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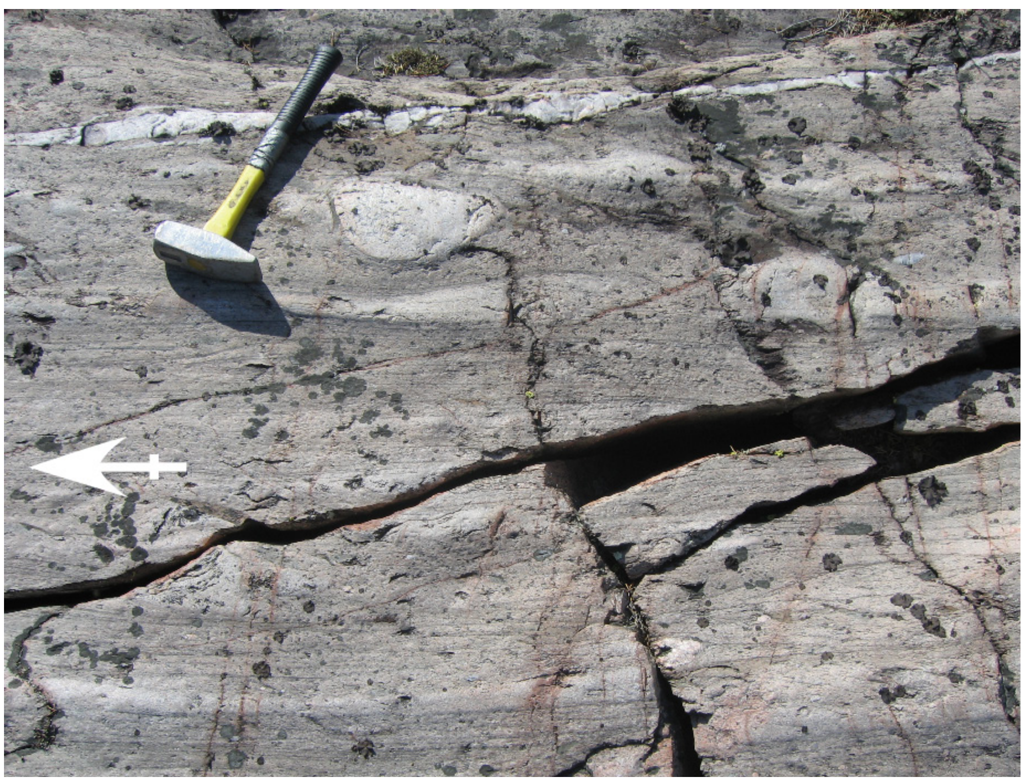

Energy Resources Conservation Board

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Overview of the Geological Evolution of the Canadian Shield in the Andrew Lake Area Based on New Field and Isotope Data, Northeastern Alberta (NTS 74M/16)

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Energy Resources Conservation Board
Alberta Geological Survey

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Abstract

In spite of systematic mapping by the Alberta Geological Survey (AGS) and the Geological Survey of Canada (GSC), aided by petrology and geochronology studies, the geological evolution of the Canadian Shield in the Andrew Lake–Waugh Lake area of northeastern Alberta is still controversial. In particular, the inferred distribution, nomenclature and geological significance of the greenschist-facies rocks near Waugh Lake vary substantially between different authors. The area was originally mapped in detail by Godfrey (1961, 1963, 1966), who separated map units on petrographic grounds. Iannelli et al. (1995) reinterpreted most of the greenschist-facies rocks in the Waugh Lake area as a stratigraphic succession; defined four formations (Doze Lake, Sederholm Lake, Johnson Lake and Niggli Lake formations), some of which he further subdivided into members; and assigned them to the Waugh Lake Group. McDonough et al. (2000) retained the name Waugh Lake Group but did not subdivide it into formations and members. Of interest to AGS is whether or not the assignment of formal stratigraphic nomenclature to rocks within the Waugh Lake area is appropriate and should be continued in future AGS publications.

In the summer of 2006, AGS carried out geological mapping and sampling near Andrew and Waugh lakes to better understand the geological context and key lithological contacts with direct metallogenic and cartographic implications for the Canadian Shield within Alberta. This report includes a summary of previous work, brief descriptions of the main map units, a re-evaluation of previous isotope data, new geochemical and isotope data, and a discussion of their significance. Based upon the review of the current state of knowledge and new field and analytical data, the author recommends that the formal stratigraphic rank of group be replaced by the informal term complex. Rocks in the investigated area are assigned to the following major map units:

a) Taltson basement complex, dominated by orthogneiss;

b) migmatite Rutledge River Complex;

c) greenschist-facies Waugh Lake Complex; and

d) Taltson plutons.

The Waugh Lake Complex yielded Sm-Nd data that coincide, within analytical error, with the adjacent Taltson plutons. Dating of a deformed igneous pod within the Waugh Lake Complex returned a U-Pb zircon emplacement age of ca. 1973 Ma, similar to the Taltson plutons. Argon/argon dates between ca. 1840 and 1820 Ma from four phyllonite samples collected on the east side of Waugh Lake are contemporaneous with the late phases of strike-slip displacement on major shear zones overprinting the Taltson magmatic zone. Field and microscope observations suggest that parts of the greenschist-faces Waugh Lake Complex may be directly derived from the adjacent Taltson magmatic zone crust through shearing and metamorphic re-equilibration. Other parts of the assemblage could be derived from a volcano-sedimentary cover of the Taltson crust incorporated in this linear zone of strain concentration and metamorphism. This tectonite zone, which is steeply dipping and roughly northerly trending, projects north and south into similar rocks that define a narrow belt, 1–2 km wide and tens of kilometres long, that appears to be part of a network of low-grade rocks stretching more than 100 km from west of Harper Lake in Saskatchewan to Hill Island Lake and beyond in the Northwest Territories. There is a distinct possibility that the low-grade metamorphism of the Waugh Lake Complex and its correlatives is related to tectonic deformation.

This report includes preliminary results and recognizes that well-documented conclusions on the geological evolution of the Andrew Lake area require further detailed field and systematic analytical work.

1 Introduction

1.1 Scope of Work

In an effort to update geological knowledge, and better assess the mineral potential, of the Canadian Shield in northeastern Alberta (Alberta shield), Alberta Geological Survey (AGS) has re-examined selected outcrops and previous exploration sites of the 1960s and 1970s near Andrew Lake. In the summer of 2005, AGS conducted a three-week reconnaissance geological field program (Pană et al., 2006). Between July 14 and 28 of 2006, a crew varying in size from two to five AGS staff members did float-plane– and motorboat-supported geological mapping and sampling near Andrew and Waugh lakes. The objective of the work was to review the general geology of the Andrew Lake area, based on the original geological reports of J.D. Godfrey (Godfrey, 1958, 1961, 1963) and the updated Geological Survey of Canada Map 1953A (McDonough et al., 2000b), and to examine the evidence for existing tectonic interpretations with direct cartographic implications for the Alberta shield. The report includes

- summaries of previous work (Section 1.3) and tectonic interpretations (Section 2) pertinent to the investigated area;
- brief descriptions of the main map units, focused on previously existing isotope data that led to current interpretations of the geological evolution (Section 3);
- an expanded discussion on the characteristics of the controversial low-grade (greenschist-facies) Waugh Lake Complex (Section 4); and
- summary tables of previous isotope work, as well as new geochemical and isotope data.

This report includes preliminary results and recognizes that well-documented conclusions on the regional geology of the Andrew Lake area require further detailed field and analytical work.

1.2 Location, Access and Physiography

The Andrew Lake map area (NTS 74M/16) is situated about 60 km north of Lake Athabasca in the extreme northeastern corner of Alberta, which adjoins Saskatchewan and the Northwest Territories (Figure 1). The area under investigation lies between latitudes 59°45′ and 60°N, and longitudes 110° and 110°30′W. Float planes can be chartered from Fort Chipewyan in Alberta, Fort Smith in the Northwest Territories and Uranium City in Saskatchewan for access to the map area, where many scattered lakes are suitable for landing. An airstrip at the south end of Andrew Lake allows access to the area by small wheel-equipped planes. The general elevation varies from 330 to 470 m above sea level. Pleistocene glacial scouring has left numerous rocky-basin lakes and a locally rugged surface, with a maximum relief of about 115 m. At some localities, clean fresh bedrock surfaces are found in the form of low wide aprons bordering lakes. Glacially smoothed outcrops, highly polished striated rock surfaces, sand plains, drumlins, eskers and glacial erratics provide evidence of Pleistocene glaciation. The lakes are either poorly connected, with westward drainage to Charles Lake, or disconnected, thus necessitating portages when travelling cross-country by water.

1.3 Previous Work

The first geological information on the Alberta shield came from J.B. Tyrrell and F.J. Alcock, who made the first canoe traverses along the north shore of Lake Athabasca, and from A.E. Cameron and H.S. Hicks, who conducted reconnaissance surveys north of Lake Athabasca (Tyrrell, 1896; Alcock, 1915, 1917; Cameron, 1930; Cameron and Hicks, 1931; Hicks, 1930, 1932). The area adjoining Andrew Lake to the east, in Saskatchewan, was mapped at a scale of 1:253 440 (1 inch to 4 miles) after gold was discovered at Goldfields (Alcock, 1936). North of the Alberta border, the Fort Smith map area was published at a scale of 1:253 440 by Wilson (1941).

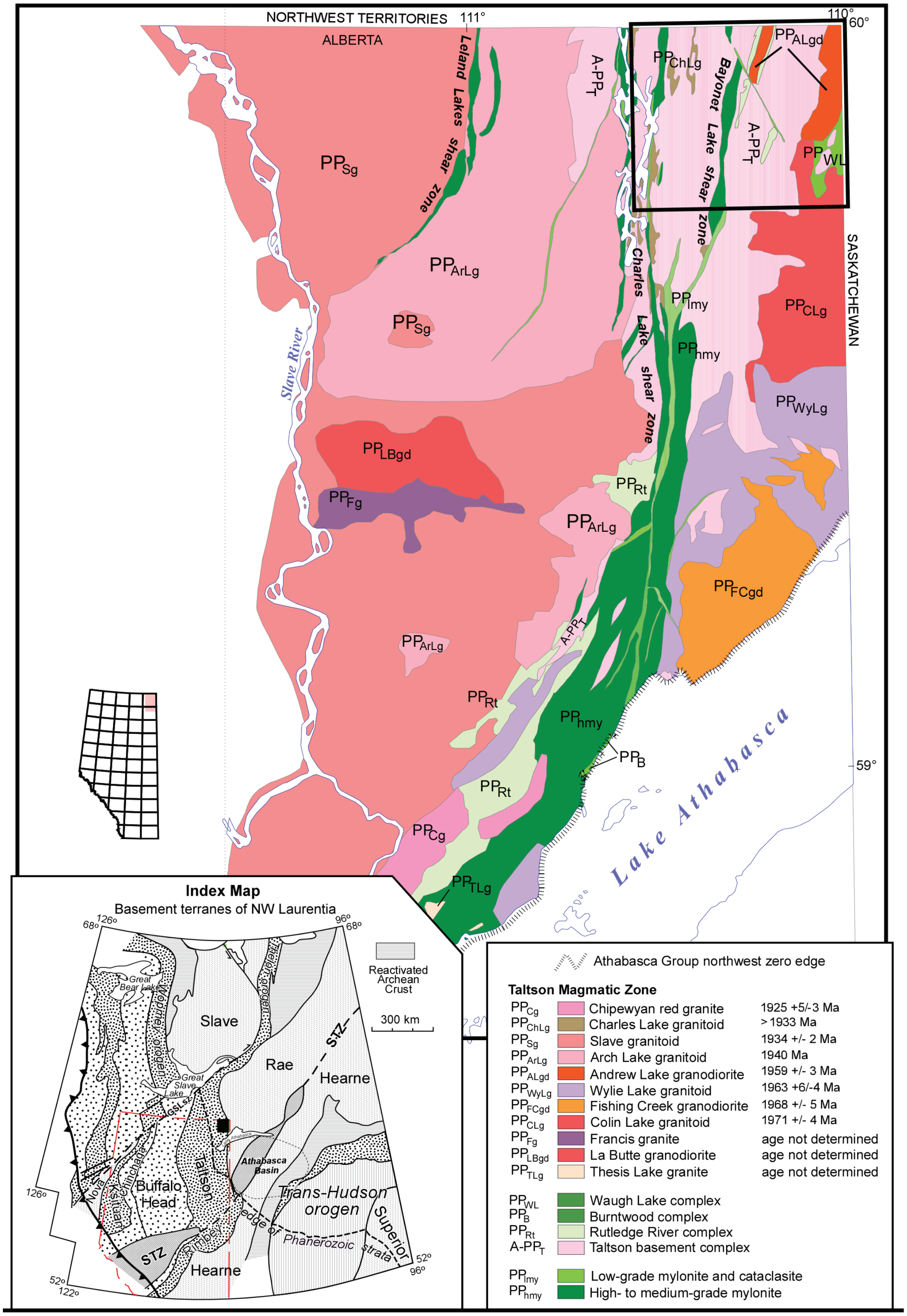


Figure 1. Simplified geology of the Alberta shield (after Godfrey, 1986a); index map compiled from Hoffman (1988) and Ross et al. (1994); black rectangle is area under investigation. See Pană (2010) for an updated map of the Alberta shield at a scale of 1:250 000.

In 1959, the Geological Survey of Canada carried out a reconnaissance survey of the Alberta shield and published a map at 1:253 440 scale, with marginal notes (Riley, 1960). In 1960, the Saskatchewan Department of Mineral Resources carried out a mapping program at a scale of 1:63 360 in an area immediately east of the Andrew Lake map area (Koster, 1961). Between 1957 and 1975, the Alberta shield was systematically mapped by J. Godfrey of the Alberta Research Council. Large portions of the Alberta shield were selectively remapped at a scale of 1:50 000 by the Geological Survey of Canada in the 1990s, aided by isotope dating (McDonough et al., 2000c; McNicoll et al., 2000), thermobarometry (Grover et al., 1997) and satellite-imagery interpretation (Schetselaar, 2000). Results of surficial mapping of the Alberta shield have been published by Bednarsky (1999). Regional geophysical data for the Alberta shield have been published by Sprenke et al. (1986), Charbonneau et al. (1994) and Lyatsky and Pană (2003).

The geology of the Alberta shield and various types of mineral occurrences previously reported there have been compiled on a 1:250 000 scale map (Godfrey, 1986a, b). These mineral occurrences as well as uraniferous boulders, pits, bogs and uranium exploration drill-holes reported along the north shore of Lake Athabasca, have been recently digitally compiled by Alberta Geological Survey (Pană and Olson, 2009).

The first geological reporting specific to the Andrew Lake area is credited to Cameron and Hicks (1931), who started one of their canoe traverses of the Canadian Shield from Andrew Lake. Low-grade uranium mineralization was found in the Andrew Lake area in the course of uranium prospecting and exploration during the 1950s (e.g., Ferguson, 1953), which led to a small uranium prospecting program of the Alberta shield by the Alberta Research Council (Collins and Swan, 1954). J. Godfrey started his systematic mapping of the Alberta shield with the Andrew Lake area (1957–1959).

The publication of the report by Godfrey (1958) on mineral showings in the area, and of his subsequent reports accompanied by detailed geological maps (Godfrey, 1961, 1963), led to intense staking activity and filing of about 150 mineral claims. A geological map at a scale of 1:15 840, credited to R. Watanabe and J. Godfrey (1960) and included in Watanabe's M.Sc. thesis (Watanabe, 1965), covers the Waugh Lake area and extends a few kilometres into Saskatchewan. The Alberta portion of this map was incorporated, with more detail, into the 'Andrew Lake, south district' map area (Godfrey, 1963) and included in the geological report on the area (Godfrey, 1963).

In 1963, an aeromagnetic survey of the area was flown by Aero Survey Ltd. on behalf of the federal government (Geological Survey of Canada, 1964a, b). In 1969 and the early 1970s, Hudson Bay Oil and Gas Co. Ltd. carried out an airborne survey over the area, conducted limited ground geophysics and completed four small trenches across an electromagnetic conductor discovered on the eastern shore of Waugh Lake (Burgan, 1971). Exploration (including blasting, trenching and drilling) for uranium was carried out in the 1970s near Andrew Lake, including the area west of Carrot Lake and near the north end of Spider Lake. No significant volume of economic mineralization was found and no further exploration was carried out at the time.

Under the auspices of the Canada-Alberta Partnership Agreement on Mineral Development, Alberta Geological Survey examined the Andrew Lake area and the previously reported mineralized zones (Langenberg et al., 1993; Langenberg and Eccles, 1996). In 1993, map units and mineral occurrences were outlined in an area of approximately 24 km^2 near Waugh Lake that was remapped at a scale of 1:10 000 (Salat et al., 1994). The revision of this work in 1994 resulted in a very different map and a formal stratigraphic nomenclature for the low-grade metamorphic rocks, now assigned to the 'Waugh Lake Group' (Iannelli et al., 1995). Selective remapping and reinterpretation of the local geology, aided by isotope dating, is included in the 1:50 000 scale map 1953A by McDonough et al. (2000b).

2 Tectonic Setting

The Andrew Lake area is part of the Taltson magmatic zone (TMZ), which makes up the southern segment of the approximately 3000 km long, Paleoproterozoic (2.0–1.9 Ga) Thelon-Taltson orogenic belt (see index map in Figure 1; e.g., van Breemen et al., 1987; Hoffman, 1988, 1989; Ross et al., 1991, 1994; Berman and Bostock, 1997; McDonough et al., 2000c; McNicoll et al., 2000; Ross, 2002; Ross and Eaton, 2002). The exposed segment of the TMZ extends from the Great Slave Lake shear zone in the Northwest Territories to Lake Athabasca in northeastern Alberta. Based on aeromagnetic mapping and drillhole data, the TMZ has been traced another 300 km farther south, beneath the late Paleoproterozoic Athabasca Group and the Phanerozoic cover of the Western Canada Sedimentary Basin, where it appears to be truncated by the Snowbird Tectonic Zone (see index map in Figure 1; e.g., Hoffman, 1988; Ross et al., 1991). The TMZ includes many ca. 1.99–1.92 Ga granitoid bodies that intrude and rework Archean to Paleoproterozoic crust at the western margin of the Rae Terrane (Figure 1; e.g., McNicoll et al., 2000; McDonough et al., 2000b, c).

Petrographic, geochemical and isotope data were interpreted to indicate that the TMZ evolved from a continental magmatic arc to the core of a collisional orogen (Figure 2; e.g., Ross et al., 1991; McDonough et al., 2000c; Ross, 2002). In this scenario, eastward subduction of ca. 2.3 Ga oceanic crust was followed by the development of the Rutledge River basin and the smaller and slightly younger Waugh Lake basin on continental crust, and resulted in the intrusion of the early, subduction-related granitoid plutons in the eastern part of the TMZ (including the area of deposition of the Waugh Lake Group); during subsequent collision, anatectic granitoid plutons intruded the western TMZ. Major eastward-directed thrusting would have juxtaposed hot rocks from lower to mid-crustal levels and the Waugh Lake Group strata, leading to metamorphism of the latter under low-grade conditions.

Additional geochemical and isotopic data revealed that the early granitoid suite of the TMZ lacks the mantle component that is apparent in Phanerozoic magmatic arcs, making it unlikely that the suite is subduction related (Chacko et al., 2000; De et al., 2000). Instead, both early and late suites of the TMZ granitoid bodies in northeastern Alberta appear to have an intracrustal origin, interpreted to indicate that the Taltson-Thelon tectonomagmatic zone evolved in a plate-interior setting, either indirectly related to subduction (Chacko et al., 2000; De et al., 2000) or triggered by mantle processes unrelated to subduction (Pană et al., 2007).

In Alberta, the exposed TMZ includes a curving, linear body of high- to medium-grade metamorphic tectonite referred to as the Taltson basement complex (TBC), which is considered to be the westernmost expression of the Churchill crust (Rae Terrane), flanked by several distinct, moderately to strongly foliated granitoid plutons with minor TBC roof pendants (Figure 1; Godfrey, 1986a; McDonough et al., 2000c). The TBC consists of a composite succession of Archean (ca. 3.2, 3.1 and 2.6 Ga) and Paleoproterozoic (ca. 2.4–2.1 Ga) metaplutonic gneiss (Goff et al., 1986; McNicoll et al., 2000). Subordinate biotite±aluminosilicate gneiss was inferred to be derived from clastic sedimentary rocks of the 2.13–2.09 Ga Rutledge River basin (e.g., Bostock and van Breemen, 1994; McDonough et al., 2000c). Minor but widespread amphibolite and hornblendite, and rare metagabbro (possibly metamorphosed ophiolite remnants) within the Taltson basement complex might represent the initial phase of rifting that subsequently formed the Rutledge River basin (McNicoll et al., 2000).

As mentioned above, the origin and tectonic setting of the TMZ granitoid plutons is controversial. Based on initial petrological data, age and position relative to the TBC, the granitoid bodies in Alberta have been assigned to two groups (e.g., Bostock et al., 1991; McDonough et al., 2000c; McNicoll et al., 2000): an earlier, 1.99–1.96 Ga, subduction-related, weakly peraluminous to metaluminous suite (I-type) east of the TBC; and a later, ca. 1.96–1.93 Ga, collision-related, peraluminous suite (S-type) west of the TBC

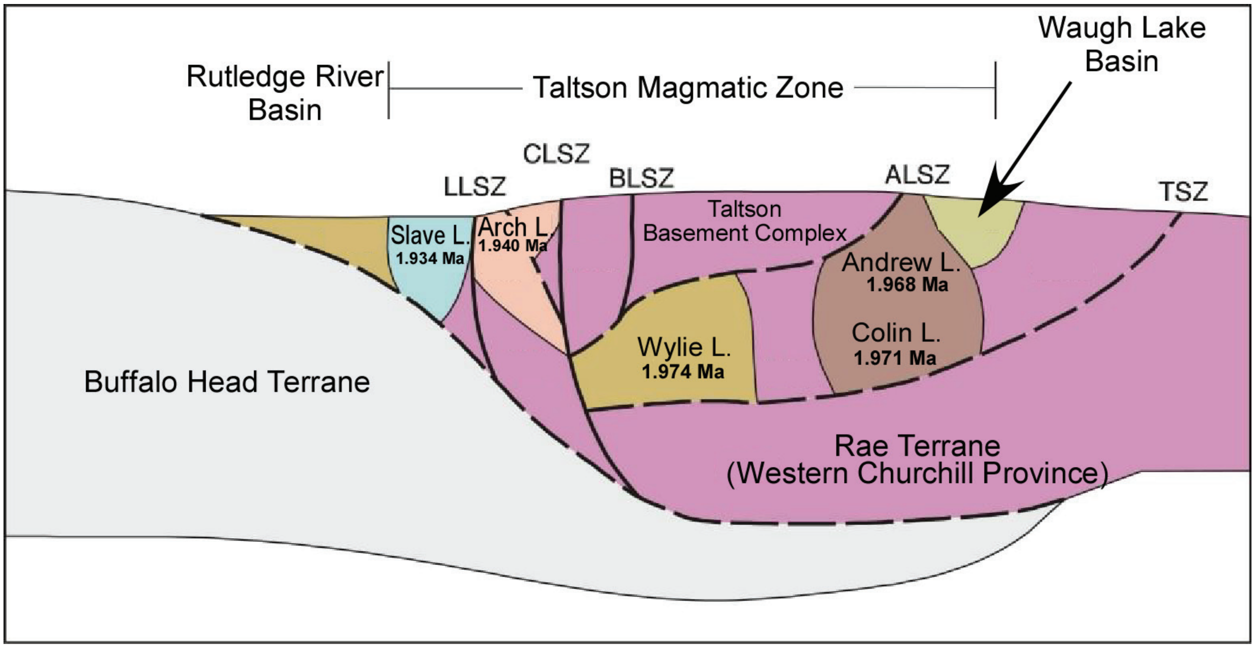


Figure 2. Conceptual cross-section of the Alberta shield (after McDonough et al., 2000b). Major shear zones, from west to east: LLSZ, Leland Lakes shear zone; CLSZ, Charles Lake shear zone; BLSZ, Bayonet Lake shear zone; ALSZ, Andrew Lake shear zone; TSZ, Tazin Lake shear zone. Abbreviation: L., Lake.

(Figure 1). The early TMZ plutons include 1) the ca. 1971 Ma Colin Lake granodiorite to quartz diorite; 2) the ca. 1963 Ma Wylie Lake suite of granodiorite, quartz diorite and quartz monzonite; and 3) the ca. 1962–1959 Ma Andrew Lake suite of granodiorite to diorite. The late S-type plutons, characterized by abundant mafic clots of biotite, garnet, andalusite, hercynite and cordierite, include 1) the ca. 1960–1934 Ma, polyphase Slave monzogranite; and 2) the ca. 1.938 Ga Arch Lake quartz monzo- to syenogranite. Several smaller, variously foliated plutons of uncertain geochemical affiliation are enclosed in the TBC: 1) the Charles Lake granite suite with monazite metamorphic ages of ca. 1933–1919 Ma; and 2) the ca. 1925 Ma Chipewyan and possibly Thesis syenogranite to quartz monzonite suite. Subsequent petrological data, including chemical and isotope analyses, indicate that the entire TMZ plutonic suite may be derived from the recycling (re-melting) of the essentially igneous TBC crust (Chacko et al., 2000; De et al., 2000; Pană et al., 2007).

From west to east, the following major, northerly-trending, high- to low-grade shear zones cross the Alberta–Northwest Territories border: the Leland Lakes (LLSZ), Charles Lake (CLSZ) and Bayonet Lake shear zones (Figure 1; Godfrey, 1958; Langenberg, 1983; Godfrey, 1986a; McDonough et al., 2000c). These three shear zones are all characterized by the presence of extensive mylonite zones (Godfrey, 1986a; McDonough et al., 2000b, c). The Andrew Lake Shear Zone (ALSZ) inferred by McDonough et al. (2000b, c) could not be confirmed by the author's work and will be discussed in Section 4. The crosscutting relationships between shear zones and granitoid bodies are very complex, indicating concurrent deformation and granitoid intrusion during the ca. 2.0–1.9 Ga Taltson tectonothermal event (e.g., McDonough et al., 2000c).

3 Main Map Units in the Andrew Lake Area

The dominant geological feature of the Andrew Lake area (Figure 3) is the alternation of northerly trending map units, ranging from several metres to 5 km in width. In current interpretations, these map

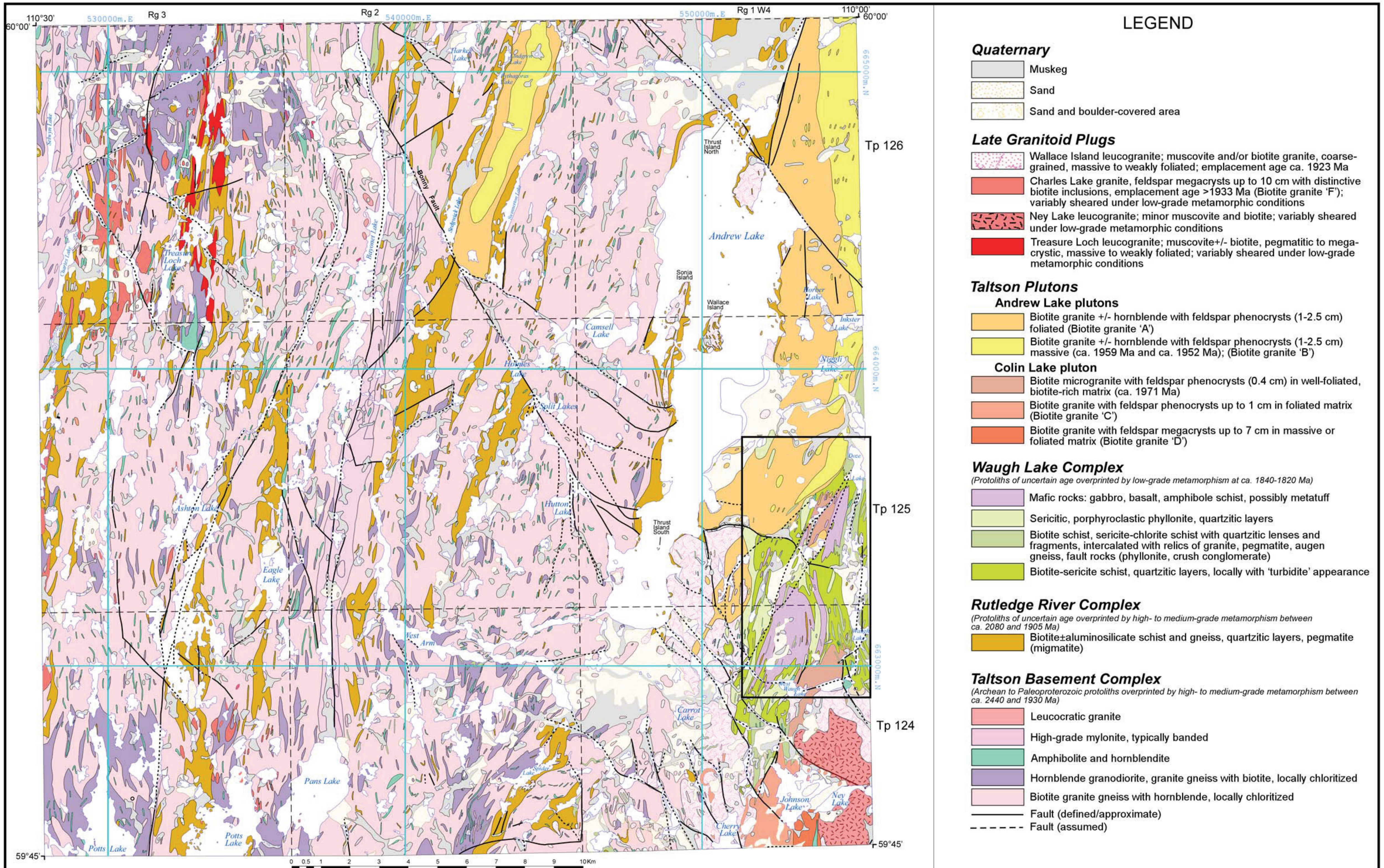


Figure 3. Compiled geological map of the Andrew Lake area; lithological contacts and faults from Godfrey (1961; 1963; 1966a and 1966b). Legend includes additional information from McDonough et al. (2000a, b) Bostock and van Breeman (1994); Berman and Bostock (1997) and new analytical work from this report. Where appropriate, granite type denoted by capital letters as originally separated by J. Godfrey are given in parentheses. The boxed area is shown in greater detail in Figure 7.

units are in chronological order, of the following principal rock groups (Godfrey 1958, 1961, 1963; Bostock and van Breemen, 1994; McDonough et al., 2000b, c; McNicoll et al., 2000):

- the Taltson basement complex (TBC), an orthogneiss suite of various granitoid gneiss units derived through Paleoproterozoic recycling of the Archean Rae crust;
- the upper amphibolite Rutledge River Complex (RRC), with granulite facies relicts and variable greenschist-facies overprint;
- the greenschist- to subgreenschist-facies Waugh Lake Complex (WLC) in the easternmost part; and
- the Taltson plutonic complex of variably deformed porphyritic feldspar-biotite granitoid bodies.

Rock types of the Rutledge River and the Waugh Lake metamorphic successions have been previously interpreted to have clastic or volcaniclastic protoliths (e.g., ‘pelitic gneiss’, ’phyllite’, ’meta-arkose’), which lead to their interpretation as metamorphosed stratigraphic successions and assignment of formal names: ‘Rutledge River Group’ (e.g., McDonough et al., 2000b, c) and ‘Waugh Lake Group’ (e.g., Iannelli et al., 1995; McDonough et al., 2000b, c). Both the protolith-based rock names and the formal stratigraphic nomenclature are at odds with the recommendations of the International Union of Geological Sciences (IUGS) and are briefly discussed below.

The medium- or high-grade Rutledge River and the low-grade Waugh Lake successions consist of metamorphic tectonites interlayered with igneous rocks. A metamorphic tectonite is defined as a rock formed through tectonism under various metamorphic conditions, and characterized by a fabric (foliation and lineation) that indicates solid (ductile) flow (S, L or L-S tectonites). Within the two successions discussed here, metamorphic tectonites are often in gradational contact with adjacent igneous rocks. The transition through strain, hydration and metamorphic reaction gradients indicates that at least some of the metamorphic tectonite layers are metamorphic derivatives of the adjacent igneous rocks. The present level of knowledge of the two metamorphic successions does not justify the previously popular use of sedimentary protolith–based rock names and of the formal stratigraphic nomenclature.

The IUGS Subcommission on the Systematics of Metamorphic Rocks (SCMR) accepts the continued usage of protolith-based names only for very low to low-grade metamorphic grades (particularly those subjected to little deformation), although “it provides a poor basis for a comprehensive…terminology”. The use of the prefix ‘meta’, in combination with a protolith-based name may be used “only when the protolith is easily identifiable or obvious”, but “more precise terms should be used whenever possible (e.g., epidote bearing actinolite-chlorite schist)” (Arkay et al., 2003, p. 7). Following the SCMR recommendations, the author has replaced previously used protolith-based rock names for the medium- or high-grade Rutledge River succession and the controversial Waugh Lake rocks with petrographic names based on the minerals present and the structure of the rock, considered “the most obvious major parameters for rock classification or nomenclature” (Schmid et al., 2004, p. 3).

Further, the Rutledge River and the Waugh Lake lithological assemblages are here redefined as complexes because they do not satisfy the criteria for formal lithostratigraphic terms ‘group’ or ’formation’ of a conventional stratigraphic hierarchy, as recommended by the IUGS Subcommission on Stratigraphic Classification (Salvador, 1994). The term ‘group’ “is applied most commonly to a sequence of two or more contiguous or associated formations with significant and diagnostic lithological properties in common” and “establishing groups without constituent formations in anticipation of possible action of future workers should be avoided” (Salvador, 1994, p. 35). Most importantly, a ‘formation’ should have a “clear and precise standard definition based on the fullest possible knowledge of its lateral and vertical variations”, a stratotype (type section) and a stratigraphic column (Salvador, 1994, p. 33–34 and 36–37). None of these criteria are satisfied by the Rutledge River and the Waugh Lake lithological assemblages.

No formations have been separated within the 'Rutledge River Group' of McDonough et al. (2000c), and the formations defined in a limited area by Iannelli et al. (1995) for the 'Waugh Lake Group' are highly suspect and irrelevant at regional scale (see Section 4). Both assemblages crop out as steeply dipping domains of medium- or low-grade tectonite and, where well exposed, their contacts to adjacent assemblages are gradational, through strain and metamorphic reactions. Hence, no clear stratigraphic relationships can be discerned due to high strain, folding and transposition.

Although "stratified volcanic rocks and bodies of metamorphic rocks that can be recognized as of sedimentary and/or extrusive origin can be treated in every respect as unmetamorphosed sedimentary lithostratigraphic units because their characteristic lithological features, original layering and stratigraphic relationships are readily distinguishable", it is incorrect to do the same "in the case of nonlayered intrusive rocks and of the bodies of metamorphic rocks that are deformed and/or recrystallized so that their original layering and stratigraphic succession may no longer be ascertained" (Salvador, 1994, p. 41). The Rutledge River lithological assemblage crops out as more or less continuous linear domains of migmatite up to 2 km wide and 10 km long, consisting of mainly biotite-garnet±aluminosilicate gneiss intimately interlayered and folded with metamorphic pegmatite and quartz segregations. The Waugh Lake lithological assemblage consists of belts of various types of biotite and/or chlorite-sericite-quartz phyllonite, fragmental rocks of uncertain origin (sedimentary or tectonic) and possibly metasedimentary rocks, separated by fine- or medium-grained igneous rocks. In both lithological assemblages, the metamorphic foliation is everywhere subvertical and concordant to the adjacent orthogneiss domains, no stratigraphic "up" can be recognized, and the fold axes are steeply plunging and parallel to the local structural trend.

According to the IUGS Subcommission on Stratigraphic Classification, the term 'complex' defines a map unit "composed of diverse types of any class or classes of rock (sedimentary, igneous, metamorphic) and characterized by irregularly mixed lithology or by highly complicated structural relationships to the extent that the original sequence of the component rocks may be obscured and the individual rocks or rock sequence cannot be readily mapped" (Salvador, 1994, p. 36). Consequently, in this report, the formal names Rutledge River Complex and Waugh Lake Complex are used to indicate that "the stratigraphic relations of the individual lithologies forming the body of rock are poorly known or identifiable and that the body, therefore, cannot be subdivided on stratigraphic grounds" (Salvador, 1994, p. 42).

3.1 Taltson Basement Complex

Most of the western half of the Andrew Lake map area is underlain by granitoid gneiss of the Taltson basement complex, which consists of foliated to layered to banded, mylonitic, biotite-hornblende granite to granodiorite gneiss and hornblende diorite gneiss (Figure 4). This complex also includes minor lenses of biotite schist, quartzitic layers, amphibolite and garnetiferous layers, and is pervasively intruded by dispersed bodies of granite pegmatite; massive, medium-grained or porphyritic pink granite; and granodiorite dikes and sills. Its contacts with the interspersed Rutledge River Complex are transitional (Godfrey, 1961, 1963). Local migmatitic structures and frequent granite dikes and sills indicate partial melting of the crust. Consistent belts of well-banded mylonite with sparsely preserved subhorizontal stretching lineations, amphibolite pull-apart structures and ductilely deformed feldspar (Figure 5) define the Charles Lake and Bayonet Lake high- to medium-grade mylonite zones, variably overprinted by low-grade mylonite (Figure 3 and Appendix 1).

Numerous bands of biotite amphibolite, amphibolite and hornblendite are scattered throughout the granite gneiss, many of them too small to be shown on a regional-scale map (Godfrey, 1961, 1963). A larger hornblendite lens that was followed approximately 200 m along strike occurs on the ridge between Split and Hutton lakes. Occasionally observed transitions from mafic granulite and hornblendite to amphibolite,

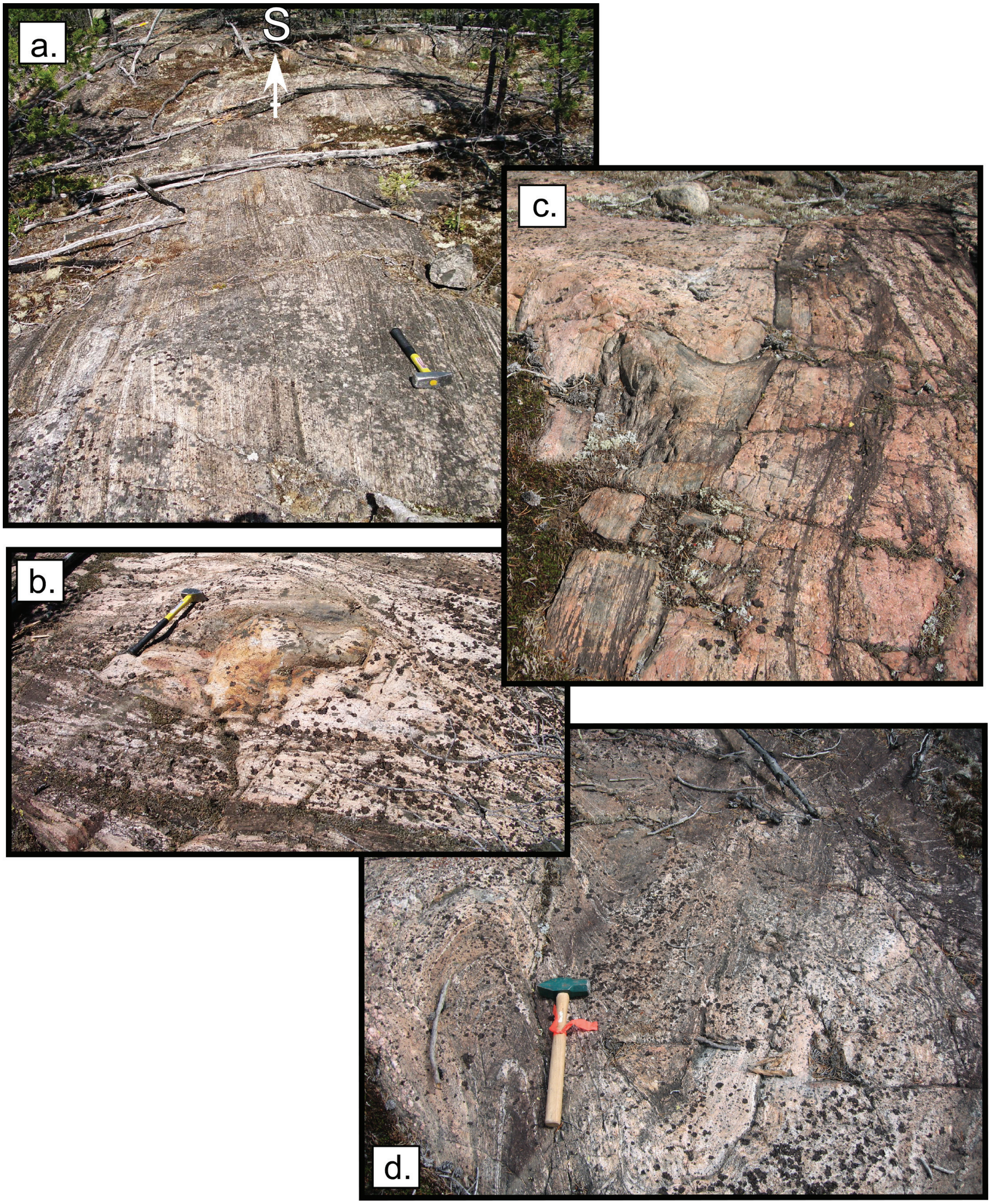


Figure 4. Typical outcrops of orthogneiss in the Taltson basement complex: a) vertical plane-parallel layering; b) granite gneiss pod within the Taltson orthogneiss; c) and d) outcrop-scale folds with subvertical fold axes interpreted to represent loci of ductile flow instability during regional-scale northerly-trending transcurrent displacement.

Figure 5. High-grade mylonite in the Bayonet Lake shear zone overprinting orthogneiss of the Taltson basement complex near the north end of the west arm of Andrew Lake; all types of foliation are northerly trending and vertical.

and from hornblende granite or diorite to amphibolitic gneiss, suggest the amphibolite bodies are the result of metamorphic re-equilibration of mafic igneous and granulite protoliths to medium-grade metamorphic conditions. Amphibolitic masses are commonly bordered by biotite-rich schistose layers that may represent a metamorphic reaction rim where biotite formed at the expense of hornblende.

Initial K-Ar isotope ages reported from the high- to medium-grade metamorphic rocks were rather inconclusive, as they were in the same range as those reported from Taltson plutons and the low-grade Waugh Lake Complex, which are clearly younger based on field relationships (Appendix 3, Table 9; Godfrey and Baadsgaard, 1962; Baadsgaard and Godfrey, 1967, 1972). Subsequent multisystem isotope data from the Taltson basement complex and Taltson plutons in areas adjacent to Andrew Lake (Appendix 4, Table 10) and elsewhere on the Alberta shield show that most of the previously reported K-Ar dates from hornblende and biotite are consistent with slow uplift and cooling of the TMZ through the hornblende (ca. 530±40°C) and biotite (280±30°C) closure temperatures between ca. 1.9 Ga and ca. 1.8 Ga (Appendices 3 and 4). Previous K-Ar muscovite dates (1750 and 1732 Ma) younger than the biotite dates from similar rocks likely record partial argon loss following cooling through the muscovite closure temperature (ca. 350°C).

Previous $\varepsilon_{Nd(2.2. Ga)}$ values of –1.6 to –14.3 and T_{DM} model ages of 3.73–2.59 Ga from Taltson basement rocks are similar to those obtained from the Buffalo Head Terrane, and overlap with values for the Archean Rae Terrane (De et al., 2000; McNicoll et al., 2000). Because most samples from Taltson granitoid gneiss layers yielded U-Pb zircon (emplacement) ages ranging from ca. 2.56 to 2.14 Ga and include layers dated at ca. 3.19 Ga and ca. 3.08 Ga, it is inferred that the Taltson basement complex formed through Paleoproterozoic recycling of the Archean Rae Terrane (McNicoll et al., 2000). Within the Andrew Lake map area, the ca. 2335 Ma age of a tabular, weakly deformed, pink pegmatitic granite that truncates layered gneissic rocks is the only U-Pb zircon age from TBC orthogneiss (Appendix 5, Figure 28). Three orthogneiss samples on the east side of the Charles Lake shear zone, collected from the ‘Potts Lake block’ and ‘Andrew Lake block’, as defined by McNicoll et al. (2000), could be relevant to the orthogneiss ages within the Andrew Lake map area. From north to south, these are as follows:

1) A biotite-bearing amphibolite gneiss with a tonalitic migmatite component, collected just north of the Alberta–Northwest Territories border, from the ‘Potts Lake block’, yielded an upper intercept age of ca. 2295 Ma, interpreted as the age of migmatization of the amphibolite gneiss country rock; other analyses from this sample show either clear inheritance or younger Pb loss (Appendix 5, Figure 29).

2) A well-layered biotite tonalite about 1 km east of CLSZ, within the ‘Andrew Lake block’, has an estimated U-Pb zircon age of ca. 3076 Ma, which supports the Archean inheritance of at least some of the Taltson granitoid gneiss units, and monazite metamorphic ages ranging between ca. 1930 and 1917 Ma (Appendix 5, Figure 30).

3) A biotite-rich tonalite gneiss from an isolated inlier of layered basement gneiss at Lapworth Point on north shore of Lake Athabasca, within the ‘Andrew Lake block’, yielded a well-constrained U-Pb zircon upper intercept age of ca. 2562 Ma (Appendix 5, Figure 31).

3.2 Rutledge River Complex

Mappable bands and minor unmappable lenses of biotite schist interlayered with granite and pegmatite lenses or dikes, milky quartz pods, augen gneiss, and small amphibolite and mafic lenses are scattered throughout the granite gneiss domain that underlies the western part of the Andrew Lake map area (Figure 6). The biotite-rich lenses have been routinely described as pelitic and semipelitic gneiss and metaquartzite, essentially based on their well-developed layering and the frequent presence of aluminosilicate mineral phases (± pink to red garnet, graphite and tourmaline).

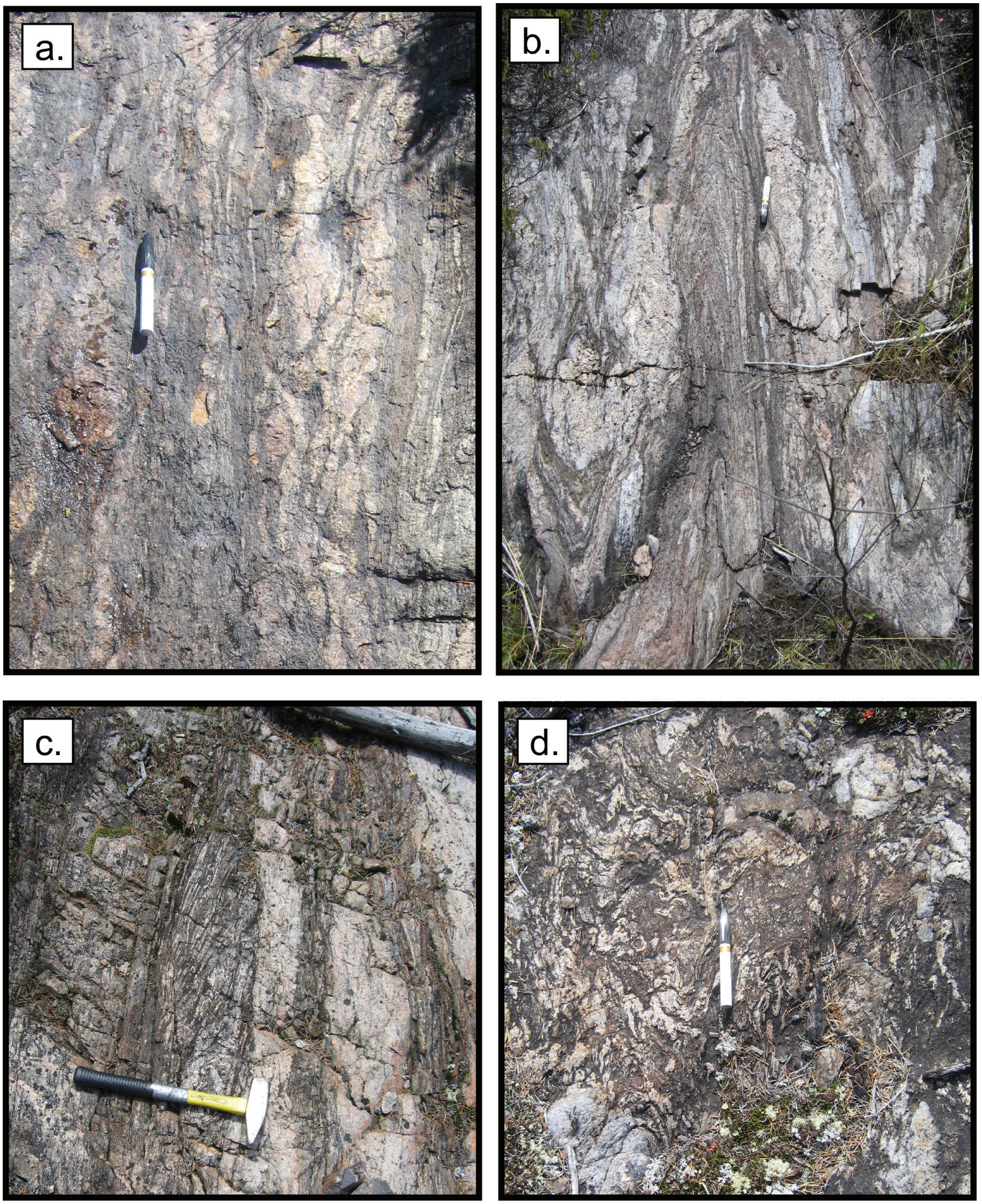


Figure 6. Biotite-garnet±aluminosilicate gneiss and schist ('metasediments'), invariably associated with intense pegmatite veining in the Rutledge River Complex: a) and b) common aspects of folded plane-parallel foliation; c) dismembered pod of biotite gneiss with internal layering obliquely oriented to regional layering (pseudo-crossbedding); d) convoluted foliation in intensely microfolded domains. Foliation and layering are northerly trending (except in fold hinges) and vertical.

Following Bostock and van Breemen (1994), McDonough et al. (2000b, c) interpreted these rocks as sediments deposited within the Rutledge River basin before ca. 2.13–2.09 Ga and later tectonically incorporated in the Rae crust and metamorphosed together under amphibolite- to granulite-grade metamorphic conditions. Larger belts of such biotite-dominated schist and gneiss, which commonly include up to 50% migmatite and intense quartz veining, are found

- immediately west and north of Andrew Lake,
- south of Holmes Lake,
- at Spider Lake, and
- east of Spider Lake (Figure 3).

Minor amphibolite and mafic lenses were noted in these bands and are particularly conspicuous toward the north end of the Spider Lake band. Internal structures in these bands record high strain and commonly exhibit extreme contortion, crenulation and shear structures (Figure 6). Godfrey (1961) interpreted the layering, particularly the interlayering of quartzite and biotite-sericite schist, and the colour banding in quartzite to represent transposed primary bedding, but noted that “no evidence of graded bedding, crossbedding, minor erosion surfaces, and similar features was found” (Godfrey, 1961, p. 12), and described the biotite gneiss and schist “with primary sedimentary features being absent” (Godfrey, 1963, p. 12). Godfrey (1961) identified many quartzite units, which range in thickness from <1 m to >10 m, that are, in fact, banded siliceous mylonite bounded by sheared quartzofeldspathic rocks, with flaser and mortar mylonitic texture.

The author’s preliminary observations indicate that common mineral assemblages in the Rutledge River Complex include quartz–potassium feldspar–biotite–garnet–aluminosilicate (sillimanite and/or andalusite) ±cordierite±graphite. Rock types consisting of various proportions of these minerals represent the distinctive characteristic of the Rutledge River Complex and will be further described with the generic term biotite±aluminosilicate gneiss. Isolated occurrences of spinel inclusions in cordierite and garnet (Swinnerton Lake), and of pyroxene in biotite-amphibole gneiss (Holmes Lake), suggest that the apparent amphibolite-facies Rutledge River Complex is the result of retrogression from granulite-facies rocks.

Grover et al. (1997) reported stable spinel-quartz mineral assemblages from the Rutledge River Complex in the west part of the Alberta shield, calculated peak metamorphic conditions of ~6 kbar and 800–900°C, and suggested successive overprinting of granulite-grade assemblages by amphibolite- and greenschist-grade assemblages along zones of strain concentration and hydration. Even higher temperature mineral assemblages, including orthopyroxene-spinel-quartz, and peak pressure temperature conditions of ~7 kbar and 920–1045°C have been documented for correlative gneiss units in Northwest Territories (Berman and Bostock, 1997).

The occurrence of zoned pyroxene relicts within the biotite-amphibole gneiss south of Holmes Lake (Pană et al., 2006) suggests that the peak metamorphic conditions in the Andrew Lake area may have reached the upper-granulite facies documented in the Northwest Territories (Berman and Bostock, 1997) but re-equilibrated under amphibolite or lower metamorphic conditions. The six-phase mineral assemblage of the typical Rutledge River schist and gneiss mapped in the Andrew Lake area is consistent with the retrograde reaction due to fluid access at the granulite–upper amphibolite transition:

$$\text{garnet} + \text{cordierite} + \text{potassium feldspar} + H_2O \leftrightarrow \text{biotite} + \text{sillimanite} + \text{quartz}$$

Further retrogression of the biotite±aluminosilicate gneiss to interlayered quartz-biotite schist, chlorite-sericite schist, phyllonite and granoblastic quartz-rich layers occurs in narrow bands (Appendix 1). Minor retrogressive shear zones are locally marked by hematite and pyrite, and highlighted by weathering as gossans and rusty zones.

Rutledge River schist and gneiss units are commonly interlayered and folded with, or crosscut by, granite or pegmatite dikes, lenses, pods or stringers to form migmatite characterized by tight isoclinal or occasionally chevron folds, drag folds and other more complicated forms indicative of plastic flow (Figure 6).

The contacts of the Rutledge River Complex with adjacent granitoid gneiss of the Taltson basement complex are concordant and gradational, either through migmatitic mixed-rock zones or shear contacts. In the shear contacts, bands of granite gneiss and biotite±aluminosilicate gneiss alternate over distances of a few metres to one hundred metres, and small masses of sliced-off granite gneiss are interfingered with the biotite±aluminosilicate gneiss. The migmatitic gradational contact consists of a 10–100 m wide zone of contorted migmatite that invades both the granitoid gneiss and the biotite±aluminosilicate gneiss.

Uranium-lead analyses of single metamorphic monazite grains from granulite-facies ('pelitic') gneiss near the Leland Lakes shear zone have $^{207}Pb/^{206}Pb$ ages that cluster in the 1926–1923 Ma range and can be interpreted to record cooling through the 725°C isotherm following high-temperature deformation (Appendix 5, Figure 32; McDonough et al., 2000c; Heaman and Parish, 1991). The $^{207}Pb/^{206}Pb$ age of ca. 1925 Ma yielded by one monazite grain and two ages of ca. 1919 Ma yielded by two other monazite grains from a biotite-spinel ('pelitic') gneiss near the Charles Lake shear zone may indicate protracted deformation and exhumation of this more easterly shear zone until ca. 1919 Ma, or alternatively may represent crystallization below the closure temperature of Pb in monazite (McDonough et al., 2000c).

Additional, nearly concordant monazite analyses from Taltson plutons and orthogneiss, corroborated by concordant zircon ages and titanite ages, constrain the age of the granulite-facies metamorphism to slightly older than ca. 1932 Ma. Younger concordant zircon (1910 Ma), monazite (1913 Ma) and titanite (1916 Ma) ages probably represent mineral growth at or below the respective mineral closure temperature (McDonough et al., 2000c).

These data from the Alberta portion of the TMZ are consistent with and complemented by existing data from the TMZ segment to the north in Northwest Territories, where various rock types assigned to the ‘Rutledge River Group’ record post high-grade metamorphism cooling at ca. 1905–1910 Ma (‘Thuburn Lakes pelitic paragneiss’), ca. 1920 Ma (‘Hill Island impure quartzite’), and an older phase at ca. 2060 Ma (‘Mama Moose pelitic paragneiss’ and ‘Tsu Lake quartzite’; Bostock and van Breemen, 1994). In these samples from Northwest Territories, U-Pb monazite dates are augmented by analyses of igneous zircon (interpreted as detrital zircon), which yielded ages ranging from roughly 2000 to 2600 Ma. This range overlaps with the range of the most common ages obtained from zircon grains of Taltson basement complex orthogneiss and suggests a close genetic relationship between the Rutledge River Complex and Taltson basement complex. The lack of Taltson age igneous zircon (ca. 2000–1900 Ma) in the Rutledge River Complex indicates that the medium- to high-grade metamorphic complex formed prior to the Taltson plutonism.

3.3 Waugh Lake Complex

In the area of Doze and Waugh lakes near the Alberta-Saskatchewan border, a distinct complex of biotite-rich quartzitic schist, phyllonite, quartzite and minor milky quartzite pods was originally mapped by Godfrey as ‘Waugh Lake Formation’; it also includes medium- to coarse-grained chlorite- and biotite-rich mafic schist and fine-grained, intermediate to mafic igneous rocks that have been interpreted as volcanic rocks deformed at greenschist- to subgreenschist–facies metamorphic conditions (Godfrey, 1963).

Watanabe (1965) expanded the mapping of the low-grade rocks east of the Alberta border and showed that they are correlative with a belt of low-grade units mapped in Saskatchewan (Koster, 1961, 1971). A formal nomenclature was introduced by Iannelli et al. (1995), who described members and formations of the ‘Waugh Lake Group’.

McDonough and McNicoll (1997) described the ‘Waugh Lake Group’ as a small outlier of strata deposited in an intra-arc basin near the eastern periphery of the Taltson magmatic arc (Figure 2). Uranium-lead dates on 'detrital' zircon from a 'metaconglomerate' layer (mostly in the range 2.0–2.3 Ga, with one grain of ca. 2.7 Ga) and an emplacement age of ca. 1971 Ga for the Colin Lake granitoid, interpreted to intrude the basal units of the volcano-sedimentary sequence, were inferred to constrain the timing of deposition of the lower part of the ‘Waugh Lake Group’ to between ca. 2020 and 1971 Ma.

The Andrew Lake ductile thrust zone was postulated to have developed around ca. 1932 Ma and to have carried Taltson basement complex gneiss in its hangingwall northeastward over plutonic suites (Andrew Lake, Colin Lake and Wylie Lake granitoid plutons) and the ‘Waugh Lake Group’ (McDonough et al., 2000c). In this interpretation, hangingwall gneiss is at upper amphibolite to granulite facies, whereas the metamorphic conditions in footwall rocks ranged from amphibolite facies at the thrust to lower-greenschist facies in the ‘Waugh Lake Group’, 2 km to the east. Several lines of evidence are inconsistent with this interpretation of the low-grade succession in the Waugh Lake area. A brief review of the data that led to the current interpretations and a summary of recent observations and analytical data collected by AGS in recent years are included in Section 4.

3.4 Taltson Plutonic Rocks

In the Andrew Lake area, the Taltson magmatic event is represented by the emplacement of relatively voluminous Andrew Lake and Colin Lake granitoid plutons and by smaller, muscovite-bearing granite intrusions, previously interpreted as phases of the adjacent larger plutons. The author’s field observations suggest that the muscovite-bearing granitic rocks are typically associated with zones of strain concentration and may represent late-stage igneous phases emplaced at shallower structural level than the hornblende and biotite granitoid plutons.

The Colin Lake pluton consists of several granitoid phases that crop out over a large area along the Alberta-Saskatchewan border, from Waugh Lake in the north to Colin Lake, some 30 km to the south (Godfrey, 1986a). Around Colin Lake, the dominant phase is a megacrystic granodiorite, whereas, at Waugh Lake, the Colin Lake pluton is represented by a medium-grained, equigranular diorite. The emplacement age of the pluton was inferred based on U-Pb zircon analyses from a weakly deformed equigranular biotite quartz diorite from the north shore of the west arm of Waugh Lake (McDonough and McNicoll, 1997). A regression line through five single-grain analyses yielded an upper intercept age of 1971±4 Ma, which roughly coincides with $^{207}Pb/^{206}Pb$ ages of the two most concordant zircon grains (Appendix 6, Figure 33).

Two north-south–elongated plutons of Andrew Lake biotite-hornblende granodiorite have been previously mapped and dated by the U-Pb zircon method (Godfrey, 1961, 1986a; McDonough et al., 2000c): one of them, located east of Andrew Lake, yielded a crystallization age of 1962 +16/–10 Ma; and the second, located west of Swinnerton Lake, yielded a crystallization age of 1959 ±3 Ma (Appendix 6, Figures 34 and 35). The cores of the plutons are relatively massive to weakly foliated biotite granite and may include a more feldspathic and massive grey granite phase. The peripheries of the plutons are foliated, with K-feldspar megacrysts set in a foliated matrix of feldspar, biotite, quartz and minor hornblende. The strained peripheries of the Andrew Lake plutons grade into high-strain rocks and are commonly intruded by syn- to late-kinematic pegmatite, microgranite and aplite bodies ranging from 1 cm to more than 3 m in width. The foliation is northerly trending, steeply dipping and typically well expressed by coarse-grained biotite and lenticular quartz matrix with aligned feldspar porphyroblasts/clasts.

Spatially associated with the Charles Lake and Bayonet Lake high-grade mylonite belts and their peripheries of Rutledge River rocks, a series of sub-kilometre–size bodies of massive to foliated, occasionally megacrystic granite have been previously assigned to the Charles Lake granite (McDonough et al., 2000b, c; Figure 3). This granitoid rock type has been variably deformed to amphibolite- and greenschist-grade, strike-lineated protomylonite and mylonite. Several multigrain fractions from a sample of weakly foliated Charles Lake granite yielded very discordant data points, indicating severe Pb loss (Appendix 6, Figure 36). The least discordant zircon fraction has a $^{207}Pb/^{206}Pb$ age of ca. 1933 Ma, which coincides with the age of three slightly discordant monazite grains and has been taken as the best estimate of the minimum emplacement age. A highly discordant zircon fraction provides evidence of inheritance from ca. 2.1 Ga, whereas a monazite grain with an age of ca. 1919 ±2 Ma is interpreted to record the timing of amphibolite-grade metamorphism and deformation (McDonough et al., 2000c).

In the same setting, between the CLSZ and the BLSZ (Figure 3), a white or light grey to pink, muscovite-bearing (with rare biotite), coarse-grained to megacrystic granite described by Godfrey (1966) was named the Treasure Loch granite and assigned to a group of late, minor, muscovite-bearing granite plugs that mark release bends along regional sinistral shear zones (Pană et al., 2006). Its age has not yet been determined, but the similar muscovite-bearing Benna Thy granite to the north along the CLSZ in the Northwest Territories has been dated at ca. 1906 Ma (Bostock and Loveridge, 1988). To the south, the Chipewyan red granite (Figure 1), with a locally well-expressed chlorite-biotite tectonic fabric that records the late deformation along the CLSZ, yielded a crystallization age of ca. 1925 Ma (Appendix 6, Figure 37).

Within the Andrew Lake–Cherry Lake area, Godfrey (1961) mapped a considerable area of massive biotite leucogranite (with white to pink to red feldspars and minor sericite) containing local areas of massive muscovite granite (with abundant white to pink feldspars and minor biotite) but did not assign a name to these rocks. In a subsequent 1:250 000 compilation, Godfrey (1986a) showed these rocks as forming the northern part of the Colin Lake granitoid bodies, whereas McDonough et al. (2000b) remapped some of them as the Colin Lake white granite (distinct from and younger than the Colin Lake granite). In this report, the massive biotite±muscovite granite, as originally mapped by Godfrey (1961), is referred to as the Wallace Island granite (after an island in Andrew Lake). Analyses of tiny, poor-quality zircons from a sample collected from the prominent peninsula near the north end of Andrew Lake revealed evidence of both inheritance and Pb loss, with the least discordant analysis yielding a 1923 ±2 Ma age. Three monazite grains yielded concordant to slightly discordant ages of ca. 1922, 1917 and 1913 Ma. The inferred emplacement age is 1923–1921 Ma, with monazite growth during cooling between 1921 and 1913 Ma (McDonough et al., 2000c). A fourth, larger monazite grain with different morphology and lower U content yielded an age of ca. 1933 Ma and was interpreted as a xenocryst from the adjacent Andrew Lake granodiorite (Appendix 6, Figure 38).

The late muscovite-bearing granite bodies are cut by pink granite, muscovite and/or feldspar pegmatite dikes, which are typically associated with the contact between the Rutledge River Complex and Taltson basement complex or Taltson plutons.

South of Waugh Lake and on strike with the main body of schist within the Waugh Lake Complex, the leucocratic Ney Lake granite contains minor muscovite and biotite and is overprinted by small shear zones parallel to an incipient northerly-trending foliation marked by sericite and chlorite. This foliation is concordant and at the same metamorphic grade as the foliation of the Waugh Lake Complex, which suggests that the Nay Lake granite was overprinted by the same tectonometamorphic event as the Waugh Lake Complex. The emplacement age of the Nay Lake granite remains unknown. Godfrey (1963) described its contact with the biotite-dominated granitoid phases of the Colin Lake pluton as gradational. However, to the north in Northwest Territories, a belt of greenschist-facies rocks similar to the Waugh

Lake Complex along the Tazin River shear zone is plugged by the ca. 1934 Ma Natael muscovite granite at Hill Island Lake and by the ca. 1813 Ma Thekulthili granite a few kilometres farther north (Bostock and van Breeman, 1994).

The muscovite-bearing granite bodies intrude and crosscut ductile shear fabrics and are massive to weakly foliated. They are likely late to post kinematic relative to the ductile strain recorded by Rutledge River and Waugh Lake complexes, and may have been emplaced independently of the main Taltson plutons. Field relationships suggest a decompression melting origin for the muscovite-bearing granite bodies within zones of ductile shearing during late phases of tectonothermal evolution of the TMZ, accompanied by progressive exhumation of the crust and concentration of strain in major shear zones.

4 Waugh Lake Complex: New Data and Their Significance

4.1 Evolution of Ideas

The low-grade metamorphic rocks of the Waugh Lake area were originally mapped and described by Godfrey (1958, 1961, 1963), who noted the following:

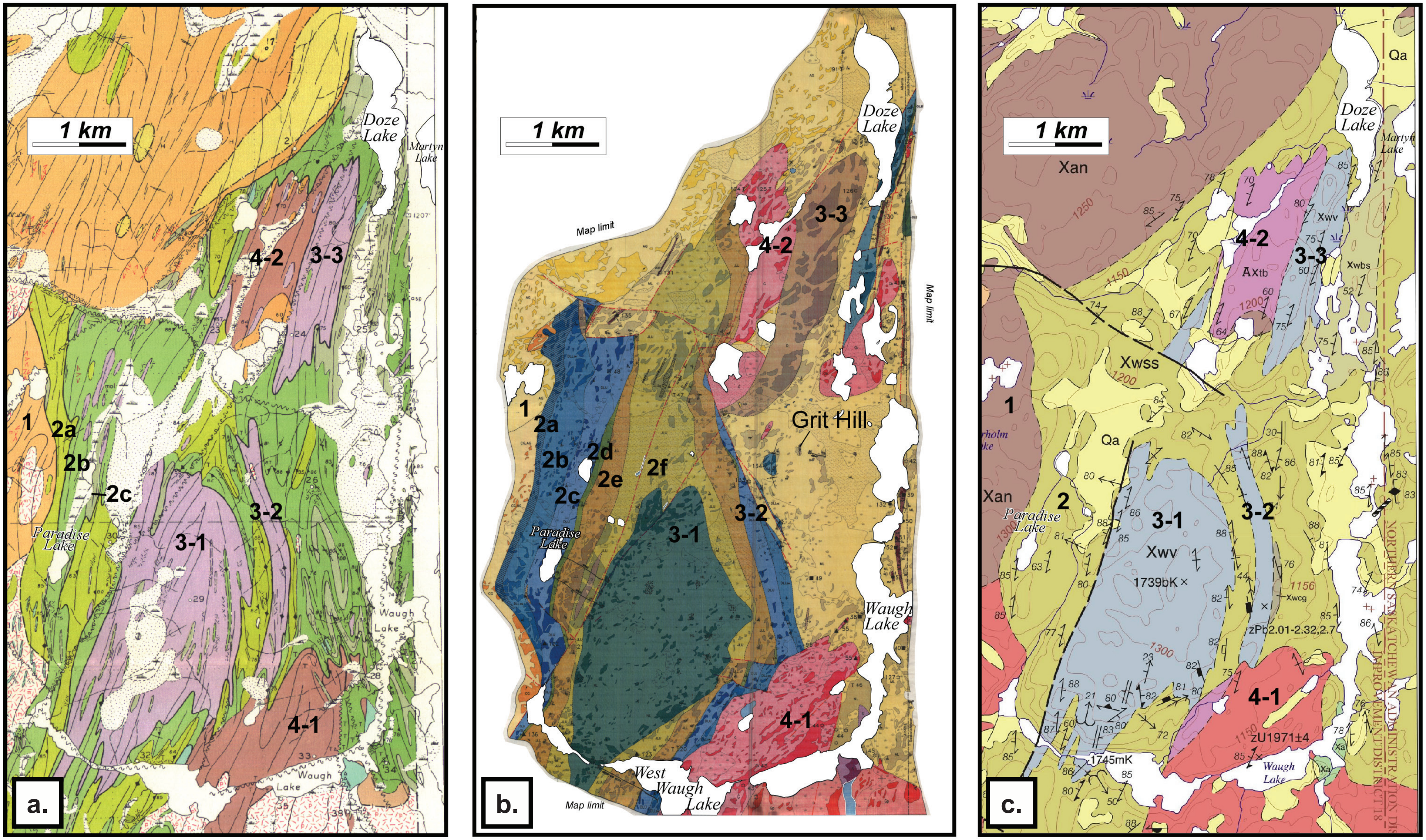
"The mode of origin of these rocks is open to speculation. The cataclastic features of the rock largely obscure the critical information needed to determine its origin and history. The large fragments and pebbles may be of clastic origin, but cataclastic deformation of a granitic phase, such as porphyroblastic biotite granites A and B [author's note: foliated and massive Andrew Lake granitoid, respectively], could also give rise to crush conglomerate and to a sericitic, porphyroclastic phyllonite. A combination of these modes of origin is also possible." (Godfrey, 1963, p. 11)

With this cautionary note, Godfrey went on to describe the 'Waugh Lake Formation' as a volcano (30%)–sedimentary (60%) sequence consisting of interlayered metasedimentary (impure quartzite, siltstone, biotite schist and sericitic porphyroclastic phyllonite) and metavolcanic rocks (basaltic flows and tuffs). However, Godfrey's reservations regarding the nature of the protoliths are well reflected in the legend of his 1:31 680 scale geological maps (Godfrey, 1961, 1963). The three map units of possible sedimentary origin in the Waugh Lake area include 'quartzite', 'biotite schist' and 'sericitic porphyroclastic phyllonite', in which various proportions of strained granite and pegmatite are interfingered with quartzite, fine-grained schist, and 'crush conglomerate and minor conglomerate?' (Table 1). Apparently, Godfrey recognized the tectonic origin of some fragmental rocks and had doubts about the origin of other ones, for which he could not rule out a sedimentary origin.

Initial K-Ar dates (Godfrey and Baadsgaard, 1962; Baadsgaard and Godfrey, 1967, 1972; Appendix 3) from Godfrey's 'Waugh Lake Formation' (1860–1740 Ma) were within the age range of the adjacent high-grade gneiss (1940–1620 Ma) and granitoid (1930–1740 Ma), so they could not be used to constrain the geological history of the area. A 1760 Ma K-Ar biotite date from a 'basalt flow' and its low-grade metamorphism were interpreted to indicate that the Waugh Lake sequence is younger than the adjacent Taltson basement orthogneiss (Baadsgaard and Godfrey, 1972). Langenberg et al. (1993) proposed an unconformity contact between low-grade rocks and their inferred higher grade basement, the Taltson basement complex, whereas Salat et al. (1994) pointed out that this unconformity could not be found and the basement of Waugh Lake sequence is therefore unknown. The low-grade rocks around Waugh Lake were considered to be strongly tectonized with deformation "expressed by extensive folding and mylonitization", and rock units were "largely interpreted as tectonites rather than as sedimentary and volcanic rocks" (Salat et al., 1994, p. 2 and 3, respectively). For most map units defined within the Waugh Lake Group, Salat et al. (1994) found the interpretation of the origin problematic. Sedimentological and, alternatively, tectonic processes were invoked, with the tectonic ones being considered "at the moment

Table 1. Comparison of map unit descriptions by different authors in the portion of the Waugh Lake area represented in Figure 7.

	Godfrey (1961)	Iannelli et al. (1995)	McDonough et al. (2000)
1	Foliated biotite granite A	Andrew Lake granite	Andrew Lake granodiorite
	Low grade	***Waugh Lake Group***	***Waugh Lake Group***
2a	- sericitic, porphyroclastic phyllonite - crush conglomerate and minor conglomerate?	(2a) Doze Lake Fm., Lower Member: layered porphyroblastic gneiss with pegmatitic to porphyritic granite dikes	- paragneiss: medium- to coarse-grained gneiss sheared under low-grade conditions (sericite- or chlorite-rich) - pebble to granule conglomerate, metagreywacke; foliated, muscovitic, feldspathic - biotite schist, phyllite, phyllonite, minor quartzite, pegmatite
2b	- quartzite, phyllite, biotite schist - ferruginous, garnetiferous and graphitic zones - feldspar augen, granite and pegmatite lenses	(2b) Doze Lake Fm., Lower Member: medium-grained to pebbly subarkose to sublitharenite with minor conglomerate (2c) Doze Lake Fm., Upper Member: rhyolitic to dacitic flows and tuffs with minor reworked tuffs	
2c	- biotite schist, sericite-chlorite schist - slate, phyllite, phyllonite, quartzite lenses, fragments - ferruginous, garnetiferous and graphitic zones - feldspar-augen granite and pegmatite lenses - crush conglomerate and minor conglomerate**?**	(2d) Sederholm Lake Fm: actinolite- and biotite-bearing, fine-grained to pebbly subarkosic wacke and sublitharenite with minor conglomerate and tuff	
		(2e) Johnson Lake Fm., Lower Member: interlayered medium-grained to pebbly subarkose and sublitharenite with minor conglomerate and tuff	
	mostly 2a	(2f) Johnson Lake Fm., Upper Member: interlayered felsic tuffs, lapilli and reworked tuffs, minor subarkose to sublitharenite	
3-1	various basic rocks: greenstone, amphibolite, basalt, gabbro and possibly metatuff	Niggli Lake Fm: mafic pyroclastic breccia, reworked mafic to intermediate tuff and sublitharenite	Waugh Lake volcanic rocks: medium- to coarse-grained mafic gneiss sheared under greenschist to sub-greenschist grade (chlorite- and biotite-rich)
3-2	"	Doze Lake Fm.: Lower and Upper members	"
3-3	"	hornblende-biotite diorite	"
4-1	foliated biotite microgranite	equigranular hornblende-biotite granite	Colin Lake granite (main phase) foliated, lineated, mylonitic
4-2	"	equigranular hornblende biotite granite (2e) Johnson Lake Fm.	Taltson basement complex: layered biotite-hornblende granite to granodiorite, hornblende diorite gneiss



Godfrey (1963) Iannelli et al. (1995) McDonough et al. (2000)

Figure 7. Published geological maps of the main area encompassed by Waugh Lake Complex in Alberta, showing significant discrepancies in rock identification, their grouping in map units, extent of map units, and intensity of deformation. Legends for the Waugh Lake map units in each map are in Table 1, and labelled units are discussed in the text. See Figure 3 for regional geological context.

not the preferred explanations" (Salat et al., 1994, p. 18). The stratigraphic approach to mapping of metamorphic tectonite, which at the time was rapidly gaining popularity, led to the mapping of "diverse facies assemblages that can be arranged into a coherent stratigraphic system" to define the Waugh Lake Group (Iannelli et al., 1995, p. 2). Their postulated stratigraphic succession included five formations:

1. Martyn Lake Formation, consisting of rhythmically bedded, turbiditic rocks;
2. Doze Lake Formation, which constitutes the 'lower' sedimentary and volcanic assemblage;
3. Sederholm Lake Formation, including distinctive green-, brown- to black-grey–weathering mafic sedimentary rocks;
4. Johnson Lake Formation, a second sedimentary and volcanic assemblage; and
5. Niggli Lake Formation, dominated by mafic volcanic flows and breccia (Table 1).

Comparison of Figure 7a and b shows how the purely stratigraphic approach to mapping adopted by Iannelli et al. (1995) resulted in significant departure from the original petrographic mapping of Godfrey (1961). The original map depicts intense internal deformation within the Waugh Lake assemblage, including spectacular infolding of igneous and schist map units (e.g., the southern portion of units 3-1 and 3-3, or the northeastern part of unit 4-1), as well as shreds of schist within the main igneous bodies and minor pods of igneous rocks within schist units. In contrast, the stratigraphic mapping depicts a syncline with Niggli Lake Formation (unit 3-1) in its axial zone. The limbs of the postulated syncline would be defined by the original stratigraphic succession of clastic and volcanic strata. In this interpretation, unit 3-2, previously represented as a lens of mafic rocks, has become the 'Doze Lake Formation', whereas four map units immediately west of the granite gneiss (unit 4-2) are lumped together into the 'Johnson Lake Formation' (see Figure 7). The underlying concept was that

1. biotite-grade metamorphism had not significantly affected original bedding and intrusive contacts,
2. planar structural elements represent bedding, and
3. observable textures were essentially syndepositional.

The interpretation of the rocks in the Waugh Lake area as a stratigraphic succession has been incorporated in the 1:50 000 scale map by McDonough et al. (2000b). Whereas the most recent map rejected the syncline interpretation (e.g., see their unit 3-2) and recognized the internal deformation of the Waugh Lake assemblage (see the southern portion of unit 3-1 and the northern margins of units 4-2 and 3-3), it has also changed the original interpretation of unit 4-2 and the western periphery of unit 4-1 by describing them as Taltson orthogneiss (Figure 7c). However, this designation could also have resulted from a colouring error in the map (McDonough, pers. comm., 2007). Although the direct correlation of the low-grade Waugh Lake assemblage mapped in Alberta with the 'western conglomerate' belt mapped in Saskatchewan was already established (Watanabe, 1965; Koster, 1971), the Waugh Lake Group was still described as "a small outlier…in the southern Taltson magmatic zone" deposited in "a small intra-arc basin" (McDonough and McNicoll, 1997, p. 102 and 101, respectively).

4.2 New Observations and Analytical Data from the Waugh Lake Complex

Geological mapping was conducted along several traverses in the area of Waugh Lake during August 2005 and July 2006. Particular attention was paid to the contacts between map units, type of strain (pure vs. simple shear) and nature of protoliths (sedimentary vs. igneous/metamorphic). In addition, an attempt was made to integrate geological knowledge from both sides of the Alberta-Saskatchewan border.

The most striking feature of the low-grade sequence is the occurrence of intense deformation, expressed by

1) heterogeneously penetrative, roughly northerly-trending and steeply dipping foliation associated with

2) locally obvious high rotational strain,

3) lithological transitions through strain gradients and metamorphic reactions, and

4) strain partitioning.

The purely stratigraphic interpretation of the low-grade rocks and their representation on the map as strata and formations assigned to the Waugh Lake Group (Iannelli et al., 1995) appears tenuous in light of the observed lack of continuity of mappable units and widespread interlayering and folding of massive to foliated igneous rocks with various low-grade schist units. In addition, even accepting the tenuous local correlations proposed by Iannelli et al. (1995) over an area approximately 5 km across west of the Saskatchewan border, the postulated stratigraphy does not take into account the entire belt of low-grade rocks extending more than 100 km in Saskatchewan (Koster, 1971) and the Northwest Territories (Mulligan and Taylor, 1969), of which the Alberta section is only a small segment (see Figure 3 and Section 4.3).

The low-grade metamorphic succession near Waugh Lake is unified in a general way by lithological features (structural/deformational features and metamorphic grade), mutual relations and evolution, but there is insufficient information to justify its designation as a formal stratigraphic unit. Therefore, the low-grade assemblage of igneous, metamorphic and possibly sedimentary rocks mapped in the Waugh Lake area are more appropriately described by an informal term, such as 'complex', 'lithozone', 'lithotectonic unit' or 'lithotectonic assemblage' (Neuendorf et al., 1997). In this report, the term Waugh Lake Complex is assigned to this assemblage of rocks.

4.2.1 Rock Types

Except for rocks with obvious massive igneous texture, foliated rocks have been differently interpreted, described and grouped in map units by different authors (Table 1). Godfrey's (1961,1963) original petrographic descriptions have been modified by Iannelli et al. (1995) to genetic terms based on the interpretation of textures in foliated rocks as primary sedimentary features. In most cases, the interpretation of the textures, both in the field and in thin sections, is equivocal. In particular, the designation of fine-grained schist units with no distinctive textural features, as well as various types of fragmental rocks, as sedimentary and the identification of rock type based on grain size may be inaccurate and misleading. Similar textures characterize the brittle-ductile transition zone and the overlying zone of low-grade metamorphism within crustal-scale shear zones (e.g., Sibson, 1977, 2001). Convergence of distinct genetic processes towards similar textures has been long recognized, although the resulting rocks often remain contentious.

In most cases, careful outcrop examination at and around 'quartz pebble conglomerate' units reveals several precursor stages of deformation of fine-grained protoliths, with polyphase segregation of quartz veins aligned or rotated into parallelism with the vertical tectonic layering, folding and dismembering of these veins into independent quartz 'pebbles' and 'cobbles' within a quartz-sericite-chlorite (phyllonite) matrix (Figure 8). The protoliths of these rocks may be of sedimentary origin (silt to fine-grained sandstone) or of igneous origin (felsic to intermediate, microcrystalline subvolcanic plugs at dilational jogs along a crustal-scale shear zone). Vertical layers of fragmental rocks consist of slices of variously sheared granitoid with length to thickness ratios of up to 7:1, variously coloured (white, grey, greenish grey or brick red) and strained quartz fragments within a green chloritic fluidal matrix (Figure 9) or brick red limonitic cataclastic matrix (Figure 10).

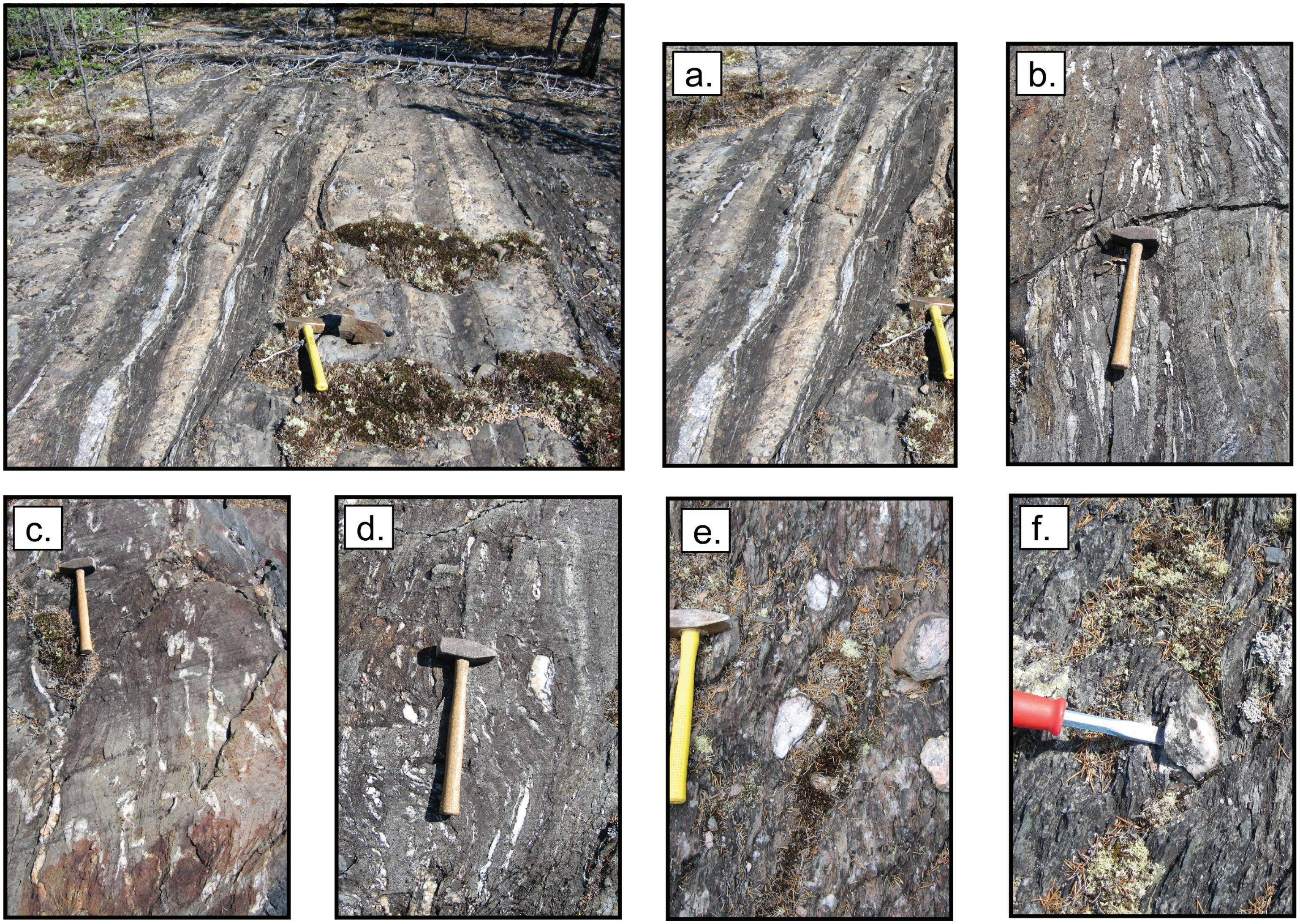


Figure 8. Successive stages in the development of quartz-pebble pseudoconglomerate within the Waugh Lake Complex along the Alberta-Saskatchewan border, near Doze and Martyn lakes: a) and b) quartz segregations within sheared domains of a fine-grained protolith become lenses and boudins through progressive stretching; c) dismembering of quartz veins through folding; d) lens- and pebble-shaped remnants of quartz veins; e) and f) quartz-pebble pseudo-conglomerate.

Initially mapped as ‘crush conglomerates’ by Watanabe (1965), these rocks were reinterpreted as ‘polymictic conglomerate’ or ‘diamictite’ (Salat et al., 1994; Iannelli et al., 1995). These units have the typical characteristics of fault rocks formed at slightly different structural levels within a shear zone: a chlorite matrix is typical for the structural level of the brittle-ductile transition and the next few kilometres above it, whereas the hematitic/limonitic cataclastic matrix is typical for the structural level shallower than 3–4 km. Their lack of continuity along strike (tens to a few hundred metres at most) makes their map representation difficult (included in the description of a unit in text or legend but not shown on map) and their stratigraphic interpretation tenuous. A very particular type of fragmental rock was encountered a few hundred metres southeast of Paradise Lake. Heterogeneous strain partitioning within a leucogranite led to the formation of alternating domains of ductile to cataclastic flow alternating with ‘ribs’ of less deformed granite (Figure 11a, b). As deformation advanced, the ‘ribs’ were consumed and dismembered into undigested ‘pods’ (Figure 11b–d) or ‘eggs’ (Figure 11e, f) in the ductilely flowing mylonitic matrix.

Near the western periphery of the Colin Lake biotite granodiorite, sheared rocks with rotated K-feldspar clasts, σ structures and chlorite-lined shear bands have been interpreted as crossbedded strata (Figure 13) and assigned to the Doze Lake Formation (Iannelli et al., 1995). Field observations indicate that these rocks grade with decreasing deformation into relatively undeformed Colin Lake granodiorite.

Several other rock types, however, show internal textures and outcrop-scale structures that are strikingly similar to sedimentary textures and may indeed have sedimentary protoliths. At a site about 10 m from the east shore of Paradise Lake (UTM Zone 12, 551954E, 6631411N, NAD83; Figure 7), a generally fine-grained green unit locally displays internal structures very similar to crossbedding along vertical layering (Figure 13). The presence in the same outcrop, only a few metres along strike to the south, of spectacular ‘δ’ structures, indicating intense dextral ductile shear along the vertical layers, and of tight isoclinal folds with subvertical fold axes, typical of both the Taltson and Waugh Lake complexes, casts doubt on the sedimentological interpretation of the rock. Moreover, the adjacent unit to the east is a locally sheared diorite, with the same grain size and colouration. The formation of the crossbedding structure through cataclastic flow within a deforming mafic protolith, of which the adjacent dolerite-diorite may be a relict, cannot be ruled out. At the end of his detailed mapping of the Waugh Lake assemblage, Godfrey (1963, p. 9) noted that “unequivocal crossbedding was not observed”. Outcrops that appear to consist of volcanic agglomerate have been encountered between Doze and Martyn lakes (Figure 7, about 2.5 km north of the northern end of Waugh Lake), north of western Waugh Lake, and between the ‘egg’ conglomerate and the large mafic unit, east of Paradise Lake (Figure 14).

A distinctively layered, fine-grained rock unit with alternating sheared and apparently massive planar layers 3–15 cm thick looks like a rhythmically bedded turbidite sequence (Figures 15, 16). Two massive siliceous rock samples (311A and 311B) and one foliated argillaceous rock sample (6610) were collected at a large outcrop approximately 500 m west of northern segment of Waugh Lake from this banded succession assigned to the Martyn Lake Formation by Iannelli et al. (1995). The analytical results for SiO_2, Al_2O_3 and Zr are presented in Table 2, along with some published turbidite analyses for comparison (McLennan et al., 1990). The data indicate that, within the outcrop, colour banding evident on the weathered surface (Figure 16) reflects the juxtaposition of layers with markedly different chemical compositions, consistent with a turbidite interpretation. The high SiO_2 and Zr contents of samples 311A and 311B are within the range of the published data for sandy turbidite layers, which may have elevated zircon concentrations due to sedimentary recycling (McLennan et al., 2003). Alternatively, shearing, transposition and strain partitioning within a felsic to intermediate aplitic protolith may have resulted in the texturally and chemically distinct layers.

Figure 9. Various fragmental rocks of uncertain origin within the Waugh Lake Complex, on the divide between Doze Lake in Alberta and Martyn Lake in Saskatchewan; strained 'fish-like' pebbles and cobbles consist of various stages of previously deformed and retrogressed granite and pegmatite, as well as white quartz from quartz veins and segregations; these rock fragments float within a quartzofeldspathic mylonitic schist with chlorite, sericite and iron oxide/hydroxide.

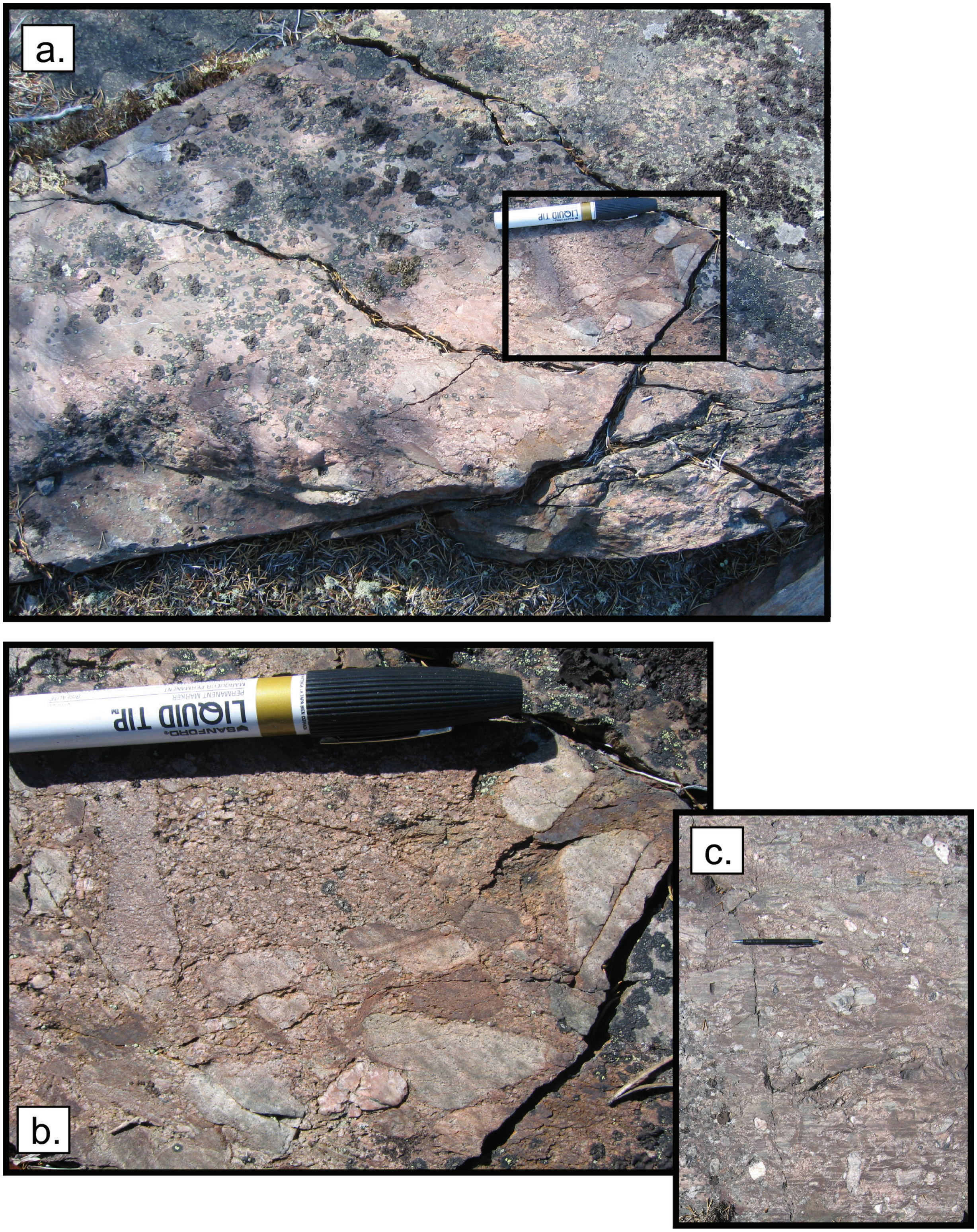


Figure 10. Layer of tectonic breccia within Waugh Lake Complex at ‘Grit Hill’ (near UTM Zone 12, 554194E, 6632714N, NAD83), 750 m west of North Waugh Lake: a) angular and subrounded clasts of variably sheared and retrogressed granite gneiss and vein quartz in a cataclastic hematitic/limonitic matrix; b) detail of boxed area in (a); and c) detail of the cataclastic matrix.

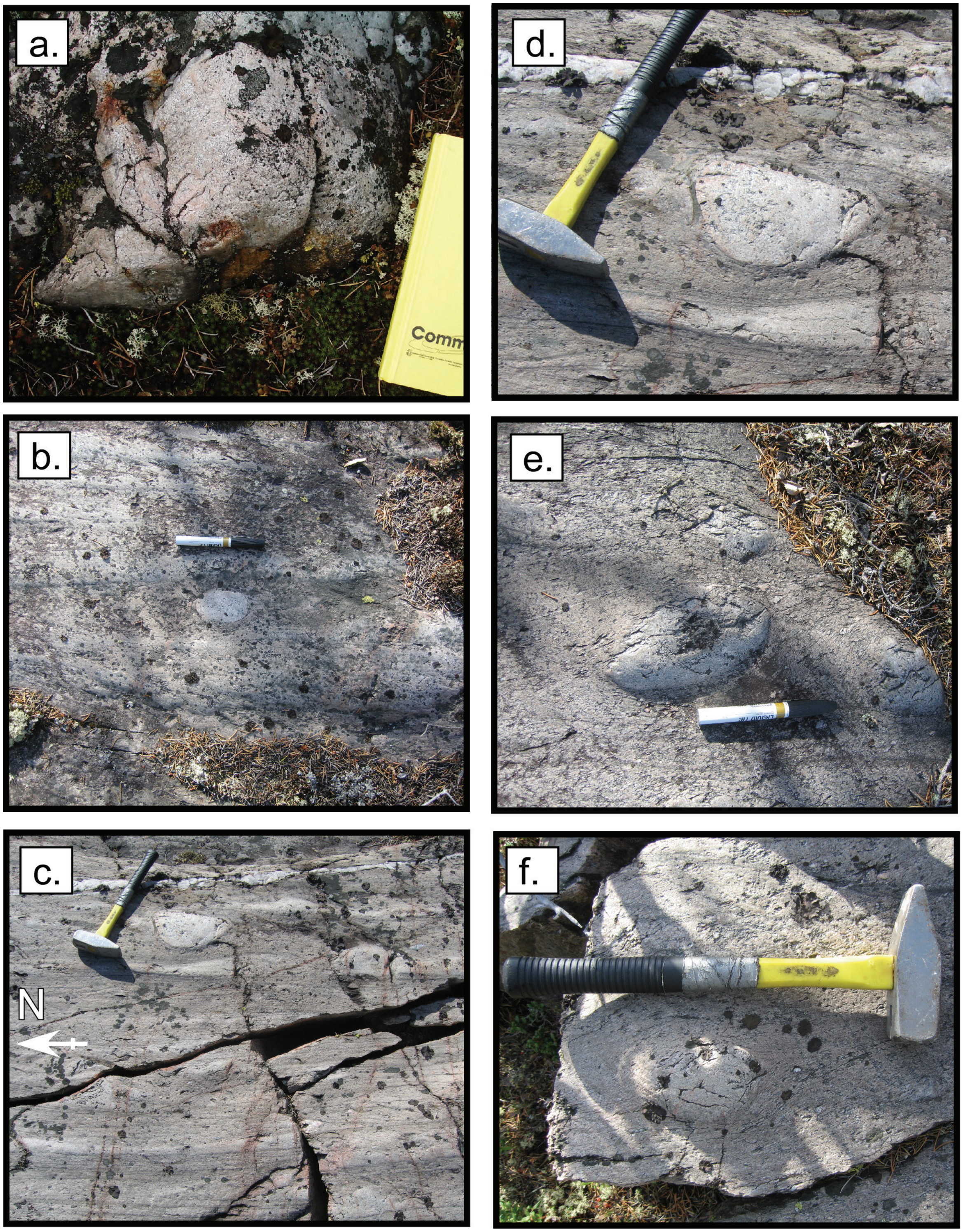


Figure 11. Different stages of strain in leucogranite within the Waugh Lake Complex (near UTM Zone 12, 552014E, 6631376N, NAD83) southeast of Paradise Lake: a) massive leucogranite; b) initial yielding of the massive granite with the development of alternating domains of more (darker) and less (lighter) intense shearing; c) pervasive ductile flow with remnants of less sheared granite as narrow white ribs in the lower half of the photo and dismembered pods in the upper part of the photo; d) detail of (c); e) and f) egg-shaped remnants of granite in the ductilely deforming matrix. Foliation is everywhere subvertical.

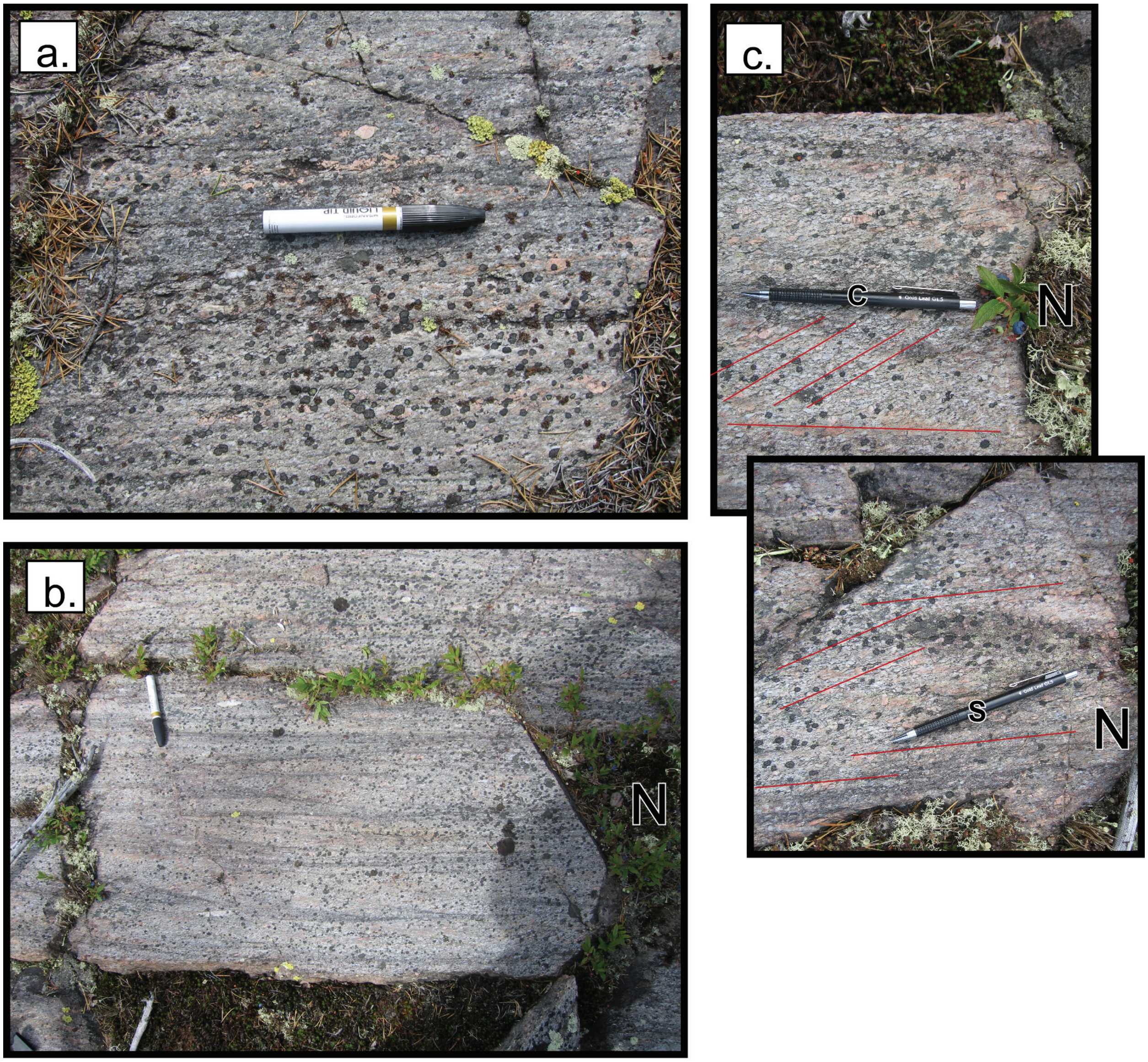


Figure 12. Foliated quartzofeldspathic rock of uncertain origin within the Waugh Lake Complex, south of western Waugh Lake ('Doze Lake Formation' of Iannelli et al., 1995): a) 'sigma' structure suggesting dextral shearing; b) colour banding previously interpreted as crossbedding; c) 'c/s' structure suggesting dextral shearing.

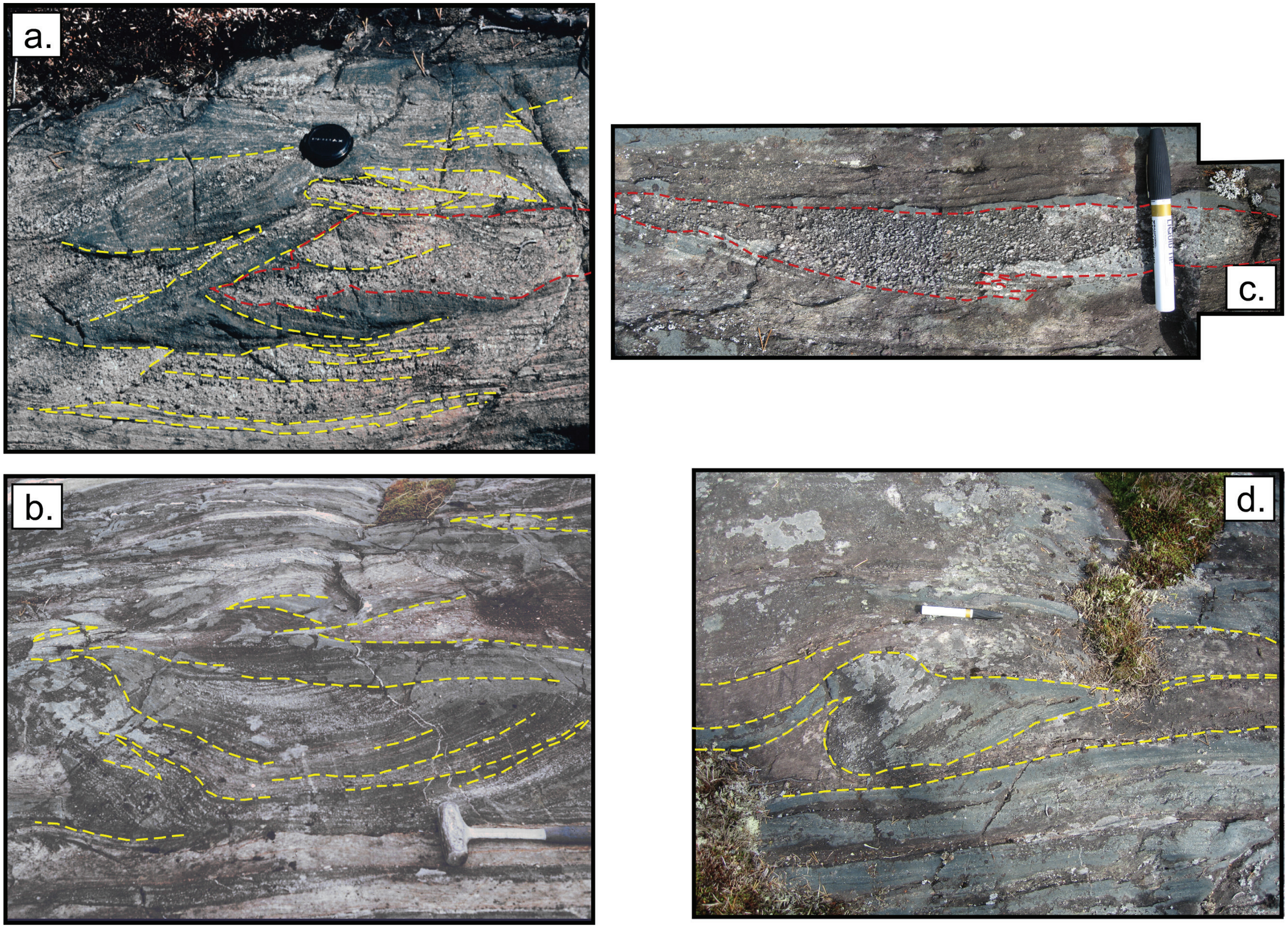


Figure 13. Outcrop-scale structures of possible sedimentary origin associated with structures typical of ductile deformation within the Waugh Lake Complex on the east shore of Paradise Lake (near UTM Zone 12, 551826E, 6631492N, NAD83): a), b) and c) internal domains within layers are locally oriented obliquely to the regional foliation and look similar to crossbedding; note the lack of continuity of the irregular or lens-shaped coarser grained domains highlighted in red in (a) and (c); d) 'delta' object indicating dextral shearing. All foliation types are everywhere subvertical.

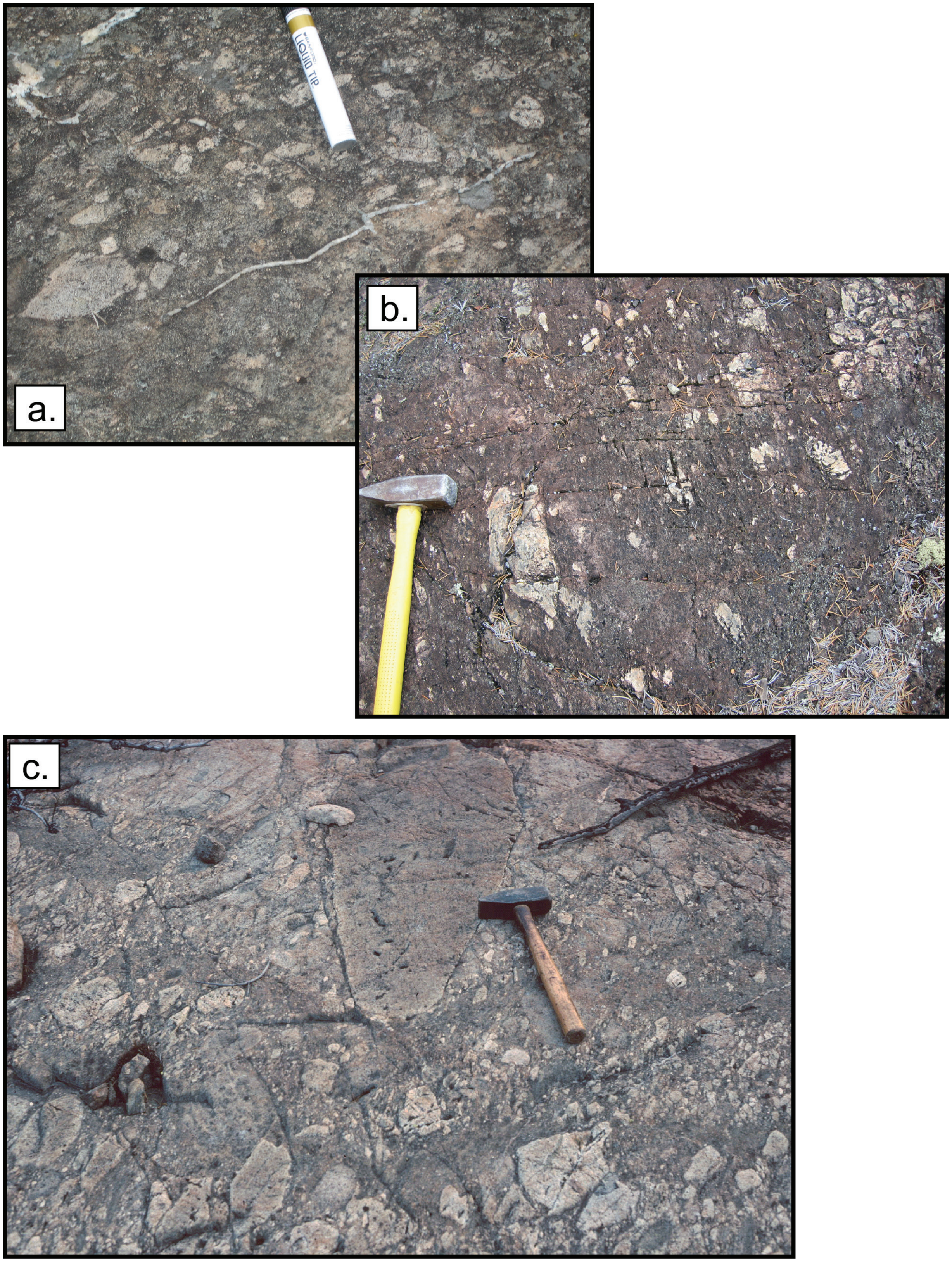


Figure 14. Fragmental rocks resembling volcanic agglomerate, Waugh Lake Complex: a) southeast of Paradise Lake; b) and c) divide between Doze Lake in Alberta and Martyn Lake in Saskatchewan.

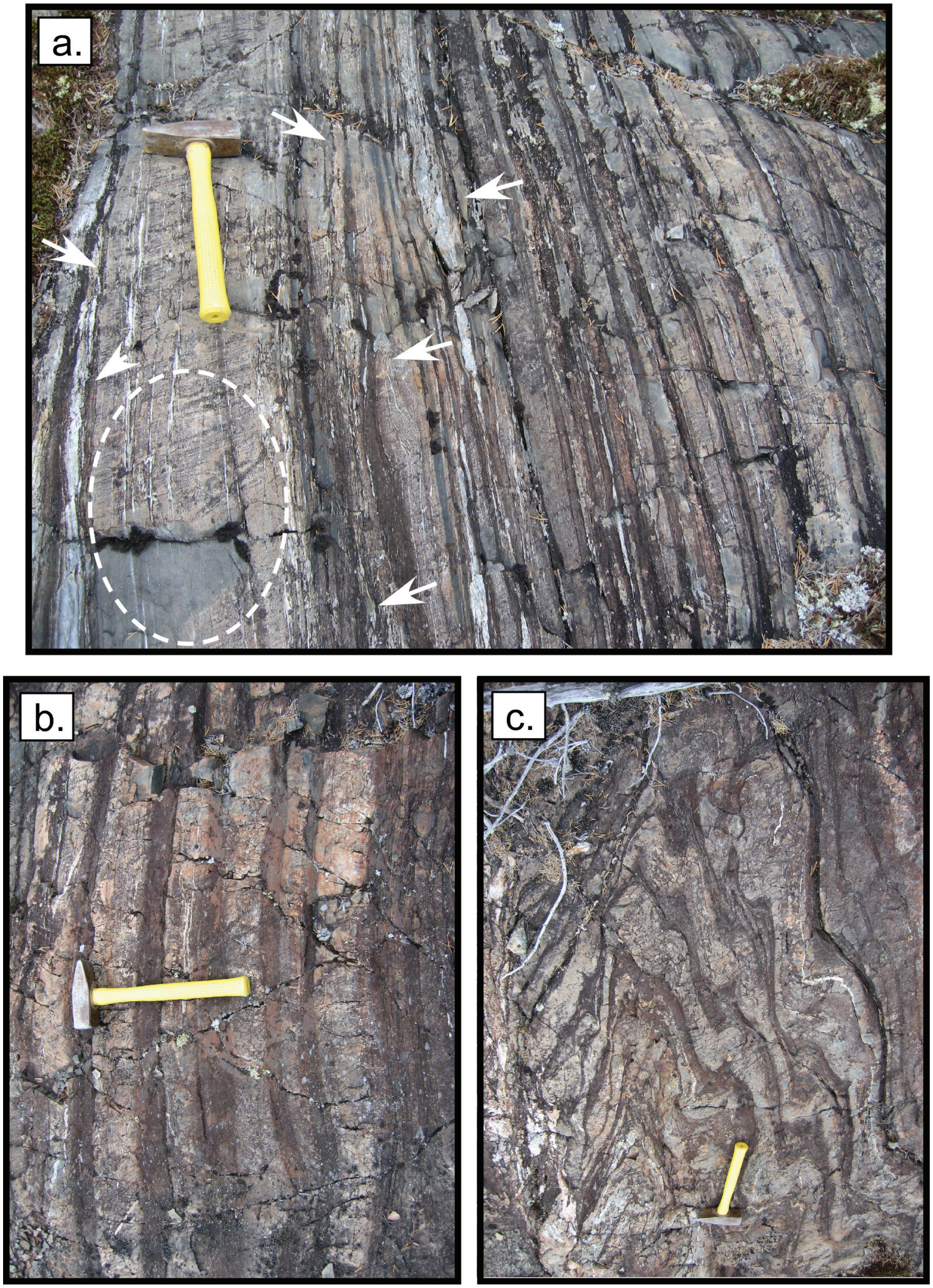


Figure 15. Layered rocks routinely interpreted as 'turbidite strata' in the Waugh Lake Complex, north of Martyn Lake, Saskatchewan: a) note the greenish grey massive layer on the left side (ellipse) and the lack of continuity of marked layers; b) and c) the northerly trending vertical layering and the vertical fold axes suggest strike-slip displacement, similar to layering in the Taltson basement complex. Alternatively, these rocks may record shearing of a relatively homogeneous intrusive rock of acidic-intermediate chemical composition (e.g., trachyte, microgranite) accompanied by quartz segregation (white layers) and phyllosilicate neoformation within linear domains of strain/fluid concentration (darker layers).

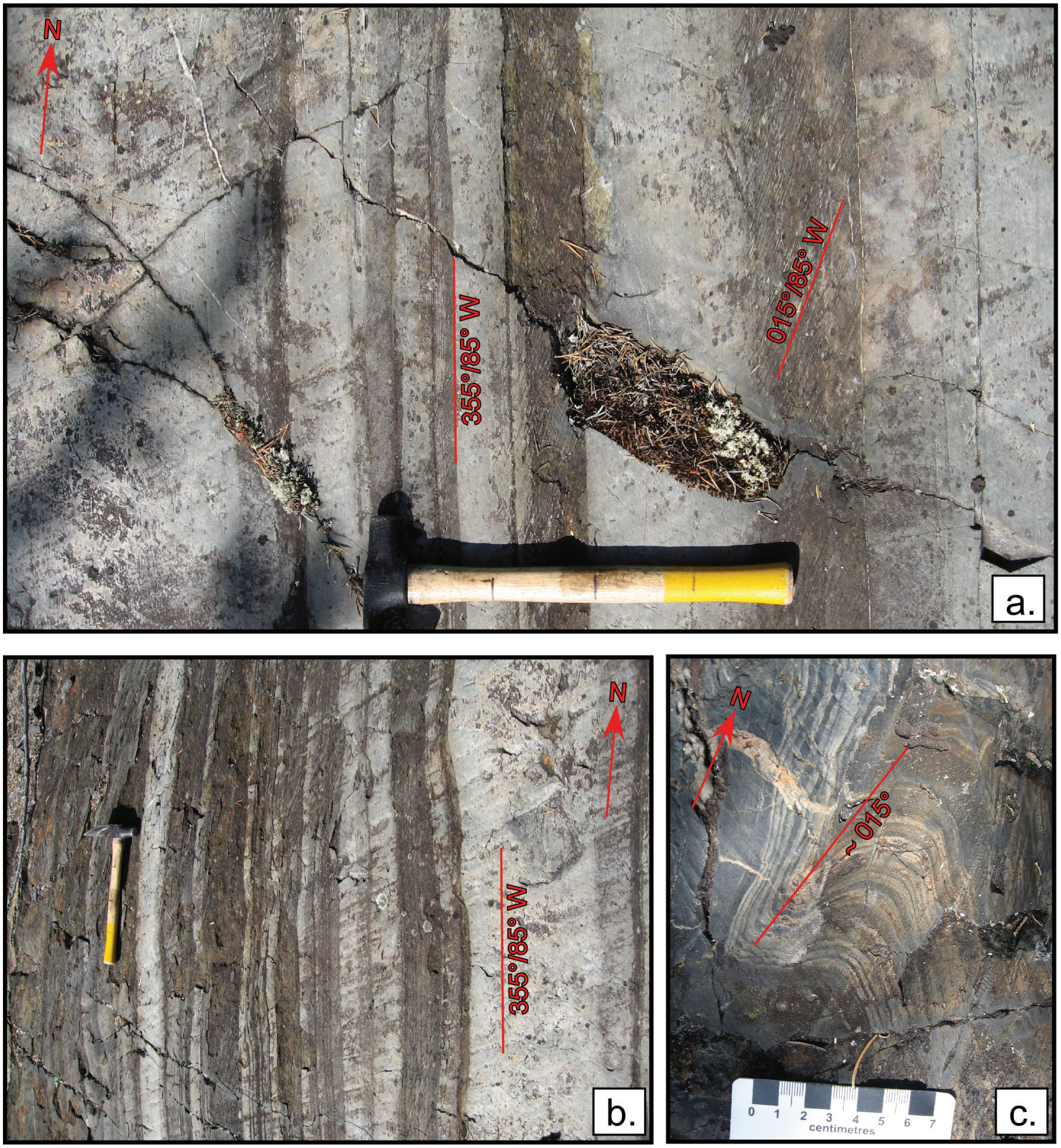


Figure 16. Plan views of possible 'turbidite' outcrop in the Waugh Lake Complex, west of North Waugh Lake (UTM Zone 12, 554800E, 6632700N, NAD83): a) and b) variable-width compositional layers dip steeply (85º) to the west and strike northerly (355º); decimetre-thick layers show lateral continuity at outcrop scale, whereas thinner layers are poorly defined and affected by layer-parallel extension (area below hammer in (b)); cleavage in darker layers strikes 015º; black lines on hammer handle are 10 cm apart; c) minor fold with axial plane striking approximately 015º; glacial striae, visible but not labelled in (a) and (b), trend approximately 060º.

Table 2. Concentrations of SiO_2, Al_2O_3 and Zr in the 'turbidite' layers of the Waugh Lake Complex and typical turbidite strata.

Sample	Source	Description	SiO_2 (%)	Al_2O_3 (%)	Zr (ppm)
311A	this report	siliceous ' turbidite' layer	83.95	8.05	449
311B	this report	siliceous 'turbidite' layer	78.64	10.16	584
6610	this report	argillaceous 'turbidite' layer	54.81	23.87	152
TE1-S	McLennan et al. (2003)	sandy turbidite layer	84.2	2.95	228
TE2-S	McLennan et al. (2003)	sandy turbidite layer	65.5	8.75	424
TE4-S	McLennan et al. (2003)	sandy turbidite layer	81.8	5.98	998
TE8-S	McLennan et al. (2003)	sandy turbidite layer	64.9	12.5	163
TE7-M	McLennan et al. (2003)	muddy turbidite layer	52.7	17.78	n/a
TE8-M	McLennan et al. (2003)	muddy turbidite layer	49.1	15.76	n/a

The intense folding of these layered rocks over hundreds of square metres around the north end of Martyn Lake (Figure 15c) may have acted as a strain-hardening domain leading to the branching off and deflection of the secondary belt of highly strained rocks on the west side of Doze Lake, continuing to Paradise Lake and dying out southward near Johnson Lake (Figure 3).

4.2.2 Contacts

A geological traverse from Sederholm Lake to Paradise Lake and western Waugh Lake provided good exposures across the western contact of the Waugh Lake Complex with the Andrew Lake granite. On the east shore of Sederholm Lake, the Andrew Lake granite is foliated and intruded by pegmatite. Farther to the east, the Andrew Lake granite was intruded by muscovite-bearing pegmatitic leucogranite and displays a penetrative, variably dipping, plane-parallel foliation trending 45°. Still farther east, the Andrew Lake granite, locally with a higher proportion of K-feldspar, becomes more intensely foliated and frequently displays augen gneiss textures. The next interval of approximately 50 m is a subvertical succession of biotite gneiss, locally with augen textures and textures suggestive of crossbedding and graded bedding, and with the general foliation trending along 30°. This interval roughly corresponds to unit 2a in Figure 7, in which Iannelli et al. (1995) correctly described a layered porphyroblastic gneiss with pegmatitic to porphyritic granite dikes and rafts in their 'Lower Member of the Doze Lake Formation'. However, their stratigraphic approach required that these rocks be derived from (Iannelli et al., 1995, map legend) "medium-grained to pebbly subarkose to sublitharenite with minor conglomerate" (i.e., the result of prograde metamorphism of their sedimentary succession interpreted to the east).

Field and microscopic evidence indicates that, in fact, the sense of metamorphic reactions is retrograde, with a relatively abrupt increase in retrogression to the east. The next outcrops towards Paradise Lake expose a slightly foliated mafic intrusion, an amphibolite and a succession of fine-grained biotite-sericite schist, in which more massive domains of biotite microgranite-diorite, with feldspar grains up to 0.5 cm, are sheared and retrogressed to chlorite-sericite schist.

This unit corresponds to unit 2b in Figure 7 and has been described by Godfrey (1961) as a mylonitic series ("sericitic, porphyroclastic phyllonite, feldspar augen, typically sheared, mylonitic to crushed matrix; crush conglomerate and minor conglomerate?"; legend to accompanying map in Godfrey, 1961), whereas Iannelli et al. (1995) identified the same interval as “medium-grained to pebbly subarkose to sublitharenite with minor conglomerate” in the legend of their map. The author argues that the contact is, in fact, a gradational transition through shearing and retrogression from the foliated Andrew Lake biotite granite and associated pegmatite bodies through augen gneiss to protomylonite and phyllonite derived from igneous rocks.

Two geological traverses were completed across the eastern margin of the Waugh Lake Complex, from Alberta into Saskatchewan, one immediately northeast of Martyn Lake and the other between the northern shore of Ney Lake in Alberta and Unger Lake in Saskatchewan (Figure 3 and Appendix 2). Along the Martyn Lake traverse, the contact is not exposed but can be inferred at the slope change between a lowland area with isolated outcrops of fine-grained rocks and spectacular plane-parallel vertical layering routinely interpreted as a turbidite sequence, and a hill slope to the east. The first outcrops on this hill slope consist of biotite-garnet±sillimanite followed by leucocratic granitic protomylonite and augen gneiss, intruded by metre-size pegmatite bodies partly rotated into parallelism with the mylonitic foliation in the granite host.

Along the southern traverse, exposures of granite on high land near Ney Lake are followed eastward by lowland/wetland with isolated ‘islands’ of high ground that expose low-grade sericite-chlorite schist, occasionally crenulated, with folded and locally sheared pegmatite bodies and including sheared granite lenses. The topographic high on the last ~300 m before Unger Lake exposes a medium-grained, equigranular, hornblende granite–granodiorite that is locally foliated and, at one location, sheared to limonite-stained phyllonite and fragmental fault rocks, identical to those mapped as conglomerate by Iannelli et al. (1995) on ‘Grit Hill’, west of Waugh Lake (Figure 7).

Between the two traverses, rocks near the eastern contact of the Waugh Lake Complex were mapped by Watanabe (1965) about 2 km east of the Alberta-Saskatchewan border (Appendix 2). In his view, the main belt of low-grade rocks includes units 2a and 2b of Godfrey (1961), consisting of various schist units described as sedimentary rocks based on their grain size (‘slate’, ‘phyllite’, ‘siltstone’, ‘greywacke’) intercalated with igneous lenses, ‘feldspar augen, granite and pegmatite lenses’ and fault rocks (‘bands of mylonite’, ‘phyllonite’, ‘crush conglomerate’). The next map unit to the east has been described in the legend of his map as “granitic metasediments rich in biotite, pink and white feldspar; highly injected, granitized; abundant pegmatitic and granitic lenses, pods, bands,” followed by a sheared, medium- to fine-grained leucogranite with muscovite. The geological contacts traced by Watanabe (1965) are largely subjective, such that Koster (1971) assigned the western half of the ‘granitic metasediments’ to his “western meta-sedimentary and meta-volcanic complex” (Waugh Lake Complex) and their eastern half to his “red gneisses complex” invaded by pegmatite and migmatite (Appendix 2).

Note that granitization involves metasomatic processes by which sedimentary rocks are transformed into granite and is equivalent to prograde metamorphism. Based on the author’s field observations, the association of rocks described by Watanabe (1965) is a mylonite-protomylonite assemblage derived from a leucogranite body intruded by pegmatite, and most likely represents the passage from a highly strained granitoid-pegmatite domain (part of the Waugh Lake Complex) to gradually less strained granite to the east. The intensity of shearing associated with retrogressive metamorphic reactions (e.g., andesine to albite, amphibole and biotite to chlorite) increase towards the Waugh Lake Complex at the expense of recognizable remnants of the gneiss-granite-pegmatite domain mapped to the east.

Northward, the belt of low-grade rocks crops out north of Martyn Lake and continues in a zone of lowland north of Doze Lake and then departs eastward from the Alberta-Saskatchewan border, then north into the Northwest Territories (Appendix 2). To the south, the Waugh Lake Complex is represented from west to east by 1) a southerly trending belt of foliated rocks between the westerly trending segment of

Waugh Lake and Johnson Lake; 2) a granitoid body consisting of several distinct phases (Godfrey, 1961) and assigned to the Colin Lake pluton (Godfrey, 1986a; McDonough et al., 2000b); and 3) the main belt of low-grade rocks that can be followed on either side of the northerly trending segment of Waugh Lake and continues southeasterly into Saskatchewan, where it has been examined on the geological traverse between Ney and Unger lakes (Figure 3 and Appendix 2).

4.2.3 Age of Sedimentation

4.2.3.1 K-Ar, $^{40}Ar/^{39}Ar$ and U-Pb Ages

The low metamorphic grade (greenschist) of rocks in the Waugh Lake area led Godfrey (1961) to infer that these rocks were younger than the high- to medium-grade Taltson orthogneiss. The first sets of K-Ar isotope dates obtained from the Waugh Lake Complex ranged from 1860 to 1740 Ma, which are within the wider ranges of 1940–1620 Ma and 1930–1790 Ma obtained from adjacent orthogneiss and granitoid plutons, respectively (Appendix 3). However, an age of 1760 Ma was inferred to best approximate the age of deposition of the Waugh Lake Formation (Baadsgaard and Godfrey, 1967, 1972).

A sample of “moderately foliated muscovite-bearing feldspathic wacke with centimetre-scale clasts” (McDonough and McNicoll, 1997, p. 105) was collected for U-Pb zircon dating from the Doze Lake Formation of Iannelli et al. (1995), which is unit 3-2 in Figure 7. The concordant to slightly discordant single-grain analyses have $^{207}Pb/^{206}Pb$ ages that are dominantly in the 2.3–2.0 Ga range, with one concordant analysis at 2738 ±41 Ma (Appendix 7, Figure 39). The youngest detrital zircon analyzed yielded an age of 2006 ±15 Ma (fraction D), so the age of the rock was interpreted to be younger than ca. 2.02 Ga (McDonough and McNicoll, 1997); implicitly, their ‘Waugh Lake Group’ would be younger than the ‘Rutledge River Group’, deposited between 2.13 and 2.08 Ga (Bostock and van Breemen, 1994).

Intriguingly, the cluster of slightly discordant data points and the concordant fraction J (with a $^{207}Pb/^{206}Pb$ age of 2318 ±4 Ma) from the analyzed Waugh Lake 'feldspathic wacke' or 'conglomerate' (Appendix 7, Figure 39) are very close to the inferred emplacement ages for the granitoid protoliths of the Taltson orthogneiss samples: ca. 2295 Ma, ca. 2330 Ma, ca. 2382 Ma and ca. 2393–2381 Ma (Appendix 7, Figures 40–43). The ‘subrounded to rounded’ shapes of zircon described in this rock of the Waugh Lake assemblage cannot be taken as evidence for a detrital origin of the grains, because similar ‘somewhat rounded’ zircon shapes are not uncommon in the Taltson orthogneiss (e.g., samples MSB 93-114 and MSB 94-33; McNicoll et al., 2000). There is a distinct possibility that the ‘feldspathic wacke’ described by McDonough and McNicoll (1997) is, in fact, a ca. 2320 Ma granite gneiss of the Taltson basement complex overprinted by strain under low-grade conditions in the Waugh Lake area. This interpretation is consistent with the identification by the same authors of Taltson basement orthogneiss within the Waugh Lake Complex (unit 4-2 in Figure 7c).

A second element considered by McDonough and McNicoll (1997) to constrain the age of sedimentation is the ca. 1971 Ma emplacement age of an equigranular quartz diorite (Appendix 6, Figure 33), initially described as biotite microgranite by Godfrey (1963), assigned to the Colin Lake pluton and interpreted to intrude the basal strata of the Waugh Lake Group defined by Iannelli et al. (1995). The author’s observations along and across this contact south of western Waugh Lake do not support this interpretation.

First, granitoid emplacement would, in principle, have induced contact metamorphism in the sedimentary rocks. There is no evidence of contact metamorphism around any of the granitoid plutons or plugs. Second, the supposed intrusive contact is, in fact, an obvious gradual transition through strain gradients accompanied by metamorphic reactions from massive or slightly foliated to strongly foliated granitoid, with synkinematic quartz segregations whose internal mylonitic laminae are parallel to the local foliation

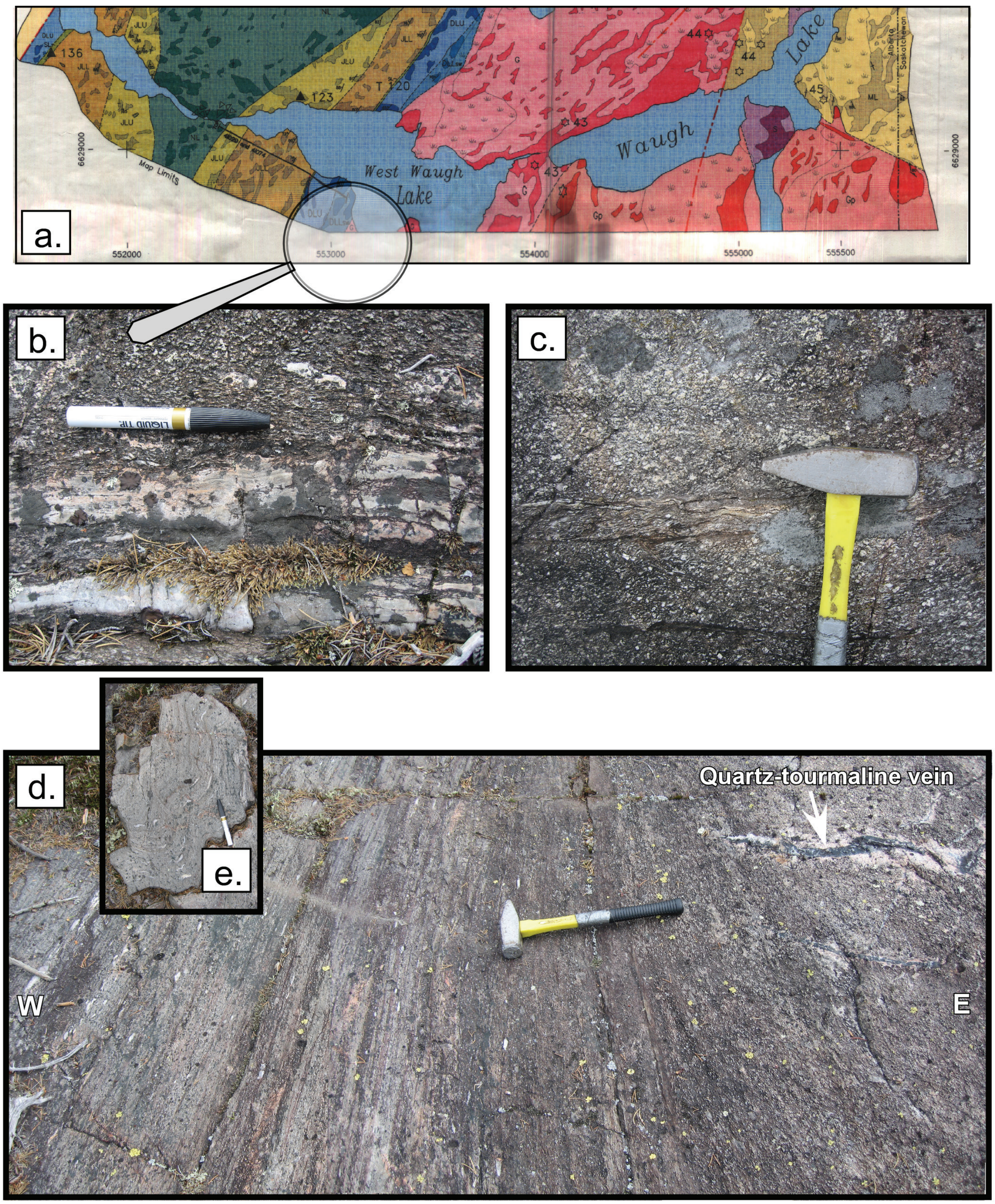


Figure 17. Gradational transition through strain partitioning and strain gradients accompanied by mineral phase transformations from the massive ca. 1971 Ma Colin Lake granitoid to strongly foliated rocks of the Waugh Lake Complex, south of western Waugh Lake: a) examined contact zone; b) mylonitic laminae in segregated quartz vein; c) centimetre-wide shear zone within the pluton; d) zone of gradual transition through textural and metamorphic re-equilibration from penetratively foliated granitoid (E) to a layered quartzofeldspathic rock of the Waugh Lake Complex (W); and e) microfolding of the mylonitic layering and dismembering of quartz segregations.

of the granite (Figure 17b); within 1–2 m, the granite gradually loses its identity into a highly sheared quartzofeldspathic rock, (Figure 17d, c and Figure 12) previously interpreted as a medium-grained to pebbly arkose in the 'Lower Member of the Doze Lake Formation' (Iannelli at al., 1995).

Intense shearing under low-grade metamorphic conditions obviously postdated granite emplacement and resulted in the development of metamorphic-mylonitic layering. At least at this particular location, the foliation/layering is unambiguously tectonic, so the inference of a sedimentation age from the field relationships (intrusion crosscutting beds) is untenable. The author argues that the structural elements, mineral paragenesis and strain distribution within the Waugh Lake assemblage are consistent with tectonothermal evolution of a transcurrent shear zone, at and above the brittle-ductile transition zone. The ca 1.971 Ga Colin Lake diorite from the western Waugh Lake area and the ca. 1.973 Ga Martyn Lake granitoid (see below), which are enveloped by low-grade mylonite belts, represent undigested pods and lozenge of the adjacent TMZ crust.

In an attempt to constrain both the timing of metamorphism and the protoliths of the layered rocks, several samples of the igneous rocks were collected during the 2006 field work for U-Pb dating from within the Waugh Lake Complex and from the granite and gneiss units on the east side of the low-grade belt, in Saskatchewan. Preliminary U-Pb geochronology work has been carried out by both thermal ionization mass spectrometry (TIMS) and laser ablation at the University of Alberta, under the supervision of L. Heaman. Samples from the Waugh Lake Complex include a granodiorite plug at the northeast end of Martyn Lake, assigned to a 'meta-basalt' unit by Koster (1971), and a coarse-grained amphibolite on the south side of the Waugh Lake elbow (Figure 18). In addition, a sample from granitic gneiss on the west side of Harper Lake, representing the high-grade crust east of the Waugh Lake belt, was provided by K. Ashton of the Saskatchewan Geological Survey.

Preliminary results of TIMS U-Pb analysis of three zircon fractions from the small Martyn Lake granodiorite body (DP5-55; UTM Zone 12, 556037E, 6636040N, NAD83) and three fractions from the Waugh Lake elbow amphibolite (DP5-81; UTM Zone 12, 555130E, 6629404N, NAD83) are included in Table 3 and Figure 19. The results of 60 laser-ablation analyses on zircon grains from the Harper Lake orthogneiss are provided in Figure 20 and Appendix 8. The three highly discordant data points obtained from the small Martyn Lake granodiorite body indicate substantial Pb loss but allow the calculation of a regression line with a concordia upper-intercept age of ca. 1973 ±3 Ma, interpreted as an emplacement age. This age is similar, within analytical error, to the 1971 ±4 Ma age of the quartz diorite phase of the Colin Lake pluton west of the main belt of low-grade tectonite. The high discordance of the analyzed fractions from sample DP5-55 (Table 3) indicates significant strain-related Pb loss in the small Martyn Lake granodiorite body surrounded by the Waugh Lake low-grade tectonite.

The three highly discordant data points obtained from the amphibolite sample DP5-81 (Table 3, Figure 19) also indicate significant strain-related Pb loss. The amphibolite from which sample DP5-81 was collected is situated immediately west of the main belt of low-grade tectonite within WLC.

The sample from the west side of Harper Lake (UTM Zone 12, 573894E, 6613105N, NAD83) is a coarse-grained to K-feldspar–porphyritic granite with a high- to medium-grade mylonitic texture from the 'granitic-granodioritic orthogneiss and migmatite unit' (unit Zgo) of the Zemlak Domain at the western margin of Rae Terrane (Ashton, 2009). The results of some 60 laser-ablation analyses on zircon grains are shown in the concordia diagram and the probability plot in Figure 19. The concordia diagram may be somewhat misleading because many of the analyses are relatively discordant, whereas the probability plot appears more relevant because it clearly shows a peak in $^{207}Pb/^{206}Pb$ ages at 2707 Ma, as well as some younger peaks at ca. 2575 and 2350 Ma. The main peak shows that the Archean Rae rocks of the Nolan Domain (van Schmus et al., 1986) extend well into the Zemlak Domain and possibly dominate the

Table 3. Summary of U-Pb data for the Martyn Lake granitoid plug and the amphibolite body south of Waugh Lake.

Sample/ fraction	U (ppm)	Th (ppm)	Pb (ppm)	Th/U	TCPb (pg)	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	Disc %
DP5-55													
1 z	209	74	22.6	0.36	5	4336	0.10461 ±13	0.96570 ±278	0.06695 ±17	641.4 ±0.8	686.2 ±1.4	836.3 ±5.1	24.5
2 z	315	129	97.6	0.41	18	907	0.27365 ±37	4.38204 ±1266	0.11614 ±28	1559.3 ±1.9	1709.0 ±2.5	1897.6 ±4.2	20.1
3 z	946	381	278.3	0.40	73	1631	0.26727 ±29	4.18437 ±513	0.11355 ±7	1526.9 ±1.5	1671.0 ±1.1	1856.9 ±1.1	19.9
DP5-81													
1 z	1070	1066	404.8	1.00	448	398	0.27742 ±33	4.00958 ±910	0.10482 ±21	1578.4 ±1.7	1636.1 ±2.0	1711.2 ±3.7	8.8
2 z	817	885	288.4	1.08	84	1638	0.27710 ±31	4.16038 ±496	0.10889 ±6	1576.7 ±1.6	1666.3 ±1.0	1781.0 ±1.0	12.9
3 z	769	751	308.0	0.98	585	227	0.26956 ±37	3.99376 ±1412	0.10746 ±37	1538.5 ±1.9	1632.9 ±3.1	1756.7 ±6.3	14.0

Note:

z, zircon; 1z, first zircon fraction; col, colourless number in parentheses corresponds to total number of grains analyzed

TCPb, total common lead; All errors in this table reported at 1σ level

DP5-55 – Martyn Lake granitoid (556037 E, 6636040 N, NAD83, Zone 12)

1 z, col subhedral larger irregular grains with mineral inclusions, magnetic separation, 6 grains, total weight 17 μg;

2 z, small subhedral lt. tan 3:1 prisms, magnetic separation, best 6 grains, total weight 3 μg;

3 z, small subhedral 3:1 prisms, magnetic separation, 2nd best 18 grains, total weight 7 μg.

DP5-81 – Amphibolite (555130 E, 6629404 N, NAD83, Zone 12)

1 z, parts of col prisms magnetic separation, 5 grains, total weight 9 μg;

2 z, equant parts of col prisms, magnetic separation, 5 grains, total weight 10 μg;

3 z, col parts of prisms magnetic separation, 2nd best 3 grains, total weight 10 μμg.

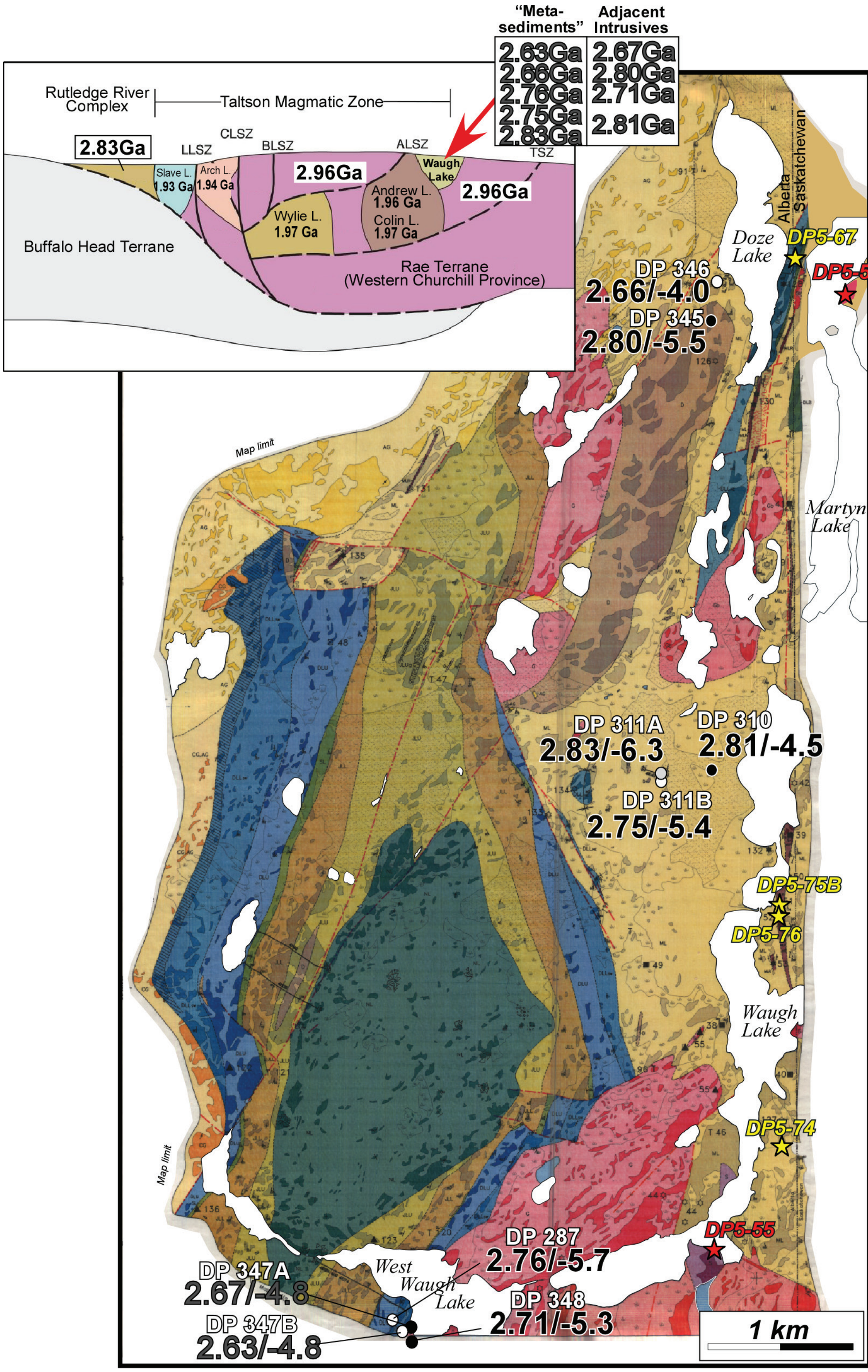


Figure 18. New isotope data from the Waugh Lake Complex. $T_{DM(1950)}$ model ages and ε_{Nd} values for selected sheared and/or layered rocks (light coloured dots) from the Waugh Lake Complex and immediately adjacent massive or weakly deformed igneous rocks (black dots). White labels are site locations; black labels are T_{DM}/ε_{Nd} values. Geological map from Iannelli et al. (1995); conceptual cross-section after McDonough et al. (2000b).

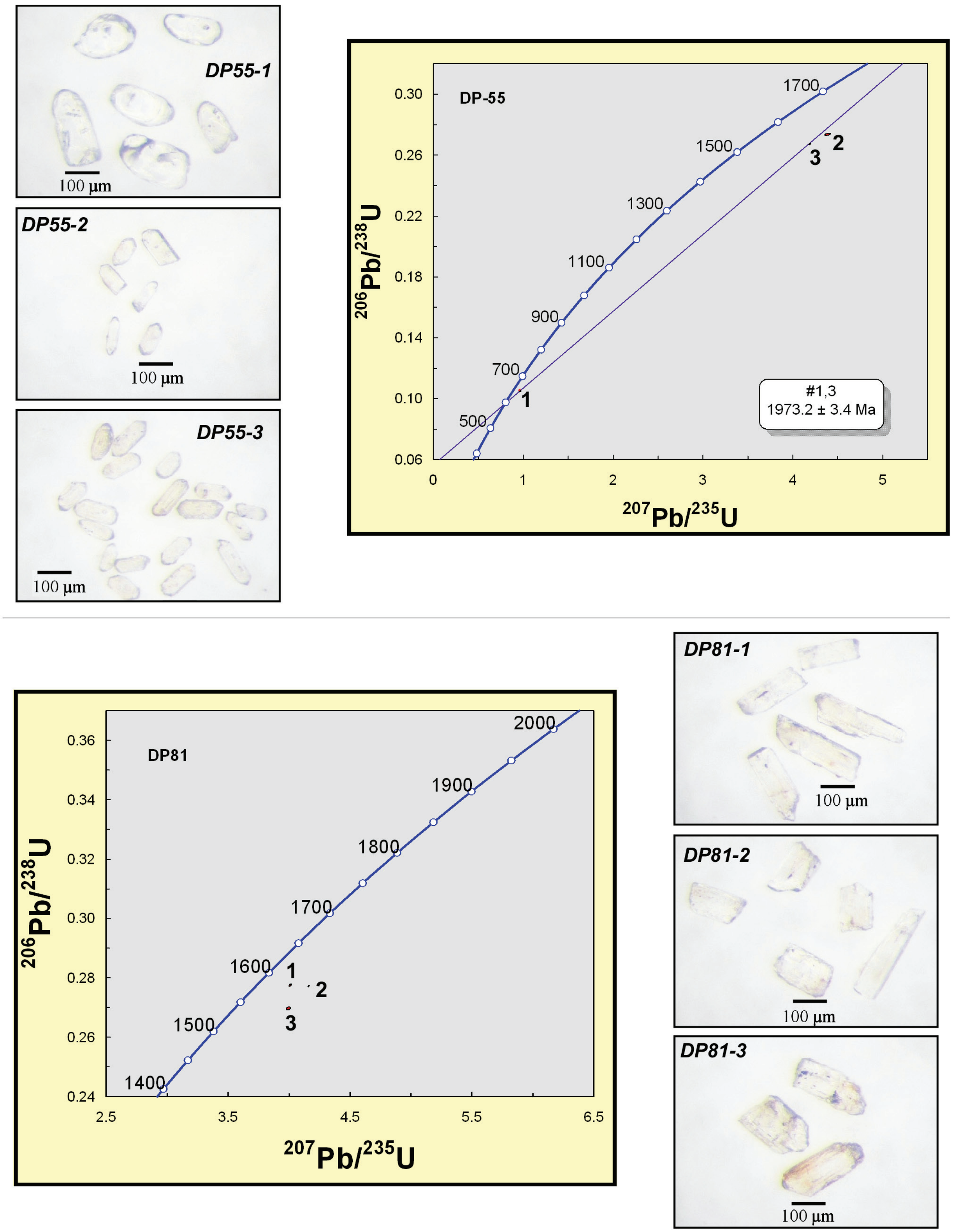


Figure 19. Concordia diagrams for the Martyn Lake granitoid plug (DP5-55; UTM Zone 12, 556037E, 6636040N, NAD83) and the amphibolite body south of Waugh Lake elbow (DP5-81; UTM Zone 12, 555130E, 6629404N, NAD83), within Waugh Lake Complex.

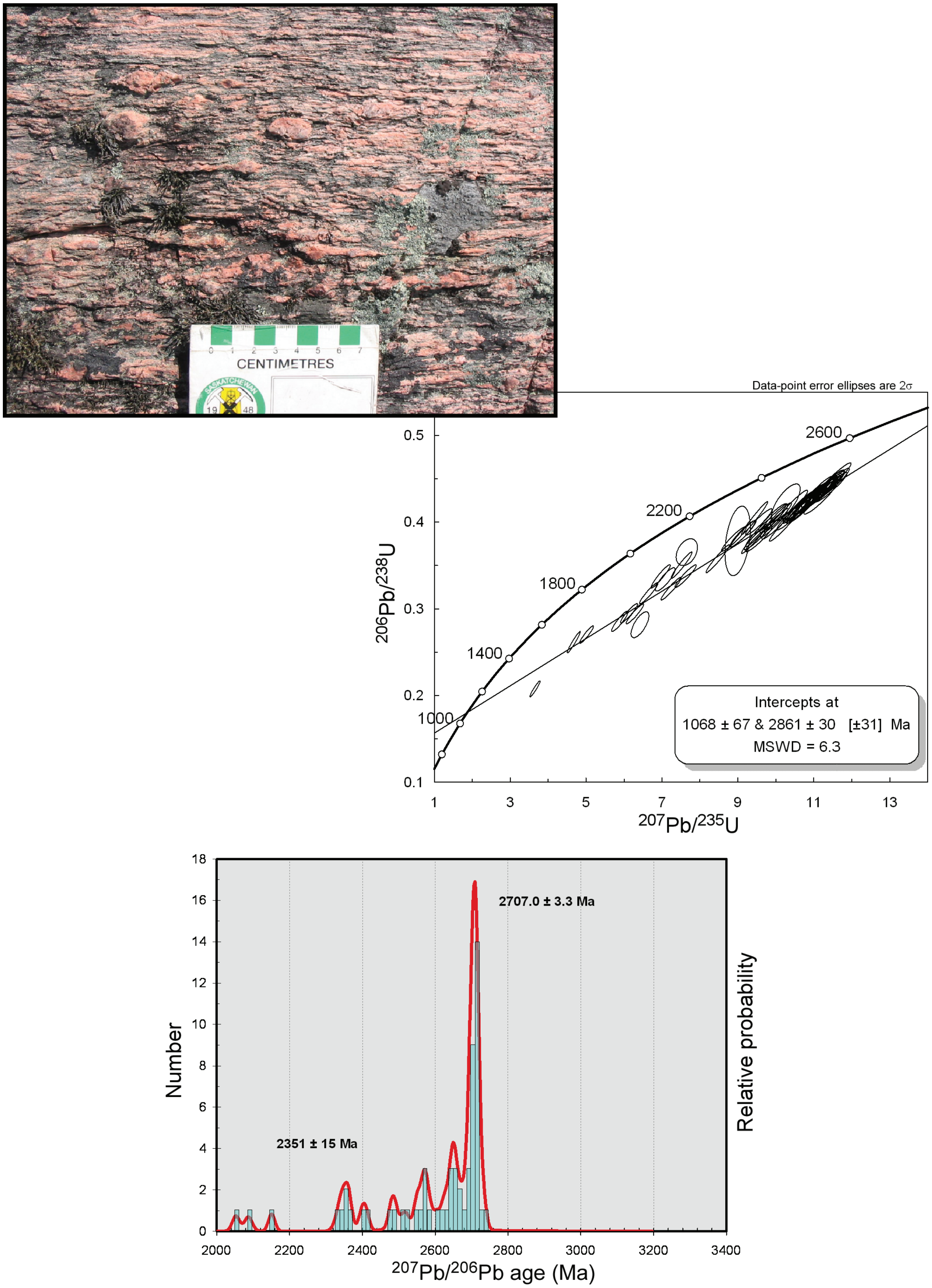


Figure 20. Concordia diagram and probability plot for the mylonitic porphyritic granite near Harper Lake, Saskatchewan (UTM Zone 12, 573894E, 6613105N, NAD83) assigned by Koster (1971) to the 'Red gneiss complex' and recently included in 'Zemlak Domain' (Ashton, 2009).

high- to medium-grade grade crust on the east side of the Waugh Lake low-grade tectonite belt. The ca. 2575 Ma peak correlates with an age of ca. 2562 ±6 Ma reported from the Taltson basement complex at Lapworth Point on the north shore of Lake Athabasca in Alberta (Appendix 5, Figure 31). The ca. 2350 Ma peak correlates with the inferred protolith ages of orthogneiss samples from the Taltson basement complex to the west in Alberta (McNicoll et al., 2000), and to the east with U-Pb zircon crystallization ages for a number of granitoid plutons in the Beaverlodge area of Saskatchewan (Hartlaub et al., 2007; Appendix 4; Table 11).

A pre-Taltson Paleoproterozoic tectonomagmatic event between 2.4 and 2.3 Ga was initially suggested based on K-Ar ages (Stockwell, 1982) and redefined as the Arrowsmith Orogeny (Berman et al., 2005). Hartlaub et al. (2007) interpreted the suite of Paleoproterozoic granite bodies in the Beaverlodge area to represent relicts of a ca. 2.3 Ga magmatic belt related to the Arrowsmith Orogeny. The time-correlative granitoid protoliths of some Taltson orthogneiss units point to a significant pre-Taltson tectonothermal event at the western margin of the Rae Terrane (Appendix 4, Table 11). In contrast, based on the widespread 2.5–2.0 Ga magmatic, metamorphic, detrital, and xenocrystic U-Pb zircon ages from Laurentia, Aspler and Chiarenzelli (1998) argued that the earliest Paleoproterozoic ages do not define a linear orogenic belt and are not restricted to a relatively short time interval.

The analyzed Archean porphyritic granite appears to have acquired its mylonitic texture at ca. 2350 Ma during the Arrowsmith tectonothermal event. Remarkably, none of the analyzed zircon grains yielded Taltson ages (2.0–1.9 Ga), although Taltson-correlative igneous rocks and tectonic activity are known farther east in the Beaverlodge Lake area of Saskatchewan (Feil Lake and Uranium City granitoid bodies; Hartlaub et al., 2005).

4.2.3.2 Sm-Nd Isotope and Geochemical Data

The pre-Taltson sedimentary origin of the 'Waugh Lake Group', which was inferred to have been deposited between 2.02 and 1.97 Ga (Iannelli et al., 1995; McDonough and McNicoll, 1997; McDonough et al., 2000c), implies erosion of the Archean crust (e.g. ca. 2.7 Ga Nolan Domain) and remobilized Archean crust (parts of the Zemlak Domain and the ca. 2.3 Arrowsmith tectonothermal event) to the east, and Taltson basement complex and Rutledge River Complex to the west. The geochemical and Nd isotopic signatures of the Arrowsmith granitoid bodies in northern Saskatchewan are suggestive of crustal derivation (Hartlaub et al., 2005). Therefore, ε_{Nd} values and T_{DM} ages of the 'Waugh Lake Group' sedimentary rocks should be similar to their essentially Archean source and distinct from the ca. 1.96 and 1.97 Ga Andrew Lake and Colin Lake plutons, respectively. In previous tectonic models, both the Andrew Lake and Colin Lake granitoid plutons have been interpreted as early, subduction-related plutons (e.g., McDonough et al., 2000b, c; Ross, 2002; Ross and Eaton, 2002) that intruded the 'Waugh Lake Group' (e.g., McDonough and McNicoll, 1997). Because subduction-related plutons include a certain amount of ca. 1.97 Ga juvenile material, Sm-Nd concentrations should yield less negative ε_{Nd} values and younger T_{DM} ages than the Archean-sourced sedimentary rocks of the 'Waugh Lake Group'.

In an attempt to trace the origin of the controversial layered rocks, a set of nine samples has been collected for Sm-Nd analysis from several rock types previously interpreted as 'metasediments' of the 'Waugh Lake Group' and from adjacent igneous rocks. A summary of existing ε_{Nd} values and T_{DM} ages from both pre-Taltson crust and Taltson plutons is included in Table 4 for comparison with the new set of data in Table 5. Additional geochemical data for the recently collected samples are included in Tables 6, 7 and 8. Samarium-neodymium sample locations are marked on the stratigraphic map of Iannelli et al. (1995), shown in Figure 18. The analytical method is described in Creaser et al. (1997).

The average T_{DM} model age and ε_{Nd} value for orthogneiss samples of the Taltson basement complex (recycled Archean crust), calculated at the time of emplacement of the Colin Lake pluton (ca. 1.97

Table 4. Samarium-neodymium concentrations, ratios and averages in rocks of the Taltson magmatic zone, Alberta.

Lithology	Range $\varepsilon_{Nd}T$	Range T_{DM} (Ga)	T (Ga)	Average $\varepsilon_{Nd}T$ (no. of samples)	Average T_{DM} age (no. of samples)	Publication
TBC	–1.6 to –14.3	2.59 to 3.73	2.20	–5.07 (27)	2.96 (22)	McNicoll et al. (2000)
TBC excluding amphibolite samples			2.20	–6.67 (20)	2.99 (20)	
TBC	–3.1 to –17.9	2.50 to 3.60	1.97	–7.96 (9)	2.86 (9)	De et al. (2000)
Amphibolite	–4.3 to +3.9	2.59 to 2.78	2.20	–0.80 (7)	2.69 (2)	McNicoll et al. (2000)
Banded mafic orthogneiss	–3.3 to –9.9	2.68 to 3.34	2.20	–5.57(4)	2.92 (4)	McNicoll et al. (2000)
Banded tonalite-granite-granodiorite	–1.6 to –12.5	2.72 to 3.36	2.20	–6.66 (9)	2.99 (9)	McNicoll et al. (2000)
Migmatitic gneiss	–4.2 to –14.3	2.90 to 3.73	2.20	–8.95 (4)	3.21 (4)	McNicoll et al. (2000)
Metasedimentary rocks	–5.1 to –5.5	2.8 to 2.9	1.97	–5.33 (3)	2.83 (3)	De et al. (2000)
Subduction-related plutons						
Colin Lake	–3.9 to –6.0	2.60 to 2.80	1.97	–5.08 (4)	2.725 (4)	De et al. (2000)
Wylie Lake	–3.4 to –6.0	2.60 to 2.80	1.97	–4.83 (4)	2.700 (4)	De et al. (2000)
Collision-related plutons						
Arch Lake	–5.2 to –9.8	2.60 to 3.00	1.97	–7.17 (14)	2.77 (14)	De et al. (2000)
Slave Lake	–3.8 to –8.9	2.60 to 3.50	1.97	–5.72 (14)	2.87 (14)	De et al. (2000)

Table 5. Samarium-neodymium concentrations and ratios in rocks of the Waugh Lake Complex.

Sample	Field station	Rock type	Sm	Nd	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	Uncertainty	$^{145}Nd/^{144}Nd$	T_{DM}	T	ε_{Nd}^{T}
			(ppm)	(ppm)			±2σ		(Ga)	(Ma)	
6704	DP 347A	Granodiorite	4.43	27.27	0.0983	0.511130	0.000007	0.348352	2.67	1950	–4.8
6705	DP 347B	Green-grey qz-chlorite schist	6.18	40.80	0.0916	0.511046	0.000007	0.348372	2.63	1950	–4.8
6706	DP 345	Gabbro-diorite	5.19	28.49	0.1102	0.511250	0.000005	0.348390	2.80	1950	–5.5
6707	DP 346	Black schist	2.56	14.70	0.1055	0.511263	0.000007	0.348391	2.66	1950	–4.0
6708	DP 348	K-feldspar granitoid	3.96	24.07	0.0996	0.511121	0.000006	0.348397	2.71	1950	–5.3
6709	DP 287	K-feldspar 'metesedim' rock	1.79	10.55	0.1027	0.511142	0.000004	0.348360	2.76	1950	–5.7
310	DP 310	Gabbro-diorite	5.35	26.88	0.1204	0.511430	0.000006	0.348404	2.81	1950	–4.5
311A	DP 311	Dark grey siliceous turbidite	2.22	12.61	0.1064	0.511157	0.000006	0.348397	2.83	1950	–6.3
311B	DP 312	Light grey siliceous turbidite	2.56	14.95	0.1036	0.511166	0.000009	0.348380	2.75	1950	–5.4

Note: Nd correction used is Nu α = 0.5122558, except for sample 6709, which uses Nu α = 0.5122431
See Table 6 for locations

Ga), are 2.86 Ga and –7.96, respectively, which are slightly older than the average 2.72 Ga T_{DM} and significantly more negative than the ε_{Nd} value of –5.08 yielded by this pluton (Table 4). These values suggest that the Colin Lake pluton formed through the anatexis of the variably recycled Archean orthogneiss crust, with a limited contribution of ca. 1.97 Ga juvenile material. Sediments older than Colin Lake pluton might be expected to yield T_{DM} model ages and ε_{Nd} values showing a stronger Archean component, similar to their pre-Taltson source rocks (TBC and possibly RRC). Instead, some of the low-grade tectonite samples from the Waugh Lake Complex show T_{DM} model ages (2.63 Ga, 2.66 Ga and 2.76 Ga) and εNd values (–4.8, –4.0 and –5.7) much closer to the Colin Lake and other Taltson-age plutons (Tables 4 and 5).

Moreover, each of these low-grade tectonite samples shows Sm-Nd concentrations and ratios almost identical to the immediately adjacent plutonic rocks. Thus, the green-grey quartz-chlorite schist sample collected at site DP 347B, within 10 m of the slightly foliated Colin Lake quartz diorite at site DP 347A on the south shore of western Waugh Lake, yielded a T_{DM} model age of 2.63 Ga, slightly younger than the age of 2.67 Ga for the adjacent Taltson-age intrusive rock, and an identical ε_{Nd} value of –4.8 (Table 5). A pink, layered quartzofeldspathic rock has been collected from site DP 287, about 200 m from the same massive igneous body, for comparison with a slightly foliated K-feldspar–bearing phase from site DP 348, approximately 50 m within the igneous body. The 2.76 Ga T_{DM} model age and the –5.7 ε_{Nd} value of the foliated rock are very similar to the 2.71 Ga T_{DM} age and –5.3 ε_{Nd} value yielded by the K-feldspar granitoid (Table 5). Both layered rocks, assigned by Iannelli et al. (1995) to the Doze Lake Formation of the Waugh Lake Group, appear to be derived from the adjacent Taltson-age plutonic rock. Their T_{DM} model ages of 2.76 and 2.63 Ga are considerably younger than the 3.21–2.83 Ga average T_{DM} ages calculated for the Taltson basement complex, and the –5.7 and –4.8 ε_{Nd} values are within the same range as, and slightly less negative than, the approximately –5 and –8 average ε_{Nd} values calculated for the Taltson basement complex (Tables 4 and 5).

A gabbro-diorite body on the west side of Doze Lake was sampled at site DP 345A for comparison with the surrounding dark grey to black, fine-grained schist unit sampled at site DP 346. The schist shows concentrations of SiO_2, Fe_2O_3, TiO_2 and V very similar to those of the nearby gabbro-diorite, with higher Al_2O_3, K_2O and B, and substantially lower MnO, MgO, CaO and Na_2O (Tables 6 and 7). Once again, the 2.66 Ga T_{DM} model age of the schist is younger than the 2.80 Ga T_{DM} model age of the adjacent gabbro-diorite, and the –4 ε_{Nd} value is less negative than the –5.5 value yielded by the igneous rock (Table 5).

Samples from the well-layered, fine-grained quartzitic rocks interpreted as turbidite beds were taken at sites DP 311 and DP 312 for comparison with the closest igneous rock, a gabbro-diorite from site DP 310. The dark-coloured quartzitic rock yielded a T_{DM} model age of 2.83 Ga, similar to the 2.81 Ga of the igneous rock, whereas the lighter coloured quartzitic rock yielded a younger T_{DM} model age of 2.75 Ga. The ε_{Nd} values from the quartzitic layers (–6.3 and –5.4) are more negative than the –4.5 ε_{Nd} value of the gabbro-diorite.

The general similarity of ε_{Nd} and T_{DM} model ages between the layered rocks and nearby Taltson plutons and, in particular, the T_{DM} model ages of the sheared rocks similar to and slightly younger than the adjacent intrusive rocks (Figure 18) are in conflict with the previous interpretation of the sheared rocks as pre-Taltson metasedimentary rocks (i.e., with no contribution of ca. 1.99–1.93 Ga Taltson-age juvenile material).

4.2.4 Age of Metamorphism

Initial isotope geochronology data reported in the 1960s and 1970s from the Taltson magmatic zone in the northeastern part of the Alberta shield included K-Ar, Rb-Sr and U-Pb dates (Godfrey and Baadsgaard,

Table 6. Major-element concentrations (in per cent) in selected rocks of the Waugh Lake Complex; analysis by fusion inductively coupled plasma–mass spectrometry (FUS-ICP).

Sample Number	Site	Rock Type	East	North	SiO_2	Al_2O_3	$Fe_2O_3^T$	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	LOI	Total
					0.01	*0.01*	*0.01*	*0.001*	*0.01*	*0.01*	*0.01*	*0.01*	*0.001*	*0.01*	*0.01*	*0.01*
6704	DP347A	grey-green granite	553025	6629060	67.97	15.57	3.75	0.056	1.47	2.70	3.40	3.65	0.517	0.15	1.031	100.30
6705	DP347B	grey-green qz-chl schist	553014	6629074	81.20	10.56	2.28	0.013	0.40	0.14	0.10	3.87	0.217	0.05	1.588	100.40
6706	DP345	gabbro-diorite	555086	6635825	58.16	15.59	7.27	0.120	5.57	5.59	2.46	3.01	0.754	0.18	1.792	100.50
6707	DP346	black schist	555266	6636113	61.80	19.55	5.29	0.026	1.64	0.22	0.59	6.82	0.618	0.08	3.885	100.50
6708	DP348	K-fsp granite	552963	6628896	63.33	17.41	4.49	0.059	2.14	1.65	3.69	4.62	0.567	0.17	1.547	99.66
6709	DP287	K-fsp sediment	552933	6629058	80.57	10.58	1.46	0.011	0.49	0.19	0.52	5.10	0.161	0.06	1.109	100.30
6610	na	argillite/siltstone	554799	6632686	54.81	23.87	5.64	0.021	1.47	0.24	0.78	6.84	0.852	0.11	4.710	99.34
6611	na	diorite with dissemin. pyrrhotite	555026	6630910	56.86	12.34	6.91	0.166	6.89	4.14	1.61	6.38	0.816	0.70	2.490	99.29
310	DP310	diorite	555147	6632685	55.46	14.44	9.04	0.137	6.74	5.02	1.52	3.75	0.875	0.18	2.625	99.80
311A	DP311A	dark colour turbidite/trachite (?)	554840	6632717	83.95	8.05	2.78	0.016	0.64	0.64	1.99	1.33	0.376	0.07	0.829	100.70
311B	DP311B	light colour turbidite/aplite (?)	554840	6632717	78.64	10.16	4.39	0.033	0.93	0.95	2.81	1.05	0.531	0.09	1.066	100.70
6710	DP344	Unger Lake granite	557513	6626722	69.23	15.34	3.21	0.046	1.02	2.96	3.63	3.37	0.429	0.15	0.553	99.95
6711	PD342	Colin Lake megacrystic granite	546481	6660179	70.36	14.41	2.96	0.056	1.59	1.39	2.83	3.90	0.334	0.12	1.951	99.90
6712	DP336	sheared aplite	556834	6626275	57.57	22.69	5.51	0.041	2.80	0.20	0.45	6.07	0.796	0.12	4.186	100.40

Note:
GPS locations are given in UTM co-ordinate system, NAD83, Zone 12.
FUS-ICP - Fusion Inductively Coupled Plasma Optical Emission Spectrometry.

Table 7. Trace-element concentrations (in ppm) in selected rocks of the Waugh Lake Complex. Values in ppm except for S (%), Au (ppb) and Ir (ppb).

Analyte Symbol	Au	Ag	Cu	Pb	Zn	Mo	V	Ni	Cr	Co	As	Ba	S	Cd	Sc	Be	Zr	Rb	Sr	Y	Sn	Sb
Detection Limit	*2*	*0.3*	*1*	*5*	*1*	*2*	*5*	*1*	*5*	*1*	*0.5*	*3*	*0.001*	*0.5*	*0.1*	*1*	*4*	*2*	*2*	*2*	*1*	*0.5*
Analysis Method	INAA	TD-ICP	TD-ICP	TD-ICP	TD-ICP	FUS-MS	FUS-ICP	TD-ICP	INAA	FUS-MS	INAA	FUS-ICP	TD-ICP	TD-ICP	INAA	FUS-ICP	FUS-ICP	FUS-MS	FUS-ICP	FUS-ICP	FUS-MS	FUS-MS
6704	< 2	0.8	18	45	76	< 2	59	8	23	6	2.6	1000	0.092	< 0.5	6.0	11	162	187	319	12	28	0.8
6705	< 2	0.4	4	10	35	< 2	38	9	41	4	1.9	468	0.006	< 0.5	4.0	2	79	205	28	12	3	< 0.2
6706	< 2	0.5	14	26	119	< 2	159	41	191	21	3.9	1056	0.012	< 0.5	19.0	2	142	112	375	17	2	< 0.2
6707	< 2	0.6	22	32	52	8	188	11	83	3	49.0	1202	0.049	< 0.5	15.7	3	165	189	45	34	5	0.9
6708	< 2	1.2	6	31	81	< 2	72	10	24	7	2.8	1292	0.020	< 0.5	7.7	6	188	194	261	12	11	< 0.2
6709	< 2	0.7	36	51	34	6	28	4	36	1	2.5	805	0.009	< 0.5	2.7	2	84	152	67	7	3	< 0.2
6610	-2	0.8	15	25	49	4	153	12	109	3	20.8	1604	0.020	0.5	21.8	4	152	197	59	28	8	3.7
6611	-2	1.8	24	27	144	< 2	137	177	399	19	8.8	5439	1.540	< 0.5	15.9	4	278	188	649	22	4	< 0.5
310	< 2	0.4	25	11	68	3	215	37	226	23	19.7	1221	0.067	< 0.5	25.1	2	139	144	255	25	2	0.5
311A	3	2.3	7	28	27	3	40	7	27	3	5.8	295	0.012	< 0.5	5.6	2	449	56	99	14	3	< 0.2
311B	< 2	3.2	11	38	45	< 2	54	7	52	3	5.9	199	0.031	< 0.5	7.9	3	584	50	147	12	2	< 0.2
6710	< 2	0.8	7	41	64	< 2	53	4	11	6	3.7	1372	0.013	< 0.5	4.6	2	165	125	463	9	2	< 0.2
6711	< 2	0.9	33	24	56	< 2	41	10	50	6	2.9	567	0.030	< 0.5	4.9	4	141	223	179	11	5	< 0.2
6712	68	1.0	12	126	166	< 2	131	13	73	5	3.7	1329	0.019	< 0.5	20.7	3	143	200	48	17	9	< 0.2

Analyte Symbol	Nb	Ta	W	Tl	Bi	Br	Ga	Ge	In	Ir	Se	Hf	Cs	Th	U	B
Detection Limit	*1*	*0.1*	*1*	*0.1*	*0.4*	*0.5*	*1*	*1*	*0.2*	*5*	*3*	*0.2*	*0.5*	*0.1*	*0.1*	*2*
Analysis Method	FUS-MS	FUS-MS	FUS-MS	FUS-MS	FUS-MS	INAA	FUS-MS	FUS-MS	FUS-MS	INAA	INAA	FUS-MS	FUS-MS	FUS-MS	FUS-MS	PGNAA
6704	9	1.1	4	1.3	< 0.4	< 0.5	22	2	< 0.2	< 5	< 3	4.5	53.8	25	8.4	56
6705	8	2.2	14	1.1	< 0.4	< 0.5	13	2	< 0.2	< 5	< 3	2.7	11.4	26.9	1.6	105
6706	7	0.6	< 1	0.7	< 0.4	< 0.5	23	1	< 0.2	< 5	< 3	4.1	6.9	11.9	2.9	48
6707	12	1.4	9	1.0	< 0.4	< 0.5	24	2	< 0.2	< 5	< 3	4.2	9.9	22.1	5.5	666
6708	11	1.7	6	1.2	< 0.4	< 0.5	23	2	< 0.2	< 5	< 3	5.3	19.9	22.8	11.5	23
6709	6	1.1	8	0.8	2.1	< 0.5	13	1	< 0.2	< 5	< 3	2.5	9.5	17	3.1	65
6610	19	1.7	6	0.7	0.6	1445	33	2	< 0.2	-5	-3	4.8	4.7	23.2	5.6	34
6611	13	0.8	13	1.4	4.0	3995	18	2	< 0.2	-5	-3	7.4	6.4	27.5	5.9	9
310	8	0.7	1	1.0	0.7	< 0.5	19	2	< 0.2	< 5	< 3	3.9	5.2	11.4	2.9	82
311A	10	1.1	3	0.3	< 0.4	< 0.5	11	1	< 0.2	< 5	< 3	11.8	3.1	19	3.7	8
311B	13	1.5	1	0.3	< 0.4	< 0.5	12	1	< 0.2	< 5	< 3	16.1	2.8	24.2	4.2	< 2
6710	8	0.8	< 1	0.9	< 0.4	< 0.5	23	1	< 0.2	< 5	< 3	4.6	4.3	17.7	4.1	na
6711	11	1	< 1	1.6	< 0.4	< 0.5	20	1	< 0.2	< 5	< 3	4.3	3.5	25.3	6.3	na
6712	20	2.1	8	0.8	< 0.4	< 0.5	34	2	< 0.2	< 5	< 3	4.3	5.1	27.7	5.2	na

Note:
Table 7 includes only concentrations obtained through the analytical method with highest precision for a particular element.
FUS-ICP - Fusion Inductively Coupled Plasma Optical Emission Spectrometry.
FUS-MS - Fusion Inductively Coupled Plasma Mass Spectrometry.
TD-ICP - Total Digestion Inductively Coupled Plasma Optical Emission Spectrometry.
INAA - Instrumental Neutron Activation Analysis.
PGNAA - Prompt Gamma Neutron Activation Analysis.
Analyses for some trace elements are included in different analytical packages and have been determined twice. Duplicate analyses obtained through analytical methods with lower precision are in Appendix 9.

Table 8. Concentrations (in ppm) of lanthanide group elements in selected rocks of the Waugh Lake Complex; analysis by fusion mass spectrometry (FUS-MS).

Analyte Symbol	**La**	**Ce**	**Pr**	**Nd**	**Sm**	**Eu**	**Gd**	**Tb**	**Dy**	**Ho**	**Er**	**Tm**	**Yb**	**Lu**
Detection Limit	*0.1*	*0.1*	*0.05*	*0.1*	*0.1*	*0.05*	*0.1*	*0.1*	*0.1*	*0.1*	*0.1*	*0.05*	*0.1*	*0.04*
6704	44.6	85.7	8.56	29.7	4.70	1.16	3.5	0.5	2.4	0.5	1.3	0.20	1.3	0.19
6705	54.2	113	11.4	38.8	6.00	1.08	3.9	0.5	2.6	0.5	1.4	0.22	1.4	0.21
6706	35.0	71.7	7.65	28.9	5.20	1.40	4.4	0.7	3.6	0.6	1.9	0.28	1.8	0.26
6707	22.3	47.5	4.77	14.4	2.50	0.56	2.4	0.6	4.5	1.1	3.9	0.66	4.6	0.70
6708	34.4	68.2	6.76	22.7	3.80	1.37	3.4	0.5	2.5	0.5	1.4	0.23	1.5	0.23
6709	12.4	40.3	3.02	9.9	1.70	0.43	1.5	0.2	1.4	0.3	1.0	0.16	1.1	0.17
6610	25.6	54.4	6.04	21.0	3.9	0.62	3.2	0.7	4.7	1.0	3.3	0.57	3.8	0.58
6611	62.3	121.0	14.50	56.2	9.9	2.39	7.2	1.0	4.7	0.8	2.3	0.33	2.1	0.31
310	30.2	64.5	7.04	27.7	5.50	1.38	4.8	0.8	4.9	1.0	2.9	0.44	2.8	0.42
311A	15.6	35.1	3.68	13.1	2.30	0.46	1.8	0.4	2.3	0.5	1.9	0.34	2.4	0.39
311B	21.4	45.5	4.79	17.4	2.90	0.56	2.1	0.3	2.1	0.5	1.8	0.33	2.4	0.41
6710	43.6	84.3	8.4	28.9	4.80	1.25	3.2	0.4	1.9	0.3	1.0	0.13	0.8	0.12
6711	37.8	79.1	8.37	30.1	5.00	0.90	3.8	0.5	2.4	0.4	1.2	0.16	1.0	0.15
6712	11.8	27.5	2.84	9.4	1.90	0.38	1.9	0.4	3.2	0.7	2.5	0.46	3.2	0.49

Note:
FUS-MS - Fusion Inductively Coupled Plasma Mass Spectrometry.

1962; Baadsgaard and Godfrey, 1967, 1972; Appendix 3). The ‘Waugh Lake Group’ benefited only from K-Ar dates, which included a hornblende age of 1840 Ma, biotite ages of 1820–1740 Ma and muscovite ages of 1830–1770 Ma. These dates largely overlap the K-Ar dates reported from the Taltson basement complex and Taltson plutons (Appendix 3). Based on the isotope data available at the time and on field relationships, the low-grade metamorphism of the ‘Waugh Lake Group’ was interpreted to be part of a second metamorphic cycle that affected the polymetamorphic rocks of the Alberta shield (e.g., Godfrey and Langenberg, 1978; Langenberg and Nielsen, 1982). The first granulite-grade metamorphic cycle was inferred to be Archean, with the second cycle starting under granulite-grade conditions and passing through amphibolite- and finally greenschist-grade metamorphic conditions assigned to the late Aphebian (1900–1790 Ma). In this interpretation, the ‘Waugh Lake Group’ should have been metamorphosed at ca. 1790 Ma, with a minimum metamorphic age constrained by the beginning of deposition of the unmetamorphosed Athabasca Group (ca. 1750 Ma; Rainbird et al., 2007). However, subsequent work, including extensive isotope dating, does not support the interpretation of a regional metamorphic event of this age within the western Rae crust. The last part of this interval (ca. 1850 - 1800 Ma) corresponds to the Trans-Hudson tectonothermal event on the opposite side of the Churchill Province, across central Saskatchewan and beyond.

Based on selective remapping, aided by great progress in geochronology techniques, McDonough et al. (2000b, c) proposed a different tectonometamorphic evolution of the Alberta shield (see Section 2). In their interpretation, towards the end of the Taltson tectonomagmatic event around ca. 1932 Ma, a major thrust zone developed at midcrustal levels, allowing a thrust sheet of granulite- to upper amphibolite–facies rocks to advance northeastward and to juxtapose the hot rocks on top of the ca. 1962 Ma Andrew Lake pluton (ca. 1962 Ma) and the ‘Waugh Lake Group’. New K-Ar muscovite (1745 Ma) and biotite (1739 Ma) dates (Appendix 4, Table 10), contributed by McDonough et al. (2000b), are consistent with previous dates obtained with the same analytical technique from the low-grade tectonite, but their ambiguous geological significance (cooling or tectonism) persisted. The following three lines of evidence appear to contradict the thrust zone interpretation: 1) although shallowly dipping foliation has been occasionally observed, it does not define consistent zones at map scale and does not contain thrust kinematic indicators; 2) the foliation throughout the Taltson basement complex, Taltson granitoid plutons and Waugh Lake Complex is subvertical and more or less concordant; and 3) the Waugh Lake Complex is nowhere closer than 2 km to the postulated thrust zone, and no metamorphic and strain zonation exists.

These lines of evidence are discussed in more detail below:

1. The author’s examination of outcrops at key locations along seven traverses across the thrust zone traced by McDonough et al. (2000b) could not confirm the presence of a regionally consistent thrust zone (Figure 3). The elongated Andrew Lake pluton, exposed between Lindgren and Swinnerton lakes, has been interpreted as a ‘tectonic window’. On its western side, the contact between vertically foliated granitoid and supracrustal rocks near Lindgren Lake is marked by vague, strike-parallel mineral stretching; the contact of the pluton with Rutledge River Complex on the east side, at Swinnerton Lake, is defined by westerly dipping foliation with no obvious mineral stretching (Figure 21a). The trace of the main segment of the thrust zone was postulated to occur beneath northerly trending Andrew Lake and the muskeg area north of it, and to crop out on two islands marked as ‘Thrust Island North’ and ‘Thrust Island South’ in Figure 3. No clear evidence of a major thrust zone was found. The rocks exposed along the inferred trace of the Andrew Lake thrust at the north end of the lake are biotite gneiss with quartzofeldspathic and quartz segregations and with the typical vertical foliation found throughout the Taltson basement complex (Figure 21b). On the southern island, the foliation in well-exposed orthogneiss is somewhat shallower, but no obvious thrust-related stretching lineation could be found. An isolated quartz segregation shows crenulation folds with relatively steeply plunging axes whose orientation corresponds to the inferred northeasterly directed

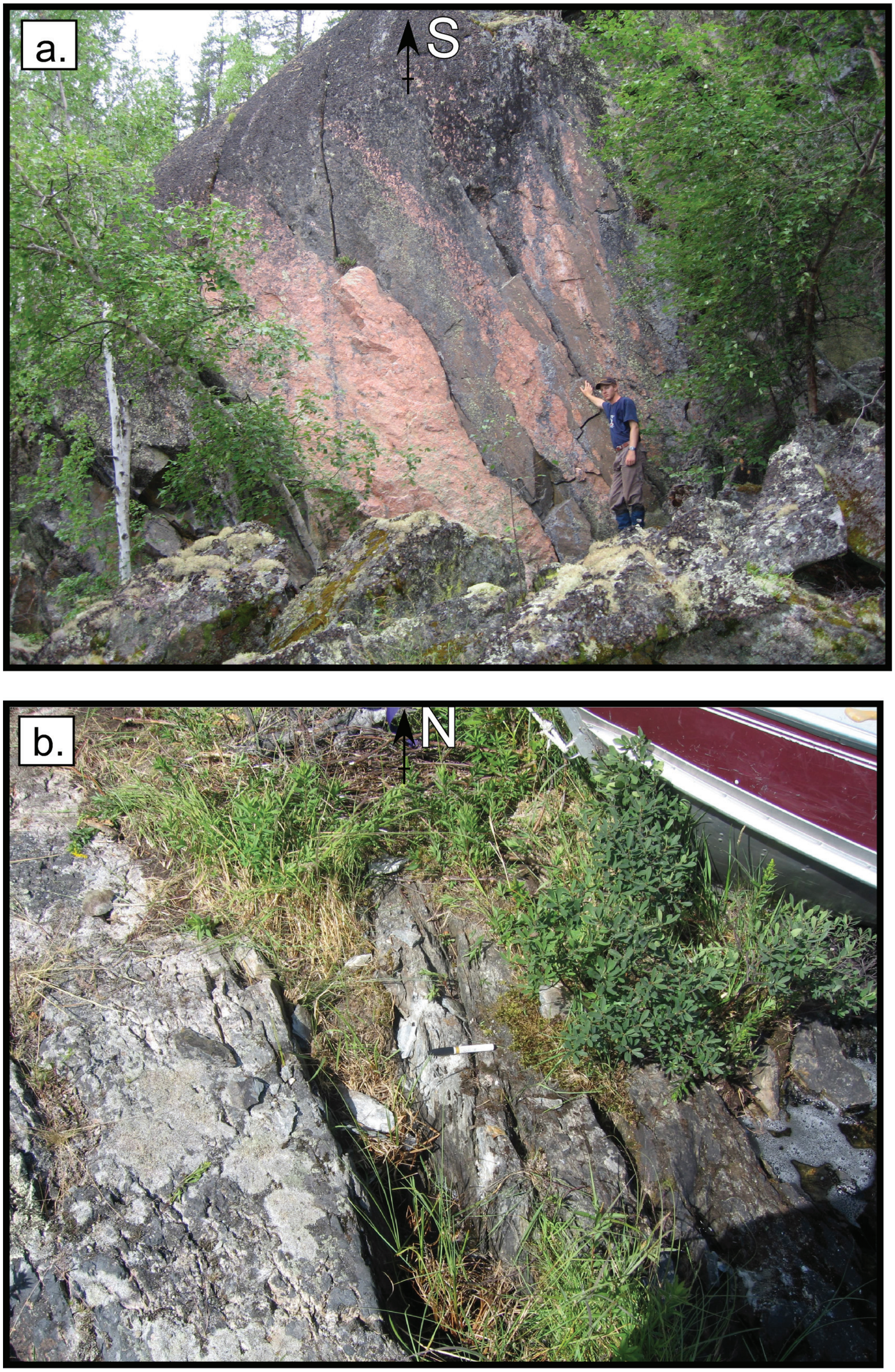


Figure 21. Outcrops along the 'Andrew Lake thrust' of McDonough et al. (2000b): a) biotite gneiss with sub-concordant granite pegmatite bodies dipping west on the west shore at the north end of Swinnerton Lake; b) vertical foliation with no obvious lineation in fine-grained gneiss on the island at the north end of Andrew Lake.

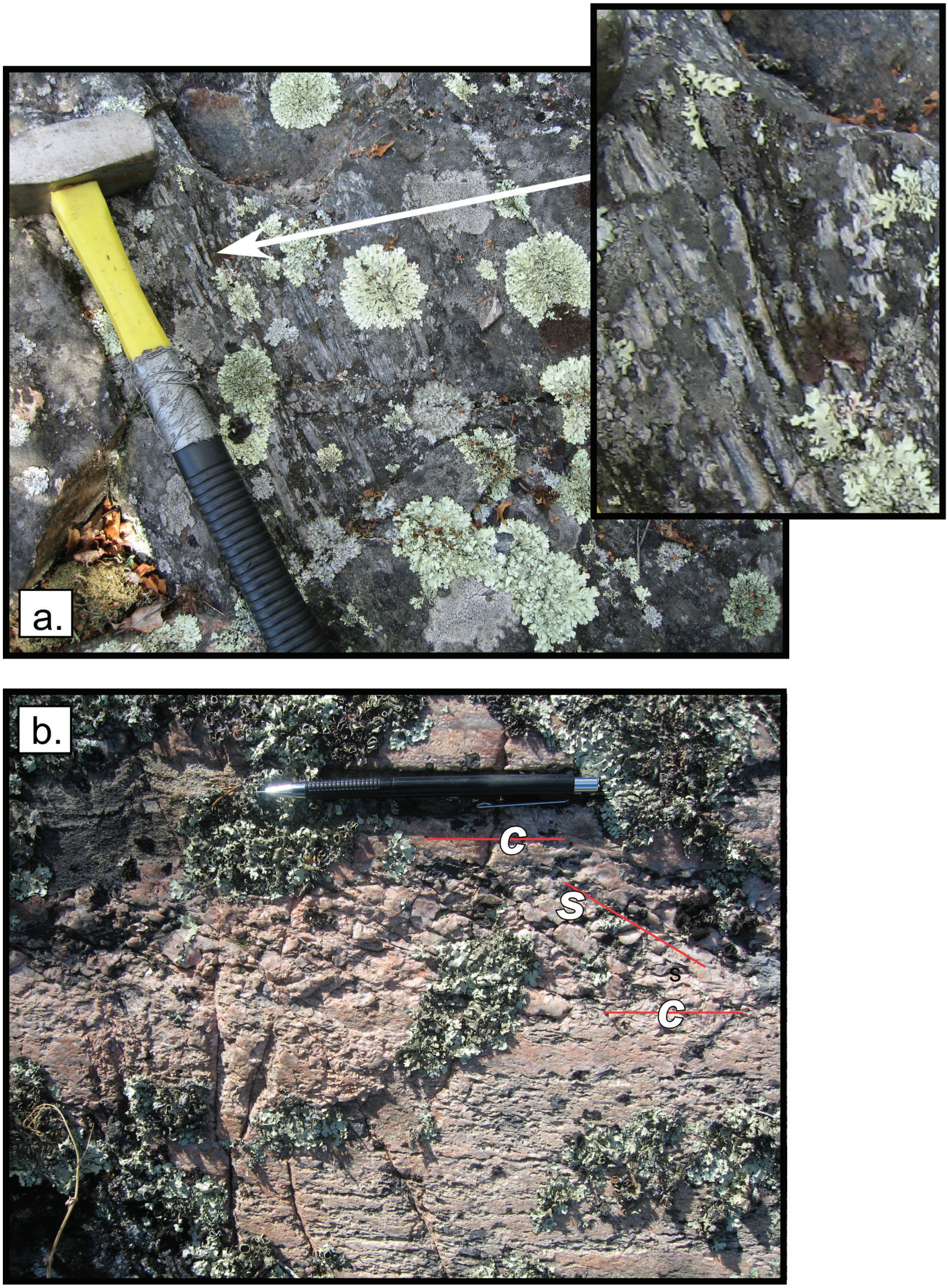


Figure 22. Kinematic indicators in orthogneiss along the 'Andrew Lake thrust' of McDonough et al. (2000b) exposed on the small island in the southern part of Andrew Lake near UTM Zone 12, 548652E, 6633883N, NAD83: a) down-dip lineation formed by crenulation of a quartz segregation (trend and plunge 255°/30° in a foliation oriented 275°/37°N (hammer lies on foliation); b) 'c/s'-type structure in a granite gneiss, suggesting sinistral displacement.

thrust (Figure 22a), but the involved mechanism is likely similar to the steeply dipping folds observed throughout the area, rather than being thrust related. Other outcrops show 'c/s' fabrics that provide unambiguous evidence of a sinistral component of movement (Figure 22b).

2. The observed foliation in the high-grade Taltson basement complex and Rutledge River Complex near the postulated thrust sheet is typically northerly trending and subvertical. The Andrew Lake granitoid plutons just below the thrust zone show no thrust-related deformation; instead, the foliation developed preferentially along their peripheries is concordant with the foliation of the surrounding metamorphic rocks. They have accommodated strain and formed kilometre-scale pods within the late phases of ductile deformation of the Taltson basement complex.

3. In Alberta, the Waugh Lake Complex encompasses an area about 4 km wide, in which the only index minerals are biotite and chlorite, with no prograde metamorphic zonation towards the postulated 'Andrew Lake thrust'. The remaining interval to the trace of the thrust consists mainly of the Andrew Lake pluton, in which the only noticeable zonation is the ductile deformation along its peripheries that surround a less deformed to massive granitoid core (Figure 3). The Waugh Lake biotite-chlorite-sericite schist assemblage has been followed southeasterly along the traverse between Ney and Unger lakes, where it continues directly along strike into Koster's (1971) 'western metasedimentary and metavolcanic complex' (see Section 4.3). This belt of low-grade rocks not only departs from, but also continues beyond the southern extension of the 'thrust', so its metamorphism is unambiguously independent of the 'thrust'.

Re-evaluation of existing geochronological data from the Taltson basement complex and Taltson plutons shows that K-Ar dates record uplift and cooling. A hornblende date of 1940 Ma and biotite dates in the 1820–1780 Ma range (120–160 Ma apart) from the Taltson basement complex, hornblende dates of 1930 and 1900 Ma and a biotite date of 1810 Ma from the Andrew Lake Pluton (80–120 Ma apart) all suggest slow uplift of the presently exposed structural level through the approximate 530°C isotherm (closure of the argon system in hornblende) and the approximate 280°C isotherm (closure of the argon system in biotite). Subsequent K-Ar analyses of hornblende-biotite pairs from the Taltson basement complex (1920 Ma and 1797 Ma), and from the Andrew Lake (1872 and 1784 Ma) and Colin Lake (1803 and 1778 Ma) plutons allow the calculation of uplift rates of approximately 0.04, 0.07 and 0.24 mm/a, respectively (Appendix 4, Table 10). Although the K-Ar dates, and implicitly the calculated uplift rates, may include significant errors, they are consistent with slow uplift and cooling of the crust, somewhat accelerated by the emplacement of the Taltson plutons.

The K-Ar hornblende age of 1840 Ma reported by Baadsgaard and Godfrey (1972) is from a granitoid body with a U-Pb zircon emplacement age of ca. 1971 Ma (McDonough and McNicoll, 1997). Like other Taltson plutons and the Taltson basement complex, the igneous and high-grade mineral parageneses, respectively, are clearly older than the ages recorded by the K-Ar system; hence, they only record uplift and cooling. However, the significance of K-Ar dates from the low-grade Waugh Lake tectonite is less clear. Quartz-albite-biotite-sericite schist of the Waugh Lake area has been formed in the low-grade temperature range that includes the closure temperatures of biotite, sericite and feldspar. Thus, reported biotite and muscovite dates from the foliated low-grade rocks of the Waugh Lake assemblage that range between 1830 and 1740 Ma may record syntectonic mineral growth during foliation development at temperatures of approximately 350–280°C. Thus, foliation development in the Waugh Lake Complex appears to be quasicontemporaneous with the passage of the adjacent crust through the muscovite-biotite isotherms at ca. 1.84–1.73 Ga (i.e., approximately 120–240 Ma after emplacement of the Andrew Lake [ca. 1.96 Ga] and Colin Lake [ca. 1.97 Ga] granitoid plutons in the Andrew Lake area).

To better constrain the timing of low-grade foliation development, four samples of sericite-muscovite schist (phyllonite) were collected along the Alberta-Saskatchewan border (Figure 18) for $^{40}Ar/^{39}Ar$ dating:

DP5-67: Zone 12, 555680E, 6636276N (NAD83)
DP5-74: Zone 12, 555612E, 6630083N (NAD83)
DP5-75B: Zone 12, 555603E, 6631715N (NAD83)
DP5-76: Zone 12, 555615E, 6631788N (NAD83)

These samples have been analyzed at the Geological Survey of Canada and the following comments extracted from O. van Breemen's report, submitted to AGS in May 2006. The gas-release spectrum for sample DP5-67 (Figure 23a) shows a well-defined plateau for the eight highest temperature gas-release steps, representing 54% of the total released ^{39}Ar. The combined weighted average $^{40}Ar/^{39}Ar$ age for these Ar release fractions is 1826 ±7 Ma. The four preceding lower temperature steps yielded slightly older ages that may be indicative of excess argon. However, a reverse isochron plot (Roddick et al., 1980) including these fractions yields an age that is not significantly different from the plateau age. For sample DP5-74 (Figure 23b), a weighted average age of 1820 ±7 Ma was obtained for the four highest temperature steps, which account for 29% of the total ^{39}Ar released. The lower temperature steps yielded slightly younger ages, which is indicative of argon loss. For sample DP5-75B (Figure 23c), a plateau age was obtained for the eight highest temperature steps, with the preceding steps giving significantly younger ages. A weighted average age of the last eight heating steps, representing 98% of the total ^{39}Ar released, is 1829 ±7 Ma. For sample DP5-76 (Figure 23d), a plateau age of 1840 ±7 Ma was obtained for seven of the eight highest temperature steps, representing 85% of the total ^{39}Ar released. The two preceding heating steps showed minor excess ^{39}Ar. The age 'dropout' at 5% laser power is probably due to degassing of inclusions. Although the uncertainties do not overlap, there is overall agreement between the $^{40}Ar/^{39}Ar$ ages, suggesting they relate to the same event in the 1840–1820 Ma age range. For moderate cooling rates, the closure temperature of Ar in muscovite is generally assumed to be about 350°C (Purdy and Jäger, 1976; Jäger, 1979). These $^{40}Ar/^{39}Ar$ ages on muscovite-rich schist samples likely date late movements along the phyllonite belt west of Waugh Lake. However, similar sericite-bearing schist is found throughout the Waugh Lake Complex (Appendix 1), which may indicate that the Waugh Lake metamorphism is, in fact, shear zone related and ca. 100 Ma younger than previously inferred.

4.3 Regional Correlations

Examination of existing geological maps of northeastern Alberta, northwestern Saskatchewan and the Northwest Territories revealed that, in fact, the Waugh Lake Complex is not a small outlier in northeastern Alberta as described by McDonough and McNicoll (1997), but forms a rather extensive belt. Thus, the eastern portion of the geological map included in Watanabe's M.Sc. thesis (1965) partly overlaps with a more regional map across the border in Saskatchewan, where the low-grade sequence has been mapped as a 25 km long by 0.8–1 km wide, westward-convex belt that extends from 59°45'N to 60°N (Figure 24 and Appendix 2). North of 60°N, the low-grade metamorphic belt projects into several belts of 'metasediments' of the Tazin Group, which have been mapped for a distance of at least 90 km to 60°43'N (Mulligan and Taylor, 1969; Bostock, 1992). Thus, the 'Waugh Lake Group', the 'western meta-sedimentary and meta-volcanic unit' and, in part, the 'Tazin Group' appear to be local designations of the same low-grade lithotectonic assemblage. These merging and diverging belts of steeply dipping, low-grade metamorphic tectonite, with widths of 0.5–1.5 km, make up a network more than 150 km long at the eastern margin of the Taltson magmatic zone. Its cartographic expression is strikingly similar to that of any major transcurrent shear zone with anastomosing branches (Figure 24).

Existing isotope dates along these belts are rather inconclusive, as the field relationships and the rock type identification are controversial. Near Waugh Lake in Alberta, igneous rocks within the Waugh Lake

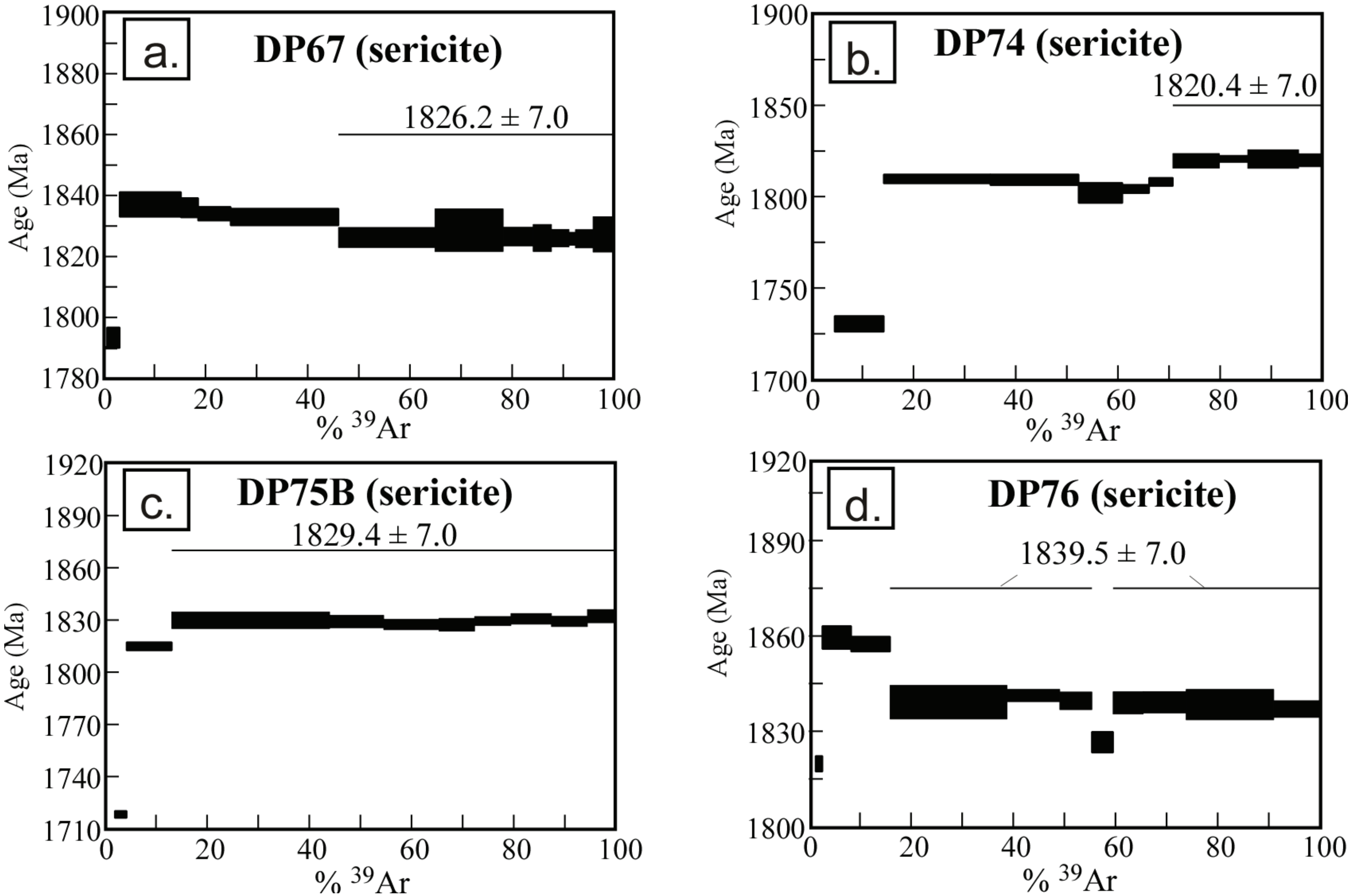


Figure 23. Argon/argon gas release spectra for samples of sericite-muscovite schist from the Waugh Lake Complex, showing the ^{39}Ar fraction at each temperature step and the 2σ uncertainty for each analysis; see text for the discussion of each diagram.

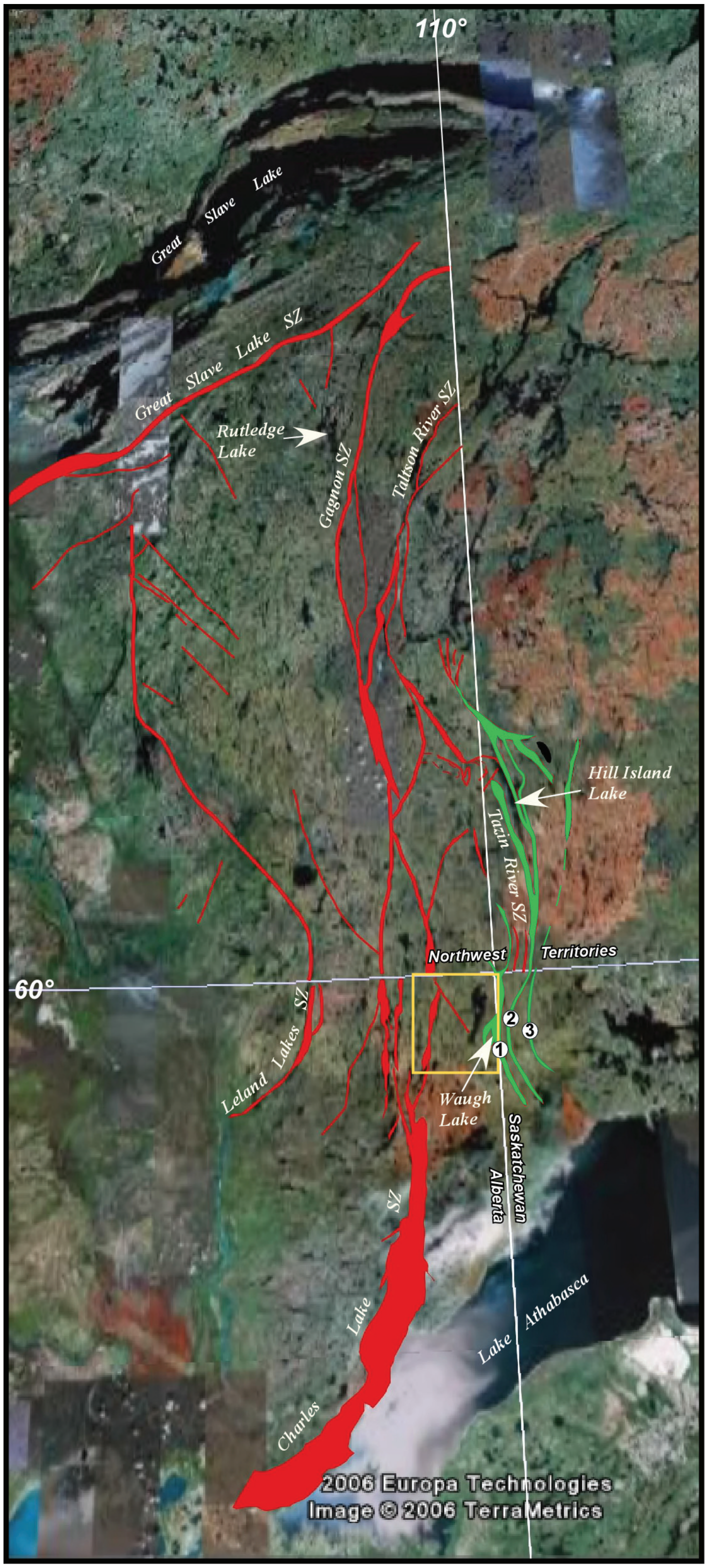


Figure 24. Probable extent of the retrogressive shear zones (in green) from northeastern Alberta and adjacent northwestern Saskatchewan into the Northwest Territories. Shear zones: 1, Waugh Lake, 2, Kornash Lake, 3, Thainka Lake. In red are granulite to greenschist shear zones known in the region (e.g., Berman and Bostock, 1997; McDonough et al., 2000c). Yellow rectangle outlines the Andrew Lake map area (NTS 74M/16).

Complex have been interpreted as either intrusions in the Waugh Lake sedimentary rocks, or as Taltson basement interfolded with the 'Waugh Lake Group' sedimentary rocks (Figure 7). Existing K-Ar and the author's $^{40}Ar/^{39}Ar$ dates discussed in Section 4.2.4 indicate that tectonism along the belt of low-grade Waugh Lake tectonite took place or continued around 1830 Ma, similar to the Charles Lake shear zone (Plint and McDonough, 1995). Zircon U-Pb ages of the Colin Lake (ca. 1971 ±4 Ma), Andrew Lake (ca. 1962 +16/–10 Ma) and Martyn (1973 ±3 Ma) granitoid plugs do not constrain the age of metamorphism of the Waugh Lake assemblage, because these and other igneous bodies have sheared peripheries and were clearly affected by the same tectonothermal event that affected the controversial layered rocks in the Waugh Lake area. The U-Pb dates of 2.7 Ga and 2.3–2.02 Ga on 'detrital' zircon from the 'muscovite-bearing feldspathic wacke with centimetre-scale clasts' reported by McDonough and McNicoll (1997) are typical for the Taltson orthogneiss, and the rock may represent a Taltson granite gneiss sliver overprinted by brittle-ductile strain.

Along Hill Island Lake in Northwest Territories, a narrow belt of low-grade 'turbiditic' rocks grades into andalusite-bearing mylonite near Tazin River fault on the east shore of the lake and into cordierite-bearing rocks on the west shore of the lake near the contact with the 1934±2 Ma Natael granite (Bostock and van Breemen, 1994). The age of metamorphism for this sequence was inferred based on a metamorphic zircon upper intercept age of 1920 +6.5/–6 Ma from a quartzitic "bed" in the low-grade zone of the "turbiditic" sequence (Bostock and van Breemen, 1994; p. 1359). This age was considered "a problem" because the "turbiditic" sequence was crosscut in other outcrops by muscovite granite pegmatite spatially related to the 1934±2 Ma Natael granite (Bostock and van Breemen, 1994).

The author suggests that the apparently conflicting U-Pb zircon ages may be explained by and testify for strain partitioning during syntectonic exhumation of the Tazin River shear zone. The texture of the quartz-rich rock described by Bostock and van Breemen (1994) is clearly mylonitic and includes two zircon populations: one consisting of 2.17 Ga and 2.13 Ga colourless zircon grains, which are not uncommon in the western Rae crust (Figures 20 and 40), and the younger brown, high-U zircon population crystallized at ca. 1.92 Ga, during shearing at elevated temperatures (van Breemen et al., 1987). The peripheries of the Hill Island sequence may record the early Tazin River shear zone tectonic overprint on the Rae crust at mid-crustal level (amphibolite facies mylonite), with foliations sealed at ca. 1.93 Ga by the Natael granite and associated muscovite-bearing pegmatite. Subsequent syntectonic exhumation of the Tazin shear zone accompanied by strain localization within the central narrower greenschist to subgreenschist facies belt along the axis of the lake may have locally resulted in the highly strained, quartz-rich rock with a muscovite-biotite-chlorite retrogressive mineral assemblage sampled by Bostock and van Breemen (1994). The zircon grains defining the regression line with a concordia intercept at ca. 1.92 Ga likely record metamorphic zircon growth in the lower amphibolite facies and Pb loss during subsequent tectonism under green schist facies conditions.

In this context, the dated sericite from the quartz clast–bearing samples would be the first reliable $^{40}Ar/^{39}Ar$ direct dating of the phyllonite texture within the Tazin River–Waugh Lake system of low-grade tectonite. The Tazin-Waugh tectonic system may represent late localization of strain under low-grade conditions during progressive exhumation of a major shear zone through the brittle-ductile and shallower structural levels. This interpretation is consistent with the evolution of other major Taltson shear zones, which record granulite to amphibolite tectonism between ca. 1940 and 1920 Ma, and greenschist-facies deformation at ca. 1800 Ma (Plint and McDonough, 1995; McDonough et al., 2000c). Some basalt and andesite bodies along the low-grade tectonite belts may represent syntectonic intrusions—the result of hydration and/or decompression melting within the shear zone. Polyphase deformation along these belts is emphasized by complex crosscutting relationships between tectonite and veins (Figure 25).

Figure 25. Spectacular quartz-tourmaline veins in the Waugh Lake Complex north of the Waugh Lake elbow.

5 Conclusions

The AGS reconnaissance field program in the Andrew Lake area consisted of selected traverses over the main lithological units and contacts, and a preliminary petrographic study supported by geochemical and isotope data.

Seven traverses across the trace of the previously inferred, west-dipping Andrew Lake thrust (McDonough et al., 2000b, c) could not confirm the presence of a regional thrust zone. The Waugh Lake Complex includes a variety of typically steeply dipping and northerly trending rocks that are highly strained under low- to very-low grade metamorphic conditions. At several places, a gradational transition has been observed between layered low-grade tectonite and foliated to massive igneous rocks. The low-grade tectonite yielded T_{DM} and ε_{Nd} values that coincide, within analytical error, with the adjacent Taltson plutons. One of the numerous igneous pods within the Waugh Lake tectonite yielded a U-Pb zircon emplacement age of ca. 1973 Ma, similar to that of the Taltson plutons. Argon/argon dates between ca. 1840–1820 Ma from four phyllonite samples collected on the east side of Waugh Lake are contemporaneous with the late phases of strike-slip displacement on major shear zones overprinting the Taltson magmatic zone. Field and microscope observations suggest that at least parts of the low-grade Waugh Lake Complex may be directly derived from the adjacent crust (i.e., the Taltson magmatic zone) through shearing and metamorphic re-equilibration.

The author's fieldwork has confirmed that the Waugh Lake Complex projects north and south into similar rocks that define narrow belt, 1–2 km wide, extending more than 100 km from west of Harper Lake in Saskatchewan to Hill Island Lake and beyond, in Northwest Territories. The low-grade metamorphism of the Waugh Lake Complex and its correlatives is most likely shear-zone related. The protoliths include mostly sheared and retrogressed rocks of the adjacent crust, and may incorporate sedimentary and volcanic units deposited atop the shear zone during its late evolution through shallow structural levels. This hypothesis is consistent with the lack of evidence for a regional metamorphism in the region after the cessation of Taltson tectonism.

The author's field and isotope data, corroborated by previous mapping and aided by the re-evaluation of previous isotope work, suggest that the geological evolution of the region includes:

1. Paleoproterozoic tectonothermal recycling of the western Archean Rae Terrane, with the development of the semicontinuous suite of ca. 2.4–2.3 Ga Arrowsmith granite bodies;
2. transcurrent tectonics under high- to medium-grade metamorphic conditions overprinting a wide linear zone in the westernmost portion of the Rae Terrane, which resulted in foliation and partial melting of the granitoid crust and development of the ca. 2.4–2.1 Ga Taltson basement complex (biotite and hornblende granite gneiss);
3. phased melting of wide domains of crust and successive emplacement of the voluminous Taltson plutons between 2.0 and 1.94 Ga, the local representatives of this event in the Andrew Lake area being the ca. 1.96 Ga Andrew Lake and ca 1.97 Ga Colin Lake granitoid bodies;
4. strain localization at midcrustal levels within the strike-slip 'dry' mylonite belts (LLSZ, CLSZ and BLSZ) and 'hydrated' belts of strain marked by the migmatitic biotite±aluminosilicate schist and gneiss, prior to ca. 1.93 Ga;
5. decompression- and/or hydration-triggered melting within and near these belts of strain concentration, recorded by local intrusion of commonly muscovite-bearing granitoid plugs (e.g., the ca. 1.93 Ga Charles Lake and ca. 1.92 Ga Wallace Island granitoid bodies);
6. gradual cessation of deformation, with annealing of the high- to medium-grade mylonite textures and

further localization of strain in narrow and discontinuous, locally anastomosing belts of greenschist- and subgreenschist-facies mylonite; and

7. transfer of strain towards the eastern periphery of the Taltson magmatic zone and concentration of deformation at its eastern periphery within the Tazin River–Waugh Lake shear zone system (ca. 1.93–1.82 Ma), which appears to be quasi-contemporaneous with the tectonothermal event along the Virgin River shear zone (1820 Ma), in the southern exposed portion of the Snowbird Tectonic Zone and with the main Trans-Hudson tectonism recorded by the uraniferous Wollaston belt (1850–1800 Ma); at this stage, it is possible that sedimentary and volcanic rocks deposited on the exhumed lower granulite–upper amphibolite facies crust (after the erosion of several kilometres of the overlying crust between ca. 1.93 and 1.84 Ga) were incorporated in the shallow structural levels of shear zones.

This report includes preliminary results and recognizes that well-documented conclusions on the geological evolution of the Andrew Lake area require further detailed and systematic field and analytical work. Systematic sampling of several varieties of controversial rocks for geochemical and multisystem isotope analysis may shed new light on their origin and evolution.

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Appendix 1 – Simplified Geological Map of the Andrew Lake Area Showing the Spatial Distribution of Low-Grade Mylonite Belts

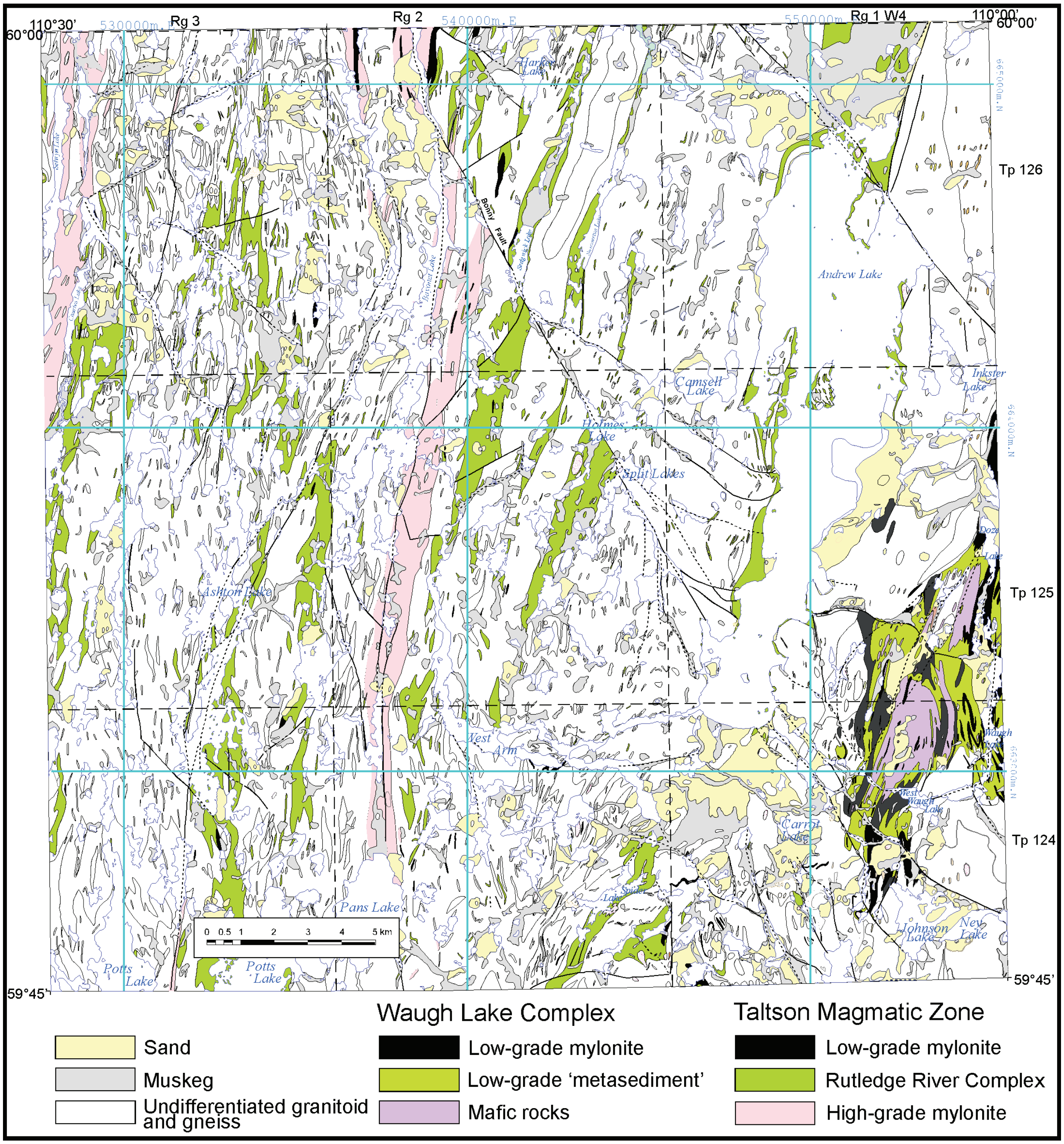


Figure 26. Simplified geological map showing the spatial distribution of low-grade mylonite bands in the Andrew Lake area (compiled from Godfrey, 1961, 1963, 1986; Iannelli et al., 1995; McDonough et al., 2000b). Within the Taltson basement complex, vertical layers of sericite-chlorite schist have been interpreted as retrogressive low-grade mylonite (phyllonite); note their association with the high-grade mylonite belts and the elongated domains of highly strained and migmatized supracrustal rocks. Within the 'Waugh Lake Group,' the same rock types have been interpreted as prograde metamorphosed strata.

Appendix 2 – Tentative Correlation to the East of the Andrew Lake Map Area, Across the Alberta-Saskatchewan Border

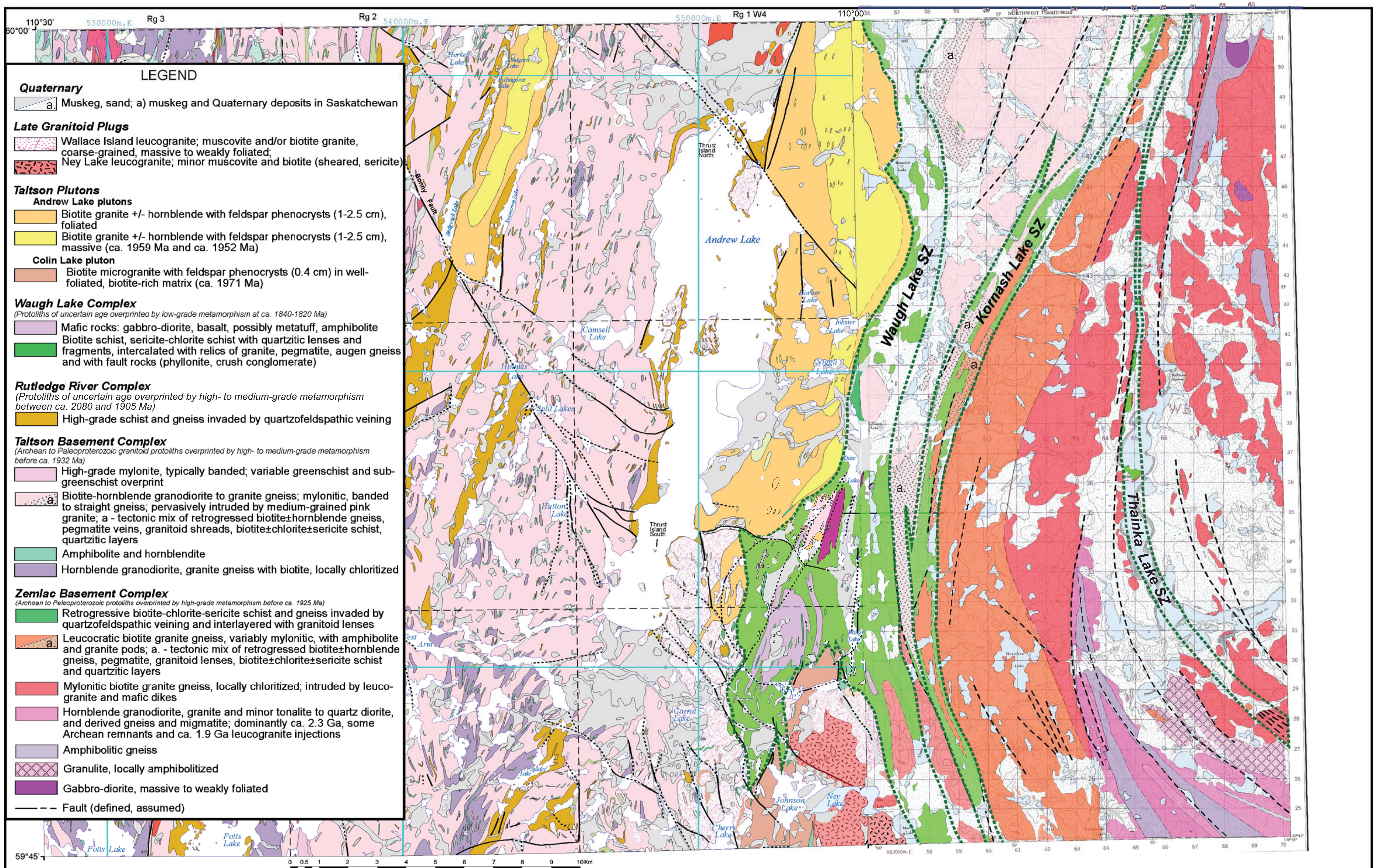


Figure 27. Tentative correlation of map units defined in the Andrew Lake area with map units separated in northwestern Saskatchewan by Koster (1971) and Ashton (2009).

Appendix 3 – Summary of Initial Isotope Dates from the Northeastern Part of the Alberta Shield

Table 9. Summary of initial isotope dates from the Taltson magmatic zone in the northeastern Alberta shield (Godfrey and Baadsgaard, 1962; Baadsgaard and Godfrey, 1967, 1972).

Sample	Easting	Northing	Rock type	Location	Dates				
					K-Ar			Rb-Sr	U-Pb
					1962	1967	1972	1972	
Taltson plutons:									
60-729-3	552157	6646099	Granodiorite	Andrew Lake peninsula		1790 B	1800 B		1950
60-729-4	552871	6646088	Porphyry	NE shore of Andrew Lake		1800 B	1810 B	1860	1900
			quartz monzonite			1910 H	1930 H		
63-95-1	553055	6645750	Foliated biotite granite	NE shore of Andrew Lake			1800 B		
90-2	553548	6647936	Porphyroblastic granite	NE of Andrew Lake	1740 B				
102-1	551370	6645944	Massive granite	NE shore of Andrew Lake	1830 B				
Waugh Lake Complex:									
60-151-11	555731	6639631	Quartz monzonite	N of Doze Lake		1810 B	1820 B	1800	2100
63-81-7	555876	6639627	Granite	N of Doze Lake			1760 B		
63-98-2	553146	6631810	Lava flow	W of Waugh Lake			1760 B		
58-546-1	552391	6630927	Lava flow	W of Waugh Lake			1790 B		
58-13-1	551538	6630399	Lava flow	W of Waugh Lake			1830 M		
58-1035-2	552701	6629446	Sericite schist	W of Waugh Lake			1770 M		
58-4-4A	552172	6627263	Biotite schist (?)	S of Waugh Lake			1780 M		
63-94-2	554538	6629560	Biotite microgranite	N shore of Waugh Lake			1740 B		
63-616-2	554180	6629110	Biotite granite 'p'	S shore of Waugh Lake			1750 B	1730	
60-128-5	555092	6629389	Mafic rock	S shore of Andrew Lake		1810 B	1820 B		1900
						1840 H	1860 H		
63-94-1	555211	6629365	Granite 'C'	S shore of Waugh Lake				1865	
Taltson basement complex:									
57-214-1	542757	6634244	Granitoid gneiss	N of western arm of Andrew Lake			1820 B		
58-43-4	541141	6631217	Metasedimentary rock	N shore of western arm of Andrew L.			1830 B		
58-46-1	539376	6631627	Metasedimentary rock	W shore of western arm of Andrew L.			1780 B		
63-617-1	541543	6644158	Granite gneiss	W of Sedgwick Lake near fault			1620 B		
63-628-4	543807	6629556	Granite gneiss	S of western arm of Andrew Lake			1820 B	1730	
63-628-5	544389	6634829	Granite gneiss	W of Hutton Lake			1800 B	1700	
59-90-2	540119	6646699	Quartz monzonite	E of Bayonet Lake		1800 B	1810 B	2240	2250
			gneiss			1740 M	1750 M		1900 A
60-149-1	544148	6645142	Quartz monzonite	E of Swinnerton Lake		1810 B	1820 B	2060	2300
			gneiss			1930 H	1940 H		
Retrogressed and mineralized Taltson basement complex									
58-44-1A	539291	6630744	Biotite schist	S of western arm of Andrew Lake		1790 B	1800 B		
58-44-1B	539291	6630744	Biotite-quartz phyllite	S of western arm of Andrew Lake			1790 B		
58-44-1C	539291	6630744	Pegmatite granodiorite	S of western arm of Andrew Lake					1900 U

Notes:
Sample numbers are Godfrey's field site numbers from Godfrey and Baadsgaard (1962) and Baadsgaard and Godfrey (1967, 1972)
UTM locations digitized from Godfrey's maps. Some isotope samples are missing from the original AGS master file with Godfrey's sample locations; an attempt has been made to match the locations from Figure 3 (geochronology sample locations) of Baadsgaard and Godfrey (1972) with samples from the master file or with sample locations shown on Godfrey's (1963) Andrew Lake South map: 58-1035-2 (32); 63-94-2 (33); 63-616-2 (35); 63-628-4 (36). Some rock types have been inferred from Godfrey's maps (Ashton Lake District, Andrew Lake North, Andrew Lake South); A, allanite; B, biotite; M, muscovite; H, hornblende; U, uraninite; U-Pb ages are from zircon unless otherwise specified; Rb-Sr ages are whole-rock apatite isochron ages

Appendix 4 – Summary of Recent Isotope Dates from Northeastern Alberta and Northwestern Saskatchewan

Table 10. U-Pb, $^{40}Ar/^{39}Ar$ and K-Ar dates from Andrew Lake (McDonough et al., 2000b) and Colin Lake (McDonough et al, 2000a) map areas, encompassing the northeastern part of the Alberta shield.

Taltson basement complex			Andrew Lake pluton			Colin Lake pluton			Waugh Lake assemblage			Mineral closure temperature
U-Pb	$^{40}Ar/^{39}Ar$	K-Ar	U-Pb	$^{40}Ar/^{39}Ar$	K-Ar	U-Pb	$^{40}Ar/^{39}Ar$	K-Ar	U-Pb	$^{40}Ar/^{39}Ar$	K-Ar	
		1603										
		1732										
		1739										
								1746				
								1778				
								1778				
					1784			1784				
								1784				
		1786						1786				
		1791			1791			1791				
								1791				
		1797						1797				
												B 280°±30°C
								1810				
		1816						1816				
											1739 M	
		1732								1840–1820	1745 M	M 350°C
	1904	1920		1899	1872			1803				H 530°±40°C
			1932			1913–1933						M 700°C
$2335^{\pm 2}$			$1959^{\pm 3}$			$1971^{\pm 4}$			2.01–2.32–2.7			Z>800°C
			$1962^{+16/-10}$									

***Notes*:**

K-Ar and $^{40}Ar/^{39}Ar$ dates are from the Geological Survey of Canada Maps 1953A (McDonough, 2000b) and 1952A (McDonough, 2000a)

B, biotite; H, hornblende; M, muscovite; Mz, monazite; Z, zircon.

U-Pb ages are from McDonough and McNicoll (1997); McDonough et al. (2000a) and McNicoll et al. (2000)

Table 11. U-Pb zircon crystallization ages, suggesting a common Paleoproterozoic magmatic event at ca. 2.4–2.3 Ga (Arrowsmith orogeny) in northeastern Alberta and northwestern Saskatchewan.

Taltson basement complex		Beaverlodge intrusions	
Rock type	**Age**	**Age**	**Rock type**
Granodiorite gneiss, Mercredi Lake block	2382 ±5 Ma		
Pegmatitic granite, Andrew Lake block	2335 ±2 Ma		
	2330 ±3 Ma	2330 ±5 Ma	Macintosh Bay monzogranite
		2321 ±3 Ma	Gunnar monzogranite
		2320 ±44 Ma	Geebee Lake tonalite
Migmatized amphibolite gneiss, Potts Lake block	2295 +6/–5 Ma	2297 ±10 Ma	Hayter Bay monzogranite
		2287 ±14 Ma	Yahyah Lake granite

Note: Data from McNicoll et al. (2000) and Hartlaub et al. (2007).

Appendix 5 – Concordia Diagrams for Relevant Orthogneiss Samples from the Taltson Basement Complex

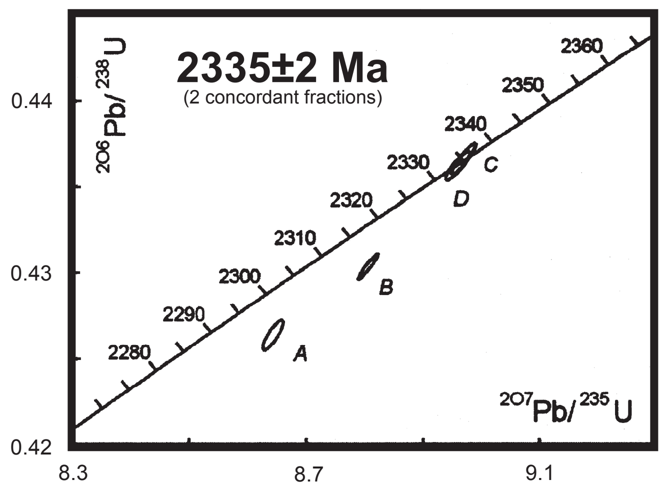


Figure 28. Discordant pegmatitic granite, 'Andrew Lake block,' west arm of Andrew Lake (MSB93-MC10, McNicoll et al., 2000).

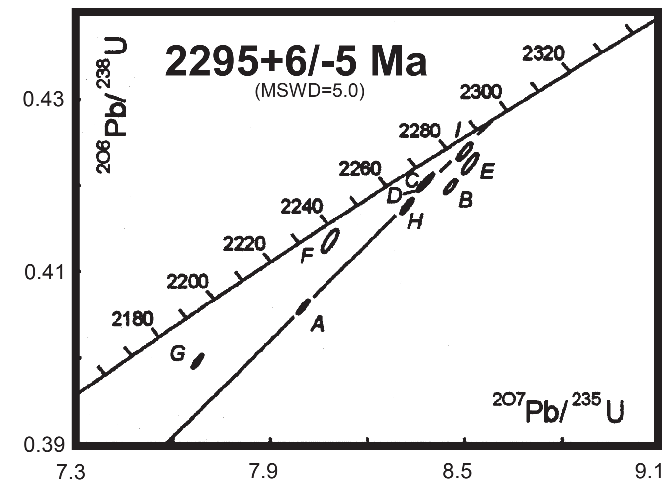


Figure 29. Migmatized amphibolite gneiss, 'Potts Lake block', Alberta–Northwest Territories border (MSB93-114, McNicoll et al., 2000).

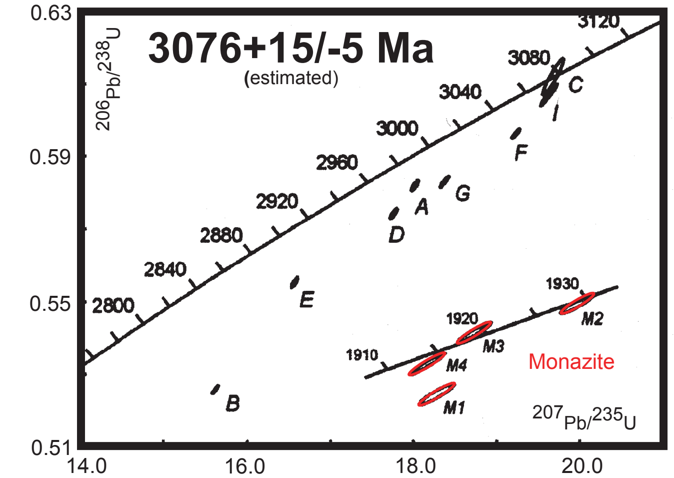


Figure 30. Layered biotite tonalite gneiss 'Andrew Lake block,' 1 km east of CLSZ (MSB93-139, McNicoll et al., 2000).

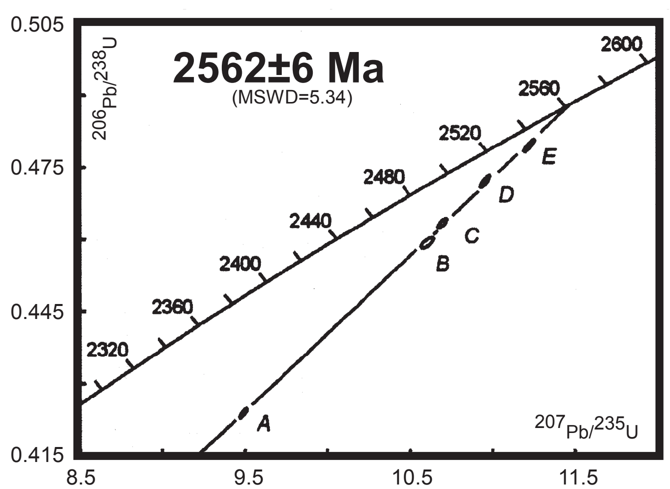


Figure 31. Layered biotite tonalite gneiss 'Andrew Lake block,' Lapworth Point (MSB94-47, McNicoll et al., 2000).

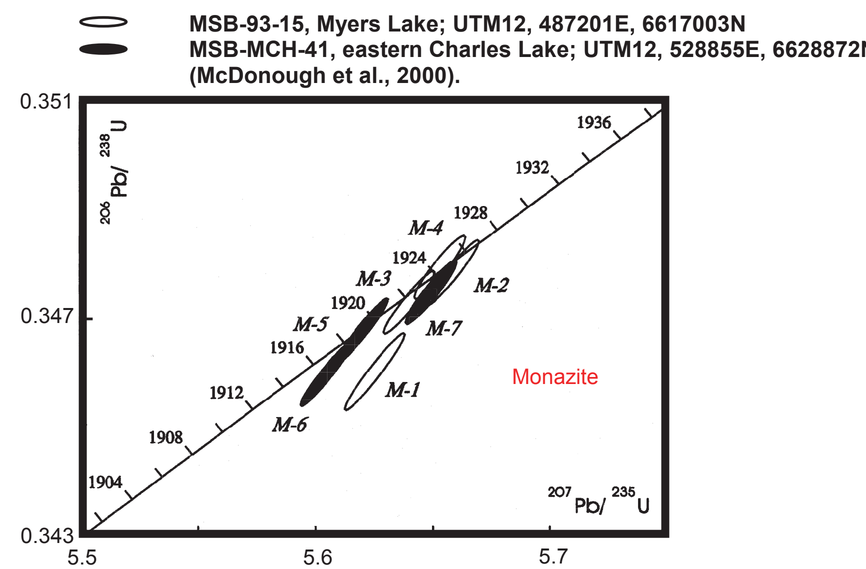


Figure 32. 'Pelitic' gneiss samples MSB-93-15, Myers Lake (UTM Zone 12, 487201E, 6617003N) and MSB-MCH-41, eastern Charles Lake (UTM Zone 12, 528855E, 6628872N).

Appendix 6 – Concordia Diagrams for Relevant Samples from the Taltson Plutons

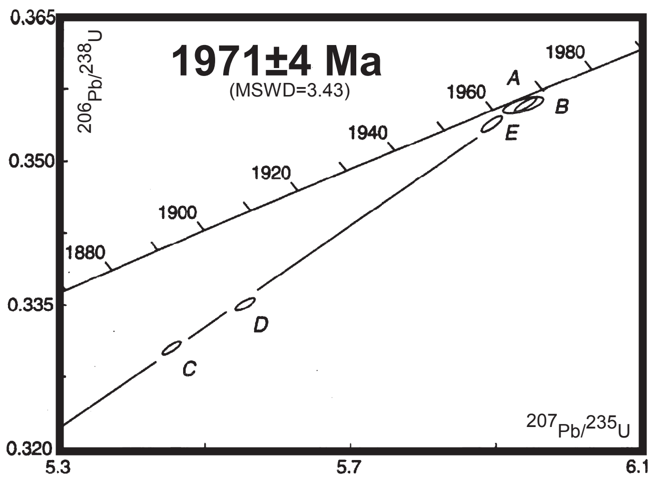


Figure 33. Colin Lake quartz diorite, north shore of Waugh Lake (sample MSB-93-148; UTM Zone 12, 554461E, 6629170N; McDonough and McNicoll, 1997).

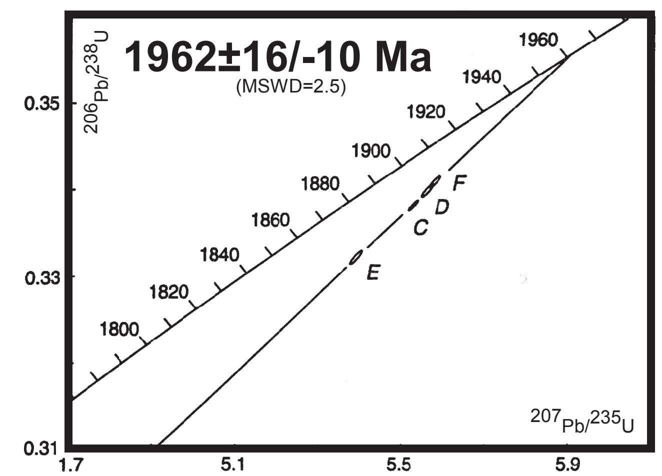


Figure 34. Andrew Lake granodiorite, east shore of Andrew Lake (sample MRB-MAN-27; UTM Zone 12, 550268E, 6635205N; McDonough and McNicoll, 1997).

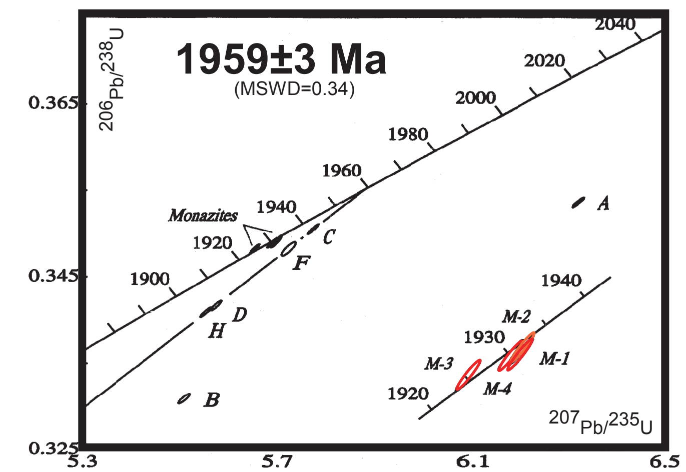


Figure 35. Andrew Lake granodiorite, inland from west shore of Swinnerton Lake (sample MRB-TAN-70; UTM Zone 12, 543262E, 6645435N; McDonough et al., 2000c).

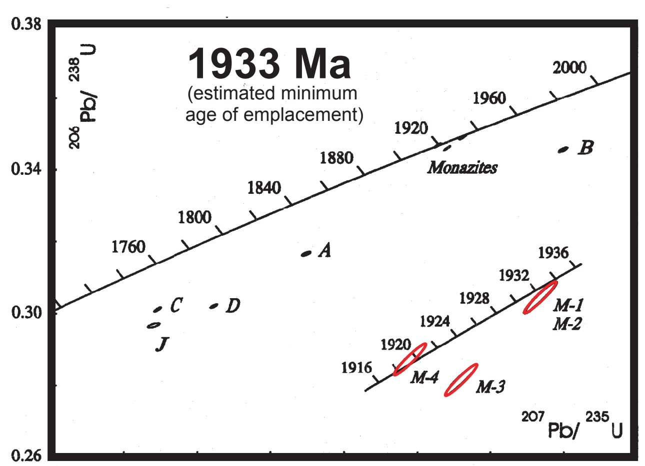


Figure 36. Charles Lake granite (sample MRB-VCH-28; UTM Zone 12, 527015E, 6637166N; McDonough et al., 2000c).

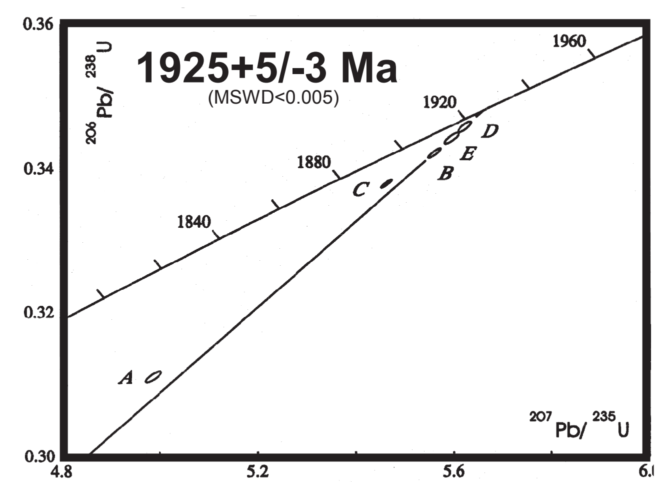


Figure 37. Chipewyan granite (sample MSB-94-55; UTM Zone 12, 498660E, 6536125N; McDonough et al., 2000c).

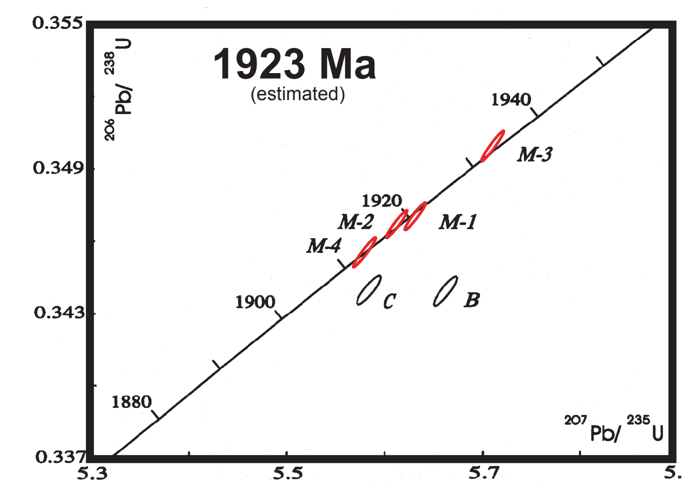


Figure 38. Colin Lake muscovite granite (Wallace Island granite; sample MRB-TAN-61; UTM Zone 12, 551561E, 6645193N; McDonough et al., 2000c).

Appendix 7 – Comparison Between Concordia Diagrams for the Waugh Lake 'Conglomerate' and Taltson Orthogneiss Units

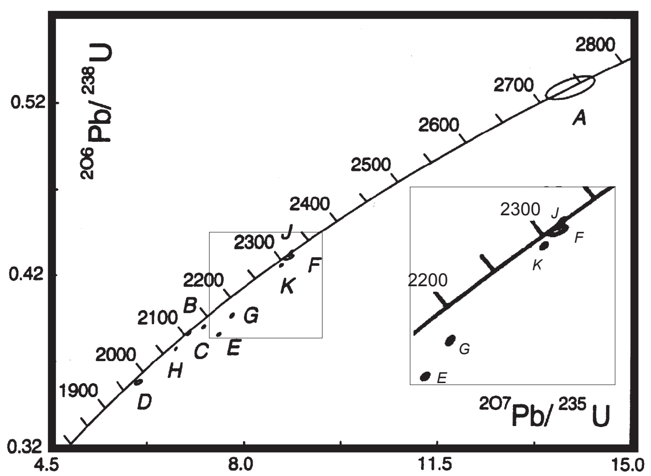


Figure 39. Waugh Lake Group conglomerate (sample MRB-MAN-14; McDonough and McNicoll, 1997).

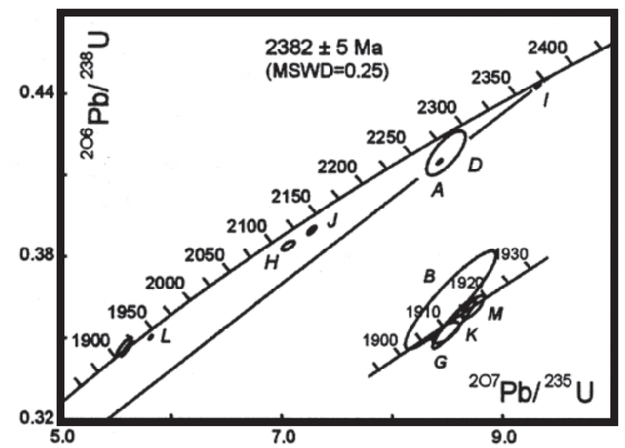


Figure 40. Granodiorite gneiss (sample MSB93-116, Mercredi Lake block; McNicoll et al., 2000).

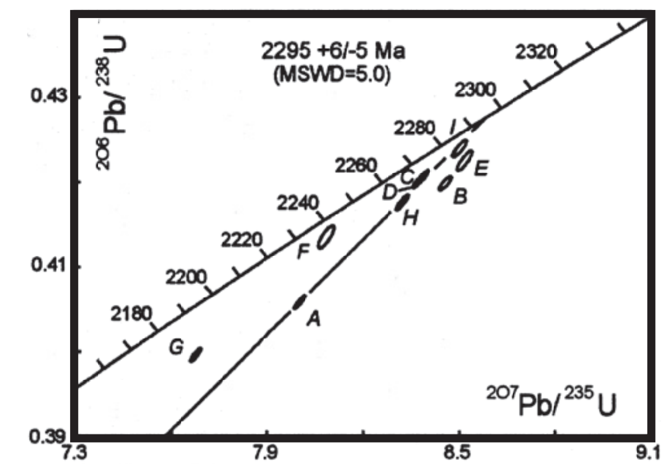
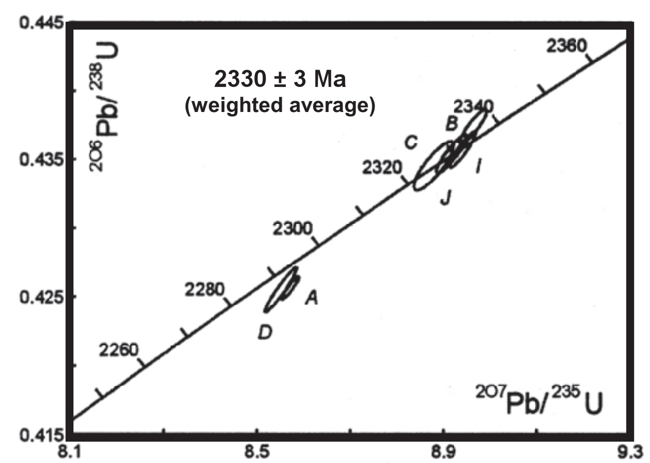


Figure 41. Amphibolite gneiss (sample MSB93-114, Potts Lake block; McNicoll et al., 2000).



.Figure 42. Syenogranite gneiss (sample MSB94-95, Dore Lake block; McNicoll et al., 2000).

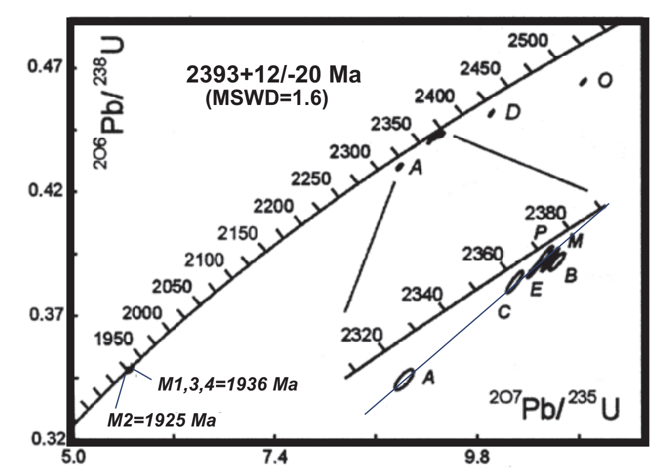


Figure 43. Mylonitic hornblende granite gneiss (sample MSB94-33, Cornwall Lake block; McNicoll et al., 2000).

Appendix 8 – Laser Ablation Raw Data

Table 12. U-Pb laser-ablation multi collector-inductively coupled plasma mass spectrometry (MC-ICP-MS) results for sample KEN-1.

		Atomic Ratios							Apparent Ages (Ma)						
	Grain ID	$^{206}Pb/^{238}U$	1s Error	$^{207}Pb/^{235}U$	1s Error	$^{207}Pb/^{206}Pb$	1s Error	rho	$^{206}Pb/^{238}U$	Error	$^{207}Pb/^{235}U$	Error	$^{207}Pb/^{206}Pb$	Error	Disc %
1	KEN-1_1-1e	0.381	0.012	8.898	0.267	0.169	0.002	0.941	2079.7	54.1	2327.6	27.0	2548.4	17.4	21.5
2	KEN-1_1-2m	0.400	0.012	9.552	0.287	0.173	0.002	0.943	2170.5	57.1	2392.6	27.2	2584.6	17.3	18.8
3	KEN-1_2-1e	0.404	0.014	9.951	0.299	0.179	0.002	0.952	2188.9	61.8	2430.3	27.3	2641.7	17.3	20.2
4	KEN-1-2-2e	0.365	0.013	7.656	0.230	0.165	0.006	0.367	2007.9	59.5	2191.4	26.6	2504.3	61.6	23.0
5	KEN-1_2-3m	0.406	0.013	9.913	0.297	0.177	0.002	0.945	2195.4	58.3	2426.7	27.3	2621.9	17.1	19.2
6	KEN-1_2-4e	0.405	0.014	10.048	0.302	0.179	0.002	0.956	2194.1	62.9	2439.2	27.3	2647.3	17.1	20.2
7	KEN-1_3-1e	0.441	0.014	11.396	0.342	0.187	0.002	0.950	2356.7	62.8	2556.1	27.6	2716.1	16.5	15.8
8	KEN-1_3-2m	0.334	0.011	7.086	0.213	0.151	0.003	0.863	1857.5	53.0	2122.2	26.4	2358.1	28.6	24.4
9	KEN-1_3-3e	0.294	0.009	6.287	0.189	0.155	0.002	0.944	1662.3	46.3	2016.7	26.0	2398.4	17.8	34.7
10	KEN-1_3-4e	0.437	0.013	11.278	0.338	0.187	0.002	0.945	2336.4	59.6	2546.4	27.6	2717.3	16.6	16.7
11	KEN-1_4-1e	0.419	0.014	10.456	0.314	0.181	0.002	0.951	2257.5	64.7	2476.0	27.4	2662.1	18.0	18.0
12	KEN-1_4-2m	0.450	0.015	11.554	0.347	0.186	0.002	0.950	2397.3	64.8	2568.9	27.7	2703.0	16.7	13.5
13	KEN-1_5-1e	0.399	0.014	9.961	0.299	0.180	0.002	0.955	2165.2	61.9	2431.1	27.3	2655.4	17.1	21.7
14	KEN-1_5-2m	0.443	0.014	11.387	0.342	0.186	0.002	0.951	2362.1	63.8	2555.4	27.6	2709.7	16.6	15.3
15	KEN-1_5-3e	0.419	0.014	10.634	0.319	0.184	0.002	0.952	2255.7	62.4	2491.7	27.5	2687.5	16.8	19.0
16	KEN-1_6-1e	0.405	0.016	9.442	0.283	0.170	0.002	0.980	2191.0	71.0	2381.9	27.2	2554.8	18.1	16.8
17	KEN-1_6-2m	0.386	0.012	9.123	0.274	0.172	0.002	0.938	2106.1	57.5	2350.4	27.1	2573.0	18.7	21.2
18	KEN-1_6-3e	0.387	0.013	9.151	0.275	0.171	0.002	0.951	2109.0	58.0	2353.2	27.1	2569.6	16.9	21.0
19	KEN-1_6-4m	0.433	0.014	11.143	0.334	0.187	0.002	0.949	2317.3	63.2	2535.1	27.6	2712.0	17.0	17.3
20	KEN-1_6-5e	0.429	0.014	10.994	0.330	0.185	0.002	0.951	2302.9	63.2	2522.6	27.5	2701.9	16.8	17.5
21	KEN-1_6-6m	0.445	0.014	11.512	0.345	0.187	0.002	0.949	2372.0	63.5	2565.5	27.7	2716.1	16.8	15.1
22	KEN-1_6-7e	0.442	0.015	11.379	0.341	0.187	0.002	0.931	2357.6	67.7	2554.7	27.6	2711.9	21.0	15.6
23	KEN-1_6-8e	0.440	0.014	11.454	0.344	0.188	0.002	0.948	2352.0	62.2	2560.8	27.6	2727.6	16.7	16.4
24	KEN-1_7-1m	0.416	0.014	10.575	0.317	0.185	0.002	0.956	2241.8	63.4	2486.5	27.5	2695.8	16.7	19.9
25	KEN-1_7-2e	0.426	0.014	10.877	0.326	0.185	0.002	0.949	2289.0	61.5	2512.7	27.5	2696.4	16.7	17.9
26	KEN-1_7-3e	0.425	0.014	10.913	0.327	0.186	0.002	0.948	2284.8	61.0	2515.7	27.5	2705.0	16.7	18.4
27	KEN-1_7-4e	0.438	0.014	11.256	0.338	0.186	0.002	0.946	2341.3	61.5	2544.5	27.6	2707.5	16.9	16.1
28	KEN-1_8-1e	0.388	0.013	9.554	0.287	0.178	0.002	0.959	2114.5	62.0	2392.8	27.2	2630.5	17.1	23.0
29	KEN-1_8-2m	0.416	0.014	10.667	0.320	0.187	0.002	0.952	2240.1	61.7	2494.5	27.5	2711.6	16.6	20.5
30	KEN-1_8-3e	0.395	0.014	10.205	0.306	0.186	0.002	0.965	2146.8	64.8	2453.5	27.4	2710.2	17.0	24.4
31	KEN-1_8-4e	0.426	0.014	11.156	0.335	0.189	0.002	0.947	2288.0	61.9	2536.3	27.6	2734.4	17.2	19.4
32	KEN-1_9-1m	0.282	0.012	6.420	0.193	0.166	0.002	0.694	1599.8	61.0	2035.0	26.0	2517.0	22.1	41.0
33	KEN-1_9-2m	0.295	0.009	6.132	0.184	0.151	0.002	0.933	1666.1	44.9	1994.9	25.9	2356.5	19.0	33.2
34	KEN-1_10-1m	0.417	0.015	10.635	0.319	0.185	0.002	0.962	2248.3	65.7	2491.8	27.5	2694.5	16.7	19.6
35	KEN-1_10-2e	0.381	0.014	9.441	0.283	0.179	0.002	0.971	2080.7	65.1	2381.9	27.2	2647.1	17.4	25.0
36	KEN-1_10-3e	0.415	0.013	10.686	0.321	0.187	0.002	0.946	2236.0	58.8	2496.2	27.5	2715.9	16.7	20.9
37	KEN-1_11-1e	0.381	0.012	9.553	0.287	0.182	0.002	0.940	2079.4	55.7	2392.7	27.2	2670.3	17.8	25.8
38	KEN-1_11-2m	0.383	0.012	9.516	0.286	0.180	0.002	0.948	2090.7	56.4	2389.1	27.2	2651.7	16.8	24.7
39	KEN-1_11-3e	0.323	0.011	7.243	0.217	0.162	0.002	0.953	1805.9	52.0	2141.8	26.4	2479.2	17.2	31.1
40	KEN-1_12-1m	0.350	0.012	7.516	0.226	0.156	0.002	0.962	1936.1	59.2	2174.9	26.5	2410.4	18.0	22.7
41	KEN-1_12-2e	0.358	0.013	8.472	0.254	0.171	0.002	0.968	1972.0	62.5	2282.9	26.9	2570.2	18.3	27.0
42	KEN-1_12-3m	0.432	0.015	11.111	0.333	0.186	0.002	0.959	2315.0	66.2	2532.5	27.6	2708.7	16.6	17.3
43	KEN-1_12-4e	0.407	0.013	10.409	0.312	0.185	0.002	0.949	2201.9	60.2	2471.8	27.4	2699.7	16.9	21.7
44	KEN-1_12-5e	0.338	0.011	6.965	0.209	0.150	0.002	0.949	1875.0	53.1	2107.0	26.3	2343.3	17.7	23.0
45	KEN-1_13-1m	0.262	0.010	4.675	0.140	0.129	0.002	0.965	1498.2	51.1	1762.8	24.8	2086.8	21.6	31.6
46	KEN-1_13-2e	0.338	0.011	7.622	0.229	0.163	0.002	0.947	1875.7	51.3	2187.5	26.6	2488.8	17.1	28.3
47	KEN-1_13-3m	0.270	0.008	5.012	0.150	0.134	0.001	0.941	1542.5	41.5	1821.4	25.1	2150.8	18.1	31.7
48	KEN-1_13-4e	0.286	0.010	5.888	0.177	0.149	0.002	0.919	1619.9	47.9	1959.5	25.7	2333.3	22.7	34.5
49	KEN-1_13-5e	0.318	0.010	6.653	0.200	0.151	0.002	0.945	1780.0	48.5	2066.4	26.1	2362.2	17.5	28.1
50	KEN-1_14-1m	0.435	0.014	11.082	0.333	0.185	0.002	0.954	2328.0	64.5	2530.0	27.6	2698.8	16.6	16.3
51	KEN-1_15-1m	0.396	0.013	9.996	0.300	0.183	0.002	0.947	2150.4	58.1	2434.4	27.3	2678.9	16.9	23.1
52	KEN-1_16-1m	0.421	0.015	10.814	0.325	0.186	0.002	0.963	2265.0	66.4	2507.3	27.5	2707.1	16.8	19.3
53	KEN-1_16-2e	0.439	0.014	11.310	0.339	0.186	0.002	0.948	2347.2	61.6	2549.0	27.6	2710.8	16.6	16.0
54	KEN-1_17-1m	0.426	0.015	10.866	0.326	0.185	0.002	0.960	2286.3	66.2	2511.7	27.5	2696.1	16.8	18.0
55	KEN-1_18-1m	0.423	0.017	10.265	0.308	0.175	0.002	0.731	2273.6	78.2	2459.0	27.4	2603.6	18.1	15.0
56	KEN-1_19-1m	0.434	0.015	11.159	0.335	0.186	0.002	0.961	2322.0	67.3	2536.5	27.6	2710.7	16.8	17.0
57	KEN-1_20-1m	0.208	0.008	3.653	0.110	0.127	0.001	0.957	1216.4	40.0	1561.1	23.7	2052.7	20.3	44.6
58	KEN-1_21-1m	0.434	0.014	11.177	0.335	0.186	0.002	0.950	2322.5	63.8	2538.0	27.6	2709.1	17.0	17.0
59	KEN-1_21-2e	0.435	0.014	11.208	0.336	0.186	0.002	0.949	2330.2	62.9	2540.6	27.6	2710.4	16.8	16.7
60	KEN-1_22-1m	0.378	0.033	8.994	0.274	0.181	0.033	0.352	2068.9	151.3	2337.4	27.4	2659.6	306.4	25.9

Appendix 9 – Duplicate Geochemical Data

Table 13. Trace-element concentrations in rocks: duplicates of concentrations reported in Table 7.

Analyte Symbol	Rock type	Cu	Ni	Zn	Ag	Pb	Sc	Cr	As	Sb
Detection Limit		*10*	*20*	*30*	*0.5*	*5*	*1*	*20*	*5*	*0.5*
Analysis Method		FUS-MS	FUS-MS	FUS-MS	FUS-MS	FUS-MS	FUS-ICP	FUS-MS	FUS-MS	FUS-MS
6704	grey-green granite	20	< 20	70	< 0.5	37	8	30	< 5	< 0.5
6705	grey-green sediment	< 10	< 20	< 30	< 0.5	7	4	60	< 5	< 0.5
6706	gabbro-diorite	10	50	110	< 0.5	22	22	240	< 5	8.1
6707	dark grey turbidite	20	< 20	40	< 0.5	29	21	90	61	< 0.5
6708	K-fsp granite	< 10	< 20	70	< 0.5	25	9	30	< 5	< 0.5
6709	K-fsp sediment	30	< 20	< 30	< 0.5	48	3	50	< 5	12.3
6610	argillite/siltstone	10	< 20	40	< 0.5	27	27	100	29	0.6
6611	diorite with disseminated pyrrhotite	10	110	120	< 0.5	31	20	390	9	0.4
310	diorite	20	50	70	< 0.5	11	30	310	33	5.9
311A	dark turbidite/trachite (?)	< 10	< 20	< 30	< 0.5	25	6	40	6	< 0.5
311B	light turbidite/aplite (?)	< 10	< 20	60	< 0.5	38	8	70	10	5.5
6710	Unger Lake granite	< 10	< 20	60	< 0.5	39	5	< 20	< 5	< 0.5
6711	Colin Lake megacrystic granite	30	< 20	50	< 0.5	20	5	60	< 5	< 0.5
6712	sheared aplite	< 10	< 20	160	< 0.5	132	24	100	< 5	< 0.5

Appendix 10 – Analytical Method for $^{40}Ar/^{39}Ar$ Geochronology

All four samples, DP67, DP74, DP75B and DP76, consisted of sericite-muscovite phyllite schist. Samples selected for analysis were crushed to granules in the range 0.25–0.50 mm. Several grains were loaded in aluminum foil packets arranged radially in an aluminum canister along with the flux monitor PP-20 hornblende (Hb3gr equivalent), with an apparent age of 1072 Ma (Roddick, 1983). The samples were then irradiated for 120 hours in the research reactor at McMaster University in a fast neutron flux of approximately 3 x 10^{16} neutrons/cm^2.

Laser $^{40}Ar/^{39}Ar$ step-heating analysis was carried out at the Geological Survey of Canada laboratories in Ottawa, Ontario. Upon return from the reactor, samples were split into several aliquots and loaded into individual 1.5 mm diameter holes in a copper planchet. The planchet was then placed in the extraction line and the system evacuated. Heating of individual sample aliquots in steps of increasing temperature was carried out using a Merchantek MIR10 10W CO_2 laser equipped with a 2 mm by 2 mm flat-field lens. The released Ar gas was cleaned over getters for 10 minutes and then analyzed isotopically using the secondary electron multiplier system of a VG3600 gas-source mass spectrometer; details of data collection protocols are in Villeneuve and MacIntyre (1997) and Villeneuve et al. (2000). Error analysis on individual steps followed numerical error analysis routines outlined in Scaillet (2000); error analysis on grouped data followed algebraic methods of Renne et al. (1998).

Neutron-flux gradients throughout the sample canister were evaluated by analyzing the hornblende flux monitors, which were interspersed among the sample packets throughout the sample canister. A linear fit was interpolated against calculated J-factor and sample position. The error in individual J-factor values is conservatively estimated at ±0.6% (at the 2σ level). Because the error associated with the J-factor is systematic and not related to individual analyses, correction for this uncertainty is not applied until calculation of weighted average plateau ages (Roddick, 1988). No significant excess ^{40}Ar was observed in any of the samples, so all regressions are assumed to pass through the $^{40}Ar/^{36}Ar$ value for atmospheric air (295.5).

Appendix 11 – Raw Analytical Data for $^{40}Ar/^{39}Ar$ Geochronology

Table 14. $^{40}Ar/^{39}Ar$ isotope raw analytical data.

Power[a]	Volume ^{39}Ar x10^{-11} cc	$^{36}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{38}Ar/^{39}Ar$	$^{40}Ar/^{39}Ar$	%^{40}Ar ATM	*$^{40}Ar/^{39}Ar$	f_{39}[b] (%)	Apparent Age Ma[c]
DP67 Sericite; J=0.03153850 (Z8803; 0.0000°N 0.0000°E)									
		Aliquot: A							
2.4	6.354	0.0032±0.0003	0.001±0.006	0.011±0.011	44.598±0.099	2.1	43.648±0.124	0.7	1561.8±3.0
2.8	21.213	0.0010±0.0001	0.044±0.003	0.003±0.011	54.262±0.158	0.5	53.970±0.160	2.5	1793.3±3.4
3	102.960	0.0011±0.0001	0.063±0.002	0.003±0.011	56.409±0.197	0.6	56.073±0.197	12	1837.1±4.1
3.3	28.682	0.0007±0.0001	0.034±0.002	0.002±0.011	56.231±0.155	0.4	56.019±0.155	3.3	1836.0±3.2
3.5	54.582	0.0004±0.0000	0.018±0.001	0.002±0.011	56.055±0.110	0.2	55.924±0.110	6.4	1834.0±2.3
3.9	180.994	0.0008±0.0000	0.038±0.002	0.002±0.011	56.093±0.134	0.4	55.871±0.134	21.1	1833.0±2.8
4.2*	162.034	0.0008±0.0000	0.039±0.002	0.002±0.011	55.779±0.160	0.4	55.547±0.161	18.9	1826.2±3.3
4.6*	114.331	0.0008±0.0001	0.048±0.002	0.002±0.011	55.909±0.340	0.4	55.662±0.340	13.3	1828.6±7.0
5*	50.349	0.0003±0.0001	0.018±0.001	0.002±0.011	55.654±0.147	0.2	55.560±0.148	5.9	1826.5±3.1
5.5*	30.955	0.0004±0.0001	0.026±0.002	0.002±0.011	55.664±0.214	0.2	55.539±0.216	3.6	1826.1±4.5
6*	29.417	0.0007±0.0001	0.029±0.002	0.002±0.011	55.732±0.138	0.4	55.534±0.139	3.4	1826.0±2.9
6.5*	10.817	0.0001±0.0001	0.001±0.004	0.002±0.011	55.546±0.102	0	55.525±0.105	1.3	1825.8±2.2
7.5*	29.419	0.0007±0.0000	0.029±0.002	0.002±0.011	55.728±0.145	0.4	55.526±0.146	3.4	1825.8±3.0
13.5*	35.594	0.0005±0.0001	0.026±0.001	0.002±0.011	55.750±0.281	0.3	55.595±0.282	4.2	1827.3±5.8
DP74 Sericite; J=0.03155480 (Z8804; 0.0000°N 0.0000°E)									
		Aliquot: A							
2.4	3.345	0.0054±0.0002	0.008±0.009	0.015±0.011	33.441±0.092	4.7	31.857±0.115	0.3	1255.2±3.3
2.8	10.014	0.0011±0.0001	0.003±0.003	0.005±0.011	31.945±0.067	1	31.610±0.074	0.8	1248.2±2.1
3	41.682	0.0006±0.0001	0.022±0.002	0.003±0.011	40.960±0.127	0.4	40.791±0.128	3.5	1492.5±3.2
3.5	116.060	0.0011±0.0001	0.048±0.003	0.002±0.011	51.334±0.197	0.6	51.007±0.198	9.6	1730.4±4.3
3.9	251.555	0.0006±0.0000	0.025±0.001	0.002±0.011	54.906±0.120	0.3	54.723±0.120	20.9	1809.7±2.5
4.2	207.806	0.0006±0.0001	0.027±0.002	0.002±0.011	54.868±0.141	0.3	54.699±0.142	17.2	1809.2±3.0
4.6	103.191	0.0012±0.0001	0.063±0.003	0.002±0.011	54.726±0.264	0.7	54.359±0.264	8.6	1802.1±5.5
5	63.240	0.0002±0.0000	0.015±0.001	0.002±0.011	54.524±0.108	0.1	54.459±0.108	5.3	1804.2±2.3
5.5	55.934	0.0004±0.0000	0.015±0.002	0.002±0.011	54.771±0.107	0.2	54.647±0.107	4.6	1808.1±2.2
6*	109.558	0.0010±0.0001	0.056±0.002	0.001±0.011	55.501±0.195	0.5	55.198±0.195	9.1	1819.6±4.10
6.5*	66.974	0.0003±0.0000	0.013±0.001	0.002±0.011	55.344±0.093	0.2	55.249±0.094	5.6	1820.7±2.0
7.5*	119.829	0.0011±0.0001	0.050±0.001	0.002±0.011	55.567±0.250	0.6	55.245±0.250	9.9	1820.6±5.2
13.5*	56.449	0.0003±0.0001	0.017±0.002	0.002±0.011	55.308±0.173	0.2	55.213±0.173	4.7	1819.9±3.6
DP75B Sericite; J=0.03196790 (Z8805; 0.0000°N 0.0000°E)									
		Aliquot: A							
2.4	3.986	0.0043±0.0003	0.003±0.010	0.013±0.011	30.789±0.133	4.1	29.514±0.153	0.8	1198.8±4.5
2.8	5.770	0.0014±0.0002	0.005±0.006	0.006±0.011	46.336±0.155	0.9	45.913±0.161	1.2	1629.7±3.8
3	11.495	0.0005±0.0001	0.001±0.003	0.003±0.011	49.944±0.088	0.3	49.798±0.091	2.3	1718.2±2.0
3.5	43.996	0.0007±0.0000	0.021±0.001	0.002±0.011	54.452±0.112	0.4	54.259±0.112	8.8	1814.9±2.4
3.9*	153.693	0.0007±0.0001	0.041±0.001	0.001±0.011	55.181±0.230	0.4	54.972±0.231	30.9	1829.8±4.8
4.2*	52.531	0.0004±0.0000	0.018±0.001	0.002±0.011	55.068±0.173	0.2	54.938±0.173	10.6	1829.1±3.6
4.6*	53.883	0.0004±0.0000	0.016±0.001	0.002±0.011	54.967±0.146	0.2	54.856±0.146	10.8	1827.4±3.1
5*	34.639	0.0005±0.0001	0.029±0.003	0.002±0.011	55.004±0.176	0.3	54.849±0.179	7	1827.3±3.8
5.5*	35.407	0.0005±0.0001	0.026±0.001	0.002±0.011	55.092±0.125	0.3	54.953±0.126	7.1	1829.4±2.6
6*	39.413	0.0004±0.0001	0.027±0.002	0.001±0.011	55.123±0.146	0.2	55.010±0.147	7.9	1830.6±3.1
6.5*	34.899	0.0006±0.0001	0.026±0.001	0.001±0.011	55.117±0.141	0.3	54.942±0.143	7	1829.2±3.0
13.5*	27.870	0.0008±0.0000	0.038±0.002	0.002±0.011	55.323±0.185	0.4	55.087±0.185	5.6	1832.2±3.9
DP76 Sericite; J=0.03199590 (Z8806; 0.0000°N 0.0000°E)									
		Aliquot: A							
2.4	3.116	0.0035±0.0002	0.001±0.012	0.018±0.011	28.418±0.155	3.6	27.395±0.167	0.5	1135.5±5.1
2.8	4.936	0.0009±0.0001	0.008±0.010	0.005±0.011	48.278±0.123	0.6	48.001±0.129	0.8	1678.7±3.0
3	7.980	0.0003±0.0001	0.003±0.005	0.003±0.011	54.530±0.112	0.2	54.446±0.117	1.3	1819.8±2.5
3.5	33.763	0.0007±0.0000	0.029±0.002	0.002±0.011	56.552±0.175	0.4	56.353±0.175	5.6	1859.5±3.6
3.9	45.930	0.0005±0.0000	0.019±0.001	0.002±0.011	56.412±0.114	0.3	56.252±0.114	7.6	1857.4±2.4
4.2	137.651	0.0009±0.0001	0.047±0.002	0.002±0.011	55.647±0.255	0.5	55.371±0.255	22.9	1839.2±5.3
4.6*	61.840	0.0003±0.0000	0.016±0.001	0.002±0.011	55.571±0.093	0.2	55.472±0.093	10.3	1841.3±2.0
5*	37.993	0.0004±0.0001	0.027±0.001	0.002±0.011	55.512±0.131	0.2	55.391±0.133	6.3	1839.6±2.8
5.5*	24.696	0.0006±0.0001	0.032±0.005	0.002±0.011	54.954±0.154	0.3	54.767±0.157	4.1	1826.5±3.3
6*	35.672	0.0004±0.0001	0.028±0.002	0.001±0.011	55.467±0.166	0.2	55.359±0.169	5.9	1838.9±3.5
6.5*	50.403	0.0004±0.0000	0.018±0.002	0.002±0.011	55.493±0.162	0.2	55.368±0.162	8.4	1839.1±3.4
7.5*	103.217	0.0011±0.0001	0.061±0.002	0.001±0.011	55.655±0.231	0.6	55.334±0.230	17.2	1838.4±4.8
13.5*	54.133	0.0004±0.0000	0.019±0.001	0.001±0.011	55.380±0.127	0.2	55.265±0.127	9	1837.0±2.6

Note:
a: As measured by laser in % of full nominal power (10W).
b: Fraction ^{39}Ar as percent of total run.
c: Errors are analytical only and do not reflect error in irradiation parameter J.
Nominal J, referenced to PP-20 Hornblende (Hb3gr equivalent = 1072 Ma) (Roddick, 1983).
All uncertainties quoted at 2s level.
* Steps included in plateau-age calculation.