**Early gold-bearing quartz veins within the Rivière-Héva
Fault Zone, Abitibi Subprovince, Quebec, Canada**

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Early gold-bearing quartz veins within the Rivière-Héva Fault Zone, Abitibi Subprovince, Quebec,
Canada

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Abstract

The Malartic Lakeshore showing is a gold-bearing quartz vein system located within the major Rivière-Héva Fault Zone (RHFZ) of the Southern Abitibi greenstone belt. This fault separates the 2700–2702 Ma felsic Héva Formation from the 2708 Ma mafic-ultramafic Dubuisson Formation. A swarm of thin diorite dykes with lamprophyric facies and gold-bearing quartz veins are only present on the Dubuisson side of the fault. The 30 to 70 cm thick gold quartz veins are boudinaged and folded. Veins are banded and associated with pyrite, chalcopyrite, galena, barite, and gold. The study area is characterized by a high degree of ductile deformation associated with the RHFZ and manifested by the southeast trending “principal schistosity” (Sp). Stretching lineations plunge moderately to shallowly toward the southeast as a result of shortening followed by late directional shearing during a transpressive deformation. A sample from the Héva Formation yielded a zircon U–Pb age of 2698.2 ± 0.8 Ma, and a diorite dyke produced an age of 2694.3 ± 2.5 Ma. Quartz veins are cross-cut by dykes and both are affected by the Sp fabric indicating an early emplacement with respect to the deformation. This situation contrasts with orogenic gold veins found in association with major fault zones. A near synvolcanic magmatic hydrothermal origin is proposed for this gold vein system. As all subvertical units in the area are south-facing, the presence of the older Dubuisson Formation over the younger Héva Formation is attributed to the RHFZ acting as a significant reverse or thrust fault.

Keywords: Archean Abitibi greenstone belt, Rivière-Héva Fault Zone, gold-bearing quartz veins, lamprophyre

38 Introduction

39 The Archean Abitibi Subprovince is known for its gold endowment especially along major fault
40 zones or breaks, such as the Cadillac–Larder Lake Fault Zone (CLLFZ) (Dubé and Gosselin 2007;
41 Bedeaux et al. 2017). Major fault zones are considered as preferential conduits for magmatic
42 intrusions and hydrothermal fluids (Robert et al. 2005). The Canadian Malartic mine is a typical
43 example of this context with several monzodiorite dykes and stocks overprinted by a hydrothermal
44 system located near the CLLFZ (De Souza et al. 2015). Typical orogenic gold-bearing quartz veins
45 are well defined in the Val-d’Or area (Dubé and Gosselin 2007), exemplified by the Sigma-
46 Lamaque deposit, a world-class type locality (Robert and Brown 1986). At Sigma-Lamaque, gold
47 quartz tourmaline veins occur in subvertically to moderately dipping shear zones associated with an
48 array of subhorizontal extensional veins developed preferentially within competent intrusive bodies.
49 This association of gold deposits within or near a major fault or shear zone is therefore a common
50 feature. However, early “pre-regional deformation” gold systems are also documented in the Val-
51 d’Or area within many deposits including the Norlartic (Couture et al. 1994) and Siscoe mines
52 (Backman 1936; Moss 1939; Sauvé et al. 1993; Robert 1994). Such early vein-type deposits can
53 form during a synvolcanic stage and may correspond to submarine magmatic hydrothermal systems,
54 e.g. the “Géant Dormant” deposit (Gaboury and Daigneault 1999) and the Doyon-Westwood mines
55 (Yergeau et al. 2015). When these types of gold veins occur in major deformation zones, their
56 synvolcanic origin is difficult to assess. Such is the case for the Malartic Lakeshore showing, which
57 is a gold vein-type (Bousquet and Carrier 2008; Guay et al. 2015, 2016) located within the major
58 Rivière-Héva Fault Zone (RHFZ) of the Southern Abitibi greenstone belt (Fig. 1). This study
59 investigates its occurrence within the RHFZ, its peculiar base metal paragenesis with Bi, Ag-
60 telluride, and barite, as well as its association with intermediate and lamprophyric dyke swarms.

61 This paper will first address the structural evolution of a major fault zone of the Abitibi belt as
62 deciphered from a type locality of the RHFZ. Second, it will illustrate how a gold vein system can
63 be misinterpreted as an orogenic system due to its spatial association with a major shear zone.
64 Finally, this paper proposes an early magmatic hydrothermal system modified by the RHFZ.

65

66 **Geological setting**

67 The study area is located in the southern portion of the Archean Abitibi Subprovince (Dimroth et al.
68 1982; Card 1990), also recognized as the Abitibi greenstone belt (Ayer et al. 2002). More precisely,
69 it is located within the Southern Volcanic Zone (SVZ; Chown et al. 1992; Mueller et al. 1996,
70 Daigneault et al. 2002, 2004) at the interface between the Malartic and Louvicourt groups (Fig. 1).

71 The 2706–2714 Ma Malartic Group (Pilote et al. 1999) is made of interdigitated ultramafic and
72 mafic flows (Pilote et al. 2015a; Pilote et al. 2000) of tholeiitic affinity. This group is subdivided
73 (from base to top) into the La Motte-Vassan, Dubuisson, and Jacola formations and is globally
74 viewed as having formed in an oceanic crust volcanic environment (Pilote et al. 2000; Scott et al.
75 2002) or in an oceanic plateau setting (Desrochers et al. 1993). The 2705–2700 Ma Louvicourt
76 Group, which lies conformably on the Malartic Group, is further divided into (from base to top) the
77 Val-d'Or and Héva formations. Intermediate to felsic flows and volcanoclastic rocks of the Val-d'Or
78 Formation display a transitional to a calc-alkaline affinity and are interpreted as having formed in an
79 arc environment (Scott et al. 2002). The 2700–2702 Ma Héva Formation contains mafic flows of
80 tholeiitic affinity and minor felsic volcanoclastic deposits of transitional to calc-alkaline affinity
81 (Scott et al. 2002). The sedimentary basins in the Abitibi belt have been divided into two distinct
82 environments by Mueller and Donaldson (1992): (i) the flysch-type turbiditic facies of the
83 Kewagama and Cadillac groups (Fig. 2) and (ii) the molasse-type Timiskaming Group located along

major fault zones such as the CLLFZ. The conglomerate-sandstone facies of the Timiskaming Group has a maximum age of 2672 Ma (Davis 2002, Pilote et al. 2015a, 2015b).

The general structural features of the SVZ can be summarized into (i) a subvertical attitude of volcanic and sedimentary strata, (ii) regional east-trending fold axial traces associated with a subvertical axial-planar schistosity, (iii) dominant dip-parallel stretching lineations, and (iv) east-trending reverse fault zones generally located along contacts between major units, such as sedimentary and volcanic units. This pattern was interpreted as the manifestation of a regional north–south shortening (Dimroth et al. 1982; Chown et al. 1992; Mueller et al. 1996, Daigneault et al. 2002). A common northeast-trending crenulation cleavage is developed locally in high-strain zones associated with major faults. A distinct variation in this pattern is recognized in the Malartic segment of the SVZ. Here the main lithological units and the major fault zones, such as the CLLFZ, exhibit a southeastern trend distinct from the regional or principal schistosity (Sp), which remain east-trending outside the influence of the fault zone (Bedeaux et al. 2017).

The Malartic region of the SVZ is cross-cut by several longitudinal fault zones commonly located at the contact between the stratigraphic units (Daigneault 1996; Daigneault et al. 2004; Fig. 1). The CLLFZ, best known for its gold endowment, separates the sedimentary Pontiac Group to the south from the volcano-sedimentary rocks of the Abitibi belt to the north (Robert 1989; Daigneault et al. 2004; Bedeaux et al. 2017). The RHFZ is considered as one of the most significant faults in this portion of the Abitibi greenstone belt (Pilote et al. 2015a). This fault, *sensu stricto*, separates the felsic units of the 2700–2702 Ma Héva Formation from the 2708 Ma mafic-ultramafic units of the Dubuisson Formation. The RHFZ can be followed for more than 100 km from west to east (Pilote et al. 2015a) (Fig. 1) transitioning from east-trending to southeast-trending in the Malartic segment and

reverting to east-trending in the Val-d'Or area to the east. The study area consists of two large stripped outcrops (Fig. 2) that cover 5600 and 5800 m², respectively. These were partly exposed for the regional study of Neumayer et al. (2000). The western outcrop is dominated by an assemblage of mafic-pillowed volcanic rocks (Fig. 3A) associated with komatiitic basalt. Pilote et al. (2015a) interpret this sequence to be part of the 2708 Ma Dubuisson Formation (Malartic Group). The eastern outcrop displays the contact between the previously mentioned assemblages and the felsic volcanoclastic rocks (Fig. 3B) belonging to the 2702 Ma Héva Formation (Louvicourt Group). The area presents an exceptional density of centimetre to decimetre thick dykes of intermediate composition (Fig. 3C) that display a lamprophyric facies (Bousquet and Carrier 2008; Guay et al. 2015, 2016; Mathieu et al. 2018). A small-volume tonalitic stock, located in the covered area between the outcrops, is intersected in boreholes (Bousquet and Carrier 2008, 2009a, 2009b) and yields a radiometric U–Pb age of 2690.7 ± 6.3 Ma on zircon grains (David et al. 2018).

The study area is characterized by upper greenschist facies metamorphism (Pilote 2013; Faure 2015) and displays the effect of strong ductile deformation related to the RHFZ manifested as a well-developed principal schistosity (Sp).

Methodology

The two outcrop zones were mapped in detail, and special emphasis was placed on the description and sampling of veins and dykes. Structural measurements were made systematically in the field. In addition, 80 oriented samples were analysed in the laboratory to extract precise measurements of linear fabrics, in particular stretching lineation. A field planar anisotropy index (PAI; Table 1) was used to qualify the degree of planar fabric development observed on the surficial outcrop plane (Daigneault 1996). This scale varies from 0 (no fabric) to 5 (strong mylonitic fabric) to estimate the

development of structural planar anisotropy. A PAI <3 represents rocks having preserved primary volcanic or sedimentary features, while a PAI >3 is attributed to rocks dominated by secondary structural features (fabrics).

Samples were prepared for whole rock geochemical analysis using lithium metaborate fusion. Major and trace element concentrations (Appendices 1 and 2) were analysed via inductively-coupled plasma mass spectrometry (ICP-MS). Both analyses were undertaken at Actlabs Ltd. using MERN standards and QA–QC controls. Mass balance was calculated for the altered samples using the neural network method for modelling precursors, as established and described in Trépanier et al. (2015). Microprobe analysis was performed at Laval University on polished sections of representative sulphide-bearing veins. Barite crystals in veins were identified using a μ XRF at the University of Québec in Chicoutimi (UQAC).

For U–Pb geochronological analysis, selected samples were prepared using standard crushing and grinding techniques. Heavy minerals were concentrated using a Wilfley table and heavy liquid techniques. Mineral separates were sorted by magnetic susceptibility using a FrantzTM isodynamic separator and then hand-picked under a binocular microscope. Analyses used individual grains of zircon that were chemically abraded following the techniques of Mattinson (2005). This included annealing the grains for 48 h at 1000 °C prior to leaching with HF at 180 °C for 10 to 20 h. Details of zircon morphology and quality are summarized in Table 2. U–Pb isotope dilution thermal ionization mass spectrometry (ID-TIMS) techniques used in this study were modified after Parrish et al. (1987) with treatment of analytical errors following Roddick (1987). Procedural blank levels were generally on the order of 0.5–1.0 pg for Pb. All other cited ages represent U–Pb determinations on zircon grains.

Results

Main lithological units

Volcanic rocks of the Dubuisson Formation

Mafic volcanic rocks constitute the main unit of the outcrop zone. This unit is strongly strained but displays locally preserved pillow shapes (Fig. 3A) that have a southwest younging direction, compatible with regional geological maps (Pilote 2013; Pilote et al. 2015a). Fine-grained basaltic andesitic units contain varying proportions of plagioclase, chlorite, epidote, and biotite. The presence of 120° triple-point junctions within the feldspar- and quartz-rich matrix is typical of recrystallization related to metamorphism. Furthermore, a tholeiitic affinity is inferred from the trace element concentrations and flat REE profiles ($\text{La/Yb} = 1$) (Figs. 4A, 4B, 4C).

A 10 m thick ultramafic unit is exposed in the eastern outcrop and consists of a fine-grained assemblage of feldspath, actinolite, epidote, and chlorite. The presence of amphibole and biotite porphyroblasts reflects alteration and an upper greenschist metamorphic grade. This unit displays a tholeiitic affinity and can be classified as a komatiitic basalt in the Jensen (1976) ternary plot (Fig. 4D). The flat REE pattern ($\text{La/Yb} = 1$), slight depletion in LREE ($\text{La/Eu} = 0.6$), and relatively high concentrations of Ni (600–700 ppm) and Cr (1700–2000 ppm) are compatible with this ultramafic composition.

Felsic volcanoclastic rocks of the Héva Formation

Felsic volcanoclastic units are present in the oriental portion of the eastern outcrop. These rocks are strongly deformed and are interpreted as a volcanoclastic deposit by Pilote (2013), mainly based on visible elongated felsic fragments (Fig. 3B) considered as lapilli following the nomenclature of

171 Fisher (1961). A few 30–50 cm thick feldspar porphyry dykes occur in this unit. The volcanoclastic
172 unit displays a distinct calc-alkaline affinity (Fig. 4B, 4C) and falls within the rhyolitic to
173 rhyodacitic field in the Winchester and Floyd (1977) discrimination diagram (Fig. 4E). The
174 fractionated REE pattern ($\text{La/Yb} = 4$) and the negative Eu anomaly are similar to other felsic units of
175 the Héva Formation (Pilote 2013).

176 ***Dyke swarm***

177 The study area is typified by an exceptional array of thin intermediate dykes (5–40 cm) spaced 2–5
178 m apart. This dyke swarm is present exclusively within the mafic assemblage of the Dubuisson
179 Formation (Fig. 2), and no such dykes were observed in the adjacent Héva Formation. The dykes are
180 fine- to medium-grained, and several contain biotite and amphibole macrocrysts within a matrix of
181 feldspar, quartz, chlorite, and epidote—features compatible with lamprophyre facies, as defined by
182 Mitchell (1994). Mathieu et al. (2018) have described some of these dykes in greater detail.

183 The swarm is composed of multiple families of dykes of varying composition (Fig. 5). For the
184 purpose of this study, these rocks are referred to as “diorite”, reflecting their overall composition.
185 Three main diorite types can be distinguished in the field by their colour (Fig. 3C, Table 3). Type 1
186 is grey, Type 2 is darker, and Type 3 is greenish-grey (Fig. 3C). The amounts of biotite and
187 amphibole vary within a given dyke type and are not discriminant indicators of type (Table 3).
188 Biotite is also abundant in the host volcanic units and is therefore associated with alteration and
189 regional metamorphism. Some amphibole crystals (actinolite) formed during regional upper
190 greenschist metamorphism, but others (hornblende) occur as primary crystals (Mathieu et al. 2018).
191 A relative chronology can be established in the field between the various types of dykes. Type 1

192 dykes are the oldest and are clearly cross-cut by types 2 and 3 (Fig. 3C). As Type 3 dykes are less
193 abundant in the study area, there can be no clear relationship established between types 2 and 3.

194 Using multi-element REE diagrams (Fig. 5A), Type 1 dykes display a La/Sm ratio of 3 and a La/Yb
195 ratio of 16. For Type 2 dykes, the La/Sm ratio is very similar to that of Type 1 dykes, but the general
196 pattern is less fractionated having a La/Yb ratio of 9. Type 3 dykes are the most distinctive with a
197 La/Yb ratio of 48. Two samples, classified as Type 2B, display a hybrid behaviour between types 2
198 and 3 with La/Sm and La/Yb values of 5.3 and 21, respectively (Table 3 and Fig. 5D). On a Y/Zr
199 binary diagram (Fig. 5B), the three dyke types separate into three distinct zones predominantly due
200 to their Y concentrations. Types 1 and 2 plot within the transitional fields (MacLean and Barrett
201 1993), whereas Type 3 plots in the calc-alkaline field (Fig. 5B). However, all dyke types fall in the
202 calc-alkaline field on the Zr/Y versus Th/Yb diagram (Fig. 5C) (Ross and Bédard 2009), but the
203 three types can still be distinguished from each other. On a Zr/TiO₂ versus Nb/Y diagram (Fig. 5E;
204 Winchester and Floyd 1977), most dykes of all three types fall within the quartz diorite field.

205 *U–Pb geochronology*

206 Two U–Pb ages on zircon grains were obtained from the study site (Fig. 6, Table 2). Sample PP-
207 2012-4001-A, a rhyolitic tuff of the Héva Formation, was collected from the eastern outcrop (Fig.
208 2). The sample has high-quality, stubby prismatic to elongate euhedral zircon grains, 200–500 µm in
209 size. The weighted mean of the ²⁰⁷Pb/²⁰⁶Pb ages of five concordant to near-concordant analyses
210 gives an age of 2698.2 ± 0.8 Ma (MSWD = 1.1), interpreted to represent the crystallization age of
211 the rhyolite. One analysis (B16–1; Table 2) is slightly discordant and is not included in the age
212 calculation.

213 Sample PP-2012-RMS (z10755) was collected from a Type 1 dyke (Fig. 2). The sample contains
214 slightly rounded, subhedral to euhedral zircon grains that are 150–400 μm in size. Six single-grain
215 analyses yielded a range of ages; the ages are interpreted as reflecting inheritance in this dyke (ca.
216 2710 Ma, 2700 Ma). A weighted mean of the youngest cluster of concordant to near-concordant
217 analyses is calculated as 2694.3 ± 2.5 Ma (MSWD = 1.3; $n = 3$), interpreted as representing the
218 maximum age of the dyke (Table 2).

219 ***Deformation***

220 The study area is characterized by a high degree of ductile deformation (RHFZ) manifested by a
221 well-developed “principal schistosity” S_p (Fig. 7). This southeast-trending penetrative fabric
222 displays a dip generally to the southwest (mean plane = $132^\circ/79^\circ$; Fig. 8a) that is representative of
223 the attitude in the Malartic region along the major fault zones (Bedeaux et al. 2017). The S_p
224 corresponds to the preferential orientation of all objects; i.e. pillows, fragments, and minerals within
225 the different rock assemblages. A slight virgation is observed on the western outcrop from 160° to
226 100° and is attributed to a wrapping of the units around a tonalitic stock in between the two outcrop
227 zones (Figs. 2 and 7). Neumayer et al. (2000) and Daigneault et al. (2002) attributed this specific
228 variation to the effect of late folding.

229 The S_p fabric is developed to a varying degree and permits the observation of primary features, such
230 as pillows in volcanic units locally. However, it becomes a mylonitic foliation in other areas. To
231 visualize the variation of this S_p development, a field PAI, described in the methodology, was
232 applied (Table 1; Daigneault 1996). The purpose of this index is not to make a direct assessment of
233 strain intensity, as strain markers are relatively rare in this volcanic context. However, this field
234 index is useful for describing fabric variability, which can be attributed to finite strain variation

related to the Sp event, especially within the same rock type. The PAI was applied to a homogeneous surface of 1–5 m² where the structural measurements were then taken (Figs. 7, 8a).

The application of the PAI produces a deformation distribution showing several anastomosed deformation corridors defined using the Sp trajectories and generalised to include stations with a PAI >3 (Fig. 7). Several of these corridors are centred on the ubiquitous quartz veins, discussed below. However, structural anisotropy culminates to the east at the contact between the mafic volcanic units and the felsic volcanoclastic unit. The latter displays a strong mylonitic fabric.

Comparisons between the different degrees of planar anisotropy were made using thin sections of the host volcanic units. Sections parallel to the stretching lineations (XZ plane) were used to illustrate the textural variation (Fig. 9). Within the less deformed mafic volcanic unit (PAI = 1), the foliation is defined by a slight preferential orientation of micaceous minerals, whereas in a severely deformed example (PAI = 5), all minerals are strongly oriented with an aphanitic matrix due to recrystallization. Secondary cleavages and folds appear in the most foliated rocks (Figs. 9C, 9D).

Stretching lineations

Stretching lineations (Ls) represent the X axis of the XY plane associated with the Sp fabric. This lineation is expressed by the preferential elongation of minerals and crystals in the deformed rocks. Field measurements of Ls are quite challenging due to the overprinting of multiple crenulation lineations on the Sp plane. Oriented samples were carefully examined using systematic cuts and thin sections to delineate and measure the pitch of the lineation on the Sp plane. Most Ls plunge toward the southeast with variations from moderate to shallow (Fig. 8b). Two examples of contrasting Ls attitude, one dip-slip (plunge 80°) and the other strike-slip (plunge 05°), are shown in thin sections that are perpendicular (YZ plane) and parallel (XZ plane) to the stretching lineations (Fig. 10). Dip-

257 slip lineations are observed generally in rocks having a moderate anisotropy index, whereas strike-
258 slip lineations are prevalent in high anisotropic deformation corridors and especially along the
259 mafic-felsic contact on the eastern outcrop (Fig. 7). A variation in the Ls from dip-slip to strike-slip
260 is observed in all dykes, as well as in the volcanic units. This feature demonstrates a common
261 history of dykes and volcanic units precluding a late introduction with respect to the deformation.

262 *Crenulation cleavages*

263 Two main families of crenulation cleavages are observed in the study area with a principal cleavage
264 trending northeast ($060^{\circ}/80^{\circ}$) (Figs. 3D and 7). The main northeast-trending family is associated
265 locally with chevron folds mostly in the high-strain zone located at the interface between the two
266 group units. A more discrete north-northeast-trending cleavage is recognized locally but lacks any
267 associated folds. No clear relationship can be established between the two cleavages as both are only
268 observed locally. A northwest-trending, horizontal shortening direction can be estimated from the
269 axial traces of the secondary folds (Fs).

270 *Gold-bearing veins*

271 Gold-bearing veins are 30–70 cm thick and form a set of subparallel veins that occur exclusively in
272 the mafic volcanic unit of the Dubuisson Formation. They are exposed mostly on the western
273 outcrop (Figs. 2 and 11). The veins exhibit a mean spacing of 5–10 m and individual veins can be
274 followed for over 100–150 m. Veins are subparallel, in plan view, with the main Sp fabric.
275 However, they display a shallower dip (60° southwest) with respect to the general Sp dip (75° – 80°).
276 Veins are dominated by quartz (70%–90%) with minor calcite (10%–20%), barite (1%–10%), and,
277 locally, biotite, chlorite, and albite. They are banded, and their margins are commonly composed of
278 a milky quartz, whereas their cores typically have smoky quartz associated with sulphides, barite,

and gold (Figs. 11, 12). In thin sections, quartz is generally granoblastic with 120° triple-point junctions. Despite deformation affecting all veins, some thicker veins display bands with preserved primary features that include the intergrowth of idiomorphic cm-scale quartz crystals interpreted as comb textures, well-formed bipyramidal quartz crystals, and druse cavities (Fig. 11F). Collectively, they represent features compatible with open-space filling fractures.

Sulphide phases may represent 1%–15% of the vein content and are made of pyrite, chalcopyrite, galena, and minor amounts of sphalerite. Some visible gold grains are observed locally. Microprobe analysis of a selection of mineralized samples (Table 4) reveals the presence of silver and bismuth telluride spatially associated with visible gold. A gold concentration of 1300 ppm was measured within the bismuth telluride and 42 800 ppm within the silver telluride. Analysed gold grains contain between 21% and 36% silver, corresponding to the electrum composition.

Another family of quartz veins occurs predominantly in the Héva Formation on the eastern outcrop. These veins are barren with only minor amounts of tourmaline and are oblique with respect to the Sp fabric.

Alteration

A 10 to 30 cm thick marginal alteration zone is recognized along gold-bearing veins (Fig. 11C). This alteration is characterized by the presence of 15%–30% biotite. A biotite stockwork is also observed in the ultramafic rocks that are located on the eastern outcrop near the contact with the felsic volcanoclastic units of the Héva Formation. In all cases, the biotite is destabilized by chlorite related to retrograde metamorphism. A mass balance calculation was performed on these alteration zones using the method of Trépanier et al. (2015). Results for the altered mafic volcanic rocks (Table 5) confirm a significant gain in K₂O (50%–100% or 1–4.5 g) and losses in FeO, CaO, MgO,

301 Na₂O, and SiO₂. Barium was also enriched significantly in these alteration zones. In addition to the
302 potassic alteration observed along the vein margins and near the felsic-mafic contact, the study area
303 is characterized by a pervasive carbonatization. Disseminated calcite (1%–10%) is present in all
304 rocks. Calcite quartz veinlets are found heterogeneously all around the outcrop zone.

305 ***Vein deformation***

306 All veins display a strong effect from RHFZ-related ductile deformation. Pinch and swell veins are
307 common, and 30–50 cm long elongated boudins, isolated from one another, display a lenticular
308 shape (Figs. 11A, 11B, 11C) and have boudin axes that are typically subvertical. The geometry of
309 the veins differs between the eastern and western outcrops. In the western outcrop, veins are mostly
310 subparallel to the Sp fabric becoming folded at the outcrop’s easternmost extremity (see inset in Fig.
311 2) and in the eastern outcrop. Fold hinges are locally observed in low-strain zones. They are open to
312 locally tight and have axes plunging at 40° to the southeast. The overall vein fold geometry
313 describes an S-shape pattern that terminates at the contact with the felsic unit. In the hinge zone, the
314 veins are cross-cut by the axial-planar Sp fabric (Fig. 11D, E). It is significant that a single vein
315 followed across the outcrop exhibits the combined effects of boudinage and folding along strike.

316 **Discussion**

317 As the major features in the study area result from or are modified by the main deformation events,
318 this section will first discuss the characteristics and impacts of the RHFZ to place all described
319 elements within a structural framework.

Rivière-Héva Fault Zone

The km-scale ductile deformation corridor associated with the RHFZ (Fig. 1) has a southeastern trend in the study area, a trend representative of the Malartic segment. However, the RHFZ has an overall easterly trend in the SVZ, as other major fault zones in the Abitibi Subprovince (Fig. 1). The RHFZ is associated with a high-strain corridor manifested by the presence of a strongly developed Sp fabric. Bedeaux et al. (2017) demonstrated that the Sp fabric is generally east-trending between the major southeast-trending faults in this portion of the SVZ, but it becomes subparallel inside the fault zones. The east-trending attitude of the Sp fabric is compatible with a north–south horizontal shortening that is documented for the entire Abitibi Subprovince (Dimroth et al. 1982; Chown et al. 1992; Daigneault et al. 2002). The same shortening direction generates a dextral strike-slip component along the southeast-trending faults (Daigneault et al. 2002; Bedeaux et al. 2017). The synthesis of Sp fabric observed in the study area (mean plane = $132^{\circ}/79^{\circ}$) favours a general southwestern dip for the RHFZ.

Detailed mapping in this area (Pilote 2013) established that all subvertical units are south-facing, in agreement with previous studies (Imreh 1984; Dimroth et al. 1982). The main displacement plane corresponding to the Rivière-Héva Fault, *sensu stricto*, is the contact between the Héva (northeast) and the Dubuisson (southwest) formations (Fig. 1). The presence of the 2708 Ma Dubuisson Formation stratigraphically above the 2700 Ma Héva Formation illustrates an inversion of the stratigraphic sequence. This inversion can be explained by RHFZ movement acting as a reverse or thrust fault.

The SVZ of the Abitibi greenstone belt is characterized by several longitudinal fault zones separating the main lithological groups. Notwithstanding the CLLFZ, and based on regional- (Pilote

2013; Pilote et al. 2015a) and local-scale (study area) patterns, the RHFZ is considered as one of the most significant faults in this portion of the Abitibi belt. As such, this fault has implications for understanding the evolution of the Abitibi belt. The RHFZ may be compatible with a large thrust system disrupting the SVZ stratigraphic sequence. Other faults within the SVZ, such as the Parfouru and La Pause faults, which host the sedimentary Kewagama Group, should be considered as candidates for producing such inversions.

Variation of the stretching lineations

Stretching lineations display significant variations within the RHFZ. The Ls are dip-slip in moderately deformed rocks but develop a moderate to shallow, southeasterly plunge within higher-strain zones. This variation in the attitude of the Ls is interpreted as a transition from a coaxial-dominated regime to a non-coaxial one along the RHFZ. Even if no clear shear sense indicators for this late event are observed in the two outcrop zones, the relationship to the east-trending Sp fabric outside the RHFZ favours a dextral strike-slip component. The variation of the stretching lineation from vertical to subhorizontal can be explained by a single event of oblique transpression or by two events with the first dominated by shortening and a reverse movement along the RHFZ followed by the second event, a strike-slip movement. Oblique transpression in triclinic high-strain zones can produce such variations of the Ls (Lin et al. 1998; Jiang and Williams 1998). A dextral transpression was also proposed for the Malartic area (Daigneault et al. 2002). For the RHFZ, this event is associated with a significant south-side up reverse component to explain the abnormal contact between the Dubuisson and Héva formations. The transition from dip-slip to strike-slip movements along the CLLFZ has been previously recognized and has been interpreted as either one progressive event (Robert 1989) or two different events or increments (Robert 1993; Wilkinson et al. 1999; Daigneault et al. 2002; Bedeaux et al. 2017). The same argument can be made for the observed

365 portion of the RHFZ. Even if a single event of dextral transpression cannot be ruled out to explain
366 the Ls variation, a progressive evolution from a dip-slip dominated event to a dextral strike-slip
367 event will also produce this variation. The displacement along the RHFZ, resulting in the
368 juxtaposition of the Dubuisson Formation with the Héva Formation, must be of a significant
369 magnitude and is therefore more compatible with a reverse movement. The presence of the
370 crenulation cleavage having a northeast strike, and the late Fs folds also support a two-phase
371 evolution.

372 The relative timing of the ductile deformation is an important issue in this portion of the Abitibi
373 greenstone belt. It has been previously described in terms of two or three events (Dimroth et al.
374 1983; Robert 1989; Desrochers et al. 1993; Neumayr et al. 2000; Bedeaux et al. 2017): (i) an initial
375 event that is not always observed followed by (ii) regional deformation responsible for the
376 development of the regional schistosity (Sp) generally attributed to a north–south shortening, and
377 finally (iii) a late dextral strike-slip event recorded mostly in longitudinal fault zones. As noted by
378 Bedeaux et al. (2018), all ductile fabrics observed in the SVZ are also recorded in the sedimentary
379 Timiskaming Group that has therefore registered both the main event of north–south shortening and
380 the late event of dextral strike-slip. If some pre-Timiskaming deformation occurred prior to the
381 regional north–south shortening event, it could be ascribed to a synvolcanic fault, tilting, or a local
382 deformational event without any specific fabric. Therefore, the regional deformation event can be
383 placed as syn- to post-Timiskaming sedimentation and, as such, is younger than 2676 Ma (Pilote et
384 al. 2015a).

385 A second scenario can also reconcile the observations made in the study area. Reverse movement
386 along the Rivière-Héva Fault can occur at a pre-ductile deformation stage of thrust or reverse

387 faulting within a more brittle regime. In this scenario, the ductile deformation associated with the
388 high-strain zone and the variation of stretching lineation can be a late overprint within a
389 transpressive regime. This interpretation was recently proposed for the CLLFZ (Bedeaux et al.
390 2018) to explain the emplacement of the Piché Structural Complex along the fault prior to the
391 imprint of ductile deformation considered to be post-Timiskaming.

392 ***Diorite dykes***

393 Three main dioritic dyke types having lamprophyric facies were identified based on their colour and
394 chemical composition. Type 1 diorite is the oldest dyke family and is cross-cut by types 2 and 3.
395 These dykes have a radiometric U–Pb age of 2694.3 ± 2.5 Ma. Even if types 2 and 3 are younger,
396 their similarities (in terms of attitude, thickness, and density, combined with their occurrence in the
397 RHFZ) are a notable feature favouring a relatively close temporal framework. Their presence within
398 the Dubuisson Formation and absence within the Héva Formation—on the other side of the fault—
399 indicates that movement along the fault occurred after their introduction. Otherwise, these dykes
400 would be present within both formations. The same deduction can be made for the feldspar porphyry
401 dykes found only within the Héva Formation.

402 ***Relationship between dykes, veins, and deformation***

403 The combined presence of dykes, gold-bearing quartz veins, and shear zones is a classic signature of
404 orogenic gold systems, such as those in the Val-d’Or area (Robert and Brown 1986). The study area
405 has similar associations, and this may explain why the Malartic Lakeshore gold veins have been
406 previously classified as an orogenic system (Neumayr et al. 2000). However, the combination of
407 these elements at the Malartic Lakeshore showing occurs in an unusual chronological order. First,
408 veins are strongly affected by deformation associated with the Sp fabric and exhibit boudinage,

recrystallization, and local folding. Second, all three dyke types present in the Dubuisson Formation are affected by both the Sp fabric and Ls, and all dykes clearly cross-cut the veins. This critical relationship is particularly evident in the hinge zone of the folded veins in the eastern portion of the outcrops of the study area (see the inset in Fig. 2). Although veins are northeast-trending in the hinge zone, the dykes continue their course with a southeasterly trend (Fig. 2). These relationships suggest an early vein emplacement with respect to peak deformation. This cross-cutting relationship by the 2694.3 ± 2.5 Ma dyke places vein formation before this event and just after the 2698.2 ± 0.8 Ma emplacement of the Héva Formation. This situation is quite different from the one expected for orogenic gold veins, which are generally interpreted as being later events during regional deformation (Groves et al. 1998; Robert and Brown 1986) and being syn- to post-Timiskaming in age (Robert 2001, Pilote et al. 2015a and 2015b). In the present context, the gold-bearing quartz veins are 25 Ma older than expected if the veins were associated with an orogenic system.

Nature of gold quartz veins and alteration

The paragenesis of the veins and their sulphide association of pyrite, chalcopyrite, sphalerite, and galena, combined with the occurrence of barite (Fig. 12), Bi and Ag tellurides, and K-alteration are features more compatible with a magmatic hydrothermal system rather than with an orogenic system. Moreover, the veins and alteration zones lack iron carbonate, a feature typical of orogenic systems. Another aspect is the open-space filling texture preserved in some veins. This is more compatible with a shallower emplacement (Vearncombe 1993; Hedenquist et al. 2000) than would be expected for an orogenic system.

The primary vein architecture was strongly modified by deformation, thereby preventing any conclusive interpretations of former geometry. The S-type folds present in the eastern portion of the

study area suggest that the veins had an oblique attitude prior to regional shortening. This attitude is different from that of the volcanic units and dykes, both of which having no such folds. The presence of folds and boudins along the strike of the same vein can be explained by a progressive deformation where the veins are in the shortening sector at the beginning of the deformation and pass, via rotation, into the extensional sector. This transition from shortening to extension for the veins can be done through a progressive non-coaxial transpression or via an evolution from a coaxial-dominated event to a non-coaxial-dominated event. The north to northeast orientation of veins in the hinge zone of the folds (eastern portion of the outcrop zone) can be used as a representative attitude of veins prior to folding.

Early “pre-regional deformation” gold systems are also documented in the Val-d’Or area in many deposits such as the Norlartic mine (Couture et al. 1994). As the interpretations presented in this study support an early gold vein system that was strongly modified by regional deformation, the specific context of vein formation remains speculative and is beyond the scope of this study. Different models could be applicable including magmatic hydrothermal fluids that form volcanogenic gold veins (Gaboury and Daigneault 1999), intrusion-related (IRGS) or porphyry-type gold veins, and epithermal-type veins. The high concentrations of base metals (Cu, Pb) may be compatible with a neutral epithermal context (Tosdal et al. 2009; Sillitoe 1993; Hedenquist et al. 2000; Garde et al. 2012), whereas metallic association (Au, Ag, Te, Bi), and vein paragenesis (quartz, calcite, barite, albite) combined with potassic alteration can be integrated into an IRGS system (Sillitoe and Thompson 1998). The presence of the nearby 2690.7 ± 6.3 Ma tonalitic stock could not be ruled out as the magmatic source related to the gold-bearing quartz vein system. Given the error margin, the age of this stock could even correspond to the introduction of the dyke families.

Evolutionary model

A three-stage model is proposed to constrain the formation of gold-bearing veins within a litho-structural framework (Fig. 13). Stage 1 begins with already subvertical units that could be attributed either to a first deformational event, a tilted sequence acquired during a synvolcanic stage along synvolcanic faults, or a combination of both processes. Gold-bearing quartz veins are emplaced into the volcanic rocks of the 2708 Ma Dubuisson Formation at a high angle with the shortening direction and are estimated to be northeast-trending (Fig. 13a). The angular relationship between dykes and veins is locally preserved in fold hinges (Fig. 11d). The various dyke families were then introduced around 2695 Ma initially with Type 1 diorite (dated at 2694 ± 2.5 Ma) followed by types 2 and 3. At this stage, there is no ductile deformation. The dykes cross-cut the veins at a high angle, a pattern only preserved within the hinge zone of folds related to the next stage.

Stage 2 represents a composite event where units of the Dubuisson Formation are transported over units of the Héva Formation (2698 ± 0.8 Ma) along a reverse or thrust fault represented by the RHFZ (Fig. 13b). The fault *sensu stricto* is located along the contact between the two units, but the associated ductile deformation zone is present within a pluridecametric corridor. The Sp fabric and its associated dip-slip Ls can be correlated to the main regional shortening event. High-angle veins are folded in accordance with the shortening component, whereas the dykes are only flattened. This cross-cutting relationship between dykes and veins is well preserved in the hinge zone of the folds.

At Stage 3 (Fig. 13c), the Ls becomes reoriented from an oblique to subhorizontal easterly plunge. This stage could be associated with the late strike-slip and transpressive event recognized along major faults of the Southern Abitibi greenstone belt (Daigneault et al. 2002, Bedeaux et al. 2017). The presence of strongly boudinaged quartz veins with subvertical axes is compatible with this

476 deformation event. The main northeast-trending family of crenulation cleavages associated with the
477 F2 axial traces of the chevron folds is attributed to a northwesterly horizontal shortening direction
478 compatible with a late dextral movement along the RHFZ.

479 **Conclusions**

480 The study area represents the type locality for contact between two major stratigraphic units of the
481 SVZ of the Abitibi greenstone belt. The RHFZ separates the 2708 Ma mafic volcanic unit of the
482 Dubuisson Formation from the 2702–2698 Ma felsic volcanic unit of the Héva Formation. This
483 produces an inversion in the stratigraphic sequence that can be explained by a thrust or reverse
484 movement along the fault zone. A swarm of thin diorite dykes, showing lamprophyric facies, are
485 only present in the Dubuisson Formation. One sample from the Héva Formation yielded a
486 radiometric U–Pb age on zircon grains of 2698.2 ± 0.8 Ma, and a Type 1 diorite dyke was dated at
487 2694.3 ± 2.5 Ma. Gold-bearing quartz veins present in the Dubuisson Formation have an open-space
488 filling texture, contain base metals (Cu, Pb), barite, Bi and Ag tellurides, and display a small-scale
489 K-alteration envelope. All units and veins are overprinted by deformation associated with the major
490 RHFZ, resulting in the development of a southeast-trending Sp foliation containing moderate to
491 shallow southeast-plunging Ls. All diorite dykes clearly cross-cut the veins and are themselves
492 overprinted by the Sp fabric, thus indicating an earlier emplacement with respect to this regional
493 deformation. As the 2694.3 ± 2.5 Ma diorite dyke cross-cuts the veins, this places vein formation
494 near the synvolcanic stage and 20 Ma earlier than the orogenic stage interpreted to be post-
495 Timiskaming and therefore post-2672 Ma. Vein paragenesis and the relative timing with respect to
496 the deformation are compatible with a relatively shallow, early vein system of magmatic
497 hydrothermal origin, possibly IRGS-type, and strongly modified by deformation associated with the
498 RHFZ.

This study demonstrates that pre-ductile deformation or early gold system features can be strongly modified by late deformation especially in major fault zones. Primary features, such as synvolcanic fractures with veins and hydrothermal alteration halos, can favour the development of large deformation corridors overprinting all original features (especially gold-bearing veins) leading to an inaccurate attribution to an orogenic gold system.

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References

- Ayer, J., Amelin, Y., Corfu, F., Kamo, S., Ketchum, J., Kwok, K., and Trowell, N. 2002. Evolution of the southern Abitibi greenstone belt based on U–Pb geochronology: autochthonous volcanic construction followed by plutonism, regional deformation and sedimentation. *Precambrian Research*, **115**: 63–95.
- Backman, O.L. 1936. Geology of Siscoe gold mine. *Canadian Mining Journal*; **57**: 467–475.
- Bedeaux, P., Pilote, P., Daigneault, R., and Rafini, S. 2017. Synthesis of the structural evolution and associated gold mineralization of the Cadillac Fault, Abitibi, Canada. *Ore Geology Reviews* **82**: 49–69.
- Bedeaux, P., Mathieu, L., Daigneault, R., and Pilote, P. 2018. Origin of the Piché Structural Complex and implications for the early evolution of the Archean crustal-scale Cadillac-Larder Lake Fault Zone, Canada. *Canadian Journal of Earth Sciences*. Accepted for publication.
- Bousquet, D., and Carrier, A. 2008. Travaux de décapage sur l'indice « RLM », propriété Malartic Lake Shore, Corporation Minière Golden Share. Ministère de l'Énergie et des Ressources naturelles, Québec; File: Travaux Statutaires, GM 64349, 69 p.

- 526 Bousquet, D., and Carrier, A. 2009a. Programme de forage 2008 sur l'indice aurifère « RLM », propriété
 527 Malartic Lake Shore, Corporation Minière Golden Share. Ministère de l'Énergie et des Ressources naturelles,
 528 Québec; File: Travaux Statutaires, GM 64938, 932 p.
- 529 Bousquet, D., and Carrier, A. 2009b. Programme de forage 2009 sur l'indice aurifère « RLM », propriété
 530 Malartic Lake Shore, Corporation Minière Golden Share. Ministère de l'Énergie et des Ressources naturelles,
 531 Québec; File: Travaux Statutaires, GM 64939, 376 p.
- 532 Card, K.D. 1990. A review of the Superior Province of the Canadian Shield, a product of Archean accretion.
 533 *Precambrian Research*, **48**: 99–156.
- 534 Chown, E., Daigneault, R., Mueller, W., and Mortensen, J. 1992. Tectonic evolution of the northern volcanic
 535 zone, Abitibi belt, Quebec. *Canadian Journal of Earth Sciences*, **29**: 2211–2225.
- 536 Couture, J.-F., Pilote, P., Machado, N., and Desrochers, J.-P. 1994. Timing of gold mineralization in the Val-
 537 d'Or district, southern Abitibi belt: evidence for two distinct mineralizing events. *Economic Geology*, **89**:
 538 1542–1551.
- 539 Daigneault, R. 1996. Couloirs de déformation de la Sous-province de l'Abitibi. Ministère des Ressources
 540 naturelles, Québec; Report MB 96–33, 132 p.
- 541 Daigneault, R., Mueller, W.U., and Chown, E.H. 2002. Oblique Archean subduction: accretion and
 542 exhumation of an oceanic arc during dextral transpression, Southern Volcanic Zone, Abitibi Subprovince
 543 Canada. *Precambrian Research*, **115**: 261–90.
- 544 Daigneault, R., Mueller, W.U., and Chown, E.H. 2004. Abitibi greenstone belt plate tectonics: the
 545 diachronous history of arc development, accretion and collision. *In* *The Precambrian Earth: Tempos and*
 546 *events. Edited by* P.G. Eriksson, W. Altermann, D.R. Nelson, W.U. Mueller, and O. Catuneanu,
 547 *Developments in Precambrian Geology*, **12**: 88–102.
- 548 David, J., Pilote, P., Hammouche, H., Leclerc, F., and Talla Takam, F. 2018. Datations U–Pb dans la
 549 Province du Supérieur effectuées au GEOTOP en 2013-2014. Ministère de l'Énergie et des Ressources
 550 naturelles, Québec; RP 2017-03.
- 551 Davis, D.W., 2002. U-Pb geochronology of Archean metasedimentary rocks in the Pontiac and Abitibi
 552 subprovinces, Quebec, constraints on timing, provenance and regional tectonics. *Precambrian Research*, **115**:
 553 97–117.
- 554 Desrochers, J.P., Hubert, C., Ludden, J.N., and Pilote, P. 1993. Accretion of Archean oceanic plateau
 555 fragments in the Abitibi greenstone belt, Canada. *Geology*, **21**: 451–454.
- 556 De Souza, S., Dubé, B., McNicoll, V.J., Mercier-Langevin, P., Creaser, R.A., and Kjarsgaard, I.M. 2015.
 557 Geology, hydrothermal alteration, and genesis of the world-class Canadian Malartic stockwork-disseminated
 558 Archean gold deposit, Abitibi, Quebec. *In* *Targeted Geoscience Initiative 4: Contributions to the*
 559 *understanding of Precambrian lode gold deposits and implications for exploration. Edited by* B. Dubé, and P.
 560 Mercier-Langevin. Geological Survey of Canada; Open File 7852, pp. 113–26.
- 561 Dimroth, E., Imreh, L., Rocheleau, M., and Goulet, N. 1982. Evolution of the south-central part of the
 562 Archean Abitibi Belt, Quebec. Part I: Stratigraphy and paleogeographic model. *Canadian Journal of Earth*
 563 *Sciences*, **19**: 1729–1758.

- Dimroth, E., Imreh, L., Rocheleau, M., and Goulet, N. 1983. Evolution of the south-central segment of the Archean Abitibi Belt, Quebec. Part 11: Tectonic evolution and geomechanical model. *Canadian Journal of Earth Sciences*, **20**: 1355–1373.
- Dubé B., and Gosselin, P. 2007. Mineral deposits of Canada: a synthesis of major deposit-types, district metallogeny, the evolution of geological provinces, and exploration methods. Geological Association of Canada, Mineral Deposits Division, Special Publication, **5**: 49–73.
- Faure, S. 2015. Relations entre les minéralisations aurifères et les isogrades métamorphiques en Abitibi / Relationship between gold mineralisations and metamorphic isogrades in Abitibi. CONSOREM research project 2013–02. 62 p. http://www.consorem.ca/production_scientifique/2013_03/2013-03_Final.pdf,
- Fisher, R.V. 1961. Proposed classification of volcanoclastic sediments and rocks. *Geological Society of America Bulletin*, **72**:1409–1414.
- Gaboury, D., and Daigneault, R. 1999. Evolution from seafloor-related to sulfide-rich quartz vein-type gold mineralization during deep submarine volcanic construction: The Géant Dormant gold mine, Archean Abitibi belt, Canada. *Economic Geology*, **94**: 1–19.
- Garde, A.A., Whitehouse, M., and Christensen, R. 2012. Mesoarchean epithermal gold mineralization preserved at upper amphibolite-facies grade, Qussuk, southern West Greenland. *Economic Geology*, **107**: 881–908.
- Groves, D.I., Goldfarb, R.J., Gebre-Mariam, M., Hagemann, S.G., and Robert, F. 1998. Orogenic gold deposits: a proposed classification in the context of their crustal distribution and relationship to other gold deposit types. *Ore Geology Reviews*, **13**: 7–27.
- Guay, F., Pilote, P., and Daigneault, R. 2015. Minéralisation aurifère et déformation sur l'indice Malartic Lake Shore, Sous-province de l'Abitibi, Québec Ministère de l'Énergie et des Ressources naturelles, Québec; Report MB 2015–14, 50 p.
- Guay, F., Pilote, P., and Daigneault, R. 2016. Minéralisation aurifère filonienne et déformation de l'indice Rive Du Lac Malartic, Sous-province de l'Abitibi, Québec. Ministère de l'Énergie et des Ressources naturelles, Québec; Report MB 2017–03, 78 p.
- Hedenquist, J.W., Arribas, R.A., and Gonzalez-Urien, E. 2000. Exploration for epithermal gold deposits. *Reviews in Economic Geology*, **13**: 245–277.
- Imreh, L. 1984. Sillon de La Motte-Vassan et son avant-pays méridional: synthèse volcanologique, lithostratigraphique et gîtologique. Ministère de l'Énergie et des Ressources naturelles, Québec; Mémoire MM 82–04, 72 p.
- Jiang, D., and Williams, P.F. 1998, High-strain zones: a unified model. *Journal of Structural Geology*. **20**: 1105–1120.
- Jensen, L.S. 1976. A new cation plot for classifying subalkalic volcanic rocks. Ontario Geological Survey, Miscellaneous Paper 66.
- Lin, S., Jiang, D., and Williams, P.F. 1998. Transpression (or $\pm$ transtension) zones of triclinic symmetry: natural example and theoretical modelling. In *Continental Transpressional and Transtensional Tectonics. Edited by R.E. Holdsworth, R. Strachanand, and J. Dewey*, Geological Society of London, Special

- 602 Publication 135, pp 41–57.
- 603 MacLean, W.H., and Barrett, T.J. 1993. Lithogeochemical techniques using immobile elements. *Journal of*
604 *Geochemical Exploration*, **48**: 109–133.
- 605 Mathieu, L., Bouchard, E., Guay, F., Liénard, A., Pilote, P., and Goutier, J. 2018. Criteria for the recognition
606 of Archean calc-alkaline lamprophyres: examples from the Abitibi Subprovince. *Canadian Journal of Earth*
607 *Sciences*, **55**: 188–205. <https://doi.org/10.1139/cjes-2017-0152>
- 608 Mattinson, J.M. 2005. Zircon U–Pb chemical abrasion (“CA-TIMS”) method; combined annealing and multi-
609 step partial dissolution analysis for improved precision and accuracy of zircon ages: *Chemical Geology*, **220**:
610 47–66.
- 611 McDonough, W.F., and Sun, S.-s. 1995. The composition of the Earth. *Chemical Geology*, **120**: 223–253.
- 612 Mitchell, R.H. 1994. The lamprophyre facies. *Mineralogy and Petrology*, **51**: 137–146.
- 613 Moss, A.E. 1939. The geology of the Siscoe gold mine, province of Québec. Ph.D. thesis, Department of
614 Geological Sciences, McGill University, Montreal, Québec, 159 p.
- 615 Mueller, W., Daigneault, R., Mortensen, J., and Chown, E.H. 1996. Archean terrane docking: Upper crust
616 collision tectonics, Abitibi greenstone belt, Quebec, Canada. *Tectonophysics*, **265**: 127–150.
- 617 Mueller, W., and Donaldson, J.A. 1992. Development of sedimentary basins in the Archean Abitibi belt,
618 Canada: an overview. *Canadian Journal of Earth Sciences*. **29**: 2249–2265.
- 619 Neumayr, P., Hagemann, S.G., and Couture, J.F. 2000. Structural setting, textures, and timing of
620 hydrothermal vein systems in the Val-d’Or camp, Abitibi, Canada: implications for the evolution of
621 transcrustal, second- and third-order fault zones and gold mineralization. *Canadian Journal of Earth Sciences*,
622 **37**: 95–114.
- 623 Parrish, R.R., Roddick, J.C., Loveridge, W.D., and Sullivan, R.W. 1987. Uranium-lead analytical techniques
624 at the Geochronology Laboratory, Geological Survey of Canada. In *Radiogenic age and isotopic studies*,
625 Report: Geological Survey of Canada Paper 87-2: 3–7.
- 626 Pilote, P., Scott, C.R., Mueller, W., Lavoie, S., and Riopel, P. 1999. Géologie des formations de Val-d’Or,
627 Héva et Jacola : nouvelle interprétation du groupe de Malartic. In *Explorer au Québec : le défi de la*
628 *connaissance. Séminaire d’information sur la recherche géologique, programme et résumés, 1999*. Ministère
629 de l’Énergie et des Ressources, Québec; Report DV 99-03, p. 52.
- 630 Pilote, P., Moorhead, J., and Mueller, W. 2000. Partie A, développement d’un arc volcanique, la région de
631 Val-d’Or, ceinture de l’Abitibi : volcanologie physique et évolution métallogénique. Ministère des
632 Ressources naturelles, Québec; Report MB 2000-09, pp. 1–20.
- 633 Pilote, P. 2013. Géologie Malartic, 32D01-NE. Ministère de l’Énergie et des Ressources naturelles, Québec;
634 Map CG-32D01D-2013–01.
- 635 Pilote, P., Daigneault, R., David, J., and McNicoll, V. 2015a. Architecture of the Malartic, Piché and Cadillac
636 groups and the Cadillac Fault, Abitibi: Geological revisions, new dates and interpretations. Abstracts of oral
637 presentations and posters, Québec Mines, 2014. Ministère de l’Énergie et des Ressources naturelles; DV
638 2015-04, p. 37.

- Pilote, P., Lacoste, P., Lafrance, V., David, J., McNicoll, V., and Daigneault, R. 2015*b*. Geology of western Malartic Group: The Lac De Montigny area (NTS 32C04-NW), phase 4). Abstracts of oral presentations and posters, Québec Mines, 2015. Ministère de l'Énergie et des Ressources naturelles, DV 2015-04, p. 58.
- Robert, F., and Brown, A.C. 1986. Archean gold-bearing quartz veins at the Sigma Mine, Abitibi greenstone belt, Quebec; Part I, Geologic relations and formation of the vein system. *Economic Geology*, **81**: 578–592.
- Robert, F. 1989. Internal structure of the Cadillac tectonic zone southeast of Val-d'Or, Abitibi greenstone belt, Quebec. *Canadian Journal of Earth Sciences*, **26**: 2661–2675.
- Robert, F. 1993. The geometrical relationship between the stretching lineation and the movement direction of shear zones: Discussion. *Journal of Structural Geology*, **15**: 239–240.
- Robert, F. 1994. Vein fields in gold districts: the example of Val-d'Or, southeastern Abitibi Subprovince, Québec. In *Current Research 1994-C*; Geological Survey of Canada, pp. 295–302.
- Robert, F. 2001. Syenite-associated disseminated gold deposits in the Abitibi greenstone belt, Canada. *Mineralium Deposita*, **36**: 503–516.
- Robert, F., Poulsen, K.H., Cassidy, K.F., and Hodgson, C.J. 2005. Gold metallogeny of the Superior and Yilgarn cratons. *Economic Geology 100th Anniversary Volume*, pp. 1001–1033.
- Roddick, J.C. 1987. Generalized numerical error analysis with applications to geochronology and thermodynamics: *Geochimica et Cosmochimica Acta*, **51**: 2129–2135.
- Ross, P.-S., and Bédard, J.H. 2009. Magmatic affinity of modern and ancient subalkaline volcanic rocks determined from trace-element discriminant diagrams. *Canadian Journal of Earth Sciences*, **46**: 823–839.
- Sauvé, P., Imreh, L., and Trudel, P. 1993. Description des gîtes d'or de la région de Val-d'Or. Ministère de l'Énergie et des Ressources, Québec; Mémoire MM 91–03, 178 p.
- Scott, C.R., Mueller, W.U., and Pilote, P. 2002. Physical volcanology, stratigraphy and lithogeochemistry of an Archean volcanic arc: evolution from plume-related volcanism to arc rifting of SE Abitibi Greenstone Belt. Val-d'Or, Canada. *Precambrian Research*, **115**: 223–260.
- Sillitoe, R.H. 1993. Epithermal models: genetic types, geometrical controls and shallow features. In *Mineral deposit modeling*; Edited by R.V. Kirkham, W.D. Sinclair, R.I. Thorpe, and J.M. Duke. Geological Association of Canada, Special Paper **40**: 403–417.
- Sillitoe, R.H., and Thompson, J.F. 1998. Intrusion-related vein gold deposits: types, tectono-magmatic settings and difficulties of distinction from orogenic gold deposits. *Resource Geology*, **48**: 237–250.
- Stacey, J.S., and Kramers, J.D. 1975. Approximation of Terrestrial Lead Isotope Evolution by a 2-Stage Model. *Earth and Planetary Science Letters* **26**: 207–221.
- Tosdal, R.M., Dilles, J.H., and Cooke, D.R. 2009. From source to sinks in auriferous magmatic-hydrothermal porphyry and epithermal deposits. *Elements*, **5**: 289–295.
- Trépanier, S., Mathieu, L., and Daigneault, R. 2015. CONSONORM_LG: new normative minerals and alteration indexes for low-grade metamorphic rocks. *Economic Geology*, **110**: 2127–2138.
- Vearncombe, J.R. 1993. Quartz vein morphology and implications for formation depth and classification of

675 Archaean gold-vein deposits. *Ore Geology Reviews*, **8**: 407–424.

676 Wilkinson, L., Cruden, A.R., and Krogh, T.E. 1999. Timing and kinematics of post Timiskaming deformation
677 within the Larder Lake Cadillac deformation zone, southwest Abitibi greenstone belt, Ontario, Canada.
678 *Canadian Journal of Earth Sciences*: **36**, 627–647.

679 Winchester, J.A., and Floyd, P.A. 1977. Geochemical discrimination of different magma series and their
680 differentiation products using immobile elements. *Chemical Geology*, **20**: 325–343.

681 Yergeau, D., Mercier-Langevin, P., Dube, B., Malo, M., and McNicoll, V.J. 2015. The Archean Westwood
682 Au deposit, southern Abitibi; telescoped Au-rich VMS and intrusion-related Au systems. *In* Targeted
683 Geoscience Initiative 4: Contributions to the understanding of Precambrian lode gold deposits and
684 implications for exploration. *Edited by* B. Dubé, and P. Mercier-Langevin. Geological Survey of Canada;
685 Open File 7852, pp. 177–191.

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Figure captions

Fig. 1. Geological map of the southern portion of the Abitibi Subprovince (modified from Québec MERN SIGEOM, Pilote et al. 2014, Daigneault et al. 2002) showing the major fault zones, the Rivière-Héva Fault Zone (RHFZ), and the location of the study area. The inset A in the lower left-hand corner presents the Abitibi Subprovince. NVZ = Northern Volcanic Zone, SVZ = Southern Volcanic Zone, NFZ = Norbénite Fault Zone, CLLFZ = Cadillac–Larder Lake Fault Zone, DPMFZ = Destor–Porcupine–Manneville Fault Zone, LPFZ = La Pause Fault Zone, DFZ = Dumagami Fault Zone, PFZ = Parfouru Fault Zone.

Fig. 2. Detailed geology of the study area divided into two main outcrops. The western outcrop is dominated by the mafic volcanic rocks of the Dubuisson Formation, while the eastern outcrop shows the contact with the felsic unit of the Héva Formation.

Fig. 3. Examples of the main lithological units. (A) Plan view showing mafic volcanic rocks on the western outcrop; some pillows are still recognizable but are flattened (principal schistosity (Sp) fabric shown by the solid white line). (B) Plan view showing strongly strained felsic volcanoclastic rocks, eastern outcrop. (C) Oblique view showing the relationship between diorite dykes. Type 3 (T3-greenish) diorite dyke cross-cutting a Type 1 (T1) dyke (in grey). A thin black Type 2 (T2) dyke is on the left side of the photo. (D) Plan view showing a mafic volcanic rock having Sp fabric and a secondary crenulation cleavage (Cs).

Fig. 4. Geochemical compositions of volcanic rocks. (A) REE patterns normalized to the primitive MORB of McDonough and Sun (1995). (B) Y/Zr diagram with the limits of MacLean and Barrett

711 (1993); TH = tholeiitic, TR = transitional, CA = calc-alkaline. (C) Zr/Y vs Th/Yb diagram of Ross
712 and Bédard (2009), TH = tholeiitic, TR = transitional, CA = calc-alkaline. (D) Jensen (1976) ternary
713 plot. (E) Zr/TiO₂ vs Nb/Y diagram (Winchester and Floyd 1977). See Appendix 1 for details.

714
715 Fig. 5. Geochemical composition of intermediate dykes separated into the three dyke types. (A) REE
716 patterns normalized to primitive MORB according to McDonough and Sun (1995). (B) Y/Zr
717 diagram with the limits of MacLean and Barrett (1993); TH = tholeiitic, TR = transitional, CA =
718 calc-alkaline. (C) Zr/Y vs Th/Yb diagram of Ross and Bédard (2009), TH = tholeiitic, TR =
719 transitional, CA = calc-alkaline. (D) La/Yb vs Zr diagram. (E) Zr/TiO₂ vs Nb/Y diagram from
720 Winchester and Floyd (1977). See Appendix 2 for details.

721
722 Fig. 6. U–Pb Concordia diagrams. (A) Felsic unit of the Héva Formation. (B) Type 1 diorite dyke;
723 see Fig. 2 for sampling locations. Dating uncertainties are presented with 2σ error.

724
725 Fig. 7. Visual estimation of principal schistosity (Sp) development using a planar anisotropy index
726 (PAI) according to criteria described in Table 1. Corridors having higher PAI values are presented in
727 pink. Sp trajectories are shown by blue lines. Stretching lineations display a dominant moderate to
728 shallow southeast-plunging trend.

729
730 Fig. 8. Stereoplots of (A) principal schistosity (Sp) with the mean Sp plane and stretching lineations
731 (Ls) measured on oriented samples. (B) Secondary crenulation cleavages (Cs) having a dominant
732 northeast-trending attitude.

733

Fig. 9. Evolution of deformation as evidenced by an increased planar anisotropy index (PAI): (A–C) Photomicrographs of three samples of volcanic rocks highlighting the evolution of deformation materialized by PAI; (A) PAI = 1/5, (B) PAI = 3/5, and (C) PAI = 5/5. Note the apparition of late microfolds expressed by biotite; (D–F) Photomicrographs of three samples of diorite dyke showing the evolution of deformation materialized by PAI; (D) PAI = 1/5, (E) PAI = 3/5, and (F) PAI = 5/5. Note the apparition of late microfolds expressed by sericite and chlorite.

Fig. 10. Comparisons between thin sections perpendicular (YZ plane) and parallel (XZ plane) to the stretching lineation (Ls) for two different rocks types: (A) and (B) Volcaniclastic units where Ls is directional or subhorizontal (A = YZ plane and B = XZ plane); (C) and (D) mafic volcanic rocks where Ls is dip-parallel or steeply plunging (C = YZ plane and D = XZ plane).

Fig. 11. Photographs of veins observed in the outcrops. (A–D) Early main mineralized veins (Type 1), (E–F) late veins (Type 2). (A) Main gold vein showing boudinage. The host rock is the mafic volcanic unit of the western outcrop (hammer for scale). (B) Isolated boudin of a vein in the western outcrop. (C) Boudinaged vein showing an alteration front in the mafic volcanic rock, western outcrop (compass for scale). (D) and (E) Vein cross-cut at a right angle by a foliated dyke, western outcrop. (F) Druse cavity preserved in the core of the vein.

Fig. 12. Mineralized banded quartz vein showing well-preserved mineralogical zonation composed of quartz (Qz) at the outer margin, quartz-calcite-feldspath (Qz+Ca+Fp) toward the core of the vein, and barite-quartz-calcite (Ba+Qz+Ca) in the core. Note that barite is the white mineral in the image.

757 Fig. 13. Schematic and idealized block model showing the relative chronology for the emplacement
758 of the main groups, veins, dykes, and deformation (see text for details). (A) Stage 1, emplacement of
759 gold quartz veins (red lines) into the volcanic rocks of the Dubuisson Formation (2708 ± 2 Ma)
760 cross-cut by diorite dykes (blue lines) introduced at around 2695 Ma. (B) Stage 2, reverse or thrust
761 movement along RHFZ putting the Dubuisson Formation over the Héva Formation (2698 ± 0.8 Ma)
762 with ductile deformation producing the subvertical Sp fabric and the dip-slip stretching lineation
763 (Ls). (C) Stage 3, late strike-slip component of shearing produces the easterly shallow-plunging
764 stretching lineation (Ls) and associated with the secondary crenulation cleavage (Cs).

765

766 Appendices

767 Appendix 1 – Chemical composition of volcanic rocks

768 Appendix 2 – Chemical composition of diorite dykes

Table 1. Field criteria used to estimate the principal schistosity (S_p) fabric as derived from a planar anisotropy index (PAI), modified from Daigneault (1996).

Index	Criteria
0	<ul style="list-style-type: none"> • No visible fabric (planar or linear) • Perfect preservation of primary rock textures
1	<ul style="list-style-type: none"> • Presence of a planar fabric (cleavage-type), weakly developed, but showing a consistent orientation • Good preservation of primary rock textures
2	<ul style="list-style-type: none"> • Penetrative planar fabric (schistosity-type) • Noticeable preferential orientation of all objects (pillows, fragments, etc.) • Primary rock textures still recognizable
3	<ul style="list-style-type: none"> • Well-developed penetrative planar fabric (schistosity-type) • Strong preferential orientation of all objects (pillows, fragments, etc.) • Primary rock textures locally recognizable but obliterated by deformation
4	<ul style="list-style-type: none"> • Strongly developed penetrative planar fabric (schistosity or mylonitic foliation-type) • Dominance of structural features over primary textures
5	<ul style="list-style-type: none"> • Strongly developed penetrative planar fabric (mylonitic foliation-type) • Complete obliteration of primary features

Table 2. U-Pb ID-TIMS analytical data.

Fraction ¹	Description ²	Isotopic Ratios ⁶													Ages (Ma) ⁸													
		Wt.	U	Pb ³	²⁰⁶ Pb ⁴	Pb ⁵	²⁰⁸ Pb	²⁰⁷ Pb	±1SE	²⁰⁶ Pb	±1SE	Corr. ⁷	²⁰⁷ Pb	±1SE	²⁰⁶ Pb	±2SE	²⁰⁷ Pb	±2SE	²⁰⁷ Pb	±2SE	%							
					²⁰⁴ Pb		²⁰⁶ Pb	²³⁵ U		Abs			²³⁸ U		Abs		Coeff.		²⁰⁶ Pb			Abs	²³⁸ U	±2SE	²³⁵ U	±2SE	²⁰⁶ Pb	Disc
					ug		ppm	ppm		pg			²⁰⁶ Pb		²³⁵ U		Abs		²³⁸ U			Abs	Coeff.	²⁰⁶ Pb	Abs	²³⁸ U	±2SE	²³⁵ U
PP-2012-3126A (Z10754)																												
A16-1 (1;CA)	Co,Clr,Eu,Pr,fln,CA16	1.3	102	57	1721	2.5	0.07	13.2425	0.0190	0.51858	0.00062	0.904	0.18520	0.00011	2693.2	5.3	2697.1	2.7	2700.0	2.0	0.3							
A16-2 (1;CA)	Co,Clr,Eu,Pr,fln,CA16	2.1	81	46	3276	1.7	0.08	13.2806	0.0162	0.51996	0.00051	0.932	0.18525	0.00009	2699.0	4.3	2699.8	2.3	2700.4	1.6	0.1							
A20-1 (1;CA)	Co,Clr,Eu,Pr,rFr,CA20	2.6	100	56	3143	2.7	0.08	13.2626	0.0162	0.51932	0.00051	0.920	0.18522	0.00009	2696.3	4.3	2698.5	2.3	2700.2	1.6	0.2							
A20-2 (1;CA)	Co,Clr,Eu,Pr,CA20	1.6	116	66	2846	2.1	0.09	13.2809	0.0168	0.52004	0.00053	0.925	0.18522	0.00009	2699.4	4.5	2699.8	2.4	2700.2	1.6	0.0							
PP-2012-RMS Parc 1 (Z10755)																												
A10-1 (1;CA)	Co,Clr,Eu,Pr,fln,fFr,CA10	1.2	39	23	1686	0.9	0.13	13.4249	0.0209	0.52244	0.00076	0.891	0.18637	0.00013	2709.5	6.4	2710.0	2.9	2710.4	2.3	0.0							
A10-2 (1;CA)	Co,Clr,Eu,Pr,fln,fFr,CA10	1.5	52	30	2177	0.8	0.11	13.1568	0.0185	0.51674	0.00064	0.915	0.18466	0.00011	2685.3	5.4	2691.0	2.7	2695.2	1.9	0.5							
C10-1 (1;CA)	Co,Clr,Eu,Pr,fFr,CA10	0.8	77	44	1220	1.6	0.09	13.2218	0.0196	0.51942	0.00070	0.899	0.18462	0.00012	2696.7	5.9	2695.6	2.8	2694.8	2.1	-0.1							
C10-2 (1;CA)	Co,Clr,Eu,Pr,fln,fFr,CA10	1.3	39	24	1060	1.6	0.17	13.2829	0.0222	0.52062	0.00077	0.906	0.18504	0.00013	2701.8	6.5	2700.0	3.2	2698.6	2.3	-0.2							
C10-3 (1;CA)	Co,Clr,Eu,Pr,fln,Fr,CA10	1.5	120	68	7413	0.8	0.08	13.1919	0.0161	0.51868	0.00052	0.932	0.18446	0.00008	2693.6	4.4	2693.5	2.3	2693.4	1.5	0.0							
C16-1 (1;CA)	Co,Clr,Eu,Pr,fln,CA16	0.6	87	48	726	2.4	0.04	13.3379	0.0241	0.52204	0.00079	0.819	0.18530	0.00019	2707.8	6.7	2703.9	3.4	2700.9	3.4	-0.3							

Notes:

¹Number in parenthesis refers to the number of zircon grains in the analysis. CA = chemically abraded.
²Fraction descriptions: Co = colourless, Clr = clear, El = elongate, Eu = euhedral, Pr = prismatic, St = stubby prismatic, rFr = Rare Fractures, fFr = Few Fractures, Fr = Fractures, fln = few inclusions, In = inclusions, CA16 = chemically abraded for 16 h.
³Radiogenic Pb
⁴Measured ratio, corrected for spike and fractionation
⁵Total common Pb in analysis corrected for fractionation and spike
⁶Corrected for blank Pb and U and common Pb, errors quoted are 1 sigma absolute; procedural blank values for this study ranged from <0.1- 0.1 pg for U and 0.5-1 pg for Pb; Pb blank isotopic composition is based on the analysis of procedural blanks; corrections for common Pb were made using Stacey and Kramers (1975) compositions
⁷Correlation coefficient
⁸Corrected for blank and common Pb, errors quoted are 2 sigma in Ma

Table 3. Characteristics of diorite dykes. Main mineral constituents: Pg = Plagioclase, Bo = Biotite, Chl = Chlorite, Am = Amphibole, Ep = Epidote. (*mean presented in parentheses)

Type	N	Colour	Thickness (cm)	Pg %	Bo%	Chl%	Am%	Ep%	La/Sm	La/Yb
1	11	Grey	20–40	42–50 (48*)	1–20 (13)	1–35 (19)	1–32 (15)	1–20 (7)	4.0	15.5
2	7	Black	5–10	30–45 (41)	1–25 (11)	3–50 (22)	5–45 (26)	2–10 (7)	3.9	9.2
3	3	Greenish	5	50–65 (58)	10	1–3 (2)	1–35 (13)	2–14 (7)	5.6	48.4

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Table 4. Microprobe analysis of sulphides within gold veins.

Phase		S	Fe	Cu	Pb	Zn	As	Ni	Mn	Co
		% wt	% wt	% wt	% wt	ppm	ppm	ppm	ppm	ppm
Chalcopyrite <i>n</i> = 21	Mean	35.32	30.45	33.1	0.13	185	263	674	209	100
	St. dev.	0.21	2.32	2.5	0.04	247	388	462	126	89
Galene <i>n</i> = 9	Mean	13.72	0.16	0.14	82.7	1360	305	567	268	170
	St. dev.	0.18	0.19	0.15	1.4	962	168	529	81	135
Pyrite <i>n</i> = 23	Mean	53.95	46.64	0.03	0.19	373	274	723	137	199
	St. dev.	0.24	0.40	0.02	0.05	204	440	369	75	96
Pyrrhotite <i>n</i> = 7	Mean	39.26	59.54	0.13	0.15	520	143	1580	366	110
	St. dev.	0.21	0.48	0.16	0.05	240	108	1899	193	-
Sphalerite <i>n</i> = 4	Mean	33.64	6.90	0.99	0.16	562468	347	645	580	130
	St. dev.	0.34	2.90	1.34	0.02	23462	100	106	577	-
Phase		Te	Au	Ag	Bi	Cu	Fe	Zn	As	Se
		%wt	ppm	%wt	ppm	%wt	%wt	%wt	ppm	ppm
Silver telluride <i>n</i> = 9	Mean	34.75	12379	59.40	220	0.15	0.26	0.08	215	323
	St. dev.	2.34	15411	1.95	-	0.08	0.23	0.06	112	146
Bismuth telluride <i>n</i> = 5	Mean	32.12	640	0.97	606898	2.11	1.90	0.09	0	574
	St. dev.	3.04	408	1.50	73441	2.07	2.46	0.10	-	342
Electrum (21% Ag) <i>n</i> = 4	Mean	0.13	731118	21.41	6438	0.11	0.04	0.36	270	290
	St. dev.	0.08	21116	0.86	923	0.10	0.03	-	259	176

Table 5. Mass balance calculations (expressed in grams) for the most altered volcanic rocks using precursor modelling based on Trépanier et al. (2011). V1 = felsic volcanic rocks, V3 = mafic volcanic rocks, V4 = ultramafic volcanic rocks.

Sample	Litho	LOI	CaO	FeOT	K ₂ O	MgO	Na ₂ O	SiO ₂
FG-81	V1	4.82	-3.96	1.43	2.41	0.14	-3.95	5.65
FG-04	V3	7.09	-0.66	0.11	2.58	0.58	0.88	-2.82
FG-09	V3	6.97	-0.88	0.80	2.81	0.29	1.28	-1.81
FG-57	V3	7.87	2.95	-2.45	1.10	-1.97	0.56	6.30
FG-58	V3	6.38	-4.12	1.90	0.94	1.89	0.35	-5.18
FG-83	V3	6.76	0.79	-2.28	0.71	-2.80	0.61	2.24
FG-84	V3	10.46	10.76	2.43	1.30	0.70	0.49	29.79
FG-85	V4	11.97	24.02	13.70	4.50	11.53	-2.56	27.93
FG-86	V4	11.04	13.47	1.42	1.18	-9.75	0.11	7.61
FG-92	V4	9.49	13.41	0.62	1.35	-9.84	0.94	3.58

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Figure 1

Figure 2

Figure 5

Figure_6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 12

Appendix 1 – Chemical composition of volcanic rocks.

Unit	Felsic volc.			Mafic volc.			Gabbro			Ultramafic volc.		
	<i>n</i> = 4		Type ex.	<i>n</i> = 9		Type ex.	<i>n</i> = 2		Type ex.	<i>n</i> = 3		Type ex.
	Avg.	St.dev.	(FG-78)	Avg.	St.dev.	(FG-54)	Avg.	St.dev.	(FG-49)	Avg.	St.dev.	(FG-85)
SiO ₂ (%)	75.01	6.32	79.10	53.50	2.23	51.52	56.13	1.27	57.03	46.47	1.89	44.35
Al ₂ O ₃ (%)	13.75	2.35	12.88	14.62	0.96	14.60	18.05	1.66	19.23	7.57	1.12	8.85
Fe ₂ O ₃ (%)	3.13	2.64	2.37	11.02	1.89	12.50	7.32	0.26	7.14	12.52	1.51	14.27
MgO (%)	1.22	1.18	0.78	5.73	1.34	5.30	3.61	2.94	1.54	10.77	1.56	9.29
MnO (%)	0.03	0.02	0.01	0.21	0.04	0.21	0.11	0.03	0.09	0.33	0.06	0.37
CaO (%)	1.26	1.29	0.15	9.57	3.29	13.34	8.46	0.50	8.11	18.12	0.55	17.83
Na ₂ O (%)	2.20	2.20	0.45	2.79	0.81	1.25	3.98	0.13	3.89	0.65	0.52	0.17
K ₂ O (%)	2.97	1.33	4.13	1.50	1.08	0.74	1.17	1.08	1.93	1.85	0.89	2.87
H ₂ O (%)	0.25	0.30	0.06	0.55	0.19	0.47	0.81	0.05	0.84	0.46	0.04	0.50
P ₂ O ₅ (%)	0.07	0.06	0.02	0.05	0.02	0.04	0.16	0.00	0.16	0.04	0.01	0.06
Cr ₂ O ₃ (%)	0.01	0.00	0.01	0.03	0.02	0.05	0.03	0.02	0.02	0.41	0.36	0.02
LOI (%)	2.60	1.51	2.23	6.50	2.34	2.15	3.67	2.96	1.58	10.83	1.25	11.97
Total (%)	99.91		99.95	99.57		100.02	99.85		99.99	99.20		98.58
Sr (ppm)	110.5	48.8	55.0	429.3	446.7	326.0	275.5	183.1	405.0	144.3	8.3	151.0
Ba (ppm)	509.8	258.8	846.0	454.9	487.1	177.0	650.0	837.2	1242.0	164.3	76.0	252.0
Nb (ppm)	5.6	3.5	6.4	4.2	7.2	4.7	8.4	5.6	4.4	2.8	3.0	1.2
V (ppm)	36.0	41.7	6.0	232.2	19.8	248.0	198.0	46.7	231.0	184.0	16.8	165.0
Co (ppm)	7.3	8.6	5.0	38.9	9.0	48.0	20.5	6.4	16.0	50.3	26.4	20.0
Ga (ppm)	18.0	2.9	18.0	13.0	1.9	15.0	16.5	6.4	21.0	11.0	0.0	11.0
Ge (ppm)	1.4	0.2	1.6	1.7	0.3	2.3	1.8	0.2	1.9	2.4	0.4	2.0
Cs (ppm)	0.6	0.4	1.1	2.7	2.2	1.4	1.2	1.3	2.1	3.5	1.5	5.3
Pb (ppm)	4.6	2.4	7.0	8.6	4.9	9.0	5.5	0.7	5.0	3.5	1.8	5.0
Hf (ppm)	2.2	1.1	1.9	1.0	0.3	0.7	2.3	0.3	2.5	0.9	0.4	1.3
W (ppm)	1.5	0.2	1.4	1.0	0.4	0.9	1.8	0.8	2.3	1.0	0.5	1.6
Rb (ppm)	75.5	30.8	105.0	55.9	42.1	28.0	36.0	35.4	61.0	73.0	33.9	112.0
Y (ppm)	10.8	6.4	7.3	19.9	2.6	21.7	17.3	0.7	17.8	10.6	3.5	14.6
Zr (ppm)	62.5	42.1	40.0	28.9	10.3	25.0	76.0	12.7	85.0	27.0	11.5	40.0
La (ppm)	6.3	3.6	3.4	2.5	0.5	2.4	9.6	0.2	9.7	1.7	0.9	2.7
Ce (ppm)	15.1	7.7	10.5	5.5	1.2	4.8	23.4	2.5	25.2	4.3	2.5	7.2
Pr (ppm)	1.8	1.1	0.8	0.8	0.2	0.7	3.2	0.3	3.5	0.7	0.4	1.1
Nd (ppm)	6.8	4.2	2.7	4.2	1.1	3.7	14.4	1.4	15.4	3.5	1.8	5.6
Sm (ppm)	1.7	1.1	0.7	1.3	0.4	1.0	3.5	0.3	3.8	1.3	0.4	1.8
Eu (ppm)	0.4	0.3	0.1	0.4	0.1	0.4	0.9	0.1	1.0	0.4	0.1	0.5
Gd (ppm)	1.6	1.0	0.9	1.9	0.5	1.7	3.1	0.1	3.2	1.6	0.5	2.2
Tb (ppm)	0.3	0.2	0.2	0.4	0.1	0.4	0.5	0.0	0.5	0.3	0.1	0.4
Dy (ppm)	1.8	1.1	1.2	3.0	0.4	2.9	3.1	0.2	3.3	2.0	0.6	2.7
Ho (ppm)	0.4	0.2	0.3	0.7	0.1	0.8	0.7	0.0	0.7	0.4	0.1	0.6
Er (ppm)	1.2	0.6	0.9	2.5	0.3	2.7	2.1	0.1	2.1	1.2	0.5	1.9
Tm (ppm)	0.2	0.1	0.2	0.4	0.1	0.4	0.3	0.0	0.4	0.2	0.1	0.3
Yb (ppm)	1.4	0.7	1.3	2.8	0.4	3.0	2.3	0.1	2.3	1.2	0.6	1.8
Lu (ppm)	0.2	0.1	0.2	0.4	0.1	0.5	0.4	0.0	0.4	0.2	0.1	0.3
Cu (ppm)	20.5	27.3	4.0	107.4	63.6	110.0	18.0	22.6	34.0	52.7	52.3	37.0
Ni (ppm)	20.8	26.4	8.0	104.0	37.5	149.0	90.0	15.6	101.0	467.7	287.0	138.0
Zn (ppm)	69.8	80.5	47.0	68.4	14.6	63.0	35.5	13.4	26.0	112.0	13.5	99.0
Au (ppb)	7.8	6.8	8.0	40.0	42.2	12.0	1.0	0.0	1.0	16.0	21.8	6.0
Cr (ppm)	25.3	17.7	14.0	92.2	57.4	180.0	106.5	2.1	105.0	1255.7	1060.5	47.0
Sc (ppm)	5.7	4.3	3.5	45.5	8.9	56.0	33.7	1.3	32.7	18.9	3.3	15.1
S (ppm)	5800	7891	5000	4556	3225	7000	1400	1414	2400	3133	1626	3700

Draft

Appendix 2 – Chemical composition of diorite dykes

Dyke	Type 1			Type 2			Type 2b			Type 3		
	n = 8		Type ex.	n = 8		Type ex.	n = 2		Type ex.	n = 4		Type ex.
Element	Avg.	St.dev.	(FG-02)	Avg.	St.dev.	(FG-10)	Avg.	St.dev.	(FG-03)	Avg.	St.dev.	(FG-07)
SiO ₂ (%)	55.63	2.34	57.18	51.47	4.10	53.39	50.64	0.12	50.73	59.18	0.40	58.90
Al ₂ O ₃ (%)	14.66	1.92	15.78	14.18	1.94	14.16	14.68	0.00	14.67	15.13	1.33	14.19
Fe ₂ O ₃ (%)	8.73	0.49	8.54	10.88	2.23	9.75	9.62	0.23	9.79	5.89	0.64	6.35
MgO (%)	6.44	2.85	4.29	7.36	0.97	7.88	5.99	0.56	5.59	4.44	1.62	5.59
MnO (%)	0.16	0.04	0.13	0.19	0.04	0.14	0.17	0.00	0.17	0.11	0.05	0.15
CaO (%)	7.81	2.22	6.59	8.91	2.71	6.53	9.83	0.23	9.99	6.92	0.71	7.42
Na ₂ O (%)	3.77	0.93	4.51	2.97	1.44	4.14	3.94	0.54	3.55	6.25	0.47	5.92
K ₂ O (%)	1.26	0.61	1.92	2.17	1.57	2.77	3.55	0.42	3.85	1.06	0.61	0.63
H ₂ O (%)	0.72	0.12	0.77	0.91	0.13	0.82	0.79	0.02	0.80	0.61	0.10	0.54
P ₂ O ₅ (%)	0.37	0.07	0.32	0.43	0.08	0.41	0.47	0.07	0.52	0.34	0.09	0.28
Cr ₂ O ₃ (%)	0.04	0.04	0.01	0.09	0.06	0.06	0.04	0.00	0.04	0.08	0.11	0.16
LOI (%)	5.68	3.94	1.39	6.26	4.06	1.59	5.75	0.52	6.12	2.19	0.66	1.72
Total (%)	99.59		100.04	99.54		100.04	99.71		99.70	100.04		100.12
Sr (ppm)	345.4	247.5	711.0	250.8	203.7	516.0	490.0	277.2	686.0	473.0	142.8	372.0
Ba (ppm)	259.6	196.1	438.0	422.3	362.6	557.0	660.0	130.1	752.0	161.5	65.8	115.0
Nb (ppm)	4.6	0.9	4.3	5.4	0.9	5.3	8.9	0.4	8.6	6.6	1.3	5.6
V (ppm)	154.9	17.3	157.0	176.0	16.7	188.0	206.5	9.2	200.0	112.0	12.7	121.0
Co (ppm)	28.8	6.2	23.0	37.5	7.9	33.0	30.5	0.7	31.0	20.5	6.4	25.0
Ga (ppm)	17.6	2.4	20.0	16.3	3.8	22.0	17.0	2.8	19.0	18.5	4.9	15.0
Ge (ppm)	1.4	0.5	1.7	1.3	0.3	1.7	1.5	0.0	1.5	1.3	0.3	1.5
Cs (ppm)	2.2	1.4	3.8	3.4	2.7	5.2	7.3	1.0	8.0	1.9	1.2	1.0
Pb (ppm)	7.6	4.8	12.0	6.0	4.1	13.0	13.5	3.5	16.0	8.0	1.4	9.0
Hf (ppm)	3.2	0.4	3.4	3.4	0.5	3.6	3.3	0.1	3.3	3.6	0.5	3.2
W (ppm)	0.9	0.4	0.7	1.0	0.5	0.6	0.7	0.1	0.6	0.5	0.3	0.7
Rb (ppm)	45.8	24.2	71.0	82.5	65.9	100.0	152.0	19.8	166.0	36.5	21.9	21.0
Y (ppm)	16.6	3.1	18.5	23.4	3.5	24.8	27.8	0.6	28.2	12.3	3.3	10.0
Zr (ppm)	113.1	17.2	139.0	116.4	14.2	124.0	134.0	4.2	131.0	134.0	24.0	117.0
La (ppm)	25.6	3.9	25.6	22.5	4.2	25.3	50.3	4.1	53.2	38.2	7.4	33.0
Ce (ppm)	56.1	7.1	55.1	51.4	8.9	56.1	107.5	4.9	111.0	81.5	20.2	67.2
Pr (ppm)	7.4	0.9	7.0	7.1	1.1	7.8	13.6	0.6	14.0	10.7	2.8	8.7
Nd (ppm)	30.5	3.5	29.0	30.5	4.7	33.1	52.5	1.1	53.3	42.2	11.4	34.1
Sm (ppm)	6.3	0.6	6.0	6.7	0.9	7.3	9.4	0.5	9.8	7.2	2.3	5.5
Eu (ppm)	1.6	0.2	1.7	1.6	0.2	1.8	2.4	0.1	2.5	1.8	0.6	1.4
Gd (ppm)	5.0	0.5	4.9	5.7	0.8	6.2	6.9	0.5	7.2	4.8	1.6	3.7
Tb (ppm)	0.6	0.1	0.6	0.8	0.1	0.9	0.9	0.1	0.9	0.5	0.1	0.4
Dy (ppm)	3.4	0.4	3.6	4.5	0.7	5.1	4.7	0.2	4.9	2.5	0.7	2.0
Ho (ppm)	0.6	0.1	0.7	0.9	0.2	1.0	0.9	0.0	0.9	0.4	0.1	0.4
Er (ppm)	1.8	0.3	2.0	2.5	0.4	2.9	2.6	0.1	2.6	1.2	0.4	0.9
Tm (ppm)	0.3	0.0	0.3	0.4	0.1	0.4	0.4	0.0	0.4	0.2	0.0	0.1
Yb (ppm)	1.5	0.3	1.9	2.3	0.4	2.6	2.4	0.1	2.4	1.0	0.2	0.9
Lu (ppm)	0.2	0.1	0.3	0.4	0.1	0.4	0.4	0.0	0.4	0.1	0.0	0.1
Cu (ppm)	17.5	14.2	35.0	32.4	31.0	45.0	78.5	31.8	101.0	35.0	39.6	7.0
Ni (ppm)	61.3	52.4	31.0	61.5	16.5	59.0	39.5	3.5	37.0	60.0	35.4	85.0
Zn (ppm)	79.0	9.0	76.0	83.4	21.1	96.0	78.0	9.9	85.0	67.0	7.1	62.0
Au (ppb)	7.3	9.3	22.0	4.8	7.0	18.0	15.0	19.8	1.0	7.0	2.8	9.0
Cr (ppm)	237.4	255.3	82.0	332.6	118.1	372.0	191.0	84.9	131.0	273.5	368.4	534.0
Sc (ppm)	18.5	4.9	14.7	23.2	3.3	24.1	20.9	4.5	17.7	12.2	4.0	15.0
S (ppm)	838	576	1800	913	1173	600	650	212	800	1200	1131	400