

# The Abitibi-Opatica transition, Superior Province, Quebec, Canada: structural analysis, 40Ar/39Ar thermochronology and implications for Archean tectonics

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## **Precambrian Research**

# The Abitibi-Opatica transition, Superior Province, Quebec, Canada: structural analysis, 40Ar/39Ar thermochronology and implications for Archean tectonics --Manuscript Draft--

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Abstract:	We present a structural and metamorphic study of the Abitibi greenstone belt (AGB) and Opatica Plutonic Belt (OPB) of the Archean Superior Province. The AGB-OPB contact is considered as an archetype example of Archean subduction, based on a LITHOPROBE seismic profile showing a North-dipping lithospheric-scale reflector interpreted as the vestige of subduction. Our mapping indicates that the AGB overlies the OPB, and that the AGB-OPB contact does not show evidence of significant shear deformation, as expected for a major upper plate-lower plate boundary. There is no metamorphic break at the AGB-OPB transition but rather a progressive increase of metamorphic grade toward the OPB. Therefore, we think that the OPB simply exposes the deepest part of a composite AGB-OPB sequence. 40 Ar/ 39 Ar ages suggest that the AGB and OPB rocks were exhumed from amphibolite-facies conditions at ~ 2678 Ma and ~ 2668 Ma, respectively, and then that both terranes share the same succession of thermal disturbances at c. 2650 Ma and 2600 Ma, the latter corresponding to shearing along the Nottaway River Shear Zone. We suggest that progressive cooling of both assemblages was accompanied by strain localisation along strike-slip shear zones and occurred when lateral flow of the lower crust became predominant over the convective overturn and vertical crustal material transfer. Comparison with adjacent areas suggests that regional metamorphism has been coeval over a large region, which is consistent with pervasive deformation and slow cooling as expected for vertical tectonic models and diapiric magmagenesis and ascent during the Archean.
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### Dear Editors,

Please find attached a paper we wish to submit to Precambrian Research.

The paper deals with the contact between two large subprovinces of the Archean Superior Province of Canada : the Abitibi greenstone belt (AGB) and the Opatica plutonic belt (OPB).

Since the middle of the 1990s, the AGB-OPB transition zone was frequently interpreted as the trace of an Archean subduction zone, mainly based on a Grenville-Abitibi Lithoprobe seismic transect acquired in the 90's. In the paper, we emphasize the fact that the ABG-OPB contact is not marked by significant shear deformation and metamorphic break, as expected for a major upper plate-lower plate boundary. In contrast, we infer that the OPB exposes the deepest part of a composite Abitibi-Opatica crust formed prior to the main regional deformation event.

We also present new Ar-Ar ages of amphibole, biotite and muscovite grains from both the AGB and OPB, the results suggesting that the two domains share a common metamorphic history from c. 2678 Ma and the same subsequent thermal disturbances.

Our structural and métamorphic observations combined with Ar-Ar age data support a model of crustal overturned in which vertical transfers of crustal material are predominant and lead to a dome-and-basin crustal geometry. Comparison with adjacent areas also suggests that regional metamorphism, that peaked around 2680-2670 Ma, has been coeval over a large region of the SE Superior province. Such characteristics are in good agreement with pervasive deformation as expected for a diapiric/doming and basins tectonic model rather than one associated with subduction/collision processes.

The paper is original because it discusses the nature of a major contact zone between two contrasted Archean domains in the Superior province, one being famous for its ore deposits (i.e. the AGB). We think that this contribution would be of interest to a large audience, because it is dealing with Archean tectonics that is currently worldwide debated. We hope that it will appear sufficiently new, documented and of large interest enough to be considered for publication in Precambrian Research

Potential reviewers are: B. Lafrance (Laurentian University, Ontario, Canada), L. Harris (INRS, Québec, Canada), Phil C. Thurston (Laurentian University, Ontario, Canada), Laurent Godin (Queen's University, Ontario, Québec) and Michael Brown (University of Maryland, College Park, Maryland, US)

Yours sincerely, Yannick Daoudene (corresponding author)

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2	<sup>40</sup> Ar/ <sup>39</sup> Ar thermochronology and implications for Archean tectonics
3	
4	Yannick Daoudene <sup>a,*</sup> , Alain Tremblay <sup>b</sup> , Gilles Ruffet <sup>c</sup> , François Leclerc <sup>a</sup>
5 6 7 8 9 10 11 12 13	<ul> <li><sup>a</sup> Ministère de l'Énergie et des Ressources naturelles, 5700, 4<sup>e</sup> Avenue Ouest, G1H 6R1, Québec, Québec, Canada.</li> <li><sup>b</sup> Université du Québec à Montréal, Département des Sciences de la Terre et de l'Atmosphère and GEOTOP, 201 avenue du Président Kennedy, H2X 3Y7, Montréal, Québec, Canada.</li> <li><sup>c</sup> CNRS (CNRS/INSU) UMR 6118, Université de Rennes 1, Géosciences Rennes, F-35042 Rennes Cedex, France</li> <li>* Corresponding author.</li> <li>E-mail address: <u>Yannick.Daoudene@mern.gouv.qc.ca</u></li> </ul>
14	ABSTRACT
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16	and Opatica Plutonic Belt (OPB) of the Archean Superior Province. The AGB-OPB contact is
17	considered as an archetype example of Archean subduction, based on a LITHOPROBE
18	seismic profile showing a North-dipping lithospheric-scale reflector interpreted as the vestige
19	of subduction. Our mapping indicates that the AGB overlies the OPB, and that the AGB-OPB
20	contact does not show evidence of significant shear deformation, as expected for a major
21	upper plate-lower plate boundary. There is no metamorphic break at the AGB-OPB transition
22	but rather a progressive increase of metamorphic grade toward the OPB. Therefore, we think
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25	at ~ 2678 Ma and ~ 2668 Ma, respectively, and then that both terranes share the same
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27	shearing along the Nottaway River Shear Zone. We suggest that progressive cooling of both
28	assemblages was accompanied by strain localisation along strike-slip shear zones and
29	occurred when lateral flow of the lower crust became predominant over the convective
30	overturn and vertical crustal material transfer. Comparison with adjacent areas suggests that
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33	magmagenesis and ascent during the Archean.
34	Keywords: Abitibi greenstone belt, Opatica plutonic belt, structures, <sup>40</sup> Ar/ <sup>39</sup> Ar
35	geochronology, Archean tectonism.
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### 37 1. INTRODUCTION

38 Orogenic sutures mark the consumption of oceanic plate lithosphere by subduction 39 and of subsequent arc-continent and/or continent-continent collision, if any. In modern plate 40 tectonics, subduction zones are linear and continuous areas of planetary extent where one 41 plate moves under another and is forced to sink due to high gravitational potential energy into 42 the mantle (e.g. Stern, 2002; Bédard, 2018; Brown et al., 2020). Typically, such areas are 43 characterized by an upper plate-lower plate geometry with contrasting geological 44 characteristics. In Phanerozoic belts, subduction sutures are a typical product of plate 45 tectonics. Although their recognition can be sometimes dubious (e.g. Dewey, 1987), they are 46 usually marked by linear belts of intense deformation, where distinct terranes with different 47 magmatic, metamorphic, structural and geochronological histories have been juxtaposed by 48 plate convergence. In the Archean, the nature of global tectonism remains debated, the main 49 problem concerning similarities and differences with present-day plate tectonics (e.g. Stern, 50 2008; Brown et al., 2020).

51 Archean cratons largely consist of greenstone belts juxtaposed to deeper crustal rocks 52 dominated by plutonic (TTG's) and orthogneissic rocks, the greenstone belts usually show 53 steep folds and shear zones whereas plutonic belts display large-scale dome-like structures 54 and fabrics. These rocks usually preserve MT/HT-LP/MP metamorphism that varies from 55 greenschist- to granulite-facies, from the upper to the lower crustal domains. The most widely 56 used arguments in favor of Archean subduction are the occurrence of calk-alkaline igneous 57 rocks, the structural style of orogenic belts, the nature and asymmetric distribution of 58 metamorphism, and seismic reflection data. In seismic reflection, there are few clear examples 59 of relict slabs beneath Archean collisional belts. An example of it is found in the Lithoprobe 60 seismic transect of the Superior Province (Calvert et al., 1995) at the contact between the Abitibi Greenstone Belt and the Opatica Plutonic Belt. In scientific literature, it has been and 61 62 is still interpreted as a fossilized subduction zone and a key argument to support Archean 63 plate tectonics (Calvert et al., 1995; Calvert and Ludden, 1999; Ludden and Hynes, 2000; 64 Wyman et al., 2002; van der Velden and Cook, 2005; Cawood et al, 2006; Percival, 2007; van 65 Hunen and Moyen, 2012; among others).

The Archean Superior Province (Fig. 1) consists of a series of volcano-sedimentary
and plutonic domains, subprovinces and/or terranes that amalgamated during the Late
Archean Kenoran orogeny (e.g. Percival, 2007). The occurrence of unconformable
sedimentary rocks as old as 2.48 Ga (i.e. the Huronian Supergroup) over the volcanic and

70 plutonic rocks indicates that tectonic accretion occurred prior to 2.5 Ga. Although the internal 71 geological characteristics of the domains/subprovinces/terranes are relatively well known, the 72 nature of their boundaries remains poorly constrained and, sometimes, debated. Current views 73 on the tectonic evolution of the Superior Province are that the orogenic collage of 74 Mesoarchean continental fragments and Neoarchean oceanic plates occurred during a 75 complex history of subduction-driven accretionary collision (e.g. Bleeker, 2003; Percival, 76 2007; Percival et al., 2006; 2012, among others). Hence, several authors have interpreted a 77 series of high-strain deformation corridors and structural lineaments of the Superior Province 78 as the trace of potential orogenic sutures (Ludden et al., 1986; Mueller et al., 1996; 79 Daigneault et al., 2002, 2004). The Abitibi-Opatica contact, for instance, is considered as the 80 surface expression of south-verging crustal-scale thrust faults rooting into an inferred North-81 dipping subduction zone, the existence of that subduction being attested by a lithospheric-82 scale seismic reflector extending from c. 40 to c. 70 km-depth.

83 In this contribution, we present a structural and metamorphic study of rocks lying on 84 both sides of the Abitibi-Opatica contact in northern Quebec for a strike length of c. 300 km, 85 between the Lac-au-Goéland area to the west and the town of Chibougamau to the east (Fig. 86 2). The main objectives of fieldwork have been to test (1) how the overall geometry of 87 regional structures fits with south-verging collisional tectonism and thrusting of the OPB over 88 the AGB, and (2) if the structural and metamorphic characteristics of both belts are consistent 89 with an upper-lower plate geometry, as inferred by the subduction model. The determination 90 of the age and duration of structural and metamorphic event(s) that have affected both 91 terranes being critical for a better understanding of their respective geological evolution and 92 mutual relationships, we have performed a series of single-grain  $^{40}$ Ar / $^{39}$ Ar analyses for 93 representative samples of both belts.

94

### 95 2. GEOLOGICAL SETTING

96 The Superior Province forms the principal Archean basement of North America (Fig. 97 1a, upper left inset). It is made up of several tectonostratigraphic terranes or subprovinces 98 grouped into four major tectonostratigraphic types/domains (Card, 1990; Card and Ciesielski, 99 1986; Hocq, 1994; Stott et al., 2010), (1) metasedimentary rocks, (2) volcanic and plutonic 100 rocks, (3) tonalite-trondjhemite-granite (TTG)-dominated domains, and (4) high-grade 101 metamorphic rocks. Most tectonic models suggest that these tectonostratigraphic terranes 102 were amalgamated, regionally deformed and metamorphosed during multiple episodes of

- 103 subduction/collision (e.g. Kimura et al., 1993; Daigneault et al., 2002; Percival et al., 2006,
- 104 2012) between ca. 2.72 Ga and 2.68 Ga (Percival, 2007). The southeastern part of the
- 105 Superior Province comprises, from north to south, the Opatica, Abitibi and Pontiac
- subprovinces, the former two being designated here as the Opatica Plutonic Belt (OPB) and
- 107 the Abitibi Greenstone Belt (AGB) (Fig. 1b).
- 108

### 109 2.1 The Opatica Plutonic Belt

110 The OPB is located in the northern part of the southeastern Superior Province. It is 111 separated in two structural domains of plutonic rocks, north and south of the Frotet-Evans 112 greenstone belt that extends east-west over a strike length of approximately 250 km (Figs. 1b 113 and 2). The detailed geology of the OPB is poorly documented, mainly due to difficult road 114 access and a limited mineral resources interest as compared to the adjacent Abitibi 115 subprovince. Existing regional studies indicate that the OPB is a large domain of igneous 116 rocks essentially consisting of foliated tonalites and diorites and equivalent gneisses. The 117 latter were designated as «grey gneisses» by Sawyer and Benn (1993). They are characterized 118 by well-developed compositional layering and locally abundant occurrences of diatexite 119 migmatites in their more fractionated members, mainly in the southern part of the belt (Benn 120 et al., 1992; Sawyer and Benn, 1993; Hocq, 1994; Sawyer, 1998). The gneissic tonalites and 121 diorites are intruded by abundant, slightly- to undeformed plutons of granodiorite and granite. 122 The crystallization age of the OPB intrusive rocks is not precisely constrained but the 123 available geochronological data suggest that they formed mostly between 2.83 and 2.68 Ga 124 (Gariépy and Allègre, 1985; Davis et al., 1992; 1994, 1995; Pedreira-Perez et al., 2020). 125 According to Sawyer (1998), partial melting and migmatization occurred at 2.68-2.67 Ga.

126 In the Lac-au-Goéland area (Fig. 3), the Nomans tonalite, which is surrounded by 127 metavolcanic and metasedimentary rocks of the AGB, possesses petrological and structural 128 characteristics that are identical to those observed in tonalitic rocks of the OPB (Daoudene et 129 al., 20134). It represents an antiformal window of the OPB within the AGB (Goutier, 2005; 130 Daoudene et al., 20134; see below). Previous mapping (i.e. Hocq, 1994), as well as our own 131 observations, shows that the OPB contains km-scale, scattered «lenses» of greenstone-type 132 rocks (Figs. 2 and 3) consisting of amphibolite-grade mafic volcanic rocks and mafic-133 ultramafic intrusive rocks (Figs. 4a and b). The texture of these mafic volcanic rocks varies

134 from primary, with locally preserved breccia and pillow structures (Fig. 4a), to migmatitic.

The Frotet-Evans greenstone belt (Fig. 2) is a large-scale example of metavolcanic 135 136 belts surrounded by plutonic rocks of the OPB. It is divided into four segments, (cf. Boily, 137 2000) from west to east, the Evans-Ouagama, Storm-Evans, Assinica and Frotet-Troilus 138 segments (cf. Boily, 2000). Figure 2 shows the two easternmost segments, the Frotet-Troilus 139 and Assinica segments. The Frotet-Evans greenstone belt has been interpreted as the erosional 140 remnant of an allochthonous thrust sheet by Sawyer and Benn (1993) but it rather represents 141 an autochthonous synformal sequence within the OPB (see below). It preserves evidence for 142 amphibolite-facies metamorphic conditions, except for its central part where greenschist-143 facies metasedimentary rocks are predominant (Simard, 1987; Brisson et al., 1997a and b). 144 Volcanic and sedimentary rocks of the Frotet-Troilus segment are included into the Frotet 145 Group; U-Pb zircon ages from felsic tuff layers and a felsic dyke yielded crystallization ages 146 of  $2780 \pm 28$  Ma and  $2750 \pm 28$  Ma (Thibault, 1985), and  $2782 \pm 6$  Ma (Pilote et al., 1997), 147 respectively. The Assinica segment represents the western extension of the Frotet-Troilus 148 segment (Fig. 2). It comprises a lower unit of mafic to felsic volcanic rocks, the Assinica 149 Group, and an upper unit of sub-concordant, clastic sedimentary rocks, the Broadback Group. 150 There are currently no isotopic data to constrain the age of the Assinica Group but correlation 151 with the Rabbit Formation of the Evans Group in the Storm-Evans segment to the west, 152 suggests that it is probably ca.  $2758 \pm 4$  Ma (U-Pb on zircon; David, 2018 in Groulier et al., 153 2020), which indicates that the overlying clastic sequence of the Broadback Group is ca. 2755 154 Ma or younger. segment.

155

### 156 2.2 The Abitibi Greenstone Belt

157 The AGB is made up of ultramafic, mafic, felsic volcanic rocks and syn-volcanic 158 intrusions of Late Archean age. It consists of several, east-west-trending assemblages of 159 volcanic, volcanoclastic and sedimentary rocks (Fig. 1b) that have been deformed and 160 metamorphosed at greenschist- to amphibolite-facies conditions (e.g. Daigneault et al., 2002). 161 The volcanic domains are separated by narrow and linear belts of clastic sedimentary rocks 162 that are frequently associated with major fault zones (Ludden et al., 1986; Percival, 2007). 163 The youngest sequence of these clastic sedimentary rocks, the Temiskaming Group, was 164 deposited above angular unconformities between ca. 2679 and ca. 2670 Ma (e.g. Robert et al., 165 2005). Based on the age of the various volcanic sequences, the AGB is subdivided into the 166 Northern and Southern Volcanic zones originally separated by the Destor-Porcupine fault 167 zone (Ludden et al., 1986; Chown et al., 1992; Daigneault et al., 2004) but relocated

168 approximately 50 km northward by Thurston et al. (2008; see their Figure 2). In the Southern 169 Volcanic zone, corresponding to the Tisdale (2710-2704 Ma) and Blake River (2704-2695 170 Ma) assemblages of Thurston et al. (2008; Fig. 1b), peak of regional metamorphism has been 171 bracketed at 2680-2660 Ma by Powell et al. (1995), and at 2665-2660 Ma by Robert et al. 172 (2005) and Tremblay et al. (2020). The AGB consists of several, east west trending 173 assemblages of volcanic, volcanoclastic and sedimentary rocks (Fig. 1) that have been 174 deformed and metamorphosed at greenschist- to amphibolite-facies conditions (e.g. 175 Daigneault et al., 2002). In the southern part of the belt, peak of regional metamorphism has 176 been bracketed at 2680-2660 Ma by Powell et al. (1995), and at 2665-2660 Ma by Robert et 177 al. (2005) and Tremblay et al. (2020). The volcanic and sedimentary rocks, which together 178 represent about 50-60% of the AGB, bound large plutonic domains dominated by TTG-type 179 plutons (Fig. 1b), and both the volcano-sedimentary assemblages and plutonic domains are 180 intruded by syn- and post-tectonic granite, granodiorite and syenite plutons that span ages as 181 young as ca. 2650 Ma syn-volcanic to syn- and post-tectonic intrusions (Feng et al., 1992; 182 Daigneault et al., 2002; Robert et al., 2005). In the Val d'Or area (Fig. 1b), the Preissac-183 Lacorne batholith is a typical example of these syn- to late-orogenic intrusions. It is made up 184 of a composite series of intrusions in which two distinct types and ages of granitic rocks are 185 present: an early, subalkaline to alkaline, monzodiorite-monzonite-granodiorite-syenite series 186 emplaced between 2690 and 2670 Ma; and a younger, S-type garnet-muscovite granite series 187 dated between 2660 and 2642 Ma, and probably derived from a major metasedimentary 188 source (Feng et al., 1992; Ducharme et al., 1997; Daigneault et al, 2002). Based on the age of 189 the various volcanic sequences, the AGB is subdivided into the Northern and Southern 190 Volcanic zones separated by the Destor Porcupine fault zone (Ludden et al., 1986; Chown et 191 al., 1992; Daigneault et al., 2004), the Southern Volcanic zone corresponding to the Tisdale 192 (2710-2704 Ma) and Blake River (2704-2695 Ma) assemblages of Thurston et al. (2008: Fig. 193 <del>1)</del>. 194

In the study area (Fig. 32), the volcanic rocks of the AGB belong to the Northern Volcanic zone. It mainly consists of two volcanic cycles formed between ca. 2.8 and 2.7 Ga (e.g. Leclerc et al., 2011), and belongs to 2711-2750-2711 Ma volcanic rocks of the Kidd-Monro, Stoughton-Roquemaure, Deloro and Pacaud assemblages with local >2750 Ma occurrences (Fig. 21b; Thurston et al., 2008). In the Chibougamau-area (Fig. 2), the base of the volcanic sequence belongs to the Chrissie Formation, which is currently one of the oldest volcanic units of the AGB at 2791.4 +3.7/-2.8 Ma (U-Pb zircon age; David et al., 2011). It

201 consists of mafic to felsic volcaniclastic rocks, calk-alkaline rhyolites, and tholeiitic-to-202 andesitic basalts. The Roy Group overlies the Chrissie Formation. It includes two volcanic 203 and sedimentary cycles and is estimated to be up to 13 km-thick (Allard et al., 1985; Dimroth 204 et al., 1984; Daigneault et al., 1990). The lower cycle consists of tholeiitic basalts, the 205 Obatogamau Formation, overlain by tholeiitic rhyolites and calk-alkaline volcaniclastic.rocks 206 of the ca. 2732 Ma to ca. 2726 Ma Waconichi Formation (Daigneault and Allard, 1990; 207 Legault, 2003; Leclerc et al., 2011; David et al., 2012). The upper cycle of the Roy Group 208 comprises, from base to top, the Bruneau, Blondeau, Scorpion and Bordeleau formations, 209 which consist of basalt and andesitic basalt at the base overlain by a well-bedded sequence of 210 volcaniclastic and sedimentary rocks (Leclerc et al., 2011). In the Chibougamau area, a lapilli 211 tuff of the Scorpion Formation yielded U-Pb crystallization age on zircon of 2716,4 ±1,0 Ma 212 (David et al., 2012). In the Lac-au-Goéland area (Fig. 3), rocks correlatives to the Bordeleau 213 Formation belong to the Bell River and Dussieux formations and yielded zircon U-Pb ages of 214 **a** 2725-2717 Ma (Goutier et al., 2004; David, 2005; David et al., 2006).

215 The Opémisca Group (Fig. 2) is a sequence of clastic sedimentary and volcanic rocks 216 overlying the Roy Group (Charbonneau et al., 1991; Leclerc et al., 2008; Leclerc and Houle, 217 2013). From base to top, it consists of the La Trêve, Daubrée, Stella and Haüy formations 218 (Charbonneau et al., 1991). In the Chibougamau area, the Stella Formation has been 219 interpreted as a lateral equivalent to the La Trêve and Daubrée formations (Charbonneau et 220 al., 1991). The La Trêve Formation consists mainly of polygenic conglomerates with 221 heterogeneous clasts, including TTG granitoids and mafic-to-felsic volcanics, whereas the 222 Daubrée Formation is basically made up of turbidites (Fig. 4c). The Haüy Formation 223 corresponds to a volcanic sequence mainly composed of K-rich porphyric andesites with 224 lenses of conglomerate. The Haüy lava flows and associated volcaniclastic sediments have 225 been interpreted as the result of evolved arc volcanism and backarc basin sedimentation 226 (Mueller, 1991). In the Stella and Haüy formations, detrital zircons provided maximum ages 227 of deposition of  $2704 \pm 2$  Ma (Leclerc et al., 2012) and  $2691.7 \pm 2.9$  Ma (David et al., 2007), 228 respectively. In the Lac-au-Goéland area (Fig. 3), the Daubrée Formation is intruded by the 229 Inconnue River Diorite, which yielded a U-Pb zircon age of  $2693.6 \pm 2.9$  Ma (McNicoll and 230 Goutier, 2008).

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### 232 **3. REGIONAL STRUCTURES**

7

233 According to Sawyer and Benn (1993), the OPB records three generations of 234 structures. The oldest is related to  $D_1$  and a penetrative, moderately dipping foliation ( $S_1$ ) 235 hosting broadly distributed, WSW-ENE-trending mineral and stretching lineations (Fig. 4b). 236  $S_1$  corresponds to the regional gneissic foliation (Fig. 4c) and defines a metamorphic layering 237 (Fig. 4c) that is axial-planar to intrafolial  $F_1$  folds.  $D_1$  has been attributed to a top-to-the SW 238 verging deformation. Sawyer and Benn (1993) argued that  $S_1$  was folded during a  $D_2$ 239 deformational event, marked by East-West trending, steeply North- or South-dipping 240 compressive shear zones. S<sub>2</sub> is a steeply dipping biotite foliation recording amphibolite-241 greenschist facies metamorphic conditions, although it may locally attain the amphibolite 242 facies. It frequently forms a composite  $S_{1-2}$  fabric with  $D_1$  structures.  $D_1$  and  $D_2$  has been 243 attributed to the progressive development of an overall SSE-verging fold-and-thrust belt (and 244 inferred backthrusting) along the AGB-OPB boundary and southward (Benn et al., 1992; 245 Sawyer and Benn, 1993; Lacroix and Sawyer, 1995). Sawyer and Benn (1993) also 246 recognized a D<sub>3</sub> event, related to NE-SW and NW-SE-trending strike-slip shear zones cutting 247 across both the Frotet-Evans and the Abitibi greenstone belts (see Fig. 2).

248 In the AGB of the study area, existing structural studies of regional extent have 249 documented four deformational phases (e.g. Dimroth et al., 1986, Daigneault et al., 1990; 250 Leclerc et al., 2012; Mathieu et al., 2020), the youngest one (D<sub>4</sub>) being related to the 251 Mesoproterozoic Grenvillian orogenic cycle (Daigneault et al., 1990). D<sub>1</sub> has been attributed 252 to an early-formed series of drape folds devoid of axial-planar schistosity (Daigneault et al., 253 1990). D<sub>2</sub> is related to syn-metamorphic folding and the development of the regional EW-254 trending schistosity ( $S_2$ ), it corresponds to  $F_{2a}$  and  $F_{2b}$  folds as described by Leclerc et al. 255 (2012).  $D_1$  and  $D_2$  of the AGB correlate with the Benn's et al. (1993)  $D_{1-2}$  phase of the OPB. According to Mathieu et al. (2020), the D<sub>3</sub> phase corresponds to the waning stage of the main 256 257 deformation event that reactivated E- W- striking D<sub>2</sub> reverse faults as transcurrent 258 strike- slip faults. In contrast, Leclerc et al. (2012) described D3 structures as NE-striking 259 shear zones cutting across earlier E-W trending D<sub>2</sub> shears/faults (Fig. 2). 260 Our comprehension and interpretation of structures in the study area are similar to

those described above. We think that  $D_{1-2}$  and  $D_3$  belong to progressive and long-lasting deformational events. In the following sections, we describe fabrics and structures related to those  $D_{1-2}$  to  $D_3$  events in both the AGB and the OPB. These descriptions are based on our own field observations and from previous studies.

265

### 266 **3.1 D<sub>1-2</sub> phase of deformation**

267 D<sub>2</sub> deformation structures are regionally-developed and associated with folding of the 268 volcanic and sedimentary rocks of the AGB (Figs. 2 and 3) whereas the D<sub>1</sub> metamorphic 269 layering is restricted to the lowermost stratigraphic part of that belt. Regional folds  $(F_2)$  have 270 variable wavelength and their axial traces strike East-West, although they are locally 271 reoriented due to younger D<sub>3</sub> strike-slip shears, such as the Nottaway shear zone (Fig. 2). In 272 the Lac-au-Goéland area (Fig. 3), for instance, to the southwest of the Nottaway shear zone, 273 S<sub>2</sub> trajectories within the AGB are NW-SE trending and define a F<sub>3</sub> syncline, the Maicasagi 274 syncline (Fig. 3; Goutier, 2005). Overall, the regional pattern of folding of the AGB is 275 characterized by a north-south succession of  $F_2$  synclines and anticlines. The core of synclines 276 is usually occupied by the Opémisca Group, whereas the anticline crests expose the 277 underlying volcanic rocks of the Roy Group which are, locally, crosscut by large pre- to syn-278 tectonic intrusions such as the Opémisca and Chibougamau plutons (Fig. 2).

279 The volcanic and sedimentary rocks of the AGB are characterized by a weakly to 280 moderately developed regional schistosity  $(S_2)$ , although some rock exposures seem to beare 281 devoid of structural fabrics development (Fig. 4d). When present,  $S_2$  is pervasive, strikes 282 East-West and dips steeply either to the north or the south (Figs 5 and 6a). Towards the 283 contact with the OPB, the volcanic rocks of the Roy Group are more intensely foliated, the 284 dominant foliation  $(S_1)$  is there tightly folded by  $F_2$  (Fig. 4e), although that, most of the time, 285 it forms a composite  $S_{1-2}$  fabric and consists of a metamorphic layering marked by 286 metamorphic minerals such as biotite and hornblende (Fig. 4b, see section 4), that 287 corresponds northward to the flat-lying S<sub>1</sub> gneissic fabric of the OPB gneisses. Our 288 interpretation is that the OPB gneissic foliation is verticalized and transposed into S<sub>2</sub>, it forms 289 there then a  $S_{1-2}$  metamorphic layering that is subparallel to the contact between the OPB and 290 the AGB (Fig. 65) and is absent farther south in the up-stratigraphy, lower-grade rocks of the 291 AGB. The Bbedding plane, when visible, is usually parallel to the regional S<sub>2</sub> that is marked 292 by the alignment of amphibole and biotite in amphibolite-grade rocks (as for  $S_1$ ) and by 293 chlorite, epidote and albite in greenschist and sub-greenschist facies rocks of the AGB. 294 East-West-trending shear zones cut the volcanic and sedimentary sequences of the

AGB. These sub-vertical shear zones are poorly exposed. Previous studies (e.g. Dimroth et al., 1986; Daigneault et al., 1990; Leclerc et al., 2012) indicate they are mainly located on the limbs of regional-scale  $F_2$  folds (Fig. 2), some of the latter being synformal troughs occupied by lower grade sedimentary rocks of the Opémisca Group (Fig. 4ed). As for  $F_2$  folds, these shear zones are deflected in the vicinity of younger NE- and NW-trending strike-slip shear zones (Figs. 2 and 3). S<sub>2</sub> increases in intensity toward and within these D<sub>2</sub> shear zones and hosts steeply plunging stretching and/or mineral lineation, suggesting significant subvertical elongation (Dimroth et al., 1986; Daigneault et al., 1990). Shear-sense indicators are rare but when present, they suggest that the shear zones correspond to north or south side-up reverse faults.

305 In the OPB, dominant deformation fabrics are attributed to  $D_1$  and expressed by a 306 weakly- to well-developed metamorphic foliation  $(S_1)$  showing variable orientations (Fig. 5). 307 S<sub>1</sub> trajectories in the OPB define an East-West trending, elongated dome-and-basin geometry, 308 consistent with the overall trend of  $D_2$  regional folds and structures in the adjacent AGB (Fig. 309 5). Mineral lineations, defined by the alignment of quartz-feldspar aggregates, are more 310 discrete and of variable orientation as compared to the AGB (Fig. 6b). These lineations 311 plunge at low angle within the regional foliation. In the area located between the Frotet-Evans 312 and Abitibi greenstone belts (Fig. 2), there are abundant decametre-scale of foliated, 313 amphibolite-grade lenses of basaltic metavolcanic rocks within the OPB gneisses. The 314 foliation in these metavolcanic rocks trends parallel to the S<sub>1</sub> metamorphic layering in the 315 surrounding gneisses.

316 The AGB-OPB contact is not clearly exposed in the area of Figure 2. Rock exposures 317 are sparse along that contact which is locally masked by syn- and post-orogenic intrusions, 318 such as the Barlow pluton in the Chibougamau area (Fig. 2) and the Canet and Maicasagi 319 plutons (Sawyer and Benn, 1993; Goutier, 2005; Daoudene et al., 20134) in the Lac-aux-320 Goéland area (Fig. 3). In the latter area, the Nomans tonalite, located to the south of the 321 inferred AGB-OPB contact, shows the same structural style as the gneissic rocks of the OPB. 322  $S_1$  trajectories within the tonalite (Fig. 5) indicate that it is a  $D_2$ -related structural dome that is 323 attached to the OPB at depth (Goutier, 2005; Daoudene et al., 20134). The occurrence of 324 foliation-parallel, amphibolite lenses along the AGB-OPB contact has been considered as 325 evidence for the structural imbrication of both terranes and the allochthonous nature of the AGB (Sawyer and Benn, 1993). We do not think that this boundary corresponds to a major 326 327 displacement zone (see section 4 below), although S<sub>1</sub> appears there to be more intensely 328 developed in both the metavolcanic rocks of the AGB and the OPB tonalites. S<sub>1</sub> trajectories 329 are parallel to the AGB-OPB contact (Fig. 5), suggesting that it formed, or was transposed, 330 during D<sub>1</sub>. There are relatively few lineation measurements in the study area (Fig. 6). Sawyer 331 and Benn (1993) reported SE-plunging mineral lineations in the hanging wall of the Lac

332 Matagami shear zone, a late (D<sub>3</sub>) sinistral fault located west of the Lac-au-Goéland area, as

333 well as down-dip lineations along the Barlow fault in the Chibougamau area. In this latter

- area, stretching lineations plunge moderately to the south, but locally show significant
- variations of orientation due to the proximity of syn- and post-tectonic intrusions (Bedeaux etal. 2021).

The Abitibi and Frotet-Evans greenstone belts share a similar internal geometry and
relationships with the underlying plutonic rocks of the OPB. However, the Frotet-Evans
greenstone belt is of higher metamorphic grade and structurally more complex than the AGB
(see Gosselin, 1996). It occupies the core of a large-scale series of east-west-trending F<sub>2</sub>
synclines (Figs. 2 and 5) within the OPB gneisses, some of them being marked, as for the
AGB, by steeply dipping reverse faults on their limbs (Simard, 1987; Gosselin, 1996; Brisson
et al., 1997a and 1997b).

In the study area, we did not observe unambiguous shear-sense criteria within the
AGB-OPB contact zone. Along the AGB-OPB contact, there are numerous tonalitic to
granitic dykes crosscutting basaltic amphibolites of the AGB (Fig. 4f). Most of these dykes
are in irregular and diffuse contacts with the host rocks and are slightly to moderately foliated,
suggesting high-temperature high temperature syntectonic emplacement.

349

### **350 3.2 D**<sub>3</sub> **structures**

351 The  $D_3$  phase of regional deformation is more localized as compared to  $D_{1-2}$ . It is 352 essentially associated with strike-slip displacement along NE- and NW-trending shear zones. 353 The Nottaway River shear zone (Figs 2 and 3) is the largest of the NW-SE-trending  $D_3$  shear 354 zones in the study area. It cuts across the  $D_{1-2}$  structures of both the AGB and OPB. It delimits 355 the western boundary of the Frotet-Evans greenstone belt, which is also cut by another major 356 NW-trending shear in its middle part, the Lucky Strike River shear zone (Fig. 2). A mylonitic 357 S<sub>3</sub> schistosity is pervasively developed along these shear zones (Fig. 7a). F<sub>3</sub> dextral folds, with 358 S<sub>3</sub> as an axial-planar foliation, can be observed in the vicinity of D<sub>3</sub> shear zones (Fig. 7b). S<sub>3</sub> 359 hosts a well-developed, sub-horizontal stretching lineation (Fig. 7c), indicating that  $D_3$  is 360 dominated by strike-slip motions. Abundant macroscopic and microscopic shear-sense 361 indicators, such as shear bands, S-C fabrics (Fig. 7a and 7d), mica fish (Fig. 7e), drag folds 362 and asymmetrical boudinage (Fig. 7f) are widely present and attest for dextral displacement along the NW-SE-trending shear zones. NE-trending shears, such as the Lac Lamark shear 363 364 zone (Fig. 2), are characterized by the same type of ductile fabrics but are sinistral, as

- indicated by regional mapping and the offset of marker horizons in the Chibougamau area
- 366 (e.g. Chown et al., 1992; Daigneault et al., 2002). The two sets of strike-slip shear zones,
- 367 NW- and NE-trending, form a conjugate system.
- 368

### 369 **4. METAMORPHISM**

In this section, we describe the metamorphic characteristics of rocks in the study area,
from top to base of the AGB stratigraphy, and from the lower-grade rocks of the AGB
towards the high-grade metamorphic units of the OPB.

373 The sedimentary rocks of the Opémisca Group, which occupy the core of regional F<sub>2</sub> 374 synclines, do not show macroscopic evidence of metamorphic recrystallization. Primary 375 sedimentary structures, such as bedding (Fig. 4ed), and pillow structures in volcanic rocks of 376 the underlying Roy Group are frequently well-preserved (Fig. 8a). Low-grade metamorphic 377 minerals, such as chlorite, sericite, epidote, and actinolite-tremolite progressively appear 378 down into the stratigraphic pile and can be relatively abundant at the margins of synclines, 379 indicating that the lowermost part of the Opémisca Group attained greenschist-facies. The 380 metamorphic grade of these sedimentary rocks even reaches the amphibolite facies towards 381 the AGB-OPB contact, specifically along the northern boundary of the Waconichi Lake 382 syncline (Figs 2 and 3) where the Opémisca metapelites contain biotite, garnet, and staurolite 383 (Fig. 8b). Staurolite occurs as mcm-long, oblong-shaped and corroded crystals defining the S<sub>2</sub> 384 foliation. Lithologies of the Opémisca Group also locally contain late- to post-kinematic 385 metamorphic minerals such as, for instance, randomly oriented chloritoïds (Fig. 8c) 386 suggesting late-stage growth.

387 The mafic rocks of the Roy Group were dominantly metamorphosed to the 388 greenschist-facies (Allard et al., 1985). These rocks are pervasively chloritized and contain 389 typical low-grade metamorphic minerals such as albite, epidote and actinolite-tremolite 390 defining the S<sub>2</sub> schistosity. In the vicinity to the contact with the OPB, they contain, however, 391 metamorphic minerals of higher temperature characteristic of the amphibolite facies. 392 Actinolite-tremolite is there replaced by hornblende and biotite. Garnet is present and usually 393 forms irregular patches that are preferentially distributed along the pillow margins of mafic 394 volcanic flows (Fig. 8d). Garnet is frequently retrogressed into a mixture of epidote, 395 amphibole and chlorite. Some basaltic amphibolites contain epidote porphyroblasts overprinting the composite S<sub>1-2</sub> foliation as well as randomly oriented crystals of actinolite, 396 397 suggesting retrogressive conditions after peak metamorphism.

398 Metamorphic conditions recorded by the OPB are more difficult to evaluate precisely, 399 mainly because these rocks essentially consist of intermediate and felsic plutonic rocks. 400 However, most of the mafic rocks of the Opatica contain hornblende+andesine assemblages, 401 which, with the occurrence of rare garnet- or clinopyroxene, imply amphibolite-facies 402 conditions (Sawyer, 1998). The intermediate and felsic gneissic rocks are commonly 403 plastically deformed and show coarse-grain recrystallization textures. Feldspar and quartz 404 crystals display evidence for syn-kinematic growth and recrystallization. Grain boundaries are 405 cuspate and undulated (Fig. 8e), which is indicative of extensive high-temperature grain 406 boundary migration during deformation (Gower and Simpson, 1992). Myrmekitic textures 407 preserved in some coarse-grained K-feldspars also indicate metamorphic temperatures in 408 excess of 500-550°C (Simpson, 1985; Simpson and Wintsch, 1989). All these observations 409 indicate that the regional D<sub>1</sub> deformation structures in the OPB gneisses were acquired at 410 amphibolite-facies conditions of metamorphism, or higher. Some felsic rocks display quartz 411 grains with a chessboard texture, which is typical of intra-crystalline slip textures developed 412 at subsolidus temperatures (i.e. above 600°C; Blumenfeld et al., 1986; Gapais and Barbarin, 413 1986; Mainprice et al., 1986; Schmid and Casey, 1986). Evidence for anatectic melting is 414 abundant in the deepest crustal domains of the OPB and marked by the occurrence of 415 migmatites (Fig. 8f). Anatectic melt occurs in D<sub>2</sub> shears zones located near to the center of the 416 OPB (Sawyer and Benn, 1993). However, according to Sawyer (1998), melting is late- to 417 post- $D_1$ , as suggested by  $S_1$  disappearing within neosome patches and paleosome material 418 showing remnants of the  $S_1$  foliation.

419 There are very few pressure-temperature (P-T) studies for metamorphic rocks of the 420 AGB. At the scale of the belt, regional metamorphism ranges in grade from subgreenschist-421 facies to the greenschist-amphibolite facies transition. In the Blake River and Tisdale 422 assemblages (Fig. 1b), low-pressure regional metamorphism is indicated both by the 423 occurrence of the actinolite-oligoclase zone, and the persistence of pre-regional 424 metamorphism and alusite. According to Powell et al. (1995), pressure of regional 425 metamorphism and of late to post-orogenic contact metamorphism were similar (360-400 426 MPa), indicating that low-pressure conditions persisted throughout the tectonic evolution of 427 the AGB. Regionally, isograds dip shallowly to the north and trend subparallel to lithological 428 and structural trends (Powell et al., 1995; Faure, 2015). In the Chibougamau area (Fig. 2), Bedeaux et al. (2020) suggested P-T conditions of 550±50°C and 6±1.2 kbar for basaltic 429 430 amphibolites of the AGB cropping out in the vicinity of the contact with the OPB. In the

- 431 OPB, Sawyer (1998) determined a peak metamorphism temperature of 760°C for the
- 432 migmatite facies rocks, and pressures of 5.2-6.2 and 6.3-7.1 kbar for diatexite migmatites and
- 433 adjacent amphibolites, respectively. Pressure measured in the Canet pluton of the Lac-au-
- 434 Goéland area (Fig. 3) were significantly lower, at 1.7-2.6 kbar (Sawyer, 1998), suggesting
- that the emplacement of this pluton occurred during exhumation of the OPB gneisses.
- 436 Figure 9 presents a series of north-south trending, interpretative structural profiles 437 between the Lac-au-Goéland (profile A-A') and the Chibougamau areas (profile D-D'). These 438 profiles show that the OPB exposes deeper crustal rocks and higher-grade metamorphic rocks 439 as compared to the adjacent and overlying volcanic rocks of both the Frotet-Evans and Abitibi 440 greenstone belts, and that both the OPB and AGB originally constituted an almost continuous 441 middle-lower (the OPB) to upper crust sequence (the AGB). This contrasts with Sawyer and 442 Benn's (1993) interpretation showing the Frotet-Evans greenstone belt as a far-travelled, 443 south-directed allochthonous nappe thrusted over the OPB gneisses, and both units forming a 444 basement-cored nappe thrusted southward over the AGB (see their Ffigures 6 and 8). We 445 think that their structural interpretation is unlikely for the following reasons, (i) in typical 446 crustal-scale thrust systems (Boyer and Elliott, 2002; Butler and Bond, 2020), older rock units 447 and/or higher grade (deeper) metamorphic rocks are transported over younger and/or lower 448 grade (shallower) metamorphic rocks, which is obviously not the case here, and (ii) it would 449 imply that the Frotet-Evans greenstone belt represents a klippen that roots 50 km, or more to 450 the north, in the Colomb-Chabouillé Group of the Nemiscau subprovince along its southern 451 the boundary with the OPB (see Fig. 2); there are, however, no metamorphic break and major 452 suture between the Nemiscau and Opatica subprovinces, and regional mapping and structural 453 analyses rather suggest the relative exhumation of the OPB as compared to the Nemiscau 454 (Pedreira-Perez et al., 2020), which does not fit the Sawyer and Benn's (1993) thrusting 455 model.
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- 456

### 457 **5.** <sup>40</sup>Ar/<sup>39</sup>Ar GEOCHRONOLOGY

In the following section, we present thirty-one <sup>40</sup>Ar/<sup>39</sup>Ar laser probe experiments on single grains of amphibole, biotite and muscovite from eighteen rock samples from both the AGB and OPB.

461

462 **5.1 Analytical method and procedures** 

Single grains of amphibole, biotite and muscovite were handpicked under a binocular microscope from 250 $\mu$ m to 1000 $\mu$ m mineral fractions. Separated minerals were irradiated with Cd-shielding, in the 5C high neutron flux facility of the McMaster Nuclear Reactor (Hamilton, Ontario, Canada). This process lasted 298 h with a global efficiency (J/h) of 5.83x10<sup>-5</sup> h<sup>-1</sup>. The irradiation standard was amphibole Hb3gr (Turner et al. 1971; Roddick 1983, Jourdan et al. 2006; Jourdan and Renne 2007); 1081.0 ± 1.2 Ma according to Renne et al. (2010, 2011).

Single grains of amphibole, biotite and/or white mica were analyzed by the  ${}^{40}$ Ar/ ${}^{39}$ Ar 470 471 method in step-heating using a CO<sub>2</sub> laser probe coupled with a MAP 215 mass spectrometer. 472 The analytical procedure is described by Ruffet et al. (1991, 1995, 1997). The five argon 473 isotopes and the background baselines were measured in eleven cycles, in peak-jumping 474 mode. Blanks were performed routinely each first or third/fourth run and subtracted from the 475 subsequent sample gas fractions. All isotopic measurements account for K and Ca isotopic 476 interferences, mass discrimination and atmospheric argon contamination. Apparent age errors are plotted at the  $1\sigma$  level and do not include the errors on the  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}\kappa$  ratio, age of the 477 478 monitor and decay constant.

A plateau is obtained when calculated  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{K}$  ratios of at least three consecutive 479 steps, containing a minimum of 70% of the <sup>39</sup>Ar released, agree with the weighted mean 480 calculated  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{\text{K}}$  ratio of the plateau segment. This 70% threshold is a requirement 481 482 criterion that we use in all our geochronological studies, and it is much more restrictive than 483 the 50% threshold set subjectively (in their own words) used by Fleck et al. (1977) and still adopted by most users of the <sup>40</sup>Ar/<sup>39</sup>Ar method (e.g. Schaen et al., 2021). Since the amount of 484 <sup>39</sup>Ar<sub>K</sub> released is not a disqualifying criterion, we use the concept of pseudo-plateau (Cheilletz 485 et al., 1999), which meets the same statistical criteria except for the cut-off threshold (<70% 486 487 of  ${}^{39}Ar_{\rm K}$ ), and which is much more efficient for the analysis of complex mineral phases. In 488 our detailed analysis of spectra, we also use a plateau-like descriptive terminology which 489 refers to a weighted average age calculated over segments of apparent ages, not necessarily 490 joined (e.g. edges of a saddle-shaped segment), but which are considered to be geologically significant. The errors on the  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{K}$  ratio and age of the monitor and decay constant are 491 492 included in the final calculation of the error margins on the pseudo- and plateau ages or for apparent ages individually cited.  ${}^{40}\text{Ar}/{}^{39}\text{Ar}$  ages are provided with 1 $\sigma$  errors. Analytical data, 493 494 parameters used for calculations (isotopic ratios measured on K and Ca pure salts; mass

495 discrimination; atmospheric argon ratios; J parameter; decay constants, etc.) and reference 496 sources are available in the complementary data repository.

497

#### 498 **5.2 Isotopic closure temperatures**

499 A radiochronological system is the combination of the crystal lattice of a mineral 500 phase and an isotopic pair consisting of a radiogenic element and its radioactive precursor, 501 e.g. <sup>40</sup>Ar\* and <sup>40</sup>K, the latter (K) being or not a constituent of the lattice. The analysis of isotopic systems, particularly those involving noble gases, has long focused on the behaviour 502 503 of the radiogenic element, in this case <sup>40</sup>Ar\*, the host minerals being themselves considered 504 inert and the fate of the isotopic system being exclusively governed by the diffusion of the gas 505 under temperature control, as described by Fick's laws. This led to the notion of isotopic 506 blocking temperature initially introduced by Jäger et al. (1967), formalized as the isotopic 507 closure temperature (T<sub>c</sub>) by Dodson (1973) and then nuanced with the notion of isotopic 508 closure window by Dunlap (1997). Therefore, during cooling and for a given isotopic system 509 (mineral and isotopic pair), the isotopic chronometer initializes when diffusion becomes 510 negligible because the temperature has dropped below its isotopic closure temperature.

511 T<sub>c</sub> for amphibole is difficult to estimate. In addition to the effect of cooling rate and 512 size of diffusion domains, amphibole diffusivity is influenced by ionic porosity (Fortier and 513 Giletti, 1989; Dahl, 1996). On the basis of natural hornblende compositions, Dahl (1996) 514 suggested a T<sub>c</sub> range of 480-550°C, which has been re-evaluated at 550-560°C by Villa 515 (1998) on the basis of Kamber's et al. (1995) experiments suggesting that, depending on the 516 crystal lattice characteristics, hornblende may form a closed system to Ar diffusion at 517 temperature as high as 580°C and a cooling rate of 0.7°K/Ma (Villa et al., 1996). In this study, 518 we will therefore consider a median T<sub>c</sub> range value of 550-600°C for amphiboles.

519 T<sub>c</sub> for biotite ranges from 250°C to 400°C (Harrison et al., 1985; Montigny, 1985; 520 McDougall et Harrison, 1988; Spear, 1993). In the absence of deformation and fluid-related 521 recrystallization, it may be at least as high as ~450°C (e.g. Villa and Puxeddu, 1994). With 522 diffusion domains in the range 100-1000  $\mu$ m and a cooling rate lower than 100°C/Ma, T<sub>c</sub> for 523 biotite is ca. 400°C (e.g. Grove and Harrison, 1996). We will here consider a T<sub>c</sub>-biotite range 524 value of 350-450°C.

525 Purdy and Jäger (1976) originally estimated T<sub>c</sub> of white micas at 350°C, by calibration 526 of isotopic ages versus metamorphic grade. This estimate was erroneous and underestimated 527 (Villa, 1998). Subsequent studies have shown that in the absence of deformation and fluid528 enhanced recrystallisation, T<sub>c</sub> for muscovite may reach 500°C (Villa, 1998; Bosse et al., 529 2005), and even temperatures higher than 550°C for phengite under blueschist facies 530 metamorphic conditions (Lister and Baldwin, 1996). Using experimental muscovite diffusion 531 coefficients yielded by Harrison et al. (2009), Pitra et al. (2010) calculated a 390-590°C 532 isotopic closure temperature range for white micas, using various effective diffusion radii 533 (100 to 1000 µm) and cooling rates (1 to 1000°C/Ma). Considering diffusion domain sizes of 534 100-1000 µm and cooling rates between 1°C and 10°C/Ma, a more reasonable and realistic 535 assessment for the cooling of Archean rocks (e.g. Chardon et al., 2009), isotopic closure of 536 white micas would occur in the temperature range 400-500°C.

537 The use of above  $T_c$  range estimates should only concern crystals unaffected by 538 deformation and/or recrystallization processes, (e.g. Villa, 1998). In contexts where these 539 phenomena occur, the other component of the radiochronological system, the crystal lattice of 540 the mineral carrying the istopic system, must also be considered because it may no longer 541 remain inert. Deformation can damage crystal lattices, enhances Ar diffusion, and promotes 542 the formation of intra-crystalline domains with lower diffusion length-scales and lower T<sub>c</sub>. In 543 the presence of fluids, minerals (or intracrystalline sub-domains) may recrystallize and neo-544 grains may possibly grow, two configurations which, if they occur at  $T \leq T_c$  and after initial 545 isotopic closure, will induce the recording of younger ages. Consequently, distinct intra-546 crystalline domains with different chemical and isotopic compositions, and thus distinct ages, 547 may coexist within the same mineral/grain.

548 Cheilletz et al. (1999), Alexandrov et al. (2002) or Tartese et al. (2011) identified and 549 detailed a process of sequential fluid-assisted resets of K-Ar isotopic chronometer of white 550 mica from leucogranites. Tremblay et al. (2020) showed that K-Ar isotopic system of white 551 micas from auriferous quartz veins and hosting shear zones of the Bourlamaque pluton (Val 552 d'Or mining district, Abitibi) was driven by sequential dynamic recrystallization 553 (dissolution/recrystallization/neocrystallization/aggregation) linked to successive periods of 554 fluid circulation and deformation which occurred below T<sub>c</sub>. Dunlap (1997) suggested that, 555 during ductile deformation, K-Ar isotopic system of white micas from mylonites could also 556 be partly driven by neocrystallization processes below T<sub>c</sub>, at 250-350°C. <sup>40</sup>Ar/<sup>39</sup>Ar amphibole 557 analyses (e.g. Tremblay et al., 2020) suggest that they also behave in the same way. 558

559 **5.3 Data processing** 

17

560 Metamorphic minerals such as amphiboles and micas extracted from rocks deformed 561 or disturbed by fluids are usually texturally complex (recrystallized, composite grains, etc). The <sup>40</sup>Ar/<sup>39</sup>Ar analysis of such crystals is very useful, as information it provides integrates the 562 563 effects of the disturbing processes and therefore the timing of their occurrence, even if this 564 occurs below Tc (e.g. Cheilletz et al. 1999; Hames et al., 2008; Castonguay et al., 2007; 565 Tremblay et al. 2020). Hence, due to the complexity of  ${}^{40}$ Ar/ ${}^{39}$ Ar results that arise from the 566 analysis of such materials, which are common in old rock series such as the Archean, their 567 processing must go beyond the simple visual examination of age spectra.

568 De Putter et al. (2015), De Putter and Ruffet (2020) and Tremblay et al. (2020) presented and developed tools for the visualization and processing of complex <sup>40</sup>Ar/<sup>39</sup>Ar 569 570 systems. Three of these tools are classical: the plateau (Fleck et al., 1977) and pseudoplateau 571 age (PPA) (Cheilletz et al., 1999) concepts, and the probability density diagrams, which have been first used for <sup>39</sup>Ar-<sup>40</sup>Ar data by Deino & Potts (1992). Other tools are more atypical; (i) 572 573 the degassing and the weighted age spectra, which were designed to visualize Ar degassing kinetics of  ${}^{40}\text{Ar}/{}^{39}\text{Ar}$  experiments, and (ii) the resizing of age spectra which was developed to 574 575 ensure their statistical representativeness.

576 Amphibole <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> spectra transcribe the evolution of the Ca/K ratio (<sup>37</sup>Ar<sub>Ca</sub> # Ca 577 and <sup>39</sup>Ar<sub>K</sub> # K with <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> = CaO/K<sub>2</sub>O / 2.179; Deckart et al., 1997) during step-heating 578 (degassing) experiments. Hence, a flat <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> segment (homogeneous Ca/K ratio) 579 expresses the degassing of an amphibole component devoid of compositional zoning and/or 580 inclusion of another mineral species.

581

### 582 5.4 Sampling strategy

To document and compare the tectonothermal history of the AGB and OPB on both sides of the contact zone, we performed a series of step-heating <sup>40</sup>Ar/<sup>39</sup>Ar analyses of micas and amphiboles collected in metamorphic and plutonic units of both terranes. Sixteen samples are from the Lac-au-Goéland area, and two from the AGB-OPB contact in the Chapais-Chibougamau area. Sample locations are shown in figures 3 and 4 whereas their geographic coordinates are in Table 1.

589 Four amphibolite samples (YD0009A, YD0048A, YD0039A, and YD0100A) and a 590 garnet-staurolite micaschist sample (JF4065A) are from the AGB. These rocks show a well-591 developed S<sub>2</sub> foliation marked by the preferred orientation of amphibole and/or biotite. In thin 592 sections, hornblendes from sample YD0009A and tremolites/actinolites from sample 593 YD0039A are homogeneous and unaltered (Fig. 8e), whereas dark green hornblendes 594 YD0048A and YD0100A show light-green rims or patches suggesting intracrystalline 595 zonation during crystal growth or recristallization. Biotites from samples YD0009A and 596 YD0100A are green-colored euhedral crystals whereas those from sample JF4065A (a 597 metasedimentary rock) are dark-brown-colored euhedral-(Fig. 8c). There is no optically 598 discernible occurrence of secondary chlorite in these samples. Sample YD0037, from which 599 were extracted amphibole and biotite grains, is a slightly deformed diorite that shows 600 petrographic characteristics similar to the Rivière Inconnue Diorite (Fig. 3), which yielded a 601 U-Pb zircon age of  $2693.6 \pm 0.6$  Ma (McNicolls and Goutier, 2008).

YD0018B, YD0020A, YD0063A, YD0035A, JG1245A, and YD0101A are tonalite
samples from the OPB. These tonalites are mainly composed of quartz and plagioclase and
contain various amounts of amphibole and/or biotite, which mark the regional foliation (S1).
Amphiboles YD0018B, YD0020A, YD0063A and YD0101A are sub-automorphous. All
samples contain green- to brown-colored biotites that are commonly euhedral and only rarely
chloritized.

Four samples of quartz-feldspar porphyry dykes were selected in both the AGB (YD0023A, YD0048B, and YD0009B) and the OPB (YD0023A). These dykes contain randomly oriented, coarse-grained muscovite crystals lacking microscopic evidence of recrystallization and/or deformation. A biotite grain from sample YD0009B was also analysed.

Two mylonite rock samples, YD0055A and YD0059A, from the Nottaway River shear zone were analysed. Sample YD0055A is a felsic rock belonging to the AGB. It shows a welldeveloped mylonitic foliation and contains abundant asymmetric muscovite crystals showing mica fish structures indicative of dextral shearing. These muscovite crystals also show secondary rims of neoformed micas typical of recrystallization. Sample YD0059A is a foliated tonalite from the OPB, it contains coarse-grained biotites that appear slightly altered and corroded. Most of these biotite grains also show fine rims of recrystallization.

Finally, we analyzed one sample of syenite of the Montviel alkaline intrusion (Goutier, 2005), which is associated to a carbonatite facies that has been dated at  $1894 \pm 4$  Ma (U-Pb on zircons; David et al., 2006).

623

624 **5.5 Results.** 

19

625 The step-heating  ${}^{40}$ Ar/ ${}^{39}$ Ar results are presented on Figures 10 to 14. The distinct age 626 spectra provided by duplicate analyses of amphibole and mica presented in the following 627 sections confirm, as already stressed by Tremblay et al. (2020), that the interpretation of 628 Archean material using the  ${}^{40}$ Ar/ ${}^{39}$ Ar method is challenging and requires a detailed analysis of 629 results.

630

### 631 5.5.1 The Abitibi Greenstone Belt.

The <sup>40</sup>Ar/<sup>39</sup>Ar results for the AGB amphiboles and biotites are show in Figure 10. 632 633 Analysis of amphibole 1 from sample YD0037A provides a rather flat age spectrum, with 634 nevertheless a deflection of apparent ages in the intermediate temperature steps (Fig. 10a). 635 The apparent ages, which are concordant on both sides of that deflection, allow the calculation of a plateau-like age at 2680.1  $\pm$  3.6 Ma (56.3% of the total <sup>39</sup>Ar<sub>K</sub> released), 636 637 whereas the base of the deflection yields  $2637.2 \pm 9.0$  Ma. The deflection of apparent ages is 638 interpreted as the expression of partial recrystallization due to a disturbing event younger than 639 c. 2637 Ma of an amphibole grain whose isotopic system initially closed at c. 2680 Ma during 640 cooling of the diorite below Tc. Degassing peaks expressed during this experiment confirm 641 the coexistence of these 2 radiogenic components. The higher degassing rate of the oldest 642 component, at c. 2680 Ma, in the first part of the spectrum and the homogeneous isotopic 643 composition on that segment indicate that it is perfectly individualised, which is, however, 644 apparently not the case for the younger component associated with recrystallization in the 645 intermediate temperature domain of the spectrum where the two radiogenic components 646 overlap. The concomitant decrease of apparent ages (from c. 2749 Ma to c. 2681 Ma), combined with calculated  ${}^{37}\text{Ar}_{\text{Ca}}/{}^{39}\text{Ar}_{\text{K}}$  (#Ca/K) ratios (from c. 9.7 to 7.2) in the low to 647 intermediate temperature steps of age, the  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  spectra of amphibole 2 of sample 648 649 YD0037A and the overall saddle shape of the age spectrum suggest the incorporation of 650 excess argon. Calcic domains constitute reservoirs for excess argon as shown by Ruffet et al. 651 (1990) for rimmed and patchy-zoned hornblendes. However, the convergence between the 652 pseudo-plateau age calculated for the bottom of the saddle, at  $2683.7 \pm 4.9$  Ma (15.4% of the total <sup>39</sup>Ar<sub>K</sub> released), and the plateau-like age calculated for amphibole 1 supports the above 653 654 estimated age (c. 2680 Ma) of an initial isotopic closure of amphiboles from YD0037A. The 655 concordant plateau age at 2678.5  $\pm$  2.9 Ma (85.6% of the total <sup>39</sup>Ar<sub>K</sub> released), calculated 656 from age spectrum of a biotite from sample YD0037A (Fig. 10a), is consistent with that 657 interpretation and suggests a fast cooling path for that poorly-deformed diorite.

658 The two amphiboles of sample YD0039A, although appearing homogeneous and 659 unaltered, also yielded distinct age spectra (Fig. 10b). Amphibole 1 yields a plateau age at  $2673.1 \pm 4.0$  Ma (86.6% of the total <sup>39</sup>Ar<sub>K</sub> released), slightly younger than the inferred age of 660 661 initial closure for amphibole YD0037A, whereas amphibole 2 yielded a disturbed age 662 spectrum showing an overall increase of apparent ages from c. 2548 Ma to c. 2620 Ma. These 663 discrepancies are reflected in the respective  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  spectra of the two experiments, 664 which differ in terms of ratio values and overall trend for low to intermediate temperature 665 steps. Microscopic observations reveal that some amphibole grains of YD0039A contain 666 biotite lamellae along cleavage planes, which is attested by the presence of a small degassing 667 peak in the first heating steps of the staircase-shaped age spectrum (amphibole 2 of Fig. 10b). 668 If these biotite lamellae are as young as the biotite of amphibolite YD0009A (c. 2564 Ma, see 669 below), as suggested by a low temperature pseudo plateau at  $2548.3 \pm 5.6$  Ma, their 670 degassing, partly concomitant with that of amphibole, may account for the staircase shape of 671 the age spectrum. Since amphiboles from sample YD0039A are tremolite/actinolite rather than hornblende, their calculated  ${}^{37}\text{Ar}_{\text{Ca}}/{}^{39}\text{Ar}_{\text{K}}$  ratios are significantly higher (>20) than those 672 673 of the other amphiboles (hornblende), which are usually <10 (Figs. 10a, 10c, 10d and 10e).

674 Except for a weak deflection of apparent ages in the intermediate to high temperature 675 steps, the age spectrum of amphibole YD0009A (Fig. 10c) yields a plateau-like age of 2677.3  $\pm$  3.6 Ma (77.8% of the total <sup>39</sup>Ar<sub>K</sub> released), which is concordant with the c. 2675-2680 Ma 676 677 amphibole cooling age calculated for samples YD0037A and YD0039A. A green biotite grain 678 from YD0009A yields a plateau age approximately 110 m.y. younger at 2564.3  $\pm$  3.4 Ma 679 (97.7% of the total  $^{39}$ Ar<sub>K</sub> released; Fig. 10c). However, the increase of apparent ages of its 680 high temperature steps, up to c. 2571 Ma (Fig. 10c), possibly testifies to a nearly complete 681 reset of its isotopic system at or after c. 2560 Ma and subsequent to its initial closure. The 682 above-mentioned subtle deflection of amphibole age spectrum probably correlates with the 683 younger apparent ages (from c. 2612 to c. 2630 Ma) of the low temperature steps, which are characterized by  ${}^{37}\text{Ar}_{\text{Ca}}/{}^{39}\text{Ar}_{\text{K}}$  (#Ca/K) ratios lower than those of the main phase at c. 2678 Ma 684 685 (c. 5 vs. c. 6.5). These low-temperature steps define a specific degassing peak that probably 686 express, as for amphibole YD0039A (see above), the degassing of scarce biotite lamellae 687 plated in amphibole cleavage planes (e.g. Castonguay et al., 2001).

688 The age spectrum of amphibole YD0048A (Fig. 10d) is consistent with thin section 689 examination which shows light-green rimmed and/or patchy zoned minerals indicative of 690 patchy recrystallisation of amphiboles. The isotopic record of cooling at c. 2675-2680 Ma

- 691 shown by samples YD0037A, 39A and 0009A was erased, at least locally on a grain scale, 692 during a recrystallisation event at  $2643.9 \pm 3.5$  Ma. The analysis of the degassing rate of 693 amphibole YD0048A supports this interpretation because, unlike amphibole YD0037A, the
- 694 main radiogenic component (i.e. expressed by the highest degassing rate; Fig. 10d) is the
- 695 radiogenic component associated with recrystallisation.
- 696 The light-green rimmed and/or patchy zoned amphibole from sample YD0100A, 697 texturally similar to those of YD0048A, provides a roughly flat but bumpy age spectrum (Fig. 698 10e) that yields a plateau-like age at  $2598.3 \pm 3.0$  Ma (67.3% of the total 39ArK released), 699 younger than amphibole cooling and/or disturbance ages of previous samples. It may 700 represent the best estimate of the age of the post-initial cooling disturbance recorded by 701 amphibole 1 from diorite YD0037A. A green euhhedral biotite grain from sample YD0100A 702 provides a hump-shaped age spectrum with maximum apparent ages at c. 2494 Ma. This age 703 spectrum shape indicates that the biotite grain is chloritized (Ruffet et al., 1991), which casts 704 doubt on the reliability of this age.
- 705 A brown biotite grain from sample YD4065A yields a staircase-shaped age spectrum 706 whose flat segment defined by high temperature steps provide a pseudo-plateau age at 2644.3 707  $\pm$  3.6 Ma (Fig. 10f). Such a staircase-shaped age spectrum is typical of <sup>40</sup>Ar\* loss by volume 708 diffusion induced by a thermal event subsequent to initial crystallization and/or cooling (e.g. Turner, 1968). The apparent age at  $2571.8 \pm 9.4$  Ma of the low-temperature step provides a 709 maximum estimate for the age of that <sup>40</sup>Ar\* loss event. Its concordance with the plateau age 710 711 yielded by biotite from sample YD0009A suggests that <sup>40</sup>Ar\* loss most likely occurred around c. 2560 Ma. In this scenario of <sup>40</sup>Ar\* loss, the age at c. 2644 Ma calculated from the 712 713 high-temperature steps should only be a minimum estimate of the age of isotopic closure 714 during initial cooling. Nevertheless, its perfect agreement with the recrystallization age of 715 amphibole YD0048A, also at c. 2644 Ma, argues for its validity.
- 716

#### 717 5.5.2 The Opatica Plutonic Belt.

#### 718 Amphibole and biotite age spectra from the OPB are show in Figure 11. All samples 719 were collected in tonalites.

720 Amphibole age spectra of samples YD0018B, YD0020A, YD0063A, and YD0101A 721 are more disturbed than those of the AGB. The effects of thermal disturbances following 722 initial isotopic closure appear to be quite significant, especially those related to an event 723 equivalent to the c. 2644 Ma disturbance recorded by the AGB amphiboles. The analysis of

- degassing patterns of amphiboles YD0020A and YD0063A (Figs. 11a and b) indicates that
- the radiogenic component associated with partial recrystallization induced by this disturbing
- event constitutes the main radiogenic component of these amphiboles. Related segments of
- age spectra allow the calculation of concordant plateau ages at  $2644.5 \pm 3.8$  Ma (72.6 % of
- the total <sup>39</sup>Ar<sub>K</sub> released) and 2648.4  $\pm$  3.0 Ma (69.6 % of the total <sup>39</sup>Ar<sub>K</sub> released),
- respectively.

730 Several elements of the <sup>40</sup>Ar/<sup>39</sup>Ar experiments suggest that this latter event affected 731 amphiboles with initial ages of 2670-2665 Ma, which correspond to the apparent age bracket 732 yielded by high-temperature steps of amphiboles YD0020A and YD0063A (Figs. 11a and b), 733 at 2668.8  $\pm$  8.9 Ma and 2672.3  $\pm$  8.9 Ma, respectively. The validity of these high temperature 734 apparent ages is supported by consistencies with pseudo-plateau ages at  $2665.6 \pm 4.4$  Ma and 735  $2668.9 \pm 5.3$  Ma calculated from the disturbed age spectra of amphiboles YD0018B and 736 YD0101A (Figs. 11c and d). The degassing patterns of the latter amphiboles, especially 737 YD0101A (Fig. 11d), indicate that the related radiogenic component is a major component of 738 their respective isotopic system. However, some aspects of the age spectrum shapes of 739 YD0018B and YD0101A remain enigmatic; for instance, the breakdown of the high-740 temperature apparent ages of amphibole YD0018B (down to  $2588.8 \pm 9.2$  Ma, Fig. 11c) coinciding with an abrupt and unusual drop of the  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratio (down to c. 2.69), which 741 742 may be due to the degassing of a non-amphibolitic component. In addition, the concomitant increase of the high temperature  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  (#Ca/K) ratios of amphibole YD0101A and 743 744 decrease of apparent ages (Fig.11d) are unexpected. Such unusual behaviours may be due to a 745 vounger disturbing event, as suggested by biotite plateau ages at  $2600.3 \pm 3.0$  Ma (94.8 % of the total <sup>39</sup>Ar<sub>K</sub> released), 2583.4  $\pm$  4.5 Ma (70.7 % of the total <sup>39</sup>Ar<sub>K</sub> released) and 2588.2  $\pm$ 746 747 3.5 Ma (96.3 % of the total  $^{39}$ Ar<sub>K</sub> released) yielded by samples YD0020A, YD0063A and 748 YD0035A, respectively (Figs. 11a, b and e). The existence of even younger disturbing events 749 is suggested by biotites YD1245A (pseudo-plateau age at  $2544.3 \pm 3.5$  Ma) and YD0101A 750 (plateau age at  $2525.4 \pm 3.1$  Ma) (Figs. 11d and f). However, the chlorite alteration of some biotite grains (hump-shaped age spectrum YD1245A) and/or the apparent occurrence of <sup>40</sup>Ar\* 751 752 loss (staircase-shaped age spectrum YD0101A) makes these latter age estimates hazardous. 753

754 5.5.3 Granitic dykes.

Three muscovite single grains from quartz-feldspars granitic dykes collected in the
 Lac-au-Goéland area, YD0009B, YD0023A and YD0048B, provided rather disparate results

757 (Fig. 12). The saddle-shaped age spectrum of muscovite YD0009B (Fig. 12a) suggests the 758 presence of two radiogenic components whose coexistence probably relates to a fluid-induced 759 partial recrystallization, (e.g. Cheilletz et al., 1999; Alexandrov et al., 2002; Castonguay et al., 760 2007; Tartèse et al., 2011; Tremblay et al., 2011; 2020). In such type of spectra, the plateau-761 like age at 2602.1  $\pm$  4.5 Ma (85.6 % of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 12a) yields a minimum 762 age estimate of the former muscovite, which was partially recrystallized at or after 2567.2  $\pm$ 763 6.2 Ma, i.e. the apparent age yielded by the base of the saddle. The degassing pattern and the 764 weighting of this age spectrum indicate that the older component is overwhelmingly the main 765 radiogenic component to which it is reasonable to attribute an age of c. 2603 Ma. (Fig. 12a).

Muscovite YD0048B provides a plateau age at  $2517.2 \pm 2.4$  Ma (93% of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 12b) but a detailed examination of that age spectrum reveals a complex thermal history for this mineral. Apparent ages increase steadily from low to high temperature steps, suggesting the occurrence, around 2500 Ma, of a disturbing event which induced <sup>40</sup>Ar\* loss by volume diffusion (Turner, 1968) on a muscovite grain older than c. 2524 Ma (Fig. 12b).

The first 30% of the  ${}^{39}$ Ar<sub>K</sub> degassing of muscovite YD0023A is highly disturbed (Fig. 772 773 12c) with erratic apparent ages which could result from the occurrence of interfoliar chlorite. 774 Overall, this age spectrum shows a saddle shape, as for muscovite YD0009B, which allows the calculation of pseudo-plateau ages at  $2603.2 \pm 4.3$  Ma (21.6% of the total <sup>39</sup>Ar<sub>K</sub> released) 775 for the base of the saddle and at 2627.1  $\pm$  3.7 Ma (47.4% of the total <sup>39</sup>Ar<sub>K</sub> released) for the 776 777 high temperature steps (Fig. 12c). Our preferred interpretation is that muscovite YD0023A 778 crystallized at the latest at c. 2627 Ma and has been affected by a fluid-induced partial 779 recrystallization event at c. 2603 Ma, which has been coeval with initial crystallization of 780 muscovite YD0009B (Fig. 12a).

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- 782

### 2. 5.5.4 The Nottaway River shear zone.

The analyzed muscovite of sample YD0055A belongs to a population of mica-fish muscovite grains (Fig. 7e) extracted from a felsic mylonite of the Nottaway River shear zone. As for muscovite YD0023A, the first 20% of the degassing spectra of YD0055A is highly disturbed (Fig. 13a), which can also reflect the occurrence of interfoliar chlorite. The remaining 80% of the degassing produces a multi-stepped age spectrum with a global staircase shape interspersed with three flat segments. Similar muscovite age spectra have been described by Tremblay et al. (2020) for Archean rocks of the southern Abitibi suprovince; they were interpreted as the result of the aggregation, via dissolution/recrystallization/

- neocrystallization of increasingly younger white mica sub-grains as deformation increments
- and coeval fluid inflows progress. Pseudo-plateau ages calculated from the flat segments
- suggest periods of aggregation/remobilisation of muscovite YD0055A at  $2565.2 \pm 3.7$  Ma and
- 794  $2504.0 \pm 5.0$  Ma, at the expense of a muscovite grain that would have initially crystallized at

795  $2595.6 \pm 5.4$  Ma (Fig. 13a). These aggregation/remobilisation processes, which partially 796 preserve the previous or initial states of white micas, occur below their isotopic closure

- 797 temperature, at temperatures as low as 250-350°C.
- Despite a possible plateau age at 2527.8  $\pm$  2.9 Ma (76.2% of the total <sup>39</sup>Ar<sub>K</sub> released), the age spectrum of biotite YD0059A, with recrystallized rims (Fig. 13b), shows a regular increase of apparent ages from low to high temperature steps, suggesting the occurrence of a disturbing event at c. 2505.5  $\pm$  8.5 Ma of an initial grain that would be older than c. 2531 Ma. The highly disturbed age spectrum of an amphibole from that sample is useless (Fig. 13b).
- 803

### 804 5.5.5 The Montviel Syenite.

805 Two amphiboles and a biotite from sample YD5143A from the Montviel Syenite (see 806 Fig. 3 for location) provide highly contrasted but significant results (Fig. 14). The age spectrum of amphibole 1, a hornblende according to its  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratio values of 4.0-4.2, 807 808 allows the calculation of a plateau age at 2653.9  $\pm$  2.7 Ma (83.6% of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 14). The age and  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  spectra of amphibole 2 are clearly different in the sense 809 810 that, from low to high temperature steps, both the apparent ages and  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratios increase in unusual proportions, from c.  $1934.4 \pm 7.3$  to  $2911.4 \pm 10.8$  Ma and 2.1 to 101.6, 811 812 respectively, which is not complying with what can be expected for hornblendes. The unrealistic age of the fusion step (at c. 2911 Ma) and its  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratio higher than 100 813 814 suggest the incorporation of excess argon, most likely by a calcic component resulting from 815 the alteration of the amphibole during a disturbing event which would have occurred at c. 816 1934 Ma or after (Fig. 14).

817 The age spectrum of biotite YD5143A is also strongly discordant with that of 818 amphibole 1. However, apparent ages of the base of its saddle shape, in the range c.1967-819 1940 Ma (Fig. 14), are consistent with the timing of the disturbance that presumably 820 destabilized amphibole 2. The destabilisation process possibly results from the emplacement 821 of the nearby Montviel Carbonatite, whose age is constrained by a U–Pb zircon age of 1894  $\pm$ 822 4 Ma (David et al., 2006). 823

### 824 **5.6 Interpretation and significance of** <sup>40</sup>**Ar**/<sup>39</sup>**Ar ages**

A synthetic compilation of all <sup>40</sup>Ar/<sup>39</sup>Ar ages measured during this study is show in
Figure 15.

827 The probability density diagram established from resized age spectra for amphiboles 828 and biotites of the AGB (Fig. 15a) allows a synthetized interpretation of analytical results. On 829 Figure 165a, the principal age peak is at  $2678.1 \pm 1.7$  Ma, as defined by amphiboles YD0009A 830 and YD0039A, as well as amphibole and biotite of sample YD0037A. The concordance of 831 biotite and amphibole ages suggests that they are cooling ages, and that this cooling has been relatively fast between 600°C and 400°C. The asymmetrical and stepped base of that peak 832 833 toward younger ages reflects the impact of subsequent disturbing events on amphiboles after 834 initial cooling. The first of these events occurs around c. 2644 Ma. It induced a partial 835 recrystallization of amphibole YD0048A and probably the total reset of biotite YD4065A 836 (Fig. 10f). The second disturbing event occurred around c. 2598 Ma (Fig. 15a) and induced 837 the complete recrystallization of amphibole YD0100A (Fig. 10e). It would have been 838 followed by a third event (with a frequency peak at c. 2564 Ma; Fig. 15a) that would have 839 caused an almost complete age resetting of biotite YD0009A and a partial reset for micaschist 840 biotite YD4065A (Figs. 10c, f and 15a).

Figure 15b suggests that OPB amphiboles initially cooled down to Tc at c. 2668 Ma,
before a partial recrystallization event that occurred at c. 2647 Ma. Hence, there seems to be a
ca. 10 m.y. time lag between the AGB and the OPB for initial cooling below Tc for
amphiboles, at approximately 2678 Ma and 2668 Ma, respectively (Figs. 15a and b).
The OPB tonalites also record a younger disturbing event around c. 2593 Ma (Fig. 15b).
These two latter post-cooling events correlate within errors to those recorded by the AGB
amphiboles at c. 2644 Ma and c. 2598 Ma (compare Fig. 15a and b).

The three analyzed muscovite grains from granitic porphyry dykes (Fig. 15c) are indicative of a complex and polyphased thermal history that can not be precisely understood here due to the small number of analyses. This thermal evolution may have started earlier than c. 2627 Ma and ended around c. 2495-2500 Ma (Fig. 15c). The effective occurrence of a disturbing event at c. 2602.7  $\pm$  3.1 Ma, as reported on the probability density diagram of Figure 15c, remains however proven.

The initial crystallisation of muscovite YD0055A, around 2600 Ma (Figs. 13a and 15d), can be confidently attributed to shear deformation along the Nottaway River Shear

- 856 Zone. <sup>40</sup>Ar/<sup>39</sup>Ar age data suggest that the shear zone was discontinuously active for, at least,
- ca. 100 m.y., from  $2595.6 \pm 5.4$  Ma, the minimum age estimate for white mica initial
- 858 crystallisation (or the complete resetting of the Ar «clock» of pre-existing muscovites) to final
- 859 shear increments around c. 2500 Ma. However, this does not exclude the possible initiation of
- the Nottaway River Shear Zone before 2600 Ma.
- <sup>40</sup>Ar/<sup>39</sup>Ar analyses suggest that the Montviel Syenite (Figs. 14 and 15e) and the nearby
  carbonatite represent two different magmatic pulses, an older one related to syenite
  emplacement at c. 2654 Ma and a younger one corresponding to the carbonatite intrusion at c.
  1894 Ma. This is, however, in contradiction with a model proposed by Nadeau et al. (2014) in
  which both the syenite and the carbonatite are attributed to a single phase of magmatism at
  1894 Ma.
- 867

### 868 6. DISCUSSION

869 The tectonic evolution of Archean terranes involves extensive rifting/plume/arc 870 magmatism and multiple, or progressive phases of deformation and metamorphism (e.g. Cawood et al., 2009; Bédard, 2018; Brown et al., 2020; Palin et al., 2020). Understanding the 871 872 nature of the related geodynamic regime(s), and identifying the various alternative model(s) 873 to the classical plate tectonics interpretation, if any, has important implications for a better 874 comprehension of the early Earth dynamics, the processes that led to continental crust 875 formation and structural evolution, including the mineral deposit potential of Archean 876 orogenic belts (e.g. Thébaud and Rey, 2013).

877

### 878 **6.1 The AGB-OPB contact – not a subduction suture.**

879 Subduction is a geological process by which the lithosphere is recycled into the Earth's 880 mantle at convergent plate margins, creating an upper plate-lower plate (UP-LP) boundary. 881 Among the different mountain belts generated by plate convergence and collision during the 882 Phanerozoic, a significant number took place in an environment of continuing subduction and 883 accretion along sites of oceanic lithosphere consumption. These belts are known as 884 accretionary orogens, zones of type-B subduction, or Cordilleran-, Pacific- and Andean-type 885 collisions (see Cawood et al., 2009 for a review). A similar type of plate setting has been 886 suggested for the AGB-OPB boundary, with the AGB acting as the down-going plate. 887 Phanerozoic accretionary orogens are variably deformed and metamorphosed, and 888 characteristically occur as paired and parallel high-temperature (HT) and high-pressure (HP)

889 belts with metamorphic grade up to granulite and blueschist-to-eclogite facies, respectively 890 (e.g. Brown, 1993; 2010; Ernst et al., 1994). Such orogens typically include wedge-shaped 891 accretionary prisms (or subduction mélanges) that formed at the UP-LP interface. These 892 accretionary prisms are organized in thrust sheets and attributed to the offscraping and 893 underplating of material from the descending oceanic plate and of subduction-related 894 volcanic-magmatic arc system in the upper plate (e.g. Ernst et al., 1994; Cawood et al., 2009; 895 Wakabayashi and Dilek, 2011; Kusky et al., 2020; Palin et al., 2020; among many others). A 896 fore-arc basin generally occurs in between. Accretionary prisms are usually elongated and 897 depending on the plate velocity, the dip of the subduction zone, the backstop geometry of the 898 wedge and the volume amount of sediments on the down-going plate, they can be relatively 899 voluminous. The Franciscan subduction complex, for instance, is at least 50 km-wide along a 900 strike-length of c. 900 km (e.g. Ernst et al., 1994; Wakabayashi, 2017). Accretionary prisms 901 are essentially made up of tectonic slices of island-arcs, backarcs, forearcs, dismembered 902 ophiolites, granulite-facies and/or ultrahigh-pressure (UHP)-HP metamorphic rocks, and pre-903 to syn-accretion clastic sedimentary sequences. Thrusting and reverse faulting usually led to 904 the imbrication of the upper plate rocks with those of the lower plate, leading to complex field 905 relationships.

906 The subduction model proposed for the AGB-OPB boundary is mainly based on the 907 Lithoprobe seismic profile (e.g. Calvert and Ludden, 1999) and field structures attributed to a 908 south-verging fold-and-thrust belt in the AGB (Sawyer and Benn, 1993; Lacroix and Sawyer, 909 1995). However, the AGB-OPB transition does not show any of the characteristic features of 910 UP-LP boundary as expected for a subduction suture. There are no accretionary prism and no 911 metamorphic break along the AGB-OPB contact. Metamorphism varies progressively from 912 low- to high-grade from the uppermost units of the AGB toward the plutonic rocks of the 913 OPB (Fig. 9). There are, obviously, local fault contacts between greenschist- and amphibolite-914 grade rocks of the AGB, such as the Barlow fault in the Chibougamau area (Bedeaux et al., 915 2020), but these are late reverse faults related to regional folding. Our field observations and 916 structural analysis indicate that the volcanic and sedimentary sequences of the AGB overly 917 the TTG-type orthogneisses of the OPB (Fig. 9), suggesting that both domains likely formed a 918 single structural unit prior to regional deformation (e.g. Benn, 2006; Benn and Moyen, 2008). 919 Hence, we do not think that the AGB-OPB transition represents a subduction-related, upper 920 plate-lower plate boundary.

921

It may be argued, however, that the subduction suture is rather located south of the

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- 922 AGB-OPB boundary, at the contact between the Northern and Southern volcanic zones of the
- 923 AGB, for instance, that it is marked by the Destor-Porcupine-Manneville (DPM) fault zone
- 924 (e.g. Daigneault et al., 2002). Although the DPM fault zone represents a major deformation
- 925 corridor, it does not bear the characteristic lithologies and metamorphic features of a
- 926 fossilized, crustal scale subduction suture. Moreover, a fossil subduction model for the DMP
- 927 fault zone implies that the Northern Volcanic zone (the inferred upper plate) would be
- 928 significantly older than the Southern Volcanic zone (the inferred lower plate), i.e. 2730–2710
- 929 Ma for the former and 2710 2695 Ma for the latter (Daigneault et al., 2002; see also Thurston
- 930 et al., 2008). It is difficult, if not impossible, to imagine how the subduction of a «young»
- 931 volcanic sequence can generate an older one in the overlying plate.
- 932

### 933 **6.2** The formation of composite OPB-AGB – speculative models.

The structural relations and metamorphic characteristics of the AGB and OPB are
reminiscent of various tectonic settings such as (i) metamorphic core complexes (MCC, Lister
and Davies, 1989; Whitney et al., 2013), (ii) suprastructure-infrastructure transition (e.g.
Culshaw et al., 2006; Stern, 2017), two models in which the upper crust is tectonically
juxtaposed to the deep crust and/or upper mantle, and (iii) the convective overturn of crustal
material (Collins et al, 1998; van Kranendonk, 2011; Bédard, 2018; Wiemer et al., 2018;
Vanderhaeghe et al., 2019).

941

### 942 6.2.1 The OPB as a metamorphic core complex.

943 As compared to MCC's, the OPB shares similar features: (1) synkinematic granitoids, 944 pegmatites and migmatites restricted to domal core domains; (2) sheet-like granitoids and 945 discontinuous high-strain migmatites towards the contact with the overlying upper crust; (3) a 946 décollement horizon (i.e. potentially, the OPB-AGB contact) between lower crustal gneisses 947 and the upper crust; and (4) a normal metamorphic zonation. However, MCC's are generally 948 less than 10 km in diameter, asymmetrical (e.g. the detachment being folded due to the 949 bowing upward of the lower crust) and usually related to major continental extension (see 950 Lister and Davies, 1989). In contrast to detachments of typical MCC's, the OPB-AGB contact 951 has been involved in progressive reorientation and repeated late-intrusion. Moreover, the 952 ductility and metamorphic contrasts across the OPB-AGB interface are not extreme as in 953 typical MCC's in which the detachment separates rocks of highly contrasting structural and 954 metamorphic character, with deformation in the cover rocks being typically brittle. Finally,

the abundant syn-tectonic granitoids of the OPB-AGB are not confined to the proximity of thedetachment, as for MCC's, but extend into the interior of both the OPB and the AGB.

957

### 958 **6.2.2** Superstructure-infrastructure transition.

959 In superstructure-infrastructure (S-I) transition zones, superstructure represents a low-960 grade upper crust level with early, typically upright structures, whereas infrastructure 961 constitutes a high-grade migmatitic lower crust level with late, gently inclined structures overprinting the early structures of the overlying upper crust (Culshaw et al., 2006). In typical 962 963 examples, a crustal-scale detachment shear zone is observed between the upper and lower 964 tectonic levels (Tikoff et al., 2002; Culshaw et al., 2006; Stern, 2017). The infrastructure 965 gneissic rocks generally yield radiometric ages younger than superstructure rocks, indicating 966 that the infrastructure was hot, partially molten and weak during the deformation of the 967 superstructure above it (e.g. Culshaw et al., 2006; Stern, 2017). The OPB presents evidence 968 for anatectic melting and syn-kinematic migmatites formation at ca. 2.68-2.67 Ma (Sawyer, 1998), a melting event that, according to our  ${}^{40}\text{Ar}/{}^{39}\text{Ar}$  data (Figs. 15a and b), has been almost 969 970 coeval with or slightly followed the ca. 2680 Ma metamorphism in the AGB. This suggests a 971 maximum time lapse of 10 m.y. for metamorphic conditions between upper (AGB) and lower 972 (OPB) crustal levels, which could fit with the S-I model data presented by Culshaw et al. 973 (2006) for the Western Superior Province. In such interpretation, the amphibolite-facies rocks 974 occurring towards the base of the AGB would define a low-angle detachment fault at the 975 contact with deeper and hotter rocks of the OPB, a detachment that would has been 976 subsequently folded and verticalized. In the study area, these amphibolites are several km-977 thick (Bedeaux et al., 2020), and may thus reasonably represent a mid-crustal decoupling zone between a ca. 15 km-thick superstructure (the AGB) and a plutonic, TTG-rich infrastructure 978 979 (the OPB). However, the occurrence of  $F_2$  folds affecting the low-angle metamorphic  $S_1$ 980 layering in the OPB gneisses invalidates a typical S-I transition model. The metamorphic age 981 variations can be simply related to the differential uplift and exhumation of both belts (the 982 ABG first and then the underlying OPB). A S-I transition model is also inconsistent with age 983 constraints from the Attic Complex (located in the southern part of the AGB; see Fig. 1b) 984 where U-Pb zircon dating suggests a 20 m.y gap between the formation of migmatites in 985 tonalitic gneisses (TTG's) of the lower crust (at ca.  $2716.1 \pm 2.9$  Ma) and those in adjacent mafic volcanic rocks (at ca.  $2695 \pm 1$  Ma) (Revelli, 2020). This has been attributed to the 986 987 diapiric ascent and crustal melting of lower crustal intrusive material and the related burial

988 (and prograde metamorphism) of cold and dense greenstones of the upper crust (Revelli,989 2020), which agrees with a convective crustal overturn as described below.

990

### 991 6.2.3 Convective crustal overturn.

992 The third tectonic setting involves the partial convective overturn of the crust (Fig. 993 16). It was originally proposed by Collins et al. (1998) for Archean rocks of the Pilbara craton 994 (Australia). In such a setting, the OPB gneissic domes represent the diapiric response of 995 density inversion created by the accumulation of a thick (ca. 15 km) greenstone cover (the 996 AGB) erupted over a previously stable and more-or-less continuous sialic crustal basement  $\geq$ 997 2.80 Ga, both domains being intruded by syn-volcanic and syn-tectonic TTG plutons ranging 998 from c. 2.76 to c. 2.65 Ga (Davis et al. 1995; Sawyer, 1998; Daoudene et al., 20156, Pedreira-999 Perez et al., 2020) (Figs. 16a and b). Diapirism is considered here as part of a larger process 1000 involving the partial convective overturn of the crust (Collins et al., 1998; Wiemer et al., 1001 2018). We suggest that deformation initiated by sinking and subsidence of the thick cover of 1002 greenstones within a hot and weak (ductile), laterally flowing sialic lower crust that generated 1003 the S<sub>1</sub> metamorphic layering both in the OPB and the lowermost part of the AGB (Fig. 16a). 1004 This produced strong horizontal temperature gradients and related changes of metamorphic 1005 facies as, for instance, in thermal models of Thébaud and Rey (2013). Due to density 1006 contrasts, both the gneissic lower crust and crosscutting TTG's ultimately migrated upward 1007 passively as accommodation structures. Material flow into the AGB synclines would has been 1008 compensated by lateral and vertical migration of the less dense sialic substrate into the domal 1009 crests of the OPB (Fig. 16b), generating a pattern of partial convective overturns in the crust 1010 (Collins et al., 1998; Weinberg, 1997). The downflow of greenstones would has been almost 1011 entirely compensated by TTG uprises (Fig. 16c). Shallow interdome depressions developed 1012 and progressively filled with clastic sediments, and progressively filled with clastic 1013 sediments; s such clastic sequences occupying the core of greenstone synclines and being 1014 represented by sedimentary rocks of the Opémisca and Broadback groups within the Abitibi 1015 and Frotet-Evans belts, respectively. Some of these clastic sequences possibly formed as early 1016 as ca. 2750 Ma, as suggested by stratigraphic relationships between the Broadback Group and 1017 the underlying Frotet-Evans volcanics (Brisson et al., 1998; see section 2.1). In the Opémisca 1018 Group, Mmaximum sedimentation age of the Haüy Formation in the Opémisca Group ( $\leq$ 1019  $2691.7 \pm 2.9$  Ma; David et al., 2007) is coeval, at least in part, with regional metamorphism in

1020 the AGB (see section 5.5). Part of the Opémisca Group therefore deposited during regional

1021 deformation, as proposed for most sedimentary sequences of the Superior Province (e.g. 1022 Leclair et al., 1993; Parmenter et al., 2006; Lin et al., 2013). As the greenstones and overlying 1023 sedimentary rocks progressively flow deeper into the synclines, strain intensity increased and 1024 was accommodated by horizontal shortening, the S<sub>2</sub> foliation became more pervasive and 1025 faults and shear zones with down-dip lineations developed (Fig. 16b). The Barlow fault 1026 (Bedeaux et al., 2021) is one of these; in the Chibougamau area, it is located along the 1027 northern flank of the Waconichi F<sub>2</sub> syncline (Waconichi basin on Fig. 2) and marked by upper 1028 crustal, south-dipping seismic reflectors (Mathieu et al., 2020). The Barlow fault is interpreted 1029 as a reactivated structure (Bedeaux et al., 2021), and it possibly belongs to a series of basin-1030 bounding faults that accommodated the exhumation of the OPB gneisses and inverted during 1031 the following contraction (e.g. Mathieu et al., 2020), which is consistent with the 1032 interpretation shown on Figure 16. With time, the early generation of TTG's were 1033 progressively deformed and rotated into subconcordance with the steepening domal 1034 boundaries (Fig. 16c). Other TTG's were syn-kinematic and formed sheeted intrusions. Some 1035 of these did not move far from their source, remained heterogeneous, and can be better 1036 described as `intrusive diatexites' (Sawyer, 1998), whereas later intrusions tend to be larger, 1037 more discordant, less strongly foliated bodies and are more obvious near the AGB-OPB 1038 boundary, where they cut the OPB-AGB interface. Such an interpretation fits well with (1) the 1039 relative downflow of the AGB volcanic rocks as compared to the OPB gneisses (Fig. 9), (2) 1040 the distribution of  $S_1$  foliations and lineations in the OPB domes (Fig. 2 and Fig. 6b), (3) the 1041 up-stratigraphy increasing proportion of plutonic clasts in polygenic conglomerates of both 1042 the Broadback and Opémisca groups (Brisson et al., 1997a; 1998; Charbonneau et al., 1991), 1043 which reflects the progressive erosion of volcanic cover rocks and underlying plutonic rocks 1044 in the source area, and (4) slightly higher pressure and temperature preserved in the OPB 1045 migmatites and amphibolites (6.3-7.1 kb and 750°C, Sawyer, 1998) as compared to the 1046 amphibolite facies rocks of the AGB (c. 6 kb and c. 550°C, Bedeaux et al., 2021).  $^{40}$ Ar/ $^{39}$ Ar age measurements performed during our study are in good agreement with 1047 1048 that model. Amphibole ages suggest that the lowermost part of the AGB exhumed around 1049  $2678.1 \pm 1.7$  Ma (Fig. 15a). Ages from the OPB, at  $2667.8 \pm 3.0$  Ma (Fig. 15b), suggest that it 1050 was more-or-less coeval with, or quickly followed the exhumation of the AGB, an event that 1051 we suggest has been coeval with decompression melting (Weinberg, 1997; Whitney et al.,

1052 2004) and the genesis of syn- to late- $D_1$  migmatites in the cores of OPB gneiss domes at 2.67-

1053 2.68 Ga (Sawyer, 1998; Fig. 16c). Then, both terranes share the same succession of thermal
1054 disturbances at c. 2650-2645 Ma and 2600-2590 Ma, including the inferred onset of shearing 1055 along the Nottaway River Shear Zone at ca. 2600 Ma, as recorded by related mylonites (Fig. 1056 15d) and quartz-feldspar porphyry dykes (Fig. 15c). Such a series of events is also 1057 documented in the Val d'Or mining district of the Abitibi subprovince (Tremblay et al., 1058 2020). Besides, at the scale of the Superior Province, there is accumulating geochronological 1059 evidence for major crustal deformation event(s) and related metamorphism spreading over a 1060 timeframe much larger than suggested in literature (e.g. Percival, 2007). Easton (2000), for 1061 instance, argued that the distribution of metamorphic grade and age of metamorphism in the 1062 western Superior reflect a series of tectonic events between 2710 and 2640 Ma, which is in 1063 agreement with U-Pb zircon age constraints for metamorphism, syn-orogenic magmatism and 1064 deformation in the southern Abitibi subprovince (see Tremblay et al., 2020 for a review). 1065 North of the OPB, in the Opinaca subprovince, granulite-facies metamorphism spreads over 1066 ages varying from c. 2666 Ma to 2636 Ma (Morfin et al., 2013) whereas it even lasted down 1067 to ca. 2620–2600 Ma in supracrustal rocks of the La Grande subprovince (Fontaine et al., 1068 2017, 2018). Combined with geochronological data from our study, this is a strong indication 1069 that peak and duration of regional metamorphism have been more-or-less coeval and long-1070 lasting over a large region of the Superior Province, if not all over the Pontiac, Abitibi and 1071 Opatica subprovinces and beyond.

1072

### 1073 **6.3 Significance of the Lithoprobe seismic reflector.**

1074 Seismic profiles showing shallow sub-horizontal to slightly dipping reflectors beneath 1075 Archean and/or Paleoproterozoic greenstone belts are relatively common. Some of these 1076 reflectors may be due, obviously, to tectonic events that are younger than, and not directly 1077 related to, the formation of the greenstone belts and, until the age of such reflectors can be 1078 determined, their bearing on the greenstone belt evolution remains speculative. However, 1079 assuming that the Lithoprobe seismic reflector of the Superior Province shown in Figure 1 is 1080 an Archean artefact (e.g. Cawood et al., 2006) and that our interpretation of the AGB-OPB 1081 relations is correct, what can be the origin and significance of that reflector then? We envision 1082 two possibilities that we briefly discuss below, (i) mantle lithosphere imbrication (Fig. 17), 1083 and (ii) bending of a flat subduction rooted southward at the Pontiac-Abitibi interface (Fig. 1084 18).

1085 Gray and Pysklywec (2010) performed a series of forward numerical modeling of the 1086 thermo- mechanical evolution of continental lithosphere undergoing collision and orogenesis

1087 under Neoarchean- like conditions for stratified mafic-to-felsic crustal domains overlying the 1088 mantle lithosphere. They have identified three dominant modes of mantle lithosphere 1089 deformation, (1) pure- shear thickening; (2) mantle imbrication; (3) and underplating. Their 1090 result for the mantle imbrication mode is reproduced on Figure 18. For that mode, numerical 1091 modeling shows that, due to a «hot» and weak lower crust, there is a decoupling between the 1092 mantle and the crust at a depth of approximately 40 km. The model shows that the 1093 lithospheric mantle becomes imbricated along major shear zones along which slices of crustal eclogites may occur, whereas the crustal domain itself essentially absorbs deformation by 1094 1095 pure-shear thickening and folding. Gray and Pysklywec (2010) inserted a weak zone in the 1096 initial configuration of their numerical experiments in order to favor the onset and localization 1097 of deformation, and to mimic an original plate boundary (see their Figure 1) as well. Although 1098 that such a weak zone may has influenced the location of some shear zones within the mantle 1099 lithosphere, it did not lead to the formation of any shear zones (that could have been 1100 interpreted as a nascent subduction plane) in the upper crust. The modeled imbrication shown 1101 on Figure 17 is obviously unrelated to subduction since the crustal part of the lithosphere is 1102 not recycled into the mantle as in typical subduction settings. We think that mantle 1103 imbrications as shown by these results likely formed during the Archean and can account for 1104 deep seismic reflectors such as the Abitibi-Opatica Lithoprobe transect.

1105 An alternative hypothesis that cannot be ignored is that the Lithoprobe seismic 1106 reflector may not represent the original location of an Abitibi-Opatica subduction zone, but 1107 the trace of a Pontiac-Abitibi subduction zone interface that flattened for several hundreds of 1108 km at depth and steepen at or close to the Abitibi-Opatica contact. Such shallow subduction 1109 flattening has been already invoked as a predominant tectonic regime during the Archean due 1110 to thicker and more buoyant oceanic lithosphere (e.g. Cawood et al., 2006). Flat subduction is 1111 documented on modern Earth. A typical example of it is the Mexican subduction zone where 1112 the young Cocos plate is shallowly subducting beneath central Mexico (Fig. 18a; Kim et al., 1113 2012). This is well illustrated by the teleseismic imaging of a 500 km long transect between 1114 the Middle America trench to the south, and the Trans-Mexican volcanic belt (TMVB, the 1115 continental volcanic arc) to the north (Fig. 18b). This shows that subduction of the Cocos oceanic plate dips c. 15° north for about 80 km along the coast, and then horizontally 1116 1117 underplates the continental crust at c. 50 km-depth for c. 250 km northward. The slab then abruptly changes to a steeply dipping geometry at approximately 75° as it bends right below 1118 1119 the TMVB. The similarity of the geometry of the Cocos plate subduction zone and the

1120 reinterpreted Abitibi segment of the Grenville-Abitibi seismic transect is striking (Fig. 18c), 1121 both in terms of depth and lateral extent of seismic reflectors. Hence, it is possible that the 1122 steeply north-dipping seismic reflector of the Abitibi Lithoprobe transect roots into the 1123 Pontiac-Abitibi interface rather than the AGB-OPB. The Pontiac-Abitibi contact would be 1124 then interpreted as a surface of former subduction and recycling of the oceanic crust that 1125 originally existed between the Pontiac and Opatica «basements», causing the underthrusting 1126 and burial of the Pontiac block below the Abitibi subprovince and accounting for ensuing 1127 tectono-metamorphism and magmatism (e.g. Dimroth et al., 1983; Card, 1990; Camiré and 1128 Burg, 1993; Davies, 2002; Frieman et al., 2017, among many others). This may possibly 1129 explain the contrasting pre-2700 Ma plume-driven constructive stage of the AGB as 1130 compared to its post-2700 Ma geodynamic evolution as depicted by Thurston et al. (2008) 1131 and Mole et al. (2021); the latter stage being mainly characterized, according to these authors, 1132 by subduction-related compressional lithospheric deformation, flysch sedimentation and 1133 more-or-less coeval emplacement of alkalic granodiorite, granite and syenite plutons (such as 1134 in the Preissac-Lacorne batholith). However, such an this-uniformitarian interpretation for the 1135 Pontiac-Abitibi relations has been recently-disputed by Piette-Lauzière et al. (2019) that argue 1136 for a multi-stage tectono-metamorphic evolution involving extensional tectonism and crustal 1137 delamination followed by re-accretion and subcretion of ribbons continents as suggested by 1138 Bédard (2018) for the Northern Superior.

1139

# 1140 CONCLUSION

1141 Our structural and metamorphic analysis of the transition between the Abitibi and 1142 Opatica subprovinces does not confirm the former existence of north-dipping subduction 1143 between these two lithological assemblages, despite the occurrence of a major seismic 1144 reflector at depth. We think that both terranes originally formed a single and composite 1145 crustal sequence, the Abitibi and Frotet-Evans greenstone belts and associated sedimentary 1146 rocks representing the upper crust and the Opatica Plutonic Belt being the remnant of middle-1147 lower crust plutonic domains. This implies that the Frotet-Evans belt originally formed an 1148 upper crust sequence that was continuous with the AGB. Although there is a lack of precise 1149 geochronological constraints for the Frotet-Evans belt, such an interpretation is consistent 1150 with available U-Pb age data from both greenstone belts. However, it does not mean that the 1151 Frotet-Evans belt has ever been as thick as the AGB (c. 10-15 km), since there are erosional 1152 unconformities on top of each volcanic sequences that may has removed more volcanic

1153 material from the former belt (due to the progressive exhumation of the OPB; see Fig. 16) as 1154 compared to the AGB. Our interpretation also means that OPB-like orthogneissic massifs are 1155 likely to be exposed south of the OPB-AGB transition. We are not saying here that OPB 1156 gneisses are systematically present at depth below the AGB but that discontinuous masses of 1157 orthogneissic TTGs, more-or-less attached to each other and to the Opatica, are likely to occur 1158 and to be expose along the crest of antiformal/domal culminations within the Abitibi belt. Examples of such metamorphic domes may be represented by the Marest and Bernetz 1159 gneissic plutons in central Abitibi (Faure, 2015), which are currently interpreted as the crustal 1160 1161 roots of volcanic arc(s) (Chown et al., 2002), and by the Attic Complex in the Lebel-sur-1162 Quévillon area (Fig. 1b) that has been recently interpreted as the result of diapiric and 1163 sagduction-related processes (Revelli, 2020).

1164 As for most Archean cratons, the structural evolution of the Opatica and Abitibi belts 1165 is the result of several episodes of regional deformation and metamorphism that spans several tens, if not hundred's, of million years. The main events,  $D_1$  and  $D_2$ , are predominantly 1166 1167 developed in the OPB and the AGB, respectively.  $S_1$  is a penetrative amphibolite-grade 1168 foliation in the OPB gneisses and in metavolcanic rocks lying at the base of both the Frotet-1169 Evans and Abitibi greenstone belts whereas S<sub>2</sub> is a medium- to low-grade regional schistosity 1170 of variable intensity in both the AGB and OPB. S<sub>1</sub> is subparallel to the OPB-AGB contact 1171 although it is frequently transposed into a composite  $S_{1-2}$  fabric within the AGB.  $S_1$ 1172 trajectories in the OPB define a dome-and-basin geometry that is the result of vertical 1173 transfert of middle- to lower-crustal material. The counterpart progressive burial of the AGB 1174 into synformal troughs generated the abrupt EW-trending S<sub>2</sub> schistosity and related down-dip mineral and stretching lineations. Amphiboles and micas <sup>40</sup>Ar/<sup>39</sup>Ar ages measured in both the 1175 1176 OPB and AGB suggest that exhumation of the latter started at  $\geq$  2678 Ma and was almost 1177 coeval with, or slightly followed, by the uplift and related migmatization of OPB gneisses at 1178 c. 2668 Ma or earlier. This is consistent with a diapiric convective overturn of the crust as 1179 illustrated on Figure 16. Both the OPB and AGB then shared the same magmatic and 1180 structural evolution, including the onset of dextral and sinistral transcurrent shearing and 1181 faulting at ca. 2600 Ma that may have lasted for c. 100 m.y. Comparison with adjacent areas 1182 suggests that magmatism and metamorphism have been coeval over a large region (Easton, 1183 2000), which is consistent with pervasive deformation and slow cooling as expected for 1184 vertical tectonic models and diapiric magmagenesis and ascent during the evolution of the 1185 Archean Superior Province (e.g. Bédard and Harris, 2014; Bédard, 2018).

1186

## **Declaration of Competing Interest.**

- 1188 The authors declare that they have no known competing financial interests or personal
- 1189 relationships that could have appeared to influence the work reported in this paper.
- 1190

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- 1669

1670 **Figure captions.** 

1671 Figure 1. a) Principal tectonostratigraphic domains of the Archean Superior Province.
1672 Modified from Card (1990) and Benn et al. (1992). b) Geological map of the southeastern

- 1673 part of the Superior Province with focus on the Abitibi greenstone belt and the surrounding
- 1674 areas. AC: Attic Complex; BP: Bernetz gneissic pluton; MP: Marest gneissic pluton.
- 1675 Modified from Thurston et al. (2008). See a) for location.

1676 Figure 2. Geological map of the northeastern part of the northeastern part Abitibi
1677 Subprovince and the eastern part of the Opatica Subprovince including the eastern portion of
1678 the Frotet-Evans greenstone belt. Samples YD101A and YD0100A used for <sup>39</sup>Ar-<sup>40</sup>Ar dating
1679 are localized on the map. BSZ, Barlow shear zone; GF, Gwillim fault; KSZ, Kapunapotagen
1680 shear zone; LF, Lamarck shear zone; LFSZ, La France river shear zone; LLSZ, La Trève lake

- 1681 shear zone; LNSZ, Lac à l'Eau-Noire shear zone; LSRSZ, Lucky Strike river shear zone;
- 1682 NRSZ, Nottaway river shear zone. See Fig. 12 for location. Modified from Dimroth et al.
- 1683 (1986) and Daigneault et al. (1990).
- Figure 3. Geological map of the Lac-au-Goéland area and location of samples used
   for <sup>40</sup>Ar-<sup>39</sup>Ar for this study. See Fig. 2 for location. Modified from Goutier (2005).
- Figure 4. Field photographs of the AGB and OPB rocks. a) Amphibolite-grade mafic
   volcanic rock showing well-preserved pillow structure. b) Strongly deformed gabbro showing
   quart-feldspar aggregates defining a steeply plunging mineral lineation. c) Un-
- 1689 metamorphosed succession of sandstone and siltstone of the Opémisca Group (Daubrée
- 1690 Formation) in the Chibougamau area Typical gneissic tonalite of the OBP characterized by a
- 1691 penetrative S<sub>1</sub> foliation. d) Un-metamorphosed succession of sandstone and siltstone of the
- 1692 Opémisca Group (Daubrée Formation) in the Chibougamau area Typical gneissic tonalite of
- 1693 the OBP characterized by a penetrative  $S_1$ -foliation. e) AGB amphibolite located along the
- southern boundary of the Noman tonalite (see Fig. 3), showing the S<sub>1</sub> metamorphic layering
  tightly folded by F2 with an axial-planar S2 foliation. f) Granitic dykes crosscutting
  amphibolitic horizons within the OPB.
- Figure 5. Structural map of the study area showing foliation trajectories and dip
  directions. Structural data were compiled from the geological mapping data available in
  SIGEOM information system of the Ministère des Ressources naturelles du Québec
  (http://sigeom.mines.gouv.qc.ca). Trajectories of regional schistosity/foliation were drawn
  from structural data acquired in the field and high-resolution magnetic imaging.
- Figure 6. Stereographic projections (equal area, lower hemisphere) for dominant
  planar and linear structural fabrics of the Abitibi and Opatica subprovinces. Data for the AGB
  correspond to S<sub>2</sub> and composite S<sub>1-2</sub> whereas those from the OPB correspond to S<sub>1</sub>.
- 1705 **Figure 7.** Field photographs of fabrics related to the D<sub>3</sub> deformation event. a) 1706 Mylonitic fabric developed in a mafic volcanic rock of the AGB. b) Dextral F<sub>3</sub> fold affecting 1707 the regional schistosity S<sub>2</sub> within a mafic volcanic rock. c) Subhorizontal stretching lineation 1708 along the Nottaway River shear zone. d) Shear bands indicating dextral shearing in a 1709 mylonitized orthogneiss of the Nottaway River shear zone (view perpendicular to the 1710 subvertical S<sub>3</sub> foliation and parallel to the subhorizontal lineation). e) Thin section cuts 1711 parallel to the XY plane showing dextral mica fish fabrics in a felsic mylonite. f) Boudinaged 1712 felsic dyke cutting across an amphibolite horizon of the OPB.
- 1713 **Figure 8.** Photographs illustrating regional metamorphism and associated deformation

1714 in the study area. a) Low-grade volcanic rocks of the Roy Group showing well-preserved

- 1715 pillow structures. b) Garnet-staurolite micaschist from the northwestern flank of the
- 1716 Waconichi syncline. Staurolite forms large, corroded crystal preferentially aligned along the

1717 foliation plane (cross-polarized light). c) Large chloritoid phenocryst overprinting the regional

schistosity marked by the preferred orientation of actinolite in the fine-grained matrix (cross-

1719 polarized light). d) Outcrop of pillowed basalt showing garnet-rich pillow margins in the

vicinity of the AGB-OPB transition. e) Typical lobate boundaries between quartz and feldspar
grains in a coarse-grained and weakly-deformed tonalite of the OPB. The foliation is marked
by the preferred orientation of biotite (plane-polarized light). f) Field example of migmatitic
gneiss of the OPB.

Figure 9. Structural profiles of the study area. See figure 3 for location. Eastsoutheast-looking for profile A-A', East-looking for the other profiles.

1726Figure 10. Single-grain  ${}^{39}$ Ar/ ${}^{40}$ Ar results for amphiboles and biotites of the AGB.1727Apparent age error bars are at the 1 $\sigma$  level; errors in the J-parameter are not included. Plateau1728and pseudo-plateau ages (1 $\sigma$  uncertainties including errors in the J-parameter) are given when1729applicable. Sample locations are show in Figures 3 and 4. See text for explanations.

Figure 11. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for amphiboles and biotites for tonalites of
 the OPB. See Figures 3 and 4 for locations. Analytical details as for Figure 11. See text for
 explanations.

Figure 12. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for muscovite of quartz-feldspar porphyry
dykes of the Lac-au-Goéland area. Analytical details as for Figure 11. See Figure 4 for
locations. See text for explanations.

Figure 13. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for muscovite, amphibole and biotite from
the Nottaway River shear zone. Analytical details as for Figure 11. See Figure 4 for locations.
See text for explanations.

Figure 14. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for amphibole and biotite from the Montviel
Syenite. Analytical details as for Figure 11. See Figure 4 for locations and text for
explanations.

Figure 15. Synthesis of the validated ages of the study area (plateau and pseudoplateau) obtained on amphibole and micas grouped by terranes and/or rock types on probability density diagrams of the resized apparent ages for each group. See text for explanation.

1746 **Figure 16.** Schematic illustration of a partial crustal overturn model for the OPB-AGB

transition. See text for discussion and details on age constraints for the different panels.

- Figure 17. Extract of numerical models performed by Gray and Pysklywec (2010).
  This model suggests that due to a hot and rheologically-weak lower crust, the mantle and the
  crust become decoupled at a depth of approximately 50 km and that the lithospheric mantle
  becomes imbricated along major shear zones whereas the crustal domain itself 2 of Gray and
  Pysklywec (2010). See text for discussion.
- Figure 18. a) Teleseismic imaging of the subduction zone interface of the Cocos
  slab beneath Mexico. Modified from Kim et al. (2012). b) Alternative interpretation for the
- 1755 AGB showing flat subduction of the Pontiac terrane below a composite AGB-OPB terrane.
- 1756 Note the striking geometrical similarity with the Cocos subduction zone. See text for
- 1757 discussion.

### **Response to comments**

#### Dear Editor,

We sincerely thank you and the two reviewers for the time you dedicated for critical review of the second version of our manuscript. From these reviews, we made several corrections as detailed below and shown in red in the revised version of the manuscript. We hope that it now responds to the standard quality for publication in Precambrian Research.

**Reviewer #1:** The final manuscript should mention the fact that some of the 2.6 Ga and 2.5 Ga ages may reflect mid crustal processes as noted in the review by reviewer 1.

The attached Word document lists a small number of strictly grammatical points that should be repaired.

The authors have done a good job of addressing comments by reviewer 2 and there is but one area of reviewer 1's comments needing a response.

To our knowledge, and according to Thurston et al. (2008) and Mole et al. (2021) as well, there are no significant intrusions younger than 2.6 Ga in the Superior Province. 2.6 to 2.5 ages. When present, young Ar-Ar cooling ages can be attributed to progressive exhumation or localized late shearing in both the AGB and OPB (see Tremblay et al. 2020). This being said, we have added much more information regarding 2700-2650 Ma magmatism in the AGB (see revised pages 5-6, 157-188). We also have made all the grammatical corrections listed by the reviewer.

#### Reviewer #3: General comments.

This is an interesting manuscript which addresses the tectonic history of the Archean Abitibi and Opatica subprovinces by focusing on their complex structural and metamorphic contact relationships. The manuscript is much improved based on the two previous reviews and is in my view publishable with moderate revisions.

The main unresolved issues of the revised version is that it does not address the comment of reviewer #1 in that the AGB's post 2700 Ma geodynamic history is significantly different than from its pre 2700 Ma constructive stage. Pre 2700 Ma the AGB's evolution was clearly autochthonous with the seven major volcanic assemblages developed sequentially over ca. 100 Ma with conformable to disconformable contacts and only minor episodes of chemical or pelagic sedimentation, nor is there any evidence of the development of tectonic fabrics at this time (see Ayer et al., 2002, Thurston et al 2008). However, post 2700 Ma there was a major change in the geodynamic environment with development of numerous sequential compressional episodes of folding and faulting often with associated regional tectonic fabrics developed. This later compressional phase was coeval with emplacement of plutons of various sizes and compositions, but also with a change in the chemical composition of felsic magmatism (in both volcanic and synvolcanic plutons) from the pre-2700 Ma tonalitic plutons being calciumrich, changing post-2700 Ma to more alkalic compositions of granodiorite, granite and syenite and showing chemical evidence of being sourced from a different, more metasomatised mantle source. This period of transition was also accompanied by the onset of earlier flysch followed by molasse sedimentation, both lying unconformably on the older volcanic and plutonic substrate.

All these features indicate a major geodynamic change occurred around 2700 Ma which suggests a change from early diapirism to later subduction (see Mole et al 2021). The above features need to be better addressed in the manuscript.

Reviewer 3 made very useful and pertinent comments regarding the AGB that help to better focus some sections of our manuscript, to add important lithological information that was lacking and to precise some of our interpretations. In short, these reviewer's comments are:

1- <u>The contrasting pre-2700 Ma and post-2700 Ma magmatic and geodynamic evolution of</u> <u>the AGB as described by Thurston et al. (2008), should be better explained and address</u> <u>in the Ms.</u>

To take account of this issue, we added a better description of the AGB in the revised section 2.2 (see revised page 5, lines 180-188), in order to add information on post-2700 Ma magmatism (essentially characterized by the emplacement of the Preissac-Lacorne batholith) and sedimentation. We now also mention this contrasting pre- vs post-2700 Ma geological evolution on section 6.3 about the significance of the Seismic reflector (see revised lines 1126-1136).

2- The absence of pre-2700 Ma fabric in the AGB.

This is true that there is no occurrences of pre-2700 Ma fabric in most of AGB rocks, but it is consistent with our interpretation; this absence is due to the fact that the AGB exposes the upper crust of an AGB-OPB composition lithosphere and most, if not all, the evidence for early flowage of the lower crust is mostly preserved in OPB gneisses and correlative lower crust rocks in the AGB, such as the Attic Complex, for instance, in which syn-kinematic migmatites yielded ages at ca. 2716 Ma, older than those of the AGB volcanic rocks (see revised lines 980-987).

3- Numerous episodes of sequential folding and faulting in the AGB.

We disagree with that statement. There is only one major event of regional folding and pervasive fabric development in the AGB (i.e. the so-called  $D2_{AGB}$  that we suggest is correlative to the D1- $2_{OPB}$ ), which is already clearly stated in the manuscript. Younger deformation, that may be locally associated to younger fabrics in both the AGB and OPB are spatially restricted to reverse fault zones (late D2) and strike-slip shears (D3).

4- <u>Post-2700 Ma deformation associated with extensive magmatism of more alkalic</u> <u>composition in the AGB.</u>

We agree with the fact that this important magmatic phase has been overlooked in our manuscript. We are now clearly indicating the existence of such magmatism (mostly represented by the Preissac-Lacorne batholith) in lines 157-188 (see comment #1 above) and 1126-1136.

5- <u>The onset of major geodynamic change marked by flysch sedimentation starting during</u> <u>the post-2700 Ma period.</u>

We do not totally agree with that comment, mainly because the occurrence of flysch-like sediments (or turbidites) is not limited to the post-2700 Ma period. It also occurs much earlier, in the Frotet-Evans greenstone belt, for instance, where age constraints for the Broadback Group (resulting from the erosion of underlying volcanic and plutonic rocks, and most likely due to their progressive exhumation) is ca. 2750 Ma or younger, and has likely been a long-lasting process, at least in the OPB. In the revised manuscript, we better described these critical stratigraphic and geochronological constraints in lines 150-154.

#### Specific comments

Lines 162: remove "separated by the Destor-Porcupine fault zone"

This fault is no longer considered to be the line dividing the Abitibi into northern and southern zones based on evidence that Tisdale and Blake River assemblages occur further to the north, the subdivision was moved northwards to the southern margin of the Scapa metasedimentary belt (see Thurston et al., 2008).

We kept the original reference for that fault zone but added that the Northern-Southern limit of the AGB has been revised by Thurston et al. (2008).

Lines 415-425. This is a long and complicated sentence. Break it up into smaller segments to clarify.

We disagree; we consider that this sentence (even though it is long) remains comprehensive.

Lines 890-900. Remove this paragraph. It is inaccurate and not relevant to your topic. The DPM is no longer considered to mark the contact between the north and south zones (see above). *We agree – this paragraph has been removed.* 

Highlights :

- The Abitibi-Opatica contact does not show significant shear deformation;
- There is no metamorphic break at the AGB-OPB transition;
- The Opatica exposes the deepest part of a composite Abitibi-Opatica crust;
- <sup>40</sup>Ar/<sup>39</sup>Ar ages exhumation of AGB and OPB rocks at c. 2678 Ma and c. 2668 Ma;
- Our model shows the importance of vertical transfer of Archean crustal material.

1	The Abitibi-Opatica transition, Superior Province, Quebec, Canada: structural analysis,
2	<sup>40</sup> Ar/ <sup>39</sup> Ar thermochronology and implications for Archean tectonics
3	
4	Yannick Daoudene <sup>a,*</sup> , Alain Tremblay <sup>b</sup> , Gilles Ruffet <sup>c</sup> , François Leclerc <sup>a</sup>
5 6 7 8 9 10 11 12 13	<ul> <li><sup>a</sup> Ministère de l'Énergie et des Ressources naturelles, 5700, 4<sup>e</sup> Avenue Ouest, G1H 6R1, Québec, Québec, Canada.</li> <li><sup>b</sup> Université du Québec à Montréal, Département des Sciences de la Terre et de l'Atmosphère and GEOTOP, 201 avenue du Président Kennedy, H2X 3Y7, Montréal, Québec, Canada.</li> <li><sup>c</sup> CNRS (CNRS/INSU) UMR 6118, Université de Rennes 1, Géosciences Rennes, F-35042 Rennes Cedex, France</li> <li>* Corresponding author.</li> <li>E-mail address: <u>Yannick.Daoudene@mern.gouv.qc.ca</u></li> </ul>
14	ABSTRACT
15	We present a structural and metamorphic study of the Abitibi greenstone belt (AGB)
16	and Opatica Plutonic Belt (OPB) of the Archean Superior Province. The AGB-OPB contact is
17	considered as an archetype example of Archean subduction, based on a LITHOPROBE
18	seismic profile showing a North-dipping lithospheric-scale reflector interpreted as the vestige
19	of subduction. Our mapping indicates that the AGB overlies the OPB, and that the AGB-OPB
20	contact does not show evidence of significant shear deformation, as expected for a major
21	upper plate-lower plate boundary. There is no metamorphic break at the AGB-OPB transition
22	but rather a progressive increase of metamorphic grade toward the OPB. Therefore, we think
23	that the OPB simply exposes the deepest part of a composite AGB-OPB sequence. ${}^{40}\text{Ar}/{}^{39}\text{Ar}$
24	ages suggest that the AGB and OPB rocks were exhumed from amphibolite-facies conditions
25	at ~ 2678 Ma and ~ 2668 Ma, respectively, and then that both terranes share the same
26	succession of thermal disturbances at c. 2650 Ma and 2600 Ma, the latter corresponding to
27	shearing along the Nottaway River Shear Zone. We suggest that progressive cooling of both
28	assemblages was accompanied by strain localisation along strike-slip shear zones and
29	occurred when lateral flow of the lower crust became predominant over the convective
30	overturn and vertical crustal material transfer. Comparison with adjacent areas suggests that
31	regional metamorphism has been coeval over a large region, which is consistent with
32	pervasive deformation and slow cooling as expected for vertical tectonic models and diapiric
33	magmagenesis and ascent during the Archean.
34	Keywords: Abitibi greenstone belt, Opatica plutonic belt, structures, <sup>40</sup> Ar/ <sup>39</sup> Ar
35	geochronology, Archean tectonism.
36	

1

### 37 1. INTRODUCTION

38 Orogenic sutures mark the consumption of oceanic plate lithosphere by subduction 39 and of subsequent arc-continent and/or continent-continent collision, if any. In modern plate 40 tectonics, subduction zones are linear and continuous areas of planetary extent where one 41 plate moves under another and is forced to sink due to high gravitational potential energy into 42 the mantle (e.g. Stern, 2002; Bédard, 2018; Brown et al., 2020). Typically, such areas are 43 characterized by an upper plate-lower plate geometry with contrasting geological 44 characteristics. In Phanerozoic belts, subduction sutures are a typical product of plate 45 tectonics. Although their recognition can be sometimes dubious (e.g. Dewey, 1987), they are 46 usually marked by linear belts of intense deformation, where distinct terranes with different 47 magmatic, metamorphic, structural and geochronological histories have been juxtaposed by 48 plate convergence. In the Archean, the nature of global tectonism remains debated, the main 49 problem concerning similarities and differences with present-day plate tectonics (e.g. Stern, 50 2008; Brown et al., 2020).

51 Archean cratons largely consist of greenstone belts juxtaposed to deeper crustal rocks 52 dominated by plutonic (TTG's) and orthogneissic rocks, the greenstone belts usually show 53 steep folds and shear zones whereas plutonic belts display large-scale dome-like structures 54 and fabric. These rocks usually preserve MT/HT-LP/MP metamorphism that varies from 55 greenschist- to granulite-facies, from the upper to the lower crustal domains. The most widely 56 used arguments in favor of Archean subduction are the occurrence of calk-alkaline igneous 57 rocks, the structural style of orogenic belts, the nature and asymmetric distribution of 58 metamorphism, and seismic reflection data. In seismic reflection, there are few clear examples 59 of relict slabs beneath Archean collisional belts. An example of it is found in the Lithoprobe 60 seismic transect of the Superior Province (Calvert et al., 1995) at the contact between the Abitibi Greenstone Belt and the Opatica Plutonic Belt. In scientific literature, it has been and 61 62 is still interpreted as a fossilized subduction zone and a key argument to support Archean 63 plate tectonics (Calvert et al., 1995; Calvert and Ludden, 1999; Ludden and Hynes, 2000; 64 Wyman et al., 2002; van der Velden and Cook, 2005; Cawood et al, 2006; Percival, 2007; van 65 Hunen and Moyen, 2012; among others).

The Archean Superior Province (Fig. 1) consists of a series of volcano-sedimentary
and plutonic domains, subprovinces and/or terranes that amalgamated during the Late
Archean Kenoran orogeny (e.g. Percival, 2007). The occurrence of unconformable
sedimentary rocks as old as 2.48 Ga (i.e. the Huronian Supergroup) over the volcanic and

70 plutonic rocks indicates that tectonic accretion occurred prior to 2.5 Ga. Although the internal 71 geological characteristics of the domains/subprovinces/terranes are relatively well known, the 72 nature of their boundaries remains poorly constrained and, sometimes, debated. Current views 73 on the tectonic evolution of the Superior Province are that the orogenic collage of 74 Mesoarchean continental fragments and Neoarchean oceanic plates occurred during a 75 complex history of subduction-driven accretionary collision (e.g. Bleeker, 2003; Percival, 76 2007; Percival et al., 2006; 2012, among others). Hence, several authors have interpreted a 77 series of high-strain deformation corridors and structural lineaments of the Superior Province 78 as the trace of potential orogenic sutures (Ludden et al., 1986; Mueller et al., 1996; 79 Daigneault et al., 2002, 2004). The Abitibi-Opatica contact, for instance, is considered as the 80 surface expression of south-verging crustal-scale thrust faults rooting into an inferred North-81 dipping subduction zone, the existence of that subduction being attested by a lithospheric-82 scale seismic reflector extending from c. 40 to c. 70 km-depth.

83 In this contribution, we present a structural and metamorphic study of rocks lying on 84 both sides of the Abitibi-Opatica contact in northern Quebec for a strike length of c. 300 km, 85 between the Lac-au-Goéland area to the west and the town of Chibougamau to the east (Fig. 86 2). The main objectives of fieldwork have been to test (1) how the overall geometry of 87 regional structures fits with south-verging collisional tectonism and thrusting of the OPB over 88 the AGB, and (2) if the structural and metamorphic characteristics of both belts are consistent 89 with an upper-lower plate geometry, as inferred by the subduction model. The determination 90 of the age and duration of structural and metamorphic event(s) that have affected both 91 terranes being critical for a better understanding of their respective geological evolution and 92 mutual relationships, we have performed a series of single-grain  $^{40}$ Ar / $^{39}$ Ar analyses for 93 representative samples of both belts.

94

### 95 2. GEOLOGICAL SETTING

The Superior Province forms the principal Archean basement of North America (Fig.
1a). It is made up of several tectonostratigraphic terranes or subprovinces grouped into four
major tectonostratigraphic types/domains (Card, 1990; Card and Ciesielski, 1986; Hocq,
1994; Stott et al., 2010), (1) metasedimentary rocks, (2) volcanic and plutonic rocks, (3)
tonalite-trondjhemite-granite (TTG)-dominated domains, and (4) high-grade metamorphic
rocks. Most tectonic models suggest that these tectonostratigraphic terranes were
amalgamated, regionally deformed and metamorphosed during multiple episodes of

103 subduction/collision (e.g. Kimura et al., 1993; Daigneault et al., 2002; Percival et al., 2006,

104 2012) between ca. 2.72 Ga and 2.68 Ga (Percival, 2007). The southeastern part of the

105 Superior Province comprises, from north to south, the Opatica, Abitibi and Pontiac

subprovinces, the former two being designated here as the Opatica Plutonic Belt (OPB) and

107 the Abitibi Greenstone Belt (AGB) (Fig. 1b).

108

# 109 2.1 The Opatica Plutonic Belt

110 The OPB is located in the northern part of the southeastern Superior Province. It is 111 separated in two structural domains of plutonic rocks, north and south of the Frotet-Evans 112 greenstone belt that extends east-west over a strike length of approximately 250 km (Figs. 1b 113 and 2). The detailed geology of the OPB is poorly documented, mainly due to difficult road 114 access and a limited mineral resources interest as compared to the adjacent Abitibi 115 subprovince. Existing regional studies indicate that the OPB is a large domain of igneous 116 rocks essentially consisting of foliated tonalites and diorites and equivalent gneisses. The 117 latter were designated as «grey gneisses» by Sawyer and Benn (1993). They are characterized 118 by well-developed compositional layering and locally abundant occurrences of diatexite 119 migmatites in their more fractionated members, mainly in the southern part of the belt (Benn 120 et al., 1992; Sawyer and Benn, 1993; Hocq, 1994; Sawyer, 1998). The gneissic tonalites and 121 diorites are intruded by abundant, slightly- to undeformed plutons of granodiorite and granite. 122 The crystallization age of the OPB intrusive rocks is not precisely constrained but the 123 available geochronological data suggest that they formed mostly between 2.83 and 2.68 Ga 124 (Gariépy and Allègre, 1985; Davis et al., 1992; 1994, 1995; Pedreira-Perez et al., 2020). 125 According to Sawyer (1998), partial melting and migmatization occurred at 2.68-2.67 Ga.

126 In the Lac-au-Goéland area (Fig. 3), the Nomans tonalite, which is surrounded by 127 metavolcanic and metasedimentary rocks of the AGB, possesses petrological and structural 128 characteristics that are identical to those observed in tonalitic rocks of the OPB (Daoudene et 129 al., 2014). It represents an antiformal window of the OPB within the AGB (Goutier, 2005; 130 Daoudene et al., 2014; see below). Previous mapping (i.e. Hocq, 1994), as well as our own 131 observations, shows that the OPB contains km-scale, scattered «lenses» of greenstone-type 132 rocks (Figs. 2 and 3) consisting of amphibolite-grade mafic volcanic rocks and mafic-133 ultramafic intrusive rocks (Figs. 4a and b). The texture of these mafic volcanic rocks varies

134 from primary, with locally preserved breccia and pillow structures (Fig. 4a), to migmatitic.

The Frotet-Evans greenstone belt (Fig. 2) is a large-scale example of metavolcanic 135 136 belts surrounded by plutonic rocks of the OPB. It is divided into four segments, from west to 137 east, the Evans-Ouagama, Storm-Evans, Assinica and Frotet-Troilus segments (cf. Boily, 138 2000). Figure 2 shows the two easternmost segments, the Frotet-Troilus and Assinica 139 segments. The Frotet-Evans greenstone belt has been interpreted as the erosional remnant of 140 an allochthonous thrust sheet by Sawyer and Benn (1993) but it rather represents an 141 autochthonous synformal sequence within the OPB (see below). It preserves evidence for 142 amphibolite-facies metamorphic conditions, except for its central part where greenschist-143 facies metasedimentary rocks are predominant (Simard, 1987; Brisson et al., 1997a and b). 144 Volcanic and sedimentary rocks of the Frotet-Troilus segment are included into the Frotet 145 Group; U-Pb zircon ages from felsic tuff layers and a felsic dyke yielded crystallization ages 146 of  $2780 \pm 28$  Ma and  $2750 \pm 28$  Ma (Thibault, 1985), and  $2782 \pm 6$  Ma (Pilote et al., 1997), 147 respectively. The Assinica segment represents the western extension of the Frotet-Troilus 148 segment (Fig. 2). It comprises a lower unit of mafic to felsic volcanic rocks, the Assinica 149 Group, and an upper unit of sub-concordant, clastic sedimentary rocks, the Broadback Group. 150 There are currently no isotopic data to constrain the age of the Assinica Group but correlation 151 with the Rabbit Formation of the Evans Group in the Storm-Evans segment to the west, 152 suggests that it is probably ca.  $2758 \pm 4$  Ma (U-Pb on zircon; David, 2018 in Groulier et al., 153 2020), which indicates that the overlying clastic sequence of the Broadback Group is ca. 2755 154 Ma or younger.

155

## 156 2.2 The Abitibi Greenstone Belt

157 The AGB is made up of ultramafic, mafic, felsic volcanic rocks and syn-volcanic 158 intrusions of Late Archean age. It consists of several, east-west-trending assemblages of 159 volcanic, volcanoclastic and sedimentary rocks (Fig. 1b) that have been deformed and 160 metamorphosed at greenschist- to amphibolite-facies conditions (e.g. Daigneault et al., 2002). 161 The volcanic domains are separated by narrow and linear belts of clastic sedimentary rocks 162 that are frequently associated with major fault zones (Ludden et al., 1986; Percival, 2007). 163 The youngest sequence of these clastic sedimentary rocks, the Temiskaming Group, was 164 deposited above angular unconformities between ca. 2679 and ca. 2670 Ma (e.g. Robert et al., 165 2005). Based on the age of the various volcanic sequences, the AGB is subdivided into the 166 Northern and Southern Volcanic zones originally separated by the Destor-Porcupine fault 167 zone (Ludden et al., 1986; Chown et al., 1992; Daigneault et al., 2004) but relocated

168 approximately 50 km northward by Thurston et al. (2008; see their Figure 2). In the Southern 169 Volcanic zone, corresponding to the Tisdale (2710-2704 Ma) and Blake River (2704-2695 170 Ma) assemblages of Thurston et al. (2008; Fig. 1b), peak of regional metamorphism has been 171 bracketed at 2680-2660 Ma by Powell et al. (1995), and at 2665-2660 Ma by Robert et al. 172 (2005) and Tremblay et al. (2020). The volcanic and sedimentary rocks, which together 173 represent about 50-60% of the AGB, bound large plutonic domains dominated by TTG-type 174 plutons (Fig. 1b), and both the volcano-sedimentary assemblages and plutonic domains are 175 intruded by syn- and post-tectonic granite, granodiorite and syenite plutons that span ages as 176 young as ca. 2650 Ma (Feng et al., 1992; Daigneault et al., 2002; Robert et al., 2005). In the 177 Val d'Or area (Fig. 1b), the Preissac-Lacorne batholith is a typical example of these syn- to 178 late-orogenic intrusions. It is made up of a composite series of intrusions in which two distinct 179 types and ages of granitic rocks are present: an early, subalkaline to alkaline, monzodiorite-180 monzonite-granodiorite-syenite series emplaced between 2690 and 2670 Ma; and a younger, 181 S-type garnet-muscovite granite series dated between 2660 and 2642 Ma, and probably 182 derived from a major metasedimentary source (Feng et al., 1992; Ducharme et al., 1997; 183 Daigneault et al, 2002).

184 In the study area (Fig. 2), the volcanic rocks of the AGB belong to the Northern 185 Volcanic zone. It mainly consists of two volcanic cycles formed between ca. 2.8 and 2.7 Ga 186 (e.g. Leclerc et al., 2011), and belongs to 2750-2711 Ma volcanic rocks of the Kidd-Monro, 187 Stoughton-Roquemaure, Deloro and Pacaud assemblages with local >2750 Ma occurrences 188 (Fig. 1b; Thurston et al., 2008). In Chibougamau (Fig. 2), the base of the volcanic sequence 189 belongs to the Chrissie Formation, which is currently one of the oldest volcanic units of the 190 AGB at 2791.4 +3.7/-2.8 Ma (U-Pb zircon age; David et al., 2011). It consists of mafic to 191 felsic volcaniclastic rocks, calk-alkaline rhyolites, and tholeiitic-to-andesitic basalts. The Roy 192 Group overlies the Chrissie Formation. It includes two volcanic and sedimentary cycles and is 193 estimated to be up to 13 km-thick (Allard et al., 1985; Dimroth et al., 1984; Daigneault et al., 194 1990). The lower cycle consists of tholeiitic basalts, the Obatogamau Formation, overlain by 195 tholeiitic rhyolites and calk-alkaline volcaniclastic.rocks of the ca. 2732 Ma to ca. 2726 Ma 196 Waconichi Formation (Daigneault and Allard, 1990; Legault, 2003; Leclerc et al., 2011; 197 David et al., 2012). The upper cycle of the Roy Group comprises, from base to top, the 198 Bruneau, Blondeau, Scorpion and Bordeleau formations, which consist of basalt and andesitic 199 basalt at the base overlain by a well-bedded sequence of volcaniclastic and sedimentary rocks 200 (Leclerc et al., 2011). In the Chibougamau area, a lapilli tuff of the Scorpion Formation

yielded U-Pb crystallization age on zircon of 2716,4 ±1,0 Ma (David et al., 2012). In the Lacau-Goéland area (Fig. 3), rocks correlatives to the Bordeleau Formation belong to the Bell
River and Dussieux formations and yielded zircon U-Pb ages of 2725-2717 Ma (Goutier et
al., 2004; David, 2005; David et al., 2006).

205 The Opémisca Group (Fig. 2) is a sequence of clastic sedimentary and volcanic rocks 206 overlying the Roy Group (Charbonneau et al., 1991; Leclerc et al., 2008; Leclerc and Houle, 207 2013). From base to top, it consists of the La Trêve, Daubrée, Stella and Haüy formations 208 (Charbonneau et al., 1991). In the Chibougamau area, the Stella Formation has been 209 interpreted as a lateral equivalent to the La Trêve and Daubrée formations (Charbonneau et 210 al., 1991). The La Trêve Formation consists mainly of polygenic conglomerates with 211 heterogeneous clasts, including TTG granitoids and mafic-to-felsic volcanics, whereas the 212 Daubrée Formation is made up of turbidites. The Haüy Formation corresponds to a volcanic 213 sequence mainly composed of K-rich porphyric andesites with lenses of conglomerate. The 214 Haüy lava flows and associated volcaniclastic sediments have been interpreted as the result of 215 evolved arc volcanism and backarc basin sedimentation (Mueller, 1991). In the Stella and 216 Haüy formations, detrital zircons provided maximum ages of deposition of  $2704 \pm 2$  Ma 217 (Leclerc et al., 2012) and 2691.7  $\pm$  2.9 Ma (David et al., 2007), respectively. In the Lac-au-218 Goéland area (Fig. 3), the Daubrée Formation is intruded by the Inconnue River Diorite, 219 which yielded a U-Pb zircon age of  $2693.6 \pm 2.9$  Ma (McNicoll and Goutier, 2008).

220

## 221 **3. REGIONAL STRUCTURES**

222 According to Sawyer and Benn (1993), the OPB records three generations of 223 structures. The oldest is related to D<sub>1</sub> and a penetrative, moderately dipping foliation (S<sub>1</sub>) 224 hosting broadly distributed, WSW-ENE-trending mineral and stretching lineations (Fig. 4b). 225 S<sub>1</sub> corresponds to the regional gneissic foliation and defines a metamorphic layering that is 226 axial-planar to intrafolial F1 folds. D1 has been attributed to a top-to-the SW verging 227 deformation. Sawyer and Benn (1993) argued that S<sub>1</sub> was folded during a D<sub>2</sub> deformational 228 event, marked by East-West trending, steeply North- or South-dipping compressive shear 229 zones. S<sub>2</sub> is a steeply dipping biotite foliation recording amphibolite-greenschist facies 230 metamorphic conditions, although it may locally attain the amphibolite facies. It frequently 231 forms a composite  $S_{1-2}$  fabric with  $D_1$  structures.  $D_1$  and  $D_2$  has been attributed to the 232 progressive development of an overall SSE-verging fold-and-thrust belt (and inferred 233 backthrusting) along the AGB-OPB boundary and southward (Benn et al., 1992; Sawyer and Benn, 1993; Lacroix and Sawyer, 1995). Sawyer and Benn (1993) also recognized a D<sub>3</sub> event,
related to NE-SW and NW-SE-trending strike-slip shear zones cutting across both the FrotetEvans and the Abitibi greenstone belts (see Fig. 2).

237 In the AGB of the study area, existing structural studies of regional extent have 238 documented four deformational phases (e.g. Dimroth et al., 1986, Daigneault et al., 1990; 239 Leclerc et al., 2012; Mathieu et al., 2020), the youngest one (D<sub>4</sub>) being related to the 240 Mesoproterozoic Grenvillian orogenic cycle (Daigneault et al., 1990). D<sub>1</sub> has been attributed 241 to an early-formed series of drape folds devoid of axial-planar schistosity (Daigneault et al., 242 1990). D<sub>2</sub> is related to syn-metamorphic folding and the development of the regional EWtrending schistosity ( $S_2$ ), it corresponds to  $F_{2a}$  and  $F_{2b}$  folds as described by Leclerc et al. 243 244 (2012).  $D_1$  and  $D_2$  of the AGB correlate with the Benn's et al. (1993)  $D_{1-2}$  phase of the OPB. 245 According to Mathieu et al. (2020), the  $D_3$  phase corresponds to the waning stage of the main 246 deformation event that reactivated E- W- striking D<sub>2</sub> reverse faults as transcurrent 247 strike- slip faults. In contrast, Leclerc et al. (2012) described D<sub>3</sub> structures as NE-striking 248 shear zones cutting across earlier E-W trending D<sub>2</sub> shears/faults (Fig. 2).

Our comprehension and interpretation of structures in the study area are similar to those described above. We think that  $D_{1-2}$  and  $D_3$  belong to progressive and long-lasting deformational events. In the following sections, we describe fabric and structures related to those  $D_{1-2}$  to  $D_3$  events in both the AGB and the OPB. These descriptions are based on our own field observations and from previous studies.

254

# 255 **3.1 D1-2 phase of deformation**

256 D<sub>2</sub> deformation structures are regionally-developed and associated with folding of the 257 volcanic and sedimentary rocks of the AGB (Figs. 2 and 3) whereas the D<sub>1</sub> metamorphic 258 layering is restricted to the lowermost stratigraphic part of that belt. Regional folds (F<sub>2</sub>) have 259 variable wavelength and their axial traces strike East-West, although they are locally 260 reoriented due to younger D<sub>3</sub> strike-slip shears, such as the Nottaway shear zone (Fig. 2). In 261 the Lac-au-Goéland area (Fig. 3), for instance, to the southwest of the Nottaway shear zone, 262 S<sub>2</sub> trajectories within the AGB are NW-SE trending and define a F<sub>3</sub> syncline, the Maicasagi 263 syncline (Fig. 3; Goutier, 2005). Overall, the regional pattern of folding of the AGB is 264 characterized by a north-south succession of F<sub>2</sub> synclines and anticlines. The core of synclines 265 is usually occupied by the Opémisca Group, whereas the anticline crests expose the

- underlying volcanic rocks of the Roy Group which are, locally, crosscut by large pre- to syntectonic intrusions such as the Opémisca and Chibougamau plutons (Fig. 2).
- 268 The volcanic and sedimentary rocks of the AGB are characterized by a weakly to 269 moderately developed regional schistosity  $(S_2)$ , although some rock exposures seem to be 270 devoid of structural fabric development (Fig. 4d). When present, S<sub>2</sub> is pervasive, strikes East-271 West and dips steeply either to the north or the south (Figs 5 and 6a). Towards the contact 272 with the OPB, the volcanic rocks of the Roy Group are more intensely foliated, the dominant 273 foliation  $(S_1)$  is there tightly folded by  $F_2$  (Fig. 4e), although that, most of the time, it forms a 274 composite S<sub>1-2</sub> fabric and consists of a metamorphic layering marked by metamorphic 275 minerals such as biotite and hornblende (see section 4), that corresponds northward to the flat-276 lying S<sub>1</sub> gneissic fabric of the OPB gneisses. Our interpretation is that the OPB gneissic 277 foliation is verticalized and transposed into  $S_2$ , it forms there a  $S_{1-2}$  metamorphic layering that 278 is subparallel to the contact between the OPB and the AGB (Fig. 65) and is absent farther 279 south in the up-stratigraphy, lower-grade rocks of the AGB. The bedding plane, when visible, 280 is usually parallel to the regional  $S_2$  that is marked by the alignment of amphibole and biotite 281 in amphibolite-grade rocks (as for  $S_1$ ) and by chlorite, epidote and albite in greenschist and 282 sub-greenschist facies rocks of the AGB.
- 283 East-West-trending shear zones cut the volcanic and sedimentary sequences of the 284 AGB. These sub-vertical shear zones are poorly exposed. Previous studies (e.g. Dimroth et 285 al., 1986; Daigneault et al., 1990; Leclerc et al., 2012) indicate they are mainly located on the 286 limbs of regional-scale F<sub>2</sub> folds (Fig. 2), some of the latter being synformal troughs occupied 287 by lower grade sedimentary rocks of the Opémisca Group (Fig. 4d). As for F<sub>2</sub> folds, these 288 shear zones are deflected in the vicinity of younger NE- and NW-trending strike-slip shear 289 zones (Figs. 2 and 3). S<sub>2</sub> increases in intensity toward and within these D<sub>2</sub> shear zones and 290 hosts steeply plunging stretching and/or mineral lineation, suggesting significant subvertical elongation (Dimroth et al., 1986; Daigneault et al., 1990). Shear-sense indicators are rare but 291 292 when present, they suggest that the shear zones correspond to north or south side-up reverse 293 faults.
- In the OPB, dominant deformation fabric are attributed to D<sub>1</sub> and expressed by a
  weakly- to well-developed metamorphic foliation (S<sub>1</sub>) showing variable orientations (Fig. 5).
  S<sub>1</sub> trajectories in the OPB define an East-West trending, elongated dome-and-basin geometry,
  consistent with the overall trend of D<sub>2</sub> regional folds and structures in the adjacent AGB (Fig.
  Mineral lineations, defined by the alignment of quartz-feldspar aggregates, are more

- discrete and of variable orientation as compared to the AGB (Fig. 6b). These lineations
- 300 plunge at low angle within the regional foliation. In the area located between the Frotet-Evans
- 301 and Abitibi greenstone belts (Fig. 2), there are abundant decametre-scale of foliated,
- 302 amphibolite-grade lenses of basaltic metavolcanic rocks within the OPB gneisses. The
- 303 foliation in these metavolcanic rocks trends parallel to the S<sub>1</sub> metamorphic layering in the
- 304 surrounding gneisses.

305 The AGB-OPB contact is not clearly exposed in the area of Figure 2. Rock exposures 306 are sparse along that contact which is locally masked by syn- and post-orogenic intrusions, 307 such as the Barlow pluton in the Chibougamau area (Fig. 2) and the Canet and Maicasagi 308 plutons (Sawyer and Benn, 1993; Goutier, 2005; Daoudene et al., 2014) in the Lac-aux-309 Goéland area (Fig. 3). In the latter area, the Nomans tonalite, located to the south of the 310 inferred AGB-OPB contact, shows the same structural style as the gneissic rocks of the OPB. 311  $S_1$  trajectories within the tonalite (Fig. 5) indicate that it is a  $D_2$ -related structural dome that is 312 attached to the OPB at depth (Goutier, 2005; Daoudene et al., 2014). The occurrence of 313 foliation-parallel, amphibolite lenses along the AGB-OPB contact has been considered as 314 evidence for the structural imbrication of both terranes and the allochthonous nature of the 315 AGB (Sawyer and Benn, 1993). We do not think that this boundary corresponds to a major 316 displacement zone (see section 4 below), although  $S_1$  appears there to be more intensely 317 developed in both the metavolcanic rocks of the AGB and the OPB tonalites. S<sub>1</sub> trajectories 318 are parallel to the AGB-OPB contact (Fig. 5), suggesting that it formed, or was transposed, 319 during D<sub>1</sub>. There are relatively few lineation measurements in the study area (Fig. 6). Sawyer 320 and Benn (1993) reported SE-plunging mineral lineations in the hanging wall of the Lac 321 Matagami shear zone, a late (D<sub>3</sub>) sinistral fault located west of the Lac-au-Goéland area, as 322 well as down-dip lineations along the Barlow fault in the Chibougamau area. In this latter 323 area, stretching lineations plunge moderately to the south, but locally show significant 324 variations of orientation due to the proximity of syn- and post-tectonic intrusions (Bedeaux et 325 al. 2021). 326 The Abitibi and Frotet-Evans greenstone belts share a similar internal geometry and

- 327 relationships with the underlying plutonic rocks of the OPB. However, the Frotet-Evans
- 328 greenstone belt is of higher metamorphic grade and structurally more complex than the AGB
- 329 (see Gosselin, 1996). It occupies the core of a large-scale series of east-west-trending  $F_2$
- 330 synclines (Figs. 2 and 5) within the OPB gneisses, some of them being marked, as for the
AGB, by steeply dipping reverse faults on their limbs (Simard, 1987; Gosselin, 1996; Brisson
et al., 1997a and 1997b).

In the study area, we did not observe unambiguous shear-sense criteria within the AGB-OPB contact zone. Along the AGB-OPB contact, there are numerous tonalitic to granitic dykes crosscutting basaltic amphibolites of the AGB (Fig. 4f). Most of these dykes are in irregular and diffuse contacts with the host rocks and are slightly to moderately foliated, suggesting high temperature syntectonic emplacement.

338

## **339 3.2 D**<sub>3</sub> structures

340 The  $D_3$  phase of regional deformation is more localized as compared to  $D_{1-2}$ . It is 341 essentially associated with strike-slip displacement along NE- and NW-trending shear zones. 342 The Nottaway River shear zone (Figs 2 and 3) is the largest of the NW-SE-trending  $D_3$  shear 343 zones in the study area. It cuts across the D<sub>1-2</sub> structures of both the AGB and OPB. It delimits 344 the western boundary of the Frotet-Evans greenstone belt, which is also cut by another major 345 NW-trending shear in its middle part, the Lucky Strike River shear zone (Fig. 2). A mylonitic 346 S<sub>3</sub> schistosity is pervasively developed along these shear zones (Fig. 7a). F<sub>3</sub> dextral folds, with 347 S<sub>3</sub> as an axial-planar foliation, can be observed in the vicinity of D<sub>3</sub> shear zones (Fig. 7b). S<sub>3</sub> 348 hosts a well-developed, sub-horizontal stretching lineation (Fig. 7c), indicating that D<sub>3</sub> is 349 dominated by strike-slip motions. Abundant macroscopic and microscopic shear-sense 350 indicators, such as shear bands, S-C fabric (Fig. 7a and 7d), mica fish (Fig. 7e), drag folds and 351 asymmetrical boudinage (Fig. 7f) are widely present and attest for dextral displacement along 352 the NW-SE-trending shear zones. NE-trending shears, such as the Lac Lamark shear zone 353 (Fig. 2), are characterized by the same type of ductile fabric but are sinistral, as indicated by 354 regional mapping and the offset of marker horizons in the Chibougamau area (e.g. Chown et 355 al., 1992; Daigneault et al., 2002). The two sets of strike-slip shear zones, NW- and NE-356 trending, form a conjugate system.

357

### 358 **4. METAMORPHISM**

In this section, we describe the metamorphic characteristics of rocks in the study area, from top to base of the AGB stratigraphy, and from the lower-grade rocks of the AGB towards the high-grade metamorphic units of the OPB.

The sedimentary rocks of the Opémisca Group, which occupy the core of regional F2
 synclines, do not show macroscopic evidence of metamorphic recrystallization. Primary

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364 sedimentary structures, such as bedding (Fig. 4d), and pillow structures in volcanic rocks of 365 the underlying Roy Group are frequently well-preserved (Fig. 8a). Low-grade metamorphic 366 minerals, such as chlorite, sericite, epidote, and actinolite-tremolite progressively appear 367 down into the stratigraphic pile and can be relatively abundant at the margins of synclines, 368 indicating that the lowermost part of the Opémisca Group attained greenschist-facies. The 369 metamorphic grade of these sedimentary rocks even reaches the amphibolite facies towards 370 the AGB-OPB contact, specifically along the northern boundary of the Waconichi Lake 371 syncline (Figs 2) where the Opémisca metapelites contain biotite, garnet, and staurolite (Fig. 372 8b). Staurolite occurs as mcm-long, oblong-shaped and corroded crystals defining the S<sub>2</sub> 373 foliation. Lithologies of the Opémisca Group also locally contain late- to post-kinematic 374 metamorphic minerals such as, for instance, randomly oriented chloritoïds (Fig. 8c) 375 suggesting late-stage growth.

376 The mafic rocks of the Roy Group were dominantly metamorphosed to the 377 greenschist-facies (Allard et al., 1985). These rocks are pervasively chloritized and contain 378 typical low-grade metamorphic minerals such as albite, epidote and actinolite-tremolite 379 defining the S<sub>2</sub> schistosity. In the vicinity to the contact with the OPB, they contain, however, 380 metamorphic minerals of higher temperature characteristic of the amphibolite facies. 381 Actinolite-tremolite is there replaced by hornblende and biotite. Garnet is present and usually 382 forms irregular patches that are preferentially distributed along the pillow margins of mafic 383 volcanic flows (Fig. 8d). Garnet is frequently retrogressed into a mixture of epidote, 384 amphibole and chlorite. Some basaltic amphibolites contain epidote porphyroblasts 385 overprinting the composite  $S_{1-2}$  foliation as well as randomly oriented crystals of actinolite, 386 suggesting retrogressive conditions after peak metamorphism.

387 Metamorphic conditions recorded by the OPB are more difficult to evaluate precisely, 388 mainly because these rocks essentially consist of intermediate and felsic plutonic rocks. 389 However, most of the mafic rocks of the Opatica contain hornblende+andesine assemblages, 390 which, with the occurrence of rare garnet- or clinopyroxene, imply amphibolite-facies 391 conditions (Sawyer, 1998). The intermediate and felsic gneissic rocks are commonly 392 plastically deformed and show coarse-grain recrystallization textures. Feldspar and quartz 393 crystals display evidence for syn-kinematic growth and recrystallization. Grain boundaries are 394 cuspate and undulated (Fig. 8e), which is indicative of extensive high-temperature grain 395 boundary migration during deformation (Gower and Simpson, 1992). Myrmekitic textures 396 preserved in some coarse-grained K-feldspars also indicate metamorphic temperatures in

397 excess of 500-550°C (Simpson, 1985; Simpson and Wintsch, 1989). All these observations 398 indicate that the regional  $D_1$  deformation structures in the OPB gneisses were acquired at 399 amphibolite-facies conditions of metamorphism, or higher. Some felsic rocks display quartz 400 grains with a chessboard texture, which is typical of intra-crystalline slip textures developed 401 at subsolidus temperatures (i.e. above 600°C; Blumenfeld et al., 1986; Gapais and Barbarin, 402 1986; Mainprice et al., 1986; Schmid and Casey, 1986). Evidence for anatectic melting is 403 abundant in the deepest crustal domains of the OPB and marked by the occurrence of 404 migmatites (Fig. 8f). Anatectic melt occurs in D<sub>2</sub> shear zones located near to the center of the 405 OPB (Sawyer and Benn, 1993). However, according to Sawyer (1998), melting is late- to 406 post- $D_1$ , as suggested by  $S_1$  disappearing within neosome patches and paleosome material 407 showing remnants of the  $S_1$  foliation.

408 There are very few pressure-temperature (P-T) studies for metamorphic rocks of the 409 AGB. At the scale of the belt, regional metamorphism ranges in grade from subgreenschist-410 facies to the greenschist-amphibolite facies transition. In the Blake River and Tisdale 411 assemblages (Fig. 1b), low-pressure regional metamorphism is indicated both by the 412 occurrence of the actinolite-oligoclase zone, and the persistence of pre-regional 413 metamorphism and alusite. According to Powell et al. (1995), pressure of regional 414 metamorphism and of late to post-orogenic contact metamorphism were similar (360-400 415 MPa), indicating that low-pressure conditions persisted throughout the tectonic evolution of 416 the AGB. Regionally, isograds dip shallowly to the north and trend subparallel to lithological 417 and structural trends (Powell et al., 1995; Faure, 2015). In the Chibougamau area (Fig. 2), 418 Bedeaux et al. (2020) suggested P-T conditions of 550±50°C and 6±1.2 kbar for basaltic 419 amphibolites of the AGB cropping out in the vicinity of the contact with the OPB. In the OPB, Sawyer (1998) determined a peak metamorphism temperature of 760°C for the 420 421 migmatite facies rocks, and pressures of 5.2-6.2 and 6.3-7.1 kbar for diatexite migmatites and 422 adjacent amphibolites, respectively. Pressure measured in the Canet pluton of the Lac-au-423 Goéland area (Fig. 3) were significantly lower, at 1.7-2.6 kbar (Sawyer, 1998), suggesting 424 that the emplacement of this pluton occurred during exhumation of the OPB gneisses.

Figure 9 presents a series of north-south trending, interpretative structural profiles between the Lac-au-Goéland (profile A-A') and the Chibougamau areas (profile D-D'). These profiles show that the OPB exposes deeper crustal rocks and higher-grade metamorphic rocks as compared to the adjacent and overlying volcanic rocks of both the Frotet-Evans and Abitibi greenstone belts, and that both the OPB and AGB originally constituted an almost continuous

430 middle-lower (the OPB) to upper crust sequence (the AGB). This contrasts with Sawyer and 431 Benn's (1993) interpretation showing the Frotet-Evans greenstone belt as a far-travelled, 432 south-directed allochthonous nappe thrusted over the OPB gneisses, and both units forming a 433 basement-cored nappe thrusted southward over the AGB (see their figures 6 and 8). We think 434 that their structural interpretation is unlikely for the following reasons, (i) in typical crustal-435 scale thrust systems (Boyer and Elliott, 2002; Butler and Bond, 2020), older rock units and/or 436 higher grade (deeper) metamorphic rocks are transported over younger and/or lower grade 437 (shallower) metamorphic rocks, which is obviously not the case here, and (ii) it would imply 438 that the Frotet-Evans greenstone belt represents a klippen that roots 50 km, or more to the 439 north, in the Colomb-Chabouillé Group of the Nemiscau subprovince along its southern the 440 boundary with the OPB (see Fig. 2); there are, however, no metamorphic break and major 441 suture between the Nemiscau and Opatica subprovinces, and regional mapping and structural 442 analyses rather suggest the relative exhumation of the OPB as compared to the Nemiscau 443 (Pedreira-Perez et al., 2020), which does not fit the Sawyer and Benn's (1993) thrusting 444 model.

445

### 446 **5.** <sup>40</sup>Ar/<sup>39</sup>Ar GEOCHRONOLOGY

In the following section, we present thirty-one <sup>40</sup>Ar/<sup>39</sup>Ar laser probe experiments on
single grains of amphibole, biotite and muscovite from eighteen rock samples from both the
AGB and OPB.

450

### 451 **5.1 Analytical method and procedures**

Single grains of amphibole, biotite and muscovite were handpicked under a binocular microscope from 250 $\mu$ m to 1000 $\mu$ m mineral fractions. Separated minerals were irradiated with Cd-shielding, in the 5C high neutron flux facility of the McMaster Nuclear Reactor (Hamilton, Ontario, Canada). This process lasted 298 h with a global efficiency (J/h) of 5.83x10<sup>-5</sup> h<sup>-1</sup>. The irradiation standard was amphibole Hb3gr (Turner et al. 1971; Roddick 1983, Jourdan et al. 2006; Jourdan and Renne 2007); 1081.0 ± 1.2 Ma according to Renne et al. (2010, 2011).

459 Single grains of amphibole, biotite and/or white mica were analyzed by the <sup>40</sup>Ar/<sup>39</sup>Ar
460 method in step-heating using a CO<sub>2</sub> laser probe coupled with a MAP 215 mass spectrometer.
461 The analytical procedure is described by Ruffet et al. (1991, 1995, 1997). The five argon
462 isotopes and the background baselines were measured in eleven cycles, in peak-jumping

463 mode. Blanks were performed routinely each first or third/fourth run and subtracted from the 464 subsequent sample gas fractions. All isotopic measurements account for K and Ca isotopic 465 interferences, mass discrimination and atmospheric argon contamination. Apparent age errors 466 are plotted at the 1 $\sigma$  level and do not include the errors on the <sup>40</sup>Ar\*/<sup>39</sup>Ar<sub>K</sub> ratio, age of the 467 monitor and decay constant.

468 A plateau is obtained when calculated  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{K}$  ratios of at least three consecutive steps, containing a minimum of 70% of the <sup>39</sup>Ar released, agree with the weighted mean 469 calculated  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{\text{K}}$  ratio of the plateau segment. This 70% threshold is a requirement 470 criterion that we use in all our geochronological studies, and it is much more restrictive than 471 472 the 50% threshold set subjectively (in their own words) used by Fleck et al. (1977) and still adopted by most users of the <sup>40</sup>Ar/<sup>39</sup>Ar method (e.g. Schaen et al., 2021). Since the amount of 473 474  $^{39}$ Ar<sub>K</sub> released is not a disqualifying criterion, we use the concept of pseudo-plateau (Cheilletz et al., 1999), which meets the same statistical criteria except for the cut-off threshold (<70% 475 of  ${}^{39}Ar_{\rm K}$ ), and which is much more efficient for the analysis of complex mineral phases. In 476 477 our detailed analysis of spectra, we also use a plateau-like descriptive terminology which 478 refers to a weighted average age calculated over segments of apparent ages, not necessarily 479 joined (e.g. edges of a saddle-shaped segment), but which are considered to be geologically significant. The errors on the  ${}^{40}\text{Ar}*/{}^{39}\text{Ar}_{\text{K}}$  ratio and age of the monitor and decay constant are 480 481 included in the final calculation of the error margins on the pseudo- and plateau ages or for apparent ages individually cited.  ${}^{40}$ Ar/ ${}^{39}$ Ar ages are provided with 1 $\sigma$  errors. Analytical data, 482 483 parameters used for calculations (isotopic ratios measured on K and Ca pure salts; mass 484 discrimination; atmospheric argon ratios; J parameter; decay constants, etc.) and reference 485 sources are available in the complementary data repository.

486

### 487 **5.2 Isotopic closure temperatures**

488 A radiochronological system is the combination of the crystal lattice of a mineral 489 phase and an isotopic pair consisting of a radiogenic element and its radioactive precursor, e.g. <sup>40</sup>Ar\* and <sup>40</sup>K, the latter (K) being or not a constituent of the lattice. The analysis of 490 491 isotopic systems, particularly those involving noble gases, has long focused on the behaviour of the radiogenic element, in this case <sup>40</sup>Ar\*, the host minerals being themselves considered 492 493 inert and the fate of the isotopic system being exclusively governed by the diffusion of the gas 494 under temperature control, as described by Fick's laws. This led to the notion of isotopic 495 blocking temperature initially introduced by Jäger et al. (1967), formalized as the isotopic

closure temperature (T<sub>c</sub>) by Dodson (1973) and then nuanced with the notion of isotopic
closure window by Dunlap (1997). Therefore, during cooling and for a given isotopic system
(mineral and isotopic pair), the isotopic chronometer initializes when diffusion becomes
negligible because the temperature has dropped below its isotopic closure temperature.

500 T<sub>c</sub> for amphibole is difficult to estimate. In addition to the effect of cooling rate and 501 size of diffusion domains, amphibole diffusivity is influenced by ionic porosity (Fortier and 502 Giletti, 1989; Dahl, 1996). On the basis of natural hornblende compositions, Dahl (1996) 503 suggested a T<sub>c</sub> range of 480-550°C, which has been re-evaluated at 550-560°C by Villa 504 (1998) on the basis of Kamber's et al. (1995) experiments suggesting that, depending on the 505 crystal lattice characteristics, hornblende may form a closed system to Ar diffusion at 506 temperature as high as 580°C and a cooling rate of 0.7°K/Ma (Villa et al., 1996). In this study, 507 we will therefore consider a median  $T_c$  range value of 550-600°C for amphiboles.

508  $T_c$  for biotite ranges from 250°C to 400°C (Harrison et al., 1985; Montigny, 1985; 509 McDougall et Harrison, 1988; Spear, 1993). In the absence of deformation and fluid-related 510 recrystallization, it may be at least as high as ~450°C (e.g. Villa and Puxeddu, 1994). With 511 diffusion domains in the range 100-1000 µm and a cooling rate lower than 100°C/Ma,  $T_c$  for 512 biotite is ca. 400°C (e.g. Grove and Harrison, 1996). We will here consider a  $T_c$ -biotite range 513 value of 350-450°C.

514 Purdy and Jäger (1976) originally estimated T<sub>c</sub> of white micas at 350°C, by calibration 515 of isotopic ages versus metamorphic grade. This estimate was erroneous and underestimated 516 (Villa, 1998). Subsequent studies have shown that in the absence of deformation and fluid-517 enhanced recrystallisation, T<sub>c</sub> for muscovite may reach 500°C (Villa, 1998; Bosse et al., 518 2005), and even temperatures higher than 550°C for phengite under blueschist facies 519 metamorphic conditions (Lister and Baldwin, 1996). Using experimental muscovite diffusion 520 coefficients yielded by Harrison et al. (2009), Pitra et al. (2010) calculated a 390-590°C 521 isotopic closure temperature range for white micas, using various effective diffusion radii 522 (100 to 1000 µm) and cooling rates (1 to 1000°C/Ma). Considering diffusion domain sizes of 523 100-1000  $\mu$ m and cooling rates between 1°C and 10°C/Ma, a more reasonable and realistic 524 assessment for the cooling of Archean rocks (e.g. Chardon et al., 2009), isotopic closure of 525 white micas would occur in the temperature range 400-500°C.

The use of above T<sub>c</sub> range estimates should only concern crystals unaffected by
deformation and/or recrystallization processes, (e.g.Villa, 1998). In contexts where these
phenomena occur, the other component of the radiochronological system, the crystal lattice of

529 the mineral carrying the istopic system, must also be considered because it may no longer 530 remain inert. Deformation can damage crystal lattices, enhances Ar diffusion, and promotes 531 the formation of intra-crystalline domains with lower diffusion length-scales and lower T<sub>c</sub>. In 532 the presence of fluids, minerals (or intracrystalline sub-domains) may recrystallize and neo-533 grains may possibly grow, two configurations which, if they occur at  $T \leq T_c$  and after initial 534 isotopic closure, will induce the recording of younger ages. Consequently, distinct intra-535 crystalline domains with different chemical and isotopic compositions, and thus distinct ages, 536 may coexist within the same mineral/grain.

537 Cheilletz et al. (1999), Alexandrov et al. (2002) or Tartese et al. (2011) identified and 538 detailed a process of sequential fluid-assisted resets of K-Ar isotopic chronometer of white 539 mica from leucogranites. Tremblay et al. (2020) showed that K-Ar isotopic system of white 540 micas from auriferous quartz veins and hosting shear zones of the Bourlamaque pluton (Val 541 d'Or mining district, Abitibi) was driven by sequential dynamic recrystallization 542 (dissolution/recrystallization/neocrystallization/aggregation) linked to successive periods of 543 fluid circulation and deformation which occurred below T<sub>c</sub>. Dunlap (1997) suggested that, 544 during ductile deformation, K-Ar isotopic system of white micas from mylonites could also 545 be partly driven by neocrystallization processes below T<sub>c</sub>, at 250-350°C. <sup>40</sup>Ar/<sup>39</sup>Ar amphibole 546 analyses (e.g. Tremblay et al., 2020) suggest that they also behave in the same way. 547

#### 548 **5.3 Data processing**

549 Metamorphic minerals such as amphiboles and micas extracted from rocks deformed 550 or disturbed by fluids are usually texturally complex (recrystallized, composite grains, etc). 551 The <sup>40</sup>Ar/<sup>39</sup>Ar analysis of such crystals is very useful, as information it provides integrates the 552 effects of the disturbing processes and therefore the timing of their occurrence, even if this 553 occurs below Tc (e.g. Cheilletz et al. 1999; Hames et al., 2008; Castonguay et al., 2007; Tremblay et al. 2020). Hence, due to the complexity of <sup>40</sup>Ar/<sup>39</sup>Ar results that arise from the 554 555 analysis of such materials, which are common in old rock series such as the Archean, their 556 processing must go beyond the simple visual examination of age spectra.

557 De Putter et al. (2015), De Putter and Ruffet (2020) and Tremblay et al. (2020) 558 presented and developed tools for the visualization and processing of complex <sup>40</sup>Ar/<sup>39</sup>Ar 559 systems. Three of these tools are classical: the plateau (Fleck et al., 1977) and pseudoplateau 560 age (PPA) (Cheilletz et al., 1999) concepts, and the probability density diagrams, which have 561 been first used for <sup>39</sup>Ar-<sup>40</sup>Ar data by Deino & Potts (1992). Other tools are more atypical; (i) the degassing and the weighted age spectra, which were designed to visualize Ar degassing
 kinetics of <sup>40</sup>Ar/<sup>39</sup>Ar experiments, and (ii) the resizing of age spectra which was developed to
 ensure their statistical representativeness.

565 Amphibole <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> spectra transcribe the evolution of the Ca/K ratio (<sup>37</sup>Ar<sub>Ca</sub> # Ca 566 and <sup>39</sup>Ar<sub>K</sub> # K with <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> = CaO/K<sub>2</sub>O / 2.179; Deckart et al., 1997) during step-heating 567 (degassing) experiments. Hence, a flat <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> segment (homogeneous Ca/K ratio) 568 expresses the degassing of an amphibole component devoid of compositional zoning and/or 569 inclusion of another mineral species.

570

### 571 5.4 Sampling strategy

To document and compare the tectonothermal history of the AGB and OPB on both sides of the contact zone, we performed a series of step-heating <sup>40</sup>Ar/<sup>39</sup>Ar analyses of micas and amphiboles collected in metamorphic and plutonic units of both terranes. Sixteen samples are from the Lac-au-Goéland area, and two from the AGB-OPB contact in the Chapais-Chibougamau area. Sample locations are shown in figures 3 and 4 whereas their geographic coordinates are in Table 1.

578 Four amphibolite samples (YD0009A, YD0048A, YD0039A, and YD0100A) and a 579 garnet-staurolite micaschist sample (JF4065A) are from the AGB. These rocks show a well-580 developed S<sub>2</sub> foliation marked by the preferred orientation of amphibole and/or biotite. In thin 581 sections, hornblendes from sample YD0009A and tremolites/actinolites from sample 582 YD0039A are homogeneous and unaltered, whereas dark green hornblendes YD0048A and 583 YD0100A show light-green rims or patches suggesting intracrystalline zonation during crystal 584 growth or recristallization. Biotites from samples YD0009A and YD0100A are green-colored 585 euhedral crystals whereas those from sample JF4065A (a metasedimentary rock) are dark-586 brown-colored euhedral. There is no optically discernible occurrence of secondary chlorite in 587 these samples. Sample YD0037, from which were extracted amphibole and biotite grains, is a 588 slightly deformed diorite that shows petrographic characteristics similar to the Rivière 589 Inconnue Diorite (Fig. 3), which yielded a U-Pb zircon age of  $2693.6 \pm 0.6$  Ma (McNicolls 590 and Goutier, 2008).

YD0018B, YD0020A, YD0063A, YD0035A, JG1245A, and YD0101A are tonalite
samples from the OPB. These tonalites are mainly composed of quartz and plagioclase and
contain various amounts of amphibole and/or biotite, which mark the regional foliation (S<sub>1</sub>).
Amphiboles YD0018B, YD0020A, YD0063A and YD0101A are sub-automorphous. All

samples contain green- to brown-colored biotites that are commonly euhedral and only rarelychloritized.

597 Four samples of quartz-feldspar porphyry dykes were selected in both the AGB 598 (YD0023A, YD0048B, and YD0009B) and the OPB (YD0023A). These dykes contain 599 randomly oriented, coarse-grained muscovite crystals lacking microscopic evidence of 600 recrystallization and/or deformation. A biotite grain from sample YD0009B was also 601 analysed.

Two mylonite rock samples, YD0055A and YD0059A, from the Nottaway River shear zone were analysed. Sample YD0055A is a felsic rock belonging to the AGB. It shows a welldeveloped mylonitic foliation and contains abundant asymmetric muscovite crystals showing mica fish structures indicative of dextral shearing. These muscovite crystals also show secondary rims of neoformed micas typical of recrystallization. Sample YD0059A is a foliated tonalite from the OPB, it contains coarse-grained biotites that appear slightly altered and corroded. Most of these biotite grains also show fine rims of recrystallization.

Finally, we analyzed one sample of syenite of the Montviel alkaline intrusion (Goutier, 2005), which is associated to a carbonatite facies that has been dated at  $1894 \pm 4$  Ma (U-Pb on zircons; David et al., 2006).

612

#### 613 **5.5 Results.**

The step-heating  ${}^{40}$ Ar/ ${}^{39}$ Ar results are presented on Figures 10 to 14. The distinct age spectra provided by duplicate analyses of amphibole and mica presented in the following sections confirm, as already stressed by Tremblay et al. (2020), that the interpretation of Archean material using the  ${}^{40}$ Ar/ ${}^{39}$ Ar method is challenging and requires a detailed analysis of results.

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### 620 5.5.1 The Abitibi Greenstone Belt.

The  ${}^{40}$ Ar/ ${}^{39}$ Ar results for the AGB amphiboles and biotites are show in Figure 10. Analysis of amphibole 1 from sample YD0037A provides a rather flat age spectrum, with nevertheless a deflection of apparent ages in the intermediate temperature steps (Fig. 10a). The apparent ages, which are concordant on both sides of that deflection, allow the calculation of a plateau-like age at 2680.1 ± 3.6 Ma (56.3% of the total  ${}^{39}$ Ar<sub>K</sub> released), whereas the base of the deflection yields 2637.2 ± 9.0 Ma. The deflection of apparent ages is interpreted as the expression of partial recrystallization due to a disturbing event younger than 628 c. 2637 Ma of an amphibole grain whose isotopic system initially closed at c. 2680 Ma during 629 cooling of the diorite below Tc. Degassing peaks expressed during this experiment confirm 630 the coexistence of these 2 radiogenic components. The higher degassing rate of the oldest 631 component, at c. 2680 Ma, in the first part of the spectrum and the homogeneous isotopic 632 composition on that segment indicate that it is perfectly individualised, which is, however, 633 apparently not the case for the younger component associated with recrystallization in the 634 intermediate temperature domain of the spectrum where the two radiogenic components 635 overlap. The concomitant decrease of apparent ages (from c. 2749 Ma to c. 2681 Ma), combined with calculated  ${}^{37}\text{Ar}_{\text{Ca}}/{}^{39}\text{Ar}_{\text{K}}$  (#Ca/K) ratios (from c. 9.7 to 7.2) in the low to 636 intermediate temperature steps of age, the  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  spectra of amphibole 2 of sample 637 638 YD0037A and the overall saddle shape of the age spectrum suggest the incorporation of 639 excess argon. Calcic domains constitute reservoirs for excess argon as shown by Ruffet et al. 640 (1990) for rimmed and patchy-zoned hornblendes. However, the convergence between the 641 pseudo-plateau age calculated for the bottom of the saddle, at  $2683.7 \pm 4.9$  Ma (15.4% of the total <sup>39</sup>Ar<sub> $\kappa$ </sub> released), and the plateau-like age calculated for amphibole 1 supports the above 642 643 estimated age (c. 2680 Ma) of an initial isotopic closure of amphiboles from YD0037A. The 644 concordant plateau age at 2678.5  $\pm$  2.9 Ma (85.6% of the total <sup>39</sup>Ar<sub>K</sub> released), calculated from age spectrum of a biotite from sample YD0037A (Fig. 10a), is consistent with that 645 646 interpretation and suggests a fast cooling path for that poorly-deformed diorite.

647 The two amphiboles of sample YD0039A, although appearing homogeneous and 648 unaltered, also yielded distinct age spectra (Fig. 10b). Amphibole 1 yields a plateau age at 649  $2673.1 \pm 4.0$  Ma (86.6% of the total <sup>39</sup>Ar<sub>K</sub> released), slightly younger than the inferred age of 650 initial closure for amphibole YD0037A, whereas amphibole 2 yielded a disturbed age 651 spectrum showing an overall increase of apparent ages from c. 2548 Ma to c. 2620 Ma. These 652 discrepancies are reflected in the respective  ${}^{37}\text{Ar}_{Ca}/{}^{39}\text{Ar}_{K}$  spectra of the two experiments. 653 which differ in terms of ratio values and overall trend for low to intermediate temperature 654 steps. Microscopic observations reveal that some amphibole grains of YD0039A contain 655 biotite lamellae along cleavage planes, which is attested by the presence of a small degassing 656 peak in the first heating steps of the staircase-shaped age spectrum (amphibole 2 of Fig. 10b). 657 If these biotite lamellae are as young as the biotite of amphibolite YD0009A (c. 2564 Ma, see 658 below), as suggested by a low temperature pseudo plateau at  $2548.3 \pm 5.6$  Ma, their 659 degassing, partly concomitant with that of amphibole, may account for the staircase shape of 660 the age spectrum. Since amphiboles from sample YD0039A are tremolite/actinolite rather

- 661 than hornblende, their calculated  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratios are significantly higher (>20) than those 662 of the other amphiboles (hornblende), which are usually <10 (Figs. 10a, 10c, 10d and 10e).
- 663 Except for a weak deflection of apparent ages in the intermediate to high temperature 664 steps, the age spectrum of amphibole YD0009A (Fig. 10c) yields a plateau-like age of 2677.3  $\pm$  3.6 Ma (77.8% of the total <sup>39</sup>Ar<sub>K</sub> released), which is concordant with the c. 2675-2680 Ma 665 666 amphibole cooling age calculated for samples YD0037A and YD0039A. A green biotite grain 667 from YD0009A yields a plateau age approximately 110 m.y. younger at  $2564.3 \pm 3.4$  Ma (97.7% of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 10c). However, the increase of apparent ages of its 668 669 high temperature steps, up to c. 2571 Ma (Fig. 10c), possibly testifies to a nearly complete 670 reset of its isotopic system at or after c. 2560 Ma and subsequent to its initial closure. The 671 above-mentioned subtle deflection of amphibole age spectrum probably correlates with the 672 younger apparent ages (from c. 2612 to c. 2630 Ma) of the low temperature steps, which are characterized by  ${}^{37}\text{Ar}_{\text{Ca}}/{}^{39}\text{Ar}_{\text{K}}$  (#Ca/K) ratios lower than those of the main phase at c. 2678 Ma 673 674 (c. 5 vs. c. 6.5). These low-temperature steps define a specific degassing peak that probably 675 express, as for amphibole YD0039A (see above), the degassing of scarce biotite lamellae 676 plated in amphibole cleavage planes (e.g. Castonguay et al., 2001).
- 677 The age spectrum of amphibole YD0048A (Fig. 10d) is consistent with thin section 678 examination which shows light-green rimmed and/or patchy zoned minerals indicative of 679 patchy recrystallisation of amphiboles. The isotopic record of cooling at c. 2675-2680 Ma 680 shown by samples YD0037A, 39A and 0009A was erased, at least locally on a grain scale, 681 during a recrystallisation event at  $2643.9 \pm 3.5$  Ma. The analysis of the degassing rate of 682 amphibole YD0048A supports this interpretation because, unlike amphibole YD0037A, the 683 main radiogenic component (i.e. expressed by the highest degassing rate; Fig. 10d) is the 684 radiogenic component associated with recrystallisation.
- 685 The light-green rimmed and/or patchy zoned amphibole from sample YD0100A, texturally similar to those of YD0048A, provides a roughly flat but bumpy age spectrum (Fig. 686 687 10e) that yields a plateau-like age at  $2598.3 \pm 3.0$  Ma (67.3% of the total 39ArK released), 688 younger than amphibole cooling and/or disturbance ages of previous samples. It may 689 represent the best estimate of the age of the post-initial cooling disturbance recorded by amphibole 1 from diorite YD0037A. A green euhhedral biotite grain from sample YD0100A 690 691 provides a hump-shaped age spectrum with maximum apparent ages at c. 2494 Ma. This age 692 spectrum shape indicates that the biotite grain is chloritized (Ruffet et al., 1991), which casts 693 doubt on the reliability of this age.

- 694 A brown biotite grain from sample YD4065A yields a staircase-shaped age spectrum 695 whose flat segment defined by high temperature steps provide a pseudo-plateau age at 2644.3  $\pm$  3.6 Ma (Fig. 10f). Such a staircase-shaped age spectrum is typical of <sup>40</sup>Ar\* loss by volume 696 697 diffusion induced by a thermal event subsequent to initial crystallization and/or cooling (e.g. 698 Turner, 1968). The apparent age at  $2571.8 \pm 9.4$  Ma of the low-temperature step provides a 699 maximum estimate for the age of that <sup>40</sup>Ar\* loss event. Its concordance with the plateau age yielded by biotite from sample YD0009A suggests that <sup>40</sup>Ar\* loss most likely occurred 700 around c. 2560 Ma. In this scenario of <sup>40</sup>Ar\* loss, the age at c. 2644 Ma calculated from the 701 high-temperature steps should only be a minimum estimate of the age of isotopic closure 702 703 during initial cooling. Nevertheless, its perfect agreement with the recrystallization age of 704 amphibole YD0048A, also at c. 2644 Ma, argues for its validity.
- 705

## 706 5.5.2 The Opatica Plutonic Belt.

Amphibole and biotite age spectra from the OPB are show in Figure 11. All sampleswere collected in tonalites.

709 Amphibole age spectra of samples YD0018B, YD0020A, YD0063A, and YD0101A 710 are more disturbed than those of the AGB. The effects of thermal disturbances following 711 initial isotopic closure appear to be quite significant, especially those related to an event 712 equivalent to the c. 2644 Ma disturbance recorded by the AGB amphiboles. The analysis of 713 degassing patterns of amphiboles YD0020A and YD0063A (Figs. 11a and b) indicates that 714 the radiogenic component associated with partial recrystallization induced by this disturbing 715 event constitutes the main radiogenic component of these amphiboles. Related segments of 716 age spectra allow the calculation of concordant plateau ages at  $2644.5 \pm 3.8$  Ma (72.6 % of 717 the total <sup>39</sup>Ar<sub>K</sub> released) and 2648.4  $\pm$  3.0 Ma (69.6 % of the total <sup>39</sup>Ar<sub>K</sub> released),

respectively.

Several elements of the <sup>40</sup>Ar/<sup>39</sup>Ar experiments suggest that this latter event affected 719 720 amphiboles with initial ages of 2670-2665 Ma, which correspond to the apparent age bracket 721 yielded by high-temperature steps of amphiboles YD0020A and YD0063A (Figs. 11a and b), 722 at 2668.8  $\pm$  8.9 Ma and 2672.3  $\pm$  8.9 Ma, respectively. The validity of these high temperature 723 apparent ages is supported by consistencies with pseudo-plateau ages at  $2665.6 \pm 4.4$  Ma and 724  $2668.9 \pm 5.3$  Ma calculated from the disturbed age spectra of amphiboles YD0018B and 725 YD0101A (Figs. 11c and d). The degassing patterns of the latter amphiboles, especially 726 YD0101A (Fig. 11d), indicate that the related radiogenic component is a major component of

- their respective isotopic system. However, some aspects of the age spectrum shapes of
- 728 YD0018B and YD0101A remain enigmatic; for instance, the breakdown of the high-
- temperature apparent ages of amphibole YD0018B (down to  $2588.8 \pm 9.2$  Ma, Fig. 11c)
- coinciding with an abrupt and unusual drop of the  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratio (down to c. 2.69), which
- may be due to the degassing of a non-amphibolitic component. In addition, the concomitant increase of the high temperature  ${}^{37}\text{Ar}_{Ca}/{}^{39}\text{Ar}_{K}$  (#Ca/K) ratios of amphibole YD0101A and
- 733 decrease of apparent ages (Fig.11d) are unexpected. Such unusual behaviours may be due to a
- younger disturbing event, as suggested by biotite plateau ages at  $2600.3 \pm 3.0$  Ma (94.8 % of
- 735 the total <sup>39</sup>Ar<sub>K</sub> released),  $2583.4 \pm 4.5$  Ma (70.7 % of the total <sup>39</sup>Ar<sub>K</sub> released) and  $2588.2 \pm$
- 736 3.5 Ma (96.3 % of the total  $^{39}$ Ar<sub>K</sub> released) yielded by samples YD0020A, YD0063A and
- YD0035A, respectively (Figs. 11a, b and e). The existence of even younger disturbing events
- is suggested by biotites YD1245A (pseudo-plateau age at  $2544.3 \pm 3.5$  Ma) and YD0101A
- (plateau age at  $2525.4 \pm 3.1$  Ma) (Figs. 11d and f). However, the chlorite alteration of some
- biotite grains (hump-shaped age spectrum YD1245A) and/or the apparent occurrence of <sup>40</sup>Ar\*
- 741 loss (staircase-shaped age spectrum YD0101A) makes these latter age estimates hazardous.
- 742

#### 743 5.5.3 Granitic dykes.

744 Three muscovite single grains from quartz-feldspars granitic dykes collected in the 745 Lac-au-Goéland area, YD0009B, YD0023A and YD0048B, provided rather disparate results 746 (Fig. 12). The saddle-shaped age spectrum of muscovite YD0009B (Fig. 12a) suggests the 747 presence of two radiogenic components whose coexistence probably relates to a fluid-induced 748 partial recrystallization, (e.g. Cheilletz et al., 1999; Alexandrov et al., 2002; Castonguay et al., 749 2007; Tartèse et al., 2011; Tremblay et al., 2011; 2020). In such type of spectra, the plateau-750 like age at 2602.1  $\pm$  4.5 Ma (85.6 % of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 12a) yields a minimum 751 age estimate of the former muscovite, which was partially recrystallized at or after 2567.2  $\pm$ 752 6.2 Ma, i.e. the apparent age yielded by the base of the saddle. The degassing pattern and the 753 weighting of this age spectrum indicate that the older component is overwhelmingly the main 754 radiogenic component to which it is reasonable to attribute an age of c. 2603 Ma. (Fig. 12a). 755 Muscovite YD0048B provides a plateau age at  $2517.2 \pm 2.4$  Ma (93% of the total  $^{39}$ Ar<sub>K</sub> released; Fig. 12b) but a detailed examination of that age spectrum reveals a complex 756

thermal history for this mineral. Apparent ages increase steadily from low to high temperature
 steps, suggesting the occurrence, around 2500 Ma, of a disturbing event which induced <sup>40</sup>Ar\*

loss by volume diffusion (Turner, 1968) on a muscovite grain older than c. 2524 Ma (Fig.12b).

The first 30% of the  ${}^{39}$ Ar<sub>K</sub> degassing of muscovite YD0023A is highly disturbed (Fig. 761 762 12c) with erratic apparent ages which could result from the occurrence of interfoliar chlorite. 763 Overall, this age spectrum shows a saddle shape, as for muscovite YD0009B, which allows 764 the calculation of pseudo-plateau ages at 2603.2  $\pm$  4.3 Ma (21.6% of the total <sup>39</sup>Ar<sub>K</sub> released) for the base of the saddle and at 2627.1  $\pm$  3.7 Ma (47.4% of the total <sup>39</sup>Ar<sub>K</sub> released) for the 765 766 high temperature steps (Fig. 12c). Our preferred interpretation is that muscovite YD0023A 767 crystallized at the latest at c. 2627 Ma and has been affected by a fluid-induced partial 768 recrystallization event at c. 2603 Ma, which has been coeval with initial crystallization of 769 muscovite YD0009B (Fig. 12a).

770

# 771 5.5.4 The Nottaway River shear zone.

772 The analyzed muscovite of sample YD0055A belongs to a population of mica-fish 773 muscovite grains (Fig. 7e) extracted from a felsic mylonite of the Nottaway River shear zone. 774 As for muscovite YD0023A, the first 20% of the degassing spectra of YD0055A is highly 775 disturbed (Fig. 13a), which can also reflect the occurrence of interfoliar chlorite. The 776 remaining 80% of the degassing produces a multi-stepped age spectrum with a global 777 staircase shape interspersed with three flat segments. Similar muscovite age spectra have been 778 described by Tremblay et al. (2020) for Archean rocks of the southern Abitibi suprovince; 779 they were interpreted as the result of the aggregation, via dissolution/recrystallization/ 780 neocrystallization of increasingly younger white mica sub-grains as deformation increments 781 and coeval fluid inflows progress. Pseudo-plateau ages calculated from the flat segments 782 suggest periods of aggregation/remobilisation of muscovite YD0055A at 2565.2  $\pm$  3.7 Ma and 783  $2504.0 \pm 5.0$  Ma, at the expense of a muscovite grain that would have initially crystallized at 784  $2595.6 \pm 5.4$  Ma (Fig. 13a). These aggregation/remobilisation processes, which partially 785 preserve the previous or initial states of white micas, occur below their isotopic closure 786 temperature, at temperatures as low as 250-350°C.

Despite a possible plateau age at  $2527.8 \pm 2.9$  Ma (76.2% of the total <sup>39</sup>Ar<sub>K</sub> released), the age spectrum of biotite YD0059A, with recrystallized rims (Fig. 13b), shows a regular increase of apparent ages from low to high temperature steps, suggesting the occurrence of a disturbing event at c.  $2505.5 \pm 8.5$  Ma of an initial grain that would be older than c. 2531 Ma. The highly disturbed age spectrum of an amphibole from that sample is useless (Fig. 13b). 792

### 793 5.5.5 The Montviel Syenite.

794 Two amphiboles and a biotite from sample YD5143A from the Montviel Syenite (see 795 Fig. 3 for location) provide highly contrasted but significant results (Fig. 14). The age spectrum of amphibole 1, a hornblende according to its  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratio values of 4.0-4.2, 796 797 allows the calculation of a plateau age at 2653.9  $\pm$  2.7 Ma (83.6% of the total <sup>39</sup>Ar<sub>K</sub> released; Fig. 14). The age and  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  spectra of amphibole 2 are clearly different in the sense 798 799 that, from low to high temperature steps, both the apparent ages and  ${}^{37}Ar_{Ca}/{}^{39}Ar_{K}$  ratios 800 increase in unusual proportions, from c.  $1934.4 \pm 7.3$  to  $2911.4 \pm 10.8$  Ma and 2.1 to 101.6, 801 respectively, which is not complying with what can be expected for hornblendes. The 802 unrealistic age of the fusion step (at c. 2911 Ma) and its <sup>37</sup>Ar<sub>Ca</sub>/<sup>39</sup>Ar<sub>K</sub> ratio higher than 100 803 suggest the incorporation of excess argon, most likely by a calcic component resulting from 804 the alteration of the amphibole during a disturbing event which would have occurred at c. 805 1934 Ma or after (Fig. 14).

The age spectrum of biotite YD5143A is also strongly discordant with that of
amphibole 1. However, apparent ages of the base of its saddle shape, in the range c.19671940 Ma (Fig. 14), are consistent with the timing of the disturbance that presumably
destabilized amphibole 2. The destabilisation process possibly results from the emplacement
of the nearby Montviel Carbonatite, whose age is constrained by a U–Pb zircon age of 1894 ±
4 Ma (David et al., 2006).

812

# 813 **5.6 Interpretation and significance of** <sup>40</sup>Ar/<sup>39</sup>Ar ages

A synthetic compilation of all <sup>40</sup>Ar/<sup>39</sup>Ar ages measured during this study is show in
Figure 15.

816 The probability density diagram established from resized age spectra for amphiboles 817 and biotites of the AGB (Fig. 15a) allows a synthetized interpretation of analytical results. On 818 Figure 15a, the principal age peak is at  $2678.1 \pm 1.7$  Ma, as defined by amphiboles YD0009A 819 and YD0039A, as well as amphibole and biotite of sample YD0037A. The concordance of 820 biotite and amphibole ages suggests that they are cooling ages, and that this cooling has been 821 relatively fast between 600°C and 400°C. The asymmetrical and stepped base of that peak 822 toward younger ages reflects the impact of subsequent disturbing events on amphiboles after 823 initial cooling. The first of these events occurs around c. 2644 Ma. It induced a partial 824 recrystallization of amphibole YD0048A and probably the total reset of biotite YD4065A

825 (Fig. 10f). The second disturbing event occurred around c. 2598 Ma (Fig. 15a) and induced

the complete recrystallization of amphibole YD0100A (Fig. 10e). It would have been

827 followed by a third event (with a frequency peak at c. 2564 Ma) that would have caused an

828 almost complete age resetting of biotite YD0009A and a partial reset for micaschist biotite

829 YD4065A (Figs. 10c, f and 15a).

Figure 15b suggests that OPB amphiboles initially cooled down to Tc at c. 2668 Ma,
before a partial recrystallization event that occurred at c. 2647 Ma. Hence, there seems to be a

ca. 10 m.y. time lag between the AGB and the OPB for initial cooling below Tc for

amphiboles, at approximately 2678 Ma and 2668 Ma, respectively (Figs. 15a and b).

The OPB tonalites also record a younger disturbing event around c. 2593 Ma (Fig. 15b).

835 These two latter post-cooling events correlate within errors to those recorded by the AGB

amphiboles at c. 2644 Ma and c. 2598 Ma (compare Fig. 15a and b).

The three analyzed muscovite grains from granitic porphyry dykes (Fig. 15c) are indicative of a complex and polyphased thermal history that can not be precisely understood here due to the small number of analyses. This thermal evolution may have started earlier than c. 2627 Ma and ended around c. 2495-2500 Ma (Fig. 15c). The effective occurrence of a disturbing event at c. 2602.7  $\pm$  3.1 Ma, as reported on the probability density diagram of Figure 15c, remains however proven.

The initial crystallisation of muscovite YD0055A, around 2600 Ma (Figs. 13a and 15d), can be confidently attributed to shear deformation along the Nottaway River Shear Zone.  $^{40}$ Ar/ $^{39}$ Ar age data suggest that the shear zone was discontinuously active for, at least, ca. 100 m.y., from 2595.6 ± 5.4 Ma, the minimum age estimate for white mica initial crystallisation (or the complete resetting of the Ar «clock» of pre-existing muscovites) to final shear increments around c. 2500 Ma. However, this does not exclude the possible initiation of the Nottaway River Shear Zone before 2600 Ma.

<sup>40</sup>Ar/<sup>39</sup>Ar analyses suggest that the Montviel Syenite (Figs. 14 and 15e) and the nearby
carbonatite represent two different magmatic pulses, an older one related to syenite
emplacement at c. 2654 Ma and a younger one corresponding to the carbonatite intrusion at c.
1894 Ma. This is, however, in contradiction with a model proposed by Nadeau et al. (2014) in
which both the syenite and the carbonatite are attributed to a single phase of magmatism at
1894 Ma.

856

#### 857 6. DISCUSSION

858 The tectonic evolution of Archean terranes involves extensive rifting/plume/arc 859 magmatism and multiple, or progressive phases of deformation and metamorphism (e.g. 860 Cawood et al., 2009; Bédard, 2018; Brown et al., 2020; Palin et al., 2020). Understanding the 861 nature of the related geodynamic regime(s), and identifying the various alternative model(s) 862 to the classical plate tectonics interpretation, if any, has important implications for a better 863 comprehension of the early Earth dynamics, the processes that led to continental crust 864 formation and structural evolution, including the mineral deposit potential of Archean 865 orogenic belts (e.g. Thébaud and Rey, 2013).

866

# 867 6.1 The AGB-OPB contact – not a subduction suture.

868 Subduction is a geological process by which the lithosphere is recycled into the Earth's 869 mantle at convergent plate margins, creating an upper plate-lower plate (UP-LP) boundary. 870 Among the different mountain belts generated by plate convergence and collision during the 871 Phanerozoic, a significant number took place in an environment of continuing subduction and 872 accretion along sites of oceanic lithosphere consumption. These belts are known as 873 accretionary orogens, zones of type-B subduction, or Cordilleran-, Pacific- and Andean-type 874 collisions (see Cawood et al., 2009 for a review). A similar type of plate setting has been 875 suggested for the AGB-OPB boundary, with the AGB acting as the down-going plate. 876 Phanerozoic accretionary orogens are variably deformed and metamorphosed, and 877 characteristically occur as paired and parallel high-temperature (HT) and high-pressure (HP) 878 belts with metamorphic grade up to granulite and blueschist-to-eclogite facies, respectively 879 (e.g. Brown, 1993; 2010; Ernst et al., 1994). Such orogens typically include wedge-shaped 880 accretionary prisms (or subduction mélanges) that formed at the UP-LP interface. These 881 accretionary prisms are organized in thrust sheets and attributed to the offscraping and 882 underplating of material from the descending oceanic plate and of subduction-related 883 volcanic-magmatic arc system in the upper plate (e.g. Ernst et al., 1994; Cawood et al., 2009; 884 Wakabayashi and Dilek, 2011; Kusky et al., 2020; Palin et al., 2020; among many others). A 885 fore-arc basin generally occurs in between. Accretionary prisms are usually elongated and 886 depending on the plate velocity, the dip of the subduction zone, the backstop geometry of the 887 wedge and the volume of sediments on the down-going plate, they can be relatively 888 voluminous. The Franciscan subduction complex, for instance, is at least 50 km-wide along a strike-length of c. 900 km (e.g. Ernst et al., 1994; Wakabayashi, 2017). Accretionary prisms 889 890 are essentially made up of tectonic slices of island-arcs, backarcs, forearcs, dismembered

ophiolites, granulite-facies and/or ultrahigh-pressure (UHP)-HP metamorphic rocks, and preto syn-accretion clastic sedimentary sequences. Thrusting and reverse faulting usually led to
the imbrication of the upper plate rocks with those of the lower plate, leading to complex field
relationships.

895 The subduction model proposed for the AGB-OPB boundary is mainly based on the 896 Lithoprobe seismic profile (e.g. Calvert and Ludden, 1999) and field structures attributed to a 897 south-verging fold-and-thrust belt in the AGB (Sawyer and Benn, 1993; Lacroix and Sawyer, 898 1995). However, the AGB-OPB transition does not show any of the characteristic features of 899 UP-LP boundary as expected for a subduction suture. There are no accretionary prism and no 900 metamorphic break along the AGB-OPB contact. Metamorphism varies progressively from 901 low- to high-grade from the uppermost units of the AGB toward the plutonic rocks of the 902 OPB (Fig. 9). There are, obviously, local fault contacts between greenschist- and amphibolite-903 grade rocks of the AGB, such as the Barlow fault in the Chibougamau area (Bedeaux et al., 904 2020), but these are late reverse faults related to regional folding. Our field observations and 905 structural analysis indicate that the volcanic and sedimentary sequences of the AGB overly 906 the TTG-type orthogneisses of the OPB (Fig. 9), suggesting that both domains likely formed a 907 single structural unit prior to regional deformation (e.g. Benn, 2006; Benn and Moyen, 2008). 908 Hence, we do not think that the AGB-OPB transition represents a subduction-related, upper 909 plate-lower plate boundary.

910

### 911 **6.2** The formation of composite **OPB-AGB** – speculative models.

912 The structural relations and metamorphic characteristics of the AGB and OPB are
913 reminiscent of various tectonic settings such as (i) metamorphic core complexes (MCC, Lister
914 and Davies, 1989; Whitney et al., 2013), (ii) suprastructure-infrastructure transition (e.g.
915 Culshaw et al., 2006; Stern, 2017), two models in which the upper crust is tectonically
916 juxtaposed to the deep crust and/or upper mantle, and (iii) the convective overturn of crustal
917 material (Collins et al, 1998; van Kranendonk, 2011; Bédard, 2018; Wiemer et al., 2018;
918 Vanderhaeghe et al., 2019).

919

# 920 6.2.1 The OPB as a metamorphic core complex.

As compared to MCC's, the OPB shares similar features: (1) synkinematic granitoids,
pegmatites and migmatites restricted to domal core domains; (2) sheet-like granitoids and
discontinuous high-strain migmatites towards the contact with the overlying upper crust; (3) a

924 décollement horizon (i.e. potentially, the OPB-AGB contact) between lower crustal gneisses 925 and the upper crust; and (4) a normal metamorphic zonation. However, MCC's are generally 926 less than 10 km in diameter, asymmetrical (e.g. the detachment being folded due to the 927 bowing upward of the lower crust) and usually related to major continental extension (see 928 Lister and Davies, 1989). In contrast to detachments of typical MCC's, the OPB-AGB contact 929 has been involved in progressive reorientation and repeated late-intrusion. Moreover, the 930 ductility and metamorphic contrasts across the OPB-AGB interface are not extreme as in 931 typical MCC's in which the detachment separates rocks of highly contrasting structural and 932 metamorphic character, with deformation in the cover rocks being typically brittle. Finally, 933 the abundant syn-tectonic granitoids of the OPB-AGB are not confined to the proximity of the 934 detachment, as for MCC's, but extend into the interior of both the OPB and the AGB.

935

## 936 6.2.2 Superstructure-infrastructure transition.

937 In superstructure-infrastructure (S-I) transition zones, superstructure represents a low-938 grade upper crust level with early, typically upright structures, whereas infrastructure 939 constitutes a high-grade migmatitic lower crust level with late, gently inclined structures 940 overprinting the early structures of the overlying upper crust (Culshaw et al., 2006). In typical 941 examples, a crustal-scale detachment shear zone is observed between the upper and lower 942 tectonic levels (Tikoff et al., 2002; Culshaw et al., 2006; Stern, 2017). The infrastructure 943 gneissic rocks generally yield radiometric ages younger than superstructure rocks, indicating 944 that the infrastructure was hot, partially molten and weak during the deformation of the 945 superstructure above it (e.g. Culshaw et al., 2006; Stern, 2017). The OPB presents evidence 946 for anatectic melting and syn-kinematic migmatites formation at ca. 2.68-2.67 Ma (Sawyer, 1998), a melting event that, according to our <sup>40</sup>Ar/<sup>39</sup>Ar data (Figs. 15a and b), has been almost 947 948 coeval with or slightly followed the ca. 2680 Ma metamorphism in the AGB. This suggests a 949 maximum time lapse of 10 m.y. for metamorphic conditions between upper (AGB) and lower 950 (OPB) crustal levels, which could fit with the S-I model data presented by Culshaw et al. 951 (2006) for the Western Superior Province. In such interpretation, the amphibolite-facies rocks 952 occurring towards the base of the AGB would define a low-angle detachment fault at the 953 contact with deeper and hotter rocks of the OPB, a detachment that would has been 954 subsequently folded and verticalized. In the study area, these amphibolites are several km-955 thick (Bedeaux et al., 2020), and may thus reasonably represent a mid-crustal decoupling zone 956 between a ca. 15 km-thick superstructure (the AGB) and a plutonic, TTG-rich infrastructure

957 (the OPB). However, the occurrence of  $F_2$  folds affecting the low-angle metamorphic  $S_1$ 958 layering in the OPB gneisses invalidates a typical S-I transition model. The metamorphic age 959 variations can be simply related to the differential uplift and exhumation of both belts (the 960 ABG first and then the underlying OPB). A S-I transition model is also inconsistent with age 961 constraints from the Attic Complex (located in the southern part of the AGB; see Fig. 1b) 962 where U-Pb zircon dating suggests a 20 m.y gap between the formation of migmatites in 963 tonalitic gneisses (TTG's) of the lower crust (at ca. 2716.1  $\pm$  2.9 Ma) and those in adjacent 964 mafic volcanic rocks (at ca.  $2695 \pm 1$  Ma) (Revelli, 2020). This has been attributed to the 965 diapiric ascent and crustal melting of lower crustal intrusive material and the related burial 966 (and prograde metamorphism) of cold and dense greenstones of the upper crust (Revelli,

- 967 2020), which agrees with a convective crustal overturn as described below.
- 968

## 969 6.2.3 Convective crustal overturn.

970 The third tectonic setting involves the partial convective overturn of the crust (Fig. 971 16). It was originally proposed by Collins et al. (1998) for Archean rocks of the Pilbara craton 972 (Australia). In such a setting, the OPB gneissic domes represent the diapiric response of 973 density inversion created by the accumulation of a thick (ca. 15 km) greenstone cover (the 974 AGB) erupted over a previously stable and more-or-less continuous sialic crustal basement  $\geq$ 975 2.80 Ga, both domains being intruded by syn-volcanic and syn-tectonic TTG plutons ranging 976 from c. 2.76 to c. 2.65 Ga (Davis et al. 1995; Sawyer, 1998; Daoudene et al., 20156, Pedreira-977 Perez et al., 2020) (Figs. 16a and b). Diapirism is considered here as part of a larger process 978 involving the partial convective overturn of the crust (Collins et al., 1998; Wiemer et al., 979 2018). We suggest that deformation initiated by sinking and subsidence of the thick cover of greenstones within a hot and weak (ductile), laterally flowing sialic lower crust that generated 980 981 the  $S_1$  metamorphic layering both in the OPB and the lowermost part of the AGB (Fig. 16a). 982 This produced strong horizontal temperature gradients and related changes of metamorphic 983 facies as, for instance, in thermal models of Thébaud and Rey (2013). Due to density 984 contrasts, both the gneissic lower crust and crosscutting TTG's ultimately migrated upward 985 passively as accommodation structures. Material flow into the AGB synclines would has been 986 compensated by lateral and vertical migration of the less dense sialic substrate into the domal 987 crests of the OPB (Fig. 16b), generating a pattern of partial convective overturns in the crust 988 (Collins et al., 1998; Weinberg, 1997). The downflow of greenstones would has been almost 989 entirely compensated by TTG uprises (Fig. 16c). Shallow interdome depressions developed

990 and progressively filled with clastic sediments, such clastic sequences occupying the core of 991 greenstone synclines and being represented by sedimentary rocks of the Opémisca and 992 Broadback groups within the Abitibi and Frotet-Evans belts, respectively. Some of these 993 clastic sequences possibly formed as early as ca. 2750 Ma, as suggested by stratigraphic 994 relationships between the Broadback Group and the underlying Frotet-Evans volcanics 995 (Brisson et al., 1998; see section 2.1). In the Opémisca Group, maximum sedimentation age 996 of the Haüy Formation ( $\leq 2691.7 \pm 2.9$  Ma; David et al., 2007) is coeval, at least in part, with 997 regional metamorphism in the AGB (see section 5.5). Part of the Opémisca Group therefore 998 deposited during regional deformation, as proposed for most sedimentary sequences of the 999 Superior Province (e.g. Leclair et al., 1993; Parmenter et al., 2006; Lin et al., 2013). As the 1000 greenstones and overlying sedimentary rocks progressively flow deeper into the synclines, 1001 strain intensity increased and was accommodated by horizontal shortening, the S<sub>2</sub> foliation 1002 became more pervasive and faults and shear zones with down-dip lineations developed (Fig. 1003 16b). The Barlow fault (Bedeaux et al., 2021) is one of these; in the Chibougamau area, it is 1004 located along the northern flank of the Waconichi F<sub>2</sub> syncline (Waconichi basin on Fig. 2) and 1005 marked by upper crustal, south-dipping seismic reflectors (Mathieu et al., 2020). The Barlow 1006 fault is interpreted as a reactivated structure (Bedeaux et al., 2021), and it possibly belongs to 1007 a series of basin-bounding faults that accommodated the exhumation of the OPB gneisses and 1008 inverted during the following contraction (e.g. Mathieu et al., 2020), which is consistent with 1009 the interpretation shown on Figure 16. With time, the early generation of TTG's were 1010 progressively deformed and rotated into subconcordance with the steepening domal 1011 boundaries (Fig. 16c). Other TTG's were syn-kinematic and formed sheeted intrusions. Some 1012 of these did not move far from their source, remained heterogeneous, and can be better 1013 described as `intrusive diatexites' (Sawyer, 1998), whereas later intrusions tend to be larger, 1014 more discordant, less strongly foliated bodies and are more obvious near the AGB-OPB 1015 boundary, where they cut the OPB-AGB interface. Such an interpretation fits well with (1) the 1016 relative downflow of the AGB volcanic rocks as compared to the OPB gneisses (Fig. 9), (2) 1017 the distribution of  $S_1$  foliations and lineations in the OPB domes (Fig. 2 and Fig. 6b), (3) the 1018 up-stratigraphy increasing proportion of plutonic clasts in polygenic conglomerates of both the Broadback and Opémisca groups (Brisson et al., 1997a; 1998; Charbonneau et al., 1991), 1019 1020 which reflects the progressive erosion of volcanic cover rocks and underlying plutonic rocks 1021 in the source area, and (4) slightly higher pressure and temperature preserved in the OPB 1022 migmatites and amphibolites (6.3-7.1 kb and 750°C, Sawyer, 1998) as compared to the

1023 amphibolite facies rocks of the AGB (c. 6 kb and c. 550°C, Bedeaux et al., 2021).

1024  $^{40}$ Ar/ $^{39}$ Ar age measurements performed during our study are in good agreement with 1025 that model. Amphibole ages suggest that the lowermost part of the AGB exhumed around 1026  $2678.1 \pm 1.7$  Ma (Fig. 15a). Ages from the OPB, at  $2667.8 \pm 3.0$  Ma (Fig. 15b), suggest that it 1027 was more-or-less coeval with, or quickly followed the exhumation of the AGB, an event that 1028 we suggest has been coeval with decompression melting (Weinberg, 1997; Whitney et al., 1029 2004) and the genesis of syn- to late-D<sub>1</sub> migmatites in the cores of OPB gneiss domes at 2.67-1030 2.68 Ga (Sawyer, 1998; Fig. 16c). Then, both terranes share the same succession of thermal 1031 disturbances at c. 2650-2645 Ma and 2600-2590 Ma, including the inferred onset of shearing 1032 along the Nottaway River Shear Zone at ca. 2600 Ma, as recorded by related mylonites (Fig. 1033 15d) and quartz-feldspar porphyry dykes (Fig. 15c). Such a series of events is also 1034 documented in the Val d'Or mining district of the Abitibi subprovince (Tremblay et al., 1035 2020). Besides, at the scale of the Superior Province, there is accumulating geochronological 1036 evidence for major crustal deformation event(s) and related metamorphism spreading over a 1037 timeframe much larger than suggested in literature (e.g. Percival, 2007). Easton (2000), for 1038 instance, argued that the distribution of metamorphic grade and age of metamorphism in the 1039 western Superior reflect a series of tectonic events between 2710 and 2640 Ma, which is in 1040 agreement with U-Pb zircon age constraints for metamorphism, syn-orogenic magmatism and 1041 deformation in the southern Abitibi subprovince (see Tremblay et al., 2020 for a review). 1042 North of the OPB, in the Opinaca subprovince, granulite-facies metamorphism spreads over 1043 ages varying from c. 2666 Ma to 2636 Ma (Morfin et al., 2013) whereas it even lasted down 1044 to ca. 2620–2600 Ma in supracrustal rocks of the La Grande subprovince (Fontaine et al., 1045 2017, 2018). Combined with geochronological data from our study, this is a strong indication 1046 that peak and duration of regional metamorphism have been more-or-less coeval and long-1047 lasting over a large region of the Superior Province, if not all over the Pontiac, Abitibi and 1048 Opatica subprovinces and beyond.

1049

#### 1050 **6.3 Significance of the Lithoprobe seismic reflector.**

1051 Seismic profiles showing shallow sub-horizontal to slightly dipping reflectors beneath 1052 Archean and/or Paleoproterozoic greenstone belts are relatively common. Some of these 1053 reflectors may be due, obviously, to tectonic events that are younger than, and not directly 1054 related to, the formation of the greenstone belts and, until the age of such reflectors can be 1055 determined, their bearing on the greenstone belt evolution remains speculative. However, assuming that the Lithoprobe seismic reflector of the Superior Province is an Archean artefact
(e.g. Cawood et al., 2006) and that our interpretation of the AGB-OPB relations is correct,

- 1058 what can be the origin and significance of that reflector then? We envision two possibilities
- that we briefly discuss below, (i) mantle lithosphere imbrication (Fig. 17), and (ii) bending of
- 1060 a flat subduction rooted southward at the Pontiac-Abitibi interface (Fig. 18).

1061 Gray and Pysklywec (2010) performed a series of forward numerical modeling of the 1062 thermo- mechanical evolution of continental lithosphere undergoing collision and orogenesis 1063 under Neoarchean- like conditions for stratified mafic-to-felsic crustal domains overlying the 1064 mantle lithosphere. They have identified three dominant modes of mantle lithosphere 1065 deformation, (1) pure- shear thickening; (2) mantle imbrication; (3) and underplating. Their 1066 result for the mantle imbrication mode is reproduced on Figure 18. For that mode, numerical 1067 modeling shows that, due to a «hot» and weak lower crust, there is a decoupling between the 1068 mantle and the crust at a depth of approximately 40 km. The model shows that the 1069 lithospheric mantle becomes imbricated along major shear zones along which slices of crustal 1070 eclogites may occur, whereas the crustal domain itself essentially absorbs deformation by 1071 pure-shear thickening and folding. Gray and Pysklywec (2010) inserted a weak zone in the 1072 initial configuration of their numerical experiments to favor the onset and localization of 1073 deformation, and to mimic an original plate boundary (see their Figure 1) as well. Although 1074 that such a weak zone may has influenced the location of some shear zones within the mantle 1075 lithosphere, it did not lead to the formation of any shear zones (that could have been 1076 interpreted as a nascent subduction plane) in the upper crust. The modeled imbrication shown 1077 on Figure 17 is obviously unrelated to subduction since the crustal part of the lithosphere is 1078 not recycled into the mantle as in typical subduction settings. We think that mantle 1079 imbrications as shown by these results likely formed during the Archean and can account for 1080 deep seismic reflectors such as the Abitibi-Opatica Lithoprobe transect.

1081 An alternative hypothesis that cannot be ignored is that the Lithoprobe seismic 1082 reflector may not represent the original location of an Abitibi-Opatica subduction zone, but 1083 the trace of a Pontiac-Abitibi subduction zone interface that flattened for several hundreds of 1084 km at depth and steepen at or close to the Abitibi-Opatica contact. Such shallow subduction 1085 flattening has been already invoked as a predominant tectonic regime during the Archean due 1086 to thicker and more buoyant oceanic lithosphere (e.g. Cawood et al., 2006). Flat subduction is 1087 documented on modern Earth. A typical example of it is the Mexican subduction zone where 1088 the young Cocos plate is shallowly subducting beneath central Mexico (Fig. 18a; Kim et al.,

1089 2012). This is well illustrated by the teleseismic imaging of a 500 km long transect between 1090 the Middle America trench to the south, and the Trans-Mexican volcanic belt (TMVB, the 1091 continental volcanic arc) to the north (Fig. 18b). This shows that subduction of the Cocos 1092 oceanic plate dips c. 15° north for about 80 km along the coast, and then horizontally 1093 underplates the continental crust at c. 50 km-depth for c. 250 km northward. The slab then 1094 abruptly changes to a steeply dipping geometry at approximately 75° as it bends right below 1095 the TMVB. The similarity of the geometry of the Cocos plate subduction zone and the 1096 reinterpreted Abitibi segment of the Grenville-Abitibi seismic transect is striking (Fig. 18c), 1097 both in terms of depth and lateral extent of seismic reflectors. Hence, it is possible that the 1098 steeply north-dipping seismic reflector of the Abitibi Lithoprobe transect roots into the 1099 Pontiac-Abitibi interface rather than the AGB-OPB. The Pontiac-Abitibi contact would be 1100 then interpreted as a surface of former subduction and recycling of the oceanic crust that 1101 originally existed between the Pontiac and Opatica «basements», causing the underthrusting 1102 and burial of the Pontiac block below the Abitibi subprovince and accounting for ensuing 1103 tectono-metamorphism and magmatism (e.g. Dimroth et al., 1983; Card, 1990; Camiré and 1104 Burg, 1993; Davies, 2002; Frieman et al., 2017, among many others). This may possibly 1105 explain the contrasting pre-2700 Ma plume-driven constructive stage of the AGB as 1106 compared to its post-2700 Ma geodynamic evolution as depicted by Thurston et al. (2008) 1107 and Mole et al. (2021); the latter stage being mainly characterized, according to these authors, 1108 by subduction-related compressional lithospheric deformation, flysch sedimentation and 1109 more-or-less coeval emplacement of alkalic granodiorite, granite and syenite plutons (such as 1110 in the Preissac-Lacorne batholith). However, such an uniformitarian interpretation for the 1111 Pontiac-Abitibi relations has been recently-disputed by Piette-Lauzière et al. (2019) that argue for a multi-stage tectono-metamorphic evolution involving extensional tectonism and crustal 1112 1113 delamination followed by re-accretion and subcretion of ribbons continents as suggested by 1114 Bédard (2018) for the Northern Superior.

1115

### 1116 CONCLUSION

1117 Our structural and metamorphic analysis of the transition between the Abitibi and 1118 Opatica subprovinces does not confirm the former existence of north-dipping subduction 1119 between these two lithological assemblages, despite the occurrence of a major seismic 1120 reflector at depth. We think that both terranes originally formed a single and composite 1121 crustal sequence, the Abitibi and Frotet-Evans greenstone belts and associated sedimentary

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1122 rocks representing the upper crust and the Opatica Plutonic Belt being the remnant of middle-1123 lower crust plutonic domains. This implies that the Frotet-Evans belt originally formed an 1124 upper crust sequence that was continuous with the AGB. Although there is a lack of precise 1125 geochronological constraints for the Frotet-Evans belt, such an interpretation is consistent 1126 with available U-Pb age data from both greenstone belts. However, it does not mean that the 1127 Frotet-Evans belt has ever been as thick as the AGB (c. 10-15 km), since there are erosional unconformities on top of each volcanic sequences that may has removed more volcanic 1128 1129 material from the former belt (due to the progressive exhumation of the OPB; see Fig. 16) as 1130 compared to the AGB. Our interpretation also means that OPB-like orthogneissic massifs are 1131 likely to be exposed south of the OPB-AGB transition. We are not saying here that OPB 1132 gneisses are systematically present at depth below the AGB but that discontinuous masses of 1133 orthogneissic TTGs, more-or-less attached to each other and to the Opatica, are likely to occur 1134 and to be expose along the crest of antiformal/domal culminations within the Abitibi belt. 1135 Examples of such metamorphic domes may be represented by the Marest and Bernetz 1136 gneissic plutons in central Abitibi (Faure, 2015), which are currently interpreted as the crustal 1137 roots of volcanic arc(s) (Chown et al., 2002), and by the Attic Complex in the Lebel-sur-1138 Quévillon area (Fig. 1b) that has been recently interpreted as the result of diapiric and 1139 sagduction-related processes (Revelli, 2020).

1140 As for most Archean cratons, the structural evolution of the Opatica and Abitibi belts 1141 is the result of several episodes of regional deformation and metamorphism that spans several 1142 tens, if not hundred's, of million years. The main events,  $D_1$  and  $D_2$ , are predominantly 1143 developed in the OPB and the AGB, respectively. S<sub>1</sub> is a penetrative amphibolite-grade 1144 foliation in the OPB gneisses and in metavolcanic rocks lying at the base of both the Frotet-1145 Evans and Abitibi greenstone belts whereas S<sub>2</sub> is a medium- to low-grade regional schistosity of variable intensity in both the AGB and OPB. S<sub>1</sub> is subparallel to the OPB-AGB contact 1146 1147 although it is frequently transposed into a composite  $S_{1-2}$  fabric within the AGB.  $S_1$ 1148 trajectories in the OPB define a dome-and-basin geometry that is the result of vertical 1149 transfert of middle- to lower-crustal material. The counterpart progressive burial of the AGB 1150 into synformal troughs generated the abrupt EW-trending S<sub>2</sub> schistosity and related down-dip mineral and stretching lineations. Amphiboles and micas <sup>40</sup>Ar/<sup>39</sup>Ar ages measured in both the 1151 1152 OPB and AGB suggest that exhumation of the latter started at  $\geq$  2678 Ma and was almost coeval with, or slightly followed, by the uplift and related migmatization of OPB gneisses at 1153 1154 c. 2668 Ma or earlier. This is consistent with a diapiric convective overturn of the crust as

- 1155 illustrated on Figure 16. Both the OPB and AGB then shared the same magmatic and 1156 structural evolution, including the onset of dextral and sinistral transcurrent shearing and 1157 faulting at ca. 2600 Ma that may have lasted for c. 100 m.y. Comparison with adjacent areas 1158 suggests that magmatism and metamorphism have been coeval over a large region (Easton, 1159 2000), which is consistent with pervasive deformation and slow cooling as expected for 1160 vertical tectonic models and diapiric magmagenesis and ascent during the evolution of the 1161 Archean Superior Province (e.g. Bédard and Harris, 2014; Bédard, 2018). 1162 1163 **Declaration of Competing Interest.** 1164 The authors declare that they have no known competing financial interests or personal 1165 relationships that could have appeared to influence the work reported in this paper. 1166 1167 ACKNOWLEDGEMENTS 1168 Y.D. benefited of a financial support from Mitacs-Acceleration and the Ministère de 1169 l'Énergie et des Ressources (MERN) du Québec for a postdoctoral fellowship at UQAM in 1170 2014-16. The Natural Science and Engineering Council of Canada (NSERC) has provided 1171 research grants to A.T. (NSERC-PG105669). Thanks to A. Chabot-Bergeron, A.-S. 1172 Corriveau, P. Lamontagne-Hallé, A. Moïse, A. Nolet-Regaudie, T. Plasmodon-Tremblay, and 1173 C. Vézina for help and support during field work. Thanks are also due to Michèle Laithier for 1174 drawing some of the figures. We thank Dr Phil Thurston and three anonymous reviewers for 1175 thoughtful comments that significantly improved the manuscript. 1176 1177 **REFERENCES** 1178 Alexandrov, P., Ruffet, G., Cheilletz, A., 2002. Muscovite recrystallization and saddle-shaped <sup>40</sup>Ar/<sup>39</sup>Ar age spectra: Example from the Blond granite (Massif Central, France). 1179
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- 1641

## 1642 **Figure captions.**

Figure 1. a) Principal tectonostratigraphic domains of the Archean Superior Province.
Modified from Card (1990) and Benn et al. (1992). b) Geological map of the southeastern
part of the Superior Province with focus on the Abitibi greenstone belt and the surrounding

- 1646 areas. AC: Attic Complex; BP: Bernetz gneissic pluton; MP: Marest gneissic pluton.
- 1647 Modified from Thurston et al. (2008). See a) for location.
- 1648 Figure 2. Geological map of the northeastern part of the northeastern part Abitibi
  1649 Subprovince and the eastern part of the Opatica Subprovince including the eastern portion of

1650 the Frotet-Evans greenstone belt. Samples YD101A and YD0100A used for <sup>39</sup>Ar-<sup>40</sup>Ar dating

are localized on the map. BSZ, Barlow shear zone; GF, Gwillim fault; KSZ, Kapunapotagen

1652 shear zone; LF, Lamarck shear zone; LFSZ, La France river shear zone; LLSZ, La Trève lake

1653 shear zone; LNSZ, Lac à l'Eau-Noire shear zone; LSRSZ, Lucky Strike river shear zone;

1654 NRSZ, Nottaway river shear zone. See Fig. 12 for location. Modified from Dimroth et al.

1655 (1986) and Daigneault et al. (1990).

Figure 3. Geological map of the Lac-au-Goéland area and location of samples used
 for <sup>40</sup>Ar-<sup>39</sup>Ar for this study. See Fig. 2 for location. Modified from Goutier (2005).

1658 Figure 4. Field photographs of the AGB and OPB rocks. a) Amphibolite-grade mafic 1659 volcanic rock showing well-preserved pillow structure. b) Strongly deformed gabbro showing 1660 quart-feldspar aggregates defining a steeply plunging mineral lineation. c) Typical gneissic 1661 tonalite of the OBP characterized by a penetrative  $S_1$  foliation. d) Un-metamorphosed 1662 succession of sandstone and siltstone of the Opémisca Group (Daubrée Formation) in the Chibougamau. e) AGB amphibolite located along the southern boundary of the Noman 1663 1664 tonalite (see Fig. 3), showing the  $S_1$  metamorphic layering tightly folded by F2 with an axial-1665 planar S2 foliation. f) Granitic dykes crosscutting amphibolitic horizons within the OPB.

Figure 5. Structural map of the study area showing foliation trajectories and dip
directions. Structural data were compiled from the geological mapping data available in
SIGEOM information system of the Ministère des Ressources naturelles du Québec
(http://sigeom.mines.gouv.qc.ca). Trajectories of regional schistosity/foliation were drawn
from structural data acquired in the field and high-resolution magnetic imaging.

Figure 6. Stereographic projections (equal area, lower hemisphere) for dominant
 planar and linear structural fabric of the Abitibi and Opatica subprovinces. Data for the AGB
 correspond to S<sub>2</sub> and composite S<sub>1-2</sub> whereas those from the OPB correspond to S<sub>1</sub>.

1674 Figure 7. Field photographs of fabric related to the D<sub>3</sub> deformation event. a) Mylonitic 1675 fabric developed in a mafic volcanic rock of the AGB. b) Dextral F<sub>3</sub> fold affecting the 1676 regional schistosity S<sub>2</sub> within a mafic volcanic rock. c) Subhorizontal stretching lineation 1677 along the Nottaway River shear zone. d) Shear bands indicating dextral shearing in a 1678 mylonitized orthogneiss of the Nottaway River shear zone (view perpendicular to the 1679 subvertical  $S_3$  foliation and parallel to the subhorizontal lineation). e) Thin section cuts 1680 parallel to the XY plane showing dextral mica fish fabric in a felsic mylonite. f) Boudinaged 1681 felsic dyke cutting across an amphibolite horizon of the OPB.

1682 **Figure 8.** Photographs illustrating regional metamorphism and associated deformation

1683 in the study area. a) Low-grade volcanic rocks of the Roy Group showing well-preserved 1684 pillow structures. b) Garnet-staurolite micaschist from the northwestern flank of the 1685 Waconichi syncline. Staurolite forms large, corroded crystal preferentially aligned along the 1686 foliation plane (cross-polarized light). c) Large chloritoid phenocryst overprinting the regional 1687 schistosity marked by the preferred orientation of actinolite in the fine-grained matrix (cross-1688 polarized light). d) Outcrop of pillowed basalt showing garnet-rich pillow margins in the 1689 vicinity of the AGB-OPB transition. e) Typical lobate boundaries between quartz and feldspar 1690 grains in a coarse-grained and weakly-deformed tonalite of the OPB. The foliation is marked 1691 by the preferred orientation of biotite (plane-polarized light). f) Field example of migmatitic 1692 gneiss of the OPB.

Figure 9. Structural profiles of the study area. See figure 3 for location. East-southeast-looking for profile A-A', East-looking for the other profiles.

**Figure 10.** Single-grain  ${}^{39}$ Ar/ ${}^{40}$ Ar results for amphiboles and biotites of the AGB. Apparent age error bars are at the 1 $\sigma$  level; errors in the J-parameter are not included. Plateau and pseudo-plateau ages (1 $\sigma$  uncertainties including errors in the J-parameter) are given when applicable. Sample locations are show in Figures 3 and 4. See text for explanations.

Figure 11. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for amphiboles and biotites for tonalites of
 the OPB. See Figures 3 and 4 for locations. Analytical details as for Figure 11. See text for
 explanations.

Figure 12. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for muscovite of quartz-feldspar porphyry
dykes of the Lac-au-Goéland area. Analytical details as for Figure 11. See Figure 4 for
locations. See text for explanations.

Figure 13. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for muscovite, amphibole and biotite from
the Nottaway River shear zone. Analytical details as for Figure 11. See Figure 4 for locations.
See text for explanations.

Figure 14. Single-grain <sup>39</sup>Ar/<sup>40</sup>Ar results for amphibole and biotite from the Montviel
Syenite. Analytical details as for Figure 11. See Figure 4 for locations and text for
explanations.

Figure 15. Synthesis of the validated ages of the study area (plateau and pseudoplateau) obtained on amphibole and micas grouped by terranes and/or rock types on probability density diagrams of the resized apparent ages for each group. See text for explanation.

1715 Figure 16. Schematic illustration of a partial crustal overturn model for the OPB-AGB

transition. See text for discussion and details on age constraints for the different panels.

- Figure 17. Extract of numerical models performed by Gray and Pysklywec (2010).
  This model suggests that due to a hot and rheologically-weak lower crust, the mantle and the
  crust become decoupled at a depth of approximately 50 km and that the lithospheric mantle
  becomes imbricated along major shear zones whereas the crustal domain itself 2 of Gray and
  Pysklywec (2010). See text for discussion.
- Figure 18. a) Teleseismic imaging of the subduction zone interface of the Cocos
  slab beneath Mexico. Modified from Kim et al. (2012). b) Alternative interpretation for the
- 1724 AGB showing flat subduction of the Pontiac terrane below a composite AGB-OPB terrane.
- 1725 Note the striking geometrical similarity with the Cocos subduction zone. See text for
- 1726 discussion.

































## From c. 2695 Ma...



Undifferentiated OPB gneiss and TTG









## **Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

⊠The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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The paper presents the result of a postdoctoral study performed by the first author. YD made field observations, the structural and metamorphic analyses, and the lithological sampling. He wrote the paper and drawn most of the figures. As research director, AT supervised YD, and participated in several field trips, discussions in many ways and the writing of the paper. GR performed the Ar-Ar analyses, wrote the section 40Ar/39Ar Geochronology and participated in the elaboration of the discussion. FL accompanied the first author on the field and participated in discussions about regional geology.

All authors read and approved the final manuscript.