makes younging or structural facing determinations suspect. In an exposure between Swill and Mills Lakes, pillow shapes suggest southerly younging (Stop B14), and northerly younging has been reported west of Gaug Lake (W. Bates, Granges Inc., pers. comm., 1992). In view of the polyphase folding in the Manitouwadge belt, these few younging determinations contribute little to regional interpretation.

The typical mineral assemblage in mafic schists consists of hornblende-plagioclase $\pm$ garnet $\pm$ biotite $\pm$ clinopyroxene with quartz, magnetite, titanite, sulphide minerals and apatite in accessory or trace amounts. Laminations are defined by plagioclase-rich lamellae. Epidote occurs in calc-silicate lenses or boudins. Garnetiferous zones (up to 40% garnet), in some cases with quartz, magnetite and cummingtonite, are particularly common in mafic schists in the outer volcanic belt. Near the northern contact to metagreywacke, mafic rocks of the outer belt are involved in a complex zone of interleaved felsic schist, foliated tonalite, garnetiferous hornblende-biotite $\pm$ magnetite gneiss, iron formation and orthoamphibole-plagioclase $\pm$ garnet $\pm$ magnetite $\pm$ hornblende gneiss (Stops B1 and B20).

The intensity of deformation generally makes it difficult to unambiguously identify the physical protolith to mafic rocks. Metagabbros interlayered with schists may be massive flows, massive bases of flows, or sills. Some laminated mafic schists may have had a tuffaceous or volcanoclastic origin, but highly strained pillowed flows could also be present. Orthoamphibole-bearing rocks near the northern contact of the outer belt are interpreted as equivalent to the much more extensive orthoamphibole-cordierite-garnet gneiss of the inner belt, and hence as metamorphosed synvolcanic alteration (see Unit 2). However, at least some of the garnet $\pm$ cummingtonite zones of the outer belt are apparently discordant to tectonic fabrics, and hence, of later origin (Stop B20).

Mixed mafic-felsic metavolcanic rocks (Unit 4) of the inner volcanic belt lie mostly along the transitional contact between mafic rocks (Unit 3) to the north, and felsic (Units 6 and 8) and orthoamphibole-bearing rocks (Unit 2) to the south. Mixed mafic-felsic rocks also form inclusions and screens in synvolcanic trondjemite (Unit 12) north and east of orthoamphibole-bearing rocks. Layering of mafic and felsic components (cm–1 m width) might suggest contemporaneous mafic and felsic volcanism, or tectonic interleaving. Mafic layers are more abundant than felsic, and consist of dark, fine grained, strongly foliated, hornblende-plagioclase schist, with or without magnetite, quartz, cummingtonite, biotite and garnet and generally similar to mafic schist of Unit 3. Layers rich in biotite, cummingtonite and garnet are more common near the southern contact. Intercalations of quartz-rich felsic and leucofelsic schist contain plagioclase, biotite and, locally, quartz phenocrysts, magnetite and garnet. West of Fox Creek, outcrops of felsic schist with elongate hornblende-richer patches or 'clasts' are interpreted as an intrusion breccia of mafic volcanic inclusions in a trondjemite matrix.

Two areas of mafic rocks north of Garnet Lake and west of orthoamphibole-bearing Unit 2 were also grouped with mixed mafic-felsic rocks. These are somewhat atypical of the unit in that the rocks are characterized by abundant calc-silicate minerals including Ca-amphibole, epidote, clinopyroxene, titanite and plagioclase, as well as quartz. Microcline, garnet and calcite are present in minor amounts. Diffuse semicontinuous layers or lenses (typically 0.5–30 cm in width) are defined by variations in the abundance and grain size of calc-silicate and felsic minerals. In some cases, quartz eyes resemble relict phenocrysts. The origin of these rocks and their relationship to more typical Unit 4 is uncertain; but it may be that they are the result of calc-silicate alteration of felsic volcanic rocks, possibly with interleaved mafic rocks.

Intermediate to mafic metavolcanic rocks (Unit 5) comprise a heterogeneous group of undifferentiated rocks, dominantly hornblende-plagioclase, hornblende-plagioclase-biotite and hornblende-plagioclase-garnet $\pm$ clinopyroxene $\pm$ magnetite schists, and foliated diorites. The unit defines the attenuated northern

limb of the Manitouwadge synform and related map-scale folds to the north (Blackman Lake antiform, Jim Lake synform). Between Rabbitskin Lake and the railway west of Fox Lake, exposure is poor and inaccessible; however, aeromagnetic trends suggest that the unit is continuous east to One Otter Lake. West of Rabbitskin Lake, fine to medium grained, mafic to intermediate schist and minor felsic schist are homogeneous, laminated or layered, and host many intrusions of foliated pegmatite, granite and tonalite. Garnetiferous zones and patches, in some cases with 30–40% garnet porphyroblasts and lesser amounts (<10%) of magnetite, are common. South of Fox Lake, laminated to layered (1 mm–10 cm), fine grained, mafic to intermediate metavolcanic rocks are present both north and south of the Nama Creek pluton, and as inclusions. The mineral assemblage includes hornblende-plagioclase $\pm$ clinopyroxene $\pm$ quartz $\pm$ biotite, with epidote locally in knots, and accessory titanite, magnetite and apatite. Southeast of Fox Lake, exposures of heterogeneous interleaved (1 cm–1 m) foliated pegmatitic granite, biotite and hornblende-biotite felsic schist locally grade to garnetiferous strongly magnetic gneiss. Garnet porphyroblasts comprise up to 50% of the gneiss, occurring in lenses (2–3 cm wide) between anastomosing mafic layers of fine grained hornblende, clinopyroxene and magnetite. In some cases, quartz eyes and aggregates are abundant. The magnetic gneisses undoubtedly contribute to the high aeromagnetic signature of the area.

In the vicinity of the Blackman Lake and Jim Lake folds, the Manitouwadge belt comprises thinly layered, fine to medium grained, mafic to intermediate metavolcanic rocks, as well as foliated diorite and gabbro that may be recrystallized homogeneous volcanic or subvolcanic rocks. The rocks are commonly strongly magnetic and, in addition, are interleaved with minor iron formation (Stop E6). The Jim Lake synform is defined by mappable zones of mafic to intermediate screens and inclusion swarms in foliated tonalite and pegmatite. Locally, rotated mafic to intermediate blocks (with discordant foliation trends) in an intrusive matrix resemble agmatite or intrusion breccia.

At least two nearly continuous zones of mafic rocks can be mapped inside the Manitouwadge synform, corresponding to zones of high variable aeromagnetic relief. These rocks, here called the Dead Lake suite, comprise a complex association of interleaved foliated gabbro, diorite, and layered mafic to intermediate rocks of probable supracrustal origin. The Dead Lake suite includes a distinctive group of strongly magnetic rocks characterized by hornblende-magnetite $\pm$ plagioclase $\pm$ garnet $\pm$ clinopyroxene $\pm$ sulphide minerals, with variable amounts of quartz (minor to 50%) in quartz eyes and irregular lenticles (Stops C3–C5). Titanite and apatite are present in minor amounts (1–3%), and allanite/epidote, zircon and monazite in trace amounts. The hornblende-magnetite $\pm$ garnet rocks are usually fine to medium grained and homogeneous, at least at hand-sample scale. The more leucocratic varieties resemble synvolcanic trondhjemite (Unit 12), but with disseminated hornblende, garnet and/or magnetite. However, in many cases, thin layering (1–5 cm) and grading, interpreted as modified bedding, is defined by variations in grain size and mineral abundances (Stop C4). Layered rocks are associated with magnetite-bearing quartzites (metachert?) or with mafic metavolcanic rocks with garnet-cummingtonite(?) concentrated in interconnected ganglions. Field observations suggested an origin as metamorphosed and digested (by intrusion) ferruginous chert (G. Stott, Ontario Geological Survey, pers. comm., 1994) and iron formation. These proposals are supported by an exposure on the western shore of Wowun Lake, in which foliated granite contains inclusions of mafic rock and weakly magnetic quartz-rich layered iron formation, the latter dispersed over a width of 2 to 3 metres. However, the high TiO₂ and Zr geochemistry of magnetite quartzite, hornblende-magnetite $\pm$ garnet rocks and garnet-hornblende-bearing trondhjemite are more consistent with sedimentary concentration of heavy minerals, rather than iron formation (see Geochemistry).

The relationship of the Dead Lake suite to the main supracrustal assemblage is problematic. The two zones of the suite are indistinguishable and may represent a structural repetition. They are similar to mafic rocks elsewhere in the Manitouwadge belt, especially garnetiferous mafic rocks, but hornblende-magnetite $\pm$ garnet rocks and associated layered magnetite quartzite are unusual. Aeromagnetic trends corresponding to the Dead Lake suite continue easterly both north and south of the central Loken Lake pluton (Unit 13b). On the southern side, the aeromagnetic trend tends to converge with the main southern limb of the Manitouwadge synform. Rocks similar to the garnet-hornblende 'trondhjemite' occur east of Banana Lake in association with mafic and orthoamphibole-garnet-cordierite screens in plutonic rocks (Stop D4). North of the Loken Lake pluton, a pronounced aeromagnetic anomaly extends north of Straight, Loken and Thompson Lakes, and apparently continues around the hinge of the Blackman Lake antiform. The same distinctive rock type (grouped with the Dead Lake suite of Unit 5) is exposed east of Larry Lake, enclosed by trondhjemite to the north of the main supracrustal belt. Along the northern contact of the Loken Lake pluton, a zone of mafic rocks associated with subeconomic Zn mineralization (Noranda's Straight Lake zone) may also belong to the Dead Lake suite.

Felsic to intermediate metavolcanic rocks (Unit 6) are among the most unsatisfying and frustrating of units to map, consisting of subtly heterogeneous and transitional lithologies including: fragmental rocks; calc-silicate, sillimanite-muscovite, and felsic-intermediate schists; straight gneiss (see Unit 11); and foliated

tonalite, but generally without marker units to allow useful subdivision. The unit contains lithologies typical of Units 1, 7, 8, 11 and 14, as well as minor exposures typical of Units 2 and 3, but not on a mappable scale.

In the inner hinge region of the Manitouwadge synform, some of the complexity is probably due to cryptic early faults and folds (see Structural Geology). Homogeneous to weakly layered, quartz-phyric and aphyric felsic schists with minor biotite±muscovite are predominant. Microcline is heterogeneously distributed, in some cases absent entirely, or in others, confined to millimetre-wide microcline-rich layers. In the vicinity of the two iron formation layers south of the Willecho deposit, felsic rocks sporadically contain large zoned sillimanite knots identical to those of Unit 1b (sillimanite-knot felsic schist) described below. Fragmental-looking rocks have lenticular aphyric felsic enclaves in an intermediate matrix containing biotite±hornblende±garnet±magnetite. The fragmental appearance may be partly an artifact of tectonism or alteration. Where the matrix is more felsic in composition (less altered?), felsic enclaves are more defined and resemble clasts in an volcanoclastic or epiclastic breccia. The outer hinge region from Nama Creek to Swill Lake, is underlain by felsic to intermediate metavolcanic rocks and biotite schist (mostly metasedimentary?) invaded by foliated tonalite.

Aphyric felsic metavolcanic rocks (Unit 7) are present in the outer volcanic belt in three main areas; interlayered with mafic rocks (Unit 3) in the Swill-Mills Lakes area, north of Gaug Lake, and north of Manitouwadge Lake near the Fox Creek fault. In the Swill-Mills Lakes area, two semicontinuous felsic units form useful markers outlining map-scale symmetrical and asymmetrical folds near the hinge region of the Manitouwadge synform (Fig. B3). The southernmost unit, up to 50 metres thick, can be recognized locally as a highly deformed monolithologic felsic breccia, consisting of fine grained to aphanitic lapilli-size felsic clasts in a matrix of biotite-muscovite or garnet-hornblende schist (Stops B12 and B13). The breccia is associated with minor iron formation and, locally, disseminated pyrite and pyrrhotite (Stop B14).

North of Gaug Lake, aphyric felsic rocks and interleaved iron formation (Unit 9) lie along or near the contact between mafic metavolcanic rocks (Unit 3) and metagreywacke (Unit 10). The felsic rocks include homogeneous schist, and monolithologic and heterolithic volcanoclastic breccias. Monolithologic clasts mostly vary from 2 to 10 centimetres in size, but some breccias have large (cm-m scale) angular clasts (Stop B4). Some large clasts have reentrant angles, and some nearest neighbours have complementary shapes suggestive of *in situ* fragmentation of larger blocks. The matrix is felsic to intermediate, normally containing biotite, but in some cases, dominated by hornblende and garnet porphyroblasts. The breccias were interpreted as proximal volcanic deposits, possibly derived from phreatic (W. Bates, Granges Inc., pers. comm., 1992) or hydrothermal explosions involving limited transportation. Heterolithic breccias contain felsic and hornblende-garnet-rich clasts (generally 2 to 10 centimetres in size), typically in a biotite-hornblende-garnet-bearing matrix. The hornblende-garnet clasts resemble matrix material of monolithologic breccias, suggesting that reworked proximal deposits were a source for heterolithic breccias (Stop B5).

North of Manitouwadge Lake and south of the Agam Lake fault, exposures of white laminated quartz-rich felsic schists are interpreted as volcanic or intrusive rocks. To the north, they are bounded by a topographic depression marking the trace of the Agam Lake fault. On the north side of the fault, the biotite schist is typical of metasedimentary rocks of Unit 10. The felsic rocks have a strong fabric, possibly related to the fault, defined by partially annealed quartz ribbons, minor biotite and muscovite.

Quartz-phyric felsic metavolcanic rocks (Unit 8), in three bodies in the inner volcanic belt, are spatially associated with the known mineral deposits. Quartz-phyric rocks are interlayered with iron formation (Unit 9), massive sulphide deposits hosted by iron formation, and sillimanite-muscovite-quartz schist (Unit 1); at least some of the interlayering is interpreted to be due to structural repetition by early faults and folds (see Structural Geology). The unit is characterized by abundant quartz phenocrysts (up to 20%, typically 1–3 mm in size), commonly forming lenticular augen, in a fine grained biotite felsic to leucofelsic matrix. The matrix consists of quartz, plagioclase, microcline and biotite, with accessory or trace amounts of muscovite, epidote and garnet. Magnetite porphyroblasts (1–3 mm in size) are prominent. As in felsic rocks of Unit 6, microcline and muscovite are heterogeneously distributed, even on a thin-section scale, suggesting that their presence is related to potassic alteration rather than primary magmatic composition (see Geochemistry). In some cases, quartz-phyric rocks grade to aphyric rocks, especially toward contacts.

The three quartz-phyric bodies include homogeneous schists and fragmental rocks, the latter both monolithologic and heterolithic. In one location, angular monolithologic quartz-phyric felsic clasts, ranging from large (>0.5 m) to small (<1 cm), are unsorted and apparently matrix-supported, in a microcline-biotite-epidote-hornblende matrix (Stop A18). The breccia could be a proximal volcanoclastic deposit, or a flow-top or flow-foot breccia. More typically, volcanoclastic rocks contain lenticular lapilli-size fragments (<1–30 cm long), and both fragments and matrix contain quartz phenocrysts (Stop A3). The clasts show subtle variation in colour and grain size that could be attributed to variable response to alteration (e.g. devitrification, diagenesis, alkali-exchange) of originally comagmatic crystalline and glassy components. In the Willroy-Geco area, the matrix contains muscovite, biotite, garnet and magnetite. North of Willecho, fragmental rocks are

more difficult to identify unambiguously. Diffuse lenticular quartz-phyric felsic enclaves are present in a calc-silicate-rich matrix including plagioclase, Ca-amphibole, epidote and garnet, as well as magnetite, muscovite and microcline.

In all three quartz-phyric bodies, the abundance of calc-silicate minerals tends to increase toward contacts with iron formation. In the Willroy area, outcrops dominated by the calc-silicate minerals; plagioclase, clinopyroxene, Ca-amphibole, garnet, epidote and titanite, can be traced laterally into quartz-phyric felsic rocks (Stop A4). The calc-silicate rocks were interpreted as metasomatically altered.

Metamorphosed iron formation (Unit 9) includes three subunits; a) quartz-magnetite iron formation, b) silicate iron formation, and c) sulphidic iron formation. These occur in both the inner and outer volcanic belts, interlayered mostly with felsic metavolcanic rocks (Units 6, 7 and 8) and sillimanite-bearing schist (Unit 1). Most of the Cu-Zn orebodies are associated with quartz-magnetite iron formation that grades laterally to sulphidic iron formation and to massive sulphide (Pye, 1957; Timms and Marshall, 1959; Friesen et al., 1982).

Quartz-magnetite iron formation, by far the most abundant type, is characterized by alternating white coarse grained (recrystallized) quartz layers and dark magnetite±grunerite±actinolite±garnet±clinopyroxene layers. The relative proportions of iron-rich minerals is variable and even magnetite is present only in trace amounts, or absent, in some cases. Sulphide minerals (pyrrhotite, pyrite, sphalerite, chalcopyrite) and carbonate may be present in minor or trace amounts, as well as stilpnomelane, apparently of retrograde origin. Layering, interpreted as a modification of chemically precipitated bedding, varies from dark lamellae in a quartz-dominated rock to dark and light layers of 10 centimetres or more in width. Iron silicate minerals, especially grunerite, vary from medium to very coarse grained (1–5 cm length), typically with weakly developed tectonic fabrics. However, at least in some localities, grunerite defines a foliation axial planar to D₂ folds.

Layering is commonly disrupted by breccia zones and outcrop-scale folding. A tectonic origin is indicated for minor folds by the observation that systematic changes in asymmetry are related to map-scale D₂ folds. Breccia zones of angular and rotated quartzose fragments in a dark magnetite-silicate matrix are sandwiched between coherent layers (Stop A22), features typical of intraformational breccias. However, the apparently competent behaviour of quartzose fragments suggests that brecciation occurred after diagenesis or low grade metamorphism. Tectonic fragmentation is also suggested by local breccia zones in the hinges of minor folds, and by quartzose fragments that preserve fold hinges. In the case of breccias related to folds, dark iron-rich material encloses quartzose fragments, evidently having migrated into thickened hinge regions.

Silicate iron formation is dark and homogeneous, resembling the dark layers in quartz-magnetite iron formation, except that garnet tends to be coarser and more abundant. Minor amounts of quartz may be present, dispersed among the iron-rich minerals. Silicate iron formation occurs mainly near the contacts of quartz-magnetite iron formation. It may represent an original depositional unit or stratiform alteration related to iron formation contacts. Alternatively, its similarity to iron-rich material in quartz-magnetite iron formation, and evidence of outcrop-scale remobilization, suggest that silicate iron formation may represent large-scale remobilization and segregation of material derived from quartz-magnetite iron formation.

Sulphidic iron formation is generally similar to quartz-magnetite iron formation, but with disseminated or stringer pyrite, pyrrhotite or sphalerite. It rarely forms a mappable unit, occurring mainly in the transition from quartz-magnetite iron formation to massive sulphide (Stop A3).

In the inner volcanic belt along the southern limb of the Manitouwadge synform, iron formation defines four main belts. Southwest of the Nama Creek deposit, the most southerly iron formation is thickened by D₂ folding (see Structural Geology). Silicate iron formation is present along its northern contact and forms a thicker zone on the west side of a sliver of iron formation that extends toward Garnet Lake. The sliver is separated from the main belt of iron formation by an early fault. Some of the thin iron formations in the Willroy area similarly represent structural repetitions (see Structural Geology). In the Geco area and west of Fox Creek, the southernmost iron formation is heavily invaded by concordant sheets of foliated tonalite, mostly not separable at the scale of mapping; hence, the true volume of iron formation is exaggerated. The southern contact between iron formation and metagreywacke is mostly obscured by tonalite sheets.

North of Garnet Lake, iron formations are tectonically thinned, in some cases occurring as boudin trails or pinching out entirely. In the Willecho area and to the north, iron formation defines map-scale folds and is thickened by folding.

Iron formation was also mapped on the limbs and near the hinge of the Blackman Lake antiform. North of Straight Lake, zincian sulphidic quartz-rich rocks (Noranda's Jim Lake zone), associated with orthoamphibole-bearing rocks, were interpreted as iron formation. West of One Otter Lake and between Jim and Davis Lakes, magnetic rocks grouped with iron formation are hornblende-plagioclase-rich with thin layers (mm–cm width) defined by varying proportions of fine grained mafic and felsic minerals and magnetite (Stop E6). They are mainly associated with mafic to intermediate metavolcanic rocks and may represent chemical precipitates with a considerable detrital or tuffaceous component, or altered volcanic rocks. Minor exposures of more typical quartz-magnetite iron formation are locally present in the area, and also in the Dead Lake suite in an exposure

on Wowun Lake. West of the Manitouwadge belt and Kern (Blackman) Lake, two lenses of quartz-magnetite iron formation are present as inclusions in tonalite of the Black Pic batholith in an area of strong aeromagnetic striping.

Metasedimentary rocks (Unit 10) lie between the inner and outer metavolcanic belts along the southern limb of the Manitouwadge synform, central to the D₂ 'Manitouwadge syncline' (see Structural Geology). Mostly comprising monotonous metagreywacke, they are dominated by homogeneous, grey, foliated, fine grained, poorly to strongly layered, biotite-quartz-feldspar±hornblende±muscovite schist. Locally, more pelitic layers contain garnet and/or sillimanite. Layering or modified bedding (0.5–30 cm in width) is defined by variation in mafic mineral abundance, and locally graded. Concordant and semiconcordant intrusions of plagioclase-phyric tonalite are common. East of Agam Lake, metagreywacke contains quartz- and plagioclase-crystal clasts, interpreted as evidence for a tuffaceous volcanogenic component. Chaotic folds, some eye-shaped, spanning a thickness of several layers (totalling 5–20 cm) and sandwiched between undisturbed layers, are interpreted as the result of transposed soft-sediment deformation (Stop A1).

The maximum depositional age of the Manitouwadge metagreywackes is at least 25 Ma younger than the age of felsic volcanism associated with Cu-Zn mineralization (Zaleski et al., 1995). The metagreywackes are involved in map-scale D₂ folds south of the Nama Creek deposit and the dominant foliation is ascribed to D₂ deformation. The contact between metagreywackes and metavolcanic rocks to north and south, largely obscured by intrusions and deformation, is interpreted to be either an unconformity or an early (pre- to syn-D₂) fault.

Tectonic rock units

Straight gneiss (Unit 11), or laminated felsic gneiss, was identified in isolated exposures and in semicontinuous map-scale layers or lenses between quartz-phyric felsic bodies (Unit 8) in the Willroy area, associated with iron formation (Unit 9) and sillimanite-knot felsic schist (Unit 1b) in the Willecho area, and associated with iron formation and sillimanite-muscovite-quartz schist (Unit 1a) south of Wowun Lake. In the Willroy area, zones of straight gneiss coincide with map-scale truncations of iron formation and repetition of a distinctive lithological sequence (see Structural Geology). Locally, straight gneiss is associated with high strain in iron formation, manifested as either straight alternating quartz-rich and dark magnetite-rich laminations (Stop D1) or concordant zones of iron formation slivers and boudins. On the basis of these observations, straight gneiss was interpreted as annealed mylonite (see Hanmer, 1988 for a definition) lying on early ductile faults.

Straight gneiss is fine grained to aphanitic, and characterized by continuous to streaky lamellae (<1–1 mm width) of quartz and feldspar, and at some localities, micaceous or hornblende-rich lamellae. Petrographic examination shows that straight gneiss has an annealed granular texture but, in some cases, strained quartz ribbons were interpreted as remnants of an original mylonitic fabric.

The protolith to straight gneiss apparently mostly comprised felsic rocks and pegmatite. On the north side of the Willecho 3 pit, a transition from pegmatite to straight gneiss is exposed over several metres (Stop A26). With increasing intensity of strain, near-massive pegmatite grades to sheared pegmatite with strongly lineated sillimanite on shear surfaces, to porphyroclastic pegmatite, to streaky fine grained straight gneiss. In porphyroclastic pegmatite, enclaves of coarse grained quartz and feldspar are enclosed by fine grained laminated gneiss containing lineated sillimanite.

Metasomatically altered rock units

Extensive zones of orthoamphibole-bearing gneiss and sillimanite-bearing schist near the mafic-felsic transition in the inner volcanic belt and, to a limited extent, in the outer volcanic belt, are interpreted as zones of synvolcanic hydrothermal alteration, modified by high grade regional metamorphism and deformation.

Orthoamphibole-garnet±cordierite gneiss (Unit 2), on the basis of surface observations and subsurface data, forms a sheet of regional extent, mantling synvolcanic trondhjemite (Unit 12) and folded by the Manitouwadge synform. Orthoamphibole-garnet gneiss extends continuously for at least 30 km from Rabbitskin Lake on the northern limb of the Manitouwadge synform, around the hinge of the fold to the Willroy-Geco area, and east to the Hucamp and Falconbridge zones of subeconomic mineralization (Fig. 3). East of the Hucamp zone, exposure is poor; however, a prominent aeromagnetic anomaly continues the same trend and the unit was intersected in drill holes. North of Straight Lake, orthoamphibole-garnet rocks and sphalerite-bearing iron formation, on the attenuated northern limb of the Manitouwadge synform, are correlated with the same horizon. East of Thompson and Banana Lakes, orthoamphibole-cordierite-garnet gneiss, defining a series of map-scale folds with a northerly trending enveloping surface, is interpreted as a repetition of the same horizon, possibly in the 'keel' of a D₂ fold involved in a D₂/D₃ fold interference pattern (Peterson and Zaleski, 1994b; Zaleski et al., 1995). Orthoamphibole-garnet rocks have also been intersected at depth by deep drill holes penetrating plutonic rocks inside the Manitouwadge synform. For example, the 'Geco deep hole', drilled

to test the down-plunge projection of the Willroy deposits, intersected about 300 metres of orthoamphibole- or sillimanite-bearing rocks.

In the area of known economic mineralization, orthoamphibole-garnet rocks lie north of mineral deposits at Nama Creek, Willroy and Geco, along the contact between synvolcanic trondhjemite and extrusive volcanic rocks. The precursors to alteration, still recognizable in less altered enclaves mostly on the margins of the unit and to the northwest further from known deposits, were mafic (Unit 3) and interlayered mafic-felsic rocks (Unit 4) near the transitional mafic-felsic contact. In the outer volcanic belt, orthoamphibole-garnet assemblages are found locally in mafic metavolcanic rocks in a discontinuous zone near the northern contact to felsic metavolcanic rocks and metagreywacke (Stop B1). Orthoamphibole-bearing zones are much less extensive in the outer belt than in the inner; however, garnetiferous zones are common in mafic rocks.

The unit is defined primarily by the presence of orthoamphibole. Orthoamphibole-bearing rocks are typically layered, with first-order metre-scale layering defined by variations in metamorphic mineral assemblages, proportions, textures and grain size (Stops A6, A13, A14, D3 and D4). Second-order layering consists of semi-continuous quartz-rich lamellae or fine felsic bands, which locally define rootless intrafolial isoclinal folds. In the Willroy area, garnet-orthoamphibole layers that anastomose around hornblende-rich enclaves resemble deformed pillow selvages or fracture-controlled alteration. Local exposures of semi-concordant hornblende-cummingtonite-plagioclase layers, and linear swarms of lenticular felsic enclaves in an orthoamphibole-garnet matrix, have been interpreted as remnants of dykes. In some areas, orthoamphibole-garnet rocks enclose lenticular or snake-like garnetiferous enclaves in a structure suggesting segmented layering and reminiscent of a coarse breccia (Stop D4). In general, the layering is interpreted as the result of transposition of syngenetic structures including tuffaceous bedding, possible pillows or early fractures, dykes or sills, and domains of more and less altered rocks.

The orthoamphibole commonly displays blue iridescence, characteristic of cryptocrystalline exsolution from a composition above the gedrite-anthophyllite solvus (Robinson et al., 1982). It occurs in a variety of assemblages, which include garnet $\pm$ cordierite $\pm$ sillimanite $\pm$ plagioclase and cummingtonite $\pm$ hornblende $\pm$ plagioclase $\pm$ garnet. Hornblende-plagioclase $\pm$ cummingtonite $\pm$ garnet assemblages interlayered with orthoamphibole-bearing assemblages are interpreted as indicative of lower intensity hydrothermal alteration of mafic volcanic rocks. Orthoamphibole-garnet-cordierite assemblages dominate in the Willroy-Geco area, interleaved (on a 10 cm–10 m scale) with sillimanite-cordierite-biotite $\pm$ garnet layers (Stop A5). Quartz and magnetite are ubiquitous, and small amounts of sulphide minerals, staurolite, and gahnite are common. More exotic minerals are found in the Geco mine, including corundum, cassiterite, högbomite [(Ti,Sn)(Fe,Mg,Zn,Mn)₆(Al,Fe)₁₆O₃₂] and nigerite [(Sn,Ti)₂(Zn,Mg,Fe,Mn)₄(Al,Fe)₁₆O₃₂] (Spry, 1982; Petersen, 1986). In orthoamphibole-bearing rocks, biotite habits suggest both metamorphic and retrograde generations, the latter forming pseudomorphs of coarse orthoamphibole sprays (Stop A5). Plagioclase is absent or present in small amounts in orthoamphibole-bearing assemblages, particularly in the Willroy-Geco area.

Sillimanite-muscovite-quartz and sillimanite-knot felsic schist (Unit 1) are subdivisions representing 'end-members' of Unit 1, the former found mainly to the south of orebodies in the Willroy-Geco area, and the latter in the Willecho area and north. The unit also includes transitional quartzites interleaved with micaceous schist. Sillimanite-muscovite-quartz schist occurs in close proximity to massive sulphide deposits and envelopes the Geco main orebody. Abundant muscovite and/or sillimanite and quartz are typical of the unit, and plagioclase, biotite, K-feldspar, garnet and magnetite may also be present. Felsic volcanic rocks lying along strike and in less altered enclaves are interpreted as the protolith to alteration.

Along the southern limb of the Manitouwadge synform in the Willroy-Geco area, two zones of sillimanite-bearing felsic schist converge to the east near the Geco mine, where they are separated by an early fault (see Structural Geology). The northern zone grades easterly from sillimanite-knot schist (Stop A19) to quartz-muscovite-sillimanite schist consisting of finely interlayered quartz and muscovite with sillimanite sprays on the foliation surface (Stop A6). Near Willroy, the southern belt consists of a thinly layered felsic schist that becomes more muscovitic and silicic to the east, merging with the northern belt and grading to a sulphidic muscovite-biotite quartzite in the Geco area.

Near the hinge region of the Manitouwadge synform in the Willecho area, sillimanite-knot felsic schist is interleaved (on a metre scale) with non-sillimanitic felsic schist, straight gneiss and sheared pegmatite (Stop A24). Similar rocks occur sporadically to the south of Willecho, associated with iron formation that can be traced continuously to the Willroy area. The sillimanite-bearing rocks contain quartz, plagioclase, microcline and biotite and are generally more feldspathic than those in the Willroy-Geco area. Sillimanite occurs in coarse knots (1–8 cm in diameter), typically zoned with greenish cores mantled by white fibrous sillimanite. In general, the knots are disseminated and vary in abundance from sparse to about 20 percent. At a few localities, thinly layered quartz-phyric felsic tuff or tuffaceous metasediment contains abundant sillimanite that coalesces into sillimanitic layers. In these cases, the distribution of sillimanite is stratabound, concentrated near the tops(?) of modified beds. A northward increase in metamorphic grade is indicated by a change from

sillimanite-muscovite to sillimanite-microcline assemblages.

Synvolcanic intrusions

Foliated trondhjemite-hornblende granodiorite (Unit 12) underlies a large area inside the Manitouwadge synform, forming an elliptical ring cored by the younger Loken Lake pluton (Unit 13b). Granodiorites and tonalites of similar appearance extend around the Blackman Lake anticline, to the hinge region of the Jim Lake synform. The area of the unit displays a high variable aeromagnetic signature, probably related to the abundance of magnetite and to the presence of supracrustal screens, for example the Dead Lake suite. A deep drill hole, collared about one kilometer north of the Geco mine, penetrated mostly trondhjemite to granodiorite for the uppermost 1470 metres, before encountering fine grained mafic metavolcanic schist.

Along its outer perimeter, trondhjemite intrudes mafic (Unit 3) and mixed mafic-felsic metavolcanic rocks (Unit 4), and orthoamphibole-cordierite-garnet gneiss (Unit 2). The outermost trondhjemite is a quartz-rich (30–50%) leucocratic tonalite with minor biotite, magnetite and, locally, garnet. It varies from fine to coarse grained, but typically contains coarse quartz grains up to 5 millimetres in diameter. Minor or trace amounts of microcline, titanite, apatite, zircon and allanite/epidote may also be present. In the Willroy-Geco area, the abundance of biotite and garnet tends to increase toward the southern contact with orthoamphibole-bearing rocks, and locally, the trondhjemite is host to narrow seams of biotite, orthoamphibole, cordierite and garnet (Stop A7). The seams were interpreted as pre-metamorphic fracture-controlled alteration related to synvolcanic hydrothermal activity, implying that trondhjemite is a synvolcanic intrusion. Our field interpretation of synvolcanic intrusion was confirmed by geochronology (2720 ± 3 Ma, see Geochronology).

Despite some variations in modal abundances, Unit 12 is characterized by abundant coarse grained quartz, disseminated magnetite porphyroblasts (1–3 mm), minor biotite, and a weakly to moderately developed fabric. Away from its outer perimeter in the hinge region of the Manitouwadge synform, foliated trondhjemite grades to more granodioritic and granitic compositions, although not in any regular or systematic fashion. Younger foliated granites, pegmatites and aplites (and probably younger foliated tonalites) are also present, and these are not always easy to distinguish from the synvolcanic intrusion. In the inner synform, the Dead Lake suite is intruded by foliated multiphase tonalite, granite and pegmatite-aplite, and at least some of the tonalite is similar to the synvolcanic trondhjemite (Stop C6). However, most tonalites in proximity to the Dead Lake suite, while resembling those of the outer margin in textures, and in magnetite and quartz content, have a higher proportion of mafic minerals, including hornblende. Some of the hornblende, present as coarse grained (0.5–1 cm), randomly to moderately oriented poikiloblasts, looks secondary. In the Dead Lake suite, leucocratic to melanocratic hornblende-magnetite-plagioclase $\pm$ garnet rocks, with conspicuous quartz eyes or lenticles, resemble altered or contaminated tonalite (Stops C3 and C5). In some cases, blocky inclusions of hornblende-magnetite-plagioclase $\pm$ garnet rock in tonalite have narrow (1–10 cm) transitional contact zones, apparently due to reaction or mixing of the two rock types.

In general, although the more potassic and hornblende-bearing intrusive rocks toward the central Manitouwadge synform, and north and west of the Loken Lake pluton, are texturally and modally similar to synvolcanic trondhjemite, it is not clear in outcrop whether they are comagmatic, or belong to different and possibly younger intrusions. However, geochemical analyses suggest that at least some are comagmatic (see Geochemistry). Foliated trondhjemite has not been observed in the outer volcanic belt; however, amongst undivided foliated intrusive rocks (Unit 14), a layer of strongly magnetic felsic rocks, lying within the mafic sequence striking through Gaug Lake, has some textural and geochemical (see Geochemistry) similarities to the quartz-eye hornblende-magnetite-plagioclase $\pm$ garnet $\pm$ clinopyroxene rocks and magnetite quartzite of the Dead Lake suite.

Syn- to post-tectonic intrusions

Foliated K-feldspar porphyritic granitoid (Unit 13) could be grouped with the Black Pic batholith; however, two mappable bodies can be distinguished from the multiphase plutonic suit (Fig. 3). These lie near the margins of the greenstone belt (including subvolcanic intrusions), and have been involved in D₃ folds. The northwestern contact between supracrustal rocks and foliated tonalites of the Black Pic batholith, inside the Blackman Lake antiform, is mantled by a long sinuous body of foliated hornblende-biotite (up to 35%) granitoid, here called the Nama Creek pluton (Unit 13a). Microcline phenocrysts, typically 1 to 2 cm in length, vary in abundance from sparse to about 25%. The matrix varies from tonalitic to granitic in modal composition, in some cases, containing little K-feldspar. Quartz content varies from about 10–20%, titanite is commonly present in minor amounts (1%), and apatite, zircon, opaque minerals and allanite are trace constituents. A moderate to strong tectonic fabric, defined by mafic minerals and augen-shaped phenocrysts, is folded by the D₃ Blackman Lake antiform and outcrop-scale Z-folds (D₃). The folded fabric, and the presence of folded pegmatite-aplite dykes, suggest that the intrusion is pre- to syn-D₂.

Another microcline phenocrystic granitoid intrudes the innermost core of the Manitouwadge synform (Fig. 3) and is here called the Loken Lake pluton (encompassing the Wowun-Luckyshoe Lakes mass of Williams

et al., 1992). With the exception of the western part of this body, exposure is poor, and the contacts were interpreted to coincide with an elliptical area of low flat aeromagnetic expression. The body is characterized by microcline megacrysts, typically 5 to 15 cm long, varying in abundance from sparse to 25%. Large areas of the intrusion consist of non-porphyrific rock resembling the matrix of porphyritic varieties; however, K-feldspar phenocrysts or augen have been observed close to contacts to the west and northeast of Dead Lake, and south of Thompson Lake. In comparison to the Nama Creek pluton, the Loken Lake pluton has larger phenocrysts and is more leucocratic, containing about 5% biotite. Quartz is abundant (25–40%) and apatite, titanite, opaque minerals, zircon and, in some cases, hornblende are typically present in trace amounts. Where K-feldspar phenocrysts are not present, the rock looks disconcertingly similar to synvolcanic trondhjemite of Unit 12, but geochronology confirms a post-volcanic age of $2687 \pm 2/-3$ Ma (see Geochronology). A strong tectonic fabric, commonly $L > S$, is delineated by quartz, feldspars, biotite, and microcline augen (Stop C1). Variations in fabric orientation suggest folding by the D_3 Manitouwadge synform implying that, like the Nama Creek pluton, the intrusion is pre- to syn- D_2 .

Undivided foliated intrusive rocks (Unit 14) include foliated multiphase plutonic rocks of the Black Pic batholith, as well as foliated tonalitic intrusions within the supracrustal suite of the Manitouwadge belt. Near the southern margin of the Manitouwadge belt, at least 3 intrusive phases can be recognised in the Black Pic batholith. The contact is a transitional zone of interleaved mafic metavolcanic rocks, diorite and younger tonalite, all involved in tight intrafolial folds. The oldest intrusive phase ($2687 \pm 3/-2$ Ma, see Geochronology), medium to coarse grained, foliated hornblende-biotite diorite to monzodiorite with plagioclase augen, dominates some outcrops and forms blocky inclusions in younger phases. Quartz is present from 5–20%, microcline from <2 –15%, and hornblende plus biotite comprise 15–35%. The rock contains minor to trace amounts of titanite and opaque minerals, and trace amounts of apatite, allanite/epidote and zircon. A strong schistosity is defined by mafic minerals and plagioclase augen. Younger foliated monzodiorite (2677 ± 2 Ma) is predominant in much of the area north of the Banana Lake antiform (distribution based on petrographic observations on thin sections made available by F. Breaks, Ontario Geological Survey), that is the area between the easterly extensions of the inner and outer volcanic belts.

Inclusions of the oldest diorite are engulfed by foliated biotite tonalite to granite, possibly representing more than one intrusive phase. The youngest phase on the southern margin of the Manitouwadge belt cuts across diorite, tonalite and granite, as dykes of weakly foliated granite, aplite and pegmatite and, locally as more discrete bodies. These youngest leucocratic rocks have minor to trace amounts of biotite, and trace amounts of opaque minerals and apatite. Diffuse transitions between granite and pegmatitic enclaves can be observed locally.

Tonalitic rocks of the Black Pic batholith extend west of the Manitouwadge belt, inside the hinge region of the Blackman Lake antiform (D_3) and north of the Jim Lake synform (D_3). Near the Blackman Lake antiform, weakly foliated to massive, medium to coarse grained tonalite is commonly present on the short limbs of minor folds and outcrop-scale shear zones. Diffuse contacts to the foliated host rocks are suggestive of localized coarsening and recrystallization, or of anatectic mobilizates that migrated into dilational zones. Evidence of migmatization increases northward toward the Quetico boundary. Northwest of the Manitouwadge belt, septa of mafic to intermediate supracrustal rocks, likely derived from the main supracrustal sequence, correspond to pronounced aeromagnetic striping parallel to foliation trends. A few screens of quartz-magnetite iron formation are present; also mafic inclusions are commonly strongly magnetic. The striping and the linear map pattern are transitional across the Quetico boundary, becoming straighter and trending more nearly east-west toward the north.

Irregular bodies and dykes of homogeneous, foliated, fine to medium grained, granodiorite and tonalite intrude the main supracrustal belt. Tonalites are most common, typically containing plagioclase phenocrysts, and accessory biotite and/or hornblende. The hinge of the Manitouwadge synform in the vicinity of Cadawaja Lake is dominated by granodiorite-tonalite containing inclusions of biotite schist (Unit 10) and felsic to intermediate metavolcanic rocks (Unit 6). In general, the contacts of this body are gradational and their position is poorly constrained. Tonalite is pervasive near the contact between iron formation and metasedimentary rocks in the inner volcanic belt, and in mafic rocks of the outer belt near their contact to metasedimentary rocks. Concordant to slightly discordant tonalite dykes (<10 cm–1.5 m) intrude all supracrustal rock units, in some cases, cutting across straight gneiss (annealed mylonite) or early folds. In most cases, tonalite foliations are parallel to the dominant fabric of the host rocks and at least some tonalite dykes can be interpreted as syn- D_2 (Stop A23).

Pegmatite, aplite and foliated granite (Unit 15) are present as foliated and massive dykes, sheets and irregular subconcordant bodies, a few of which can be shown at the scale of mapping. The largest is a foliated granite intruding supracrustal rocks and synvolcanic trondhjemite immediately north of the Geco mine. In general, the unit comprises undivided leucocratic intrusions of various ages, composed of quartz, K-feldspar and plagioclase, with accessory muscovite or biotite. Garnet is present locally in pegmatite, mainly near

contacts to iron formation. North and west of the Nama Creek deposit, pegmatites have sheared contacts and surfaces coated by aligned sillimanite, for example as described in the transition to straight gneiss north of the Willecho 3 pit (Stop A26). In many cases, pegmatites show tectonic fabrics only near contacts or in fine grained aplitic zones. Locally, discordant pegmatite dykes and dyke swarms show little evidence of deformation, excepting quartz-filled gashes at a high angle to contacts.

Alkalic rocks (Unit 16) including hornblende syenite, hornblendite and lamprophyre, both massive and foliated, were encountered sporadically throughout the map area. These generally form dykes or irregular bodies, too small to show at the scale of mapping, and in some cases, with complex contacts or internal relationships that were not mapped in detail. A massive hornblendite body of about 250 metres in diameter, lying northeast of Cadawaja Lake, is the only one shown on our 1:25000 map, but other exposures of the unit were observed southwest of Swill Lake, northwest of Willecho, near the town of Manitouwadge, and 1.5 km south of Manitouwadge Lake just west of the extension of the Fox Creek fault. The unit generally corresponds to the 'Appinite suite' of Williams et al. (1992) and the Fox Creek locality was described by them.

On outcrop, syenite-hornblendites are usually heterogeneous, pink and green varying from leucocratic to melanocratic, with up to 80–90% hornblende. Tectonic fabrics, where recognizable, are typically weaker than in adjacent host rocks. An exception is a fine grained syenite with a strong foliation folded by asymmetric D₃ folds, exposed near the Geco gatehouse (Stop B1). The rocks are characterized by hornblende phenocrysts, up to 15 mm in length, riddled with inclusions of oriented biotite blades. The orientation of inclusions varies between phenocrysts, apparently being related to hornblende crystallographic structure rather than to any penetrative fabric. Minor clinopyroxene is also commonly present, as skeletal or corroded phenocrysts or as ragged relicts coring hornblende. Titanite and apatite are either abundant or present in trace amounts along with allanite, opaque minerals, epidote and quartz. In many cases, the rocks look like hybrid mixtures of leucocratic and melanocratic material.

Northwest of Willecho, a syenite-hornblendite dyke of about 50 metre width was traced for 500 metres. The dyke consists of dark, medium to coarse grained, hornblende porphyry, with hornblende phenocrysts in a matrix of plagioclase and/or microcline (up to 35%). Diffuse, medium grained to pegmatitic, feldspathic net veins cut the porphyry. The veins are flattened and have a tectonic lineation.

In the town of Manitouwadge, a foliated lamprophyre dyke (about 1 m wide) cuts across diorite and aplitic dykes of the Black Pic batholith. The lamprophyre dyke consists of biotite and clinopyroxene phenocrysts (up to 5 mm in size) in a quartz+plagioclase matrix with minor hornblende and apatite (about 2%).

The Fox Creek occurrence (1.5 km south of Manitouwadge Lake) of syenite-hornblendite is notable, although it was not mapped in any detail. It intrudes Black Pic tonalites of low aeromagnetic response and is itself associated with an isolated high aeromagnetic anomaly (partly masked by diabase dykes). The exposure is relatively extensive (at least 300 metres). Internal structures include orbicular gabbro (Williams et al., 1992), rhythmic layering (cm-scale) defined by mafic and felsic mineral abundance, and truncations of layering that resemble cross-bedding (Stop F5). Locally, the rock is an intrusion breccia comprising lithic clasts (1–10 cm in size) in a hornblende-porphyrific matrix typical of the unit (Stop F4). The clasts are mostly mafic (hornblende-biotite-plagioclase-titanite±epidote±opaque minerals), apparently with coarse grained hybridized margins, in some cases, with a fine grained core. Particularly notable is the presence of a single (observed) massive sulphide clast, 2 centimetres in length, consisting of pyrite, magnetite, ilmenite and chalcopyrite. Presumably the clast population of the breccia represents rocks sampled by the intrusion during its ascent and emplacement; however, the closest known massive sulphide occurrence is the Geco mine 3.5 km to the north. Rock units in this area are steeply dipping and, according to our current understanding of the subsurface structure, unlikely to extend below the intrusion breccia. The subsurface dips and extent of the syenite-hornblende body are not known, but it seems likely that the body sampled some unknown sulphide mineralization during its ascent.

Quetico subprovince

Metasedimentary rocks (Unit 10q) of the Quetico subprovince are dominated by monotonous metagreywacke and biotite schist, lithologically similar to metagreywacke in the Manitouwadge greenstone belt (Unit 10), except for the ubiquitous presence of migmatitic segregations. Quetico migmatitic metasedimentary rocks extend across the northern part of our map, defining a broadly curved subprovince boundary, concave to the south. Similar metasedimentary rocks can be traced in discontinuous outcrops from Jim Lake to east of Appelle Lake, to Davis and Larry Lakes, folded by the Jim Lake synform (D₃) and a related map-scale Z-fold in the Davis Lake area. The position of the folded subprovince boundary is based on exposures of migmatitic biotite schist (Quetico subprovince) and foliated tonalite with mafic inclusions (Wawa subprovince), especially between Appelle Lake and the railway west of Jim Lake. The mafic inclusions were interpreted as the continuation of metavolcanic rocks (Unit 5) of the Manitouwadge belt. Immediately east of Jim Lake, exposures of foliated granitoid with inclusions of biotite schist were grouped with Quetico rocks.

Quetico metagreywackes are compositionally layered (typically 2–50 cm), locally with pelitic to psammitic grading defined by variations in biotite, quartz and feldspar proportions (Stop E3). Pelitic layers commonly

contain garnet and, less commonly, sillimanite. In some cases, hornblende is associated with biotite in mafic layers, especially near contacts to tonalitic rocks with mafic inclusions. Ubiquitous migmatitic segregations, mainly pegmatitic and tonalitic leucosome, locally with garnet and cordierite, are preferentially concentrated in pelitic layers. North of Davis Lake, garnetiferous metagreywacke is locally interlayered with thin (1–2 cm), contorted and segmented, quartz-rich rusty iron formation, and associated migmatitic segregations contain coarse (to 1 cm) grunerite porphyroblasts (Stop E5). A single sample, collected north of Appelle Lake and about 200 metres beyond the northern edge of our 1:25000 map, contained orthopyroxene-biotite-garnet, while another sample about 140 metres to the south contained biotite-garnet-sillimanite-spinel. These assemblages suggest a metamorphic transition to granulite facies just north of the map area.

The western exposures of Quetico metasedimentary rocks, approximately as far east as Appelle Lake, are interlayered with medium to coarse grained gabbroic and dioritic rocks. Foliated gabbro-diorite is found on both sides of the subprovince boundary, but is locally difficult to distinguish unambiguously from tonalitic and mafic (recrystallized metavolcanic) rocks typical of the northern Wawa subprovince in the area. The gabbro-diorite (not a separate unit on our map) is partly equivalent to the Everest Lake pluton, a lenticular intrusion (about 1–1.5 km in width) lying along the Wawa-Quetico subprovince boundary grouped with the Black Pic batholith (Williams and Breaks, 1990a; Williams et al., 1992). Our observations suggest that the intrusion comprises sheets, varying considerably in thickness, that transgress the subprovince boundary. Interlayering with Quetico metasedimentary rocks apparently varies from outcrop-scale (10 m plus) to thin (20 cm) concordant sheets (Stop G2). Aeromagnetic striping in the area, parallel to and transgressing the subprovince boundary, is inferred to reflect the presence of gabbro-diorite sheets. The gabbro-diorite is foliated, but some low strain enclaves are nearly massive, for example in boudinaged remnants of extended layers (Stop G2). Locally, blocky gabbroic inclusions in a matrix of contorted migmatite may have originated as pre-metamorphic dykes.

Proterozoic intrusions

Diabase dykes (Unit 17) likely include representatives of at least three Proterozoic swarms; Matachewan, Biscotasing and Marathon dykes (K. Buchan, pers. comm., 1992). They produce pronounced aeromagnetic lineaments. The most voluminous dykes in the Manitouwadge area are northwest-trending dykes inferred to belong to the Matachewan swarm, dated at 2454 ± 2 Ma (Heaman, 1988). A second set of northeasterly trend, probably belonging to the 2166.7 ± 1.4 Ma Biscotasing swarm (Buchan et al., 1993), is mainly represented by a wide dyke (up to 80 m) striking through Swill Lake. Northerly trending dykes are inferred to belong to the 2170 Ma Marathon swarm (Fahrig and West, 1986).

STRUCTURAL GEOLOGY

Our preferred model accounts for structural observations in the Manitouwadge belt in a 4-stage history of ductile deformation. The map pattern of the belt is largely controlled by northeasterly plunging D_3 folds, including the Manitouwadge synform. D_1 and D_2 folding and faulting produced much of the internal complexity of the belt, particularly apparent in the hinge region of the Manitouwadge synform. Heterogeneously distributed D_4 deformation locally modified the map pattern and caused sporadic development of outcrop-scale structures.

D_1 deformation

D_1 encompasses all structural features that pre-date D_2 deformation and it may represent more than one phase of deformation. Most D_1 structures are interpreted from map-scale observations and are cryptic at outcrop-scale, due to overprinting by subsequent metamorphism and complex deformation. The most compelling evidence for D_1 map-scale structures is in the area of known mineralization (Fig. 4). In the Willroy-Geco and Willecho areas, a prominent discontinuity is associated with truncation of iron formation and other units, and with repetition of the lithological sequence; quartz-phyric felsic rocks, iron formation and massive sulphide mineralization, sillimanite-muscovite-quartz schist. The discontinuity is marked by several mappable zones of laminated straight gneiss (Unit 11), interpreted as annealed mylonite. An additional discontinuity is expressed as a 'finger' dominated by silicate iron formation that extends from the thick southernmost iron formation northerly toward Garnet Lake (accompanying 1:25000 map).

The straight gneiss is typically a hard laminated felsic rock, characterized in thin section by millimetre-scale quartz and feldspar lamellae, mostly annealed to a granular texture. Despite annealing, the grain size remains very fine and, in some cases, vestiges of mylonite fabric survive as quartz ribbons. North of the Willecho 3 pit (Fig. 4), pegmatite is transitional to straight gneiss, as already described (see Tectonic rock units). In the Willecho area in general, straight gneiss occurs adjacent to truncated and boudinaged iron formation. Straight gneiss in the Willroy-Geco area extends easterly and is exposed near the southern end of Wowun Lake, in contact with highly strained, straight-laminated iron formation (Stop D1). These observations led us to interpret straight gneiss as annealed mylonite lying on discontinuities related to early ductile faults.

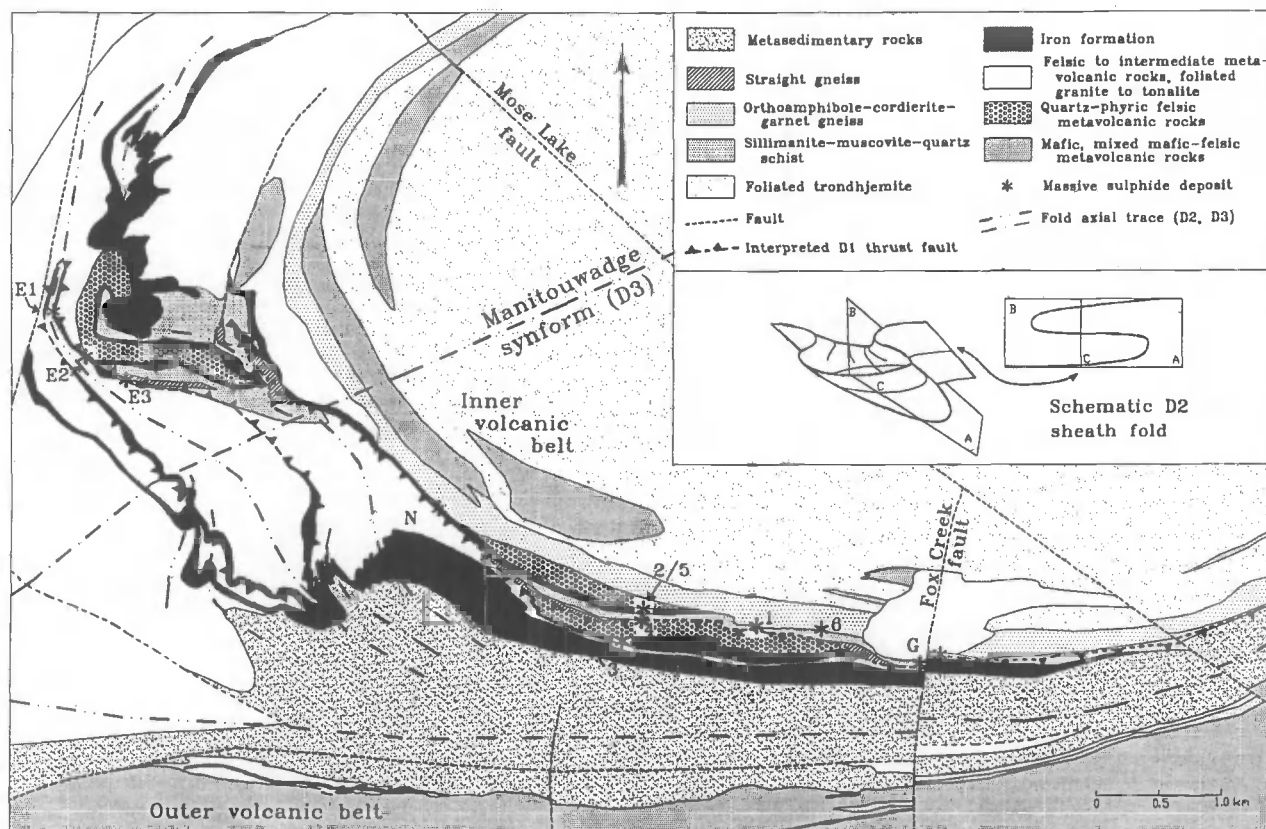


FIG. 4. Geology of the area from the Geco mine (G) in the east, to the Willecho orebodies (1 to 6), the Nama Creek mine (N) and the Willecho orebodies (E1 to E3) in the west. All the known economic Cu-Zn deposits lie in the inner volcanic belt. The inset shows a schematic view of the D₂ sheath fold interpreted to repeat the D₁ fault and mineralized sequence between the Nama Creek (N) and Willecho deposits (E1, E2, E3).

Straight gneiss mostly lacks any well defined lineation, and where a lineation is obvious, for example lineated sillimanite in sheared pegmatite, it may have developed in response to later deformation. Due to the paucity of lineations and kinematic indicators, and the difficulty in interpreting offsets based on stratigraphy, the transport direction of early ductile faults is not known; however, the low angle of the discontinuities suggests thrust faults.

Straight gneiss layering is mostly parallel to dominant D₂ foliation, except in the hinge regions of some D₂ folds. Relationships suggesting a pre-D₂ age for straight gneiss fabrics and associated faults are best preserved in the hinge region of the D₂ map-scale fold southwest of the Nama Creek deposit (Fig. 4). In this area, minor D₂ folds in felsic rocks are defined by an earlier (D₁) gneissosity that we interpret to be related to straight gneiss layering (see Stops A22-A23). The fault that follows the finger of iron formation (see above) toward Garnet Lake is deformed by map-scale D₂ folds.

Northeast of the Willecho deposits, a map-scale isoclinal fold is defined by quartz-phyric felsic rocks, iron formation and sillimanite-knot felsic schist (Fig. 4). The southern limb of the fold is apparently truncated by a D₁ fault, suggesting that the fold was produced by progressive D₁ deformation. Irregular truncation of units along D₁ fault traces in general could be the result of pre- or early D₁ folding. The map pattern of D₁ structures is also complicated by later deformation, particularly during D₂.

D₂ deformation

The dominant planar and linear fabrics, and outcrop-scale folds, ubiquitous in most rocks, are attributed to D₂ deformation. The dominant D₂ foliation is typically a pervasive moderate to strong schistosity, defined by micas, amphiboles, and deformed quartz and feldspar. Except in the hinge regions of D₂ folds, the foliation is a composite D₂-D₁ fabric, generally parallel to lithological contacts. Early elements of the fabric include

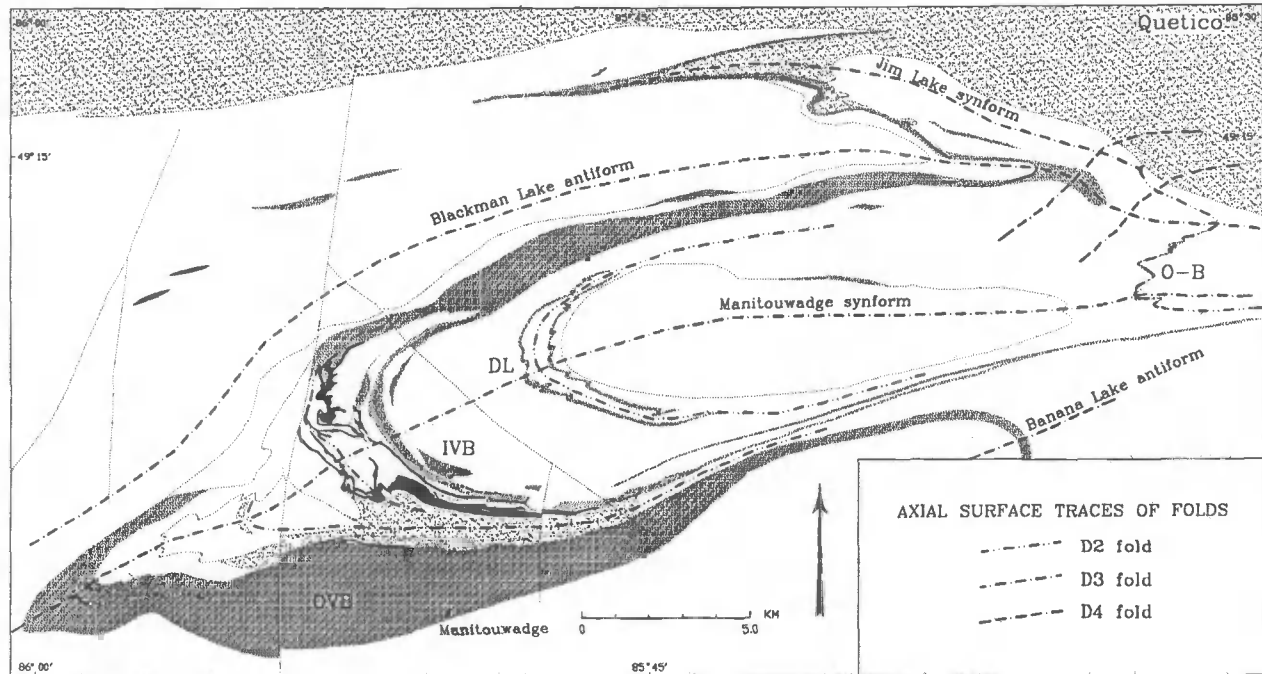


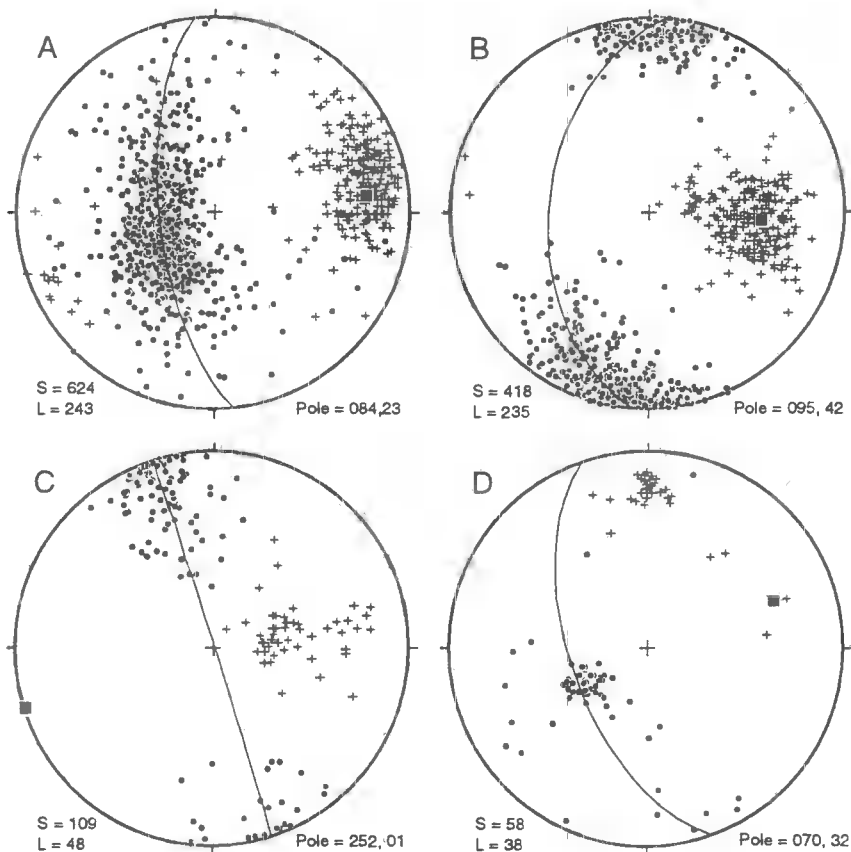
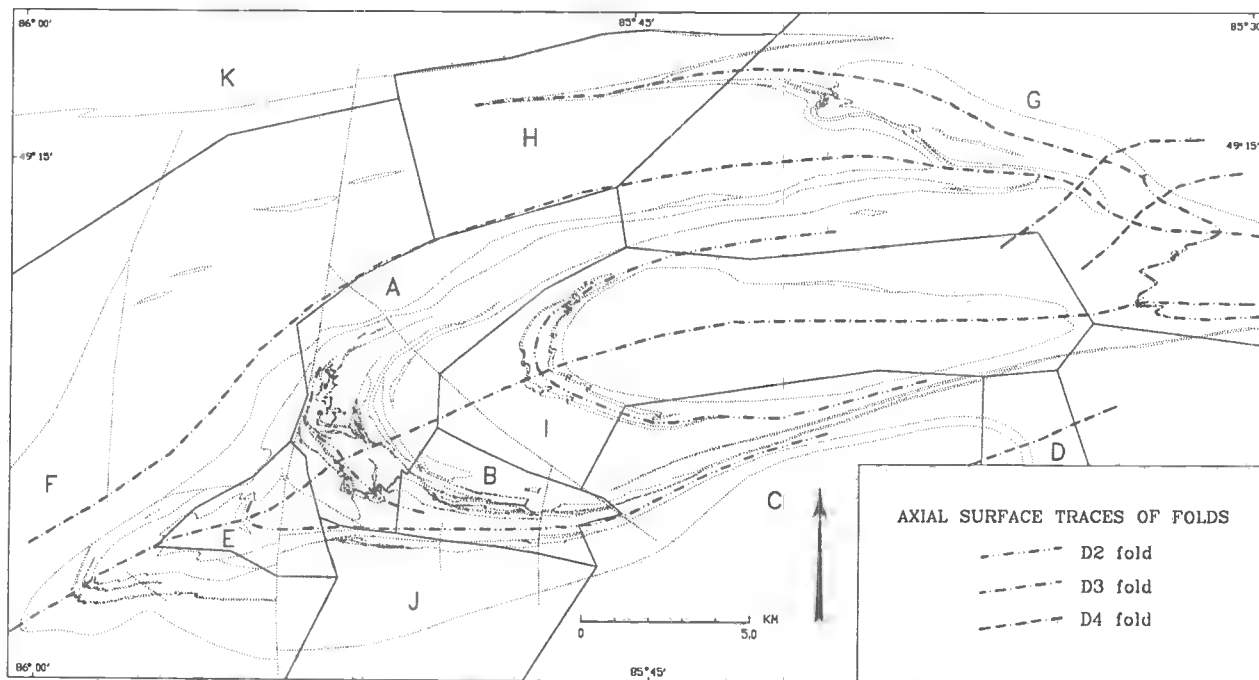
FIG. 5. Axial traces of D_2 synclines, D_3 and D_4 folds. The D_2 'Manitouwadge syncline' repeats the volcanic sequence (OVB = outer volcanic belt, IVB = inner volcanic belt) about a central zone of metagreywackes on the southern limb of the D_3 Manitouwadge synform. The Manitouwadge metagreywackes are interpreted to be correlative with Quetico rocks. Another D_2 syncline could account for the two zones of the Dead Lake suite (DL) or, alternatively, the Dead Lake suite could be equivalent to outer-belt mafic rocks repeated by the D_2 syncline centred on the metagreywackes. Orthoamphibole-cordierite-garnet gneiss and mafic rocks in the One Otter-Banana Lakes area to the east (O-B) are interpreted as repetitions of the supracrustal sequence by a D_2 synclinal 'keel' involved in D_2 - D_3 fold interference patterns.

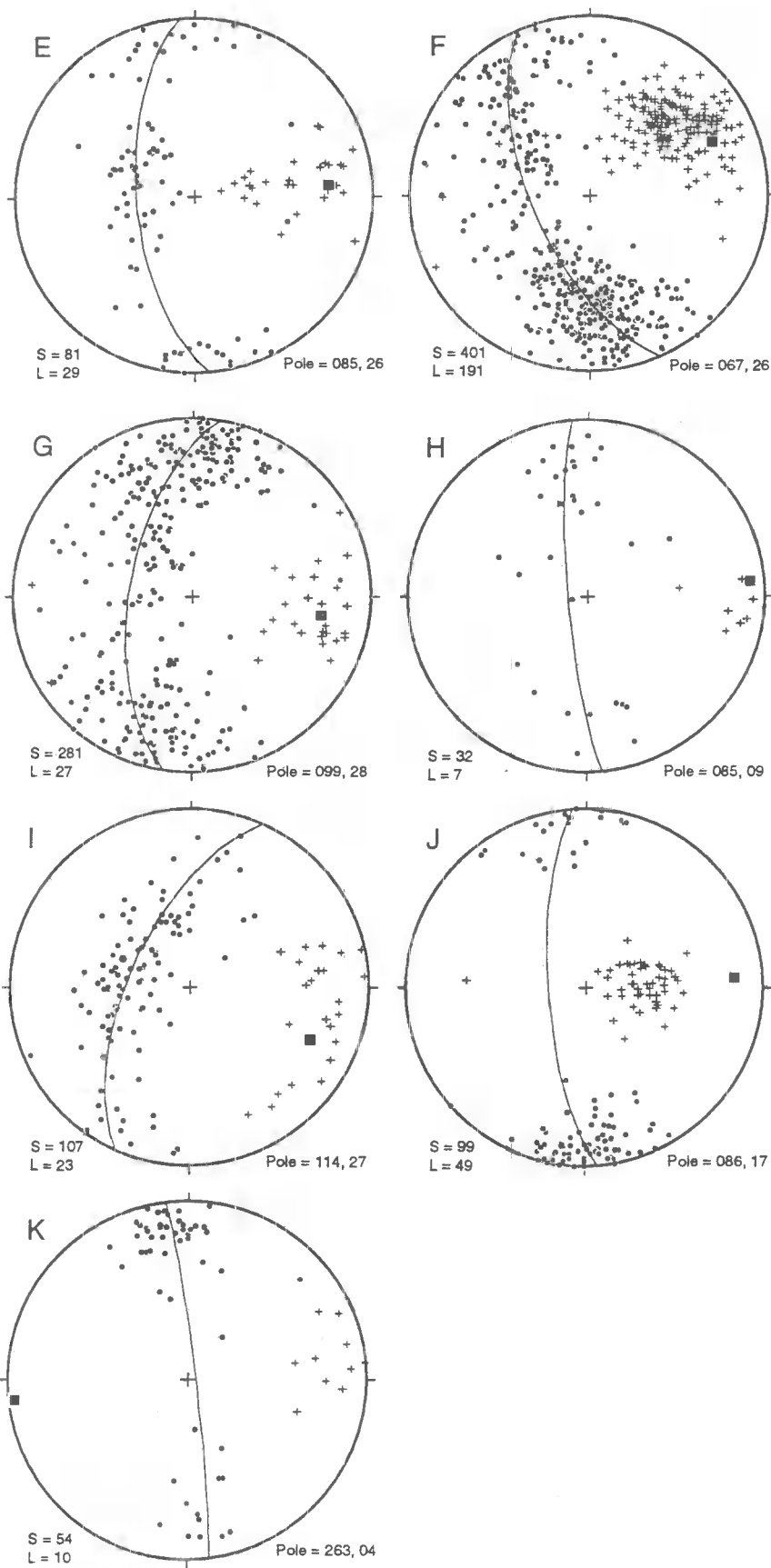
modified bedding, metamorphic layering, transposed intrusive sheets, and shortened pillows, volcanogenic fragments, mafic clots in intrusive rocks and sillimanite knots. D_2 fabrics are well developed in most intrusive rocks, including synvolcanic trondhjemite (Unit 12), early phases of the Black Pic batholith (Unit 14), the Loken Lake and Nama Creek plutons (Unit 13), and undivided intrusive rocks (Unit 14) within the supracrustal sequence. In the Quetico subprovince to the north, the dominant planar fabric is correlated with D_2 fabrics in the Manitouwadge belt. D_2 foliations are deformed by D_3 folds, including the Manitouwadge and Jim Lake synforms and the Blackman Lake antiform (Figs. 5 and 6).

Metamorphic minerals characteristic of upper amphibolite-facies, such as sillimanite, amphiboles and micas, define the orientation of the east-northeasterly plunging D_2 mineral lineation. In some places, unoriented sprays of orthoamphibole and sillimanite are present in the same outcrop as lineated grains. The juxtaposition of lineated and randomly oriented grains may reflect patterns of strain partitioning, or it may suggest post-kinematic recrystallization. Locally, strong elongation of sillimanite knots, feldspar augen, or mafic enclaves produced $L>S$ tectonites. In general across the region, the relationships between orientations of lineations and later folds are variable. In many areas, the D_2 lineation is nearly coaxial with D_3 fold axes; however, in the inner volcanic belt, the orientation of the lineation varies systematically, from the hinge region of the Manitouwadge synform, to the southern limb (Figs. 5 and 6).

D_2 foliations and east-northeasterly plunging mineral/stretching lineations are parallel to axial surfaces and axes of local D_2 folds. In the inner hinge region of the D_3 Manitouwadge synform, map- and outcrop-scale D_2 folds are defined by iron formation and felsic rocks. Locally, minor folds have circular eye shapes produced by curving hinge lines. Southwest of the Nama Creek deposit, a prominent map-scale fold of iron formation has S-asymmetry inconsistent with a parasitic relationship to the D_3 Manitouwadge synform. In the hinge region of the S-fold, outcrop-scale folds change asymmetry across the axial surfaces of map-scale minor folds of

FIG. 6. Structural subareas A to K of the Manitouwadge area, corresponding to the accompanying stereonet (this and following page). Equal area nets (A-K) show the distribution of the dominant D_2 planar (filled circle) and linear (cross) fabric elements. The structural data was parsed into approximately 30 subareas on the basis of similar orientation. The subareas on the map are composites showing the major changes in structural trends related to map-scale folding. The number of data planar (S) and linear (L) data points is shown for each net, and the great circle and pole (filled square) is the cylindrical best fit for planar data.





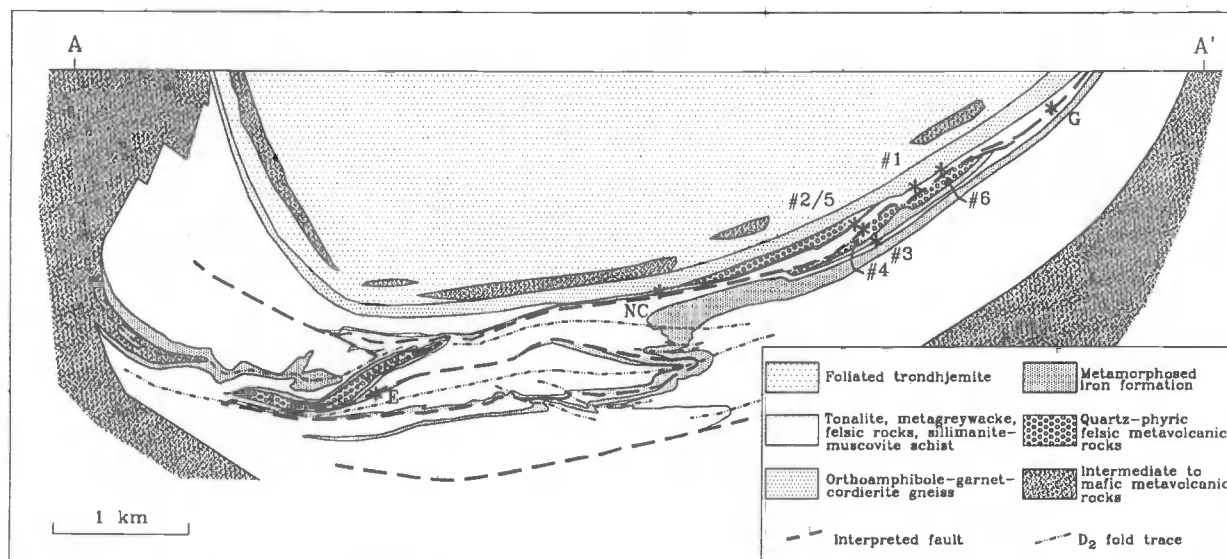


FIG. 7. Down-plunge projection of the inner hinge region of the Manitouwadge synform. The line of projection is the plunge of the synform at $065^{\circ}/25^{\circ}$. A-A' is the line of intersection between the projection plane and the horizontal at 350 metres elevation. Topographic relief (about 140 metres) is not shown. See Figure 3 for the surface trace of the projection. G = Geco mine, 1 to 6 = Willroy orebodies, NC = Nama Creek mine, E = Willecho 3 orebody.

the iron formation-felsic contact (Stops A22–A23). The dominant northeasterly plunging D_2 mineral lineation is parallel to axes of minor folds. Felsic rocks near the contact have a folded gneissosity, interpreted to be a D_1 fabric, and local micaceous enclaves have a schistosity that is oblique to the D_1 foliation and parallel to the axial planes of local D_2 folds. To the south, metagreywackes are involved in map-scale D_2 folds and the dominant foliation is ascribed to a D_2 axial-planar fabric.

Between the Nama Creek and Willecho deposits, lithological units and D_1 faults define a complex geometry interpreted as the result of folding by a map-scale D_2 sheath fold (Fig. 4). The geometry of the pre- D_3 structures in the inner hinge region of the Manitouwadge synform can be viewed in a down-plunge projection (Fig. 7). The mapped D_1 faults are interpreted to be part of a single ductile thrust fault, folded by a map-scale D_2 sheath fold that produced a continuous fault trace to the northeast and an eye fold to the southwest. The sheath fold repeats mineralized iron formation (Willecho deposits) lying on or near the D_1 fault. Alternatively, the map pattern of the area could be explained by mushroom-shaped D_2/D_3 fold interference; however, the nearly coaxial relationship of D_2 and D_3 linear fabric elements (see below) is more consistent with a northeasterly plunging sheath fold. Also, the intrafolial character, rather than a repetitive fold pattern, suggests sheaths rather than interference.

On the southern limb of the Manitouwadge synform, the progression from volcanic to sedimentary rocks (Fig. 3) was previously interpreted as a stratigraphic succession repeated by an early syncline (Suffel et al., 1971; Touborg, 1973), the 'Manitouwadge syncline'. An increase in Zn/Cu of the Geco orebodies from north to south, and the presence of orthoamphibole-cordierite-garnet gneiss (metamorphosed equivalent of footwall synvolcanic alteration) to the north, support southerly younging for the Willroy-Geco area (Suffel et al., 1971; Friesen et al., 1982). Our mapping confirms repetition of the lithological sequence; mafic metavolcanic rocks, felsic metavolcanic rocks interlayered with iron formation, metasedimentary rocks (Figs. 3 and 5). The correlation of volcanic sequences is strengthened by the ages of felsic rocks in both inner and outer belts, the same within error at 2720 Ma (see Geochronology). The extensive unit of orthoamphibole-cordierite-garnet rocks in the inner volcanic belt has a counterpart in the outer belt in sporadic orthamphibole-bearing zones in mafic rocks near their northern contact. In our interpretation, volcanic and sedimentary sequences are repeated across the southern limb of the D_3 Manitouwadge synform by a D_2 syncline with an axial trace in the central metagreywacke (Fig. 5).

In our preferred structural interpretation, map-scale D_2 folding is responsible for repetitions of supracrustal rocks along the axial trace of the Manitouwadge synform (Figs. 5 and 8). The two zones of

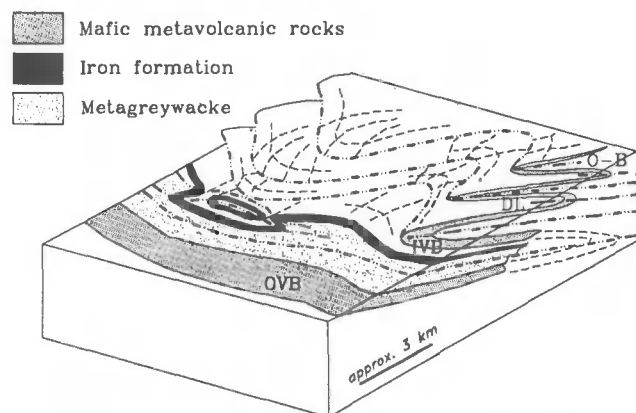


FIG. 8. Block diagram showing schematic pre-D₃ relationships of supracrustal rocks, repeated by D₂ folds. The fold trace in metagreywacke between the inner (IVB) and outer (OVB) volcanic belts represents the 'Manitouwadge syncline' (Fig. 5). The sheath fold defined by iron formation represents the Nama Creek to Willecho area in Figure 4. DL = Dead Lake suite, O-B = One Otter-Banana area.

the Dead Lake suite and exposures of mafic and orthoamphibole-bearing rocks east of One Otter and Banana Lakes may represent D₂ synclinal 'keels', involved in D₂/D₃ fold interference patterns. Alternatively, the Dead Lake suite might simply be septa of mafic volcanic and associated rocks that were included in deeper levels of the synvolcanic trondhjemite intrusion.

D₃ deformation

The regional map pattern is dominated by easterly plunging D₃ folds, from north to south; the Jim Lake synform, the Blackman Lake antiform and the Manitouwadge synform. The Banana Lake antiform could be either a D₃ or a D₄ fold or shear zone. There is a gradational change to the south, from tight fold hinges and dominantly east-west structural trends, to map-scale folds with broader hinge regions and northeasterly traces. The crude Z-asymmetry of map-scale D₃ folds and the curvature of their axial traces suggests formation during oblique dextral transpression.

The shape of the Manitouwadge synform changes from a broad rounded inner hinge to a tighter outer hinge in the Swill-Mills Lakes area (Figs. 3 and B3). In the inner hinge region, the synform has a shallow to moderate northeasterly plunge. The axial surface dips to the south, based on shallow southerly dips of the northern limb, and near-vertical to moderate southerly dips of the southern limb. D₂ planar fabrics in the Willecho-Geco area define a girdle with a moderate to shallow easterly plunging axis (Fig. 6, area A), generally parallel to the axis of the synform. D₂ lineations vary systematically from the hinge region to the southern limb of the synform, defining a small circle rotation (Fig. 6, areas A and B). This suggests a component of flexural slip during folding, possibly due to mechanical anisotropy produced by iron formation or altered rocks. In the outer hinge region, D₂ foliations in supracrustal rocks and in the Black Pic batholith define a girdle with a moderately northeast-plunging pole, and D₂ lineations consistently plunge moderately to the northeast. In agreement with field observations, orientations suggest that early fabrics are strongly transposed by D₃ folding. Local enclaves of westerly plunging D₂ lineations are likely the result of reorientation by minor folds related to the Manitouwadge synform.

D₃ map-scale minor folds are identified across the map area. In the outer hinge region of the Manitouwadge synform, D₃ folds are defined by contacts between felsic or intermediate and mafic metavolcanic rocks. In the inner volcanic belt, iron formation contacts define D₃ folds in the northern hinge region and southern limb of the Manitouwadge synform. In the One Otter-Banana area along the eastern axial trace of the synform, a thin belt of supracrustal rocks defines several folds which also deform the foliation (Fig. 5). Underground at the Geco mine, the 'Geco drag fold' (Brown et al., 1960; Friesen et al., 1982) may also be a D₃ structure. The Z-shaped, easterly plunging fold pair is defined by quartz-muscovite-sillimanite schist, which hosts the main orebody at Geco (see Setting of Mineralization). (Note that Brown et al. mistakenly refer to S-asymmetry in their discussion, but show Z-asymmetry in their figures.) The main orebody has a teardrop shape and lies within the synformal axial region of the 'Geco drag fold' (ibid.). The geometry is consistent with that expected for a parasitic fold related to the Manitouwadge synform and similar to that of D₃ outcrop-scale folds (see below).

Locally developed outcrop-scale D₃ folds vary in character depending on rock type and position with respect to larger structures. In the Swill Lake area in the outer hinge region of the Manitouwadge synform (Fig. B3), two felsic layers in the mafic volcanic sequence define map-scale symmetrical and asymmetrical D₃ folds (Stops B11-B21). The moderately open D₃ folds deform D₂ foliation and, locally, layering. Fold asymmetry changes from dominantly Z-asymmetry on the southern limb of the synform, to M-asymmetry in the hinge region, to S-asymmetry on the northwestern limb. In the vicinity of the Geco and Willroy deposits, both on surface and underground, micaceous rocks locally preserve D₃ folds, typically with shallow easterly or westerly plunges and gently curving hinge lines. Sulphide mineral are commonly concentrated in the hinge

regions of the D₃ folds in the Geco mine. In a few surface exposures, the D₂ mineral lineation is evidently folded at a low angle to the fold axes.

In general, D₃ axial planar fabrics are weakly developed. Locally, a crenulation cleavage lies in the axial regions of D₃ outcrop-scale folds, and an associated crinkle lineation is parallel to fold axes. On the limbs of map-scale D₃ folds, schistosity in micaceous layers is locally oblique at a low angle to the dominant D₂ foliation. The absence of a pervasive D₃ axial planar fabric may reflect post-peak metamorphic conditions, the presence of a strong D₂ anisotropy and the paucity of micaceous rocks.

The Blackman Lake antiform is defined by folded D₂ foliation and mafic to intermediate rocks, which mostly occur as screens in foliated plutonic rocks (Fig. 3). Along the western trace of Blackman Lake antiform in the Janet Lake area (accompanying 1:25000 map), D₂ fabrics in foliated tonalite change orientation around the fold and define a girdle with an east-northeasterly plunging axis (Fig. 6, area F). On the northerly trending short limb of the fold, D₂ foliations in migmatitic rocks are reoriented in outcrop-scale shear zones (1–3 m long) and in convolute minor folds. The shear zones have northerly strikes, steep dips, and either sinistral or dextral offset in plan view. The folds vary in style and orientation, but most have S-asymmetry and moderate northeasterly plunges, and locally, they deform an earlier (D₂?) mineral lineation. Along the easterly trending long limb of the Blackman Lake antiform, south of Blackman Lake (or Kern Lake) (Fig. 3), moderately east-northeasterly plunging, Z-shaped folds are commonly parallel to a distinct stretching lineation. Locally, pegmatites and granitic veins are concentrated in the short limbs of folds. North of Blackman Lake, migmatitic segregations and veins increase toward the Quetico boundary. Abundant screens of mafic gneiss, and locally, iron formation could be an extension of the outer volcanic belt of the Manitouwadge synform. In the One Otter area, the Blackman Lake antiform is defined by the attenuated extension of the inner volcanic belt (Fig. 5). D₂ fabrics define a girdle with an east-southeasterly plunging pole (Fig. 6, area G).

The most northerly D₃ fold, the Jim Lake synform, is defined by zones of mafic inclusions in tonalite and by migmatitic metasedimentary rocks of the Quetico subprovince (Fig. 3). Migmatitic biotite schist can be traced to the south as far as Larry Lake and is involved in a parasitic Z-fold between Jim and Davis Lakes (accompanying 1:25000 map). Away from fold hinges, dominant D₂ foliations are mostly east-west trending and steeply dipping, parallel to the axial surfaces of tight-isoclinal D₃ folds. The foliations define a crude girdle with a near-horizontal easterly plunging axis, subparallel both to lineations and D₃ fold axes.

Outcrop-scale structures in migmatitic rocks in the Quetico subprovince and near the subprovince boundary include, ubiquitous Z-folds with easterly to northeasterly plunging axes that fold the dominant planar fabric (D₂?), shear bands, foliation fish, and rotated boudinage, mostly showing dextral kinematics and interpreted to reflect oblique dextral shear. Their relationship to the dominant fabric suggests that they are correlative with D₃ deformation in the Manitouwadge belt. All of the minor structures deform migmatitic layering, but are transected by discordant migmatitic veins which, in some cases, are subparallel to the axial surfaces of folds. These observations suggest that peak metamorphism and migmatization were synchronous with progressive deformation and the D₃ map-scale folds near the subprovince boundary.

D₄ deformation

D₄ deformation has limited map-scale expression, in part consisting of gentle deflections of the Blackman Lake antiform, Jim Lake synform and the Wawa-Quetico subprovince boundary in the northeastern map area (Fig. 5). The geometry of the deflections is consistent with dextral transpression focussed on the Wawa-Quetico boundary. Another possible map-scale D₄ structure is a northeast-trending shear zone along the northwestern margin of supracrustal rocks folded by the Manitouwadge synform. The shear zone follows a pronounced northeasterly trending lineament, defined by truncated aeromagnetic trends, and generally covered by recent deposits in the Nama Creek valley. Mafic rocks of the outer metavolcanic belt thin dramatically and show increasing transposition with proximity to the shear zone, and may be truncated by the zone. High strain is manifested by spectacular L>S tectonites with a moderate northeasterly plunging stretching lineation, defined by hornblende, quartz rods (especially in folded quartz veins), and streaky mafic lenses in intermediate metavolcanic rocks. Earlier structures, including D₃ and possible D₂ folds, are strongly transposed, so that all linear features are essentially parallel to the local D₄ lineations.

D₄ kink folds and crenulation cleavage are developed locally, particularly in micaceous rocks associated with mineralization on the southern limb of the Manitouwadge synform. Intersections of crenulation cleavage and foliation produce an intersection lineation with a moderate easterly plunge. The cleavage has a variable northeasterly strike, and both kinks and cleavage typically have a Z-asymmetry. In some exposures of quartz-muscovite schist (Unit 1a), D₄(?) kinking produced dextral asymmetric quartz lenses (Stop A6). Crenulations are also present in retrograde biotite-rich zones in orthoamphibole-garnet-cordierite gneiss (Unit 2).

In the vicinity of the Jim Lake fold and in the Quetico subprovince, tight kink folds of small amplitude (<10 cm) are common, plunging moderately to both east and west. In some cases, small northeast-striking shears of foliation are transitional to kink folds. West of Jim Lake on both sides of the Wawa-Quetico boundary, an oblique foliation, usually defined by biotite, strikes northwesterly across the dominant east-west

migmatitic layering. The oblique foliation increases in intensity to the west and is nearly ubiquitous between Appelle and Hourglass Lakes. The foliation transects tight-isoclinal folds in migmatitic rocks and is deformed by D_4 kinks; hence, it post-dates D_3 , and pre-dates D_4 , deformation.

Late Faults

A number of brittle-ductile to ductile faults in the Manitouwadge area were previously identified by Pye (1957) and Milne (1974). The Agam Lake fault (Pye, 1957) forms a pronounced strike-parallel topographic lineament mostly in metasedimentary rocks on the southern limb of the Manitouwadge synform. It is primarily a brittle fault, but local occurrences of L-tectonites suggest an earlier ductile history. It appears to correspond to a change in aeromagnetic signature, from high in the south to low in the north, possibly suggesting involvement of subsurface volcanic rocks.

High-angle faults associated with brittle structures are marked by aeromagnetic and topographic lineaments. The Mose Lake fault (Pye, 1957) has a northwesterly strike, parallel to that of Matachewan dykes, the oldest Proterozoic dyke swarm in the Manitouwadge area. More nearly north-south faults include the Cadawaja fault (*ibid.*), interpreted to extend across the Wawa-Quetico boundary on the basis of left-lateral displacement aeromagnetic trends. The Slim Lake fault (*ibid.*) to the east produced little apparent horizontal displacement. The Fox Creek fault displaces massive sulphide mineralization at the Geco mine, apparently with east-side up and 60 metres of left-lateral movement (*ibid.*; Brown et al., 1960).

GEOCHRONOLOGY

Our ongoing geochronology studies are focussed on determining ages of volcanism, sedimentation, sedimentary provenance, and plutonism, and on bracketing deformational events and metamorphism in the Manitouwadge greenstone belt and the adjacent Quetico subprovince (Fig. 9, Table 2).

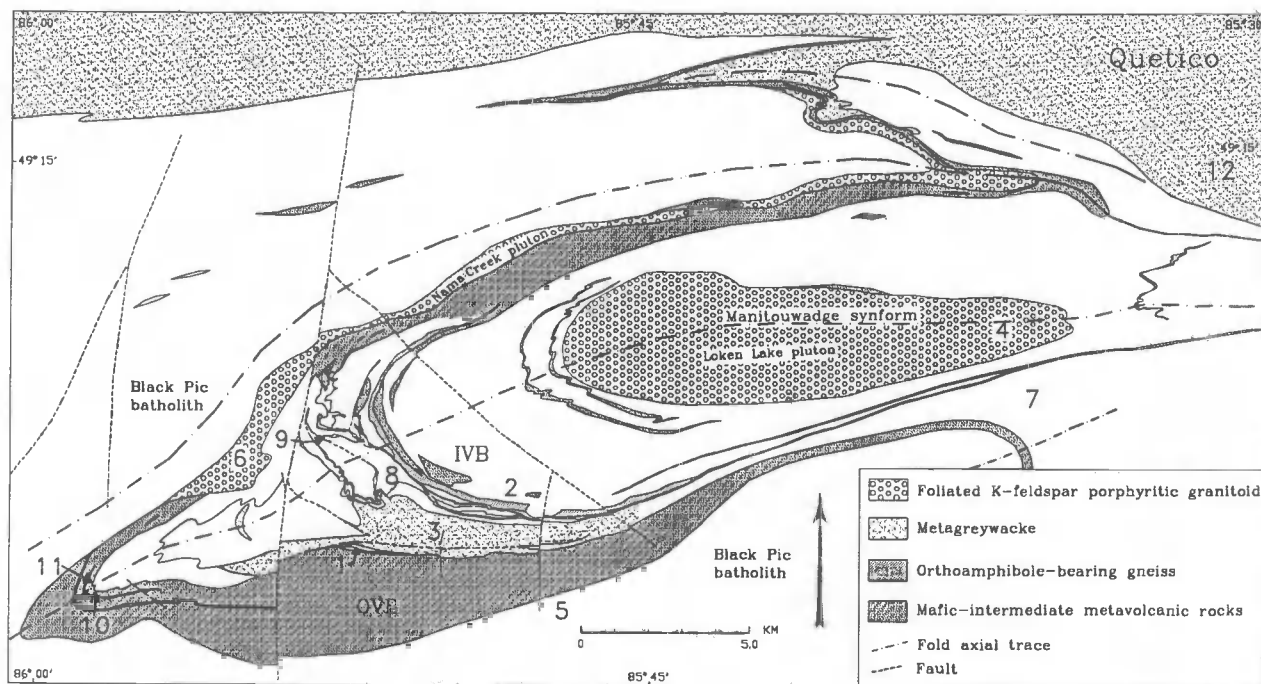


FIG. 9. Locations of 12 geochronology samples. The numbered sample locations match those in Table 2. IVB = Inner volcanic belt, OVB = Outer volcanic belt.

Zircon-Metavolcanic and metasedimentary rocks

The oldest ages thus far emerging from the Manitouwadge belt are U-Pb zircon ages of circa 2720 Ma, dating felsic volcanism in both the inner and outer volcanic belts. An aphyric felsic breccia from the barren (unmineralized) sequence of the outer belt gave an age of 2722 ± 2 Ma (Fig. 10, Stop B5), within error of 2720 ± 2 Ma muscovite schist at the Geco mine, interpreted as an altered felsic volcanic rock (Davis et al.,

1994). Our age of 2720 ± 3 Ma, for foliated trondhjemite sampled north of the Willroy and Geco mines (Fig. 9, Stop A7), confirmed our field interpretation of subvolcanic intrusion (Fig. 11). The 2720 Ma barren and mineralized felsic rocks stratigraphically overlie mafic metavolcanic rocks in the inner and outer volcanic belts and, barring unknown structural complications, mafic volcanism was either contemporaneous (within analytical uncertainty), or significantly older. We are in the process of testing these alternatives with a sample of felsic volcanic breccia interlayered with the mafic sequence between Swill and Mills Lakes (Stop B13).

TABLE 2. Summary of Geochronology

Rock type	Mineral	Age	Relationships
VOLCANISM			
Felsic breccia, outer volcanic belt (1) ¹	zircon	2722 ± 2 Ma ²	
Geco mica schist, inner volcanic belt	zircon	2720 ± 2 Ma ³	felsic volcanic protolith
Foliated trondhjemite, inner volcanic belt (2)	zircon	2720 ± 3 Ma ²	synvolcanic
SEDIMENTATION			
Manitouwadge metagreywacke (3)	zircon	$leq 2693$ Ma ⁴	depositional age, pre-D ₂ post-D ₁
PLUTONISM			
Loken Lake pluton (4)	zircon	$2687 \pm 2 / -3$ Ma	pre- to syn-D ₂
Black Pic diorite (5)	zircon	$2687 \pm 3 / -2$ Ma	pre- to syn-D ₂
Nama Creek pluton (6)	zircon	2680 ± 3 Ma ⁴	pre- to syn-D ₂
Black Pic monzodiorite (7)	zircon	2677 ± 2 Ma	pre- to syn-D ₃
METAMORPHISM and RETROGRESSION			
Geco mica schist	monazite	2675 ± 1 Ma ³	metamorphic
Black Pic monzodiorite (7)	titanite	2674 ± 2 Ma	regional cooling through closure temperature
Nama Creek pluton (6)	titanite	2672 ± 3 Ma	regional cooling through closure temperature
Syn-D ₂ tonalite dyke (8)	monazite	2671 ± 3 Ma ²	retrograde?
Pre- to syn-D ₁ pegmatite (9)	monazite	2669 ± 3 Ma ²	retrograde?
Geco biotite schist	monazite	2661 ± 1 Ma ⁵	retrograde
Syn-D ₃ tonalite dyke (10)	titanite	$2658 \pm 4 / -2$ Ma	retrograde
Syn-D ₃ granite dyke (11)	titanite	2655 ± 3 Ma	retrograde
Loken Lake pluton (4)	titanite	2652 ± 4 Ma	retrograde
Synkinematic granite dyke, Quetico subprovince (12)	monazite	2642 ± 2 Ma	retrograde?

¹Numbers in parentheses refer to locations in Figure 9.

²Zaleski et al. (1994) ³Davis et al. (1994) ⁴Zaleski et al. (1995) ⁵Schandl et al. (1991)

Metagreywacke on the southern limb of the Manitouwadge synform (Fig. 9), contained a suite of concordant or slightly discordant detrital zircons, ranging in age from 2719 ± 2 Ma to 2679 ± 11 Ma (Fig. 12, Stop A1). The youngest zircon, together with another grain of more precise age of 2692 ± 1 Ma, provide a conservative maximum limit on deposition of 2693 Ma, at least 25 Ma younger than the age of felsic volcanic rocks to the north and south. Zircon analyses tend to cluster around 2700 Ma, an age not known elsewhere in the Manitouwadge belt. The oldest zircon of 2719 ± 2 Ma could be derived from local volcanic rocks, which would imply their exposure at or before the time of sedimentation. In general, the zircon population and maximum depositional age are similar to those determined for the Quetico subprovince west of Thunder Bay (Percival and Sullivan, 1988; Davis et al., 1990). The Manitouwadge metagreywacke is lithologically indistinguishable from that of the Quetico subprovince, the main difference being in the higher metamorphic grade and presence of migmatitic segregations in the latter. We are attempting to bracket the minimum depositional age of Manitouwadge metagreywacke by dating a cross-cutting tonalite dyke (Stop A1).

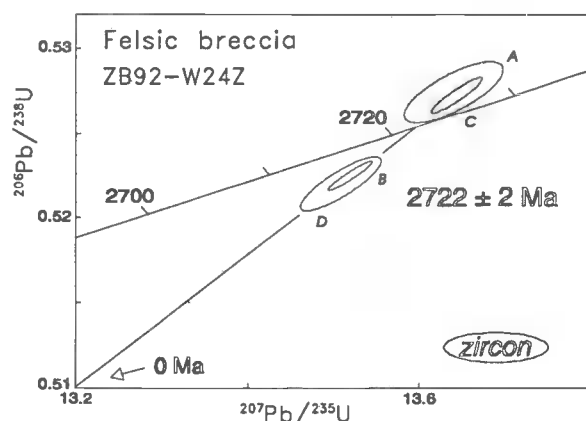


FIG. 10. Concordia diagram showing zircon analyses from felsic breccia in the outer volcanic belt (Location 1, Fig. 9).

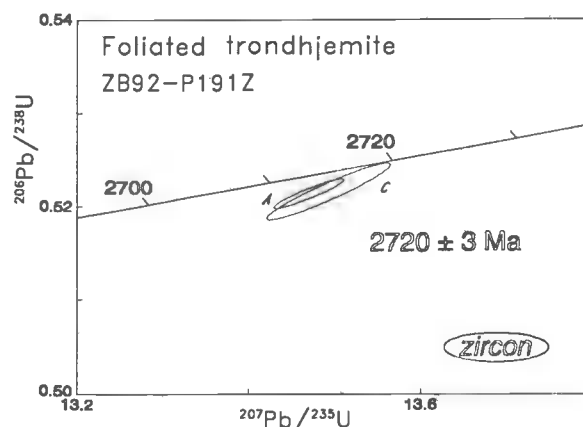


FIG. 11. Concordia diagram with zircon analyses from foliated subvolcanic trondhjemite in the inner volcanic belt (Location 2, Fig. 9).

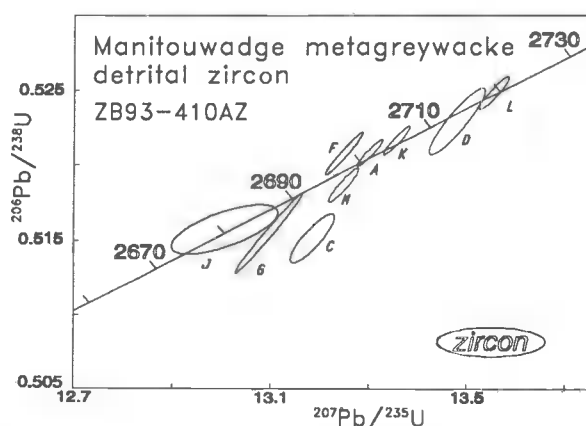


FIG. 12. Concordia diagram of detrital zircons from Manitouwadge metagreywacke between the inner and outer volcanic belts (Location 3, Fig. 9). All analyses are from single abraded crystals, except 'A' which represents a fraction of about 20 grains. The maximum age of deposition is constrained to 2693 Ma by the two youngest grains.

Zircon and monazite—Intrusive rocks and brackets on deformation

We have attempted to bracket deformation by sampling dykes that show relative timing relationships with respect to various generations of tectonic fabrics and structures. In general, we had limited success at dating these intrusions because, either they did not contain zircon, or the zircon was highly fractured and vulnerable to secondary Pb loss. In those cases in which zircon of doubtful quality was analysed, the result was a scatter of discordant points. Some dykes contain monazite, apparently of igneous morphology, analysed in the expectation that it would give primary crystallization ages. Monazite, including some prismatic grains with pyramidal terminations, and fractured zircon (not analysed), were separated from a pre- to syn-D₁ pegmatite transitional to straight gneiss along an interpreted D₁ ductile fault (see Stop A26). Two monazite analyses lie above concordia (Fig. 13), suggesting the presence of excess ²⁰⁶Pb derived from ²³⁰Th, a short-lived intermediate daughter product of ²³⁸U decay. The preferential incorporation of ²³⁰Th into monazite during crystallization commonly results in isotopic analyses that lie above concordia (Parrish, 1990). The spread of analyses down to concordia suggests subsequent minor Pb loss; hence, the concordance of a single grain (Fig. 13, 'Z') is likely due to combined excess ²⁰⁶Pb and Pb loss. The best age estimate is the ²⁰⁷Pb/²³⁵U model age of 2669 ± 3 Ma defined by the slightly discordant grains above concordia.

Southwest of the Nama Creek deposit (Fig. 9), a syn-D₂ tonalite dyke, subparallel to the axial surfaces of D₂ folds of the contact between iron formation and felsic rocks (Stop A23), has an axial planar foliation. The dyke contained only monazite, giving slightly discordant isotopic analyses that define an age of 2671 ± 3 Ma (Fig. 14). This age is within error of monazite from the syn-D₁ pegmatite and of 2675 ± 1 Ma age reported for metamorphic monazite from the Geco mine (Davis et al., 1994).

Two intrusions, interpreted to be syn-D₃, were sampled in the Swill-Mills Lakes area near the hinge of the Manitouwadge synform (Fig. 9). These intrusions have only yielded useful titanite ages (see below). A third sample, from a foliated granitic sheet both deformed by, and cutting, D₃ folds (Stop B18), is still being processed.

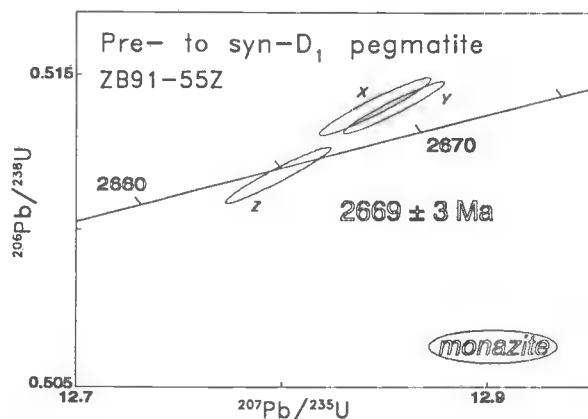


FIG. 14. Concordia diagram showing monazite analyses from a syn-D₂ tonalite dyke (Location 8, Fig. 9).

FIG. 13. Concordia diagram showing monazite analyses from pegmatite involved in a D₁ high strain zone north of the Willecho 3 deposit (Location 9, Fig. 9). Grains lying above concordia suggest the presence of 'excess' ²⁰⁶Pb derived from ²³⁰Th incorporated into the monazite during crystallization. The spread of analyses down to grain 'Z' on concordia suggests subsequent minor Pb loss and, hence, that the concordance of grain 'Z' is a fortuitous combination of excess ²⁰⁶Pb and Pb loss. In view of these features of U-Pb systematics in monazite, the best estimate of age is the ²⁰⁷Pb/²³⁵U model age of 2669±3 Ma defined by grains 'X' and 'Y'.

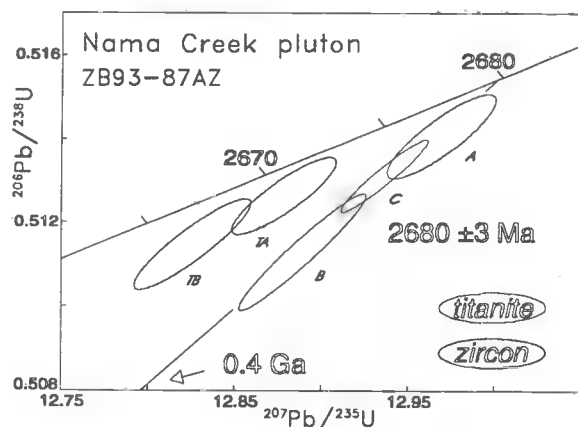
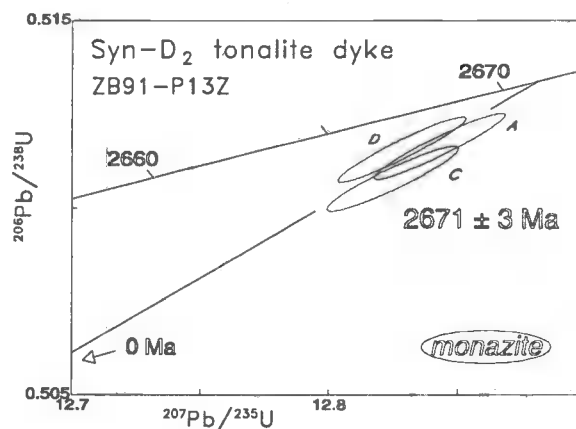


FIG. 15. Concordia diagram showing zircon and titanite (TA, TB) analyses from the pre- to syn-D₂ Nama Creek pluton (Location 6, Fig. 9). The zircon age of 2680±3 Ma is interpreted as the age of the intrusion and the maximum age of D₂ deformation. The titanite age of 2672±3 Ma is interpreted as the time of regional cooling through the closure temperature of titanite.

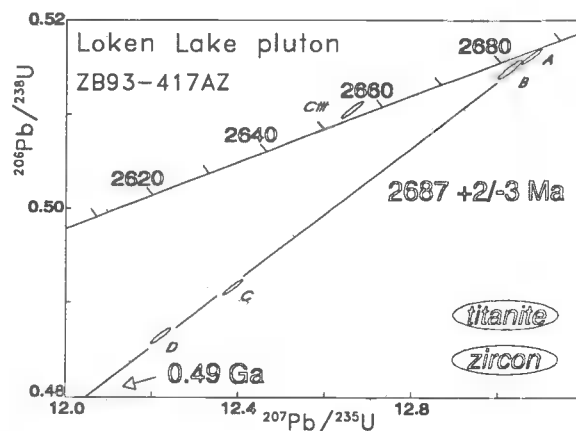


FIG. 16. Concordia diagram showing zircon and titanite (Ctit) analyses from the pre- to syn-D₂ Loken Lake pluton (Location 4, Fig. 9). The zircon age of 2687±2/-3 Ma is the age of intrusion, whereas the titanite age of 2652±4 Ma is interpreted as a late retrograde event that crystallized or reset titanite.

In the Quetico subprovince (Fig. 9), a muscovite-biotite granitic dyke was interpreted to be contemporaneous with progressive deformation (late D₃ or D₄) and, judging by the presence of muscovite, late in

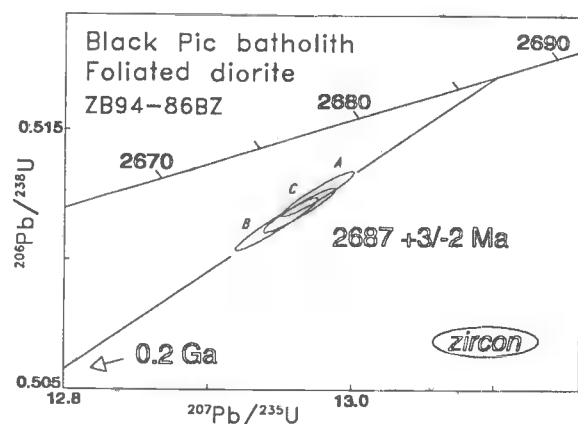


FIG. 17. Zircon concordia diagram of the oldest phase of the Black Pic batholith, foliated plagioclase porphyritic diorite (Location 5, Fig. 9). The age of $2687 \pm 3/-2$ Ma is within error of that of the pre- to syn-D₂ Loken Lake pluton.

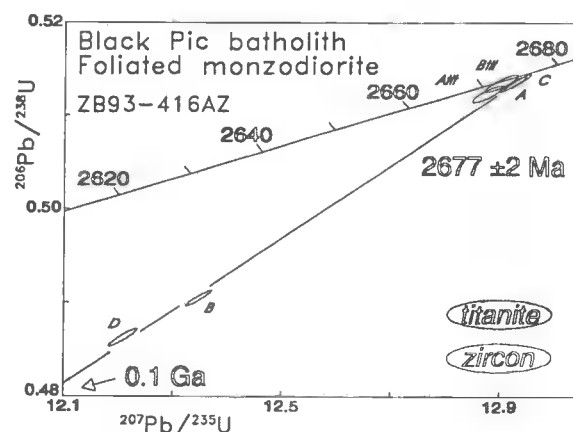


FIG. 18. Concordia diagram showing zircon and titanite (Atit, Btit) analyses from foliated monzodiorite of the Black Pic batholith (Location 7, Fig. 9). The zircon age of 2677 ± 2 Ma dates the intrusion and is within error of the Nama Creek pluton. The titanite age of 2674 ± 2 Ma indicates cooling through the closure temperature of titanite soon after intrusion.

the metamorphic and anatexis history. The dyke is subparallel to the axial planes of Z-folds in migmatitic rocks and has an axial planar foliation (Stop E2). Shear bands in the host rock and the dyke cut across the contact. Zircons from the dyke gave scattered discordant analyses, apparently in part due to inheritance. Of two monazite grains analysed, one was highly discordant, and the other slightly discordant above concordia (see discussion of excess ^{206}Pb above), giving an age of 2642 ± 2 Ma (Table 2).

We had better success in dating larger plutonic bodies. In some cases, the relative age of these with respect to deformation can be inferred with some confidence. Two K-feldspar porphyritic granitoids, the Nama Creek and Loken Lake plutons (Fig. 9), are both interpreted as pre- to syn-D₂ intrusions based on tectonic fabrics deformed by D₃ folds. Analyses of zircon from Nama Creek pluton define an age of 2680 ± 3 Ma (Stops F11-F12, Fig. 15), whereas those from the Loken Lake pluton are slightly older at $2687 \pm 2/-3$ Ma (Fig. 16). The oldest dioritic and the youngest aplitic (not processed) phases, both foliated, of the Black Pic batholith were sampled south of the Manitouwadge belt (Stop F3). Zircon analyses from the diorite yield an age of $2687 \pm 3/-2$ Ma (Fig. 17), within error of the Loken Lake pluton. Foliated monzodiorite of the Black Pic batholith, between the eastern extensions of the inner and outer volcanic belts (Fig. 9), has an age of 2677 ± 2 Ma (Fig. 18). The age is within error of the Nama Creek pluton, but the monzodiorite foliation could be a D₂ or D₃ fabric.

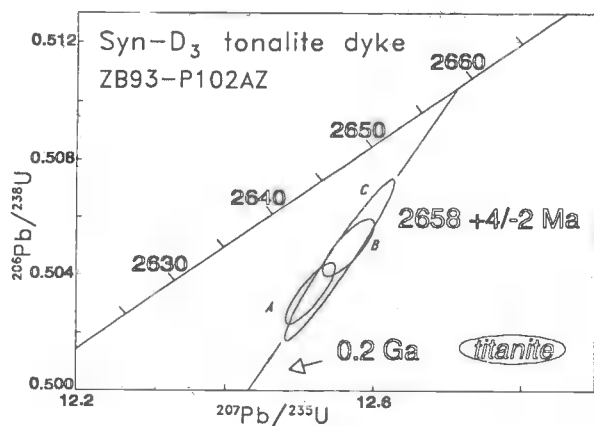


FIG. 19. Concordia diagram showing titanite analyses from a syn-D₃ tonalite dyke near the hinge region of the Manitouwadge synform (Location 10, Fig. 9), defining an age of $2658 \pm 4/-2$ Ma.

Titanite

U-Pb isotopic analyses were done on titanite from intrusive rocks in order to place minimum age constraints on magmatism and metamorphism. The analyses group around two distinct ages of 2673 and 2655 Ma. The 2680 Ma Nama Creek pluton and the 2677 Ma Black Pic monzodiorite have titanite ages of 2672 ± 3 Ma (Fig. 15) and 2674 ± 2 Ma (Fig. 18), respectively. We tentatively interpret these as the time of regional cooling through the 600°C closure temperature of titanite (Heaman and Parrish, 1991) and, for the case of the Nama Creek pluton, about 8 Ma after intrusion. Titanite from the 2687 Ma Loken Lake pluton defines an age of 2652 ± 4 Ma (Fig. 16), whereas syn-D₃ intrusions in the Swill-Mills Lakes area gave ages of $2658 \pm 4/-2$ Ma from a tonalite dyke (Fig. 19), and 2655 ± 3 Ma from a granitic sheet (Table 2). The circa 2655 Ma titanites suggest a widespread late hydrothermal event that locally crystallized or reset titanite.

SETTING OF MINERALIZATION

The Geco, Willroy, Nama Creek and Willecho massive sulphide deposits lie in the inner volcanic belt along the southern limb and hinge region of the D₃ Manitouwadge synform (Fig. 3). Our understanding of the complex pre-D₃ deformation of the area and of the probable original relationships of sulphide deposits depends on detailed lithological and structural mapping, as well as any additional clues that could be gleaned from features characteristic of individual deposits. Except for Nama Creek, each of these deposits comprised several massive, semi-massive or disseminated sulphide orebodies of widely varying Cu, Zn and Pb grades and ratios (Table 3). On the basis of Cu-Zn-Pb proportions, nature of mineralization and relationship to iron formation, the sulphide orebodies of the Manitouwadge camp can be divided into three main types. Firstly, Cu-rich stringer and disseminated orebodies are hosted by orthoamphibole-cordierite-garnet gneiss (footwall alteration) or envelop the Geco main orebody in sillimanite-muscovite-quartz schist. Secondly, massive and semi-massive Zn-Cu-(Pb) orebodies are associated with iron formation horizons interleaved with sillimanite-muscovite-quartz schist or quartz-phyric felsic rocks. Thirdly, massive and semi-massive Zn-Pb-(Cu) orebodies are hosted by iron formation south of sillimanite-muscovite-quartz schist. With some exceptions (see below), the spatial distribution of orebodies defines a tendency to more Zn-rich and Pb-rich deposits toward the south. In terms of classification based on Cu-Zn-Pb, the Manitouwadge belt collectively fits Franklin's (1986) Cu-Zn Group 1a, which includes the deposits of the Abitibi greenstone belt, the Norwegian Caledonides and other areas dominated by basaltic volcanism (Fig. 20).

TABLE 3. Types and Grades of Mineralization

Orebody	Type	%Cu	%Zn	%Pb	g/tAg
Geco 4/2 Cu	disseminated	1.8	0.3	—	16
Geco Main	massive	2.3	8.2	0.4	74
Geco Main	stringer, disseminated	1.5	0.1	—	22
Geco 8/2 Zn	massive, disseminated	0.2	4.9	—	36
Geco ZnIF	semi-massive	0.1	8.6	0.8	41
Willroy 1	disseminated, stringer	1.3	—	—	9
Willroy 2	semi-massive	<0.1	6.0	0.2	64
Willroy 3	massive	1.3	4.5	0.2	56
Willroy 4	massive, semi-massive	—	8.0	1.2	168
Willroy 5	semi-massive	—	6.0	0.2	28
Willroy 6	stringer	2.0	1.0	—	14
Willroy 7	massive	0.4	5.0	—	39
Nama Creek	semi-massive	0.8	3.9	—	28
Willecho 1	semi-massive	1.0	3.0	—	42
Willecho 2	semi-massive	0.2	5.8	—	42
Willecho 3	semi-massive	0.6	3.8	—	53

Data compiled by H. Lockwood, Noranda Inc. from unpublished company records. Willroy data from an unpublished report by Derry, Michener and Booth (1971) for Willroy Mines Inc. Willroy %Pb from Timms and Marshall (1959), or estimated assuming near constant Zn/Pb. The Willroy 7 deposit was never mined.

The sulphide mineralogy is generally simple, consisting of pyrite, pyrrhotite, chalcopyrite, sphalerite and galena (Timms and Marshall, 1959; Friesen et al., 1982). Many details of individual orebodies have been

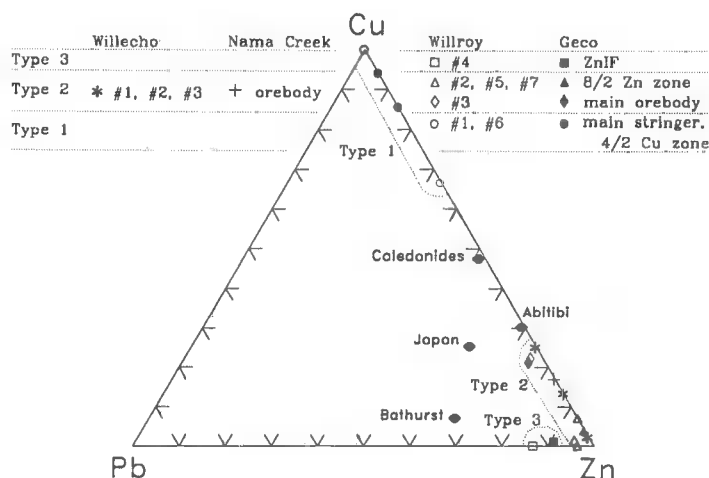


FIG. 20. Cu-Zn-Pb ternary diagram (weight %) showing the metal composition of the orebodies of the Manitouwadge camp. In the key and the dotted fields, orebodies are grouped into Types 1, 2 and 3 (see text). The average base-metal proportions in the Abitibi belt, Norwegian Caledonides, Japanese Green Tuff belt, and the Bathurst-Newcastle area of New Brunswick are from Lydon (1984).

described by Pye (1957), Timms and Marshall (1959), Brown et al. (1960) and Friesen et al. (1982) and only observations pertinent to our reinterpretation are reviewed here.

Relationships of orebodies and host rocks

The Geco 4/2 Cu zone, and Willroy 1 and 6 deposits, consist of Cu-rich, disseminated or stringer sulphides on or near the contact between orthoamphibole-cordierite-garnet gneiss and sillimanite-muscovite-quartz schist (Fig. 4, 21). On the south side of the 4/2 Cu zone, a weakly mineralized 'breccia' of flattened quartz lenses in an orthoamphibole matrix has been variably interpreted as altered pyroclastic rock (Friesen et al., 1982) or disrupted altered iron formation (Bakker et al., 1985).

The Geco main orebody lies south of the 4/2 Cu zone, enveloped by sillimanite-muscovite-quartz schist and quartzite. It is by far the largest and highest grade of the orebodies, although the proportions of Cu-Zn-Pb are similar to Willroy 3 (Fig. 20). A halo of chalcopryite disseminated in sillimanite-muscovite-quartz schist surrounds the massive zone (Table 3, Geco main stringer), especially on the lower eastern margin and south side (Friesen et al., 1982). The main orebody has been traced to the east in underground workings, where it grades and thins to Zn-rich massive sulphide interleaved with iron formation, the latter correlated with iron formation formerly exposed in the Geco quarry at surface (Geco mine, unpublished cross-sections). We are not able to confirm this relationship and, furthermore, surface mapping of the surviving exposures shows that the Geco main orebody lies to the north of the southernmost iron formation, and that the iron formation is continuous to the quarry. The subsurface relationships imply that the eastern extension of the main orebody crosses stratigraphy at a low angle (*ibid.*); but we are unable to discriminate between possible syngenetic or epigenetic relationships (see below).

Low grade economic and subeconomic Cu mineralization is continuous from the Geco main orebody to the 8/2 Zn zone (Friesen et al., 1982), near the transition from sillimanite-muscovite-quartz schist to more quartzitic varieties of the same unit. The 8/2 Zn zone and iron formation on the same horizon project to surface south of the main orebody. The southernmost orebody at the Geco mine is a zincian iron formation (Table 3, ZnIF) in micaceous quartzite, interpreted as an altered volcanic rock. Zincian iron formation attains economic grades in deeper levels of the mine near the Mose Lake Fault (Fig. 4) (Bakker et al., 1985; Williams et al., 1990).

The Geco main orebody was displaced by the steeply dipping, high-angle Fox Creek fault (Pye, 1957; Brown et al., 1960). Well away from the Fox Creek fault, evidence of tectonic remobilization of sulphide minerals is ubiquitous, in part related to layer-parallel movement (Friesen et al., 1982). For example, wall-rock breccias have a sulphide matrix, and disrupted segmented pegmatite and tonalite dykes can be traced across massive sulphide. All orebodies are conformable to the nearly vertical host rocks, and elongate with shallow to moderate easterly plunges, subparallel to the dominant mineral lineation. The massive sulphide zone of the Geco main orebody is thickest in the hinge of the easterly-plunging Z-shaped 'Geco drag fold' (*ibid.*), inferred to be a D₃ or later fold based on the involvement of folded foliation. The Z-asymmetry is consistent with the geometry expected for a parasitic fold on the southern limb of the D₃ Manitouwadge synform. Locally, Cu grades are highest in fold noses (Friesen et al., 1982) and minor (less than metre-scale) folds have pods of massive chalcopryite in dilational zones.

The Zn-rich Willroy 2, 3, 4, 5 and 7 deposits lie on iron formation horizons south of the Willroy 1 and 6 Cu-stringer deposits, intercalated with quartz-phyrlic felsic metavolcanic rocks and sillimanite-muscovite-

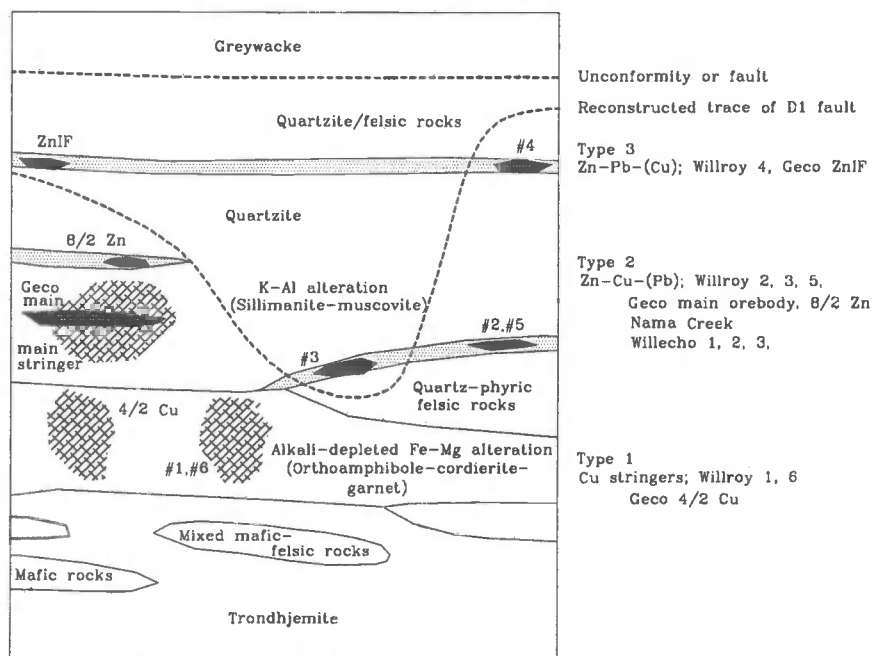


FIG. 21. Schematic reconstructed section of the probable relationships between orebodies, showing the trace of the D_1 fault that divides the area. Stringer/disseminated zones are shown as cross-hatching, and iron formations are stippled. Some ambiguity remains in detailed correlations of mineralized horizons between the Geco main and Willroy orebodies.

quartz schist. The Willroy 3 orebody, the southernmost and largest of the Willroy deposits, is zoned from a relatively Cu-rich core ($\text{Cu}=\text{Zn}$) to a Zn-rich margin, to barren pyrite-pyrrhotite sulphidic iron formation (Pye, 1957; Timms and Marshall, 1959). The Willroy 4 orebody and Geco zincian iron formation have the highest Pb grades of the Manitouwadge deposits (Table 3, Fig. 20).

The Nama Creek deposit, located near the contact between orthoamphibole-cordierite-garnet gneiss and sillimanite-muscovite-quartz schist, consisted of marginally economic Zn-rich mineralization (Table 3), hosted by iron formation and much intruded by pegmatite. North of the hinge region of the Manitouwadge synform (Fig. 4), three Willecho deposits are highly deformed Zn-rich semi-massive sulphide sheets interleaved with iron formation and altered rocks.

Structural complications— D_1/D_2 folds and faults

The interpreted D_1 fault of the Willroy-Geco area continues eastward in sillimanite-muscovite-quartz schist to highly strained iron formation and straight gneiss exposed east of Wowun Lake (Fig. 3). Our observations allow some latitude in the position of the fault with respect to the Geco main orebody; it could lie north or south of the orebody. The Willecho deposits lie near or on the D_2 fold repetition of the D_1 fault, and their original stratigraphic relationship to the Willroy-Geco orebodies is obscure.

In our interpretation, the D_1 fault divides the Nama Creek-Willroy-Geco area into 2 tectonic blocks (Fig. 21). The Willroy 3 orebody and Geco zincian iron formation lie in the southern block, and the Willroy 1, 2, 4, 5 and 6 orebodies, Nama Creek and Geco 4/2 Cu zone lie in the northern block. The Geco main orebody and 8/2 Zn zone could lie either on the northern margin of the southern block, or the southern margin of the northern block. The Cu-rich disseminated and stringer zones, the Geco 4/2 Cu zone, Willroy 1 and 6 deposits are interpreted to lie on the same horizon, and are probably related to the disseminated Cu mineralization around the Geco main orebody. Zn-rich massive and semi-massive sulphide bodies north of the D_1 fault, the Willroy 2, 5 and Nama Creek deposits, could lie on the same iron formation horizon as the Willroy 3 deposit on the south side of the fault. The Geco zincian iron formation in the southern block is interpreted to be correlative with iron formation hosting the Willroy 4 deposit in the northern block.

The Willroy 3 and Geco main orebody are similar in Zn-Cu-Pb (Fig. 20). Cu-to-Zn zoning toward iron formation, such as that exhibited by the Willroy 3 deposit, is characteristic of vent-proximal massive sulphide bodies (Lydon, 1984). Similarly, stringer sulphide zones, like those associated with the Geco main orebody, are typical of subsurface feeder zones of proximal deposits. These features suggest the Willroy 3 and Geco

main orebody precipitated close to a hydrothermal vent, and that the Geco main orebody could also lie on or near the Willroy 3, 2, 5 and Nama Creek horizon. However, we cannot discriminate between several possible interpretations of the relationships between the Geco main orebody, its iron-formation-hosted eastern extension in the Geco quarry (i.e. the southernmost iron formation), and early faults. Structural complications are possible including, early faulting juxtaposing the main orebody with an unrelated iron-formation-hosted deposit and, remobilization of sulphides along early faults and iron formation layers. The Cu-to-Zn zoning, reported in the sulphide-iron formation transition in the quarry, argues against remobilization, as does the observation that, at Geco and in other deformed massive sulphide bodies (e.g. Valenta, 1994), remobilization generally involves redistribution of chalcopyrite.

The increase in Zn/Cu of the Geco orebodies from north to south, characteristically a stratigraphic-upwards zonation in volcanogenic massive sulphide deposits, and the presence of orthoamphibole-cordierite-garnet gneiss, the metamorphosed equivalent of synvolcanic footwall alteration, to the north have both been used to infer southerly younging in the Willroy-Geco area (Suffel et al., 1971; Friesen et al., 1982). The northerly position of the synvolcanic trondhjemite intrusion is consistent with exposure of progressively deeper levels of a volcanic edifice in that direction. Southerly younging in the Willroy-Geco area implies that the northern tectonic block is, in general, stratigraphically lower than the southern tectonic block (Fig. 21).

A consequence of D₂-fold repetition of sequences on the southern limb of the Manitouwadge synform is that the mineralized and altered intervals of the Willroy-Geco area are repeated in the outer volcanic belt, assuming that they were not removed by erosion or faulting along the contact with younger greywackes. Sporadic zones of orthoamphibole, garnet and minor cordierite in mafic rocks of the outer volcanic belt are interpreted as incipient synvolcanic alteration correlative with altered rocks in the Willroy-Geco area.

GEOCHEMISTRY

A suite of 127 samples was collected to represent supracrustal and plutonic rocks of the Manitouwadge greenstone belt, the Black Pic batholith, and the Quetico subprovince. These were analysed for major elements, ferrous iron, H₂O, CO₂ and S_t, and most were also analysed for trace elements, including rare-earth elements (REE). Our data are supplemented by some major-element analyses of samples from the area of known mineral deposits, made available by Noranda Inc. Evaluation of results of our geochemistry program is ongoing and many of the interpretations presented here are incomplete or preliminary.

Metavolcanic and subvolcanic rocks

A sample suite of 54 metavolcanic and associated rocks was chosen to represent primary geochemistry, avoiding evidence of synvolcanic alteration and later retrogression as much as possible (with some exceptions discussed below). The suite represents 5 main rock types; aphyric felsic metavolcanic rocks, quartz-phyric metavolcanic felsic rocks, foliated trondhjemite-granodiorite, mafic metavolcanic rocks, and hornblende-magnetite-garnet rocks/magnetite quartzites. The latter include magnetite-rich layered rocks (see Unit 5), resembling sedimentary concentrations of heavy minerals, and quartz-richer transitional rocks interpreted as contaminated trondhjemite. To facilitate discussion, the suite can be subdivided into 6 geographic groups; the inner and outer volcanic belts, and the Dead Lake, eastern extension, One Otter-Banana Lakes and Jim-Davis Lakes areas (Fig. 22). The foliated trondhjemite-granodiorite in the inner volcanic belt is interpreted as synvolcanic; similar rocks were sampled to the east and north to test their compositional similarity and the extent of possibly comagmatic intrusive rocks.

Mafic metavolcanic rocks throughout the Manitouwadge belt are tholeiitic basalts (Fig. 23). The compositions of mafic rocks from the inner and outer volcanic belts, and from the attenuated and folded extensions of the belt, for the most part, are permissive of correlation and suggest that all mafic rocks could have been produced by the same magma system. The tendency of the outer belt to high Mg, low TiO₂ (Fig. 24), and low light REE (Fig. 26) contents, is defined by gabbros and samples from the Swill-Mills Lake area (near the hinge of the Manitouwadge synform in the southwestern part of Fig. 22). Mafic compositions in the inner and outer belts show considerable overlap, and the slight differences could reflect a bias in which gabbros and rocks deeper in the stratigraphic section were not sampled in the inner belt. The single sample from the inner belt that plots in Jensen's (1976) field of basaltic komatiite (Fig. 23) is associated with altered rocks (see below) and may have undergone incipient magnesian alteration. A sample (from the One Otter-Banana area) that lies in the field of calc-alkaline andesites, persistently shows anomalous geochemistry with respect to other mafic rocks in the Manitouwadge belt. Hornblende-magnetite-garnet rocks and magnetite quartzites in the Dead Lake suite, and a single sample from the outer volcanic belt, have distinctive high FeO_t and TiO₂, and low MgO contents (Figs. 23, 24).

Felsic volcanic rocks and synvolcanic trondhjemite are transitional calc-alkaline to tholeiitic rhyolites and dacites (Fig. 23), mostly containing 65–80 percent SiO₂ (Fig. 24). In terms of Al-(Fe_t+Ti)-Mg, quartz-phyric felsic rocks show a somewhat wider compositional range that could be due to a mafic component in heterolithic breccias, and/or incipient alteration. The narrow compositional range of synvolcanic trondhjemite in the

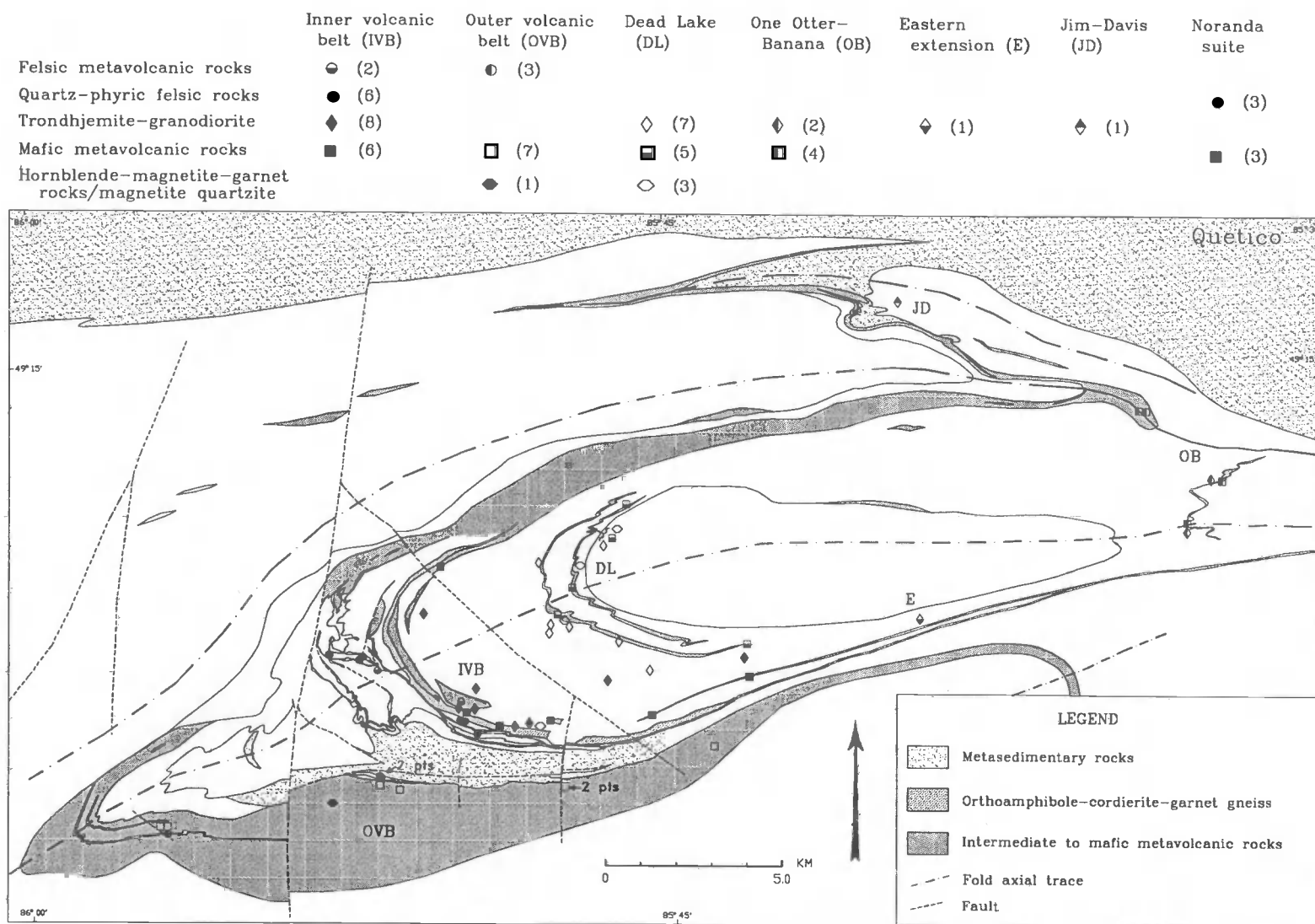


FIG. 22. Locations of metavolcanic and associated geochemical samples in the Manitouwadge greenstone belt. The key shows the symbols used to represent rock types and geographic areas, and the number of samples is in parentheses. Locations of samples analysed by Noranda Inc., all from the inner volcanic belt, are not shown. '2 pts' indicates 2 samples from the same location. The Swill-Mills Lakes area (see text) encompasses the 3 most westerly samples in the outer volcanic belt.

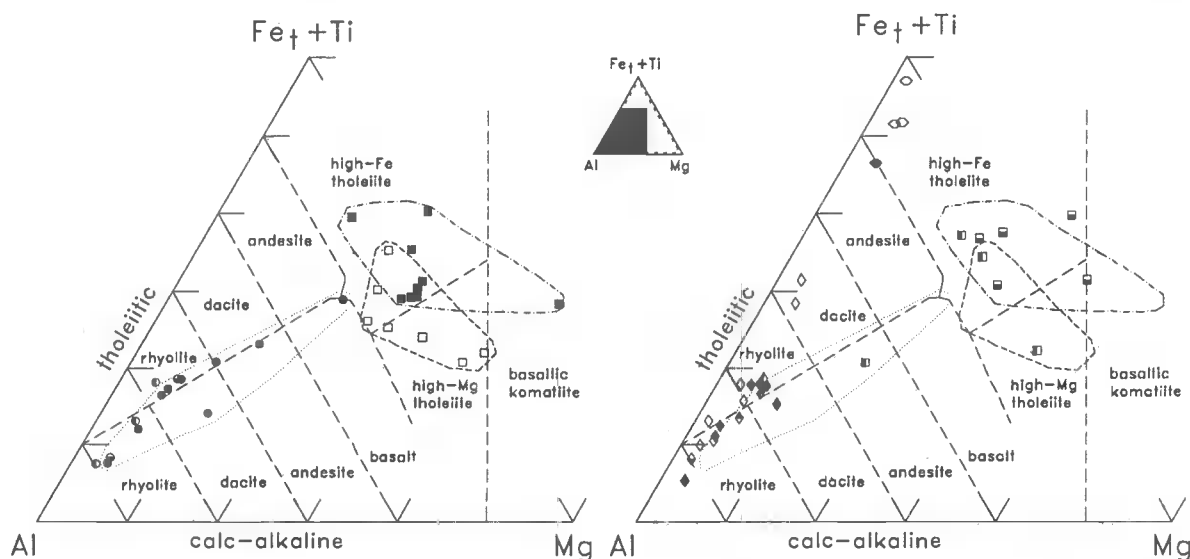


FIG. 23. Metavolcanic and associated rock compositions in terms of Jensen's (1976) definition of tholeiitic and calc-alkaline rhyolite to basalt, and basaltic komatiite, based on Al-(Fe_T+Ti)-Mg (cation %), elements less likely to be mobile during sea-floor alteration and regional metamorphism. Symbols as in Figure 22. The fields of quartz-phyric felsic rocks (dotted), and inner- (dash-dot) and outer-belt (dashed) mafic rocks are outlined as in Figure 24. The samples are divided between 2 plots to minimize overlap.

inner volcanic belt tends to encompass, or nearly so, analyses of similar-looking rocks from the Dead Lake, One Otter-Banana, and Jim-Davis Lakes areas, suggesting a comagmatic relationship. Some trondhjemite samples in the vicinity of the Dead Lake suite tend to have high FeO_T and TiO₂ contents, intermediate between that of hornblende-magnetite-garnet rocks/magnetite quartzites and inner-belt trondhjemite, consistent with contamination.

Mafic rocks in the inner volcanic belt can be distinguished from felsic rocks on the basis of TiO₂ and FeO_T content, with mafic rocks having TiO₂>1% and FeO_T>10%, and felsic rocks having TiO₂<0.4% and FeO_T<8% (Fig. 24). A single felsic sample with TiO₂=1.36% is a micaceous quartz-phyric heterolithic breccia. The majority of quartz-phyric felsic rocks have TiO₂ between 0.27 and 0.39%.

'Unaltered' felsic extrusive rocks in the Manitouwadge belt have undergone extensive alkali exchange, resulting in wide variations in K₂O, Na₂O and CaO content (Fig. 25). Quartz-phyric felsic rocks are predominantly potassic. In contrast, synvolcanic trondhjemite is sodic and has a relatively restricted range of alkalis and CaO, consistent with limited secondary redistribution. Trondhjemite samples near the Dead Lake suite extend to more calcic compositions, again suggesting contamination by hornblende-magnetite-garnet rocks/magnetite quartzites which group with mafic metavolcanic rocks.

Chondrite-normalized REE in mafic rocks in the inner and outer volcanic belts define patterns that are relatively flat or have slightly elevated light REE, with weak negative or positive Eu anomalies (Fig. 26). The inner belt has higher light REE levels (11–40×chondrite) than the outer belt (6–25×chondrite). The lowest light REE contents in the outer belt are in gabbroic rocks and samples from the Swill-Mills Lakes area, again possibly reflecting a sampling bias. Heavy REE levels in both belts are nearly identical.

In contrast to mafic rocks from the inner and outer volcanic belts, which show considerable compositional overlap, REE patterns define two distinct suites of felsic volcanic rocks. The total REE abundance in two samples of northernmost outer-belt aphyric felsic rocks is relatively low, and normalized REE define a steeply sloped pattern from light to heavy (normalized La/Yb = 8) and a moderately negative Eu anomaly (Eu/Eu* = 0.65–0.83) (Fig. 27). Zr/Y, which generally mimics the light to heavy REE slope (Leshner et al., 1986), is high at 32–44. REE and Zr/Y in northernmost outer-belt felsic rocks are typical of barren (not mineralized) felsic volcanic suites in the Wabigoon and Abitibi subprovinces (ibid.). A single sample of felsic breccia, interleaved with mafic rocks in the Swill-Mills Lakes area, has REE contents between those of felsic rocks in the inner and outer belts. Inner-belt felsic rocks, including both volcanic rocks and subvolcanic trondhjemite, are characterized by higher total REE, moderately sloping patterns (normalized La/Yb = 2.3–4.9), more

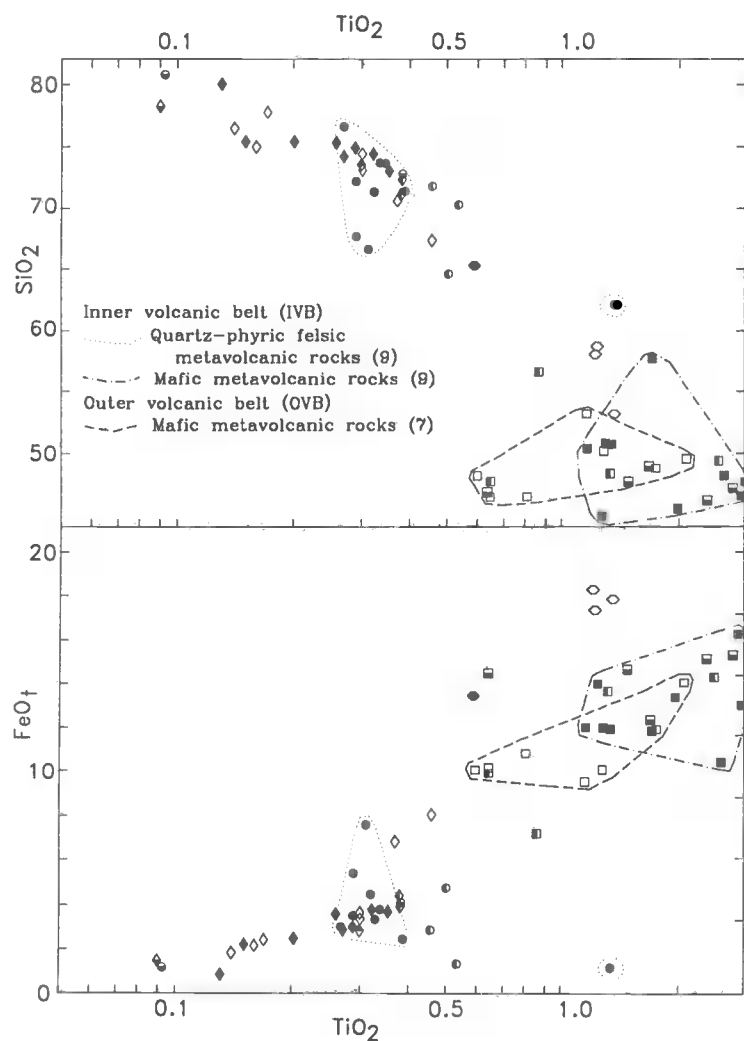
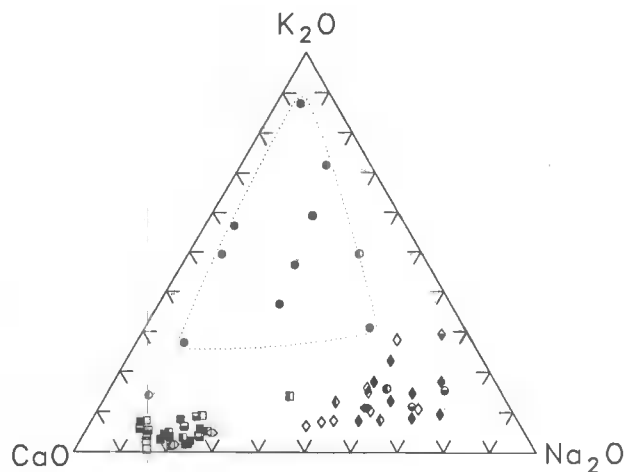


FIG. 24. SiO_2 and FeO_t (weight %) as functions of TiO_2 (weight %, logarithmic scale). Mafic and felsic rocks, especially in the inner volcanic belt, can be discriminated on the basis of TiO_2 abundance, a valuable observation as TiO_2 is relatively immobile during alteration and metamorphism. Symbols as in Figure 22. The outlined sample fields in this and other plots are intended as a visual aid, without statistical significance.

FIG. 25. $\text{CaO-K}_2\text{O-Na}_2\text{O}$ (weight %) proportions. The compositional variation of quartz-phyric felsic rocks (dotted field) suggests alkali exchange dominated by potassic alteration. Trondhjemite-granodiorite (symbols as in Fig. 22) has a relatively restricted range of sodic compositions. Hornblende-magnetite-garnet rocks and magnetite quartzites lie with calcic mafic rocks.



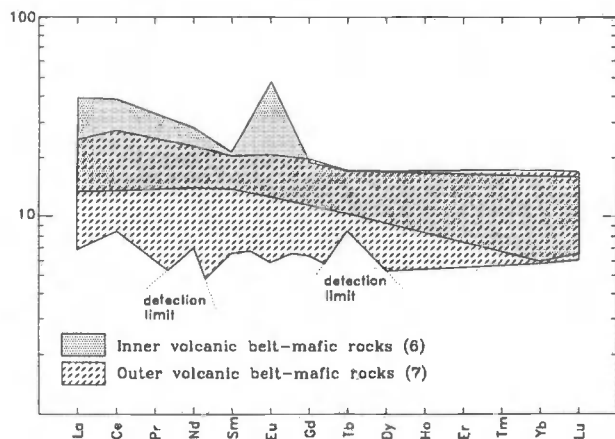
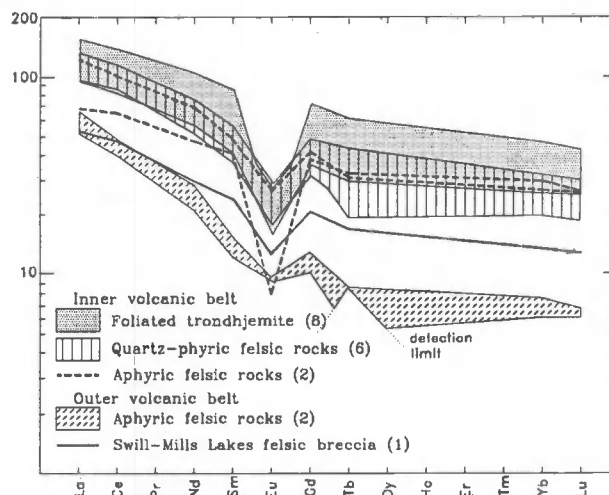


FIG. 27. Chondrite-normalized REE abundances in felsic metavolcanic and subvolcanic rocks. Normalization values and interpolated Gd as in Figure 26.

FIG. 26. Chondrite-normalized rare earth element (REE) abundances in mafic metavolcanic rocks. The positive Eu anomaly in the inner volcanic belt is defined by a single sample; all others have nearly flat or weakly negative Eu/Eu*. Gd was interpolated assuming a linear distribution between normalized Sm and Yb. (Tb was not used because of its relatively high detection limit). Normalization values are those of Taylor and McLennan (1985).



pronounced negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.20\text{--}0.58$), and Zr/Y from 1.7–11. Quartz-phyric felsic samples, collected from each of the three lenses in close proximity to massive sulphide deposits (Fig. 4), form a coherent group compositionally similar to synvolcanic trondhjemite (Fig. 27). On the basis of similarity in composition and age, the trondhjemite is interpreted to have been a reservoir for quartz-phyric felsic volcanism. In general, inner-belt felsic rocks are geochemically comparable to felsic volcanic suites associated with massive sulphide mineralization in the Wabigoon (Sturgeon Lake area), Abitibi and Uchi subprovinces (Leshner et al., 1986), in particular in the magnitude of negative Eu anomaly.

Altered rocks

In view of the unusual extent and concordance of altered rocks at Manitouwadge, especially orthoamphibole-bearing rocks, geochemistry was applied in order to define compositional trends associated with alteration, to test the consistency of our field identification of probable protoliths, and to compare compositions of altered and unaltered rocks with those reported from similar and dissimilar geological settings. In this discussion, 'alteration' and 'protolith' imply their premetamorphic equivalents and, although some element mobility is likely during metamorphism, this is presumed to have occurred on a local scale, not significant to the discussion here. Altered rocks are compared with unaltered or 'least' altered volcanic rocks sampled in the inner volcanic belt on the southern limb and in the hinge region of the Manitouwadge synform. Samples representing synvolcanic alteration, including orthoamphibole-cordierite-garnet gneiss and sillimanite-biotite-cordierite interlayers, as well as sillimanite-muscovite-quartz schist and associated quartzite, were collected in the Willroy-Geco area. The sample suite was supplemented by whole-rock major-element analyses made available by Noranda Inc., extending from the Willroy 2/5 deposit to about 1 km north of the axial trace of the Manitouwadge synform (Fig. 4). Orthoamphibole-bearing rocks and sillimanitic interlayers

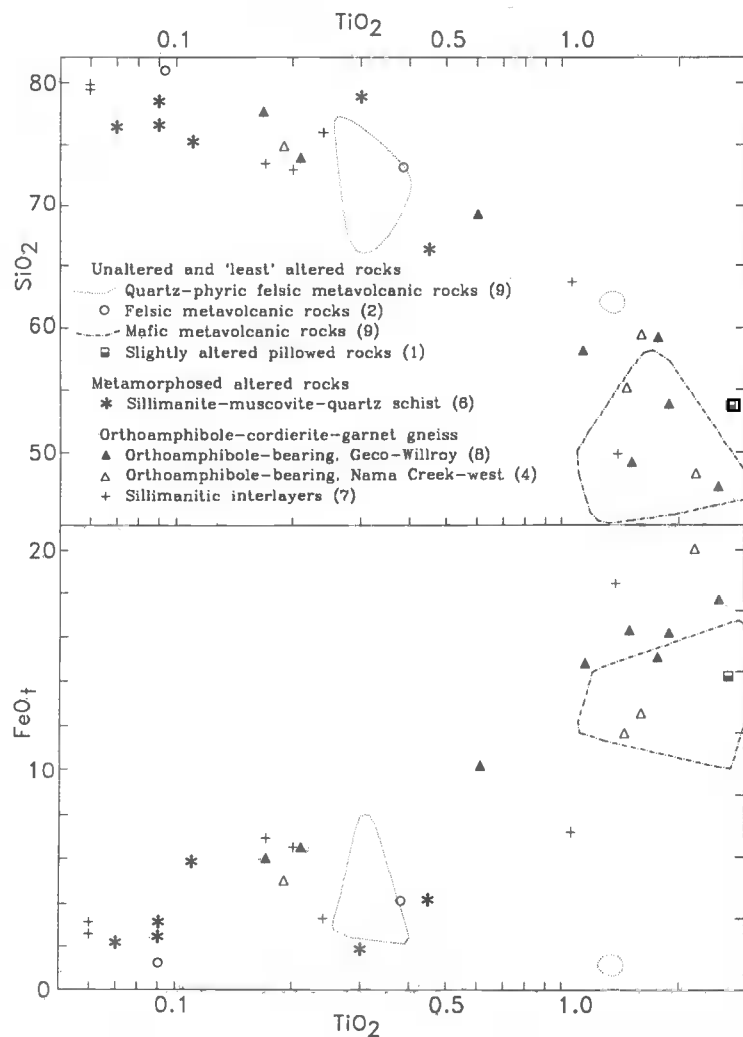


FIG. 28. SiO_2 and FeO_4 (weight %) as functions of TiO_2 (weight % logarithmic scale) for samples from the inner volcanic belt. TiO_2 abundance in unaltered or 'least' altered rocks defines a bimodal population corresponding to felsic and mafic rocks. The bimodal distribution is mimicked by altered rocks. The fields for quartz-phyric felsic rocks and mafic rocks are from Figure 24.

in the Willroy-Geco area are mostly intensely altered, whereas the area represented by the Noranda suite includes transitional or incipient alteration.

Both orthoamphibole-bearing rocks and sillimanitic interlayers have bimodal distributions of TiO_2 content similar to that of mafic and felsic rocks, suggesting that TiO_2 was relatively immobile during alteration and that, in most cases, the protolith of altered rocks can be inferred from TiO_2 abundance (Fig. 28). In low TiO_2 rocks ($<0.5\%$), TiO_2 shows a negative correlation with SiO_2 that could be interpreted as a differentiation trend. However, the high SiO_2 content (75–81%) of samples with $\text{TiO}_2 < 0.1\%$ suggests either TiO_2 loss and/or, silicic alteration accompanied by volume increase and dilution of TiO_2 . FeO_4 in mafic, felsic and altered rocks shows the same bimodal distribution as TiO_2 , and the FeO_4 content of altered rocks is generally similar to, or slightly higher, than that of 'least' altered rocks. These observations are consistent with field identification of protolith, that is, felsic volcanic rock for sillimanite-muscovite-quartz schist, and interlayered felsic and mafic rocks for orthoamphibole-cordierite-garnet gneiss. Interestingly, the distribution of orthoamphibole-bearing layers and sillimanitic interlayers in orthoamphibole-cordierite-garnet gneiss is apparently not determined by protolith.

The alteration trends can be summarized by viewing compositions of 'least' altered and altered rocks in an 'unfolded' $(\text{FeO}_4 + \text{MgO}) - (\text{Al}_2\text{O}_3/2) - (\text{K}_2\text{O}) - (\text{Na}_2\text{O} + \text{CaO})$ tetrahedron (Fig. 29) (Riverin and Hodgson, 1980; Luff et al., 1992). The mobility of alkalis in felsic rocks is apparent from their wide compositional range in projections showing $\text{K}_2\text{O} - (\text{Na}_2\text{O} + \text{CaO})$. Sillimanite-muscovite-quartz schist shows considerable overlap with felsic rocks, but also extends to proportionately more $(\text{FeO}_4 + \text{MgO})$ -rich and Al_2O_3 -rich compositions. 'Least' altered mafic rocks define a restricted compositional range. Orthoamphibole-cordierite-garnet gneiss has proportionately higher $(\text{FeO}_4 + \text{MgO})$ and Al_2O_3 , and lower $(\text{Na}_2\text{O} + \text{CaO})$, than 'least' altered rocks. Orthoamphibole-bearing rocks and sillimanitic interlayers are mainly different in $(\text{FeO}_4 + \text{MgO})/\text{Al}_2\text{O}_3$, the latter being slightly more aluminous. Transitional altered rocks, mainly from the area northwest of the

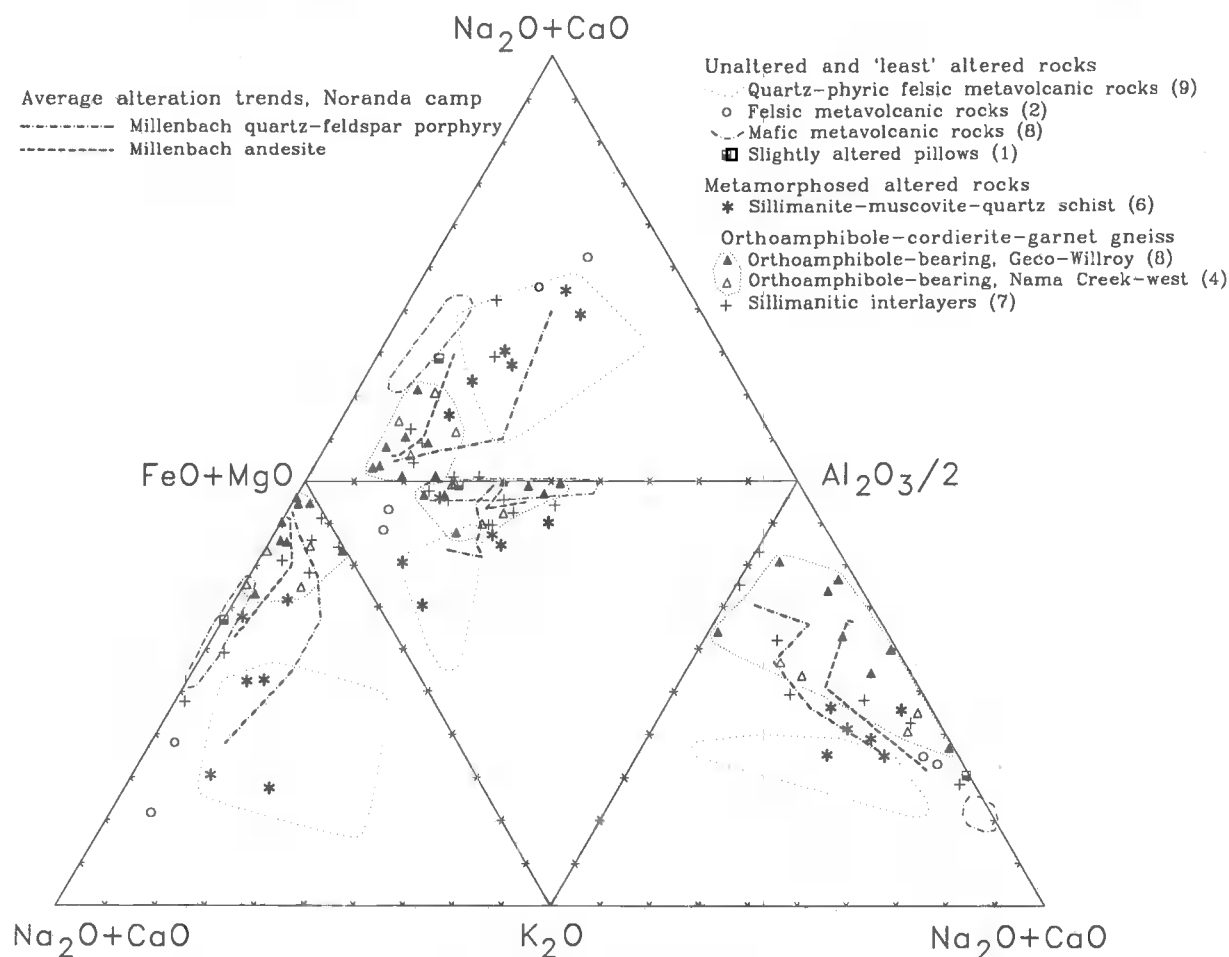


FIG. 29. Unfolded tetrahedron showing compositional trends (molar proportions) associated with unaltered, 'least' altered, and altered rocks. Orthoamphibole-bearing rocks from the Willroy-Geco area (filled triangles) are generally more intensely altered than those west of the Nama Creek deposit (open triangles). The fields of orthoamphibole-bearing rocks, quartz-phyric felsic rocks and mafic rocks are outlined. Average alteration trends for the Millenbach andesite (dashed) and quartz-feldspar porphyry (dash-dot) are shown for comparison (analyses from Table 3, Riverin and Hodgson, 1980). The form of the plot is based on Riverin and Hodgson (1980) and Luff et al. (1992).

Nama Creek deposit, tend to have compositions between more altered Willroy-Geco rocks and 'least' altered rocks, in some cases, overlapping with 'least' altered rocks. Alteration trends for the Millenbach andesite and quartz-feldspar porphyritic rhyolite associated with alteration pipes in the Millenbach mine (analyses from Table 3 of Riverin and Hodgson, 1980) are shown for comparison. Although alteration zones at Millenbach are much smaller in extent (100's of metres), compositional trends generally correspond to those of altered and 'least' altered rocks at Manitouwadge. The most intensely altered felsic rocks at both Manitouwadge and Millenbach are considerably higher in $(\text{FeO}_t + \text{MgO})$ and Al_2O_3 than chloritic and sericitic altered quartz-feldspar pyroclastic rocks from the Brunswick No. 12 deposit of the Bathurst camp (compare Fig. 15 of Luff et al., 1992).

In terms of AFM (Fig. 30), the compositions of orthoamphibole-bearing rocks from Manitouwadge mostly lie in the field defined by cordierite-anthophyllite rocks derived from altered volcanic precursors, as compiled by Reinhardt (1987). Reinhardt's field of evaporitic clays and magnesian pelites, which encompasses cordierite-anthophyllite metasedimentary rocks from the Rosebud syncline, Australia is restricted to $\text{MgO}/(\text{MgO} + \text{FeO}) > 0.7$. Cordierite-anthophyllite rocks from the Hemlo greenstone belt, interpreted to be the metamorphic equivalent of immature clastic sedimentary rocks derived from a mafic-ultramafic source (Pan et al., 1991), are predictably indistinguishable from altered volcanic rocks.

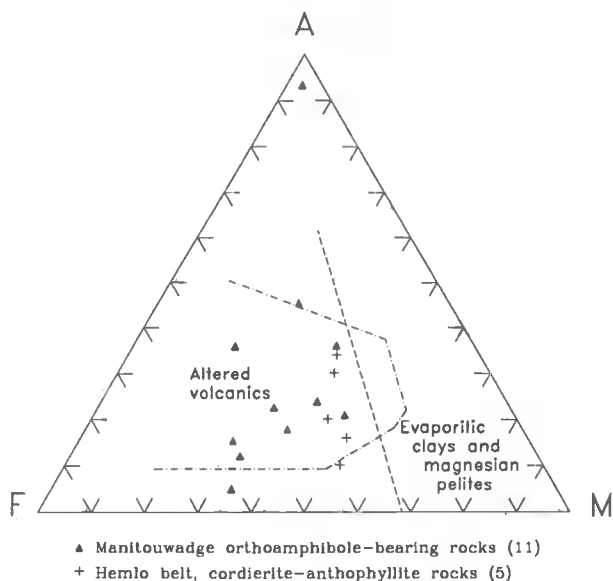


FIG. 30. AFM plot of orthoamphibole-bearing altered rocks at Manitouwadge. An additional sample (not shown) lies at negative FM proportions. The AFM values (molar proportions) are calculated after the method of Reinhardt (1987) in which;

$$A = \text{Al}_2\text{O}_3 - (\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}),$$

$$F = \text{FeO}_t - 6(1-m)\text{K}_2\text{O},$$

$$M = \text{MgO} - 6m\text{K}_2\text{O}, \text{ and}$$

$$m = \text{MgO} / (\text{MgO} + \text{FeO}_t)_{\text{whole rock}}.$$

The calculation scheme attempts to 'correct' for oxides present in biotite and feldspars, assuming that $\text{MgO}/(\text{MgO} + \text{FeO}_t)$ of biotite can be approximated by the value of the whole rock. Reinhardt's fields of altered volcanics, and evaporitic clays and magnesian pelites are shown for comparison, as are five analyses of cordierite-anthophyllite metasedimentary rocks from the Schreiber-Hemlo greenstone belt (from Table 2, Pan et al., 1991).

DISCUSSION

Depositional setting and deformation of massive sulphide deposits and alteration zones

When the presence of the D_1 fault that dissects the Willecho-Geco area is taken into account (Figs. 4 and 21), alteration zones and orebody types at Manitouwadge have systematic relationships to stratigraphic level, despite remaining ambiguities about detailed correlations of some mineralized horizons. Cu stringer and disseminated orebodies (Willroy 1 and 6, and Geco 4/2 zones, Table 3) lie in orthoamphibole-cordierite-garnet gneiss of the footwall alteration zone. The enveloping Geco main stringer zone lies higher in the stratigraphy, linking the largest main orebody, and possibly the 8/2 Zn zone, with footwall alteration. These zones have features characteristic of subsurface conduits for mineralizing fluids and may mark the locations of focussed cross-stratal flow that supplied deposits higher in the stratigraphy. Zn-Cu-(Pb) orebodies associated with iron formation (Willroy 2, 3 and 5, Geco main orebody and 8/2 Zn zone, Nama Creek and Willecho 1-3) are sea-floor precipitates or near-sea-floor replacement deposits. This group, representing voluminous base-metal precipitation and the peak of hydrothermal activity, includes the largest deposits, the Willroy 3 and Geco main orebodies. The Cu-to-Zn zoning of the Willroy 3 orebody toward the hosting iron formation, and the stringer zones associated with the Geco main orebody, are suggestive of proximal deposits centred on areas of hydrothermal upflow. In other Zn-Cu-(Pb) orebodies, possible primary features, such as layering or zoning, have not been reported. However, comparison with metal ratios in the Millenbach and Amulet orebodies in the Noranda camp (Knuckey et al., 1982), suggests that their higher Zn/Cu is a feature of deposition further from the main area of hydrothermal venting. The iron-formation-hosted Zn-Pb-(Cu) orebodies (Willroy 4 and Geco zincian iron formation) are highest in the stratigraphy and formed during the waning of hydrothermal activity.

A sedimentary association for the deposits of Manitouwadge belt has been previously emphasized, largely based on the assumption that the Manitouwadge greywackes represent a transition from volcanism to clastic sedimentation, and on the presence of iron formation and quartzites near the contact (Franklin et al., 1981; Friesen et al. 1982; Williams et al., 1990). Detrital zircon ages show that the greywacke is considerably younger than the underlying volcanic rocks, and that the contact is either an unconformity or a fault. The quartzite is a silicified felsic volcanic rock, locally gradational to sillimanite-muscovite-quartz schist (with which it is grouped as a map unit) and iron formation. We regard all of these rocks as the products of synvolcanic hydrothermal activity, either through alteration or through precipitation from hydrothermal fluids.

'Least' altered felsic volcanic rocks at Manitouwadge have all undergone alkali exchange, dominated by enrichment in K and with little effect on other elements, typical of low temperature sea-floor alteration (Lagerblad and Gorbatshev, 1985). Synvolcanic trondhjemite was not affected by widespread alkali exchange, despite the local Fe-Mg alteration near contacts, as inferred from orthoamphibole-garnet seams. Similar observations of alkali exchange in the Abitibi camp, involving the sodic Flavrian pluton and its more potassic comagmatic volcanic rocks, were interpreted as the result of sea-floor potassic alteration of extrusive rocks, and preservation of primary compositions in the subvolcanic body (Goldie, 1979).

The lateral distribution and geochemistry of sillimanite-muscovite-quartz schist suggest that these rocks

interacted with hydrothermal fluids and are not simply a more extreme example of sea-floor alkali exchange. In the Manitouwadge camp, they have been regarded as a favourable exploration tool indicating proximity to ore. This is difficult to reconcile with the stratigraphic position of sillimanite-muscovite-quartz schist in the hanging wall to orebodies, and sandwiched between semi-continuous layers of iron formation. Locally, the distribution of thin sillimanitic layers in tuffaceous rocks was apparently governed by bedding and is most consistent with sea-floor alteration. The Geco, Willroy 1 and 6 areas, interpreted as zones of probable hydrothermal upflow, are an exception in that sillimanite-muscovite-quartz schist directly overlies orthoamphibole-cordierite-garnet footwall alteration. The observations suggest lateral percolation of hydrothermal fluids from this area, either in the subsurface or on the sea floor or both. Geochemically and mineralogically, the schist resembles the muscovite-aluminosilicate rocks in the Snow Lake massive sulphide camp (Walford and Franklin, 1982; Zaleski, 1989) and sericitic alteration capping and peripheral to chloritic pipes of Noranda-type deposits (Franklin et al., 1981), interpreted as representing the outer cooler parts of a hydrothermal system.

It is generally accepted that subvolcanic intrusions provide the necessary heat for driving hydrothermal systems, and that these commonly intrude their own volcanic deposits (Campbell et al., 1981). In the Snow Lake camp, synvolcanic plutons both intrude altered rocks produced by their hydrothermal system, and show evidence of alteration; however, the distribution of alteration zones only crudely follows intrusive contacts (Fig. 11 in Galley, 1993). The very close spatial relationship between the trondhjemite contact and stratabound alteration at Manitouwadge is unusual, and the stratigraphic control is perhaps even more pronounced when the unaltered or weakly altered character of supracrustal inclusions in trondhjemite is taken into account. The semi-concordant contact and supracrustal septa suggest a sill or laccolith, or multiple semi-concordant intrusions, with alteration focussed along the upper contact. It is also possible that the trondhjemite invaded deeper parts of the hydrothermal system and that surviving inclusions of altered rocks are not exposed at the surface.

Regional semi-conformable alteration zones associated with volcanogenic massive sulphide deposits typically involve alkali-exchange, silicification and epidotization (Galley, 1993), rather than alkali-depleted Fe-Mg alteration. Despite the unusual extent and concordance of alteration zones at Manitouwadge, the geochemical trends from 'least-altered' to intensely altered, of increase in $\text{FeO}_4 + \text{MgO}$ and Al_2O_3 and depletion in alkalis and CaO, are similar to those recorded in alteration pipes in the Abitibi camp (Riverin and Hodgson, 1980). The bimodal grouping of altered rocks, defined by 'immobile' element abundances, mimics mafic and felsic volcanic precursors. Orthoamphibole-garnet-cordierite rocks and sillimanitic interlayers had both mafic and felsic protoliths and the metamorphic assemblage was determined by small variations in $(\text{FeO}_4 + \text{MgO})/\text{Al}_2\text{O}_3$, apparently due to alteration and not inheritance from the protolith. There is no evidence to suggest the presence of pelitic rocks, although some layering in altered rocks could reflect reworking and local redeposition of unconsolidated volcanoclastic material.

The unusual extent of orthoamphibole-bearing gneiss at Manitouwadge may be partly due to the high metamorphic grade. With increasing grade, orthoamphibole-hornblende stability expands the range of orthoamphibole-bearing assemblages to more calcic bulk-rock compositions (Spear, 1993, pages 478–489). By inference, bulk-rock compositions that produced orthoamphibole-bearing rocks at Manitouwadge might, at a lower metamorphic grade, be considered incipient alteration. In the greenschist- and amphibolite-facies metavolcanic rocks of the Snow Lake area, semi-continuous zones of chloritic alteration have a lateral extent of up to two to three kilometres (Galley, 1993). The closest analogues for the stratabound regional orthoamphibole-bearing alteration at Manitouwadge are found in the Bergslagen area of the Svecofennian Baltic shield. At Bergslagen, exhalative base-metal deposits and iron formation are underlain by conformable stratabound Mg-rich alteration zones with a regional extent of several, to several tens of kilometres (Trägårdh, 1988; Ripa, 1988; Baker et al., 1988). Amphibolite-facies metamorphism at Bergslagen may play a role, as at Manitouwadge, in facilitating identification of altered rocks. However, in both areas, it appears that alteration was partly focussed on aquifer horizons possibly consisting of permeable, poorly consolidated volcanoclastic deposits.

Structural and tectonic synthesis

Our preferred structural model for the Manitouwadge greenstone belt, and the adjacent Quetico sub-province, involves four phases of ductile deformation. D_1 faults (shear zones) are recognized from the coincidence of truncated lithological units, repeated mineralized sequences, and zones of straight gneiss interpreted as annealed mylonite. Although no sense of kinematics or offset has been observed, sequence repetitions, geometries consistent with low angle truncation, and the early relative age of D_1 structures, suggest thrusting. Dominant D_2 planar and linear fabrics, typically defined by high grade metamorphic minerals, suggest D_2 deformation broadly synchronous with peak metamorphism. D_2 shortening resulted in repetition of the volcanic sequence across the easterly trending 'Manitouwadge syncline' on the southern limb of the D_3 Manitouwadge synform, and possibly accounts for the presence of volcanic rocks in the Dead Lake and One Otter-Banana Lakes areas. Metagreywacke in the core of the 'Manitouwadge syncline' is folded and contains D_2 fabrics.

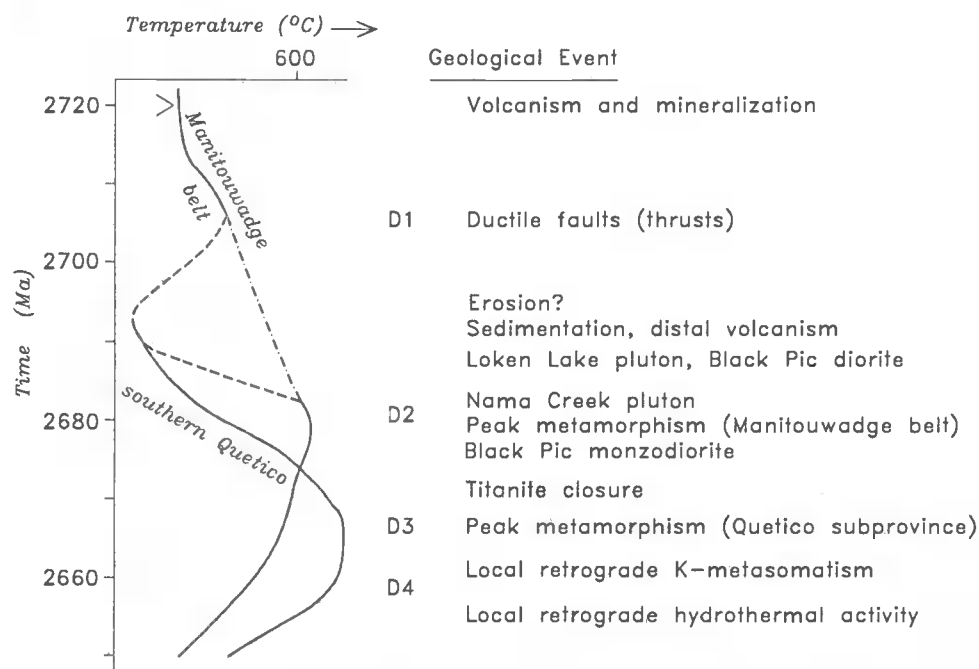


FIG. 31. Temperature-time plot showing relationships between the deformation sequence, metamorphism and plutonism in the Manitouwadge belt and Quetico subprovince. Alternative paths are shown for the Manitouwadge belt assuming either, unconformable deposition of sedimentary rocks (dashed) or, emplacement by early faulting (dash-dot).

The D₃ Manitouwadge synform, Blackman Lake antiform, and Jim Lake synform, fold the Manitouwadge belt and the Wawa-Quetico subprovince boundary. Relationships between metamorphic minerals, migmatitic segregations, and deformation fabrics suggest that D₃ was broadly coeval with peak metamorphism in the Quetico subprovince. Map-scale D₄ structures modify the geometry of D₃ folds. The dominantly Z-asymmetry of the D₃/D₄ folds, and dextral kinematic indicators, are interpreted as a response to progressive dextral transpression. In a preliminary structural model (Peterson and Zaleski, 1994a), we proposed a 5-phase deformation sequence for the Manitouwadge area. The preliminary model interpreted the Blackman Lake antiform and Jim Lake synform as D₄ structures that folded the axial trace of the D₃ Manitouwadge synform. Although our observations still permit the 5-phase model, it required a more complicated deformation scheme than our preferred simpler 4-phase model.

The Wawa-Quetico boundary in the Manitouwadge area is transitional on structural and lithological criteria. Firstly, there is a gradational change to the south, from dominantly east-west structural trends, to map-scale folds with broader hinge regions; secondly, Manitouwadge and Quetico metasedimentary rocks are indistinguishable. Zircon provenance ages constrain the maximum depositional age of Manitouwadge metagreywacke to 2693 Ma, at least 25 Ma younger than the 2720 Ma felsic volcanism (Zaleski et al., 1994; Davis et al., 1994). The metagreywacke contains upper amphibolite-facies assemblages and a D₂ tectonic fabric, hence the minimum age of deposition can be inferred to be circa 2680 Ma (see below). Provenance studies on Quetico metasedimentary rocks have established age brackets of circa 2700 to 2688 Ma (Percival and Sullivan, 1988; Davis et al., 1990), overlapping the age brackets at Manitouwadge. In contrast to the detrital zircons of Mesoarchean age recovered from Quetico rocks (ibid.), at Manitouwadge, the oldest detrital zircon of 2719±2 Ma could be derived from local volcanic rocks. The Quetico studies were done at least 200 km west of Manitouwadge, and both source areas and timing of sedimentation could be expected to vary.

In the Manitouwadge belt, uplift resulting from D₁ deformation could have contributed sediment sources, although our field observations are equivocal regarding the relationship between D₁ and sedimentation. The presence of straight gneiss (annealed mylonite) on D₁ thrust faults indicates ductile deformation and somewhat elevated temperatures. Two post-D₁ pre-D₂ time-temperature paths are possible for the Manitouwadge belt, one assuming erosion and unconformable deposition of sedimentary rocks, the other consistent with tectonic emplacement of an allochthonous sequence by pre- or syn-D₂ faulting (Fig. 31).

A major episode of magmatic activity from 2687 to 2677 Ma is indicated by the age of four plutonic

rocks; Black Pic diorite ($2687 \pm 3/-2$ Ma), the Loken Lake pluton ($2687 \pm 2/-3$ Ma), the Nama Creek pluton (2680 ± 3 Ma) and Black Pic monzodiorite (2677 ± 2 Ma). The first three of these are pre- to syn-D₂ intrusions, and hence, the maximum age of D₂ deformation and contemporaneous peak metamorphism is constrained to 2680 Ma. Magmatic activity could have contributed to metamorphic heating.

Peak metamorphic temperatures of 600–700°C in the belt (Petersen, 1984; Pan and Fleet, 1992) are close to the 700°C closure temperature of the U-Pb system in monazite, and exceeded the 600°C closure temperature of titanite (Heaman and Parrish, 1991). Metamorphic monazites from metavolcanic and altered rocks range in age from 2675–2669 Ma (Schandl et al., 1991; Davis et al., 1994; Zaleski et al., 1995), and encompass the circa 2673 Ma titanite ages from the Nama Creek pluton and Black Pic monzodiorite (Table 2). We tentatively interpret the titanite ages as the time of regional cooling through 600°C (Fig. 31); however, it is possible that the metamorphic fluids that allowed the crystallization of monazite were also responsible for crystallization or resetting of titanite. Field observations indicate that peak metamorphism and migmatization in the Quetico subprovince was synchronous with progressive deformation and the map-scale folds (D₃) near the boundary (Peterson and Zaleski, 1994a). Peak metamorphism in the Quetico subprovince could have been coeval with granitic magmatism at 2670–2650 Ma (Percival, 1989). The regional distribution of metamorphic ages and assemblages is consistent with progressively later development of peak conditions, and increasing grade, from south to north; from 2676–2678 Ma amphibolite-facies metamorphism in the Schreiber-Hemlo greenstone belt (Corfu and Muir, 1989b), to 2680 Ma upper amphibolite-facies metamorphism in the Manitouwadge belt, to 2670–2650 Ma granulite-facies metamorphism in the Quetico subprovince (Fig. 2).

A transitional boundary between volcano-plutonic and metasedimentary subprovinces of the Superior Province is not unique to the Manitouwadge area. Similar relationships have been described between the Quetico subprovince and Coutchiching metasedimentary rocks in the Rainy Lake area of the Wabigoon subprovince (Davis et al., 1989), and Kehlenbeck (1985) considered the Beardmore-Geraldton belt to be a structural and lithological transition zone between the Quetico and Wabigoon subprovinces (Fig. 1). Kehlenbeck's conclusion that the Beardmore-Geraldton belt developed through deformation involving the margins of both subprovinces is equally applicable to the Wawa-Quetico boundary in the Manitouwadge area.

Studies along the Wawa-Quetico subprovince boundary indicate that it varies in character. To the west of Manitouwadge, rocks of volcanic and sedimentary origin are mostly in fault contact, although in some areas, earlier structures can be correlated across the boundary (Percival, 1989). To the east, between the Moshkinabi belt (Fig. 2) and the Lepage fault zone (Fig. 1), sedimentary and volcanic rocks are apparently interbedded along the subprovince boundary (Berger, 1985). The variation along-strike suggests that the subprovince boundary in the Manitouwadge area preserves features of early ductile deformation that may represent a lateral transition between the conformable contact to the east and fault juxtaposition to the west.

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FIELD-TRIP STOPS

Introduction

The field trips in this guidebook are organized thematically and geographically, to represent the geology of the Manitouwadge area, and give examples of the outcrop observations critical to our structural, stratigraphic and petrological interpretations. The guidebook is partly intended for the field trip to the Manitouwadge greenstone belt sponsored by the Forty-first Annual Meeting of the Institute on Lake Superior Geology and, in the longer term, as a series of self-guided tours that can be easily followed by anyone interested. A short introductory section summarizes the geology and significance of each area. For more detail and regional synthesis, the reader is referred to the interim report that comprises the first part of this volume.

A. Known economic deposits, inner volcanic belt

All of the known economic massive sulphide deposits of the Manitouwadge greenstone belt, comprising the Geco, Willroy, Nama Creek and Willecho deposits, lie in the inner volcanic belt of the southern limb of the D₃ Manitouwadge synform (1:25000 map). The Geco mine, scheduled to close in late 1995, is the only currently producing deposit. In the case of mined-out orebodies, some spatial relationships can be inferred from the positions of open stopes and pits with respect to the hosting metavolcanic rocks, iron formation and metamorphosed alteration zones along- and across-strike. Detailed 1:5000 mapping (1991-92) in the area of known deposits was critical to our definition of lithological units, and interpretation of pre-D₃ deformation and probable depositional relationships. To the north, supracrustal rocks are bounded by a synvolcanic trondhjemite which also engulfs screens of mafic and interlayered mafic-felsic metavolcanic rocks. The trondhjemite is mantled by orthoamphibole-garnet-cordierite gneiss, interpreted as the metamorphosed equivalent of a synvolcanic hydrothermal alteration zone, developed in a protolith of intercalated mafic-felsic rocks. To the south of orthoamphibole-garnet-cordierite gneiss, felsic metavolcanic rocks and mineralized iron formation are interlayered with sillimanite-muscovite-quartz schist representing metamorphosed synvolcanically altered felsic rocks. The volcanic sequence is repeated in the outer volcanic belt by the D₂ 'Manitouwadge syncline', the axial trace of which lies in the metasedimentary rocks central to the southern limb of the D₃ Manitouwadge synform (Fig. 5). Stratigraphic younging in the Willroy-Geco area is southerly, based on generally southward increase in Pb/Zn and Zn/Cu of orebodies, and on the position of orthoamphibole-cordierite-garnet gneiss (footwall-type alteration) and synvolcanic trondhjemite to the north.

The area of the inner hinge region and southern limb of the Manitouwadge synform (D₃) preserves the best map- and outcrop-scale evidence of the pre-D₃ structural history (see Structural complications—D₁/D₂ folds and faults). The near-continuous layers of iron formation are excellent marker units, useful in linking correlative stratigraphic horizons and outlining structural features. In our structural interpretation, a D₁ ductile fault divides the area into upper (southern) and lower (northern) tectonic blocks and is responsible for the repetition of mineralized horizons. Between the Nama Creek and Willecho deposits, the D₁ fault surface is deformed by a map-scale D₂ sheath fold (Fig. 4).

The Geco-Willecho area has many hazards associated with previous mining operations. Permission for access, via a locked gate across the Willroy mine road, must be arranged in advance with Noranda Inc. Many stops in this area were previously described in the guidebook for the International Association on the Genesis of Ore Deposits (IAGOD) (Williams et al., 1990), and IAGOD stop numbers are given here to facilitate comparison with our descriptions and reinterpretations. Our station numbers with prefix 'ZB' are also given for each stop as a cross-reference to sample and analytical data in preparation for open file release.

A1-A8. Willroy-Geco area

The following field-trip stops (1:25000 map) generally proceed from younger to older (and deeper in the section) rocks, from the metagreywackes central to the D₂ 'Manitouwadge syncline', to iron formation interlayered with felsic volcanic rocks, to sillimanite-muscovite-quartz schist and orthoamphibole-cordierite-garnet gneiss, to subvolcanic trondhjemite. The general sequence was interpreted as a stratigraphic succession by Suffel et al. (1971), based on comparison with successions typical of other greenstone belts. Suffel et al. were also the first to interpret the Manitouwadge deposits as volcanogenic massive sulphides, and orthoamphibole-bearing rocks as metamorphosed footwall alteration.

Stop A1, Manitouwadge metagreywacke, ZB91-10, ZB93-410, ZB94-87. Starting at the locked gate on the Willroy mine road, drive north about 300 metres up the short hill, turning west (left) on the gravel track near the hill top. About 200 metres more brings you to a parking area on the east side of a small dam across the Slim Lake valley. Cross the foot bridge and rock dam to an outcrop on the west side. The dam contains tailings from the Willroy deposits that were dumped into Slim Lake (now mostly a meadow) from 1950's to 1970's. Lime is periodically pumped into the ponded water to neutralize acidity and precipitate metals. The north-south lineament is the topographic expression of the Slim Lake fault, a high-angle brittle fault with minor dextral offset.

The Manitouwadge metasedimentary rocks comprise biotite±garnet±sillimanite schist and biotite-hornblende schist, mostly homogeneous, but also with layering and grading defined by the abundance of mafic minerals. Metagreywackes in this exposure are thinly (2 mm–2 cm) to thickly (10 cm–2 m) layered (transposed bedding) and contain plagioclase- and quartz-crystal clasts, evidence of a tuffaceous component. Some layers have possible lithic fragments (1–10 cm long) with diffuse margins. The layering, and the sub-parallel foliation (D_2), are near-vertical and straight, and a strong mineral lineation plunges easterly. Locally developed intrafolial folds, some eye-shaped, probably originated as soft-sediment folds that were subsequently transposed during further tectonic deformation.

Based on U-Pb provenance study of detrital zircon from this site (Fig. 12), the maximum age of deposition is 2693 Ma, at least 25 Ma younger than felsic volcanism (circa 2720 Ma) in both the inner and outer volcanic belts. The involvement of metagreywacke in D_2 folds indicates that sedimentation pre-dated D_2 and that the contact is either an unconformity or a pre- to syn- D_2 fault. Field observations are equivocal regarding the relationship of D_1 deformation and sedimentation; however, uplift resulting from D_1 deformation may have contributed sediment sources. We interpret the Manitouwadge metagreywackes as a tectonic outlier of Quetico rocks.

Foliated tonalite dykes (25 cm–2 m in width) intrude metagreywacke at a low oblique angles to layering, and some show stretching and boudinage related to layer-parallel shear. One of these dykes was collected for geochronology to constrain to the minimum age of sedimentation (analysis in progress). A nearly massive muscovite-pegmatite cross-cuts foliation but has concordant aplitic apophyses; the pegmatite shows thickening in the hinge region of a minor fold.

Stop A2, Iron formation, ZB91-11, ZB92-P18, IAGOD #2. Return to the Willroy mine road and continue 200 metres north, turning west at another gravel track. Continue about 1.5 km to a microwave tower and a large outcrop area on the hill top. The space for parking and turning around is small, soft and sandy. From the rocky crest, looking to the east, the Geco headframe and the open cuts of Willroy deposits are visible, situated in a narrow interval of iron formation, quartz-phyric felsic rocks and sillimanite-muscovite-quartz schist. The outcrop on which you stand is part of the thick southernmost quartz-magnetite iron formation, apparently unmineralized in the Willroy area, and possibly correlative with zincian iron formation at the Geco mine.

The iron formation is layered (1 mm–10 cm) with alternating white to smoky quartzose layers and dark layers rich in magnetite and Fe-silicates. The layering is tightly folded (D_2 or combined D_2 /transposed D_1 ?) about easterly-plunging axes, and fold asymmetry varies across the outcrop area. The mineral lineation is parallel to fold axes and, locally, grunerite defines an axial planar foliation. The southernmost iron formation has been thickened by folding, and we interpret these outcrop-scale folds as related to map-scale D_2 folds of the iron formation/felsic contact to the west (Stop A23, Fig. A3).

The metamorphic assemblage in dark layers consists of magnetite, grunerite with epitaxial overgrowths of green amphibole (ferroactinolite–ferrohornblende), clinopyroxene (ferrosalite), and minor pyrrhotite and garnet (Alm/Sps/Grs/Pyp = 58/22/17/2). Whole-rock geochemical analyses of 3 samples containing variable proportions of quartzose and dark material, have SiO_2 from 68.2–90.7%, FeO from 5.6–26.0%, MnO from 0.3–1.3%, CaO from 0.70–2.25%, and $\text{TiO}_2 < 0.02\%$, $\text{Al}_2\text{O}_3 < 0.53\%$, $\text{Na}_2\text{O} < 0.08\%$ and $\text{K}_2\text{O} < 0.08\%$, suggesting negligible detrital or volcanic component. The base metal content is also very low; Cu < 8 ppm, Pb < 9 ppm and Zn < 28 ppm.

The iron formation is cut by foliated dykes (30 cm–1.5 m in width) including granodiorite with mafic clots and plagioclase-porphyritic tonalite-granodiorite. The dykes are generally subparallel to the axial traces of D_2 folds but, locally show asymmetric folds with very long limbs (+10 m). In fold hinges, the foliation in the dykes is discordant to dyke margins and axial planar to folds. The dykes likely belong to a suite of tonalite dykes, some of which show structural relationships suggesting syn- D_2 emplacement. A discordant pegmatite cuts both iron formation and dykes.

Stop A3, Quartz-phyric felsic rocks and iron formation near the Willroy 2, 3, 4 and 5 orebodies, ZB91-12-14, ZB92-P139, ZB92-P152–P153. Return to the Willroy mine road and proceed about 800 metres north to a large gravelled area with some decrepit core racks, all that remains of Willroy Mines #1 and #2 shafts and millsite (Fig. A1). On the south side of the gravelled clearing, a chain-link fence restricts access to the open stope of the Willroy 3 orebody (1:25000 map, Fig. A1). The road continues to the northwest side of the clearing, and down a hill toward an abandoned railway linking the Willroy and Geco properties. On the hill-top plateau to the east of the road, another chain-link fence restricts access to dangerous ground immediately above the Willroy 4, 2 and 5 orebodies. Within 50 metres on both sides of the road, there are several pavement outcrops of quartz-phyric felsic breccias and thin interleaved iron formation (mostly too thin to show on the 1:25000 map).

The more southerly outcrops consist of felsic fragmental rocks in which both matrix and fragments (<1 cm–10 cm) have quartz phenocrysts. The variation in clast lithology is not dramatic and might be attributed

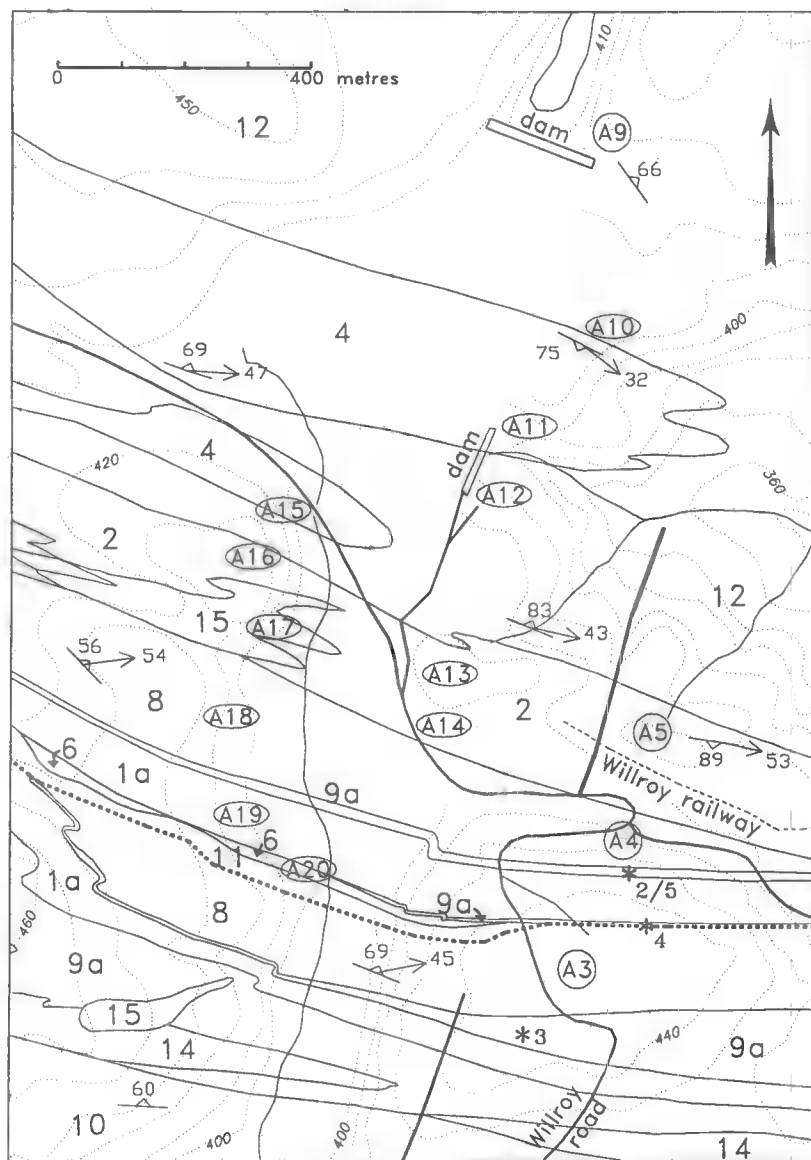


FIG. A1. Geology and field-trip stops of the Slim Lake section (A9–A20), and Willroy Stops A3–A5. Structure symbols show dominant D₂ foliations and lineations, and asterisks are the locations of the Willroy 2 to 5 orebodies. In this and the following figures, the figure locations are shown on the accompanying 1:25000 map and lithological units are shown by bold numbers that match the map legend. Circled alpha-numerics refer to field-trip stops. Dotted lines are topographic contours at 10 metre intervals.

to variable alteration of crystalline and glassy materials in a volcanoclastic deposit; but some biotite-rich streaks could represent more mafic clasts. Microcline and muscovite are both heterogeneously distributed, even on the scale of a thin section, suggesting that the K₂O abundance was determined by alkali exchange rather than primary magmatic composition. Magnetite porphyroblasts are ubiquitous and garnet is present locally. These breccias are part of the southernmost lens of quartz-phyric metavolcanic rocks in the Willroy area.

In scattered outcrops on the north side of the plateau, semi-continuous and discontinuous thin iron formations intercalated with the felsic breccia are difficult to trace laterally. A D₁ ductile fault is interpreted to pass through this area, following the northern contact of quartz-phyric felsic rocks (not exposed). The felsic rocks to the south are moderately muscovitic and contain quartz eyes and quartz pods up to 5–10 cm long. In the ditch west of the road, a small (2 m²) outcrop of iron formation in the first bushes at the top of the north-facing slope shows two types of iron formation; firstly, a breccia of quartzose fragments in a dark matrix and, secondly, homogeneous silicate iron formation. The iron formation lies on the north side of the fault, on or near the same horizon as the Zn-Pb-rich Willroy 4 orebody (Fig. 21, Table 3). Two geochemical samples from the outcrop showed elevated Zn (32 and 68 ppm), and low Cu (4 and 13 ppm) and Pb (7 and 11 ppm) abundances.

Between the road and the chain-link fence to the east, sulphidic iron formation crops out along-strike from the Willroy 2/5 orebodies. This is one of the few exposures of a transition from quartz-magnetite iron formation to sulphidic iron formation associated with an orebody. Two geochemical samples, separated by 5

metres of strike-length, show elevated levels of base metals. The more distant (western) sample has Cu=18 ppm, Pb=14 ppm and Zn=431 ppm; whereas the nearer (eastern), more sulphidic sample has Cu=132 ppm, Pb=88 ppm and Zn=474 ppm.

Along the hill top back toward the Willroy road, a small branch road leads down hill to the northwest (Fig. A1). About 20–30 metres down the road along the ditch on the north side, a biotitic variety of sillimanite-muscovite-quartz schist crops out between the projected horizons of the Willroy 4 and 2/5 orebodies. In the schist, numerous sillimanite knots are associated with plagioclase, garnet, and minor sulphide minerals. Monocrystalline and polycrystalline quartz eyes, observed in thin section, resemble relict quartz phenocrysts and suggest that quartz-phyric rocks were the protolith to premetamorphic alteration.

Stop A4, Willroy millsite calc-silicate rocks, ZB91-171–175, ZB92-P318, P123, IAGOD #3. A group of stripped glacially-polished outcrops, on the north-facing slope leading down from the plateau of Stop A3, can be accessed either by walking down the brow of the hill or driving down and walking back up (Fig. A1). These enigmatic exposures of calc-silicate rocks always generate much discussion, although they are not a separable unit on our 1:25000 map. In general in the Manitouwadge belt, calc-silicate minerals are sporadically distributed throughout felsic to intermediate rocks, and particularly concentrated in calc-silicate-rich zones adjacent to contacts with iron formation. The Willroy millsite outcrops are atypical in structure (for example, fracture-control on calc-silicate distribution) and in their evidently complex multistage history, unusual features that make them difficult to place in a regional context.

The Willroy millsite calc-silicate rocks lie in the stratigraphic footwall of the Willroy 2/5 orebodies, and grade laterally to quartz-phyric felsic rocks (the northernmost lens on the Willroy property) which we consider to be the protolith to metamorphism and calc-silicate alteration. The abundance of clinopyroxene, garnet, Ca-amphibole, plagioclase, epidote, calcite and titanite, as well as quartz and microcline, is variable. Locally, for example in exposures high on the slope, dark calc-silicate minerals occupy a deformed conjugate-fracture set in which flattened diamond-shapes (looking down-plunge) outline more leucocratic felsic domains comprising quartz, microcline, plagioclase and minor hornblende, biotite, garnet and epidote. Fractures oriented at high angle to foliation are folded (shortened), whereas those at low angle to foliation are straight and look extended. Pervasive calc-silicate zones (coarse grained, green with epidote, clinopyroxene and Ca-amphibole), of 0.25–1 metre width, are subparallel to local foliation trends and cross the fracture-controlled alteration. The pervasive zones have the appearance of breccia and, locally, some 'fragments' preserve layering or fracture-controlled calc-silicate alteration. In some cases, early structures look rotated or truncated at the contact to the pervasive calc-silicate matrix.

This group of outcrops, and a suite of calc-silicate rocks from the Geco mine (Suffel Collection), were studied by Pan and Fleet (1992), who concluded that dispersed Ca was remobilized and concentrated during metamorphism. High Cl in calcic amphibole from the Geco mine could indicate a memory of previous synvolcanic enrichment during sea-floor alteration (*ibid.*). In our view, the localization of calc-silicate zones near iron formation contacts could be attributed to either primary or secondary processes. One possibility is that calc-silicate zones formed during metamorphic redistribution of Ca during decarbonation reactions of carbonate-facies iron formation. However, arguing against this is the lack of evidence for carbonate-facies precursors for metamorphosed iron formation at Manitouwadge, and the paucity of carbonate minerals in the calc-silicate rocks. Alternatively, in a model-driven syngenetic scenario, the distribution of calc-silicate rocks could reflect synvolcanic hydrothermal alteration in which iron formation acted as a cap rock.

From the eastern base of the main outcrop, walk east along the road to another outcrop about 10 metres into the woods on the south side. Here, the rock is dominated by coarse grained garnet and hornblende, and more typical of calc-silicate zones in general. There is little evidence of a felsic precursor until, a short distance to the east (just before a gravel road branching to the south), a pavement (ZB92-P123) of quartz-phyric felsic rocks, typical of those along-strike of the Willroy millsite calc-silicate rocks.

Stop A5, Willroy railway cut, ZB91-15, IAGOD #3 (continued). Looking north from the Willroy millsite, the rock wall of the Willroy-Geco railway cut 150 metres away is subparallel to the dominant foliation of the exposed orthoamphibole-cordierite-garnet gneiss, and the easterly-plunging lineation is obvious and spectacular (Fig. A1). Orthoamphibole-cordierite-garnet gneiss (Unit 2) is interpreted to be metamorphosed synvolcanic alteration in the stratigraphic footwall to known economic deposits. The unit is continuous around the inner Manitouwadge synform from Rabbitskin Lake in the north, to the Falconbridge zone of subeconomic mineralization to the east, a distance of about 30 km (1:25000 map). (Don't spend too long at this outcrop; it's nice, but there are much better exposures of altered rocks at Stops A6 and A7.) The interlayered (metre-scale) orthoamphibole-garnet-cordierite±biotite±plagioclase±staurolite and garnet-biotite-sillimanite±cordierite assemblages are typical of the unit. Orthoamphibole is present both as aligned prisms and large (4–5 cm) radiating sprays, variation possibly resulting from heterogeneous deformation or post-kinematic recrystallization. Biotite, at least in part, is found in pseudomorphs after orthoamphibole, suggesting retrograde potassic metasomatism. Quartz pods locally contain very coarse (10 cm plus) crystals

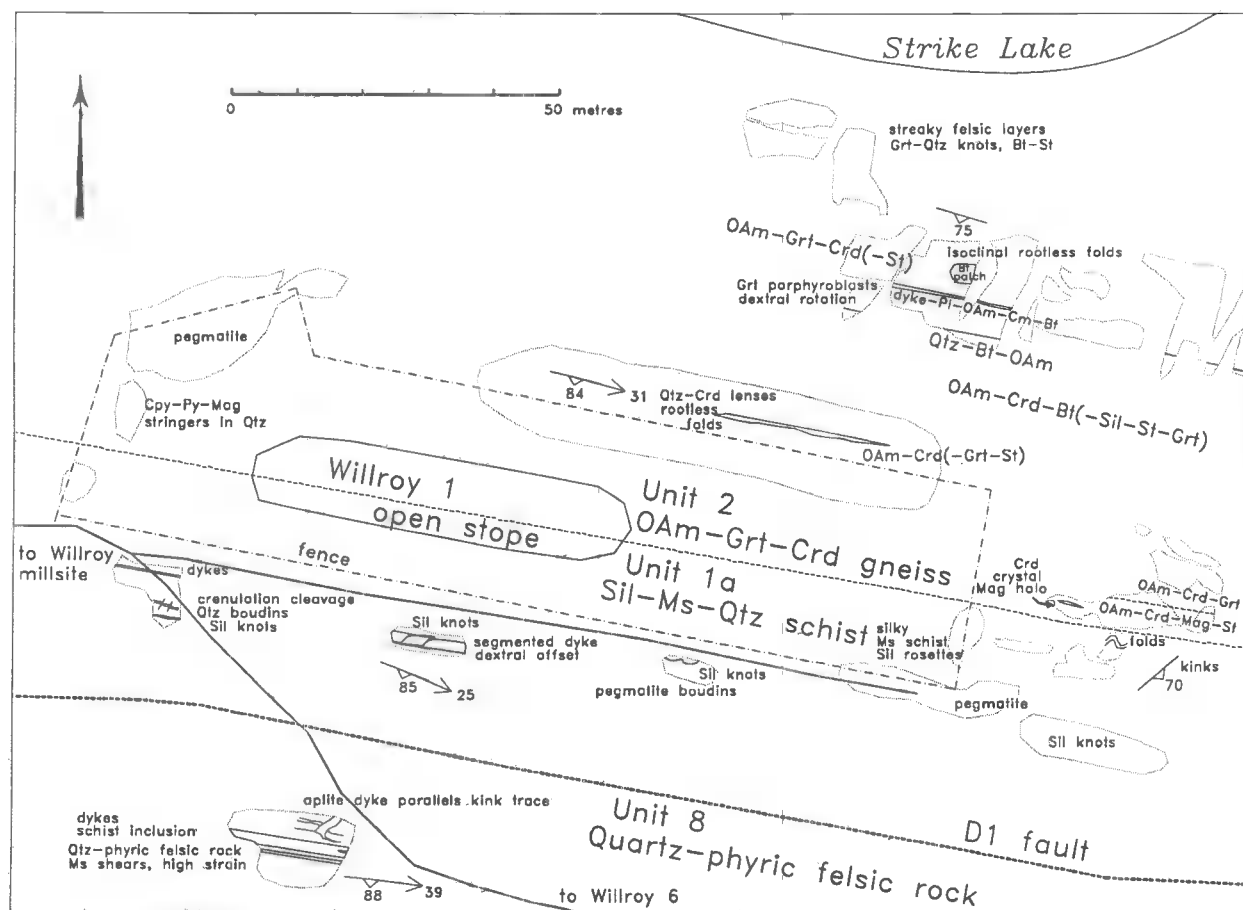


FIG. A2. Detailed geology of the Willroy 1 Cu stringer orebody (Table 3) and host rocks. Structure symbols show dominant D_2 foliations and lineations, except for post- D_2 kinks (labelled). Mineral abbreviations are after Kretz (1983). Dotted lines show outcrop areas.

of blue cordierite.

Stop A6, Willroy 1 alteration section, ZB91-146-147, 164-170, ZB92-P1-P7, IAGOD #4. From the Willroy millsite, take an overgrown gravel road east for about 1.25 km to the chain-link fence on the north side of the road. From the western end of the fence, the open stope of the Willroy 1 stringer Cu orebody is visible (Fig. A2). The rocky knolls to the north and in the fenced area are mostly pegmatite. The following outcrop descriptions are from south to north, from stratigraphic hanging wall to footwall of the Willroy 1 orebody.

Walk about 50 metres southeast along a gravel track to a rock knoll south of the track. The white weathering felsic schist, with muscovite-rich partings and shears, is a highly strained metavolcanic rock associated with the interpreted D_1 fault that passes through the Willroy-Geco area. Relict quartz phenocrysts are visible in thin section, as well as abundant microcline and minor biotite, garnet and epidote. Foliated tonalite dykes that cut the felsic schist could belong to the syn- D_2 suite. A massive aplite-pegmatite dyke intrudes along the axial surface of a high-angle kink of schistosity and has an asymmetrical tail that extends along a foliation-parallel shear. The timing relationships are ambiguous; either a pre-existing shear influenced the intrusion geometry, or the dyke was deformed during local reactivation of shears.

Return toward the fence to a rusty pavement on the track. This outcrop and several along-strike are examples of sillimanite-muscovite-quartz schist (Unit 1a), in this exposure, consisting of thin alternating layers (1-5 mm) of quartz and coarse grained muscovite. Locally, the layers define isoclinal folds (mm- to cm-scale). Sillimanite knots are present in discontinuous zones. Several features show dextral sense of motion in plan view including, an oblique muscovite foliation resembling a C-S fabric with mainly dextral asymmetry, quartz

lenses with asymmetric tails, and Z-shaped folds and crenulations of the dominant schistosity (D_2). Muscovite lineation (D_2 ?) plunges to the east, whereas the axes of Z-folds and crenulations (D_3 or later?) plunge to the west.

Continue along the south side of the fence for about 50 metres to a pavement just in the trees. The sillimanite-muscovite-quartz schist has large elongate (easterly-plunging) sillimanite knots, locally coalescing into sillimanitic layers. Dyke relationships are again interesting; a concordant foliated tonalite dyke (50 cm wide) is cut by several thin aplitic dykes. The aplitic dykes have sigmoidal traces in the tonalite (syn- D_2 ?) and tail off into asymmetrical foliation-parallel apophyses in the host schist. The tonalite dyke shows small amounts of offset along the cross-cutting aplites. In plan view, the asymmetry and offsets are consistent with dextral motion. Interestingly, sillimanite knots are absent in the host schist for about 30 cm on each side of the tonalite dyke. About 50 metres further east, another outcrop of sillimanite-muscovite-quartz schist contains a foliated tonalite dyke and aplite-pegmatite boudins.

Continue around the southeast corner of the fence. At the fence and immediately to the east, several excellent outcrops of muscovite schist are in abrupt contact (covered) with orthoamphibole-cordierite-garnet gneiss (Unit 2). The muscovite schist is silky white and has rosettes of fine sillimanite on foliation surfaces. Close to the contact, the schist is quartz-rich with asymmetric quartz lenticles and pods consistent with dextral kinematics, and kinks, crenulations and folds of the dominant foliation are well developed and pervasive. The Willroy 1 open stope is along-strike west of the contact. Orthoamphibole-bearing rocks to the north are streaky with semicontinuous quartz lamellae overgrown by porphyroblastic minerals. Orthoamphibole displays a blue iridescence, characteristic of the presence of exsolution lamellae in an originally supersolus orthoamphibole. Magnetite and staurolite are also present, the latter mantled by cordierite or by a cordierite-gahnite intergrowth, petrographic evidence of metamorphic decompression reactions.

Metre-scale compositional layering is defined by variation in proportions, textures and assemblages of metamorphic minerals; for example, on the northern part of the outcrop, garnet-poor rocks are transitional to garnetiferous rocks. First-order compositional layers and second-order lamellae probably represent a combination of primary and tectonic structures.

Continue to the north side of the fence, and follow the trail to the outcrops between the fence and Strike Lake, comprising an excellent section typical of alkali-depleted Fe-Mg rocks in the footwall to the Willroy-Geco deposits. There is a gradual northward increase in leucocratic felsic-looking lenses and semicontinuous layers, in some cases, resembling attenuated hinge zones of isoclinal folds. Quartz-rich layers and lamellae are overgrown by garnet porphyroblasts showing differential dextral rotation. Some layers are spotted by staurolite with a white mantle of partially pinitized cordierite. The presence of the assemblage, plagioclase-biotite-gedrite-cummingtonite, in a concordant intermediate dyke (20 cm wide) suggests either, that the dyke was hydrothermally altered or, that it exchanged with altered host rocks during metamorphism.

A local irregular discordant patch of biotite overgrows and obliterates layering and foliation. The biotite is weakly or randomly oriented but shows a poorly developed crenulation suggesting, as at Stop A5, that at least some biotite is of retrograde origin. Monazite inclusions in biotite, extracted from mica schist at the Geco mine, define a U-Pb isotopic age of 2661 ± 1 Ma, possibly dating a retrograde event (Schandl et al., 1991).

Stop A7, Synvolcanic trondhjemite, ZB91-20, ZB92-P191-P193. Return to the Willroy millsite and railway cut, and find the easiest access to the railway bed (1:25000 map). Continue easterly along the old railway keeping track of distances from the first outcrop on the south side (just east of a fluorescent? bright green pond). Initially the route passes exposures of orthoamphibole-cordierite-garnet gneiss (Unit 2), foliated trondhjemite (Unit 12), and enclaves of mixed mafic and felsic rocks (Unit 4), the latter interpreted to be the protolith to synvolcanic alteration. After 400 metres, foliated trondhjemite typical of Unit 12 is exposed in several outcrops. In general, trondhjemite contains magnetite and biotite and, near the contact with orthoamphibole-garnet-cordierite gneiss, garnet is locally abundant. Our field interpretation of the unit as a synvolcanic intrusion was corroborated by U-Pb zircon geochronology giving an age of 2720 ± 3 Ma, within error of the age of felsic volcanism (Figs. 10 and 11, Table 2). 'Immobile' elements in geochemical analyses of trondhjemite from this and other locations are indistinguishable from those of quartz-phyric felsic rocks (Figs. 24 and 27), suggesting intrusive and extrusive equivalents of the same magma.

Continue to 750 metres to low rock knolls on the north side of the rail bed; you are about 100 metres north of Strike Lake (not visible) and Stop A6. Trondhjemite is host to diffuse 'seams' of garnet-orthoamphibole-cordierite-biotite, some of which are folded and have an axial planar foliation. We interpret the seams as the result of synvolcanic alteration similar to that of adjacent orthoamphibole-garnet-cordierite gneiss and, by implication, we interpret the trondhjemite as a high-level synvolcanic intrusion that contributed heat to the hydrothermal system. Trondhjemite intruded its own aureole of hydrothermally altered rocks and, upon subsequent cooling, was subject to incursion by hydrothermal fluids.

Stop A8, Pillowed mafic rocks, ZB91-21, ZB92-P202. This is a special stop for keeners and sceptics

who need proof of mafic volcanic rocks north of the Willroy-Geco area. Continue for about 1.5 km along the railway cut to dark rocks exposed on both sides. On the north side, some highly deformed pillows are discernible. Geochemically, these rocks are tholeiitic basalts similar to mafic rocks in the outer volcanic belt of the Manitouwadge synform (Figs. 23–26).

A9–A20. Slim Lake section

The Slim Lake field trip visits many rock units already described, but it is useful as a near-continuous 'stratigraphic' section across the Willroy area (Fig. A1) including exposures of straight gneiss (annealed mylonite) associated with the interpreted D₁ fault. The outcrops are mainly along the margins of a tailings meadow, formerly Slim Lake. The trip is divided into two parts; firstly, north of the Willroy road along the eastern side of the meadow (A9–A14), exposures are described from north to south; secondly, south of the road along the western side of the meadow (A15–A20), descriptions continue to the south. The base of the supracrustal section (to the north) comprises interlayered mafic and felsic metavolcanic rocks, invaded by synvolcanic trondhjemite to granodiorite. To the south, about 150 metres of orthoamphibole-garnet-cordierite gneiss (Unit 2) and enclaves of less altered mixed mafic/felsic rocks (Unit 4) are overlain by a discontinuous belt or lens of quartz-phyric felsic metavolcanic rocks (Unit 8). This northerly lens of quartz-phyric felsic rocks, about 150 metres thick at Slim Lake, is interpreted to be equivalent to the southerly lens of similar rocks (e.g. Stop A3), repeated across a D₁ fault (Zaleski and Peterson, 1993a). Quartz-phyric rocks are overlain by the iron formation (not exposed) that hosts the Willroy 2 and 5 deposits to the east, and by 50–100 metres of sillimanite-muscovite felsic rocks (Unit 1a). The southernmost unit on the traverse is a laminated felsic straight gneiss interpreted to lie on the D₁ fault that truncates some map units (notably iron formation) and repeats part of the section (quartz-phyric felsic rocks, iron formation, sillimanite-muscovite-quartz schist).

Continue west on the Willroy mine road from the millsite and railway cut, turning north on the dirt road that follows the east side of the tailings meadow. Stay on the high road over a containment dam (Stop A12 is along a short eastern fork of the road somewhat below the dam). Although the road peters out after the dam, you should be able to drive (at your own risk) to the north end of the meadow to another dam. Be sure to look down over the steep northern side of the dam for a view of the remnants of Slim Lake and a perspective on the thickness of the tailings pile.

Stop A9, Potassic variant of synvolcanic trondhjemite, ZB92-P99. The Stop is east of the tailings dam along the edge of the woods (Fig. A1); more outcrops of the same rock lie along the west side of the dam. The foliated granodiorite is texturally, modally and geochemically similar to trondhjemite (e.g. Stops A7 and A12), especially in the abundance of coarse grained quartz and magnetite porphyroblasts. We interpret foliated granodiorite as a potassic variety of the synvolcanic intrusion.

Stop A10, Quartz-phyric felsic rocks in mixed mafic-felsic rocks, ZB92-P101. Retrace the route to the south approximately 300 metres to an old drill road. Walk 50 metres east along drill road to some small outcrops on the north side of the road. The felsic metavolcanic rock with a few quartz eyes is part of the package of interlayered mafic and felsic metavolcanic rocks, elsewhere affected by synvolcanic alteration.

Stop A11, Mixed mafic-felsic rocks, (optional), ZB92-P102. Retrace the route for another 100 metres south of the drill road and look for an outcrop just in the woods to the east. The exposure belongs to the same map unit as Stop A10, here characterized by intimately interlayered, fine grained mafic and felsic rocks. The interlayering suggests that mafic and felsic volcanism was partly coeval. The numerous thin cross-cutting quartz-rich and granitic veinlets have been folded.

Stop A12, Foliated trondhjemite, ZB92-P104, ZB91-4. Continue south over tailings dam to a lower eastern fork of the road; turn north and continue down a short distance to a pavement outcrop on the east. The exposure is typical of synvolcanic trondhjemite, quartz-rich leucocratic and foliated, with magnetite and local biotite-rich patches. A foliated plagioclase-porphyritic tonalite dyke (20 cm) probably belongs to the syn-D₂ suite.

Stop A13, Orthoamphibole-garnet-cordierite gneiss and mixed mafic-felsic metavolcanic rocks, ZB92-P106, P109, P441. Continue south to a broad area of pavement outcrops that form the top of the low knoll east of Slim Lake and north of the Willroy road. These extensive outcrops are part of the footwall alteration zone characterized by orthoamphibole-garnet-cordierite gneiss; however, hydrothermal alteration was apparently heterogeneous, leaving hints of the interlayered mafic-felsic protolith. Compositional layering is defined by metamorphic assemblages that include; quartz, orthoamphibole, garnet, cordierite, cummingtonite, hornblende, sillimanite, staurolite, gahnite and plagioclase. If time permits, a visit to the eastern end of the outcrops is interesting. The metamorphosed altered rocks are folded and in one place, a fold limb is truncated by a broad, diffuse zone of the same minerals, suggesting local metamorphic redistribution of orthoamphibole-bearing assemblages.

Stop A14, Orthoamphibole-garnet-cordierite gneiss, ZB92-P108 (optional). Return to the Willroy road and continue south about 50 metres to an outcrop on the east side. In this example of metamorphosed altered

rocks, layering is defined by sillimanite-cordierite and orthoamphibole-cordierite-staurolite assemblages. A biotite-rich zone is interpreted to be the result of local potassic retrogression. Note the somewhat disorganized, D₃ or D₄ crenulations and kink folds, preferentially developed in the micaceous zone. Note also the spectacular sprays of orthoamphibole nearest to the road.

Stop A15, Felsic metavolcanic rock with orthoamphibole-garnet, ZB92-P37. Take the Willroy road to the west side of the Slim Lake tailings meadow to commence second leg of section. The first brief stop is a pavement outcrop along the south side of road. This and the following stops (A15–A20) are easily accessed by walking along the western margin of the meadow.

We have back-tracked to the north and are once again in Unit 4, interlayered mafic-felsic rocks, here represented by heterogeneous felsic metavolcanic rock. Minor amounts of biotite, hornblende, magnetite, garnet, and local quartz and plagioclase phenocrysts(?) define a weak compositional layering. A few zones of orthoamphibole-garnet are typical of synvolcanic alteration, although much of the orthoamphibole has been replaced by biotite. The metavolcanic rocks are intruded by foliated tonalite. Some minor folds are present, and there is an indication of dextral, north-side-down kinematics from porphyroblasts with asymmetric tails of matrix material.

Stop A16, Contact of trondhjemite and altered rocks, ZB92-P77. Walk south along west side of meadow, keeping track of the distance from the first outcrop beyond Stop A15. At 60 metres, just before an outcrop of fine grained hornblende schist, head west up into the woods. After climbing up a steep slope over deadfall (50–60 m), go to the north end of a low ridge of outcrop.

The north end of the outcrop is medium to coarse grained trondhjemite; the contact between this and altered rocks is exposed in the central part of the outcrop under a tree limb. At the contact, the trondhjemite is finer grained, more strongly foliated and contains more garnet and biotite than usual. The altered rocks are heterogeneous, grading from very coarse grained garnet-biotite schist near the contact to finer grained garnetiferous felsic rocks, to intermediate rocks with hornblende and cummingtonite layers. The relationships suggest that incipient alteration affected trondhjemite along the contact.

Carefully scramble back down to the meadow and continue south. Take a quick look at the hornblende schist previously noted, a less altered equivalent of some of the rocks just visited.

Stop A17, Metamorphosed synvolcanic alteration, ZB92-P56, P79. Continue to a large outcrop, about 150 metres south, dominated by pegmatite but with a sheet of garnet-orthoamphibole-biotite schist with very coarse garnet. About 40 metres further on (190 m), in a small pavement outcrop fresh blue cordierite is present, with sprays of orthoamphibole, and tiny staurolite grains.

Stop A18, Quartz-phyric felsic rocks, ZB92-P81–P83, P54. Continue to several low outcrops between about 300 to 400 metres south. These rocks are part of the northern body of quartz-phyric felsic metavolcanic rocks, characterized by abundant (to 20%) quartz phenocrysts and ubiquitous magnetite porphyroblasts.

Optional for keeners and breccia lovers; head up into woods near the northernmost exposures of the quartz-phyric felsic rocks looking for the cut lines (somewhat overgrown) of the Willroy grid. Find L130+00E at approximately 25+00N and a small stripped outcrop of monolithologic quartz-phyric felsic breccia. Angular fragments, ranging in size from very small to metre-scale, are supported in a pale green matrix enriched in biotite, hornblende and epidote.

Stop A19, Sillimanite-knot felsic schist, ZB92-P86, P87. Several small outcrops between 500 and 600 metres document the presence of Unit 1.

Stop A20, Straight gneiss (annealed mylonite), ZB92-P88. At approximately 600 metres, go 10 metres into the woods to a large outcrop. The pink to white, quartzose laminated gneiss, with some thin biotite- and hornblende-rich layers, is very hard and flinty. This is an example of straight gneiss, forming part of a zone of laminated rocks, lying on the interpreted D₁ fault that passes through the Willroy-Geco area. In this case, the straight gneiss is completely annealed to a granoblastic texture. Some garnet is present locally, and very fine magnetite porphyroblasts are surrounded by leucocratic halos.

A21. Nama Creek deposit

Stop A21, Nama Creek Mines deposit, ZB91-110, IAGOD #5. From the western margin of the tailings meadow, drive northwest on the continuation of the Willroy road leading to the Nama Creek and Willecho deposits (1:25000 map). About 600 metres from the meadow, turn off on a gravel road to the southwest and continue 500 metres to the flooded open cut of the Nama Creek deposit (Fig. A3).

The Nama Creek Mines deposit (also known as the Big Nama deposit) was a marginally economic Zn-rich orebody (Table 3) hosted by iron formation immediately south of orthoamphibole-garnet-cordierite gneiss (Unit 2). The iron formation can be traced east to that hosting the Willroy 2/5 orebodies. As traced by drilling, the subeconomic extension of the Nama Creek mineralization apparently plunges easterly to converge in the subsurface with the Willroy 3 horizon (1:25000 map). The two zones are separated by the D₁ fault,

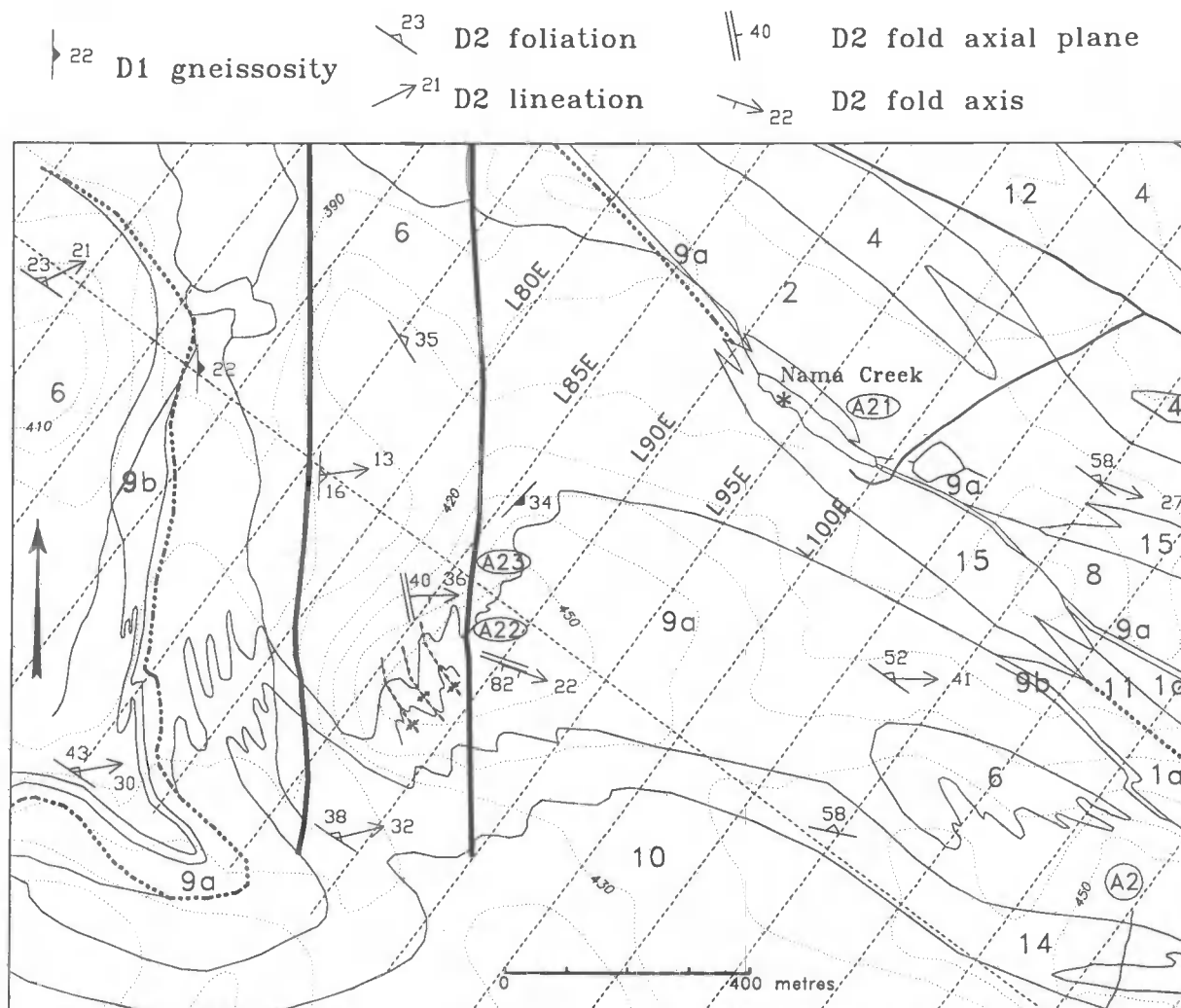


FIG. A3. Geology and field-trip stops in the area of an S-shaped D_2 fold of iron formation (Unit 9) and felsic metavolcanic rocks (Unit 6). Dashed lines are cut lines of Noranda's Willroy grid.

interpreted to lie south of the Nama Creek deposit. The deposit is extensively invaded by pegmatite-aplite sheets which tend to obscure field relationships.

On the north side of the flooded cut, orthoamphibole-garnet-cordierite gneiss has minor amounts of magnetite, ilmenite, chalcopyrite, pyrrhotite and biotite, as well as staurolite inclusions in cordierite and garnet. Orthoamphibole-bearing gneiss is intercalated with sillimanitic layers and, in some cases, sillimanite occurs in orthoamphibole-bearing assemblages. However, sillimanite is mantled by cordierite or quartz, or engulfed by skeletal garnet, textures suggestive of metamorphic decompression reactions.

At the eastern end of the cut, pegmatite is interleaved with chlorite-mica schist, foliated tonalite and semi-massive pyrite and magnetite. In the concordant pegmatite dyke, coarse grained quartz, plagioclase and sprays of sillimanite (to 5 cm) tend to be oriented perpendicular to dyke margins, looking as though they nucleated on the wall rock and grew into the pegmatite, apparently without subsequent deformation.

A22–A23. D_2 folds of iron formation/felsic metavolcanic contact

Southwest of the Nama Creek deposit, the thick southernmost iron formation (the same as that of Stop A2) of the inner volcanic belt defines a map-scale S-fold (1:25000 map, Fig. A3). Interdigitations of felsic rock and iron formation related to parasitic folds in the hinge region are well exposed. Locally, a gneissic fabric (D_1) in the felsic rocks is involved in folding. The dominant D_2 foliations and lineations are parallel to the axial planes and fold axes, respectively, of outcrop-scale folds. The axial trace of the map-scale S-fold is

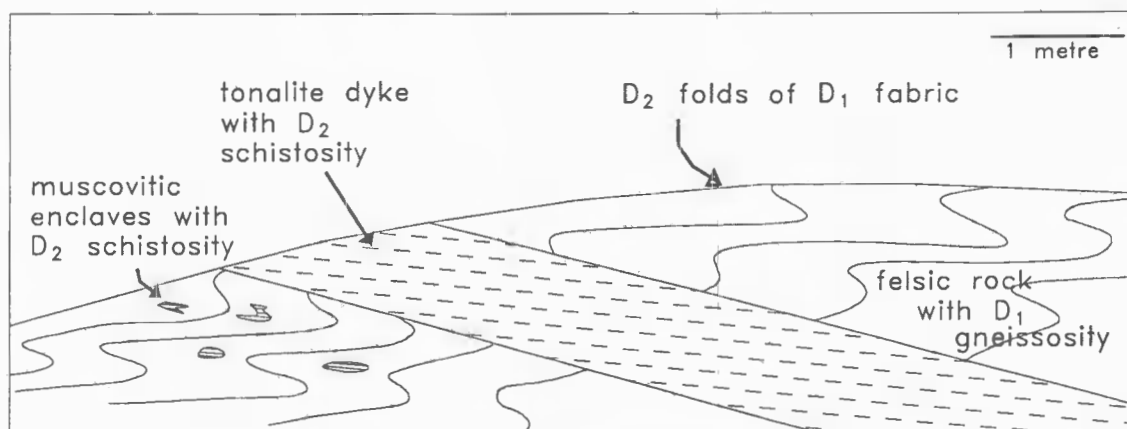


FIG. A4. Sketch of outcrop relationships (Stop A23) between D₂ folds of D₁ gneissosity in felsic metavolcanic rock, and D₂ schistosity in a syn-D₂ tonalite dyke.

deformed by the D₃ Manitouwadge synform and the S-asymmetry is inconsistent with what would be expected for a D₃ parasitic fold related to the synform. On the basis of these observations, the S-fold is interpreted to be a D₂ fold that reorients vestiges of D₁ fabric preserved in felsic rocks in the strain shadow near the hinge region. A major D₂ sheath fold is interpreted to repeat the D₁ fault and mineralized iron formation of the Willroy-Geco area between the Nama Creek and Willecho deposits (Fig. 4).

Access to the exposed contact involves about 700 metres (one way) of bushwacking on the somewhat overgrown cut lines of the Willroy grid. The exposures are worth viewing for those interested in the pre-D₃ structural history of the area. On the south side of the flooded cut of the Nama Creek deposit, look for L95E (17+00N, English units). Although the picket may be fallen and illegible, it should be possible to spot the old cut and follow it southwest (218°) about 530 metres to the Willroy baseline. Remember that your compass may not be reliable.

Stop A22, Folds in iron formation, ZB91-46. Continue northwest (308°) 170 metres on the Willroy baseline to about 20 metres beyond L90E and some excellent stripped outcrop. In the quartz-magnetite iron formation some groups of layers define S-folds (D₂) and others are disrupted by stratabound breccia zones. The breccias, consisting of segmented quartzose layers in a dark matrix, tend to be localized where quartzose layers are thin and/or proportionately less abundant than dark layers of magnetite and Fe-silicates. Quartzose fragments are preferentially oriented subparallel to the axial traces of folded layers. The apparently competent behaviour of quartzose fragments suggests that brecciation occurred after diagenesis or low grade metamorphism, conditions under which quartz is relatively competent. A pegmatite cuts across the iron formation and contains a discontinuous (<5 mm) garnetiferous zone at the contact.

Stop A23, D₂ folded contact, ZB91-50, ZB92-P13. Continue northwest on the baseline another 50–60 metres to the crest of a hill with a view down to Garnet Lake and along the baseline. On the hill top and northwest-facing slope, there are good exposures of interfingering iron formation and felsic rock. Scattered outcrops of folded iron formation in the thick bush show changes between Z- and S-asymmetry related to the axial traces of D₂ minor folds (Fig. A3).

Continue just over the hill crest north of the baseline and along L87+50E to exposures of the folded contact between iron formation and felsic rock (with sparse quartz eyes). Locally, the felsic rock has a crenulated gneissic layering (D₁) defined by quartz lamellae and biotite-garnet-rich zones. The biotite has a crystallographic-preferred orientation parallel to crenulation traces. Near the folded contact, felsic gneiss contains odd irregular muscovite-schist enclaves (some look angular, some like rootless isoclinal folds). The gneissosity in the host rock is folded, whereas, muscovite in the enclaves has an axial planar orientation (Fig. A4). We interpret the gneissosity to be a D₁ foliation preserved in the hinge region of the map-scale D₂ fold.

A tonalite dyke subparallel to the axial traces of D₂ folds (of somewhat variable orientation) near the iron formation/felsic contact has an axial planar foliation, parallel to that in the irregular muscovite-schist enclaves in adjacent felsic rocks (Fig. A4). The relationships suggest that the dyke was intruded during D₂ deformation, interpreted to be broadly coeval with peak metamorphism. On the basis of lithology, field relations and geochemistry, the dyke is one of a suite of tonalite dykes, other probable examples of which were seen at Stops A1, A2 and A5. A sample of the dyke, collected for geochronology, contained no separable

zircon. Monazite, including grains of igneous morphology, defines a U-Pb age of 2671 ± 3 Ma (Fig. 14, Table 2), within error of both 2675 ± 1 Ma metamorphic monazite from the Geco mine (Davis et al., 1994), and 2669 ± 3 Ma monazite from a pre- to syn- D_1 pegmatite (Stop A26). The estimated temperature of peak metamorphism (650°C) in the Manitouwadge belt is less than the closure temperature of the U-Pb system in monazite (700°C , Heaman and Parrish, 1991). Nevertheless, it would appear that monazite ages are dating metamorphic or late metamorphic events, or that monazite continued to exchange isotopes during protracted regional cooling and decompression.

A24–A26. Willecho 3 pit and inner hinge of the Manitouwadge synform

The three Willecho orebodies lie in an area of complex D_1/D_2 folding and faulting near the hinge region of the D_3 Manitouwadge synform (1:25000 map, Fig. 4), and it is difficult to reconstruct their exact stratigraphic relationship to orebodies in the Willroy-Geco area. North of Willecho 3, the sequence; quartz-phyrlic felsic rocks, iron formation, sillimanite-knot felsic schist, is repeated in a map-scale isoclinal fold, possibly a D_1 fold. In the Willroy-Geco area, we interpret a similar sequence as a stratigraphic succession. Sillimanite-knot felsic schist in the Willecho area (Unit 1b) is, in general, more felsic than sillimanite-muscovite-quartz schist in the Willroy area (Unit 1a, Stop A6), although the latter also has felsic variants. Typically, layers with large disseminated sillimanite knots are interleaved with non-sillimanitic layers, and sillimanite-biotite-garnet schists are of local occurrence.

In general, toward the north in the Willecho area, there is a transition from muscovite-sillimanite-microcline to sillimanite-microcline assemblages, consistent with northerly increase in metamorphic grade.

Stop A24, Sillimanite-knot felsic schist, ZB91-2, ZB92-228. From the Nama Creek junction, continue west about 2.25 km on the gravel road toward the Willecho deposit. Where the road starts uphill on the northwest side of Garnet Lake, stop at some low outcrops in the ditch on the north side (1:25000 map). The outcrops are typical of Unit 1b; diffuse layering in the dominantly felsic rocks is defined by concentrations of disseminated large (to 2 cm) zoned white-green sillimanite knots and local garnet. The sillimanite knots are strongly elongate, plunging northeasterly. Zoning in the knots commonly consists of a muscovite-sillimanite core mantled by weakly oriented by sprays of fine sillimanite, in turn mantled by strongly oriented sillimanite fibres engulfed or overgrown by coarse, either oriented or random muscovite. In one sample, the cores of sillimanite knots contain relict andalusite. The matrix has a few quartz eyes and is composed of quartz, plagioclase, microcline, biotite and muscovite. Muscovite mostly looks like late unoriented overgrowths, but commonly some blades, interleaved with foliated biotite, are concordant to schistosity.

Boudins of pegmatite wrapped by gneissic layering indicate high strain, possibly related to straight gneiss exposed at the Willecho pit (Stop A26). Sillimanite-knot felsic schist is also well exposed south of the road, along an overgrown track angling southeast toward Garnet Lake.

Stop A25, Limb of $D_1(?)$ fold, ZB92-87. Drive on about 500 metres west of Stop A24 to a gravelled clearing and the remains of building foundations marking the location of the former Willecho shaft, which intersected the down-plunge extension of the Willecho 3 orebody. Continue on the road west of the clearing about 100 metres to an overgrown track branching to the north, and walk about 50 metres to excellent stripped outcrops near L5N (may or may not be visible) of the Willecho grid. The stripping exposes the folded contact between quartz-magnetite iron formation and sillimanite-knot schist on the limb of the $D_1(?)$ map-scale fold. The enveloping surface of the contact trends about 260° ; whereas, metre-scale isoclinal S-shaped folds of the contact and layered iron formation have oblique westerly-trending axial traces. Schistosity (composite $D_1/D_2?$) defined by biotite and flattened sillimanite knots is axial planar to the folds. The asymmetry of minor folds and of fabrics is consistent with a parasitic relationship to the map-scale D_1 fold, but the schistosity defined by high grade minerals may be a D_2 fabric developed during transposition of the early fold.

Stop A26, Willecho 3 pit and straight gneiss, ZB92-55, IAGOD #6. The Willecho 3 pit is on the south side of the gravel road about 3.5 km west of the Nama Creek (Stop A21) junction. A gravel track slopes gently into the pit. Watch for hazardous loose rocks on the steep pit walls. The pit walls comprise lenticular to layered sillimanite-garnet-biotite schist, iron formation, minor orthoamphibole-garnet gneiss and semi-massive to disseminated sulphide mineralization (pyrite-pyrrhotite-sphalerite-chalcocopyrite). Locally, rotated garnet porphyroblasts wrapped by biotite-sillimanite suggest north-over-south sinistral kinematics.

Along the top of the northern side of the pit, from about midway along its length and continuing easterly, it is possible to walk out a transition from sheared pegmatite to straight gneiss over about 100 metres (at low angle to strike, true width may be <10 m). The westernmost outcrops of pegmatite are cut by anastomosing discrete shears with abundant coarse strongly lineated sillimanite on shear surfaces. Sheared pegmatite grades to porphyroclastic pegmatite with coarse quartz-K-feldspar-plagioclase enclaves (2–5 cm) in a fine grained matrix; lineated sillimanite coats shear surfaces. Closer to the eastern end of the pit, very fine grained layered rocks of felsic composition were grouped with straight gneiss (annealed mylonite), possibly of a mixed felsic metavolcanic and pegmatitic parentage. The presence of concordant and low angle pegmatite dykes suggests

multiple generations of pegmatite intrusion. Immediately below in the wall of the pit, mineralized sheets and lenses of the Willecho 3 deposit dip moderately to the north and plunge northeasterly, underneath the straight gneiss. We interpret the straight gneiss as lying on a D₁ fault, correlative with the D₁ fault of the Willroy-Geco area, repeated by a map-scale D₂ sheath fold (Fig. 4). By implication, the pegmatite involved in the straight gneiss transition was intruded before or during D₁ deformation. In a sample of pegmatite collected for U-Pb geochronology, only monazite was suitable for isotopic analysis. The age of 2669 ± 3 Ma (Fig. 13) is apparently a metamorphic age subject to the same ambiguities as the monazite age from the syn-D₂ tonalite dyke (Stop A23).

B. Outer volcanic belt

B1. Northern contact zone

Although the volcanic sequences of the inner and outer belts are correlative, the outer belt lacks voluminous synvolcanic intrusions, extensive units of orthoamphibole-bearing or sillimanite-muscovite-bearing rocks and (known) massive sulphide deposits. Throughout the outer belt, the northern contact between mafic metavolcanic rocks and felsic and metasedimentary rocks is a complex zone that includes semicontinuous garnetiferous domains (locally with orthoamphibole-cordierite), tonalite sheets and minor iron formation. The orthoamphibole-garnet-cordierite zones are interpreted as the distal equivalent of the extensive zones of metamorphosed synvolcanic alteration in the inner belt.

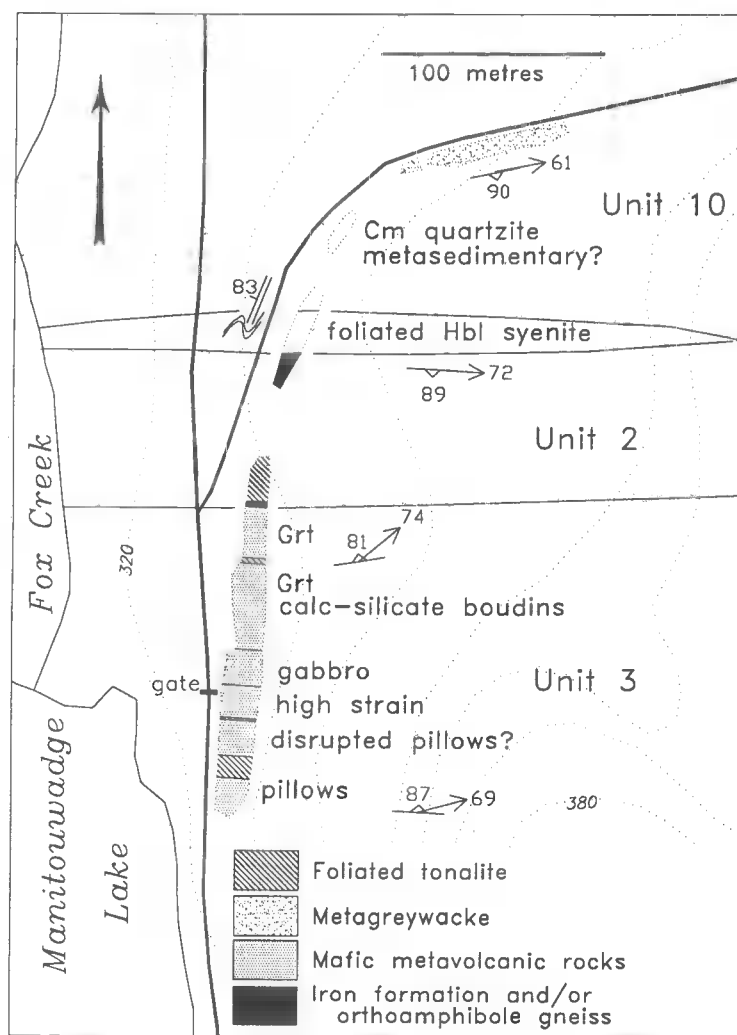


FIG. B1. Outcrop geology of the Geco gatehouse section, showing exposures of mafic and orthoamphibole-bearing rocks of the outer volcanic belt and the contact with metagreywacke. Structure symbols show dominant D₂ foliations and lineations, and the axial surface of a post-D₂ fold in foliated hornblende syenite. Mineral abbreviations from Kretz (1983).

Stop B1, Geco gatehouse section, ZB92-P249-P251, ZB93-P306, IAGOD #1. This stop encompasses outcrops that extend along the east side of the road to the Geco mine, starting across from the gatehouse at the Geco main entrance (Fig. B1). Many units typical of the outer volcanic belt, and the intruded and tectonized contact to younger metagreywacke, are exposed over a short distance. Begin at the southern end of

the outcrops (south of the gate) and work north to the eastward curve in the road sign-posted to the Geco #4 shaft. The section begins in the northern part of mafic metavolcanic rocks of the outer belt (1:25000 map). The highly deformed pillows and pillow breccias in the first outcrop were the main focus of previous field trips (Williams et al., 1990). The pillows are best examined on the glacially polished top surface of the outcrop. The pillows appear significantly shortened on both top and front surfaces of the outcrop. Pillow selvages are dark in colour and have local concentrations of garnet. Interpretation of structural facing is tempting, but ambiguous (go ahead, everybody does it).

Moving north along the outcrop, mafic metavolcanic rocks, mostly without recognizable pillows, are interleaved with foliated tonalite sills (Fig. B1). In the vicinity of the fence, foliated coarse grained hornblende-plagioclase gabbro is interleaved with mafic schist. The gabbro thickens to the west in the Gaug Lake area. North of the foliated gabbro, mafic metavolcanic rock, with garnetiferous layers and calc-silicate boudins, and foliated tonalite continue to the end of the first outcrop. Note the steep easterly plunging lineation. A thin layer of rusty, locally magnetic rock, interpreted as iron formation, is exposed at the contact between mafic rocks and tonalite near the northern end of the outcrop.

Along the road about 25 metres, the southern part of the next outcrop includes interlayered orthoamphibole-plagioclase-biotite gneiss and iron formation. Along strike just west of Manitouwadge Lake, orthoamphibole gneisses contain cordierite, garnet, staurolite, and gahnite, similar assemblages to those in altered rocks near the Willroy deposit (Stops A5, A6). The central outcrop area is dominated by fine grained, foliated hornblende-microcline syenite, possibly belonging to a hornblendite-syenite suite sporadically exposed throughout the Manitouwadge area (compare Stops F4, F5). A D_3 Z-fold of the dominant syenite foliation is exposed on the top surface of the outcrop. The northern end of this, and all of the next outcrop, are cummingtonite-bearing rocks of intermediate to felsic composition, possibly due to silicification. The cummingtonite-bearing rocks were grouped with metasedimentary rocks; however, their origin is uncertain. The remaining outcrops to the north along the inside curve of the road are more typical biotite-quartz-plagioclase schist interpreted as metagreywacke.

B2–B10. Gaug Lake area, comparison to inner volcanic belt

The Gaug Lake area lies in the outer volcanic belt on the southern limb of the Manitouwadge synform (1:25000 map) and, like the inner belt, comprises mafic and felsic rocks and iron formation (Fig. B2). Much of the area belongs to the 'Central' Granges Inc. claim group and grid. Felsic rocks and iron formation are less abundant than in the inner volcanic belt, although this may reflect preservation rather than original volumes. Felsic rocks are of the same age in both belts (circa 2720 Ma), but differ in composition, comprising aphyric rhyodacites and dacites in the outer belt. Garnet-hornblende assemblages in mafic and felsic rocks could be the result of synvolcanic calc-silicate alteration and subsequent regional metamorphism.

In the Gaug Lake area, foliations are generally steep and lineations plunge east to northeast. The area encompasses enclaves of relatively low strain where volcanoclastic deposits are remarkably well preserved. The presence of outcrop-scale D_2 , and D_3 or later folds, suggests that at least some intercalation of felsic rocks and iron formation is due to folding. The Agam Lake fault forms a pronounced east-west topographic lineament, mainly within metasedimentary rocks, but locally juxtaposing metavolcanic and metasedimentary rocks. Evidence of an early ductile history is recorded in isoclinal folds and a strong stretching lineation near the fault zone. Locally, rocks preserve mylonitic textures, unusual in that straight gneiss associated with low angle faults (D_1) in the inner belt is mostly annealed. There is also evidence of later brittle movement on the Agam Lake fault (W. Bates, Granges Inc., pers. comm., 1992).

The access is along ATV trails and cat tracks leaving from the base of the Manitouwadge ski hill. For those intending to use an ATV, the ATV trail starts from the road about 500 metres east of the ski hill parking lot. Note, however, that the cat track that branches west to the Granges area may not be suitable for ATVs. The Gaug Lake area itself is about 1.5 km beyond the ski hill, and the stops are located using the cut lines of the Granges grid (metric units). The condition of the grid and pickets is variable and it is best to measure distance along the cat track in case some grid lines are difficult to spot, or pickets illegible.

Stop B2, Deformed pillow basalt, ZB93-408. Beyond the rope tow, head up the most easterly ski run for about 500 metres to where the run crosses a powerline clearing near some smooth sloping outcrops (Fig. B2). The pillow basalts belong to the outer mafic metavolcanic belt. The pillows are small (<0.5 m) and flattened with selvages and interstices defined by coarse grained hornblende porphyroblasts oriented parallel to schistosity.

Stop B3, Mafic schist, ZB93-382. Continue up the ski run about 200 metres to the next good outcrop, more typical of mafic metavolcanic rocks in the Manitouwadge belt. Here, medium grained gabbroic augen schist, possibly originally a massive flow or base of a flow, grades to laminated (1 mm–1 cm) fine grained highly strained mafic schist. In the high-strain zone, layer truncations resulting from boudinage have a superficial

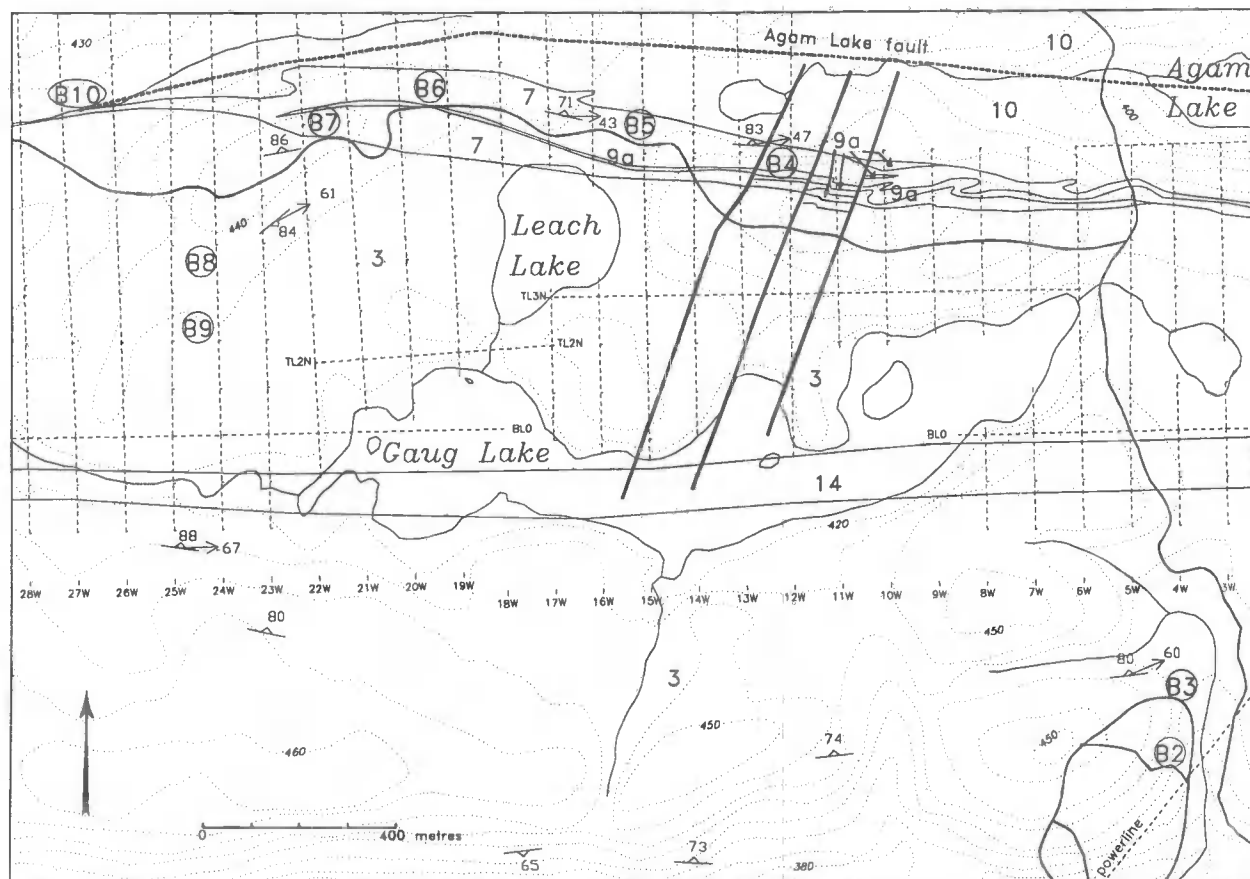


FIG. B2. Geology and field-trip stops (B2–B10) of the Gaug Lake area, outer volcanic belt. Dashed lines are cut lines of the Granges grid. Structure symbols show dominant D_2 foliations and lineations.

resemblance to cross-bedding. Layering also shows subtle rootless isoclinal folds. The layering looks somewhat tuffaceous, but could be entirely of tectonic origin.

Return down the ski run to the powerline and follow the rough trail leading northwest under the line for 200 metres to join the ATV trail. Follow the ATV trail to the north, crossing the stream at the east end of Gaug Lake after 1 km, and continuing up the hill on the north side for 100 metres. Leave the ATV trail and take the cat track (drill road) heading west. The cat track crosses cut lines (azimuth=0°) of the 'central Granges grid' at 100 metre intervals, the ATV trail being at approximately L5W.

Stop B4, Felsic metavolcanic rocks and iron formation, ZB93-366. Find L13W on the cat track at the northern edge of a drill-site clearing and continue on the line to 5+40N (about 40 metres north of the clearing). The stripped patch of felsic metavolcanic rocks, and another of quartz-magnetite iron formation about 25 metres to the north-northeast (15 metres east of L13W at 5+65N), are typical of intercalated units in the area. Continue east about 40 metres in relatively open bush to a long (20 m) northeasterly trending ledge with excellent exposures of felsic breccia. Angular monolithologic felsic clasts (up to 30 cm) are supported in a darker matrix rich in garnet, hornblende and biotite. The large size and angularity of the clasts is characteristic of proximal volcanoclastic deposits, possibly fragmented by phreatic (W. Bates, Granges Inc., pers. comm., 1992) or hydrothermal explosions. The abundance of calc-silicate minerals in the matrix suggests preferential synvolcanic alteration of permeable unconsolidated matrix material.

Stop B5, Felsic breccia, mafic-felsic contact, ZB93-367. On the cat track near L15W, monolithologic felsic breccia is exposed on ledges and flat outcrops. In this case, the matrix and clasts are similar, except that the matrix is more biotitic. U-Pb zircon geochronology on a sample from this outcrop constrains felsic volcanism in the outer volcanic belt to 2722 ± 2 Ma (Fig. 10), within error of the mineralized sequence in the inner belt.

On L15W immediately south of the cat track (6+00N), there are more excellent exposures of both monolithologic and heterolithic breccias. In the first case, coarse (2–10 cm) felsic clasts are supported by a matrix with variable amounts of hornblende and garnet, again as at Stop B4, suggesting preferential alteration of the matrix. The heterolithic breccias contain both felsic and garnet-hornblende-rich clasts, the latter similar to the matrix of monolithologic breccias. The clast types suggest that the heterolithic breccias represent erosion and redeposition of clasts and matrix material of monolithologic breccias, and by implication, that alteration of matrix material predated its incorporation into heterolithic deposits.

Continue west of the grid line about 15 metres to a low west-facing ledge that trends southerly for about 100 metres to Leach Lake and crosses the contact between felsic and mafic rocks. Initially, there is a sequence (north-to-south) from felsic rocks, to about 5 metres of weakly rusty very magnetic silicate iron formation, to heterolithic breccias again with felsic and hornblende-garnet-bearing clasts. The iron formation has about 25–35% garnet up to 1 cm, hornblende, magnetite, minor grunerite and disseminated quartz. Follow the ledge to the south to a sharp contact of felsic rock and mafic schist, the latter typical of most of the mafic metavolcanic unit. Tightly folded tonalite dykes have an axial planar foliation parallel to the schistosity (D_2) of the mafic host rock. The folds and tonalite fabric are interpreted to be D_2 structures.

Stop B6, D_3 (?) fold of iron formation and felsic rock, ZB93-409. On the cat track 30 metres west of L19W, a flat outcrop consists of homogeneous white felsic rock (volcanic or subvolcanic) and a thin (<1 m) iron formation. The iron formation and a strong foliation in the felsic rock are both deformed by an isoclinal fold, interpreted to be D_3 or later, based on folding of the dominant (D_2) fabric. The axial trace of the fold is subparallel to the dominant fabric in the area. On the southern limb, a 20-cm zone of felsic rock penetrates through the iron formation, apparently in an intrusive relationship. A thin (10 cm) concordant strongly foliated tonalite dyke also lies on the fold limb.

Stop B7, Folded zoned gabbroic dyke, garnetiferous mafic rocks, ZB93-336. Outcrops lie near L21W and the cat track, and north on L21W near 6+90N (15 m from track), the latter an excellent pavement. The immediate area is dominated by felsic rock, locally breccia with recessive weathering clasts (up to 10 cm). The felsic rock is cut by a folded (D_2 ?) composite-zoned gabbroic dyke (1.25 m in width). The dyke defines an asymmetric fold in which the northern limb cuts across the local dominant foliation, and the southern limb is generally concordant. Foliation in the dyke is parallel to that in the host. The southern limb shows some odd features; the dyke is apparently repeated against the southern limb in what looks like a sinistral offset. The composite and zoned nature of the dyke are also unusual, as is the presence of an intermediate-mafic dyke in the felsic rocks.

Continue on the cat track about 50 metres west of L21W, crossing the covered contact into mafic rocks. The first exposure is of rusty garnetiferous (25% to 2 cm), locally strongly magnetic, mafic rock. To the west and southwest, the mafic rock is streaked with quartz veinlets and mottled with irregular epidote and bleached patches. Buff leucotonalite dykes containing some mafic inclusions are tightly folded (D_2) and have thickened hinges.

Stop B8, Deformed pillow basalt, ZB93-337. Continue to 10 metres west of L24W at 3+50N (160 m south of the track) to a large stripped outcrop, unfortunately missing over. The fine to medium grained mafic rock has pillows, 0.4–1 metres in length, with rusty, quartz-rich, selvages. A few pillow shapes give hints of northerly younging, but the pillows are strongly deformed and some selvages define local S-folds.

Ten metres west of L24W at 3+10N, on a stripped outcrop, mafic rocks and tonalite dykes define tight folds (D_2) with straight limbs parallel to schistosity. About 15 metres further west, the mafic rock contains minor (<10%) garnet.

Stop B9, Unusual magnetic rocks, ZB93-338. Continue south on L24W to 2+75N to a stripped outcrop about 30 metres west of the line. These are unusual and enigmatic rocks. To the north, about 3 metres of fine grained rusty intermediate rock has thin (<15 cm) dykes of 2 types; irregular boudinaged(?) gabbroic dykes and, hybrid dykes containing 30% tonalitic streaks and patches with hornblende porphyroblasts, in a gabbroic matrix. The textures of hybrid dykes resemble those of magma mixing. To the south, about 6 metres is exposed of very heterogeneous magnetic rock (hornblende-clinopyroxene-bearing) containing 20–80% ovoid or annular dark 1-cm spots in a leucocratic matrix. The spots consist of hornblende-magnetite cores or rings in a felsic matrix. Could these be metaspherulites or metavarioles? Locally, the spotted rock grades (northerly) to something that looks like a fracture breccia cut and veined by dark magnetite-rich material. The dark material is similar to silicate iron formation, containing magnetite, grunerite(?) and minor pyrite and pyrrhotite.

Stop B10, Agam Lake fault, ZB93-420. North of the cat track (100 m) on L27W at 6+60N is a large outcrop area near a stream on the Agam Lake fault lineament. Laminated to lensy felsic rocks immediately south of the stream have intrafolial isoclinal folds, apparently steeply plunging to both west and east, possibly related to ductile movement on the Agam Lake fault. About 15 metres to the south, there is a contact to more

intermediate to mafic rocks which are cut by tonalite dykes that define S-shaped folds (D_2). Schistosity in the host rock and tonalite have an axial planar orientation.

B11–B21. Hinge region of the Manitouwadge synform near Swill Lake

In the Swill Lake area, the outer belt of mafic and felsic metavolcanic rocks is folded in the hinge region of the D_3 Manitouwadge synform (Fig. B3, 1:25000 map). The southernmost unit is a thick sequence of mafic metavolcanic rocks that include homogeneous schist, laminated and layered schist (tuffaceous?), and deformed pillowed flows. As in the Gaug Lake area, foliated gabbroic rocks, interlayered with the fine grained mafic schists, may be massive flows or bases of flows, or sills. High strain zones in metagabbro are characterized by fine grained homogeneous schist and cloud the distinction between tectonic and stratigraphic units. Highly strained pillows are recognized locally and, in one location, deformed pillow shapes suggest southerly younging. In view of the folding in the area, this isolated younging determination does not have regional significance.

Garnetiferous zones (less than 15 metres in width) occur in the mafic rocks, particularly near the northern contact to felsic-intermediate units (compare Stop B1). Locally, garnet is concentrated in pillow selvages, and just west of Swill Lake (Fig. B3), a thick garnetiferous zone is associated with minor exposures of orthoamphibole-bearing rocks. In some cases, garnetiferous zones are discordant to tectonic fabrics or are concentrated along the traces of fold hinges, suggesting synkinematic metasomatism.

Within the thick sequence of mafic rocks, two semi-continuous felsic units are useful markers defining map-scale symmetrical and asymmetrical folds. The southern unit, up to 50 metres wide, is a highly deformed monolithologic felsic lapilli tuff(?), with fine grained to aphanitic felsic clasts in a matrix of mica schist. The unit is associated with thin, discontinuous exposures of iron formation and locally hosts disseminated pyrite and pyrrhotite. North of the mafic metavolcanic rocks, fine grained felsic to intermediate rocks are strongly deformed and commonly laminated. The dominance of hornblende over biotite suggests that they had a metavolcanic protolith. Felsic to intermediate rocks are extensively invaded by foliated tonalite, which dominates in the northeast.

The dominant D_2 planar fabric, well developed in the Swill Lake area, is folded by the D_3 Manitouwadge synform. D_2 folds with axial planar fabrics are preserved locally in the east. There is a dramatic westward increase in post- D_2 strain (Fig. B3), which transposes all earlier fabrics. High strain fabrics, best exposed in the hinge region of the Manitouwadge synform, include spectacular $L>S$ tectonites and folds with a moderate northeasterly plunge, parallel to lineations. The tectonites may, in part, be the result of transposition of D_3 structures during deformation focussed on the western contact of supracrustal rocks with the Black Pic batholith.

Two coarse grained, plagioclase-megacrystic diabase dykes cut through the area (Fig. B3). The northeasterly-trending dyke is well exposed and probably belongs to the Biscotasing swarm of about 2167 Ma (Buchan et al., 1993). The location of the northwesterly-trending dyke is inferred from one outcrop exposure and aeromagnetic and topographic trends. It probably belongs to the Matachewan swarm of about 2454 Ma (Heaman, 1988).

Claims in the area are held by Noranda Inc. in the northwest, Al Turner in the central area, and Granges Inc. in the east. Access is by old logging roads north of the Caramat road leading to extensive areas of variably overgrown clear cut. Deep streams and washouts on most of the main logging roads severely limit vehicular access. Some outcrops are located using a cut grid (metric units) on the Turner claim group. To reach the Swill Lake area, follow the Caramat Road west from Highway 614 approximately 7.5 km and turn north on a logging road just beyond the landfill-site entrance on the south side of the Caramat Road. The first washout at 0.8 km may or may not be passable; if the road is wet, the slope on the far side of the washout will probably be greasy mud. Continue north along the road.

Stop B11, Folded mafic metavolcanic rocks, ZB93-2, ZB93-P50. At 750 metres north of the first washout, several outcrops lie on a small knoll to the east, beginning approximately 80 metres from the road. The mafic metavolcanic rocks comprising laminated mafic gneiss, mafic schist and foliated gabbro, are typical of the southern Swill Lake area. Structures include S-folds with associated crenulation cleavage (D_3), a possible D_2/D_3 fold interference pattern, ductile shear zones in the foliated gabbro with indeterminate (possibly dextral) shear sense, and brittle fractures, typically filled with epidote and having dextral offset in plan view.

Stop B12, Felsic breccia, ZB93-P175. Continue another 400 metres northward along the road and then northeast across a clear cut for 200 metres to the large light-coloured outcrop at the base of the cliff clearly visible from the road. This is an excellent exposure of the southern felsic unit, a highly deformed but well preserved monolithologic felsic breccia. A geochemistry sample from this outcrop, prepared by extracting clasts and carefully avoiding the high angle brittle fractures, shows the unit to be a calc-alkaline rhyolite, in contrast to the tholeiitic basalts to north and south. There is a fine, streaky lineation on foliation surfaces, and small Z-folds are common across the outcrop, although at least one S-fold was also observed.

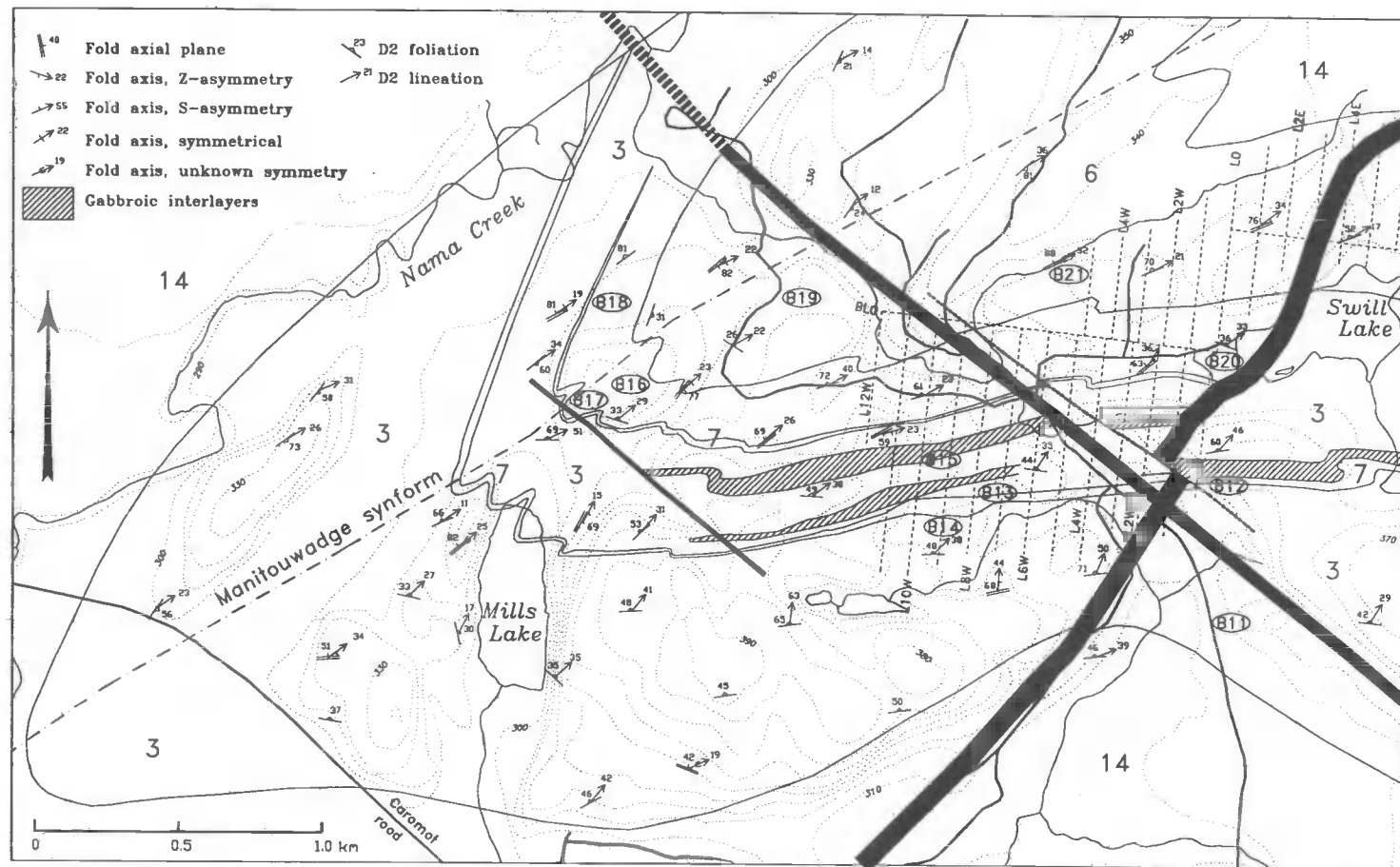


FIG. B3. Geology and field-trip stops (B11-B21) of the Swill-Mills Lakes area near the hinge region of the D₃ Manitouwadge synform. Dashed lines are the cut lines of A. Turner's Swill Lake grid.

Stop B13, Felsic breccia, stretching lineation, ZB93-P71, P72. Return to the road and continue another 200 metres, then follow a drill road branching to the southwest. After about 100 metres, take the northern fork of the drill road and continue westerly 350 metres to a low open area where several tracks branch off and the drill road intersects L7W of the Turner grid. Continue northerly (010°) along L7W. At 6+00S and about 50 metres to the east, the southern unit of monolithologic felsic breccia crops out. Note the pronounced stretching lineation. The northern end of the outcrop area is mafic schist. Just west of the line at 6+00S, the felsic unit, here containing some garnet (rimmed by plagioclase) and with clasts less distinct, is interlayered with rocks of more intermediate composition. The northern contact with mafic schist is exposed.

Stop B14, Deformed pillows, sulphidic felsic rocks, ZB93-P77-P78, P82. Continue westerly 200 metres, following the crude drill road to east-west-trending outcrops along north side of an open area just west of L9W. These are the best preserved pillows that we have seen in the Manitouwadge area, some with cusps suggesting southerly younging, although of questionable regional significance. The pillows are zoned from dark hornblende-rich selvages to lighter cores.

Walk 50 metres north to a trenched area of rusty sulphidic (mostly pyrite) rocks grouped with the southern felsic unit. The very fine grained siliceous rocks have some layers with plagioclase, biotite and hornblende. In a second trench, visible to the northwest, felsic rock is intruded by foliated leucocratic tonalite.

Continue northeast from the trench, finding L9W and an outcrop that crosses the line near 6+50S. The exposure shows a contact between very highly deformed (mylonitic), garnet-bearing felsic breccia and garnet-rich mafic schist.

Continue west to L10W, 6+50S and another exposure of the felsic breccia of the southern felsic unit, in this case, represented by a ribbon mylonite. Strongly elongate clasts are still recognizable defining a lineation that plunges moderately to the northeast. As previously, garnet tends to be rimmed by plagioclase.

Stop B15, High strain zones in metagabbro, ZB93-P81. Follow L10W northward to an outcrop west of the line at 6+00S, a beautiful example of metagabbroic layers in the thick mafic sequence. In the transition to high strain zones, the gabbro is medium to coarse grained, foliated and contains hornblende porphyroclasts. In zones of high strain, the textures and fabrics are similar to those of the fine grained homogeneous mafic schist common throughout the area, suggesting that some of the homogeneous mafic schist may be highly strained metagabbro.

Stop B16, Symmetrical M-folds in the hinge of the Manitouwadge synform, ZB93-P151, P135. Continue north on L10W about 200 metres until it intersects a logging road. Note in passing the mafic metavolcanic rocks and metagabbro along the way. Follow the road west for 700 metres to where the main road makes a broad turn to the north and a side road continues westward. Follow the side road for about 60 metres and then take the right-hand northwesterly (less travelled) fork for 300 metres to a small pavement on the south side of the trail. The first outcrop, intended as very brief stop, is dominantly homogeneous fine grained, mafic schist (highly strained gabbro?) with an inclusion of layered quartz-magnetite iron formation, both cut by foliated granite.

Continue west along the track for 100–120 metres to an outcrop area extending south of the road. The area lies in the hinge region of the Manitouwadge synform. Garnetiferous mafic metavolcanic rocks are interlayered with rocks of more intermediate composition. The layering and dominant foliation (D_2) are extensively folded, mostly into symmetrical M-shaped D_3 folds, although some S-folds occur at the northern end of the outcrop area. Some layers have abundant coarse (cm-scale) magnetite porphyroblasts.

Stop B17, Felsic rocks, D_3 folds and strong lineation, ZB93-P134 (optional). Continue 150–170 metres west of Stop B16 and 30–40 metres south of the track, climbing over logs and brambles to find the outcrop. Felsic metavolcanic rocks belonging to the northern felsic unit of the Swill Lake area are intercalated with more intermediate hornblende-bearing rocks. Tight folds (D_3) of the dominant foliation (D_2) plunge moderately to the northeast, parallel to a very strong stretching lineation.

Stop B18, D_3 folds of metavolcanic rocks and granite, ZB93-P154, P155, ZB94-84. Return east (about 550 m) to where the main road turns north. Continue northward for 220 metres to a westerly branching track. Follow the track for 150 metres to the top of a hill and extensive outcrops of strongly lineated and folded, plagioclase-hornblende-bearing intermediate metavolcanic rocks. From the hill top, continue northerly to the base of the steep north slope, then head westerly (about 300 m) in the low cleared area, keeping the slope and log piles to your left. You are heading for the top of a low hill, on the southeast side of which lies a small cleared pavement visible in the distance. The hill is about 200 metres north of a long wood pile.

Intermediate metavolcanic rocks are dominant, thinly interlayered with magnetite-rich mafic rocks. Many tight folds (D_3) reorient the foliation (D_2) and plunge northeasterly, parallel to a very strong stretching lineation. At the southern end of the hill top, and 30–40 metres west, mafic metavolcanic rocks and sheets of foliated granite are both involved in outcrop-scale M-folds. The mafic rock has a strong folded schistosity

(D₂), and a moderately developed axial planar fabric. The granite forms sheets parallel to the D₂ schistosity of the host rock and its most obvious fabric is axial planar to the D₃ folds.

The granite, interpreted as a pre-D₃ intrusion, was collected in an attempt to bracket the age of D₃ deformation by U-Pb geochronology. Analyses of zircon gave a scatter of discordant points, whereas titanite defines an age of 2655 ± 3 Ma (Table 2), within error of the age of titanite from a syn-D₃ tonalite dyke less than a kilometre away, and also within error of titanite from the Loken Lake pluton. The circa 2655 Ma titanite ages are interpreted as dating a late hydrothermal event that crystallized or recrystallized titanite.

Stop B19, Lineation in intermediate metavolcanic rock, ZB93-P157 (optional). Retrace the route eastward to the main road. Continue northward along the main road for 200 metres to a large wood pile and an easterly branching side road. Follow the side road for 70 metres to a small clean outcrop on the south side of the road. Thin streaky hornblende-rich and felsic layers, and the dominant foliation (D₂), define small folds (D₃) with a weak Z-asymmetry. A beautiful lineation, strongly developed, plunges northeasterly parallel to fold axes.

Stop B20, Garnetiferous mafic rocks, diabase, ZB93-4, P54. Return (south and then east) along the main road for about 1.5 km, crossing a narrow deep washout and stream just before intersecting a major northwest-trending logging road. This is the road on which the vehicle(s) are parked somewhere (depending how many washouts were negotiated) to the south. Cross the intersection with the main road, and follow the road leading uphill and northward, also identifiable by the deep long gully that has been eroded into it. After 150 metres, the road curves to the east. Continue to the end of the road where it branches to the east and south, 150 metres west of Swill Lake. The southeast branch leads to a trench 50 metres distant. The stripped outcrop area and trench are dominated by dark mafic metavolcanic rocks with garnetiferous zones. Garnet is most abundant in seams of 10–30 cm width, that anastomose through the dominant foliation while maintaining a generally concordant orientation. Grain size ranges from pinhead to 1–2 cm in diameter, and shape from equant to lensoid, parallel to foliation. In thin section, δ -type tails indicate a north-side-up shear sense with a slight sinistral component of motion (also discernible on outcrop). The abundance of garnet implies metasomatic alteration of the mafic protolith; however, relationships between tectonic fabrics, garnet porphyroblasts and anastomosing garnetiferous seams are most consistent with synkinematic alteration and garnet crystallization.

Walk southeast to the shore of Swill Lake. Water-washed outcrops in the vicinity of the beaver dam are dominated by coarse grained diabase with megacrysts of plagioclase, one of many exposures of the wide northeast-trending dyke that cuts across the Swill Lake area.

Return to the branch in the road and, just west of the branch, head south on L1W to a small outcrop 25 metres west of the line at 0+75S. Dark fine grained hornblende-plagioclase schist with minor garnet is intercalated with lighter-coloured orthoamphibole-cummingtonite-plagioclase-garnet-biotite schist. This is one of the few orthoamphibole-bearing rocks in the Swill Lake area and suggests a link with similar orthoamphibole-bearing rocks to the east near Manitouwadge Lake (e.g. Stop B1).

Stop B21, Intermediate metavolcanic rocks, ZB93-P179. Return 50 metres to the road and head west, then follow L5W north to 2+25N and several outcrops scattered about 35 metres west of the line. The intermediate metavolcanic rocks vary from thinly layered to laminated and, are locally quite felsic. The U-shaped fold in the northwestern part of the outcrop area may be a D₂ fold, based on the observation that it folds layering, whereas the dominant foliation appears to be axial planar. An interesting dyke, with straight sharp margins, cuts across foliation in the southern part of the outcrop area. It contains pillow-like, medium grained, hornblende-rich mafic blobs in a leucocratic matrix dominated by microcline and plagioclase. The dyke can be traced for about 200 metres to the east-northeast and there are similar exposures south of Swill Lake.

C. Inner Manitouwadge synform, Dead Lake suite

The Dead Lake suite comprises a complex association of interleaved foliated gabbro, diorite, and layered mafic to intermediate rocks of probable supracrustal origin collectively grouped with mafic to intermediate metavolcanic rocks of Unit 5. The suite can be mapped in at least two nearly continuous zones within the Manitouwadge synform corresponding to high variable aeromagnetic relief (1:25000 map). It includes a group of distinctive homogeneous or layered, strongly magnetic rocks characterized by hornblende-magnetite $\pm$ plagioclase $\pm$ garnet $\pm$ clinopyroxene $\pm$ sulphide minerals with variable amounts of quartz (minor to 50%), commonly in eyes or lenticles. Layered varieties, in particular, resemble metamorphosed iron formation or ferruginous chert; however, their geochemical composition (high FeO₄, TiO₂ and Zr) suggests sedimentary concentration of heavy minerals.

The extent of synvolcanic trondhjemite and comagmatic rocks in the inner Manitouwadge synform is not easily answered on the basis of outcrop and petrographic observations. At least some tonalite and granodiorite

intruding the Dead Lake suite is texturally and compositionally similar to synvolcanic trondhjemite (Figs. 23 and 24). However, most quartz-rich tonalites in the area have a higher content of mafic minerals, including hornblende, than those near the Willroy-Geco area. Some of the hornblende, present as coarse grained (0.5–1 cm), randomly to moderately oriented poikiloblasts, appears secondary.

The mafic rocks of the Dead Lake suite are tholeiitic basalts, geochemically very like mafic rocks in the inner and outer volcanic belts, with which we regard them to be correlative. The two zones of the suite are indistinguishable and could represent a structural repetition. The structural relationship of the suite to the main supracrustal sequence in the Manitouwadge belt is best explained by some variation on a model involving major D₂ folds. One possibility is that the entire sequence between the Loken Lake pluton and the Black Pic batholith is folded by the D₂ 'Manitouwadge syncline', the axial trace of which lies in metasedimentary rocks on the southern limb of the D₃ Manitouwadge synform (Fig. 5). This model implies that the Dead Lake suite is equivalent to the lower stratigraphic levels of the outer volcanic belt, and that trondhjemite pinches out in the subsurface between the two limbs of the fold. Trondhjemite might be represented in the outer belt by some of the thin concordant units of tonalite, for example, striking through Gaug Lake (1:25000 map). A second possibility is that the two zones of the Dead Lake suite represent a D₂ synclinal 'keel', analogous to the 'keel' of the D₂ 'Manitouwadge syncline' (Figs. 5 and 8). This implies that the Dead Lake suite is equivalent to metavolcanic screens included in trondhjemite in the Willroy-Geco area (and north to Rabbitskin Lake) and the presence of a D₂ anticlinal trace within the trondhjemite.

The Loken Lake pluton, occupying the innermost area of the Manitouwadge synform, is a biotite tonalite to granite characterized by microcline megacrysts, 5 to 15 cm long. It has a strong tectonic fabric, commonly represented by L>S tectonites. The abundance of megacrysts varies from sparse or none to 25%, although near the contacts (based on aeromagnetic signature and mapping), it is usually possible to find at least one megacryst depending on the size of the area exposed. Large areas of non-porphyrific rock, mainly more central to the pluton, would otherwise be difficult to distinguish on outcrop from synvolcanic trondhjemite. However, the rocks are geochemically distinct, and the Loken Lake pluton is significantly younger at 2687±2/–3 Ma (Fig. 16). Layering and the dominant foliation (D₂) in the Dead Lake area, including foliations in the Loken Lake pluton, are folded about an east-southeasterly plunging axis (110°/25°), interpreted to be the axis of the D₃ Manitouwadge synform. On this basis, the Loken Lake pluton was interpreted to be a pre- to syn-D₂ intrusion.

The Dead Lake area is reached by taking a logging road north from the Camp 70 road east of Manitouwadge. The Camp 70 road intersects the Geco mine road 2 km east of town, and 1 km further, crosses a lumber yard and railway line. Follow the Camp 70 road for about 9 km from the railway line, crossing the bridge across the Black River, passing a sign-posted turn for Twist road and a sand pit on the north, the latter about 1.7 km before the logging road. The logging road also leads to a garbage dump and you will know you made the right turn by the garbage on the road and to the west. About 300 metres north of the turn and on the north side of the dump, a dirt road that branches west to Wowun Lake leads to Stop D1. Drive on past the branch road and keep track of your distance from it. Watch out for probable washouts, hopefully negotiable with care.

Stop C1, Loken Lake pluton, ZB94-27. Continue on the logging road for 6 km, winding around east and north of Wowun Lake (not visible), making no turns off the main road (1:25000 map). The stop is a low rocky knoll on the north side of the road, immediately north of the water-filled ditch. The location is 400 metres east of a railway crossing (which you will not see unless you overshoot). In this exposure, the Loken Lake pluton contains about 5% megacrystic microcline augen, of about 10 cm in length, in a tonalitic matrix with minor biotite and trace amounts of hornblende. The strong southeasterly plunging lineation (L>S, composite D₂/D₃?) is partly defined by asymmetrical (both σ - and δ -type, Hanmer and Passchier, 1991) tails on microcline augen. Most of these are consistent with dextral, north-side-down, sense of rotation, although some show sinistral kinematics. The conflicting rotations could be due to variations in the original orientation of long axes of phenocrysts.

Stop C2, High strain zone, contact to Dead Lake suite, ZB93-360, ZB94-30. Cross the railway at 6.4 km and continue on to 8 km stopping at a pavement outcrop that extends from the north side of the road. This is the first exposure of the Dead Lake suite, here represented by interleaved white quartzofeldspathic, intermediate (30–40% hornblende+biotite), and mafic (60–70% hornblende) layers, 0.5 to 30 cm in width. Abundant magnetite, either fine grained or porphyroblastic to 3 mm, is concentrated in some layers. The general appearance is that of highly recrystallized and coarsened mafic to intermediate volcanic rocks with tonalite sheets.

Cross to a rocky ledge (3–4 m high) visible about 50 metres away on the south side of the road. The leucocratic tonalite, with a very strong foliation and some streaky quartz ribbons, could be a highly strained equivalent of either synvolcanic trondhjemite or a non-porphyrific variety of the Loken Lake pluton. The presence of pinhead garnet and magnetite in some layers suggests the former.

Stop C3, Dead Lake suite, quartz-garnet-magnetite-hornblende rocks, ZB93-361, ZB94-108. Continue to where the road bends to the southwest (8.3 km) just past the top of a small hill, and down hill to a gravelly area west off the road at 9 km. The stop is 150 metres further along the road on the first knoll on the west side, just beyond a swampy area. Several rock types are exposed including; 1) strongly magnetic, medium to coarse grained, garnet-magnetite-hornblende-plagioclase rock, locally with up to 50% quartz and minor sulphides and titanite, 2) fine grained hornblende schist with local biotite porphyroblasts, 3) fine to medium grained, weakly foliated, white tonalite with 2-5% disseminated magnetite, and 4) a discordant pegmatite dyke. In some places, quartz-garnet-magnetite-hornblende rock forms a transitional zone along the contact between hornblende schist and more leucocratic rocks. The orientation of the dominant fabric (D_2) is somewhat variable ($038-075^\circ/18-34^\circ$), and biotite porphyroblasts in hornblende schist look to be defining a weak oblique foliation.

From the south end of the outcrop, climb up the hill to several outcrops along its crest about 80 metres west of the road. The rock types are similar to those listed above, with more quartz-garnet-magnetite-hornblende rock, locally cut by weakly foliated tonalite and forming angular inclusions in strongly foliated tonalite (trondhjemitic?). Biotite porphyroblasts and selvages in hornblende schist are developed mostly near contacts with tonalite. Tight outcrop-scale folds of layering might include both D_2 and D_3 folds, the former with axial planes parallel to the dominant foliation (D_2), and the latter with a weak axial planar fabric developed only in biotite selvages at tonalite-hornblende schist contacts.

Stop C4, Dead Lake suite, layered quartz-garnet-magnetite-hornblende rocks, ZB94-101. Continue south on the road, passing a junction at 9.6 km with a road heading uphill to the west and, at 9.9 km, reaching a muddy flooded stretch which may not be drivable depending on water levels, vehicles and drivers. The next stop, a long (150 m) pavement outcrop along the west side of the road, is 450 metres further on and worth a short hike. The outcrop is dominated by fine grained homogeneous foliated biotite tonalite with some thin (5 cm) concordant aplitic layers. On scattered exposures at the southern end of the outcrop area, it is possible to trace a very magnetic 1-2 metre layer through a series of complex folds. The magnetic layer consists of quartz, plagioclase, magnetite, hornblende, epidote, garnet and titanite. The proportions of these minerals define second-order (cm-scale) layering and grading that looks like modified bedding and suggests a sedimentary protolith. Locally, the magnetic layer is cut by diorite and pegmatite. Geochemical analysis of a sample of the magnetic layer shows that, in addition to high FeO_4 (18.0%) and CaO (8.31%), it is relatively high in TiO_2 (1.36%) and Zr (2900 ppm). TiO_2 and Zr are typically very low in chemical precipitates, their presence being normally attributed to a detrital or volcanic component. However, the high values here may be more suggestive of sedimentary concentration of heavy minerals. To add to the enigma, preliminary interpretation of geochemical analyses of quartz-garnet-magnetite-hornblende rock, such as that seen at Stop C3, and hornblende-bearing trondhjemitic, suggest that these represent synvolcanic trondhjemitic contaminated or altered(?) by a component similar to the magnetic layer here.

Stop C5, Dead Lake suite, metagabbro, trondhjemitic, diorite relationships, ZB94-3, ZB94-229. Continue along the road 800 metres beyond the mud bath (10.7 km cumulative distance if you are still in your vehicle), to another extensive outcrop in the ditch on the west. Among dioritic and tonalitic rocks similar to those seen at previous stops, a very dark mafic variety of magnetic hornblende-magnetite-plagioclase-garnet rock contains only minor amounts of quartz. Locally, the magnetic rock is veined by fine and coarse grained felsic rocks, the latter resembling synvolcanic trondhjemitic.

Continue south for another 300 metres (11 km cumulative) to another extensive outcrop about 100 metres long and stretching west from the road for 50 metres. The rock types include; 1) medium to coarse grained metagabbro varying from 80-90% hornblende to plagioclase-spotted (30% to 5 mm, phenocrysts? or blasts?), 2) fine to medium grained magnetic hornblende-magnetite-garnet rock also locally plagioclase-spotted (up to 50%), 3) fine to coarse grained trondhjemitic, 4) fine grained foliated diorite with 50% hornblende and minor biotite porphyroblasts, and 5) aplite/pegmatite. The metagabbro is interpreted as the oldest rock; it is cut by trondhjemitic and both metagabbro and trondhjemitic are engulfed by foliated diorite. Diorite and trondhjemitic contain local inclusions of magnetic rock. Note that the metagabbro locally has nearly 5% brown spots (1-3 mm) of titanite, which in thin section, looks like wormy pods and aggregates. Although geochemically the metagabbro is a tholeiitic basalt, it has relatively high TiO_2 (2.80%) and anomalously low Zr/ TiO_2 .

Stop C6, Foliated granodiorite, synvolcanic?, ZB94-233. Continue to the south along the road for 1150 metres (12.15 km cumulative), just 300 metres north of the hairpin bend at the southern end of the road. An outcrop area stretches west of the road for 50-60 metres, dominated by coarse grained homogeneous foliated granodiorite with disseminated magnetite porphyroblasts (up to 4 mm in size) and minor biotite. The strong foliation (D_2) undulates in small folds and shears. The rock is representative of exposures south from here and, in texture, and geochemical and modal composition, resembles synvolcanic trondhjemitic, except for the presence of 10-15% microcline.

D. Eastern extension of the 'Geco horizon'

The easternmost exposure, of what is locally called the 'Geco horizon', crops out east of Wowun Lake near the Hucamp zone of subeconomic mineralization (1:25000 map). Straight gneiss in this exposure is interpreted to lie on the continuation of the D₁ fault that dissects the Willroy-Geco area (Stops A3, A6 and A20). East of the Hucamp zone, exposure is poor, but a prominent aeromagnetic anomaly continues on the same trend, and the 'Geco horizon' was intersected in drill holes extending east to the Falconbridge zone of subeconomic mineralization. Orthoamphibole-garnet-cordierite gneiss, mafic rocks and minor iron formation east of Banana Lake (Noranda's East One Otter and Banana claim groups, Stops D3–D4) are interpreted to be correlative with the 'Geco horizon', but their structural relationship is subject to interpretation. They are involved in a series of map-scale folds, that also deform the dominant foliation (D₂), with a northerly trending enveloping surface. In a preliminary interpretation, we correlated the East One Otter and Banana zones with the Falconbridge zone, across a fault with sinistral offset (Zaleski and Peterson, 1993b; Peterson and Zaleski, 1994a). This model has ramifications to regional interpretation, as it implies that the northern limb, axial trace and the inner volcanic belt of the southern limb (including the 'Geco horizon') of the Manitouwadge synform (D₃) are refolded by the Blackman Lake antiform and the Jim Lake synform, making these latter D₄ folds (see Structural Geology). Several elements of the preliminary model proved to be problematic, among them the postulated sinistral fault, requiring 2.5 km of map-view displacement, for which we could find neither field nor aeromagnetic evidence. In our preferred revised model, the East One Otter and Banana zones represent a D₂ synclinal 'keel', involved in a D₂/D₃ fold interference pattern along the axial trace of the D₃ Manitouwadge synform (Fig. 5, 1:25000 map). This model allows for a simpler more elegant interpretation of the regional structure in which the major folds that determine the map pattern of the belt (Manitouwadge synform, Blackman Lake antiform, Jim Lake synform) are all D₃ folds.

The first two stops (D1–D2) are accessed from the logging road on the east side of Wowun Lake, the same road that was used to access the Dead Lake suite (Stops C1–C6). The last two stops (D3–D4) are accessed by continuing east on the Camp 70 road for 17 km beyond the Dead Lake logging road.

D1–D2. Eastern inner volcanic belt, 'Geco horizon'

Stop D1, Easternmost surface exposure of the 'Geco horizon', ZB93-P16. Take the Camp 70 road and turn north on to the Dead Lake logging road (see C. Dead Lake suite, for more detail), after 300 metres, turn west off the Dead Lake road on to a dirt road around the north side of a garbage dump. Drive on 700 metres until a pavement outcrop is just visible through a strip of bushes on the south side of the road (1:25000 map).

On the west side of Wowun Lake, an attenuated sequence comprising (north to south); trondhjemite (synvolcanic), orthoamphibole-garnet (or cummingtonite-garnet) and sillimanite-cordierite-biotite-garnet gneisses, iron formation and metagreywacke, crops out over an interval of about 200 metres. Here, on the east side of Wowun Lake, near the Hucamp zone, the 'Geco horizon' is marked by (north to south); 1) felsic straight gneiss containing some elongate garnet, 2) a thin (2 m) straight streaky-layered highly strained iron formation, 3) <5 metres of a highly strained quartz-eye rock (possibly quartz-phyric felsic metavolcanic rock) with fine grained garnet-magnetite in the matrix, and 4) rusty felsic rock. We correlate this high strain zone with the interpreted D₁ fault in the Willroy-Geco area. Throughout the outcrop, there are concordant foliated plagioclase-porphyritic tonalite dykes, probably belonging to a syn-D₂ tonalite suite. The dykes, although highly strained, look less so than the host rocks. Note the southerly dip of the dominant foliation (D₂); here, the southern limb of the Manitouwadge synform is overturned.

Stop D2, Eastern continuation of mafic rocks of the inner volcanic belt, ZB94-60. Return to the Dead Lake logging road, turn north and continue for 650–700 metres to an extensive outcrop on the east side of the road. The southern part of the outcrop, coarse grained homogeneous foliated trondhjemite with local magnetite and biotite, is typical of synvolcanic trondhjemite. To the north, trondhjemite is interleaved with mafic rocks in a transitional contact zone to fine grained mafic and laminated schist, the latter correlated with mafic metavolcanic screens in trondhjemite to the west. The mafic schist locally contains plagioclase porphyroclasts(?), possibly derived from dismembered veins, and biotite porphyroblasts, apparently defining a weak near-horizontal foliation. As at Stop D1, layering and the dominant foliation (D₂) dip moderately to the south.

D3–D4. Orthoamphibole-garnet-cordierite rocks, Banana area

The Banana and East One Otter grid areas are dominated by multiphase granitic to tonalitic rocks with variably developed (mainly weak) fabrics. At least some intrusive rocks contain abundant (1–5%) disseminated magnetite porphyroblasts and, texturally and compositionally (Figs. 23–25), resemble synvolcanic trondhjemite north of the Willroy-Geco area. Supracrustal rocks, including orthoamphibole-garnet-cordierite rocks, mafic metavolcanic rocks and iron formation, are present as screens and inclusions defining a narrow (about 50 m) zone of map-scale folds related to the D₃ Manitouwadge synform (Fig. D1). In our preferred

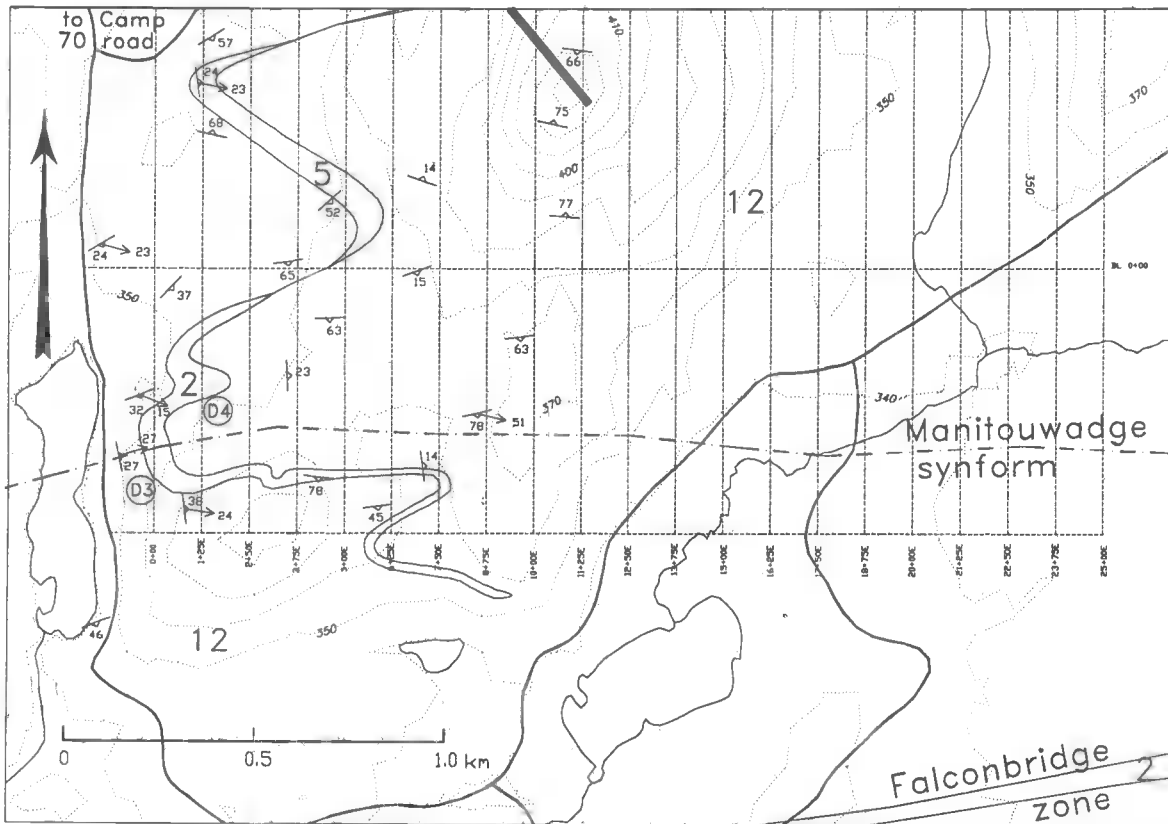


FIG. D1. Geology and field-trip stops east of Banana Lake along the eastern trace of the D₃ Manitouwadge synform. Dashed lines are cut lines of Noranda's Banana grid. Structure symbols show dominant D₂ foliations and lineations.

structural model, the zone follows the folded trace of a D₂ synform (Fig. 5). Folded fabrics, interpreted as D₂, are easiest to see and measure within the supracrustal screens. The area was the focus of mineral exploration; however, a drill hole in the East One Otter zone intersected only minor disseminated sulphide mineralization, and geophysical surveys of the Banana zone failed to show any conductive anomalies.

From the junction with the Dead Lake logging road, follow the Camp 70 road for about 17 km east and north, turning to the east at a junction east of Thompson Lake on a gravel road sign-posted for Hillsport. Continue easterly for about 3 km, taking a dirt road to the south where the main road bends sharply to the north. Continue southerly for 3 km, in part along the eastern shore of a lake (which might be flooding the road), until a cut line (running down the hill to the east) intersects the road. The line is the 7S tie line of the Banana grid. The following locations are referenced to the grid (metric units).

Stop D3, Mafic, orthoamphibole- and cummingtonite-bearing rocks, ZB93-313-315. Follow TL7S to the east uphill for 160 metres to L0, and L0 to the north to some rocky knolls and ledges from 6+00S to 3+75S on both sides of the grid line (Fig. D1). Over this distance from south to north, the sequence; trondhjemitic, orthoamphibole-bearing rocks, hornblende-biotite±cummingtonite mafic rocks, trondhjemitic, is exposed. The trondhjemitic is medium to coarse grained, contains magnetite porphyroblasts and is geochemically indistinguishable from the same unit in the Willroy-Geco area. It contains inclusions of dioritic-looking fine to medium grained metabasite (about 50% hornblende-biotite) that may be coarsened metavolcanic rock. Granitic rocks with associated pegmatite are also common, apparently representing younger(?) intrusions. At 5+65S, there is a contact to relatively leucocratic rocks with coarse sprays (to 5 cm) of orthoamphibole (or cummingtonite?), and locally biotite, in a quartz-plagioclase matrix. Locally, orthoamphibole and biotite define a strong easterly plunging lineation. The orthoamphibole rocks strike subparallel to the grid line (0°) defining the hinge of an open easterly plunging fold which we interpret as the hinge region of the Manitouwadge synform. Orthoamphibole-bearing rocks can be traced to the east to orthoamphibole-garnet-cordierite rocks on the limbs of the fold. At 5+00S about 20 metres east of the line, trondhjemitic and granite-pegmatite

with abundant mafic (hornblende±cummingtonite) screens become dominant in the core of the fold until 3+80S. At 4+15S, a sample of fine to medium grained mafic schist (about 60% hornblende-cummingtonite) has the major-element composition of a tholeiitic ferrobalt, in contrast to mafic metavolcanic rocks from the supracrustal sequence to the west, which are high and low Mg tholeiites. The high FeO₄ and presence of metamorphic cummingtonite may be due to alteration. At the base of the rocky hill at 3+80S, cummingtonite(+orthoamphibole?)-hornblende schists on the northern limb of the open fold strike at 070° and dip southerly underneath plutonic rocks, defining the northern limb of a synform.

Stop D4, ZB93-287, ZB94-93. Follow the trend (070°) of rocky ledges and knolls for about 130 metres to L1+25E, looking at the extensive exposures of orthoamphibole-garnet-cordierite rocks. Plutonic rocks form the highest knolls to the south. The orthoamphibole-garnet-cordierite rocks are coarse grained (locally with garnet to 10 cm and coarse orthoamphibole sprays) and heterogeneous with garnet-rich and orthoamphibole-rich layering or enclaves (e.g. just west of L1+25E, 3+00S). Garnet textures and distribution are fascinating, ranging to very coarse (to 10 cm) garnet porphyroblasts to fine grained garnetite, and local garnetiferous 'snakes' in an orthoamphibole-rich matrix. Locally, orthoamphibole-garnet-cordierite rocks are cut by irregular white veinlets (1-4 cm wide) characterized by quartz-sillimanite-garnet-cordierite. The margins to the host rock are diffuse and, although the veinlets have no systematic orientation and are discordant to the dominant foliation, the dominant foliation is found in both host rock and veinlet. The relationships suggest that the veinlets are the result of local channelling of metamorphic fluids.

Orthoamphibole-garnet-cordierite rocks do not continue east of L1+25E beyond 40 metres; foliation trends indicate the presence of an antiformal hinge (Fig. D1). At 2+75S, on L1+25E and 20 metres to the west, plutonic rocks contain 2 layers (1 about 5 m thick) of an unusual strongly magnetic rock consisting of abundant (50%) quartz eyes in a matrix of fine grained magnetite, hornblende, clinopyroxene, garnet and minor plagioclase. The quartz eyes are monocrystalline, flattened and strained; locally, they are sufficiently abundant to coalesce into layers. The rock closely resembles quartz-garnet-magnetite-hornblende rocks of the Dead Lake suite, and possibly originated as a contaminated or altered quartz-phyric metavolcanic rock or trondhjemite. The quartz-eye magnetite-garnet rock tends to crop out to the east of orthoamphibole-garnet-cordierite rocks in the Banana area, including immediately to the east of the last exposure of orthoamphibole-garnet-cordierite rocks in the antiformal hinge.

Orthoamphibole-garnet-cordierite rocks also crop out near L2+50E, 5+40S and L3+75E, 5+50S along the southern limb of the open fold. The road can be regained either by returning along tie line 7S or by continuing to base line 0 which also extends westerly to the road. Either of these routes crosses extensive outcrops of multiphase plutonic rocks, mainly magnetite-bearing trondhjemite and granite-pegmatite.

E. Quetico subprovince, Jim Lake synform

The Quetico subprovince in the Manitouwadge area is characterized by migmatitic biotite schists and metagreywackes, similar in texture, composition and depositional age, to those in the Manitouwadge belt (1:25000 map). The main difference is in metamorphic grade; rocks of the Quetico subprovince have been metamorphosed to granulite facies just north of the subprovince boundary and are extensively migmatized. Typically homogeneous or layered, they comprise biotite-plagioclase-quartz±garnet±cordierite±sillimanite schist with extensive concordant, folded and cross-cutting tonalitic segregations. The dominant planar fabrics are interpreted to be correlative with D₂ or D₃ structures in the Manitouwadge belt. Northeasterly plunging Z-shaped folds of foliation, with a moderately developed, northeasterly striking axial-planar cleavage (in some cases, crenulation cleavage) are interpreted to be due to D₃ deformation. Locally, the folds are associated with outcrop-scale structures (e.g. foliation fish) indicating dextral oblique shear. The D₃ structures deform, and are cut by, migmatitic layers and veins, suggesting that D₃ was approximately synchronous with peak metamorphism in the Quetico subprovince.

The D₃ Jim Lake synform is defined by mappable zones of mafic to intermediate metavolcanic screens and inclusions, with minor iron formation, enclosed in foliated to weakly foliated tonalite. Quetico metagreywackes can also be traced around the synform and along its southern limb as far as Larry Lake (1:25000 map). The Jim/Davis Lakes area lies near the hinge region, although most of the exposure lies along the southern limb and, in particular, near a map-scale Z-fold apparently parasitic to the Jim Lake synform. Quetico metasedimentary rocks in the Jim/Davis Lakes area are represented by migmatitic biotite schist with local thin iron formation. On the inside of the fold, northeast of Davis Lake, coarse grained trondhjemite to granodiorite resembles synvolcanic trondhjemite (Unit 12) in texture and composition (Figs. 23-25). South of the southern limb, foliated microcline porphyritic granitoid belongs to the Nama Creek pluton, which forms a sinuous body between supracrustal rocks and the Black Pic batholith folded by the Blackman Lake antiform. The Jim Lake synform is a tight fold and early fabrics are largely transposed and subparallel to the axial surface; however locally, dominant fabrics (D₂) are folded around the hinge region and L-tectonites have a strong stretching lineation parallel to the axes of D₃ minor folds.

The eastern part of the Quetico subprovince in the Manitouwadge area can be accessed from the Hillsport road (continuing from Stops D3–D4). The southern limb of Jim Lake synform in the vicinity of Jim and Davis Lakes is reached by continuing to the north on the Camp 70 road beyond the Hillsport junction east of Thompson Lake.

E1–E3. Quetico subprovince, eastern Manitouwadge area

Stop E1, Z-folds and kinks, ZB93-P293. From the Banana area, return 3 km north to the Hillsport road where it curved sharply north (1:25000 map). Continue north and then northeast for 4.5 km, turning right (southeast) on to a gravel road. On the north side of gravel road, 1.1 km from the intersection, a pavement of metasedimentary rocks shows well developed migmatitic layering parallel to a fine biotite foliation (D_2/D_3). The dominant planar fabrics are parallel to the axial surfaces of folds with Z-asymmetry and, in one place, folded by a kink (D_4), also with Z-asymmetry. A coarse grained cordierite-bearing segregation cuts discordantly across the outcrop.

Stop E2, Dextral kinematic indicators, ZB93-P294. Continue southeast to a large pavement outcrop north of the road 1.7 km from the intersection. The metagreywackes contain extensive concordant and cross-cutting migmatitic segregations. Outcrop structures, interpreted as D_3 , include asymmetric folds (most with Z-asymmetry), foliation fish, rotated boudins, and incipient shear bands, all indicating a dextral sense of shear. A coarse grained muscovite-biotite granite dyke intrudes subparallel to the axial surfaces of D_3 folds and has an axial planar foliation. It contains wall-rock inclusions of biotite schist, apparently rotated during incorporation into the dyke, and folded migmatitic schist. The dyke was interpreted as intruded during progressive deformation late in the metamorphic and anatexis history, hence, contemporaneous with late D_3 to D_4 . Zircon from a sample of the dyke gave a scatter of discordant analyses. A single slightly discordant monazite grain with an age of 2642 ± 2 Ma (Table 2), is difficult to interpret in isolation; it could date the intrusion, or retrograde crystallization or resetting of monazite.

Stop E3, Graded layering, Z-folds, ZB93-P287. This easy, but longish, walk to see a single outcrop of graded layering (modified bedding) and spectacular folds in Quetico metagreywacke is worth while, if time permits. Return to the Hillsport road, turning southwest to retrace the route for 250 metres to an intersection with an overgrown road on the north side of an open area. Walk along the overgrown road north and west for 725 metres, crossing a small stream, beyond which, continue south 50 metres to an open area of pavement outcrop. The outcrop is dominated by layered, fine to coarse grained, quartzofeldspathic biotite schist, locally with sillimanite and abundant garnet. Layers of more pelitic composition have more migmatitic segregations. A sharp folded contact (Z-asymmetry) between the coarse garnet-biotite-rich top(?) of one layer, and the quartz-plagioclase-rich bottom(?) of another, suggests relict graded bedding with southward younging. Folds on various scales mostly have Z-asymmetry and, locally, are associated with an axial planar crenulation cleavage. Folds with S-asymmetry dominate one area, indicating that the axial trace of a larger scale fold crosses the outcrop, although no closures are exposed.

E4–E6. Jim Lake synform, Jim/Davis Lakes area

Stop E4, Foliated trondhjemite, ZB93-P270. From the Hillsport junction, take the Camp 70 road north-westerly around the northern ends of Thompson, Loken and Straight Lakes (1:25000 map). At 7 km beyond the junction, turn north on the Jim Lake road; the single-lane Fox Lake siding road continues to the west. Continue north for 3.6 km on the Jim Lake road, turning west on to a good dirt road which leads, in 300 metres, to an area of flooded sand pits. The road is visible beyond the flooded area; it's best to park and wade across as, 200 metres further down the road, it will be necessary (allowing for the vagaries of nature) to cross a beaver dam (less than 10 m long). The road continues in good shape beyond these obstacles.

On the south side of the road, 200 metres west of the beaver dam, a large area (100 m long) of flat-lying outcrops consists of homogeneous coarse grained trondhjemite with minor biotite and magnetite, indistinguishable from synvolcanic trondhjemite in the Willroy-Geco area. The fabric is foliated to gneissic and, locally, deflected by both sinistral and dextral minor shear zones.

Stop E5, Migmatitic biotite schist and iron formation, ZB94-207. Continue along the road for 540 metres and look for a very overgrown track that heads off to the south, about 100 metres east of a low swampy area. Bushwack south for 300 metres and then southwesterly for 100 metres following the track, at which point the new growth is very thick. Continue westerly about 50 metres to a large outcrop (visible on airphoto 90/2-4909 37-15, Ministry of Natural Resources, Ontario) on the north side of the road and west of a small swamp. The biotite schist, containing pinhead garnet, was grouped with Quetico metasedimentary rocks. Tonalitic leucosome is more abundant in pelitic layers and has biotite-garnet-rich selvages, 1–5 cm in width. Leucosome apparently shows 2 generations of folding, the latest of which are east-southeasterly plunging ($121^\circ/18^\circ$) folds that deform early folds and have an axial planar biotite schistosity ($D_3?$). Thin (1–5 cm) rusty quartzose layers with minor magnetite and grunerite, the remnants of iron formation, are

segmented and contorted in the migmatitic matrix. Leucosome adjacent to the iron formation contains some coarse (to 1 cm) porphyroblasts of grunerite.

Stop E6, Mafic metavolcanic rocks, iron formation, Nama Creek pluton, ZB94-197, ZB93-P281-282, ZB94-210. Make your way back to the road and continue westerly, passing the remains of an old cabin 1070 metres west of the beaver dam. At 1345 metres west of the dam, excellent exposures south of the road comprise about 25% mafic to intermediate metavolcanic rocks included in coarse grained to pegmatitic tonalite. Fine to medium grained screens of mafic rock, locally with epidote knots, show layering defined by grain size and hornblende abundance (20–80%). The screens tend to be boudinaged or segmented and, in some cases, biotite-rich rims are related to intrusive contacts. Intermediate to felsic layered screens with disseminated magnetite and possible quartz eyes are present in lesser amounts.

Continue west 30 metres on the same outcrop area to a weakly magnetic quartzose iron formation (about 30 metres of strike-length is exposed) with minor sulphide minerals associated with mafic to intermediate rocks. The iron formation is a semi-continuous layer, up to 5 metres thick, tracable along-strike for about 150 metres on both sides of the road, and cropping out again 1 km to the west. To the south, the unit trends south-southwesterly around the map-scale Z-fold in the Davis Lake area.

Continue along the road for 1650 metres west of the beaver dam to an outcrop on the road itself and extending to the north. The exposure is typical of the 2680 Ma Nama Creek pluton, medium to coarse grained granodiorite, strongly foliated with about 10–25% microcline phenocrysts about 1 cm long. The matrix has white chalky plagioclase and about 20% biotite and hornblende. Fine grained biotite and hornblende inclusions outline growth zoning in the phenocrysts. Some minor mafic screens are present, and an extensive network of folded aplite-pegmatite dykes.

F. Black Pic batholith, supracrustal screens and major folds

F1–F5. Black Pic batholith

Supracrustal rocks of the Manitouwadge belt are enclosed by multiphase foliated to massive plutonic rocks collectively known as the Black Pic batholith (1:25000 map). Foliations in the Black Pic batholith mimic the orientations of the dominant D₂ foliations in the supracrustal suite and are folded by the D₃ Manitouwadge synform. To the south of the Manitouwadge synform, plutonic rocks are mainly tonalitic to granodioritic, but also include dioritic and granitic phases, commonly on a single outcrop. On the basis of cross-cutting relationships, mafic phases are generally older than felsic phases.

Stop F1, Black Pic/supracrustal contact zone, ZB93-61–62. Take the Caramat Industrial road west from Highway 614 for 1.2 km, turning southeast on a gravel road just west of an industrial area. In about 300 metres, a long pavement crops out north of the road under a powerline (1:25000 map). The northern outcrops are dominated by layered fine grained mafic to intermediate hornblende-biotite amphibolite (metavolcanic), locally with epidote knots and garnet. Foliated hornblende-biotite diorite, with 10–20% plagioclase augen (mostly 0.5 cm), increases in abundance toward the south. The diorite, typical of the oldest phase of the Black Pic batholith, is cut by multiphase, concordant and discordant granite, pegmatite and aplite dykes, varying from strongly foliated to massive. Both diorite and granite are involved in tight folds with metavolcanic rocks. The schistosity in the amphibolite is parallel to axial plane traces and also undulates in broad open S-folds (plan view).

Continue 100 metres south on the powerline clearing on the east side of the road. Here in a high strain zone, streaky laminated (submm–10 mm) mafic rocks are unusual in having a westerly-plunging lineation. The zone is cut by plagioclase-porphyritic diorite. Intercalated mafic and dioritic rocks and pegmatite-aplite dykes continue to the south.

Stop F2, Aplitic granite with diorite inclusions, ZB93-63. Retrace the route about 100 metres from the first outcrop of Stop F1 and turn on to a road branching southeast skirting the industrial area; 100 metres further, there are patchy pavement outcrops in the cleared area west of the road. The white weathering fine to medium grained weakly foliated aplitic granite has diffuse pegmatitic patches and more discrete veins. The granite resembles the youngest phase of Stop F1 and contains some minor inclusions of foliated plagioclase-porphyritic diorite.

Stop F3, Three phases of the Black Pic batholith, ZB93-67, ZB94-86. Stops F3, F4 and F5, are accessed from the Camp 70 road at the railway crossing east of Little Manitouwadge Lake; 100 metres east of the crossing, turn off the Camp 70 road to the southwest on to a road paralleling the tracks (1:25000 map). After 500 metres, large rocky knolls slope down on the southeast side of the road. Three main plutonic rocks are exposed, from oldest to youngest; 1) strongly foliated (D₂) coarse grained hornblende-biotite diorite correlated with the oldest phase of Stops F1 and F2, 2) fine to medium grained, buff weathering, foliated granodiorite with <5% biotite and, 3) weakly foliated pegmatitic-aplitic granite, similar to Stops F1 and F2 but more pegmatitic. The diorite contains subtle plagioclase augen (3–5 mm) and locally approaches

monzodiorite with about 10% microcline. It contains some dark fine grained elongate mafic enclaves. Diorite, sampled on this outcrop, has a U-Pb zircon age of $2687 \pm 3/-2$ Ma (Fig. 17, Table 2), within error of the Loken Lake microcline-megacrystic pluton, and both are interpreted as a pre- to syn- D_2 intrusions.

The granodiorite has a strong foliation, locally gneissic, and it contains angular and elongate inclusions of diorite and minor mafic schist. This phase is difficult to correlate between exposures, and it may be that there are several granodioritic-tonalitic phases of similar appearance that post-date diorite and pre-date aplitic granite. Aplitic granite dykes cut the older phases at low to moderate angles to foliation, and have a weak fabric apparently with the same orientation as that in host rocks. A sample of aplitic granite, collected for geochronology on this outcrop, had only poor quality zircon.

Stop F4, Syenite-hornblendite intrusion breccia, ZB93-69. Continue southwest along the road to 1.8 km, 100 metres east of a powerline, to good outcrop ledges and knolls on the north side of the road. Multiphase plutonic rocks here are similar to those of Stop F3, but the relationships are more difficult to see. Of note are the rubbly outcrops and blocks of an interesting intrusion breccia, possibly originating as a diatreme, although we have not mapped here in any detail (see Alkaline rocks (Unit 16), Fox Creek occurrence). An assortment of lithic fragments (1–25 cm), mostly angular mafic to intermediate hornblende-biotite-rich, are embedded in a coarse grained matrix comprising plagioclase, hornblende and biotite with hornblende phenocrysts (euhedral prisms to 1.5 cm) and minor to trace amounts of quartz, microcline, apatite and titanite. Locally, hornblende looks to be of metasomatic igneous origin, forming radiating sprays or oriented perpendicular to inclusion contacts, as though nucleated on inclusion surfaces. Biotite is mainly present as laths within hornblende and tends to be uniformly oriented in each phenocryst, locally looking like partial replacement of hornblende. Although the breccia looks massive, it is cut by aplitic dykes. The breccia contains at least one fragment (2 cm size) of massive sulphide (pyrite-chalcopyrite-magnetite) transported from depth? during emplacement; an interesting concept, as the contact to supracrustal rocks lies about 1 km to the north, and the closest known massive sulphides are more than 3 km to the north at the Geco mine. The area of the breccia crudely corresponds to several isolated anomalous aeromagnetic highs (1:25000 map). The breccia was grouped by Williams et al. (1992) in an 'appinite suite', with other examples intruding the Rawluk Lake pluton to the east in the Faries Lake area (Fig. 2).

Stop F5, Syenite-hornblendite internal structures, Black Pic batholith, ZB93-70. Continue directly under the powerline, south of the road, to an area of extensive outcrop. The first exposures are apparently a continuation of the syenite-hornblendite of Stop F4, here represented by a heterogeneous suite including rhythmically layered rocks, with truncations of layering analogous to cross-bedding. About 200 metres further uphill along the powerline, two main plutonic rocks more typical of the Black Pic batholith are exposed. The oldest, medium grained well foliated granodiorite with about 15% biotite and hornblende, is cut by a younger network of pink massive aplite-pegmatite dykes.

F6–F12. Western Blackman Lake antiform (D_3), Janet Lake road

In contrast to the Black Pic batholith south of the Manitouwadge belt (Stops F1–F5), plutonic rocks to the northwest contain many supracrustal screens, including mafic metavolcanic rocks and iron formation. Outcrop-scale structures, foliation and aeromagnetic trends suggest the presence of a major fold, the D_3 Blackman Lake antiform, with a northeasterly trending axial trace in the Janet Lake area (1:25000 map). In the hinge region, changes in vergence of outcrop-scale folds accompany changes in foliation orientation. The eastern limb of the Blackman Lake fold forms the contact to the main supracrustal belt and may have been the locus of post- D_2 shearing. The pre- to syn- D_2 microcline-porphyrritic Nama Creek pluton was intruded along the contact. The supracrustal screens on the western limb of the Blackman Lake antiform may have been derived from a highly attenuated hinge region of the Jim Lake synform or, alternatively, they may be a continuation of the outer volcanic belt to the south, displaced by apparent dextral shear.

The Janet Lake road is reached by following the Caramat road west from Highway 614 passing the bridge across Nama Creek (14 km), where the Caramat road turns to the north, and continuing northerly about 7.5 km. The Janet Lake road is mostly a good single-lane gravel track, but rough and slow.

Stop F6, Supracrustal screens in tonalite, ZB93-273. Drive east for 1.6 km on the Janet Lake road to an extensive pavement outcrop on the south side (1:25000 map). The foliated tonalite contains many screens of homogeneous to layered (1–10 cm), fine grained, mafic to intermediate metavolcanic rocks. Screens up to 1.5 metres in width comprise about 25% of the outcrop. Some of the layering looks like transposed veins or dykes, and some screens have intrafolial (nearly) folds apparently truncated by the tonalite host rock. It is difficult to unequivocally correlate the fabrics in screens with those in the main supracrustal sequence; but the screens may contain D_1 and D_2 fabrics and folds, whereas the foliation in the tonalite may be a D_2 or D_3 fabric.

Stop F7, Iron formation in tonalite, ZB93-K126. At 4 km east on the Janet Lake road take the road to the north for 2.1 km, then keep to the left fork for a further 300 metres. A good pavement on the west

side of the road exposes screens of iron formation and mafic to intermediate rocks in foliated tonalite. The layered quartz-magnetite iron formation, about 1.5 metres thick, is folded and segmented, and enclosed by amphibolite which is also strongly magnetic. The amphibolite consists of streaky layered, medium to coarse grained, intermediate metavolcanic rock with local smeared-out pegmatite. Some more homogeneous foliated hornblende-biotite diorite is subconcordant to the supracrustal rocks. The silicate mineralogy of the iron formation is typical of iron formation in the Manitouwadge belt; garnet, grunerite and clinopyroxene (bright green mineral?). Pegmatite invading the iron formation is loaded with coarse grained magnetite, suggesting contamination. Iron formation also crops out about 1 km to the northeast.

Stop F8, Z-fold, ZB93-137. Return to the Janet Lake road, and continue east for a further 2.8 km to a pavement on the south side of the road just west of a small stream. Layering in the fine grained foliated tonalite-diorite is defined by the hornblende abundance, which varies from about 10–60%. A strong rodding lineation is parallel to the axis of a Z-fold (D_3) with very long limbs. Locally, layers are boudinaged along the long limbs of folds. Pegmatite dykes follow the axial traces of some folds.

Stop F9, Mafic metavolcanic rock in foliated tonalite, ZB93-133. Continue for 3.8 km past the junction (Stop F7) to good outcrop on both sides of the road. The stop is near the hinge region of the Blackman Lake antiform, and fabric trends are generally variable, due to minor folds. Medium to coarse grained (coarsened?) layered mafic metavolcanic rocks (hornblende-plagioclase-quartz-biotite-clinopyroxene) are invaded by foliated tonalite to diorite. The mafic inclusions show some early folding either transected by foliation or with axial planar foliation. Locally, blocky mafic inclusions in foliated coarse grained tonalite show some rotation of schistosity.

Stop F10, Hinge region of the Blackman Lake antiform, ZB93-129. On the Janet Lake road, 4.4 km past the junction, the road bends 90° to the south and, after another 300 metres, a pavement lies east of the road. The stop is in the hinge region of the Blackman Lake fold. Multiphase foliated plutonic rocks are dominated by medium to coarse grained tonalite with diffuse layering (5 mm–10 cm) defined by 5–40% mafic minerals. The tonalite is intruded by about 10–20% pinkish granite-pegmatite-aplite which varies from concordant and folded dykes to net veins. Pegmatite tends to be present on small shears or short limbs of folds in tonalite, and locally tonalite looks coarsened and homogeneous in diffuse zones on fold limbs. Locally white tonalite is either massive or has a weak foliation (possibly axial planar) defined by about 30% coarse (3–8 mm) hornblende porphyroblasts. Fabric trends are variable and commonly swirly, averaging about 300°/45°. There is a general change across the outcrop from dominantly S-shaped folds in the southeast to dominantly Z-shaped folds in the northwest, accompanied by a change in foliation trends from northerly to westerly.

Stop F11, Nama Creek pluton, ZB93-91. Continue 2.5 km beyond the sharp bend to the south to a high knoll east of the road. See Stop F12 for a description.

Stop F12, Nama Creek pluton, ZB93-87. The stop is 4.6 km south on the road beyond the sharp bend; however, it will probably be necessary to walk the last 1–2 km, depending on how overgrown the road is. The outcrop is a rocky knoll in the bush about 50 metres west of the road. Stops F11 and F12 are both in the microcline-porphyritic Nama Creek pluton; Stop F11 is more easily accessed, but Stop F12 was sampled for geochronology. Both outcrops are typical of the relatively homogeneous intrusion, consisting of coarse grained foliated hornblende-biotite granodiorite to tonalite with 0–20% microcline phenocrysts or augen, 1–3 cm in length. The pluton is interpreted as pre- to syn- D_2 , on the basis of foliations folded around the hinge of the Blackman Lake antiform west of One Otter Lake (1:25000 map). The strong linear and planar fabrics south of Janet Lake may be related to transposition during post- D_2 deformation near the contact to the Manitouwadge belt. The geochronology sample of the Nama Creek pluton contained both zircon and titanite, the former giving an age of intrusion of 2680 ± 3 Ma (Fig. 15, Table 2), somewhat younger than the 2687 Ma pre- to syn- D_2 Loken Lake pluton and oldest diorite of the Black Pic batholith. The titanite age of 2672 ± 3 Ma suggests that regional cooling through the closure temperature of titanite (600°C, Heaman and Parrish, 1991) occurred about 8 Ma after intrusion.

G. Quetico subprovince, western and central Manitouwadge area

In general, the description for the Quetico subprovince in the eastern Manitouwadge area (Stops E1–E3) also applies to the western and central area. However, in the latter area near the Wawa-Quetico boundary, migmatitic metagreywacke is commonly interleaved with gabbroic and dioritic rocks. Diorite apparently intrudes rocks along the subprovince boundary and to the north, and field observations suggest that it pre-dates migmatization (Stop G2). Williams and Breaks (1990a) mapped a lenticular unit of melanocratic diorite (not differentiated on our 1:25000 map) called the Everest Lake pluton (Williams and Breaks, 1990b; Williams et al., 1992), extending along the western subprovince boundary as far east as Appelle Lake. They considered the Everest Lake pluton to be part of the multiphase Black Pic batholith. Much of the generally east-west trending aeromagnetic striping on both sides of the western subprovince boundary could be related to concordant diorite sheets.

The western-central Quetico subprovince is most easily accessed from the Caramat Industrial road west of the Manitouwadge belt. However for Stop G2, this assumes that a major washout, about 3 km east on the Husak road, has been repaired; the latest information (Feb. 1995) suggests that the road is passable.

Stop G1, Migmatitic metagreywacke, folds and new fabrics, ZB94-50. The stop is reached by following the Caramat Industrial road about 32 km from Highway 614 to an intersection with the Michal Lake road (to the west) and the Husak road (to the east). The intersection is about 9.75 km north on the Caramat road from the junction with the Janet Lake road (Stops F6-F12). Turn west on the Michal Lake road and drive on for about 1.3 km to an extensive area of pavement outcrops in an old clearing or road leading south, between Hourglass and Slingshot Lakes (1:25000 map). The exposure of migmatitic metagreywacke, locally with cordierite or garnet in leucosomes, shows multiple folds and complex structural relationships. Near the road, stretched and boudinaged migmatitic segregations define isoclinal folds, and some rotated boudins show dextral kinematics (in plan view). Elsewhere on the outcrop, layering and foliation show both Z- and S-shaped folds, some of which refold earlier folds of migmatitic segregations. In micaceous layers, an axial planar spaced cleavage is commonly developed, with less micaceous microlithons preserving earlier fabrics; in psammitic layers, spaced cleavage is less developed and folded fabrics dominate. Locally, in the hinge of an S-fold, asymmetric kink bands that deform the spaced cleavage were interpreted as the result of progressive deformation.

Stop G2, Migmatitic metagreywacke interleaved with diorite, ZB94-C45. Return to the intersection with Caramat road and continue through to the east on the Husak road for about 13.5 km (about 1.5 km west of the junction with the Olson Lake road). Turn left on to a small track leading northeasterly and up hill, following it as it turns to the northwest (left) after 360 metres. About 70 metres further, the track is crossed by a large (200 m long) clear area of outcrops trending north-northeasterly. Three main rock types are exposed; 1) medium to coarse grained foliated diorite, commonly with up to 40% plagioclase phenocrysts (2-5 mm) and 10-50% hornblende and biotite, 2) fine to medium grained layered heterogeneous foliated tonalite to hornblende-biotite diorite, and 3) migmatitic biotite schist with boudinaged and folded layers of tonalitic-pegmatitic leucosome. These rock types are interleaved on a scale of 10 cm to several metres. The diorite is foliated, but low strain enclaves preserve igneous textures. Diffuse coarse leucocratic enclaves, suggesting incipient migmatization, are heterogeneously distributed, in part, localized in dilational zones such as boudin necks. In many cases, coarse leucocratic zones form an interconnecting network of ganglions enclosing enclaves of dioritic protolith. The overall impression is that the diorite sheets show less strain than metagreywacke, possibly a function of competency contrast. The heterogeneous and patchy distribution of coarse leucocratic domains in diorite suggests that migmatization may have partly been driven by the introduction of fluids or fluxing components. Note also that tight to isoclinal folds of layering are transected locally by an oblique southeasterly trending foliation defined by hornblende and biotite.

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