Canadian Journal of Earth Sciences

Petrogenesis and mode of emplacement of a Neoarchean tonalite-trondhjemite-diorite (TTD) suite: the Eau Jaune Complex, Abitibi greenstone belt

Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2021-0016.R3
Manuscript Type:	Article
Date Submitted by the Author:	28-Jul-2021
Complete List of Authors:	Kieffer, Marie; Université du Québec à Chicoutimi, Applied Sciences Mathieu, Lucie; UQAC - Université du Québec à Chicoutimi, 555 Bd de l'université Bedeaux, Pierre; Université du Québec à Chicoutimi, Sciences de la Terre Gaboury, Damien; Université du Québec à Chicoutimi, Applied Sciences Hamilton, Michael; University of Toronto, Department of Earth Sciences
Keyword:	Abitibi greenstone belt, Chibougamau, Eau Jaune Complex, Neoarchean magmatism, TTD suite, TTG suite
Is the invited manuscript for consideration in a Special Issue? :	Not applicable (regular submission)

SCHOLARONE™
Manuscripts

1 Petrogenesis and mode of emplacement of a Neoarchean tonalite-trondhjemite-diorite (TTD)
2 suite: the Eau Jaune Complex, Abitibi greenstone belt

3 Marie A. Kieffer^{1*}, Lucie Mathieu¹, Pierre Bedeaux¹, Damien Gaboury¹ and Michael A.
4 Hamilton²

5 ¹**M. A. Kieffer, L. Mathieu, P. Bedeaux, D. Gaboury.** Département des Sciences Appliquées,
6 Université du Québec à Chicoutimi, QC, Canada G7H 2B1

7 ²**M. A. Hamilton.** Jack Satterly Geochronology Laboratory, Department of Earth Sciences,
8 University of Toronto, 22 Russell St., Toronto, ON, Canada M5S 3B1

9 ***Corresponding author:** Marie A. Kieffer (e-mail: marie.kieffer1@uqac.ca)

Abstract

Magmatism during the maturation phase of Archean greenstone belts produced voluminous tonalite-trondhjemite-granodiorite (TTG) suites, as well as a lesser amount of tonalite-trondhjemite-diorite (TTD) suites. Such TTD suites have recently been recognized in the Archean Abitibi greenstone belt, on the southern flank of the Superior Craton, Canada, but their source(s), differentiation processes and depths of emplacement remain poorly constrained. The Neoarchean Eau Jaune Complex (EJC) lies in the northeastern corner of the Abitibi greenstone belt and represents one of the most voluminous tonalite-dominated and diorite-bearing intrusive suites of the Chibougamau region. This TTD suite comprises six intrusive phases with distinct petrology and chemistry. All units were emplaced as laccolith-like intrusions injected along discontinuities within the volcanic succession at ca. 2724 Ma (U-Pb zircon dating), during the synvolcanic interval (i.e., construction and maturation phase), at a depth of approximately 7–8 km. The most HREE-depleted phases (granodiorite, tonalite and trondhjemite) correspond to magmas that fractionated amphibole and were likely produced by partial melting of a garnet- and titanate-bearing amphibolite, akin to TTG magmas. The least HREE-depleted phases are dioritic in composition and correspond to mantle-derived magmas that may have interacted with TTG melts. This indicates interaction between coeval mantle-derived and crustal melts during the maturation phase of the Abitibi greenstone belt. Models formulated to address the geodynamic evolution of greenstone belts must account for the coeval production of basalt-derived (TTG suites) and mantle-derived (tholeiitic magmatism) melts occasionally interacting to form TTD suites.

Keywords: Abitibi greenstone belt; Chibougamau; Eau Jaune Complex; igneous petrology; Neoarchean magmatism; structural geology; TTD suite; TTG suite; tonalite

32 **Résumé**

33 Lors de la phase de maturation des ceintures de roches vertes archéennes, l'activité magmatique a
34 produit de volumineuses suites de type tonalite-trondhjémite-granodiorite (TTG) ainsi que de
35 moindres quantités de suites de type tonalite-trondhjémite-diorite (TTD). Ces suites TTD ont
36 récemment été reconnues dans la ceinture de roches vertes archéenne de l'Abitibi, située dans la
37 partie sud du Craton de Supérieur, Canada. Or, leur(s) source(s), leurs processus de
38 différenciation ainsi que leurs profondeurs de mise en place restent à ce jour peu contraints. Le
39 Complexe d'Eau Jaune (CEJ) est situé à l'extrémité nord-est de la ceinture de roches vertes de
40 l'Abitibi. Cette intrusion néoarchéenne représente l'une des plus volumineuses intrusions
41 tonalitiques à composante dioritique de la région de Chibougamau. Cette suite TTD est constituée
42 de six phases intrusives de pétrologies et de compositions chimiques distinctes. Les phases se
43 sont mises en place le long de discontinuités dans la séquence volcanique, sous forme
44 d'intrusions de type laccolithe, à environ 2724 Ma (datation U-Pb sur zircons), soit durant
45 l'intervalle synvolcanique (correspondant aux phases de construction et de maturation), et à une
46 profondeur de 7 à 8 km. Les phases les plus appauvries en HREE (granodiorite, tonalite et
47 trondhjémite) correspondent à des magmas apparentés aux magmas de type TTG, qui ont
48 fractionné des amphiboles et ont probablement été formés par la fusion partielle d'une
49 amphibolite à grenats et titanates. Les phases les moins appauvries en HREE sont de composition
50 dioritique et correspondent à des magmas dérivés du manteau pouvant avoir interagi avec des
51 magmas de type TTG. Cela indique une interaction entre des magmas dérivés du manteau et de la
52 croûte au cours de la phase de maturation de la ceinture de roches vertes de l'Abitibi. Les
53 modèles géodynamiques expliquant l'évolution des ceintures de roches vertes doivent donc
54 inclure la génération contemporaine de magmas dérivés de basaltes (suites de type TTG) et du

manteau (magmatisme tholéiitique), qui peuvent occasionnellement interagir pour former des suites de type TTD.

Mots-clefs: ceinture de roches vertes de l'Abitibi; Chibougamau; Complexe d'Eau Jaune; géologie structurale; magmatisme Néoarchéen; pétrologie ignée; suite de type TTD; suite de type TTG; tonalite

1. Introduction

Tonalite-trondhjemite-granodiorite suites (TTG) are the main component of intrusive Archean rocks and can represent up to 50–70 vol.% of a craton (Condie 1993; Moyen and Stevens 2006; Hoffmann et al. 2011; Laurent et al. 2014). In Archean greenstone belts, TTG suites are voluminous intrusions emplaced during the main magmatic stage (maturation phase), commonly referred to as the synvolcanic period (Feng and Kerrich 1992; Condie 1993; Smithies 2000; Whalen et al. 2002; Moyen and Martin 2012; Mathieu et al. 2020a). These TTG suites are derived from the deep partial melting of a hydrated and LREE-enriched (light rare-earth elements) basalt (Martin et al. 2005; Moyen and Martin 2012), and represent the cornerstone of most geodynamic models (Moyen and Martin 2012). For instance, the Al_2O_3 , HREE (heavy rare-earth elements) and Sr content of TTG suites has been related to the depth of partial melting (Moyen and Martin 2012). Their Mg content has been used to infer possible interactions between the melt and mantle rocks and to argue for a steepening subduction angle through time (Martin et al. 2005; Martin et al. 2014).

In Archean greenstone belts, it is generally accepted that TTG magmas originated from the partial melting of LREE-enriched basalts (Martin et al. 2005; Moyen and Martin 2012) under a range of P-T conditions and different degrees of melting (F), with melts becoming more mafic (e.g.

dioritic) with increasing F (Moyen and Martin 2012). In addition, the chemical heterogeneity of the basaltic source (Adam et al. 2012; Nagel et al. 2012; Hastie and Fitton 2019) may affect the chemistry of the TTG suites (Kendrick and Yakymchuk 2020). It has been recently recognised that tonalite-trondhjemite-diorite (TTD) plutonic suites, which correspond to tonalite-dominated plutons containing a significant diorite component, are coeval with TTG magmatism (Mathieu and Racicot 2019; Mathieu et al. 2020a). However, TTD suites are less abundant (e.g. ~12% of the aerial surface covered by synvolcanic intrusions in the Abitibi greenstone belt; Mathieu et al. 2020a) and remain more poorly documented than TTG suites. Most studies refer to TTD intrusions as diorite-bearing TTG suites (Katz et al. 2017; Liou and Guo 2019; Moyen 2019) with a source that may include a significant mantle component and may correspond to TTG hybridized by mantle-derived melts (Mathieu et al. 2020a). As with all magmatic systems, fractional crystallization and other processes (e.g., contamination, magma mixing) can further modify the chemistry of TTG and TTD suites. Nonetheless, few studies have evaluated the extent of these processes (Liou and Guo 2019; Laurent et al. 2020). Magma source having a mantle component, as well as basalts, are involved in the formation of Archean cratons (Mole et al. 2013). TTD suites also indicate that magmas derived from mantle and basaltic sources are coeval and may have interacted during the synvolcanic period, which should be taken into account by geodynamic models proposed for the Archean. Insights into geodynamic processes could thus be gained from a better understanding of the petrogenesis of individual TTD intrusions.

The geodynamic evolution of the Archean crust, including the Abitibi Subprovince, remains controversial. The main geodynamic models include: (1) subduction and accretion of associated volcanic arcs (Ludden et al. 1984; Ludden and Hubert 1986; Kimura et al. 1993; Calvert et al. 1995; Calvert and Ludden 1999; Wyman et al. 2002), as was suggested for the Chibougamau area

(Polat et al. 2018); (2) mantle plume-driven crustal reworking activity with associated arc-type volcanism (Rey et al. 2003; Gerya et al. 2015), which could be followed by subduction (Ayer et al. 2002; Wyman et al. 2002); (3) delamination, including aspects of a subcretion model, implying horizontal movements followed by the imbrication of a crust that is too thick to subduct (Bédard et al. 2003; Bédard et al. 2013; Bédard 2018) and; (4) rifting of a cratonic nucleus (Mathieu et al. 2020b; Mole et al. 2021), to produce a 30-40 km thick mafic crust by the partial melting of a hot ambient mantle (Herzberg et al. 2010); however, the study of Roman and Arndt (2020) show that such a process is not plausible on Earth. Hypothetical models (3) and (4) consider that partial melting occurred at the base of the thickened crust (Van Kranendonk et al. 2015). It is beyond the scope of this paper to discuss which of the above-cited geodynamic models is the more favourable to TTD formation: all these models are generally compatible with coexistence of mantle- and basalt-derived magmas. Rather, this article aims to provide new geochemical data on the source and petrogenesis of a voluminous TTD intrusion of the Abitibi greenstone belt. The results of this study can provide new insights for future models of Archean geodynamics.

In the Superior province, some TTG and TTD suites are genetically related to magmatic-hydrothermal, porphyry-style, Au and Cu-Au mineralization, such as the Cu-Au Central Camp associated with the voluminous Chibougamau pluton (Fig. 1; Mathieu and Racicot 2019) and the Cu-Au Troilus deposit (Fraser 1993) north of the Chibougamau area, as well as the Au-(Cu) Côté Gold deposit related to the Chester Intrusive Complex (Katz et al. 2017) in the southwestern part of the Abitibi greenstone belt. In addition, some TTG suites such as the Tilly granitoid (La Grande Subprovince, James Bay area; Chapon et al. 2010) and those within the MOP-II project (Chibougamau area, Fig. 1; Lépine 2009), are associated with Mo-bearing mineralization. In

porphyry systems, it is commonly accepted that metals (e.g., Cu, Au, Mo) and volatiles (e.g., CO₂, H₂O, Cl) are concentrated in the residual melt by differentiation processes (Sillitoe 2010; Richards 2018). Intrusions associated with porphyry deposits generally lie 5–10 km below the surface, whereas the exsolved and volatile-rich magmas carrying metals form porphyry stocks at shallower crustal levels (<4 km depth; Sillitoe 2010; Richards 2018). Thus, emplacement depths and differentiation processes are of paramount importance for concentrating elements of interest. However, these and other constraints are lacking for magmatic-hydrothermal mineralizing systems associated with TTG and TTD intrusive suites. This study aims to gain a better understanding of the differentiation and emplacement depth of a tonalite-dominated and diorite-bearing (TTD) intrusive suite, which could be of use to other studies whose aims are to better comprehend Archean magmatic-hydrothermal mineralizing systems.

This study focuses on the northeastern Abitibi greenstone belt, southern Superior Province, where numerous TTG and TTD suites and several magmatic-hydrothermal systems have been identified (Pilote et al. 1998; Leclerc et al. 2017; Mathieu et al. 2020a). One of the largest intrusive suites in this region is the Eau Jaune Complex (EJC; Fig. 1). It hosts a Mo occurrence (Desgagné 2007; Grenier and Sansfaçon 2011) and has been interpreted previously as a polyphase TTG suite having a diorite component (Tait 1992). Early studies (Tait et al. 1986; Tait 1992), however, were insufficiently detailed to evaluate the petrogenesis of this intrusion. Our study aims to determine the petrogenetic evolution and mode of emplacement of the EJC through the acquisition of new and comprehensive field, petrological, and geochemical data. This has been made possible as part of the Metal Earth project and its accompanying logistical support, in parallel with geophysical studies and regional mapping conducted by the Metal Earth project in the Chibougamau area.

2. Geological setting

The EJC lies approximately 45 km southwest of the town of Chibougamau, at the eastern extremity of the Neoproterozoic Chibougamau-Matagami greenstone belt, corresponding to the northeastern Abitibi Subprovince (Fig. 1). As with most greenstone belts, the tectono-magmatic evolution of the study area can be summarized in two principal stages: (1) a synvolcanic period during which most of the volcanic and intrusive rocks were emplaced (2800–2721 Ma construction and maturation period; Ludden et al. 1984; Leclerc et al. 2011; Mathieu et al. 2020b) and; (2) a syntectonic interval corresponding to the main stage of deformation, erosion, and sedimentation (2700–2680 Ma; Daigneault and Allard 1990; Chown et al. 1992; Daigneault 1996; Leclerc et al. 2015; Mathieu et al. 2020b). In the Chibougamau area, deposition of most sedimentary rocks, i.e. the 2704–2692 Ma Opémisca Group, located approximately 10 km north of the EJC (Fig. 1) (David et al. 2007), and coeval minor alkaline magmatism, occurred during the syntectonic period (Leclerc et al. 2012). The most voluminous intrusions in the study area include the Chibougamau pluton (2718–2715 Ma; Pilote et al. 1997) and the Lac Doré Complex (ca. 2728 Ma; Mortensen 1993a), both emplaced during the synvolcanic period and lying approximately ≥ 20 km north-east of the EJC (Fig. 1). This study focuses on an intrusive episode of the synvolcanic period during which older volcanic rocks (Chrissie and Des Vents formations, see below) and volcanic cycles 1 and 2 of the Roy Group (Mortensen 1993a) also formed.

The EJC is in contact with the volcanic rocks of the Des Vents, Chrissie and Obatogamau formations (Fig. 2). The 2798.7 ± 0.7 Ma (Davis et al. 2014) Des Vents Formation represents a calc-alkaline felsic volcanic centre (Potvin 1991; Leclerc et al. 2017), while the Chrissie Formation consists of tholeiitic basaltic lava flows erupted at 2791.4 ± 3.7 Ma (David et al. 2011). The Obatogamau Formation (stratigraphic base of the Roy Group) is dominated by feldspar megacryst-bearing basalt-andesite lava flows. This poorly-dated unit was interpreted to

have formed before the ca. 2730–2727 Ma Waconichi Formation (Leclerc et al. 2011) or, in a more recent study, at ca. 2726 Ma (Boucher et al. 2020) on the basis of new U-Pb zircon dating of a felsic layer within the Obatogamau Formation. The Des Vents and Chrissie formations are part of an older volcanic cycle that preceded the emplacement of the Obatogamau and Waconichi formations of volcanic cycle 1 (Leclerc et al. 2017), while the younger Bruneau and Blondeau formations belong to volcanic cycle 2 dated from 2724.4 ± 1.2 Ma (Davis et al. 2014) to $<2721 \pm 3$ Ma (Leclerc et al. 2012; Leclerc et al. 2017).

The EJC is also in contact, to the west and north, with the tonalitic rocks of the Lapparent suite (Racicot et al. 1984). This extensive TTG suite is dominated by gneissic tonalitic rocks and was previously considered as the basement to the Chibougamau supracrustal sequence (Mueller et al. 1989; Chown and Mueller 1992); however, U–Pb zircon dating has demonstrated igneous emplacement ages of 2713.4 ± 2.5 Ma and 2711.6 ± 1.2 Ma (Mortensen 1993b; Chown et al. 2002), indicating that the Lapparent suite was coeval with cycle 2 volcanism.

The EJC has been interpreted as a TTG suite comprising a minor quartz-diorite phase and three tonalite phases (Tait 1992). However, the revised map presented here (Fig. 2) shows that the EJC comprises only three main intrusive phases: (1) EJC1 (2718.6 ± 5.5 Ma; David 2018), formed of quartz diorite, diorite, and minor tonalite; (2) EJC2, mainly consisting of tonalite, as well as tonalitic and dioritic gneisses; and (3) EJC3, dominated by trondhjemite. Racicot et al. (1984) and Midra et al. (1992) interpret that the EJC intrudes the Lapparent suite. However, this interpretation is not in agreement with available radiogenic data indicating that the EJC is, at least in part, older (2718.6 ± 5.4 Ma; David 2018) than the two samples from the Lapparent suite which were dated at ca. 2713 Ma and 2712 Ma. On the basis of field relationships, Tait (1992) also proposed that a minor part of the EJC was synvolcanic while most of the intrusion was

emplaced during the syntectonic period. A compilation by Faure (2012) places intrusion of the EJC at the transition between the synvolcanic and syntectonic periods (i.e., an intrusive event occurring at ca. 2710 Ma). Other studies, however, consider that the EJC formed exclusively during the synvolcanic period (Chown et al. 2002), a conclusion supported by the published geochronological data for phase EJC1 (see above).

The Chibougamau area was affected by several phases of deformation according to Baker (1980), Chown et al. (1992), Daigneault (1996), Daigneault and Allard (1990) and Leclerc et al. (2015). Most of the strain observed in the field can be related to a major Neoproterozoic north–south ductile shortening event, which is generally referred to as deformation 2 (D_2) in the literature, in contrast to D_1 , which is a minor, possibly local, pre-ductile deformation event (Dimroth et al. 1986; Daigneault et al. 1990). At the regional scale, the D_2 event formed subvertical east–striking folds and subvertical east- to southeast-trending faults. The most prominent structural feature is a broadly subvertical east-striking penetrative regional foliation recorded by the bulk of metasedimentary and metavolcanic rocks and, to some extent, by intrusive rocks. A D_3 event, related to northwest–southeast shortening during the waning stage of the D_2 event, reactivated the D_2 subvertical east–west-striking faults as dextral strike-slip faults (Daigneault et al. 1990). Most of the deformation (D_2 and D_3 events, minor D_1) occurred between 2707 Ma and 2692 Ma during the syntectonic period as defined by Dimroth et al. (1986) and Daigneault and Allard (1990). These first three deformation events occurred during the terrane assembly (or regional crustal shortening) leading to cratonisation and probably associated with the Kenoran orogeny (Dallmeyer et al. 1975). According to Baker (1980) and Kline (1985), a D_4 event, related to the Grenville orogeny (ca. 1.1 Ga), is responsible for amphibolite-grade metamorphism near the Grenville front, approximately 50 km east of the study area (Fig. 1), and forming north-northeast-

striking faults. Chown and Mueller (1992) inferred that the Lapparent tonalitic suite is flanked to the north and south by east–west-trending reverse faults and that it corresponds to a horst-like structure. These faults were interpreted as basement faults reactivated during deformation (Chown and Mueller 1992). Most of the Chibougamau area is overprinted by greenschist-facies regional metamorphism (Dimroth et al. 1983; Powell et al. 1995) related to the main deformation event (D_2), such that metamorphic minerals are aligned along foliation planes in volcanic and sedimentary rocks (Mueller et al. 1989; Leclerc et al. 2017). Contact metamorphism (amphibolite and hornfels facies) associated with synvolcanic intrusions is rarely documented and may have been obliterated by regional metamorphism (Daigneault et al. 1990; Faure 2012). The lack of a contact metamorphic aureole around the EJC is consistent with a synvolcanic origin (Tait et al. 1986; Tait 1992; Legault 2003; Faure 2012).

3. Sampling procedures and analytical techniques

3.1. Sampling and preparation

Documentation of 155 outcrops was accomplished during the summers 2018 and 2019 (Fig. 2). This included detailed mapping of five trenches of the Moly-Desgagné showing (previously excavated for exploration purposes) using a 1 m × 1 m grid (Fig. 3). It also included sampling of the easternmost part of the Lapparent tonalitic suite to redefine the western extent of the EJC (Fig. 2), based upon textures observed in the field as well as mineralogy and geochemistry. Sample locations and lithologies are provided in Supplementary Material S1. Structural data have been compiled from the fieldwork-based studies of Tait (1992), Midra et al. (1992), and Leclerc et al. (2015), with additional data acquired as part of this study. To characterize the petrology and

geochemistry of the EJC intrusive phases and its xenoliths, we collected 42 samples, each approximately 1–2 kg, and examined 40 thin sections.

3.2. Petrographic analyses

Petrographic observations were performed on 30 standard thin sections and 10 polished thin sections using an Olympus petrographic microscope located at the Université du Québec à Chicoutimi (UQAC). Twenty-seven samples were subjected to both whole-rock and thin-section analyses. We documented textures, and modal proportions (in vol.%, e.g. percentage of each phase visible on the surface of the thin section) were visually estimated for each sample (Supplementary Material S1). Plagioclase composition was estimated using the Michel-Lévy twin extinction angle method (Kerr 1959; Nesse 2000). Microprobe analysis performed on representative plagioclases confirmed the accuracy of the optical estimates (Supplementary Material S3). Petrographic observations were compared to CIPW normative calculations (Cross et al. 1902; Hutchison 1975) determined from whole-rock chemical compositions.

3.3. Whole-rock lithogeochemistry

Whole-rock analyses were performed by ALS Chemex Laboratory (Vancouver, Canada) on 42 selected samples. Samples were dried and finely crushed until more than 70% was reduced to a powder having <2 mm grain size. A split of up to 250 g of this powder was pulverized using a mild steel grinder until more than 85% of the powder passed through a 75 µm screen. Lithium-borate fusion and acid digestion (minor elements), as well as four-acid and Aqua Regia digestion (metallic elements), were performed prior to analysis (Supplementary Material S2). Major, minor, and metallic elements were determined using inductively coupled plasma–atomic emission spectrometry (ICP-AES); lower detection limits are 0.01 wt% for major and minor

elements and 0.001–10 ppm for metals. Rare-earth elements (REE) were analysed using inductively coupled plasma–mass spectrometry (ICP-MS), with lower detection limits of 0.01–2 ppm. Five standards (ORCA-1) and three duplicates were added into the felsic samples sequence, and four standards (LK-NIP-1) and two duplicates were added into the mafic samples sequence. The geochemical data from 31 sites from the SIGÉOM (2019) data set, acquired between 1988 and 2015, was also integrated into this study. Detection limits and complete chemical results are available in Supplementary Material S2.

Normative mineral proportions were estimated using a normative calculation derived from the CIPW method (Cross et al. 1902). The calculation is modified to consider excess water throughout the calculation (modified CIPW norm “biotite + hornblende”; Hutchison 1975). The total alkali silica (TAS) diagram of Middlemost (1985) was used to estimate the $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio for each sample, a step required for the CIPW calculation.

3.4. Mineral chemistry

Amphibole and feldspar chemistry were determined by electron microprobe at McGill University (Montreal) using a SXFiveFE field-emission electron probe micro-analyser. The operating conditions included a 20 kv accelerating voltage, a 20 nA beam current, and a 5 μm spot size. Calibration relied on natural and synthetic standards, and the associated error is $\pm 1\%$ (measured value divided by the reference material value). We used the Locock (2014) spreadsheet (version 1.9.3, 27 September 2017) to estimate amphibole cation site occupancies on the basis of 23 oxygen cations and classified the occupancies according to the International Mineralogical Association Commission on New Minerals Nomenclature and Classification (IMA–CNMNC) classification (Hawthorne et al. 2012). We based our estimates of feldspar cation site occupancies

on eight oxygen cations and classified these occupancies using the method of Greenwood and Earnshaw (1984) (Supplementary Material S3).

3.5. U–Pb geochronology

Given one U–Pb age of 2718.6 ± 5.5 Ma (David 2018) for the quartz-diorite of EJC1, a tonalite (sample 18UCD0137) and a trondhjemite (sample 18UCD0043) from the EJC2 and EJC3 phases, respectively, were dated to obtain a more complete estimate of the timing of emplacement of the broader EJC. The two samples (approximately 7 kg each) were processed and analysed at the Jack Satterly Geochronology Laboratory (University of Toronto). The samples underwent conventional jaw crushing followed by grinding using a Bico™ disk mill. The initial separation of heavy minerals was performed through multiple runs using a riffled Wilfley shaking-water table. The heavy-mineral concentrate was then magnetically separated with a Frantz isodynamic separator. Zircon grains were recovered by heavy liquid (methylene iodide) separation. The best quality, least-cracked, core- and inclusion-free grains of zircon were handpicked under alcohol using a binocular microscope.

Selected zircon separates underwent conventional U–Pb dating by isotope dilution thermal ionization mass spectrometry (ID-TIMS) following a chemical abrasion pre-treatment (CA, modified after Mattinson 2005). Grains were dissolved in a concentrated hydrofluoric (HF) acid at 200 °C for several hours. Pb and U were then isolated by an anion exchange procedure and dried. The purified sample-spike solutions were loaded onto outgassed Re filaments with a silica gel (Gerstenberger and Haase 1997), and isotopic ratios were measured with a VG354 mass spectrometer using a Daly collector in pulse-counting mode. Detailed CA-ID-TIMS U–Pb analytical procedure and results can be found in Supplementary Material S4.

4. Results

4.1. Field relationships

New, detailed fieldwork indicates that the EJC comprises six intrusive phases. Phases EJC1, EJC2, and EJC3, as defined by SIGÉOM (2018), can be subdivided on the basis of distinct colour, grain size, and mineral proportions. All phases enclose xenoliths of mafic rocks and are cut by minor aplite, pegmatite, and other medium-grained felsic dykes. The main intrusive phases have distinct mineralogical compositions and textures, and they correspond to three types of diorite (phases 1a, 1b, 3b), two types of tonalite (phases 2a, 2b), and one trondhjemite (3a) phase. The currently mapped outlines of phases 1a-b and 3a-b remain almost unchanged, the new subdivisions fitting within the respective borders of EJC1 and EJC3 as mapped by the MERN (Fig. 2). However, our new field work also suggests that the EJC2 phase (2a and 2b in this study) extends roughly 15 km further to the west than previously thought (Fig. 2), a conclusion supported by a combination of new textural, mineralogical and geochemical data.

Where observed, contacts between intrusive phases are sharp and well-defined. Enclaves of granodiorite 1a are observed in quartz diorite-tonalite 1b on one of the Moly-Desgagné outcrops (Fig. 3). Several outcrops display enclaves of granodiorite 1a and quartz diorite-tonalite 1b enclosed within tonalite 2a or biotite tonalite 2b. Enclaves of tonalite 2a and biotite tonalite 2b were also observed within trondhjemite 3a and amphibole-rich diorite 3b. We observed no contact relationships between tonalite 2a and biotite tonalite 2b, and between trondhjemite 3a and amphibole-rich diorite 3b. Thus, on the basis of these observations, we can deduce the following emplacement sequence: (1) granodiorite 1a formed first, followed by (2) quartz diorite-tonalite 1b, (3) tonalites 2a and 2b, and, lastly, (4) trondhjemite 3a and amphibole-rich diorite 3b.

The six intrusive phases host xenoliths of mafic rocks, which are more abundant along the outer margin of each intrusive phase, as well as near the contact between the EJC and the volcanic rocks that host the complex. Some of the mafic xenoliths contain plagioclase megacrysts, which are similar to those that characterize the Obatogamau Formation. Thus, mafic xenoliths likely correspond to wallrock fragments of the volcanic rocks found in the vicinity of the EJC. The xenoliths are amphibole-rich and thin- to coarse-grained. Grain coarsening is likely a result of amphibolite-grade contact metamorphism induced by the emplacement of the EJC.

4.2. Mineralogy of the main intrusive phases

Granodiorite 1a is only observed at the Moly-Desgagné showing (Figs. 2, 3). This medium-grained (2-5 mm) unit is composed mainly of quartz (30–40 vol.%) and altered (sericitized and epidotized) feldspars (30–50 vol.%), with a minor amount of mafic constituents (<5 vol.% biotite + amphibole; Figs. 4a, 4b). Quartz diorite-tonalite 1b is mainly observed in the eastern part of the EJC (Fig. 2). It contains altered feldspars (30–50 vol.%), quartz (9–30 vol.%), and up to 10 vol.% green hornblende (Figs. 4c, 4d). Quartz diorite-tonalite 1b is generally fine-grained (<1 mm), but coarse-grained equivalents are observed on the Moly-Desgagné showing. The composition of plagioclases in granodiorite 1a and quartz diorite-tonalite 1b ranges from andesine to labradorite.

Tonalites 2a and 2b (EJC2) are located in the western part of the EJC (Fig. 2). Tonalite 2a displays a characteristic porphyritic texture with 1–2 mm long feldspar grains in a fine-grained (<0.05 mm) quartz-rich matrix (Figs. 4e, 4f). Tonalite 2a contains 35 vol.% quartz and 30–50 vol.% feldspars (some samples include euhedral microcline). Biotite tonalite 2b locally displays quartz-feldspar miarolitic cavities. It also commonly contains minor amounts of white micas, likely formed as a result of local hydration. Perthitic textures and myrmekites are commonly

346 observed in thin section. Biotite tonalite 2b is medium-grained (1–5 mm) (Figs. 5a, 5b),
347 sometimes contains porphyritic feldspar grains (5 mm), and is composed of quartz (20–30 vol.%)
348 and feldspars (50 vol.%), with plagioclase being more abundant than alkali feldspar (<10 vol.%).
349 This unit 2b also commonly contains biotite clusters (up to 3 mm thick, 5–10 vol.%) and green
350 hornblende (<5 vol.%) intergrown with feldspars.

351 Intrusive phases located in the central part of the EJC (EJC3 phase) (Fig. 2) correspond to
352 trondhjemite 3a and amphibole-rich diorite 3b. Grain size and modal mineral proportions are
353 spatially more homogeneous in these phases than in the previously described units. Trondhjemite
354 3a is equigranular and medium-grained (~3 mm) (Fig. 5c). It typically contains less than 5-10
355 vol.% biotite partially replaced by chlorite and <7 vol.% euhedral green hornblende.
356 Trondhjemite 3a is quartz-rich (30–40 vol.%), contains 45 vol.% plagioclase and less than 5
357 vol.% alkali feldspar (including microcline). Feldspars are euhedral with well-defined grain
358 boundaries and co-crystallized with green hornblende. Both minerals commonly display
359 oscillatory zoning (Fig. 5d). Amphibole-rich diorite 3b is medium-grained (0.5–4 mm) and is
360 characterized by abundant and ≤ 4 mm long amphibole laths (30–35 vol.%) (Fig. 5e). Amphibole-
361 rich diorite 3b also contains 5–10 vol.% quartz and <5 vol.% biotite. Amphibole is green
362 hornblende locally replaced by actinolite and tremolite (Fig. 5f). Plagioclase grains are euhedral
363 and show intergrowths with other plagioclases, alkali feldspar (<5 vol.%) including microcline
364 (<1 vol.%), biotite (<2 vol.%), and green hornblende.

365 Granodiorite 1a, quartz diorite 1b and biotite tonalite 2b are each present at the Moly-Desgagné
366 showing and exhibit evidence of hydrothermal alteration, such as sericitized feldspars and
367 chloritized amphibole and biotite. Interstitial carbonates (ankerite) are also present in some of
368 these samples (Fig. 6). In addition, some tonalite 2a samples from the western part of the EJC

display a chloritization of biotite and pervasive hematization. Evidence of recrystallization was observed in most samples. Quartz grains can display an undulatory extinction or form subgrain domains having 120° angles (Fig. 4b). The six intrusive phases also contain euhedral to subhedral epidote grains with, in most cases, an allanite core (Figs. 5b, 6b). These epidotes are often included in amphibole or plagioclase and can reach up to 2 mm in length.

Amphiboles from all six intrusive phases are euhedral to subhedral. They display intergrowths with adjacent minerals, such as plagioclase and alkali feldspar, and commonly display a poikilitic texture. Inclusions of feldspar, quartz, biotite, or of other amphibole crystals, are anhedral. As stated above, amphibole can also include epidote grains with an allanite core. All these criteria, and particularly the latter, indicate that the amphiboles from all six intrusive phases of the EJC have a magmatic origin (Turner 1981; Schmidt and Poli 2004).

4.3. Mineral chemistry

Electron microprobe analyses ($n = 35$) of euhedral plagioclase grains from two samples of the amphibole-rich diorite 3b correspond to andesine (An_{33-49}) and labradorite (An_{51-53}) compositions. Microprobe analyses show that amphibole in the biotite tonalite 2b are magnesio-hornblende ($n = 11$). Amphibole grains from amphibole-rich diorite 3b are tschermakites ($n = 27$; Fig. 7a), and the core and rim of several amphiboles display distinct chemistries. From core to rim, a decrease in Al, Fe, Mn, and Ti (C-site) of 0.1–0.4 apfu and an increase in Na, K, and Ca (A-site) of 0.15–0.4 apfu are observed.

Amphibole chemistry was used to estimate the temperature and pressure conditions under which the EJC phases emplaced. The methods of Putirka (2016) and Anderson and Smith (1995) were used to evaluate temperature and pressure, respectively (Fig. 7b). Amphibole grains of quartz

diorite-tonalite 1b are estimated to have crystallized at 750 ± 30 °C and 2.2 ± 0.6 kbar. Most amphibole cores in amphibole-rich diorite 3b are estimated to have formed at $740\text{--}800 \pm 30$ °C and 3–4 kbar, whereas rims record higher temperature ($780\text{--}880 \pm 30$ °C) and pressure ($4\text{--}6 \pm 0.6$ kbar) conditions. As the pressure calculation method of Anderson and Smith (1995) is temperature-dependent, the temperature-independent geobarometer of Mutch et al. (2016) was also used to confirm the calculated pressures. The method of Mutch et al. (2016) returned higher pressure estimates of ca. 2.6 ± 0.5 kbar for quartz diorite-tonalite 1b and of ca. 4.5 ± 0.5 kbar and 8.1 ± 0.5 kbar for the cores and rims, respectively, of amphibole grains from amphibole-rich diorite 3b.

4.4. Deformation

According to our field observations and compiled structural data (Fig. 8), the main foliation planes strike approximately east–west, except near the contacts between phases EJC1, EJC2 and EJC3, and within the WNW-ESE deformation zone discussed below (Fig. 9a), where the main foliation becomes parallel to the contacts. A lack of significant vertical exposure prevents precise dip measurement; however, where observed, dips are subvertical. The main east–west foliation is penetrative and well-defined, on most outcrops, by aligned amphibole and plagioclase. In some thin sections, quartz grain alignment attributed to a dissolution-recrystallization process was observed, and probably indicates a solid-state deformation. A secondary north–south-striking foliation (Fig. 9a), faintly defined by preferentially oriented amphibole and plagioclase, was observed locally and is overprinted by the penetrative east–west-striking foliation. Only 18 of 118 examined outcrops exposing intrusive rocks display this early foliation.

Mafic xenoliths display two types of foliations (Fig. 9b): (1) a foliation of variable orientation, which is the result of a strong ductile strain restricted to the xenoliths and that was likely acquired before the EJC magma incorporated the xenoliths (Fig. 10a) and; (2) a foliation that mimics the contacts between the mafic rocks and the hosting EJC intrusive phase (Fig. 10b). The latter was likely acquired during or after xenoliths incorporation in the EJC magma (Fig. 10b). Foliation planes are defined by amphibole alignment, and most of these display a poikiloblastic texture characteristic of metamorphic amphiboles.

A WNW–ESE-trending higher-deformation zone transects the eastern EJC and encompasses the Moly-Desgagné showing (Fig. 8). This deformation zone is characterized by: (1) a penetrative WNW–ESE foliation affecting the magmatic phases and their mafic xenoliths, which is often expressed by a compositional layering resulting from the transposition of different lithologies within the deformation zone, and is best displayed in outcrops of the Moly-Desgagné showing (Figs. 8, 10c); (2) folding of early north–south foliation planes to form folds with ESE-trending fold axes and a plunge of $<30^\circ$ to the ESE; (3) a strong stretching lineation along a WNW–ESE direction plunging $16\text{--}35^\circ$ to the ESE and; (4) mostly sinistral, but also locally dextral, shear sense indicators (delta-shaped and sigma-shaped feldspars) in WNW-ESE foliation planes (Fig. 10d). A late east–west to WNW–ESE foliation marked by chlorite alignment and which crosscuts the pervasive WNW–ESE foliation is locally visible in the WNW–ESE deformation zone. Chronology and characteristics of the deformation features encountered by the EJC are summarized in Table 2.

4.5. Chemistry of the EJC

433 Macroscopic and petrographic observations described in the previous section point to limited
434 metamorphism and hydrothermal alteration, with the exception of some samples from the Moly-
435 Desgagné showing. Accordingly, metamorphism and hydrothermal alteration are not suspected to
436 have significantly modified the whole-rock chemistry of the studied samples. The studied rocks
437 lie within the subalkaline field of the total alkali versus silica (TAS) diagram (Le Bas et al. 1986;
438 Fig. 11a) and define a calc-alkaline trend on an AFM diagram (Irvine and Baragar 1971; Fig.
439 11b). Quartz diorite-tonalite 1b and amphibole-rich diorite 3b display the highest Fe and Mg
440 contents (Table 1). All samples are relatively K₂O-poor (mean of 1.26 wt% K₂O; Table 1) and lie
441 along the K-poor calc-alkaline differentiation trend defined by Lameyre and Bowden (1982) (Fig.
442 12). The CaO, Na₂O, and K₂O contents of the studied rocks correspond to those of experimental
443 liquids obtained by the melting of quartz-barren basalts (Moyen and Martin 2012; Fig. 11c).

444 Felsic phases (i.e., >65 wt% SiO₂; granodiorite 1a, tonalite 2a, biotite tonalite 2b and
445 trondhjemite 3a) display the lowest CaO, FeO and Fe₂O₃, MgO, P₂O₅, and TiO₂ contents (Figs.
446 13c, 13d, 13e, 13h and 13i), whereas intermediate to mafic phases (i.e., <65 wt% SiO₂; quartz
447 diorite-tonalite 1b and amphibole-rich diorite 3b, respectively) contain less alkalis (Na₂O, K₂O)
448 than the felsic phases (Figs. 13a and 13f). The Al₂O₃ content of most samples is ≥15 wt% Al₂O₃
449 and shows a slight decrease with increasing SiO₂ (Fig. 13a). The other elements, alkalis excepted,
450 decrease sharply with increasing SiO₂ content. Na₂O and K₂O values are scattered, possibly as a
451 consequence of weak alteration (Figs. 12b and 12f).

452 The six intrusive phases have distinct multi-element and REE profiles (Fig. 14). The profiles
453 show that tonalite 2a, biotite tonalite 2b and trondhjemite 3a are more fractionated (La/Yb_N
454 values of 23.18, 18.23, 13.07, respectively) than samples from phase EJC1 (1a and 1b) and
455 amphibole-rich diorite 3b (La/Yb_N values of 10.63, 5.80, and 3.18, respectively). Samples from

phase EJC1 and biotite tonalite 2b have similar LREE-enrichment, whereas samples from tonalite 2a and phase EJC3 are the least enriched in LREE. Biotite tonalite 2b and trondhjemite 3a are more depleted in HREE compared with the four other phases (Figs. 14b, 14d, 14f). All six intrusive phases display negative Nb, Ti, and Eu anomalies, although some samples display a positive Eu anomaly. Most samples also display positive Zr and Hf anomalies (Fig. 14a, 14c, 14e).

4.6. Zircon U–Pb ages

Results of chemically abraded ID-TIMS analyses of single zircon grains from phases EJC2 and EJC3 (Fig. 2) are presented below. They complement a recently published LA-ICP-MS study of igneous zircon from a EJC1 quartz diorite sample, which gave an age of 2718.6 ± 5.5 Ma (David 2018).

A biotite tonalite 2b sample (18UCD0137) yielded abundant, euhedral, colourless to medium brown and typically water-clear zircon having low-order facets. Three single grain ID-TIMS analyses cluster tightly on the concordia diagram (Fig. 15a), with $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 2724.5 and 2727.0 Ma (Supplementary Material S4). The youngest fractions, Z2 and Z3, which consist of a clear needle and an elongate broken prism, respectively, are identical in age, within error, and together yield a weighted average age of 2725.0 ± 0.9 Ma (2σ ; MSWD = 1.3), interpreted as a good representation of the age of crystallization of biotite tonalite 2b. Zircon fraction Z1 – a stubby, short broken prism fragment characterized by a slightly older $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2727.0 ± 1.4 Ma – might reflect crystallization in an earlier phase of the pluton (i.e. antecryst; Fig. 15a).

Zircons recovered from a sample of trondhjemite 3a (18UCD0043) comprise colourless to pale brown, well-faceted, short- to long-prismatic and needle-like forms that are square to flat in cross-section. Three of four analysed single grain fractions (Z1–Z3) cluster on the concordia within a narrow range of $^{207}\text{Pb}/^{206}\text{Pb}$ ratios (Fig. 15b; Supplementary Material S4). Two strongly overlapping analyses (Z1, Z3) yield a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2725.9 ± 1.1 Ma (2σ ; MSWD = 1.0). This date is interpreted as the best estimate of the age of crystallization of trondhjemite 3a. Fraction Z2, having a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2728.7 ± 1.8 Ma, may reflect a minor inheritance. A fourth fraction, Z4, with an age of 2713.5 ± 3.1 Ma is distinctly younger but slightly more discordant (0.4%) and may reflect minor secondary Pb loss.

5. Discussion

In this section, we use the characteristics of the main intrusive phases of the EJC, including field, petrological, geochemical, and geochronological data, to address: (1) the potential source compositions; (2) the number of distinct magmas that were emplaced and; (3) their timing and conditions of emplacement. Finally, the emplacement history and petrogenesis of the EJC are discussed.

5.1. Magma characteristics

Our new fieldwork and data corroborate the previously mapped boundaries of the EJC1 and EJC2 phases as defined in Fig. 2. However, new samples located in the “Tonalite de l’Est” intrusion (Fig. 2), which was delineated by the MERN, display the same textures, mineralogy and geochemical characteristics as samples from tonalite 2a or biotite tonalite 2b of the EJC. These new data thus indicate that the EJC2 phase extends 15 km further to the west than previously interpreted.

The six phases of the EJC all have a calc-alkaline affinity (Fig. 11b). Granodiorite 1a, tonalite 2a and biotite-tonalite 2b are rich in SiO_2 (>64 wt%), depleted in HREE, and display fractionated REE patterns, strong negative Nb and Ti anomalies and a less pronounced negative Ta anomaly. For these units, depletion in HREE is likely due to the presence of residual garnet and/or amphibole in the source (Moyen and Martin 2012) and, to a lesser extent, to the fractional crystallization of amphiboles (Prowatke and Klemme 2006; Szopa et al. 2013). Residual titanates (e.g., rutile, and also titanite; Briquet et al. 1984) are likely responsible for the negative Nb, Ta, and Ti anomalies (Sanematsu et al. 2016). Trondhjemite 3a, which is also silica-rich (> 64 wt% SiO_2), is slightly less depleted in HREE, less fractionated, and has less pronounced negative Nb, Ta, and Ti anomalies than the felsic phases mentioned previously. Intermediate and mafic phases 1b and 3b (SiO_2 <64 wt%) have higher HREE content and show less pronounced Nb, Ta, and Ti negative anomalies than felsic phases 1a, 2a, 2b and 3a.

Liou and Guo (2019) demonstrated that fractional crystallization of amphiboles and apatite can produce tonalite and trondhjemite from dioritic magmas. Throughout the evolution from diorite to tonalite and trondhjemite, they observed a decrease in LREE and HREE and an increase in the magnitude of the Eu negative anomaly; these chemical trends are explained by apatite, amphibole, and plagioclase fractionation, respectively. In the EJC, a decrease of the HREE content is also observed from quartz diorite-tonalite 1b and amphibole-rich diorite 3b to trondhjemite 3a, through tonalite 2a and biotite tonalite 2b (Fig. 14). This HREE decrease may correspond to amphibole fractionation from the mafic parental magmas. No major variation of the LREE content is observed, however, indicating either that the fractional crystallization of apatite is not an important process, or that other minerals modified the LREE content of the

521 magma (see below). The limited Al variation between each intrusive phase (Fig. 13a) points to
522 minor plagioclase fractionation.

523 Amphibole fractionation may explain the HREE decrease observed from granodiorite 1a to
524 tonalite 2a and biotite tonalite 2b, and then to trondhjemite 3a. This process should also have
525 increased the LREE content of most felsic phases, although such a change in LREE is not
526 observed. However, fractionation of a LREE-rich mineral such as allanite, even in small amounts,
527 is sufficient to induce a net LREE decrease and to compensate for the LREE increase caused by
528 amphibole fractionation. It is thus concluded that the fractionation of amphibole and allanite may
529 explain the observed REE profiles.

530 The major elements displayed on the Harker diagrams (Fig. 13) define a trend from the HREE-
531 richest rocks (granodiorite 1a) to the HREE-poorest rocks (tonalites 2a and 2b and trondhjemite
532 3a). This trend may be interpreted as a liquid line of descent (LLD). The units may thus be co-
533 magmatic and related by a differentiation process that resulted in an increase of Na₂O and K₂O,
534 and a decrease of all the other oxides, with increasing SiO₂. These interpretations only apply to
535 the most felsic intrusive phases. Indeed, the two dioritic phases (1b and 3b) are strongly enriched
536 in HREE compared with the other four, more felsic phases. Following these observations and
537 those made on the Chibougamau pluton (Mathieu and Racicot 2019), we propose that the HREE
538 decrease between diorites and the more felsic phases is too extensive to be solely attributed to the
539 fractionation of amphiboles.

540 These observations indicate that the most mafic (1b and 3b) and felsic (1a, 2a, 2b and 3a)
541 intrusive phases likely arose through the differentiation of two different parental magmas. These
542 two magmas could have been generated by the partial melting of the same source material at

different depths. The partial melting of a source where garnet and titanates are retained in the residuum could respectively account for the HREE depletion and negative Nb-Ti anomalies displayed by the felsic phases. The partial melting of a garnet-free source (i.e., same source but melting at a lower depth) could produce the HREE undepleted magmas that formed mafic phases 1b and 3b. In this model, the source of the EJC could be akin to that of TTG magmas and correspond to a metamorphosed and hydrated basalt (Nair and Chacko 2008; Moyen 2011; Moyen and Martin 2012; Martin et al. 2014). However, quartz diorite-tonalite 1b and amphibole-rich diorite 3b are too mafic for their source to be basalt, as this would require very high degrees of partial melting. Mixing with a more mafic component, such as a mantle peridotite, is required to account for the elevated Fe-Mg and low silica contents of these two mafic intrusive phases of the EJC. It also explains their relative HREE enrichment compared to the felsic phases derived from basaltic sources.

The model described by Polat et al. (2018) stipulates that the Doré Lake Complex, located 50 km to the north-east of the EJC (Fig. 1), formed in a subduction-collision context. However, data presented in our study are insufficient to support this hypothesis and further investigation (e.g., using stable isotopic systems) would be needed to test this model for the EJC. In the absence of such additional data, it is concluded that the coeval production of melts with both a mantle, and a basaltic source, as well as the local interaction of these melts to produce intrusive complexes such as the EJC (see above) and the slightly younger Chibougamau pluton (Mathieu and Racicot 2019) can occur within any of the tectonic settings proposed for the Neoproterozoic as listed in the introduction section.

This study concludes that the EJC is derived from two distinct magmas: (1) a first magma having a basaltic source that differentiated, by fractional crystallization or other processes, to form

granodiorite 1a, tonalite 1a, biotite tonalite 2b, and trondhjemite 3a and; (2) a second hybridized magma (mantle and basaltic source) that formed quartz diorite-tonalite 1b and amphibole-rich diorite 3b.

5.2. Petrogenesis and comparison with TTG suites

Intermediate to felsic large-volume synvolcanic intrusions in Archean greenstone belts generally correspond to TTG or TTD suites (Moyen and Martin 2012; Mathieu et al. 2020a). Samples from this study are akin to TTG suites, in that they belong to a K-poor calc-alkaline series (Fig. 12), as do most TTG suites (Moyen and Martin 2012). Also, tonalite (phases 2a and 2b) and trondhjemite (phase 3a) follow the trondhjemite trend, typical of TTG suites in the Na-Ca-K ternary space (Moyen and Martin 2012). In addition, the experimental melts obtained from a “quartz absent, ‘dry’” starting constituent have Na-Ca-K contents close to those of the EJC samples (Fig. 11c). By definition, TTG suites contain at least 64 wt% SiO₂, as well as a limited amount of Fe, Mg, Mn, and Ti, and contain no more than 0.5–2 wt% K₂O (Moyen and Martin 2012). As such, granodiorite 1a, amphibole-rich diorite 3b and some sample from diorite-tonalite 1b do not contain enough SiO₂ to be considered part of a TTG suite.

Applying the chemical criteria listed by Moyen and Martin (2012) to the EJC rocks shows that 46 of 66 samples (studied and compiled) contain enough SiO₂ (>64 wt%) to belong to a TTG suite (Table 3). These 46 samples correspond to tonalite 2a, biotite tonalite 2b, trondhjemite 3a, and to part of granodiorite 1a. The remaining 20 samples are mostly diorite and quartz diorite. Among the 46 samples with SiO₂ > 64 wt%, the Fe-Mg content of several mafic mineral-enriched samples is higher than expected for TTG suites (Table 3). Rocks from the EJC have Na₂O and Al₂O₃ contents that are within the expected range for TTG suites. Their trace-element and REE

contents are also close to those expected for TTG suites (Table 3). Tonalite 2a, biotite tonalite 2b and trondhjemite 3a of the EJC belong to the “medium REE” and “high REE” TTG groups, which correspond to melts generated at 10–20 kbar and >20 kbar, respectively (Moyen and Martin 2012) (Table 3). These interpretations are only valid if the HREE content of TTG parental magmas is not significantly modified by differentiation processes. However, recent studies demonstrated that amphibole fractionation can significantly modify the HREE content of TTG suites (Liou and Guo 2019; Laurent et al. 2020). As discussed above, amphibole fractionation likely occurred and slightly modified the HREE content of the EJC samples. Thus, the HREE content of the EJC phases is considered a poor proxy of the partial melting depth, which remains poorly constrained for the EJC.

The EJC is mainly composed of tonalite, trondhjemite, and diorite, with a minor amount of granodiorite. This comparison between EJC chemistry and TTG classification confirm the conclusion drawn previously, i.e. that the EJC is a TTD intrusive suite formed from two distinct parental magmas having different sources: (1) a magma akin to that of TTG suites, which formed by the partial melting of a garnet-titanate-bearing metamorphosed hydrated basalt (Moyen and Martin 2012), produced the HREE-depleted phases 1a, 2a, 2b and 3a and; (2) a magma extracted from a source having a mantle component and that may have interacted with TTG melt (Martin et al. 2005), responsible for the formation of less HREE-depleted quartz diorite-tonalite 1b and amphibole-rich diorite 3b.

The aim of this contribution is to provide new insights into the differentiation and emplacement depth of a tonalite-dominated (TTD) intrusive suite, which could be of use to other studies whose aims are to better understand Archean magmatic-hydrothermal mineralizing systems associated with TTG and TTD intrusive suites. Porphyry-style Au-Cu mineralization is genetically related to

synvolcanic TTG-TTD intrusions (Mathieu and Racicot 2019; Katz et al. 2017) whereas Mo mineralization is generally related to highly fractionated felsic intrusions (Sillitoe 2010; Sinclair 2017; Richards 2018). It is beyond the scope of this paper to discuss the nature and origin of the Mo mineralization in the Moly-Desgagné showing. However, as our results show that the EJC belongs to a TTD intrusive suite, it can be concluded that this magmatic system is likely not sufficiently evolved and that the most fractionated phases (trondhjemite 3a) are of a too limited volume to have generated Mo mineralization.

5.3. Timing of emplacement

The new CA-ID-TIMS U–Pb zircon geochronological data, which established crystallization ages of 2725.0 ± 0.9 Ma and 2725.9 ± 1.1 Ma for EJC2 and EJC3 intrusive phases, respectively, overlap within a narrow analytical error. This suggests essentially synchronous emplacement of EJC2 and EJC3, two of the youngest magmatic components of the EJC. The EJC1 quartz diorite-tonalite is older than EJC2 and EJC3 on the basis of field relationships, and its LA-ICP-MS zircon age, reported at 2718.6 ± 5.5 Ma (David 2018), may be closer to the upper end of its error limit at ca. 2724 Ma. Given that all three radiogenic age determinations agree within the bounds of their 2-sigma errors, we conclude that the main phases of the EJC were emplaced within a narrow interval of time at ca. 2725–2724 Ma.

In a regional context, emplacement of the EJC occurred relatively late in the evolution of volcanic cycle 1 of the Roy Group (ca. 2730–2727 Ma to ca. 2724 Ma), immediately postdating the youngest members of the Waconichi Formation (ca. 2730–2727 Ma; summarized in Leclerc et al. 2011), but occurring slightly after the intrusion of the 2728–2727 Ma Lac Doré Complex (Mortensen 1993a). Moreover, intrusions of the EJC narrowly predate or are coeval with the

onset of volcanic cycle 2 magmatism (≥ 2724 Ma; Davis et al. 2014). Our new geochronological data suggest the Obatogamau Formation formed a short time before the EJC. This is consistent with the mafic and intermediate xenoliths observed in the field, which may represent volcanic rocks of the Obatogamau, Des Vents (2798.7 ± 0.7 Ma; Davis et al. 2014) and/or Chrissie (2791.4 ± 3.7 Ma; David et al. 2011) formations.

We conclude that the EJC is dominated by at least two distinct magmas that both underwent differentiation in, possibly, the lower part of the crust. Several magma injections, originating from two or more deep magma accumulations (i.e., magma chambers *a* and *b*), successively formed, in a narrow time span, the various intrusions of the EJC. This scenario best explains the crosscutting relationships observed in the field. It explains how magma from magma chamber *a* (e.g., granodiorite 1a) could be found enclosed in a phase extracted from magma chamber *b* (e.g., quartz diorite-tonalite 1b), which can in turn be enclosed in a less evolved phase (e.g., tonalite 2a or biotite tonalite 2b) from magma chamber *a*.

5.4. Conditions of crystallization

All six intrusive phases are dominated by variable proportions of plagioclase, alkali feldspar, quartz, biotite and amphibole, with accessory epidote and allanite, all of magmatic origin. Quartz diorite-tonalite 1b and amphibole-rich diorite 3b contain up to 10 vol.% and up to 31 vol.% amphibole, mostly green hornblende, whereas the felsic phases are dominated by quartz, feldspar, with minor biotite and almost no amphibole. All intrusive phases show magmatic mineral assemblages that crystallized under amphibolite facies P-T conditions (Hammarstrom and Zen 1986; Schmidt and Poli 2004). Indeed, the assemblage of green hornblende and andesine-labradorite is characteristic of amphibolite-facies conditions. Quartz and feldspar subgrains,

655 having 120° angles, point to deformation and recrystallization (Turner 1981). In addition, some
656 rocks, mostly observed in the deformation zone, contain minor amounts of actinolite, albite,
657 chlorite, and white mica, an assemblage indicative of retrograde (greenschist) metamorphism that
658 likely accompanied exhumation of the EJC.

659 All EJC units contain magmatic epidote with allanite cores, which are embedded in plagioclase or
660 amphibole. For the EJC intrusive phases, the presence of magmatic epidote and associated
661 mineral assemblages point to P-T crystallization conditions of approximately 650–720 °C and 5–
662 14 kbar for tonalites (2a and 2b), 660–730 °C and 9–16 kbar for trondhjemite (3a), and 620–700
663 °C and 5–15 kbar for granodiorite (1a), according to the phase diagrams of Schmidt and Poli
664 (2004) for, respectively, tonalite, trondhjemite, and granodiorite. Thus, magmatic epidote
665 indicates that EJC magmas differentiated, at least in part, at depths of at least 18 km, as deduced
666 by converting P-T conditions using the lithostatic pressure estimated in the Superior Province for
667 the Archaean period (Brown and Hagemann 1995).

668 The bulk of the amphiboles are magmatic hornblende, which is stable at approximately 450–
669 700 °C and 2–10 kbar (Turner 1981; Hyndman 1992). As well, a portion of the amphibole is
670 tschermakite, according to microprobe analysis. This points to medium- to high-P conditions
671 (Hynes 1982), as tschermakite is stable between 7 and 17 kbar at 800 °C (Oba 2013). These
672 conditions correspond to depths of at least 20 km (Brown and Hagemann 1995). The amphiboles
673 (magnesio-hornblende) of quartz diorite-tonalite 1b crystallized at 700–750 °C and 2–2.5 kbar
674 (7–8 km depth), according to the method of Putirka (2016) and the Al-in-hornblende
675 geobarometer (Anderson and Smith 1995), respectively. The amphiboles (tschermakite) of
676 amphibole-rich diorite 3b crystallized at 750–800 °C and 3–5.5 kbar, using the method of

Anderson and Smith (1995), or between 4.5 ± 1.5 and 6.9 ± 3.1 kbar, according to the method of Mutch et al. (2016). This corresponds to depths of 10 and 19 km, respectively.

The P-T conditions summarized above overlap for tonalite, trondhjemite, and granodiorite phases (2a and 2b, 3a, and 1a, respectively) and correspond to P-T conditions of 650–700 °C and ca. 5–10 kbar. This suggests that the magma mostly crystallized at a depth of at least 11 km. In situ P-T estimates described above for quartz diorite-tonalite 1b and amphibole-rich diorite 3b show that these intrusive phases crystallized at depths between 7 and 19 km and likely rose as crystal-laden magmas. Magmas of granodiorite 1a, tonalite 2a, biotite tonalite 2b and trondhjemite 3a may have begun to crystallize at the same depth as quartz diorite-tonalite 1b and amphibole-rich diorite 3b, but could have risen without most amphiboles. This hypothesis is supported by petrographic evidence for low modal proportions of mafic minerals and by the HREE depletion and enrichment, respectively, of intermediate-mafic (1b and 3b) and felsic (1a, 2a, 2b and 3a) phases (Fig. 14). Indeed, amphiboles crystallizing at depth, but not carried upward by the magma, likely induced the HREE depletion displayed by phases 1a, 2a, 2b, and 3a. In contrast, quartz diorite-tonalite 1b and amphibole-rich diorite 3b contain more amphibole, and as a result are HREE-enriched, indicating that magmas entrained these minerals on ascent.

Another line of evidence for the final depth of emplacement and final crystallization of the EJC is provided by the mafic xenoliths. These mafic xenoliths are observed in every phase of the EJC and they underwent amphibolite-grade contact metamorphism, as indicated by poikiloblastic amphiboles. Microprobe analyses demonstrate that these amphiboles are mostly magnesiohornblende, which is typical of the amphibolite facies. Most mafic xenoliths contain feldspar megacrysts and likely came from the Obatogamau Formation, which is defined as a regional feldspar megacryst-bearing volcanic unit (Leclerc et al. 2017). The thickness of the older Des

Vents and Chrissie, and cycle 1 Obatogamau formations have been estimated respectively at 2–2.5 km (Mueller et al. 1989), >1.8 km (Charbonneau et al. 1991), and approximately 4 km (Leclerc et al. 2017). The total thickness of the volcanic pile prior to the onset of volcanic cycle 2 was thus 4.3–8.3 km at the time of EJC emplacement. This thickness corresponds to a pressure of 1 kbar at the base of the Obatogamau Formation or 2.5 kbar at the base of the Des Vents Formation. Even considering that these estimates of the volcanic pile thickness are approximate, the corresponding pressure range is much lower than pressure estimated in the previous section for the phases of the EJC. Thus, to incorporate xenoliths from the Des Vents, Chrissie, and Obatogamau formations, the EJC phases must complete their crystallization at a depth of <8.3 km, which is a much shallower crustal level than suggested by the chemistry of magmatic amphiboles (i.e., 7–19 km depth). Thus, the best pressure estimate for the final EJC emplacement depth is provided by quartz diorite-tonalite 1b amphiboles and is ca. 2–2.5 kbar. This pressure corresponds to a depth of 7–8 km, which is in agreement with the inferred thickness of the volcanic pile at the time of EJC emplacement. It is thus possible that the EJC phases reached their final level of emplacement as crystal-laden magmatic mushes.

In addition, amphiboles from phase 3b record a pressure increase from core to rim, which could correspond to partial amphibole recrystallization during burial. This burial may correspond to the thickening of the volcanic pile by the onset of volcanic cycle 2. Tectonic movements, such as the onset of shortening towards the end of the synvolcanic period, may also be responsible for this late pressure increase (Chown et al. 1992; Mathieu et al. 2020b).

5.5. Structural evolution

The contact metamorphic aureole associated with the EJC has been obliterated by regional greenschist metamorphism (Tait 1992) coeval with the D_2 event, which occurred during the syntectonic period (Daigneault et al. 1990). Our new geochronological data confirm that the EJC was emplaced during a brief interval of time at ca. 2724 Ma during the synvolcanic period (2730–2708 Ma; Chown et al. 1992), rather than during the younger syntectonic period (2700–2682 Ma; Chown et al. 1992), as suggested by Chown et al. (2002).

Field data indicates that the EJC area experienced several stages of deformation (Table 2). A strong ductile strain, recorded by metavolcanic xenoliths (Figs. 10a), predates the EJC emplacement at ca. 2724 Ma (i.e. pre-EJC local deformation). Metavolcanic xenoliths also record a less pronounced foliation along their rims, which was likely formed as they were incorporated into the EJC. The EJC then experienced two stages of ductile deformation reflected by an early north–south-striking foliation (EJC local deformation 1) overprinted by a main east–west-striking and subvertical foliation (EJC local deformation 2). The second foliation is pervasive and is even more pronounced in the WNW–ESE deformation zone (Fig. 10c). These two ductile foliations are specific to the EJC (i.e., they are not recorded by the volcanic hosts rocks), indicating deformation events occurred at depth during or slightly after the emplacement of the EJC. As there is a gap of >20 Myr between the EJC emplacement and the onset of the syntectonic period, it is most unlikely that the two episodes of ductile deformation experienced by the EJC correspond to the deformation events D_1 (minor and possibly local, pre-ductile event) and D_2 (major Neoproterozoic north–south shortening ductile event; Dimroth et al. 1986; Daigneault et al. 1990) of the syntectonic period. However, a late east–west to WNW–ESE foliation defined by chlorite alignment is observed in the WNW–ESE deformation zone. This late foliation is probably associated to the regional D_2 event, which likely occurred during or after the

exhumation of the EJC (Table 2). This event could have caused a late dextral reactivation of this initially sinistral deformation zone.

5.6. Emplacement model

The EJC was intruded at a final depth of 7–8 km, which is too shallow to invoke diapirism as a realistic mode of emplacement (Vigneresse and Clemens 2000). The Lapparent suite, which is composed of several chemically-distinguishable intrusive members, has as principal component the Lapparent intrusion itself in the central part, and the EJC on its easternmost flank (Figs. 1 and 2), all of which are aligned along an east–west direction. The elongated shape of the grouping of intrusions is unlikely the result of deformation, as such voluminous intrusions are hard to deform to the observed ratio. Rather, this alignment is likely the result of magma injections along a major, possibly crustal-scale, east–west structure. This scenario is supported by Chown et al. (2002) who demonstrated that the Houghton pluton, located 35 km west of the EJC, was emplaced at roughly 4 kbar along a major fault. The exhumation event registered by the EJC is in agreement with the model of the major east–west horst faults flanking the Lapparent suite to the north and south of Chown and Mueller (1992). The EJC is composed of six different intrusive phases derived from two distinct magmas and emplaced within a short period at ca. 2724 Ma. Several studies (Clemens and Mawer 1992; Petford et al. 1994; Glazner et al. 2004) have shown that multiple magma injections along faults can produce voluminous and multiphase plutons within a short time span (i.e., under 10 000–100 000 years). We propose that the EJC was likewise formed via multiple injections of magma that ascended along faults and fractures as a result of deformation and/or of an important magma influx.

The existence of an east–west structure that facilitated the ascent of EJC magmas can be inferred from the geometry of the Lapparent suite and its associated intrusions, including the EJC (Chown and Mueller 1992; Chown et al. 2002), but the volcanic rocks outcropping to the east and west of these grouping of intrusions provide no evidence for the extension of this major structure into the volcanic pile. However, other structures could have served as conduits for ascending magma. Feeder zones for plutons generally correspond to active shear zones where magmas are collected and then ascend (Moyen et al. 2003). Our field data point to two zones that concentrate numerous injections of different EJC magmas: (1) the sharp western limit of phase EJC3 in contact with EJC2; and (2) the WNW–ESE deformation zone crosscutting the eastern part of the EJC (Fig. 8). In these two zones, the intrusive phases display evidence of ductile deformation, which, as demonstrated before, occurred during the EJC emplacement and before the regional deformation events. Because injections of several intrusive phases of the EJC within narrow zones have not been observed elsewhere in the EJC, we infer that these two zones could correspond to potential feeder zones for the EJC magmas.

Intrusions formed by magma injected along fractures and invading stratified sequences commonly form sills (Menand 2008). After its ascent, magma accumulates in weak zones, such as contacts between lithological units (Roman-Berdiel et al. 1995). Additional magma input causes the sill to inflate and to form a laccolith (Menand 2008), and its inflation can locally deform the surrounding rocks (Roman-Berdiel et al. 1995). This mechanism has been proposed by Benn (2004) for the formation of Kenogamissi Complex (Swayze Subprovince, Ontario) and can also be applied to the EJC. Thus, we propose that repeated magma injections formed a sill at the interface between the older Des Vents and Chrissie formations and/or between the Chrissie and Obatogamau formations (Fig. 16a). Laccolith growth through magma input created a

pervasive foliation within the previously injected phases, and caused the foliation to become parallel to the contact between each magmatic phase. Growth of the laccolith by vertical expansion promoted uplift of the overlying Des Vents and Chrissie formations (Figs. 16b, 16c, 16d), which explains why these formations are only exposed along the rim of the EJC in the Chibougamau area. The inflation-induced deformation of the volcanic sequence do not account for the strong ductile deformation recorded by the xenoliths prior to their incorporation by the EJC, but it explains the fabric recorded along the rims of the xenoliths as they were incorporated into the EJC (Fig. 16c, 16d). Then, the EJC was buried, then exhumed, and underwent the regional D₂ deformation event and associated greenschist metamorphism.

6. Conclusions

The EJC is a TTD suite composed of six intrusive phases. Fractional crystallization dominated by amphibole fractionation affected a parental tonalitic magma, derived from the partial melting of a garnet-titanate-bearing metamorphosed basalt, and formed the HREE-depleted granodiorite 1a, tonalite 2a, biotite tonalite 2b and trondhjemite 3a phases. Another parental magma, corresponding to a mantle-derived mafic magma that may have interacted with TTG melts, formed the HREE-enriched quartz diorite-tonalite 1b and amphibole-rich diorite 3b phases.

New geochronological data, along with microprobe analyses of amphiboles and field relationships, indicate that the EJC was emplaced in rapid succession during the synvolcanic period at ca. 2725-2724 Ma and at approximately 7–8 km depth. EJC magmas ascended along faults or fractures and formed one or several sills along discontinuities at the base of (and/or within) the Chrissie, Des Vents, and Obatogamau formations. The WNW–ESE deformation zone and the western limit of the EJC3 phase represent potential magmatic conduits. Magma influx

promoted sill inflation that formed laccoliths, and later exposed the Des Vents and Chrissie formations along the margin of the EJC. The laccolith inflation also marginally deformed the host volcanic units, imparting a fabric preserved within xenoliths found in all EJC intrusive phases.

The volcanic rocks that predate the EJC have been subjected to a stage of strong ductile strain prior to the emplacement of the EJC. The EJC subsequently underwent two stages of ductile deformation during the synvolcanic period, forming an early north–south foliation followed by the pervasive east–west foliation. This second episode also strongly impacted the WNW–ESE deformation zone. The EJC was later exhumed and weakly affected by the regional D_2 deformation event and greenschist metamorphism. Brittle east–west deformation and retrograde greenschist metamorphism are best expressed in the WNW–ESE deformation zone.

Acknowledgements

This study was undertaken as part of the Metal Earth project (Laurentian University) investigation of the Chibougamau area. This research was funded by Canada First Research Excellence Funds and by federal, provincial, and industrial partners (<http://merc.laurentian.ca/research/metal-earth/>). The authors acknowledge Laurentian University for the training and logistical support provided throughout the project. The authors are grateful to the independent prospector Larry Desgagné who gave us access to his claims (the Moly-Desgagné showing), which were key for this study, and provided support for sampling. Patrick Houle (MERN) is also warmly thanked for his help throughout the summer field work in the Chibougamau area, as well as Dr. Réal Daigneault (UQAC) for his help unravelling the structural history of the EJC. The first author also thanks the field assistants and the Chibougamau Metal Earth team for the fieldwork, support, and discussions that contributed greatly to the thinking

832 behind this study. We thank editor Dr. S. Pehrsson, associate editor F. Corfu, reviewers M.
833 Sanborn-Barrie and H. Sandeman, and an anonymous reviewer for their constructive reviews
834 which greatly improved the manuscript. This paper is Metal Earth contribution number MERC-
835 ME-2020-100.

Draft

References

- Adam, J., Rushmer, T., O'Neil, J., and Francis, D. 2012. Hadean greenstones from the Nuvvuagittuq fold belt and the origin of the Earth's early continental crust. *Geology*, **40**: 363-366.
- Anderson, J.L., and Smith, D.R. 1995. The effect of temperature and fO₂ on the Al-in-hornblende barometer. *American Mineralogist*, **80**: 549-559.
- Ayer, J., Amelin, Y., Corfu, F., Kamo, S., Ketchum, J., Kwok, K., and Trowell, N. 2002. Evolution of the southern Abitibi belt based on U–Pb geochronology: autochthonous volcanic construction followed by plutonism, regional deformation and sedimentation. *Precambrian Research*, **115**: 63-95.
- Baker, D.J. 1980. The metamorphism and structural history of the Grenville Front near Chibougamau. Unpublished Ph.D., University of Georgia, Athens, GA, United States.
- Bédard, J.H., Brouillette, P., Madore, L., and Berclaz, A. 2003. Archaean cratonization and deformation in the northern Superior Province, Canada: an evaluation of plate tectonic versus vertical tectonic models. *Precambrian Research*, **127**: 61-87.
- Bédard, J.H., Harris, L.B., and Thurston, P. 2013. The hunting of the snArc. *Precambrian Research*, **229**: 20-48.
- Bédard, J.H. 2018. Stagnant lids and mantle overturns: Implications for Archaean tectonics, magmagenesis, crustal growth, mantle evolution, and the start of plate tectonics. *Geoscience Frontiers*, **9**: 19-49.
- Benn, K. 2004. Late Archaean Kenogamissi complex, Abitibi Subprovince, Ontario, Canada: doming, folding and deformation-assisted melt remobilisation during syntectonic batholith emplacement. *Transactions of the Royal Society of Edinburgh: Earth Sciences*, **95**: 297-307.
- Boucher, A., Mathieu, L., Hamilton, M.A., Bedeaux, P., and Daigneault, R. 2020. Petrogenesis and economic potential of the Obatogamau Formation, Chibougamau area, Abitibi greenstone belt. *Canadian Journal of Earth Sciences*, **in press**.
- Briqueu, L., Bougault, H., and Joron, J.L. 1984. Quantification of Nb, Ta, Ti and V anomalies in magmas associated with subduction zones: petrogenetic implications. *Earth & Planetary Science Letters*, **68**: 297-308.
- Brown, P.E., and Hagemann, S.G. 1995. Geobarometry and depth relations in Archean lode-gold deposits. XIII ECROFI Symposium, **18**: 30-31.
- Calvert, A.J., Sawyer, E.W., Davis, W.J., and Ludden, J.N. 1995. Archaean subduction inferred from seismic images of a mantle suture in the Superior province. *Nature*, **375**: 670-674.

- 870 Calvert, A.J., and Ludden, J.N. 1999. Archean continental assembly in the southeastern Superior
871 province of Canada. *Tectonics*, **18**: 412-429.
- 872 Chapon, B., Jébrak, M., Ross, P.S., and Stevenson, R. 2010. Le système porphyrique à
873 molybdène de Tilly. DIVEX (Diversification de l'Exploration Minérale au Québec),
874 rapport final 2010, sous-projet SC36.
- 875 Charbonneau, J.-M., Picard, C., and Dupuis-Hebert, L. (eds). 1991. Synthèse géologique de la
876 région de Chapais-Branssat. Ministère de l'Energie et des Ressources Naturelles, MM 88-
877 01.
- 878 Chown, E.H., Daigneault, R., Mueller, W., and Mortensen, J.K. 1992. Tectonic evolution of the
879 Northern Volcanic Zone, Abitibi belt, Quebec. *Canadian Journal of Earth Sciences*, **29**:
880 2211-2225.
- 881 Chown, E.H., and Mueller, W. 1992. Basement Influence on the Supracrustal and Plutonic
882 Evolution and Deformation of an Archean Greenstone Belt. *In* *Basement Tectonics 7 -*
883 *Proceedings of the Seventh International Conference on Basement Tectonics. Edited by R.*
884 *Mason. Robert Mason, Kingston, Ontario, Canada. pp. 465-476.*
- 885 Chown, E.H., Harrap, R., and Moukhsil, A. 2002. The role of granitic intrusions in the evolution
886 of the Abitibi belt, Canada. *Precambrian Research*, **15**: 291-310.
- 887 Clemens, J.D., and Mawer, C.K. 1992. Granitic magma transport by fracture propagation.
888 *Tectonophysics*, **203**: 339-360.
- 889 Condie, K.C. 1993. Chemical composition and evolution of the upper continental crust :
890 Contrasting results from surface samples and shales. *Chemical Geology*, **104**: 1-37.
- 891 Cross, W., Iddings, J.P., Pirsson, L.V., and Washington, H.S. 1902. A quantitative chemico-
892 mineralogical classification and nomenclature of igneous rocks. *The Journal of Geology*,
893 **10**: 556-697.
- 894 Daigneault, R., and Allard, G.O. (eds). 1990. Le complexe du Lac Doré et son environnement
895 géologique (région de Chibougamau - Sous-province de l'Abitibi). Ministère de l'Energie
896 et des Ressources Naturelles, Québec, MM 89-03.
- 897 Daigneault, R., St-Julien, P., and Allard, G.O. 1990. Tectonic evolution of the northeast portion
898 of the Archean Abitibi greenstone belt, Chibougamau area, Quebec. *Canadian Journal of*
899 *Earth Sciences*, **27**: 1714-1736.
- 900 Daigneault, R. 1996. Couloirs de déformation de al Sous-Province de l'Abitibi. Ministère des
901 Ressources Naturelles, MB 96-33, Québec.
- 902 Dallmeyer, R.D., Maybin, A.H., and Durocher, M.E. 1975. Timing of Kenoran Metamorphism in
903 the Eastern Abitibi Greenstone Belt, Quebec: Evidence From $^{40}\text{Ar}/^{39}\text{Ar}$ Ages of
904 Hornblende and Biotite From Post-Kinematic Plutons. *Canadian Journal of Earth*
905 *Sciences*, **12**: 1864-1873.

- 906 David, J., Davis, D.W., Dion, C., Goutier, J., Legault, M., and Roy, P. 2007. U-Pb age dating in
907 the Abitibi Subprovince in 2005-2006. Ministère des Ressources Naturelles, Québec, RP
908 2007-01(A).
- 909 David, J., McNicoll, V., Simard, M., Bandyayera, D., Hammouche, H., Goutier, J., Pilote, P.,
910 Rhéaume, P., Leclerc, F., and Dion, C. 2011. Datations U-Pb effectuées dans les
911 provinces du Supérieur et de Churchill en 2009-2010. Ministère des Ressources
912 Naturelles et de la Faune, Québec, RP 2011-02(A).
- 913 David, J. 2018. Datations U-Pb dans la Province du Supérieur effectuées au GEOTOP en 2015-
914 2016. Ministère de l'Énergie et des Ressources Naturelles, Québec, MB 2018-16.
- 915 Davis, D.W., Simard, M., Hammouche, H., Bandyayera, D., Goutier, J., Pilote, P., Leclerc, F.,
916 and Dion, C. 2014. Datations U-Pb effectuées dans les provinces de Supérieur et de
917 Churchill en 2011-2012. Ministère des Ressources Naturelles et de la Faune, Québec, RP
918 2014-05.
- 919 Desgagné, L. 2007. Rapport d'exploration simplifié. Ministère des Ressources Naturelles et de la
920 Faune, GM 63200.
- 921 Dimroth, E., Mueller, W., Rocheleau, M., Archer, P., Jutras, M., Piché, M., Simoneau, P.,
922 Carignan, J., Chown, E.H., Guha, J., Goulet, N., Allard, G.O., Franconi, A., and Gobeil,
923 A. 1983. Stratigraphie et évolution du bassin de transition entre les groupes de Roy et
924 d'Opémisca, région de Chibougamau-Chapais. In *Stratigraphie des ensembles volcano-
925 sédimentaires archéens de l'Abitibi : état des connaissances. Edited by MRN. Ministère
926 de l'Énergie et des Ressources, DV 83-11, Québec. pp. 21-33.*
- 927 Dimroth, E., Mueller, W., Daigneault, R., Brisson, H., Poitras, A., and Rocheleau, M. 1986.
928 Diapirism during regional compression: the structural pattern in the Chibougamau region
929 of the Archean Abitibi Belt, Quebec. *Geologische Rundschau*, **75**: 715-736.
- 930 Faure, S. 2012. Réévaluation de la géologie et des modèles d'exploration pour l'or dans le
931 segment Caopatina-Desmaraisville, sud de Chibougamau. Projet CONSOREM 2012-02.
- 932 Feng, R., and Kerrich, R. 1992. Geodynamic evolution of the southern Abitibi and Pontiac
933 terranes: evidence from geochemistry of granitoid magma series (2700 - 2630 Ma).
934 *Canadian Journal of Earth Sciences*: 2266-2286.
- 935 Fraser, R.J. 1993. The Lac Troilus Gold-Copper Deposit, Northwestern Quebec : A Possible
936 Archean Pophyry System. *Economic Geology*, **88**: 1685-1699.
- 937 Gerstenberger, H., and Haase, G. 1997. A highly effective emitter substance for mass
938 spectrometric Pb isotope ratio determinations. *Chemical Geology*, **136**: 309-312.
- 939 Gerya, T., Stern, R.J., Baes, M., Fischer, R., Sizova, E., Sobolev, S.V., and Whattam, S.A. 2015.
940 Plume tectonics and cratons formation in the early Earth. In *AGU Fall Meeting Abstracts*.

- 941 Glazner, A.F., Bartley, J.M., Coleman, D.S., Gray, W., and Taylor, R.Z. 2004. Are plutons
942 assembled over millions of years by amalgamation from small magma chambers ? GSA
943 Today, **14**: 4-11.
- 944 Greenwood, N.N., and Earnshaw, A. 1984. Chemistry of the elements. Pergamon press, Oxford.
- 945 Grenier, J., and Sansfaçon, R. 2011. Prospection 2010 et programme de forage 2011 – Moly-
946 Desgagné, 32G10. Ministère des Ressources Naturelles et de la Faune, GM 66125.
- 947 Hammarstrom, J.M., and Zen, E.-A. 1986. Aluminium in hornblende: An empirical igneous
948 barometer. American Mineralogist, **71**: 1297-1313.
- 949 Hastie, A.R., and Fitton, J.G. 2019. Eoarchaeon tectonics: New constraints from high pressure-
950 temperature experiments and mass balance modelling. Precambrian Research, **325**: 20-38.
- 951 Hawthorne, F.C., Oberti, R., Harlow, G.E., Maresch, W.V., Martin, R.F., Schumacher, J.C., and
952 Welch, M.D. 2012. IMA report - Nomenclature of the amphibole supergroup. American
953 Mineralogist, **97**: 2031-2048.
- 954 Herzberg, C., Condie, K.C., and Korenanga, J. 2010. Thermal history of the Earth and its
955 petrological expression. Earth & Planetary Science Letters, **292**: 79-88.
- 956 Hoffmann, J.E., Münker, C., Næraa, T., Rosing, M.T., Herwartz, D., Garbe-Schönberg, D., and
957 Svahnberg, H. 2011. Mechanisms of Archean crust formation inferred from high-
958 precision HFSE systematics in TTG. Geochimica et Cosmochimica Acta, **75**: 4157-4178.
- 959 Hutchison, C.S. 1975. The norm, its variations, their calculation and relationships.
960 Schweizerische mineralogische petrografische Mitteilungen, **55**: 243-256.
- 961 Hyndman, D.W. 1992. Petrology of igneous and metamorphic rocks. McGraw-Hill, New York
962 and London.
- 963 Hynes, A. 1982. A Comparison of Amphiboles from Medium- and Low-Pressure Metabasites.
964 Contributions to Mineralogy and Petrology, **81**: 119-125.
- 965 Irvine, T.N., and Baragar, W.R.A. 1971. A Guide to the Chemical Volcanic Rocks. Canadian
966 Journal of Earth Sciences, **8**: 523-548.
- 967 Katz, L.R., Kontak, D.J., Dubé, B., and McNicoll, V. 2017. The geology, petrology, and
968 geochronology of the Archean Côte Gold large-tonnage, low-grade intrusion-related Au
969 (Cu) deposit, Swayze greenstone belt, Ontario, Canada. Canadian Journal of Earth
970 Sciences, **54**: 173-202.
- 971 Kendrick, J., and Yakymchuk, C. 2020. Garnet fractionation, progressive melt loss and bulk
972 composition variations in anatectic metabasites: Complications for interpreting the
973 geodynamic significance of TTGs. Geoscience Frontiers, **11**: 745-763.
- 974 Kerr, P.F. 1959. Optical Mineralogy. McGraw-Hill, University of California.

- 975 Kimura, G., Ludden, J.N., Desrochers, J.P., and Hori, R. 1993. A model of ocean-crust accretion
976 for the Superior province, Canada. *Lithos*, **30**: 337-355.
- 977 Kline, S.W. 1985. Metamorphic mineral chemistry, petrology, and sulfide mineralogy of the
978 Dore Lake complex, Chibougamau, Quebec. Unpublished Ph.D., University of Georgia,
979 Athens, GA, United States.
- 980 Lameyre, J., and Bowden, P. 1982. Plutonic rock type series: discrimination of various granitoid
981 series and related rocks. *Journal of Volcanology and Geothermal Research*, **14**: 169-186.
- 982 Laurent, O., Martin, H., Moyen, J.-F., and Doucelance, R. 2014. The diversity and evolution of
983 late-Archean granitoids : Evidence for the onset of « modern-style » plate tectonics
984 between 3.0 and 2.5 Ga. *Lithos*, **205**: 208-235.
- 985 Laurent, O., Björnsen, J., Wotzlaw, J.-F., Bretscher, S., PSilva, M.P., Moyen, J.-F., Ulmer, P.,
986 and Bachmann, O. 2020. Earth's earliest granitoids are crystal-rich magma reservoirs
987 tapped by silicic eruptions. *Nature Geoscience*, **13**: 163-169.
- 988 Le Bas, M.J., Le Maitre, R.W., Streckeisen, A., and Zanettin, B. 1986. A Chemical Classification
989 of Volcanic Rocks Based on the Total Alkali-Silica Diagram. *Journal of Petrology*, **27**:
990 745-750.
- 991 Leclerc, F., Bédard, J.H., Harris, L.B., McNicoll, V., Goulet, N., Roy, P., and Houle, P. 2011.
992 Tholeiitic to calc-alkaline cyclic volcanism in the Roy Group, Chibougamau area, Abitibi
993 Greenstone Belt – Revised Stratigraphy and implications for VHMS exploration.
994 *Canadian Journal of Earth Sciences*, **48**: 661-694.
- 995 Leclerc, F., Harris, L.B., Bédard, J.H., van Breemen, O., and Goulet, N. 2012. Structural and
996 Stratigraphic Controls on Magmatic, Volcanogenic, and Shear Zone-Hosted
997 Mineralization in the Chapais-Chibougamau Mining Camp, Northeastern Abitibi, Canada.
998 *Economic Geology*, **107**: 963-989.
- 999 Leclerc, F., Talla Takam, F., and Guemache, M. 2015. Géologie – Lac à l'Eau Jaune. Ministère
1000 de l'Energie et des Ressources Naturelles, Québec, carte CG-2015-04.
- 1001 Leclerc, F., Roy, P., Houle, P., Pilote, P., Bédard, J.H., Harris, L.B., McNicoll, V., van Breemen,
1002 O., David, J., and Goulet, N. 2017. Géologie de la région de Chibougamau. Ministère des
1003 Ressources Naturelles et de la Faune, Québec, RG 2015-03.
- 1004 Legault, M. 2003. Environnement métallogénique du couloir de Fancamp avec emphase sur les
1005 gisements aurifères de Chevrier, région de Chibougamau, Québec. Ph.D. in Mineral
1006 Resources, Université du Québec à Chicoutimi, Chicoutimi, Quebec.
- 1007 Lépine, S. 2009. Le gîte à Au-Cu-Mo de MOP-II (Chibougamau, Québec) : un porphyre Archéen
1008 déformé. M.Sc. in Earth Sciences, Université du Québec à Montréal, Montréal.
- 1009 Liou, P., and Guo, J. 2019. Generation of Archaean TTG Gneisses Through Amphibole-
1010 Dominated Fractionation. *Journal of Geophysical Research: Solid Earth*, **124**: 3605-3619.

- 1011 Locock, A.J. 2014. An Excel spreadsheet to classify chemical analyses of amphiboles following
1012 the IMA 2012 recommendations. *Computers & Geosciences*, **62**: 1-11.
- 1013 Ludden, J.N., Francis, D., and Allard, G.O. 1984. The geochemistry and evolution of the volcanic
1014 rocks of the Chibougamau region of the Abitibi metavolcanic belt. *In* Chibougamau-
1015 stratigraphy and mineralization. *Edited by* J. Guha, Chown, E.H., Westmount. pp. 20-34.
- 1016 Ludden, J.N., and Hubert, C. 1986. Geologic evolution of the late Archean Abitibi greenstone
1017 belt of Canada. *Geology*, **14**: 707-711.
- 1018 Martin, H., Smithies, R.H., Rapp, R., Moyen, J.-F., and Champion, D. 2005. An overview of
1019 adakite, tonalite-trondhjemitic-granodiorite (TTG), and sanukitoid: Relationships and
1020 some implications for crustal evolution. *Lithos*, **79**: 1-24.
- 1021 Martin, H., Moyen, J.-F., Guitreau, M., Blichert-Toft, J., and Le Pennec, J.-L. 2014. Why
1022 Archean TTG cannot be generated by MORB melting in subduction zones. *Lithos*, **198**-
1023 **199**: 1-13.
- 1024 Mathieu, L., and Racicot, D. 2019. Petrogenetic Study of the Multiphase Chibougamau Pluton:
1025 Archean Magmas Associated with Cu-Au Magmato-Hydrothermal Systems. *Minerals*, **9**:
1026 174.
- 1027 Mathieu, L., Crépon, A., and Kontak, D.J. 2020a. Tonalite-Dominated Magmatism in the Abitibi
1028 Subprovince, Canada, and Significance for Cu-Au Magmatic-Hydrothermal Systems.
1029 *Minerals*, **10**: 242.
- 1030 Mathieu, L., Snyder, B.D., Bedeaux, P., Cheraghi, S., Lafrance, B., Thurston, P., and Sherlock,
1031 R. 2020b. Deep into the Chibougamau area, Abitibi greenstone belt: structure of a
1032 Neoproterozoic crust revealed by seismic reflection profiling. *Tectonics*, **39**: 10.
- 1033 Mattinson, J. 2005. Zircon U-Pb chemical abrasion (CA-TIMS) method: Combined annealing
1034 and multi-step partial dissolution analysis for improved precision and accuracy of zircon
1035 ages. *Chemical Geology*, **220**: 47-66.
- 1036 Menand, T. 2008. The mechanics and dynamics of sills in layered elastic rocks and their
1037 implications for the growth of laccoliths and other igneous complexes. *Earth & Planetary
1038 Science Letters*, **267**: 93-99.
- 1039 Middlemost, E.A.K. 1985. *Magmas and Magmatic Rocks: an introduction to igneous petrology*.
1040 Longman, London, New York.
- 1041 Midra, R., Chown, E.H., and Tait, L. 1992. Géologie de la région du lac Dickson (bande
1042 Caopatina-Desmaraisville). Ministère de l'Énergie et des Ressources du Québec, MB 91-
1043 30 (cf. ET 90-01).
- 1044 Mole, D.R., Fiorentini, M.L., Cassidy, K.F., Kirkland, C.L., Thebaud, N., McCuaig, T.C.,
1045 Doublier, M.P., Romano, S.S., Maas, R., Belousova, E.A., Barnes, S.J., and Miller, J.
1046 2013. Crustal evolution, intra-cratonic architecture and the metallogeny of an Archean

- 1047 craton. *In* Ore deposits in an evolving Earth. *Edited by* G.R.T. Jenkin, Lusty, P.A.J.,
1048 McDonald, I., Smith, M.P., Boyce, A.J., and Wilkinson, J.J. Geological Society, London,
1049 Special Publication. p. 393.
- 1050 Mole, D.R., Thurston, P., Marsh, J.H., Stern, R.A., Ayer, J., Martin, L.A.J., and Lu, Y.J. 2021.
1051 The formation of Neoproterozoic continental crust in the south-east Superior Craton by two
1052 distinct geodynamic processes. *Precambrian Research*, **356**: 160104.
- 1053 Mortensen, J.K. 1993a. U-Pb geochronology of the eastern Abitibi subprovince. Part 1 :
1054 Chibougamau-Matagami-Joutel region. *Canadian Journal of Earth Sciences*, **30**: 11-28.
- 1055 Mortensen, J.K. 1993b. U-Pb geochronology of the Lapparent Massif, northeastern Abitibi belt:
1056 basement or synvolcanic pluton ? *Canadian Journal of Earth Sciences*, **30**: 42-47.
- 1057 Moyen, J.-F., Nédélec, A., Martin, H., and Jayananda, M. 2003. Syntectonic granite emplacement
1058 at different structural levels : the Closepet granite, South India. *Journal of Structural*
1059 *Geology*, **25**: 611-631.
- 1060 Moyen, J.-F., and Stevens, G. 2006. Experimental constraints on TTG petrogenesis: Implications
1061 for Archean geodynamics. *Geophysical Monograph Series*, **164**: 149-175.
- 1062 Moyen, J.-F. 2011. The composite Archean gneisses: Petrological significance, and
1063 evidence for a non-unique tectonic setting for Archean crustal growth. *Lithos*, **123**: 21-
1064 36.
- 1065 Moyen, J.-F., and Martin, H. 2012. Forty years of TTG research. *Lithos*, **148**: 312-336.
- 1066 Moyen, J.-F. 2019. Archean granitoids: classification, petrology, geochemistry and origin.
1067 Geological Society, London, Special Publications, **489**.
- 1068 Mueller, W., Chown, E.H., Sharma, K.N.M., Tait, L., and Rocheleau, M. 1989. Paleogeographic
1069 and paleotectonic evolution of a basement-controlled Archean supracrustal sequence,
1070 Chibougamau-Caopatina, Quebec. *The Journal of Geology*, **94**: 339-420.
- 1071 Mutch, E.J.F., Blundy, J.D., Tattitch, B.C., Cooper, E.J., and Brooker, R.A. 2016. An
1072 experimental study of amphibole stability in low-pressure granitic magmas and a revised
1073 Al-in-hornblende geobarometer. *Contributions to Mineralogy & Petrology*, **171**: 25.
- 1074 Nagel, T.J., Hoffmann, J.E., and Münker, C. 2012. Generation of Eoarchean tonalitic-
1075 trondhjemitic-granodiorites from thickened mafic arc crust. *Geology*, **40**: 375-378.
- 1076 Nair, R., and Chacko, T. 2008. Role of oceanic plateaus in the initiation of subduction and origin
1077 of continental crust. *Geology*, **36**: 583-586.
- 1078 Nesse, W.D. 2000. *Introduction to Mineralogy*. Oxford University Press.

- 1079 Oba, T. 2013. Phase relationship of $\text{Ca}_2\text{Mg}_3\text{Al}_2\text{Si}_6\text{Al}_2\text{O}_{22}(\text{OH})_2$ – $\text{Ca}_2\text{Mg}_3\text{Fe}_{23+}\text{Si}_6\text{Al}_2$
 1080 $\text{O}_{22}(\text{OH})_2$ join at high temperature and high pressure – the stability of tschermakite.
 1081 *Journal of the Faculty of Sciences, Hokkaido University, Series IV*, **18**: 339-350.
- 1082 Petford, N., Lister, J.R., and Kerr, R.C. 1994. The ascent of felsic magmas in dykes. *Lithos*, **32**:
 1083 161-168.
- 1084 Pilote, P., Dion, C., Joanisse, A., David, J., Machado, N., Kirkham, R.V., and Robert, F. 1997.
 1085 Géochronologie des minéralisations d'affiliation magmatique de l'Abitibi, secteurs
 1086 Chibougamau et de Troilus-Frotet: implications géotectoniques. *In* Programme et
 1087 résumés, Séminaire d'information sur la recherche géologique. Ministère des Ressources
 1088 Naturelles, DV 97-03, Québec. p. 47.
- 1089 Pilote, P., Robert, F., Kirkham, R.V., Daigneault, R., and Sinclair, W.D. 1998. Partie D –
 1090 Géologie économique. Minéralisations de type porphyrique et filoniennes dans le
 1091 Complexe du Lac Doré – les secteurs du Lac Clark et de l'Île Merrill. *In* Géologie et
 1092 métallogénie du district minier de Chapais-Chibougamau : nouvelle vision du potentiel de
 1093 découverte. *Edited by* P. Pilote. Ministère de l'Energie et des Ressources Naturelles, DV
 1094 98-03. pp. 71-90.
- 1095 Polat, A., Frei, R., Longstaffe, F.J., and Woods, R. 2018. Petrogenetic and geodynamic origin of
 1096 the Neoproterozoic Doré Lake Complex, Abitibi subprovince, Superior Province, Canada.
 1097 *International Journal of Earth Sciences* **107**: 811-843.
- 1098 Potvin, R. 1991. Etude volcanologique du centre felsique du Lac des Vents, région de
 1099 Chibougamau. M.Sc. in Earth Sciences, Université du Québec à Chicoutimi, Chicoutimi,
 1100 Quebec.
- 1101 Powell, W.G., Carmichael, D.M., and Hodgson, C.J. 1995. Conditions and timing of
 1102 metamorphism in the southern Abitibi greenstone belt, Quebec. *Canadian Journal of Earth*
 1103 *Sciences*, **32**: 787-805.
- 1104 Prowatke, S., and Klemme, S. 2006. Trace element partitioning between apatite and silicate
 1105 melts. *Geochimica et Cosmochimica Acta*, **70**: 4513-4527.
- 1106 Putirka, K. 2016. Amphibole thermometers and barometers for igneous systems and some
 1107 implications for eruption mechanisms of felsic magmas at arc volcanoes. *American*
 1108 *Mineralogist*, **101**: 841-858.
- 1109 Racicot, D., Chown, E.H., and Hanel, T. 1984. Plutons of the Chibougamau-Desmaraisville Belt :
 1110 A preliminary survey. . *In* Chibougamau – Stratigraphy and mineralization. *Edited by* J.
 1111 Guha and E.H. Chown. Canadian Institute of Mining and Metallurgy, Special Volume 34.
 1112 pp. 178-197.
- 1113 Rey, P.F., Philippot, P., and Thebaud, N. 2003. Contribution of mantle plumes, crustal thickening
 1114 and greenstone blanketing to the 2.75-2.65 Ga global crisis. *Precambrian Research*, **127**:
 1115 43-60.

- 1116 Richards, J. 2018. A shake-up in the porphyry world ? *Economic Geology*, **113**: 1225-1233.
- 1117 Roman-Berdiel, T., Gapais, D., and Brun, J.P. 1995. Analogue models of laccolith formation.
1118 *Journal of Structural Geology*, **17**: 1337-1346.
- 1119 Roman, A., and Arndt, N. 2020. Differentiated Archean oceanic crust: Its thermal structure,
1120 mechanical stability and a test of the sagduction hypothesis. *Geochemica et Cosmochemica*
1121 *Acta*, **278**: 65-77.
- 1122 Sanematsu, K., Ejima, T., Kon, Y., Manaka, T., Zaw, S., Morita, S., and Seo, Y. 2016.
1123 Fractionation of rare-earth elements during magmatic differentiation and weathering of
1124 calc-alkaline granites in southern Myanmar. *Mineralogical Magazine*, **80**: 77-102.
- 1125 Schmidt, M.W., and Poli, S. 2004. Magmatic Epidote. *Reviews in Mineralogy & Geochemistry*,
1126 **56**: 339-430.
- 1127 Sigéom. 2018. Système d'Information Géominière à référence spatiale. Regroupement des
1128 données géoscientifiques aux échelles 1 : 20 000 et 1 : 50 000. Available from
1129 <http://sigeom.mines.gouv.qc.ca>.
- 1130 Sillitoe, R.H. 2010. Porphyry copper systems. *Economic Geology*, **105**: 3-41.
- 1131 Sinclair, W.D. 2017. Porphyry deposits. *In* *Mineral Deposits of Canada: A Synthesis of Major*
1132 *Deposit-Types, District Metallogeny, the Evolution of Geological Provinces, and*
1133 *Exploration Methods*. Goodefellow, W.D., Geological Association of Canada, Mineral
1134 Deposits Division, Special Publication, 5. pp. 223-243.
- 1135 Smithies, R.H. 2000. The Archaean tonalite-trondhjemite-granodiorite (TTG) series is not an
1136 analogue of Cenozoic adakite. *Earth & Planetary Science Letters*, **182**: 115-125.
- 1137 Szopa, K., Gawęda, A., Müller, A., and Sikorska, M. 2013. The petrogenesis of granitoid rocks
1138 unusually rich in apatite in the Western Tatra Mts. (S-Poland, Western Carpathians).
1139 *Mineralogy and Petrology*, **107**: 609-627.
- 1140 Tait, L., Pilote, P., and Chown, E.H. 1986. Géologie de la région du lac à l'Eau Jaune. Ministère
1141 de l'Energie et des Ressources, Québec, DP 86-09.
- 1142 Tait, L. 1992. Géologie de la région du lac à l'Eau Jaune (Territoire-du-Nouveau-Québec).
1143 Ministère de l'Energie et des Ressources du Québec, MB 91-29 (cf. ET 90-08).
- 1144 Turner, F.J. 1981. *Metamorphic petrology – Mineralogical, field, and tectonic aspects*. McGraw-
1145 Hill Book Company.
- 1146 Van Kranendonk, M.J., Smithies, R.H., Griffin, W.L., HUSTON, D.L., Hickman, A.H., and
1147 Champion, D. 2015. Making it thick: a volcanic plateau origin of Palaeoarchean
1148 continental lithosphere of the Pilbara and Kaapvaal cratons. *Geological Society, London*,
1149 *Special Publication*, **389**: 83-111.

- 1150 Vigneresse, J.L., and Clemens, J.D. 2000. Granitic magma ascent and emplacement : neither
1151 diapirism nor neutral buoyancy. *In* Shale and Igneous Diapirs in and around Europe.
1152 Geological Society, London, Special Publication, 174. pp. 1-19.
- 1153 Whalen, J.B., Percival, J.A., McNicoll, V., and Longstaffe, F.J. 2002. A mainly crustal origin for
1154 tonalitic granitoid rocks, Superior Province, Canada: Implications for late Archean
1155 tectonomagmatic processes. *Journal of Petrology*, **43**: 1551-1570.
- 1156 Wyman, D., Kerrich, R., and Polat, A. 2002. Assembly of Archean cratonic mantle Lithosphere
1157 and crust: Plume-arc interaction in the Abitibi-Wawa subduction-accretion complex.
1158 *Precambrian Research*, **115**: 37-62.
- 1159

1160 **Figure captions**

1161 Fig. 1. Simplified geological map of the Chibougamau area, modified from the Quebec Ministry
1162 of Energy and Natural Resources (*Ministère de l'Énergie et des Ressources Naturelles*, MERN)
1163 map (SIGÉOM 2018, 2019). The map projection system is UTM NAD83 Zone 18N.

1164 Fig. 2. Simplified geological map of the Eau Jaune Complex area, modified from the MERN map
1165 (SIGÉOM 2018, 2019), showing the locations of the 155 outcrops visited as part of this study.
1166 The lithogeochemical samples analyzed as part of this study and those analysed earlier by the
1167 Quebec MERN are displayed with a bold white outline. The colours correspond to the main
1168 lithology of each outcrop based upon its mineralogy and its geochemical characteristics when
1169 available. The map projection system is UTM NAD83 Zone 18N.

1170 Fig. 3. Detailed mapping, using a 1 m × 1 m grid, of the main trench of the Moly-Desgagné
1171 showing and the sample locations, created as part of this study. Three of the six Eau Jaune
1172 Complex magmatic phases (1a, 1b, and 2b) are displayed here and crosscutting relationships are
1173 clearly seen. Note that the rocks of this outcrop underwent strong deformation events and were
1174 later affected by hydrothermal alteration.

1175 Fig. 4. Pictures (a, c, and e) and microphotographs (b, d, and f) in crossed nicols showing typical
1176 textures of the Eau Jaune Complex magmatic phases. (a and b) granodiorite 1a; (c and d) quartz
1177 diorite-tonalite 1b; (e and f) tonalite 2a. Act, actinolite; Afs, alkali feldspar; Cb, carbonates; Hbl,
1178 hornblende; Mc, microcline; Myr, myrmekite; Qz, quartz; Ser, sericite; Wmca, white micas.

1179 Fig. 5. Pictures (a, c, and e) and microphotographs (b, d, and f) in crossed nicols showing typical
1180 textures of the Eau Jaune Complex magmatic phases. (a and b) biotite-tonalite 2b; (c and d)

1181 trondhjemite 3a; (e and f) amphibole-rich diorite 3b. Aln, allanite; Bt, biotite; Chl, chlorite; Ep,
 1182 epidote; Hbl, hornblende; Pl, plagioclase; Qz, quartz; Ser, sericite.

1183 Fig. 6. Microphotographs in (a) polarized light and (b) polarized and analysed light displaying
 1184 hydrothermal alteration on the Moly-Desgagné showing. Aln, allanite; Cb, carbonates; Chl,
 1185 chlorite; Ep, epidote; Fsp, feldspar; Qz, quartz; Ser, sericite.

1186 Fig. 7. (a) Amphiboles from granodiorite 1b and amphibole-rich diorite 3b displayed on the
 1187 diagram for Ca amphiboles of Hawthorne et al. (2012). X and Y axes correspond, respectively, to
 1188 C- and A-site atomic compositions in apfu, with $C^* = \sum^C(\text{Al} + \text{Fe}^{3+} + \text{Mn}^{3+} + \text{Cr} + \text{V} + 2\text{Ti} +$
 1189 $2\text{Zr}) - {}^{\text{W}}\text{O} - {}^{\text{C}}\text{Li}$ and $A^* = \sum^A(\text{Li} + \text{Na} + \text{K} + 2\text{Ca} + 2\text{Pb})$; (b) Pressure versus temperature
 1190 diagram showing the results of calculations achieved using amphibole chemistry and the method
 1191 of Putirka (2016) for temperature and Anderson and Smith (1995) for pressure. Both methods are
 1192 designed for magmatic amphiboles. Error associated with pressure is ± 0.6 kbar.

1193 Fig. 8: Simplified geological map of the Eau Jaune Complex area, modified from the MERN
 1194 (SIGÉOM 2019) and showing the main foliation measured on visited outcrops as part of this
 1195 study. Trajectories of main foliation according to the work of Tait (1992), Midra et al. (1992) and
 1196 Leclerc et al. (2015), as available on the SIGÉOM (2019), are also represented. The map
 1197 projection system is UTM NAD83 Zone 18N. Legend is as in Fig. 2.

1198 Fig. 9. Rose diagrams of main foliation measurements taken during this study in (a) magmatic
 1199 phases of the Eau Jaune Complex and (b) mafic xenoliths. Note that we chose to represent
 1200 directions rather than strike/dip because very few dip measurements were possible as the outcrops
 1201 were planar.

Fig. 10. Pictures showing deformation of the Eau Jaune Complex . (a) Obatogamau metavolcanic mafic xenolith displaying a strong ductile (note elongate feldspar glomerocrysts), which is truncated by the quartz diorite tonalite 2b, and therefore developed before their incorporation into the EJC. (b) Compositional layering of a mafic xenolith (N100/76°) and foliation developed during amphibolite-condition metamorphism (N164/88°). (c) Strong banding inside the N100–N130 deformation corridor, visible on one outcrop of the Moly-Desgagné showing. (d) Epidote ball inside an amphibolite-grade metamorphosed mafic xenolith. The ball displays a delta-shape asymmetry suggesting a sinistral rotation.

Fig. 11. Major element diagrams displaying the 37 intrusive rocks from the Eau Jaune Complex analysed as part of this study (full symbols) and 31 analyses compiled from the SIGÉOM (2018) database. (a) Total alkali versus silica (TAS) diagram (Le Bas et al. 1986); Tonalite and trondhjemite fields after the TTG compilation in the Abitibi subprovince of Mathieu et al. (2020). (b) Alkali-iron-magnesium (AFM) diagram (Irvine and Baragar 1971), with $A = Na_2O + K_2O$, $F = FeOT$, and $M = MgO$, in %. (c) Na-Ca-K (%) triangle also displaying experimental liquids, as compiled by Moyen and Martin (2012).

Fig. 12. QAP diagram (Streckeisen 1976) for normative mineral compositions distributed using the method of Le Maitre (1976). Albite is only in plagioclase, as shown by petrographic observations and in situ microprobe analysis of feldspars. Legend as in Fig. 11. Differentiation series from Lameyre and Bowden (1982) are displayed as grey arrows: T, tholeiitic; A, alkaline. The calc-alkaline series is divided into: a, K-poor; b, intermediate; c, K-rich. Numbers refer to 1, quartzolite; 2, quartz-rich granitoids; 3, alkali feldspar granite; 4, syenogranite; 5, monzogranite; 6, granodiorite; 7, tonalite and trondhjemite; 8, quartz-alkali feldspar syenite; 9, quartz syenite;

1224 10, quartz monzonite; 11, quartz monzodiorite; 12, quartz diorite; 13, alkali feldspar syenite; 14,
1225 syenite; 15, monzonite; 16, monzodiorite; 17, diorite.

1226 Fig. 13. Harker diagrams (major and minor elements versus SiO_2) displaying the 37 samples from
1227 the Eau Jaune Complex analysed as part of this study and 31 analyses compiled from the
1228 SIGÉOM (2018) database. Trends are shown using grey dashed lines. Legend is as in Fig. 11.

1229 Fig. 14. Multi-element diagrams (a, c, and e) for immobile elements, as ordered after Pearce
1230 (2008) and rare-earth element (REE) diagrams (b, d, and f) normalized to the primitive mantle
1231 (Hoffmann 1988). The diagrams display the 37 intrusive rocks from the Eau Jaune Complex
1232 analysed as part of this study and 31 analyses compiled from the SIGÉOM (2018) database. Each
1233 line corresponds to a sample and an area corresponds to minimum and maximum values.

1234 Fig. 15. Concordia diagrams showing CA-ID-TIMS results from U–Pb zircon analyses for (a)
1235 sample 18UCD0137, a biotite tonalite from Eau Jaune Complex unit 2b, host to the Moly-
1236 Desagné Mo showing, and (b) sample 18UCD0043, a trondhjemite from EJC unit 3a. Error
1237 ellipses and calculated age are shown at the 2σ level of uncertainty.

1238 Fig. 16. Conceptual diagrams illustrating the different steps leading to the Eau Jaune Complex
1239 (EJC) emplacement and its features. (a) Location of the cross-section. (b) Injection of EJC1
1240 magma along a major structure, inducing contact metamorphism and forming a sill at
1241 stratigraphic discontinuities. (c) Growth of the sill by inflation to form a laccolith and initiate the
1242 overlapping of the Chrissie and Des Vents formations. The laccolith incorporates metamorphosed
1243 xenoliths and locally deforms the surrounding volcanic rocks (details in inset). (d) Intrusion of
1244 phases EJC2 and EJC3, also forming laccoliths. These intrusions metamorphose and deform the
1245 surrounding rocks. The laccoliths incorporate volcanic xenoliths, which were metamorphosed

1246 and deformed by the previous injections. Details in inset: xenoliths incorporated in phase EJC2,
1247 which were metamorphosed and deformed by growth of the EJC1 laccolith. Vertical scale is
1248 exaggerated.

Draft

1 Table 1: Mean and standard deviation (σ) of major and trace elements within the six major
2 intrusive phases of the EJC. (La/Yb)_N corresponds to measured La and Yb values normalised to
3 the primitive mantle (Hoffmann 1988).

	1a <i>n</i> = 6	1b <i>n</i> = 18	2a <i>n</i> = 13	2b <i>n</i> = 19	3a <i>n</i> = 6	3b <i>n</i> = 4
Major elements (wt% anhydrous)						
SiO ₂	62.88 (9.24)	59.88 (5.73)	71.62 (2.89)	69.28 (2.92)	69.32 (1.32)	54.43 (4.34)
TiO ₂	0.40 (0.17)	0.73 (0.21)	0.24 (0.15)	0.37 (0.10)	0.31 (0.04)	1.05 (0.28)
Al ₂ O ₃	14.54 (2.58)	15.77 (1.33)	15.14 (0.95)	15.44 (1.09)	15.78 (0.75)	19.25 (1.15)
Fe ₂ O _{3T}	4.31 (1.33)	7.39 (2.08)	2.06 (1.18)	3.31 (0.89)	3.27 (0.69)	8.65 (1.62)
MnO	0.13 (0.18)	0.10 (0.04)	0.03 (0.02)	0.05 (0.01)	0.04 (0.01)	0.11 (0.02)
MgO	1.41 (0.57)	3.15 (1.64)	0.59 (0.43)	1.10 (0.47)	0.90 (0.15)	3.00 (0.63)
CaO	7.91 (9.77)	5.78 (2.16)	2.45 (0.77)	3.38 (0.78)	3.90 (0.31)	7.75 (1.34)
Na ₂ O	3.74 (1.20)	3.84 (0.88)	4.85 (0.38)	4.46 (0.42)	4.74 (0.39)	4.16 (0.38)
K ₂ O	1.27 (0.45)	0.95 (0.53)	2.17 (0.88)	1.42 (0.38)	0.90 (0.11)	0.44 (0.12)
P ₂ O ₅	0.12 (0.06)	0.21 (0.10)	0.09 (0.07)	0.12 (0.04)	0.09 (0.01)	0.23 (0.06)
Trace elements (ppm)						
Ba	286.7 (84.0)	255.6 (130.8)	618.2 (251.5)	417.6 (130.1)	362.0 (39.2)	145.4 (50.9)
Rb	35.3 (13.7)	24.9 (14.7)	47.0 (17.6)	45.3 (10.9)	25.2 (4.8)	9.7 (4.0)
Sr	301.8 (127.7)	230.3 (101.2)	423.9 (207.6)	358.9 (137.1)	426.8 (69.0)	392.0 (37.1)
Cr	35.0 (23.7)	62.3 (33.8)	19.4 (15.8)	21.7 (20.4)	33.5 (16.3)	24.5 (9.0)
Ni	25.7 (24.5)	41.3 (37.1)	5.4 (4.9)	11.7 (8.2)	9.7 (5.2)	23.3 (3.6)
Th	1.8 (0.8)	2.3 (1.9)	3.0 (2.3)	3.8 (1.1)	0.9 (0.4)	0.7 (0.32)
Nb	4.9 (2.3)	8.0 (3.0)	4.1 (2.1)	5.2 (1.3)	2.8 (0.5)	5.3 (0.4)
Ta	0.5 (0.1)	0.7 (0.4)	0.9 (0.8)	2.2 (2.1)	0.3 (0.1)	0.5 (0.1)
Zr	142.3 (53.3)	176.3 (82.1)	108.0 (59.0)	154.3 (47.9)	108.5 (10.0)	174.0 (45.7)
Hf	3.3 (1.3)	4.1 (1.8)	2.8 (1.1)	3.8 (1.0)	2.8 (0.3)	4.0 (0.9)
Y	7.9 (1.5)	21.0 (6.7)	5.6 (4.0)	9.3 (2.5)	5.4 (1.5)	18.9 (1.4)
La	12.4 (5.7)	18.1 (5.3)	17.0 (15.8)	20.8 (7.6)	9.1 (2.3)	8.3 (1.0)
Ce	25.0 (11.9)	39.5 (12.7)	33.6 (31.2)	40.6 (14.8)	17.1 (4.5)	19.7 (2.4)
Pr	2.9 (1.4)	4.8 (1.5)	3.8 (3.4)	4.5 (1.6)	1.8 (0.5)	2.8 (0.3)
Nd	11.0 (5.2)	19.8 (6.2)	13.5 (12.0)	16.1 (5.6)	7.2 (1.5)	13.1 (1.9)
Sm	2.2 (1.0)	4.3 (1.1)	2.2 (1.7)	2.7 (0.7)	1.4 (0.3)	3.5 (0.4)
Eu	0.7 (0.2)	1.1 (0.3)	0.6 (0.3)	0.8 (0.2)	0.6 (0.1)	1.1 (0.2)
Gd	1.8 (0.5)	4.0 (1.1)	1.6 (0.1)	2.1 (0.5)	1.2 (0.3)	3.4 (0.4)
Tb	0.3 (0.1)	0.6 (0.2)	0.2 (0.2)	0.3 (0.1)	0.2 (0.1)	0.6 (0.1)
Dy	1.5 (0.3)	3.7 (1.0)	1.1 (0.9)	1.6 (0.5)	1.0 (0.30)	3.5 (0.5)
Ho	0.3 (0.1)	0.8 (0.2)	0.2 (0.2)	0.3 (0.1)	0.2 (0.1)	0.7 (0.1)
Er	0.8 (0.2)	2.2 (0.7)	0.5 (0.4)	0.9 (0.3)	0.5 (0.5)	2.0 (0.1)
Tm	0.1 (0.0)	0.3 (0.1)	0.1 (0.1)	0.1 (0.1)	0.1 (0.0)	0.3 (0.0)
Yb	0.8 (0.1)	2.3 (0.8)	0.5 (0.3)	0.9 (0.3)	0.5 (0.2)	1.8 (0.2)
Lu	0.1 (0.0)	0.4 (0.1)	0.1 (0.1)	0.1 (0.10)	0.1 (0.0)	0.3 (0.0)

$(\text{La/Yb})_{\text{N}}$	10.6 (4.4)	5.8 (1.9)	27.0 (19.6)	17.9 (8.8)	13.1 (4.0)	3.18 (0.35)
-----------------------------	------------	-----------	-------------	------------	------------	-------------

Draft

5 Table 2: Summary and characteristics of the deformation episodes of the Eau Jaune Complex area. Three local deformation episodes
6 are specific to the EJC area and occurred prior to the regional deformation events.

	Ages	Deformation episodes	Direction	Rocks recording the deformation	Movement	Intensity	Style	Conditions
Synvolcanic period	2800–2721 Ma*	Pre-EJC local deformation	Non determined	Metavolcanic xenoliths	--	Strong	Ductile	Amphibolite facies
		EJC emplacement (ca.2724 Ma)	Along rims	Metavolcanic xenoliths	--	Moderate	Ductile	Amphibolite facies
			WNW-ESE (original dip probably not the one observed)	EJC deformation zone and sharp western limit of EJC3	--	Strong	Ductile	Amphibolite facies
		EJC local deformation 1	N-S (dip not observed)	EJC intrusive phases	--	Faint	Ductile	Amphibolite facies
		EJC local deformation 2	E-W (subvertical)	EJC intrusive phases	--	Moderate	Ductile	Amphibolite facies
			WNW-ESE (folded – variable dip)	EJC intrusive phases and metavolcanic xenoliths in EJC deformation zone	Sinistral	Strong	Ductile	Amphibolite facies
?	?	(regional D ₁ ?)	Exhumation of the Eau Jaune Complex					
Syntectonic period	2700–2680 Ma**	(regional D ₁ ?) + regional D ₂	WNW-ESE (folded – variable dip)	EJC intrusive phases and metavolcanic xenoliths in EJC deformation zone	Dextral (reactivation)	?	Ductile	Greenschist facies
			E-W (subvertical)	Only observed on Moly-Desgagné showing	--	Faint	Fragile	Greenschist facies

7 *Ludden et al. (1984); Leclerc et al. (2011); Mathieu et al. (2020)

8 **Daigneault et al. (1990); Chown and Mueller (1992); Daigneault (1996); Leclerc et al. (2015); Mathieu et al. (2020)

9 Table 3: Chemical composition of the six intrusive phases of the EJC compared with that of
 10 tonalite-trondhjemite-granodiorite (TTG) suites, as reviewed by Moyen and Martin (2012).

Criteria	Low REE TTG	Med. REE TTG	High REE TTG	Not TTG
Criterion 1 66 samples in the data set		$\text{SiO}_2 > 64 \text{ wt\%}$		Other values
	46 samples (5 1a; 3 1b; 13 2a = all; 19 2b = all; 6 3a = all)			Most samples from 1a and 1b and all 3b
Criterion 2 46 samples		$\text{Fe}_2\text{O}_{3\text{T}} + \text{MgO} + \text{MnO} + \text{TiO}_2 < 5 \text{ wt\%}$ 0.50–4.94 ($n = 26$)		Other values 5.01–8.41 ($n = 20$)
Criterion 3		$0.5 < \text{K}_2\text{O} < 2 \text{ wt\%}$ 0.55–1.79 ($n = 37$)		Other values 2.34–3.77 ($n = 9$)
Criterion 4	$\text{Na}_2\text{O} > 5 \text{ wt\%}$ 5.04–5.71 ($n = 5$)	$4 < \text{Na}_2\text{O} < 6 \text{ wt\%}$ 4–5 ($n = 34$); 5–6 ($n = 5$)		Other values 3.02–3.98 ($n = 6$)
Criterion 5		$0.3 < \text{K}_2\text{O}/\text{Na}_2\text{O} < 0.6$ 0.31–0.50 ($n = 19$)		Other values 0.16–0.30 ($n = 22$); 0.60–0.85 ($n = 5$)
Criterion 6 at 70 wt% SiO_2	$\text{Al}_2\text{O}_3 > 16 \text{ wt\%}$ 16.14–18.44 ($n = 21$)	$\text{Al}_2\text{O}_3 > 15 \text{ wt\%}$ 15.02–18.437 ($n = 30$)	$\text{Al}_2\text{O}_3 > 14 \text{ wt\%}$ 14.15–18.44 ($n = 33$)	Other values 11.64–13.98 ($n = 7$)
Criterion 7	$\text{Yb} < 1 \text{ ppm}$ 0.1–1.0 ($n = 31$)	$\text{Yb} < 1.5 \text{ ppm}$ 0.1–1.3 ($n = 43$)	$\text{Yb} < 2 \text{ ppm}$ 0.1–2.0 ($n = 44$)	Other values 2.3–2.9 ($n = 2$)
Criterion 8		$(\text{La}/\text{Yb})_{\text{N}} > 15$ 15.5–58.5 ($n = 23$)		Other values 2.5–14.3 ($n = 23$)
Criterion 9		$\text{La} > 10 \text{ ppm}$ 10.1–32.9 ($n = 34$)		Other values 2.6–9.2 ($n = 12$)
Criterion 10		Negative Ti, Nb and Ta anomalies Every sample ($n = 59$) No Sr and Eu negative anomalies Most samples, except 14 samples (2 (1a), 5 (1a), 6 (3a), and 1 (3b))		Other values Positive Eu anomaly ($n = 14$)
Criterion 11	$50 \leq \text{Sr}/\text{Y} < 500$ 50.7–288.9 ($n = 19$)	$20 < \text{Sr}/\text{Y} < 200$ 21.2–180.5 ($n = 41$)	$10 < \text{Sr}/\text{Y} < 50$ 15.8–47.8 ($n = 24$)	Other values 6–4–9.3 ($n = 2$)

11

12 References cited in table caption

- 13 Chown, E.H., and Mueller, W. 1992. Basement Influence on the Supracrustal and Plutonic
14 Evolution and Deformation of an Archean Greenstone Belt. *In* Basement Tectonics 7 -
15 Proceedings of the Seventh International Conference on Basement Tectonics. *Edited by* R.
16 Mason. Robert Mason, Kingston, Ontario, Canada. pp. 465-476.
- 17 Daigneault, R., St-Julien, P., and Allard, G.O. 1990. Tectonic evolution of the northeast portion
18 of the Archean Abitibi greenstone belt, Chibougamau area, Quebec. *Canadian Journal of*
19 *Earth Sciences*, **27**: 1714-1736.
- 20 Daigneault, R. 1996. Couloirs de déformation de la Sous-Province de l'Abitibi. Ministère des
21 Ressources Naturelles, MB 96-33, Québec.
- 22 Hoffmann, A.W. 1988. Chemical differentiation of the Earth. *Earth & Planetary Science Letters*,
23 **90**: 297-314.
- 24 Leclerc, F., Bédard, J.H., Harris, L.B., McNicoll, V., Goulet, N., Roy, P., and Houle, P. 2011.
25 Tholeiitic to calc-alkaline cyclic volcanism in the Roy Group, Chibougamau area, Abitibi
26 Greenstone Belt – Revised Stratigraphy and implications for VHMS exploration.
27 *Canadian Journal of Earth Sciences*, **48**: 661-694.
- 28 Leclerc, F., Talla Takam, F., and Guemache, M. 2015. Géologie – Lac à l'Eau Jaune. Ministère
29 de l'Energie et des Ressources Naturelles, Québec, carte CG-2015-04.
- 30 Ludden, J.N., Francis, D., and Allard, G.O. 1984. The geochemistry and evolution of the volcanic
31 rocks of the Chibougamau region of the Abitibi metavolcanic belt. *In* Chibougamau-
32 stratigraphy and mineralization. *Edited by* J. Guha, Chown, E.H., Westmount. pp. 20-34.
- 33 Mathieu, L., Snyder, B.D., Bedeaux, P., Cheraghi, S., Lafrance, B., Thurston, P., and Sherlock,
34 R. 2020. Deep into the Chibougamau area, Abitibi greenstone belt: structure of a
35 Neoproterozoic crust revealed by seismic reflection profiling. *Tectonics*, **39**: 10.
- 36 Moyen, J.-F., and Martin, H. 2012. Forty years of TTG research. *Lithos*, **148**: 312-336.

37

Fig. 1. Simplified geological map of the Chibougamau area, modified from the Quebec Ministry of Energy and Natural Resources (*Ministère de l'Énergie et des Ressources Naturelles*, MERN) map (SIGÉOM 2018, 2019). The map projection system is UTM NAD83 Zone 18N.

Fig. 2. Simplified geological map of the Eau Jaune Complex area, modified from the MERN map (SIGÉOM 2018, 2019), showing the locations of the 155 outcrops visited as part of this study. The lithogeochemical samples analyzed as part of this study and those analysed earlier by the Quebec MERN are displayed with a bold white outline. The colours correspond to the main lithology of each outcrop based upon its mineralogy and its geochemical characteristics when available. The map projection system is UTM NAD83 Zone 18N.

194x163mm (300 x 300 DPI)

Fig. 3. Detailed mapping, using a 1 m × 1 m grid, of the main trench of the Moly-Desgagné showing and the sample locations. Three of the six Eau Jaune Complex magmatic phases (1a, 1b, and 2b) are displayed here and crosscutting relationships are clearly seen. Note that the rocks of this outcrop underwent strong deformation events and were later affected by hydrothermal alteration.

238x183mm (300 x 300 DPI)

Fig. 4. Pictures (a, c, and e) and microphotographs (b, d, and f) in crossed nicols showing typical textures of the Eau Jaune Complex magmatic phases. (a and b) granodiorite 1a; (c and d) quartz diorite-tonalite 1b; (e and f) tonalite 2a. Act, actinolite; Afs, alkali feldspar; Cb, carbonates; Hbl, hornblende; Mc, microcline; Myr, myrmekite; Qz, quartz; Ser, sericite; Wmca, white micas.

Fig. 5. Pictures (a, c, and e) and microphotographs (b, d, and f) in crossed nicols showing typical textures of the Eau Jaune Complex magmatic phases. (a and b) biotite-tonalite 2b; (c and d) trondhjemite 3a; (e and f) amphibole-rich diorite 3b. Aln, allanite; Bt, biotite; Chl, chlorite; Ep, epidote; Hbl, hornblende; Pl, plagioclase; Qz, quartz; Ser, sericite.

Fig. 6. Microphotographs in (a) polarized light and (b) polarized and analysed light displaying hydrothermal alteration on the Moly-Desgagné showing. Aln, allanite; Cb, carbonates; Chl, chlorite; Ep, epidote; Fsp, feldspar; Qz, quartz; Ser, sericite.

Fig. 7. (a) Amphiboles from granodiorite 1b and amphibole-rich diorite 3b displayed on the diagram for Ca amphiboles of Hawthorne et al. (2012). X and Y axes correspond, respectively, to C- and A-site atomic compositions in apfu, with $C^* = \sum^C(\text{Al} + \text{Fe}^{3+} + \text{Mn}^{3+} + \text{Cr} + \text{V} + 2\text{Ti} + 2\text{Zr}) - {}^W\text{O} - {}^C\text{Li}$ and $A^* = \sum^A(\text{Li} + \text{Na} + \text{K} + 2\text{Ca} + 2\text{Pb})$; (b) Pressure versus temperature diagram showing the results of calculations achieved using amphibole chemistry and the method of Putirka (2016) for temperature and Anderson and Smith (1995) for pressure. Both methods are designed for magmatic amphiboles. Error associated with pressure is ± 0.6 kbar.

177x74mm (150 x 150 DPI)

Fig. 8: Simplified geological map of the Eau Jaune Complex area, modified from the MERN (SIGÉOM 2019) and showing the main foliation measured on visited outcrops as part of this study. Legend is as in Fig. 2.

Fig. 9. Rose diagrams of main foliation measurements taken during this study in (a) magmatic phases of the Eau Jaune Complex and (b) mafic xenoliths. Note that we chose to represent directions rather than strike/dip because very few dip measurements were possible as the outcrops were planar.

Fig. 10. Pictures showing deformation of the Eau Jaune Complex . (a) Obatogamau metavolcanic mafic xenolith displaying a strong ductile (note elongate feldspar glomerocrysts), which is truncated by the quartz diorite tonalite 2b, and therefore developed before their incorporation into the EJC. (b) Compositional layering of a mafic xenolith (N100/76°) and foliation developed during amphibolite-condition metamorphism (N164/88°). (c) Strong banding inside the N100–N130 deformation corridor, visible on one outcrop of the Moly-Desgagné showing. (d) Epidote ball inside an amphibolite-grade metamorphosed mafic xenolith. The ball displays a delta-shape asymmetry suggesting a sinistral rotation.

Fig. 11. Major element diagrams displaying the 37 intrusive rocks from the Eau Jaune Complex analysed as part of this study (full symbols) and 31 analyses compiled from the SIGÉOM (2018) database. (a) Total alkali versus silica (TAS) diagram (Le Bas et al. 1986); Tonalite and trondhjemite fields after the TTG compilation in the Abitibi subprovince of Mathieu et al. (2020). (b) Alkali-iron-magnesium (AFM) diagram (Irvine and Baragar 1971), with A = Na₂O+K₂O, F = FeOT, and M = MgO, in %. (c) Na-Ca-K (%) triangle also displaying experimental liquids, as compiled by Moyen and Martin (2012).

Fig. 12. QAP diagram (Streckeisen 1976) for normative mineral compositions distributed using the method of Le Maitre (1976). Albite is only in plagioclase, as shown by petrographic observations and in situ microprobe analysis of feldspars. Legend as in Fig. 11. Differentiation series from Lameyre and Bowden (1982) are displayed as grey arrows: T, tholeiitic; A, alkaline. The calc-alkaline series is divided into: a, K-poor; b, intermediate; c, K-rich. Numbers refer to 1, quartzolite; 2, quartz-rich granitoids; 3, alkali feldspar granite; 4, syenogranite; 5, monzogranite; 6, granodiorite; 7, tonalite and trondhjemite; 8, quartz-alkali feldspar syenite; 9, quartz syenite; 10, quartz monzonite; 11, quartz monzodiorite; 12, quartz diorite; 13, alkali feldspar syenite; 14, syenite; 15, monzonite; 16, monzodiorite; 17, diorite.

Fig. 13. Harker diagrams (major and minor elements versus SiO_2) displaying the 37 samples from the Eau Jaune Complex analysed as part of this study and 31 analyses compiled from the SIGÉOM (2018) database. Trends are shown using grey dashed lines. Legend is as in Fig. 11.

Fig. 14. Multi-element diagrams (a, c, and e) for immobile elements, as ordered after Pearce (2008) and rare-earth element (REE) diagrams (b, d, and f) normalized to the primitive mantle (Hoffmann 1988). The diagrams display the 37 intrusive rocks from the Eau Jaune Complex analysed as part of this study and 31 analyses compiled from the SIGÉOM (2018) database. Each line corresponds to a sample and an area corresponds to minimum and maximum values.

Fig. 15. Concordia diagrams showing CA-ID-TIMS results from U–Pb zircon analyses for (a) sample 18UCD0137, a biotite tonalite from Eau Jaune Complex unit 2b, host to the Moly-Desgagné Mo showing, and (b) sample 18UCD0043, a trondhjemite from EJC unit 3a. Error ellipses and calculated age are shown at the 2σ level of uncertainty.

92x145mm (300 x 300 DPI)

Fig. 16. Conceptual diagrams illustrating the different steps leading to the Eau Jaune Complex (EJC) emplacement and its features. (a) Location of the cross-section. (b) Injection of EJC1 magma along a major structure, inducing contact metamorphism and forming a sill at stratigraphic discontinuities. (c) Growth of the sill by inflation to form a laccolith and initiate the overlapping of the Chrissie and Des Vents formations. The laccolith incorporates metamorphosed xenoliths and locally deforms the surrounding volcanic rocks (details in inset). (d) Intrusion of phases EJC2 and EJC3, also forming laccoliths. These intrusions metamorphose and deform the surrounding rocks. The laccoliths incorporate volcanic xenoliths, which were metamorphosed and deformed by the previous injections. Details in inset: xenoliths incorporated in phase EJC2, which were metamorphosed and deformed by growth of the EJC1 laccolith. Vertical scale is exaggerated.