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1 **Relationships between basin architecture, basin closure, and occurrence of**
2 **sulphide-bearing schists: an example from Tampere Schist Belt, Finland**

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10

11 **Abstract**

12 We present field observations from the Palaeoproterozoic volcano-sedimentary Tampere
13 palaeobasin, where the primary structures have been exceptionally well preserved. We
14 use the observations to construct a new tectonic model for the southeastern margin of the
15 Tampere basin during its inversion and subsequent closure. The observed volcano-
16 sedimentary and structural features suggest a change in the local structural style from
17 thick-skinned inversion to thin-skinned thrusting, in order to accommodate the crustal
18 shortening during basin closure. Furthermore, it is suggested that there is a genetic
19 relationship between the interpreted palaeothrusts and the sulphide-bearing schist
20 horizons in the study area. On a more general note, the results infer that presently
21 subvertical mineralised shear zones may have originally been gently dipping, further
22 suggesting that the mineralised fluids may not necessarily have been sourced from great
23 depths (i.e. from deep within the basement).

24

25 **Introduction**

26 Ancient basins archive the early development of plate tectonics, life, and ore-bearing
27 systems. Especially basins involving volcanic-derived materials contain economically

28 important ore and mineral deposits (e.g. Lentz 1998; Weihed *et al.* 2005). The formation of
29 these ores is often related to hydrothermal activity and the migration of mineralising fluids:
30 this can happen both in locally extensional environments (notably the formation of the
31 volcanogenic massive sulphide deposits, 'VMS ores', which most commonly form in
32 volcanic arc settings; e.g. Ohmoto 1996; Allen *et al.* 2002) but mineralising fluids can also
33 move along strike-slip or reverse faults (e.g. Sibson *et al.* 1988; Teagle *et al.* 1990;
34 Piessens *et al.* 2002). Usually the mineralisations formed in compressional settings are
35 related to steep strike-slip or reverse faults where the fluids are sourced from great depths;
36 reported mineralisations related to more gently dipping reverse faults are less common
37 (e.g. Nguyen *et al.* 1998; Piessens *et al.* 2002).

38 The Tampere Schist Belt (TSB) in southern Finland is a c. 1.92-1.88 Ga volcano-
39 sedimentary basin that underwent inversion and closure between c. 1.89-1.88 Ga (Fig. 1;
40 Patchett & Kouvo 1986; Kähkönen *et al.* 1989; Nironen 1989a; Rutland *et al.* 2004). The
41 tectonic setting of the belt varied with time from a microcontinental rift to a volcanic arc
42 system (Lahtinen *et al.*, 2009). The closure of the basin resulted in the formation of a
43 large-scale, predominantly tight to isoclinal syncline with steeply dipping fold axial planes
44 (Figs. 1b & 2; e.g. Kähkönen 1989; Nironen 1989a). Despite the crustal-scale shortening,
45 the primary sedimentary and volcanic structures are generally remarkably well preserved.
46 The TSB, therefore, offers an excellent opportunity to examine the volcano-sedimentary
47 evolution of an ancient basin, and the mechanics of and strain distribution during its
48 subsequent closure. In addition, the TSB is valuable for understanding the evolution of
49 ore-bearing systems within volcano-sedimentary basins as it hosts several mineralised
50 deposits. Probably the best known of these are the Haveri deposit, mined in 1942-1962,
51 and the presently mined Kutemajärvi deposit (Fig. 1b). The Haveri Fm. forms the base of
52 the TSB succession and hosts an orogenic gold deposit (originally a VMS ore that was
53 later remobilised; Nironen 1994; Eilu *et al.* 2003). The Haveri deposit is hosted by schists
54 and is associated with faults and folds (Nironen 1994). Various sulphide-bearing schist
55 horizons and lenses are also contained higher in the local stratigraphy (e.g. Kähkönen &
56 Nironen 1994). The Kutemajärvi deposit is interpreted to be an epithermal ("shallow heat")
57 metamorphic deposit (Eilu *et al.* 2003). It, too, is associated with schists and prominent
58 fault zones (Poutiainen & Grönholm 1996).

59 The aim of this study is to investigate the structural development and the architecture
60 of a part of the TSB in more detail, including the relationships between the volcano-
61 sedimentary sequences, the tectonic structures, and the sulphide-bearing schist horizons.
62 Important insights are gained into understanding the mechanisms of the basin closure and
63 the localisation of the sulphide mineralisation within the basin. The suggested conceptual
64 model implies that early, gently dipping thrusts acted as both channels and traps for the
65 mineralising fluids that possibly sourced either locally or from relatively shallow depths
66 from the base of the basin infill. The continued compression caused a subsequent rotation
67 of the thrusts into their present subvertical position. Although the studied rocks have been
68 somewhat metamorphosed, we have in general dropped out the prefix “meta” from rock
69 type names.

70

71 **Geological setting**

72 The continental crust of the central Fennoscandian Shield is mainly composed of 1.95-
73 1.80 Ga rocks, forming the so-called Svecofennian domain (Fig. 1a; Gaál & Gorbatshev
74 1987). The orogenic evolution that produced this orogen was considered as the single
75 Svecofennian orogeny by Gaál & Gorbatshev (1987), while Lahtinen *et al.* (2005) divided
76 it into three separate orogenies: the Fennian arc-accretionary orogeny at c. 2.0-1.85 Ga,
77 the Andean-type Svecobaltic orogeny at c. 1.84-1.80 Ga, and the Nordic orogeny with a
78 continent-continent collision at 1.82-1.79 Ga (Fig. 1a; Lahtinen *et al.* 2005). Rutland *et al.*
79 (2004) and Williams *et al.* (2008) divide the earliest orogeny (‘Fennian’ in Lahtinen *et al.*
80 2005) into two orogenic episodes; the Early Svecofennian orogeny (or ‘pre-Bothnian’)
81 before c. 1.92 Ga, and a Middle Svecofennian orogeny (or ‘post-Bothnian’) at c. 1.88-1.85
82 Ga. The two events were separated by rifting/extension of the pre-Bothnian basement and
83 associated volcano-sedimentary deposition. They also conclude that the pre-Bothnian
84 rocks were deposited in a large marginal basin and that neither the pre-Bothnian nor the
85 post-Bothnian events were arc-accretionary. In any case, most of the igneous and
86 sedimentary rocks in central and southern Finland formed during the early “Fennian”
87 event(s), before c. 1.85 Ga (e.g. Lahtinen *et al.* 2005, and references therein).

88 The E-W striking Tampere Schist Belt (TSB) lies in the centre of the Svecofennian
89 domain (Fig. 1). The entire TSB is c. 200 km long (Fig. 1a) but its best studied central part,
90 from Viljakkala in the west to Orivesi in the east, extends for about 60 km (Fig. 1b). In the

91 north, the TSB is bounded by the igneous rocks of the c. 1.88 Ga Central Finland Granitoid
92 Complex CFGC (e.g. Nironen *et al.* 2000). To the south of the TSB lies the Vammala
93 migmatite belt VMB (also known as the Pirkanmaa belt). The TSB – VMB contact is
94 strongly sheared along the curvilinear “Tampere shear zone” (TSZ), and intruded by the c.
95 1.88-1.87 Ga Siitama batholith near Orijärvi (Figs. 1b & 2; Nironen *et al.* 2000). The
96 batholith is internally relatively undeformed but is clearly sheared along the assumed
97 continuation of the TSZ.

98 The TSB is characterised by 1.90-1.89 Ga metavolcanic and related sedimentary
99 rocks with volcanic arc -type geochemistries, and by turbiditic metasedimentary rocks
100 originally deposited between c. 1.92-1.90 Ga (e.g. Seitsaari 1951; Ojakangas 1986;
101 Kähkönen 1987, 1989; Lahtinen *et al.* 2009). The rocks are metamorphosed in low-T
102 amphibolite to transitional greenschist/amphibolite facies conditions (Kilpeläinen *et al.*
103 1994). Overall, the central TSB defines a main syncline with an E-W striking, subhorizontal
104 hinge zone and a steeply dipping axial plane (Figs. 1b & 2; Seitsaari 1951; Kähkönen
105 1989; Nironen 1989). The northern limb is dominated by volcanic rocks, while the southern
106 limb is considerably richer in sedimentary rocks, although towards the southernmost
107 margin volcanics again become more abundant (e.g. Seitsaari 1951; Kähkönen 1994,
108 1999).

109 The nature of the basement underlying the TSB volcanics and sediments is unknown,
110 but Lahtinen & Huhma (1997) imply that evolved crust and associated mantle were in
111 place in the area around the onset of the extension at c. 1.92 Ga. The lowermost
112 recognised syn-rift unit is the volcanic Haveri Formation in the central TSB, and it consists
113 of (proto)oceanic crust (Fig. 1b; Mäkelä 1980; Kähkönen & Nironen 1994; Kähkönen 2005;
114 Lahtinen *et al.* 2009). The lower part of the Haveri Fm. is characterized by pillow lavas with
115 enriched mid-oceanic ridge/within plate basalt (EMORB/WPB) geochemistry. The upper
116 part consists of cherts, sedimentary carbonates, and black schists. The Haveri Fm. is
117 overlain by the sedimentary Osara Formation, equivalent to the Myllyniemi Formation near
118 Lake Näsijärvi. These formations consist mostly of what are in the TSB literature often
119 referred to as “turbidites”; greywackes and pelites deposited in submarine fan channels
120 with dominantly westerly palaeocurrents (Fig. 1b; Mäkelä 1980; Ojakangas 1986;
121 Kähkönen 1989). The Myllyniemi Fm resembles the sedimentary rocks at Karppi, Orivesi
122 in the east and are considered to represent the same unit (Fig. 1b; e.g. Seitsaari 1951).

123 Sulphide-bearing schist layers are found locally within the lower parts of the Myllyniemi
124 Fm, including the Karppi area. Upwards in the TSB succession, the Myllyniemi sediments
125 and their lateral counterparts are overlain by subduction-related basaltic to rhyolitic
126 volcanic rocks and by conglomeratic to pelitic sedimentary rocks derived thereof (Fig. 2;
127 Seitsaari 1951; Simonen 1953; Kähkönen 1999).

128 The Myllyniemi-type sediments were deposited between 1.92-1.90 Ga. The maximum
129 age of deposition is constrained by detrital zircons in the Osara and Myllyniemi Fms
130 (minimum U-Pb ages of c. 1.92 Ga; Huhma *et al.* 1991; Claesson *et al.* 1993; Lahtinen *et*
131 *al.* 2009). The minimum age of deposition is given by the 1898 ± 8 Ma age for the volcanic
132 unit at Pirttiniemi, stratigraphically directly overlaying the Myllyniemi Fm (Kähkönen 1989;
133 Kähkönen *et al.* 2004). In addition, the volcanics-dominated unit along the southern margin
134 of the TSB, the “Tohloppi-Kiviranta-Sorila volcanics” have been dated at 1892 ± 2 Ga
135 (magmatic age from a dacite at Tesoma close to Tohloppi; Kähkönen 1994; Kähkönen *et*
136 *al.*, 2004). Despite their apparent position below the Myllyniemi Fm on the map (Fig. 1b),
137 the Tohloppi volcanics are, therefore, also younger than the Myllyniemi sediments and
138 might be equivalent with the Pirttiniemi volcanics (Kähkönen 1994). The onset of the basin
139 inversion and the timing of the final closure are not well constrained, but it is implied that
140 the crust was already thickened by c. 1.88 Ga. Two large plutons of approximately this age
141 intrude the TSB: the Värmälä pluton with a U-Pb zircon age of 1878 ± 3 (Nironen 1989b;
142 Figs. 1b & 2), and the Hämeenkyrö pluton, dated at c. 1882 ± 6 Ma (U-Pb zircon age; Fig.
143 1c; Patchett & Kouvo 1986). Minor extrusive volcanism and related deposition at this stage
144 is, however, also implied (c. 1.88 Ga; Kähkönen & Huhma, 2012).

145 The relationship between the Vammala Migmatite Belt VMB and the TSB is important
146 in order to understand the tectonic setting of the TSB deposition. The gneisses and
147 migmatites of the VMB are dominated by metasedimentary rocks of interpreted turbiditic
148 origin. Volcanogenic rocks are not common and, where present, mostly show MORB- or
149 WPB-affinities, similar to the Haveri Fm. of TSB, rather than subduction-related
150 characteristics. In addition, the VMB is considerably richer in black schists than the TSB. A
151 relatively deep basin setting for much of the VMB is implied, but sedimentary rocks more
152 typical for shelf and/or upper parts of submarine fan environments have also been found
153 (Lahtinen *et al.*, 2009). The temporal and spatial relationship between the two belts has
154 been controversial. Nironen (1989a) suggested that the TSB and VMB belong to the same

155 unit but possibly represent different crustal levels/metamorphic conditions. Kähkönen
156 (1999, 2005) considered the VMB largely as the subduction zone complex related to the
157 arc volcanism in the TSB. However, age determination results imply that VMB includes
158 pre-1.91 Ga rocks, so that the TSB sediments and volcanics were deposited in a basin
159 formed during extension of this basement (Rutland *et al.* 2004). On the other hand,
160 Lahtinen *et al.* (2009) show that the Myllyniemi-type metasediments in the lower part of the
161 TSB succession and bulk of the VMB turbidites are identical in detrital zircon age
162 populations, and that a part of the VMB sedimentary rocks have c. 1.89 Ga detrital zircon
163 probably derived from the TSB arc volcanics. Lahtinen *et al.* (2009) also emphasized
164 changes in the sedimentary and tectonic environments during evolution of the TSB and
165 VMB. In their model, the basalts and the sedimentary rocks of the Haveri Fm. were formed
166 during rifting of a c. 2.0 Ga microcontinent. As the rifting continued, the sedimentation
167 environment changed to a passive margin, and the Myllyniemi-type sediments were
168 deposited at c. 1.92-1.90 Ga. The onset of subduction and TSB arc-type volcanism at c.
169 1.90 Ga caused a change from a passive margin/continental slope setting to that of an
170 (intra-)arc system, where the 1.90-1.89 Ga TSB volcanism represent the 'proper' volcanic
171 arc and the VMB acted mainly as the accretionary complex.

172 The folding that produced the main Tampere syncline is interpreted to be associated
173 with the main basin inversion and closure event (Fig. 2; F_1 folding phase of Nironen
174 1989a). Especially the southern limb seems to contain several tight to isoclinal, upright F_1
175 folds with gently dipping to subhorizontal fold axes/hinge lines, 'parasitic' with respect to
176 the main regional syncline (Fig. 2; Nironen 1989a). The southern limb is bounded by the
177 Tampere shear zone: kinematic indicators imply thrusting toward north (i.e. reverse south-
178 side up; Figs. 1b & 2; e.g. Nironen 1989a; Kähkönen 1999). Nironen (1989a) suggests that
179 the displacement along the thrust was essentially vertical, and that the thrust formed
180 during the first deformation phase D_1 , i.e. simultaneously with F_1 . The strike-slip kinematics
181 of the TSZ are not yet described in the literature; near Tampere, asymmetric structures
182 have not been observed in the horizontal plane (Nironen 1989a). Nironen (1989a) further
183 suggests that the Hämeenkyrö and Värmälä plutons (Figs. 1 & 2) intruded during the F_1 .

184 The area of this study is located in the eastern part of the TSB (Figs. 1 & 2). The area
185 was chosen for two main reasons: i) it represents the (presumably) lower level of the
186 stratigraphy, the tectonic development and the present structural geometry of which is

187 largely unknown; and ii) it contains some sulphide-bearing schist horizons/lenses and
188 geophysical anomalies that are assumed to be associated with the schist horizons. The
189 relationships of the sulphide-bearing schists to the overall volcano-sedimentary
190 architecture and/or the tectonic history are unknown. Kähkönen (1999) identifies two units,
191 Karppi turbidites and Pohjala volcanics, in the area (Figs. 1b, 2). The Karppi turbidites in
192 the north consists of greywackes and mudstones, whereas the Pohjala volcanics in the
193 south contain various volcanoclastic rocks in addition to massive plagioclase- and
194 plagioclase-uralite-porphyrysts. The Karppi Fm. has been correlated with the Myllyniemi and
195 Osara Fms. further west (Kähkönen 1999, 2005).

196

197 **Observations**

198 The exposure in the entire TSB is limited due to gentle topography, widespread
199 Quaternary deposits, and thick soils. However, in more elevated areas, good-quality
200 outcrops are available or can be created with reasonable effort, enabling representative
201 observations. Special attention was paid to the structural and volcano-sedimentary
202 features, the architecture, and the spatial relationships of the mapped rocks.

203 Like elsewhere within the TSB, the studied supracrustal rocks show well preserved
204 primary structures, despite the evidently considerable shortening during basin closure (Fig.
205 3). In many places, the S_0 (bedding) top-of-strata indicators and various other primary
206 structures are clearly identifiable. The strata are usually strongly tilted with dips of $>70^\circ$.
207 The first secondary foliation (local S_1) is a roughly E-W to NEE-SWW striking schistosity,
208 mostly subparallel to S_0 (Fig. 4). A younger secondary schistosity S_2 is seen to strike at an
209 angle to S_1 , usually NE-SW and crenulating the S_1 (Fig. 3h). The S_1 and S_2 vary in
210 intensity depending on the rock type: the fine-grained, mica-rich mudstones often show
211 intense schistosity, while the coarser-grained sedimentary and volcanic rocks may seem
212 almost undeformed (Fig. 3). Kinematic markers (sigma clasts, S-C structures) are
213 observed occasionally, mostly in the south. On subhorizontal surfaces, they record both
214 dextral and sinistral sense of shear. However, shearing becomes more intense toward the
215 south and sinistral kinematic markers become dominant close to the implied continuation
216 of the Tampere shear zone.

217 At least two different folding styles (here called F_1 and F_2) are distinguishable in the
218 field. F_1 shows tight to isoclinal folds with subvertical fold axial planes and gently to

219 moderately (up to c. 40°) plunging fold axes/hinge lines. F_2 is observed in the southern
220 part of the mapping area, where it is seen to re-fold F_1 . F_2 displays open to closed folds
221 with steep axial planes and fold axes plunging c. 60-70° (Fig. 4).

222 The rock types are here divided into two main lithological units: the metavolcanics-
223 dominated southeastern part (the volcanic unit VU); and the metasiliciclastics-dominated
224 northern part (siliciclastic unit SCU) (Figs. 3 & 4). There is a sharp transition from one unit
225 to the other, only observable in the eastern part of the mapping area due to the c. 1.88-
226 1.87 Ga Siitama batholith intruding the units in the western part (Fig. 4).

227 The two main units SCU and VU are very distinct and display different lithological,
228 architectural, and structural characteristics. The VU is composed both of coherent
229 porphyritic rocks, evidently lavas, domes, and/or sills in origin, and of stratified volcanic
230 rocks. (Fig. 4). Individual unit thicknesses reach up to c. 150 m. Some sedimentary
231 (volcaniclastic) rocks are found in the central part of the VU. S_1 is usually (sub)parallel to
232 S_0 in the VU, and is best developed in the finer-grained layers. S_1/S_0 normally follows an
233 E-W strike throughout the VU, except at the contact with Siitama batholith. The top-of-
234 strata indicators (graded bedding, erosion surfaces) are common and dominantly towards
235 the south, except in the southern part of the VU where the younging direction is towards
236 the north (Fig. 4). The SCU rocks on the other hand are dominated by sandstones, but
237 pebble conglomerates and mudstones (shales and schists) are common as well. Individual
238 beds within the SCU are of variable thicknesses, from cm-scale to m-scale. Top-of-strata
239 indicators systematically, with one possible exception close to the VU-SCU contact,
240 indicate northward younging directions (ripples, graded bedding, erosion surfaces; Figs.
241 3g, h). As with the VU, S_1/S_0 mostly follows an E-W strike except locally in the vicinity of
242 the Siitama batholith. S_2 schistosity is commonly observed in mudstones, striking
243 mostly approximately SSW-NNE to WSW-ENE (Figs. 3h & 4).

244 Based on the southwards-younging pattern of the VU, the stratigraphic succession
245 of the VU subunits can be defined (from bottom to top): (VU1) felsic volcanoclastics, (VU2)
246 intermediate volcanoclastics, (VU3) coherent felsic plagioclase porphyrys, (VU4) coherent
247 uralite-plagioclase (ur-pl) porphyritic rocks (lavas or sills), (VU5) volcanoclastics including
248 conglomerates and sandstones (here called "southern volcanoclastics"), and (VU6) mafic
249 and felsic tuffs. The northwards-younging SCU rocks are discussed as a single unit as
250 individual beds do not define laterally coherent subunits as within the VU (Fig. 4). The

251 relative age and stratigraphic position the SCU with respect to the VU will be discussed
252 later.

253 (VU1) *The felsic volcanoclastics* are at its base composed of fine volcanic sandstones
254 (“tuffs”; Fig. 4). These are followed by volcanic conglomerates/breccias (“lapillistones”)
255 covered by coarse tuffs (Fig. 3a), both of which are interbedded with more fine tuffs. The
256 beds in this unit are relatively well sorted. S_1 is strong in the fine tuffs.

257 (VU2) *The intermediate volcanoclastics* contain two distinguishable types of strata: (a)
258 sandy tuff with few lapillis and gradual variations in grain size leading to diffuse strata
259 boundaries, and (b) graded lapillistone interlayered with sandy tuff, including some cobble-
260 sized fragments of porphyritic lavas (Figs. 3b & 4). Some of the lapillis are strongly
261 angular. Both types are rich in biotite and show strong S_1 .

262 (VU3) *The coherent felsic plagioclase porphyrys* contain feldspar phenocrysts of
263 variable sizes (typically 1-3 mm in diameter, Figs. 3c & 4). A weak S_1 schistosity is
264 observed, and networks of hydrothermal veins are abundant.

265 (VU4) *The ur-pl porphyrys* envelope the subunits VU5 and VU6 (Figs. 3d & 4). They
266 contain abundant c. 1-3 mm phenocrysts of uralite and/or plagioclase in fine-grained
267 matrix. In the northern part of this unit, uralite phenocrysts are more common, while
268 plagioclase phenocrysts gradually become more dominant toward the south. S_1 is fairly
269 weak in the northern part of the subunit but becomes strong towards the south. Both
270 dextral and sinistral sigma clasts and S-C structures are observed in the southern part of
271 the unit, but most of the observed indicators show sinistral shear.

272 (VU5) *The southern volcanoclastics* are exposed within two elongated areas,
273 separated by the subunit VU6 (Figs. 3e & 4). VU5 consists mainly of volcanic sandstones,
274 pebbly sandstones, and conglomerates, but some tuff breccias rich in angular clasts are
275 also observed. Weak S_1 is present. As is typical for other VU rocks, ripples and erosion
276 surfaces in this subunit show southward top-of-strata, but only in the exposures that are
277 located north of VU6 (Figs. 3e & 4). South of VU6, the VU5 rocks show northward top-of-
278 strata. S_1 is weak.

279 (VU6) *The mafic and felsic tuffs* consist of alternating fine-grained mafic and felsic
280 layers. The felsic material often occurs as lenses within the mafic material (Fig. 3f). F_1 tight
281 to isoclinal folds with steep axial planes and c. 40° eastward plunging fold axes are

282 observed (Fig. 3f). Close to the Siitama batholith, F_2 refolds the F_1 folds into open to
283 closed folds with fold axes plunging c. 60° toward the SE-ESE (Fig. 4).

284 (SCU) *The SCU* (Figs. 3g-l) shows laterally varying layer thicknesses in an E-W
285 direction so that the conglomerates, the sandstones and the mudstones define
286 interfingered lensoid shapes on the geological map (Fig. 4). The grain size variations
287 define rhythmic ('cyclic') patterns at the map scale, with overall fining-upward sequences
288 (younging towards north). The only folds that can be observed are very local, tight,
289 asymmetric (dextral) folds in the southern part of the metasiliciclastic unit (Fig. 4). These
290 folds show W-E to NW-SE striking, steeply dipping axial planes and roughly 40° east- to
291 southeastwards plunging fold axes.

292 Sulphide-bearing schists appear in three exposures within the SCU (Figs. 3j & 4).
293 Two of the exposures occur in the central part of the mapping area, close to the contact
294 with the Siitama batholith and the VU, and one exposure is a thin horizon in the northwest.
295 In addition, sulphide schist are also found within the Siitama batholith (Fig. 4). The
296 sulphide schists are often extensively weathered and eroded, and have rusty outcrop
297 surfaces. The bedding or other structures are not easily observable, but a few coarser-
298 grained sandstone lenses are distinguishable in places. S_1 is subparallel to S_0 (where
299 visible). The sulphide mineralisation consists mainly of small amounts of disseminated
300 pyrite and chalcopyrite visible to the naked eye, and of secondary mineralisation (mainly
301 haematite) identifiable in the fractures in thin section (Figs. 3k & l). In thin section, the S_1 is
302 strongly developed, defined by elongate micas and opaque minerals (Figs. 3k & l). It is
303 notable that the opaque minerals, presumably pyrite/chalcopyrite, and haematite, are
304 present both within the S_1 and within the fractures cross-cutting it. The economic potential
305 of the sulphide schists within the study area has not been assessed, but the grade seems
306 to be low.

307

308 **Interpretation**

309

310 *Structure*

311 The observed S_1 schistosity is relatively constant throughout the area (Fig. 4). S_1
312 is generally striking E-W and (sub)parallel to the bedding S_0 . S_1 is interpreted to represent
313 the main deformation phase D_1 (i.e. main inversion and basin closure), which resulted in

314 the large-scale folding and the formation of the TSB syncline. The intensity of the S₁
315 schistosity is directly dependent on the lithotype: it is best formed in the mudstones but
316 only weak in the sandstones and many of the coherent porphyritic metavolcanics. In the
317 absence of clear signs of significant lateral movement (shearing) in most parts of the
318 mapping area, the dominantly E-W strikes of the S₁ would imply approximately N-S
319 shortening during D₁, in accordance with previous works (e.g. Kähkönen 1989; Nironen
320 1989a). The appearance of dominantly sinistral kinematic markers toward the south,
321 affecting also the northernmost edge of the Siitama batholiths, suggests that the
322 continuation of the Tampere shear zone (TSZ) exists in this area, and that the shear sense
323 along the TSZ was sinistral, at least at the time of the emplacement of the batholith (c.
324 1.88-1.87 Ga).

325 S₂ is locally observed, and it constantly strikes at an angle of c. 40-50° counter-
326 clockwise from S₁. S₂ is usually, but not always, observed in the mudstones, where it is
327 seen to crenulate S₁. S₂ is interpreted to represent a D₂ deformation phase separate from
328 and younger than S₁. Like S₁, S₂ is not seen to be associated with pervasive lateral shear
329 in the field, so that the dominantly NE-SW strike of the S₂ suggests an approximately NW-
330 SE oriented compression during D₂. The D₂ is probably unrelated to the sinistral TSZ
331 shearing, as a NW-SE compression would have resulted in dextral shearing along the
332 TSZ.

333 The polarity observations imply a previously unknown F₁ syncline, 'the Pohjala
334 syncline', in the VU (Figs. 4 & 5). The tightly folded rocks of subunit VU6 form the core of
335 the syncline. The c. 40° eastward plunge of the F₁ fold axes in VU6 is interpreted to be
336 representative for the entire syncline. The Pohjala syncline is best defined by the subunit
337 VU5 that shows opposite bedding polarities on the north vs. south sides of the VU6 (Figs.
338 4 & 5). The rest of the subunits (VU1-VU4) form the northern limb of the Pohjala syncline
339 (a narrow zone of subunit VU4 is also present in the southern limb). The c. 40° eastward
340 plunge of Pohjala syncline is interpreted to at least partly result from the intrusion of the
341 Siitama batholith. In the hinge zone of the Pohjala syncline (in VU6), the F₂ folding event is
342 seen to refold the isoclinal F₁ folds. The F₂ folds are open to closed folds with hinge lines
343 plunging c. 60° toward the SE-ESE. They are interpreted to be related to the sinistral
344 shearing along the TSZ.

345 In the SCU, north of the VU, both the lithology and the polarity indicators abruptly
346 change (Fig. 4): igneous rocks are present only as relatively thin sills/dykes and, except at
347 one location close to the VU/SCU transition, the SCU rocks consistently show northward
348 top of strata. The SCU rocks also show mostly northward dipping S_0 . The age of the
349 dykes/sills is unknown, but they show S_1 schistosity. The tight, asymmetric folds
350 observed in the southern part of the SCU are interpreted to be F_1 folds. There is a
351 possibility that the intrusion of the Siitama batholith has caused the local folding. However,
352 due to the similarity between these folds and the F_1 folds within the Pohjala syncline
353 (similar tight to isoclinal folding with moderately eastward-plunging hinge lines), we
354 interpret both folds to have formed during the main deformation (folding) phase. It is
355 unclear whether the dextral asymmetry of the F_1 folds at the bottom of the SCU reflects a
356 relative movement of the north-side-up, or whether the folds are parasitic folds of a larger
357 fold within the metasiliciclastic unit. Because no sinistral F_1 folds have been found
358 (corresponding the other hypothetical limb of a larger fold), and because the folds are only
359 found close to an inferred fault zone (see below) we tentatively prefer the interpretation of
360 these F_1 folds as shear folds indicating north-side-up reverse movement.

361 A discontinuity plane is interpreted to offset the SCU subunits (Fig. 4). The inferred
362 fault has an apparent dextral kinematics and bends into parallelism with the S_0 in the
363 southern part of the mapped metasiliciclastic unit (Figs. 4 & 5). The age of the inferred
364 fault is uncertain. It may, therefore, represent either a late dextral fault zone, formed as a
365 response to D_2 with c. NW-SE compression or alternatively it might form a part of an
366 imbricate thrust system formed during D_1 (the dextral offset therefore being apparent).

367 The most extensive sulphide schist occurrences in the area are found in the lowest
368 parts of the SCU near the contact with the VU (Figs. 4 & 5). The sulphide mineralisation is
369 interpreted to have occurred in two phases. The first mineralisation phase occurs within
370 the S_1 (Figs. 3k & l). Therefore, the mineralisation is interpreted to be genetically linked to
371 the formation of S_1 and to the closure of the basin. The sulphide-bearing hydrothermal
372 fluids from which the minerals precipitated probably sourced from deeper within the basin.
373 The second phase occurred during a brittle deformation event and also suggests
374 hydrothermal fluid activity within the area. There is no indication in the thin sections that
375 the S_1 -related mineralisations were remobilised during this event. A possible source for
376 this second-phase fluids is the Siitama batholith. However, sulphide schist blocks have

377 been found within the batholith, so that the mineralisation within the schist must, therefore,
378 be older than the intrusion (1.88-1.87 Ga). Sulphide schist has also been observed north
379 of the study area, kilometres away from the batholith, further suggesting a remobilisation of
380 sulphides by another mechanism than magmatic fluids. Finally, not all the schists and
381 none of the other rock types are sulphidised, which would be expected with a near-by
382 magmatic source.

383

384 *Architecture*

385 The SCU and VU can be reasonably correlated with the Karppi turbidites and the
386 Pohjala volcanics, respectively, east of the mapping area (Fig. 1a; see also Seitsaari,
387 1951). We also correlate the SCU with the turbidites of the Myllyniemi Fm. (Fig. 1;
388 Kähkönen 1999) also typically located in the southern limb of the main TSB syncline (Figs.
389 1b, 2; Kähkönen 1989; Nironen 1989a). In accordance with previous works, we interpret
390 the interfingering metasiliclastic rocks of the SCU to have been formed in a submarine
391 turbidite fan coevally with the other Myllyniemi-type sedimentary rocks at c. 1.92-1.90 Ga
392 (Fig. 6). At least five fining-upward cycles at a 100-metre scale are observed in the
393 mapped part of this unit, suggesting stepwise basin subsidence/sea level transgression
394 (Fig. 4). Alternatively, the apparent repetition of the cycles might be caused by a
395 development of an imbricate system within the SCU.

396 The architecture of the VU is simpler than that of the SCU, assuming that the
397 porphyries of VU4 are completely extrusive and not sills. The tuffs, lapillistones, and
398 porphyries (VU1-VU4) are overlain by the relatively thin volcanoclastic VU5. The ripples in
399 the sandstones of VU5 are consistent with erosion and redeposition within a dynamic
400 volcanic environment (Fig. 3e). The youngest rocks within the VU are the fine VU6 tuffs in
401 the core of the Pohjala syncline (Fig. 3f). The rheological differences during deformation
402 between the mafic and the felsic material in VU6 are interpreted to be responsible for the
403 break-up of the felsic material into fragments, whereas the mafic material behaved more
404 ductilely.

405 The relative age of the VU with respect to the SCU is not immediately obvious from the
406 collected data. The age relationship of the units has consequences to the resulting
407 geological model: this is discussed in the following chapter.

408

409 **Discussion**

410 *Geological model*

411 The mapping reveals a 'double-syncline' (Figs. 4 & 5): the Pohjala syncline within the
412 VU in the south, and the larger TSB main syncline system in the north (only the
413 metasiliciclastics-dominated southern limb of which extends into the mapping area).
414 However, no anticline has been observed between the two synclines, and the transition
415 from one syncline/unit to another (and from one bed polarity trend to another) is sharp.
416 The single southward top-of-strata observation within the SCU close to the VU/SCU
417 contact may represent remnants of an anticline, but this is very local as elsewhere along
418 the contact the SCU top-of-strata indicators are systematically northward. This indicates
419 that faulting must be involved and is responsible for the juxtaposition and the present
420 configuration of the units. No such shear or fault zone is directly observed in the VU/SCU
421 contact zone (however, the contact itself is not exposed). The primary structures are often
422 very well preserved, even right next to the inferred contact zone, and the lack of
423 asymmetric kinematic markers at outcrop suggest coaxial rather than shear deformation.
424 The only possible asymmetric kinematic markers related to the inferred fault/shear zone
425 are found in the mudstones near the VU/SCU contact: the asymmetric F_1 folds could be
426 tentatively interpreted as shear folds indicating north-side-up reverse movement. The
427 absence of obvious fault or shear zone indicators in the vicinity of the contact zone does
428 not, however, as such contradict the hypothesis for faulting. Many examples exist
429 worldwide of simple shear (non-coaxial strain) being strongly partitioned into narrow zones
430 that accommodate large amounts of movement. For example, the Glencoul thrust in NW
431 Scotland forms a part of a thin-skinned thrust system that accommodates up to tens of
432 kilometres of transport along a mylonitic (non-coaxial) thrust zone that is only from some
433 decimetres to up to some metres in thickness, while the strain directly below the shear
434 zone is mostly coaxial (e.g. Butler 1984; Law 1987; Law *et al.* 2010).

435 The age relationship between VU and SCU is obscure: the polarity indicators and the
436 lithologies and, therefore, the relative ages of the units cannot be systematically carried
437 over from one unit to the other. However, if the asymmetry of the F_1 folds observed close
438 to the VU/SCU contact reflects north-side-up kinematics, then the SCU is older than the
439 VU. This interpretation is supported the stratigraphic correlation with dated units elsewhere
440 within the TSB. The VU seems to occupy an approximately similar stratigraphic position as

441 the Tohloppi-Kiviranta-Sorila volcanics further west, whereas the Myllyniemi turbidites,
442 presently north of the Tohloppi volcanics, can be correlated with the SCU (Kähkönen
443 1994; Fig. 1b). The depositional age of the Myllyniemi Fm. has been determined to 1.92-
444 1.90 Ga while the Tohloppi-Kiviranta-Sorila volcanics are younger. Based on this
445 implications that the VU is younger than the SCU, propose the following geological model
446 for the studied area (Fig. 7).

447 1) The SCU was deposited, like other parts of the Myllyniemi Fm, mainly
448 on a margin of a rifted microcontinent at c. 1.92-1.90 Ga, with the rift-related
449 Haveri-type volcanics partly at the base (Figs. 6 & 7a). The lobate architecture of
450 the SCU results from deposition largely on the microcontinental slope during
451 continuing but stepwise basin subsidence (Fig. 6; see also Kähkönen 1999).

452 2) The onset of northward subduction and related arc-type volcanism at
453 c 1.90 Ga caused a change to volcanism-dominated deposition in the TSB for the
454 period of c. 1.90-1.89 Ga. In our study area, the volcanism resulted first in the
455 deposition of the felsic and intermediate tuffs (VU1 and VU2), followed by the pl-
456 and the ur-pl porphyrys (VU3 and VU4), the southern volcanoclastics (VU5), and
457 finally the mafic to felsic tuffs (VU6).

458 3) During the continuing subduction basin inversion initiates with N-S
459 compression (the local D_1 event; Fig. 7b). Some of the extensional faults are
460 reactivated as (blind) reverse faults, leading to open folding of the VU and SCU.
461 This stage of local “thick-skinned” deformation is needed to produce the original
462 Pohjala syncline; a purely thin-skinned thrusting of sub-horizontal sedimentary
463 packages on top of each other would probably have only resulted in repetition of
464 units, without significant folding of the units in the footwall of the thrust.

465 4) As the basin closure progresses, the normal fault inversion cannot
466 accommodate the increasing amount of shortening, and a southward-propagating
467 “thin-skinned” thrust system develops (Fig. 7c). Where the thick-skinned
468 shortening is accommodated at this point is not inferable from our data. The thrust
469 planes localise approximately layer-parallel and break towards the surface when
470 they meet the buried basement highs defined by the faulted and partly inverted
471 blocks. The thrust system juxtaposes different geological units, including the
472 observed southward-younging and northward-younging units. The earlier

473 anticlines are mostly destroyed as they are thrust upwards and subsequently
474 eroded. At least one major thrust is active within the VMB: this fault will later
475 develop into the Tampere shear zone (TSZ; see also Nironen 1989a). The details
476 of the development of TSZ are outside the scope of this study, but the TSZ thrust
477 helps to accommodate the overall shortening and tightens the Pohjala syncline.

478 5) Continuing shortening leads to a tightening of the folds and in the
479 rotation of the bedding and the thrust faults into an upright position (Fig. 7d). In
480 their steeper position, the thrust faults would occupy a position that is
481 mechanically unfavourable for reverse movements. Lateral (shear) movements
482 along these previous thrust fault planes is more likely at this stage, the sense of
483 shear possibly being mostly sinistral based on the field observations. However,
484 bedding-perpendicular coaxial (pure) shear while the beds were in an upright
485 position probably accommodate some shortening, so that the S_1 probably
486 developed latest at this stage. At depth, the layer-parallel thrusts are folded
487 together with the rock units that host them.

488 6) Latest at the time of the intrusion of the Siitama batholith c. 1.88-1.87
489 Ga, the sinistral lateral kinematics in this part of the Tampere shear zone is
490 established and the structures have reached their final subvertical orientations
491 (Fig. 7e).

492 7) The second deformation phase D2, with compression from c. SSE-
493 NNW crenulates the S_1 schistosity, and leads to localised dextral faulting within
494 the metasiliciclastic unit (the curved fault, Figs. 4 & 5).

495
496 The bulk of the shortening is therefore accommodated partly by folding, partly by the
497 thrusts, and partly by the bedding-perpendicular shortening (probably including volume
498 change and compaction). The minimum amount of shortening across the VU can be
499 calculated, on the basis of the folding of the VU, to a minimum of c. 50%. This shortening
500 estimate cannot be directly extrapolated to the scale of the TSB, because VU-type folding
501 is not at present identified in the northern parts of the belt. On the other hand, the
502 interpreted overall geometry of the TSB (Fig. 2) and the implications of the structural
503 model presented above suggest a minimum shortening of a similar order of magnitude.
504 The estimate does not consider volume change, or account for the unknown total number

505 of thrusts within the TSB. The possible existence of additional thrusts and/or
506 imbricate/duplex systems within the SCU would mean that the total shortening was notably
507 larger than 50%.

508 The model largely agrees with previously published literature in terms of the overall
509 temporal and spatial development of the basin. The development of early thrusts and their
510 subsequent rotation and folding at a large scale is inferred, confirming the interpretation of
511 (Nironen 1989a). However, the model gives a more detailed suggestion for the structural
512 development and shortening of the TSB that might explain some of the observed problems
513 within the TSB (e.g. Nironen 1989a). Furthermore, contrary to Nironen (1989a), the implied
514 main thrusting direction, at least in the southern margin of the TSB, is toward the south. A
515 northward-verging thrust system cannot explain the observed structural geometries and
516 the inferred age relationships; however, if the VU is older than the SCU, a northward-
517 verging thrust system is possible. Northward structural vergences within and at the
518 northern margin of the VMB are consistent with our model, and the model in fact requires
519 that the Vammala area is uplifted (or a subduction complex/accretionary wedge develops)
520 at an early stage of the basin inversion. Tampere are would, effectively, have formed a
521 backstop to the developing Vammala wedge/complex.

522

523 *The implications to the sulphide mineralisation*

524 The sulphide mineralisation seems to be a result of a combination of sedimentary
525 processes and hydrothermal activity (remobilisation) along restricted zones within the TSB.
526 In thin section, the sulphide schist displays iron oxide and sulphide mineralisation both
527 within the schistosity, and along the second-order fractures (Figs. 3k & l). The most
528 prominent sulphide schist exposures within the study area are located close to the
529 VU/SCU contact, where the presence of an early thrust is implied in our model. This
530 spatial association suggests a genetic relationship between the localisation of the thrust
531 fault, the occurrence of the schistosity-parallel mineralisation, and the localisation of the
532 mineralised second-order fractures. The model suggests that the first mineralisation phase
533 occurred at model stage 5) when hydrothermal, mineralised fluids moved within and close
534 to the old thrust plane. The ore minerals were trapped within the developing S_1 schistosity
535 close to the old thrust plane (Figs. 3k & l). The fluids probably sourced from within the
536 lower parts of the basin, not exposed within the study area. If the base of the basin infill in

537 our mapping area consists of similar VMS rocks than those that are interpreted to source
538 at least the Haveri deposit, they might have served as a source for the dissolved metals in
539 the hydrothermal fluids. Another possible source for the fluids is magmatic, but the nearest
540 intrusion, the Siitama batholith, is younger than the sulphidisation (evidenced by the
541 presence of sulphide schist xenoliths within the intrusion). Some of the mineralisation was
542 remobilised later, or there was a new pulse of mineralised fluids, as evidenced by the
543 haematite- and sulphide-filled second-order fractures (Figs. 3k & l). The age of the
544 fractures is unknown but they localised within the inferred palaeo-thrust after the main
545 schistosity within the thrust had formed, and may reflect later strike-slip movements
546 inferred by our model. The results imply that presently subvertical mineralised fault zones
547 may have originally been more gently dipping, and that they may not necessarily have
548 sourced their minerals from great depths (i.e. from deep within the basement).

549 If the sulphide schists within the TSB are related to early thrusts, the presence of
550 sulphide schists might help in deducing the structural geometries and the basin closure
551 processes elsewhere within the basin, and possibly aid mineral exploration in the area. For
552 example, the Kutemajärvi and Haveri deposits are at least partly fault-controlled, and both
553 deposits are hosted by schists (e.g. Nironen 1994; Eilu *et al.* 2003). Furthermore, an
554 extensive sulphide schist horizon outcrops just north of the mapped area, possibly
555 indicating a presence of a prominent early thrust with similar mineralisation. A minor
556 sulphide schist lens is also found in the northeastern part of the mapping area, implying
557 that a minor thrust might exist here, possibly related to an unconfirmed imbricate system
558 (Fig. 5).

559

560 **Summary**

- 561 • The detailed structure of a part of the Palaeoproterozoic, well preserved Tampere
562 basin is described.
- 563 • The observed architecture and the structures of the volcanosedimentary rocks within
564 the southeastern Tampere basin imply a presence of hitherto unrecognised
565 palaeo-thrusts within the basin. These thrusts played an important role in
566 accommodating the overall shortening during basin closure.

- 567 • We propose a conceptual tectonic model that implies a transition from a normal fault
568 inversion-type crustal shortening, to locally ‘thin-skinned’ accommodation of
569 shortening as the basin closure continued.
- 570 • The palaeothrusts were also key in the localisation of the fracture zones that
571 transported and trapped sulphide-bearing hydrothermal fluids, as the sulphide
572 mineralisations are spatially linked to the thrusts. The proposed tectonic model
573 correlates the structural geometries observed in the field and the occurrence of the
574 sulphide-bearing fractured schists.

575

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579 **References**

580 Allen, R. L., Weihed, P. & the Global VHMS Research Project team 2002. Global
581 comparison of volcanic-associated massive sulphide districts. In: Blundell, D. J.,
582 Neubauer, F. & von Quadt, A. (eds) *The Timing and Location of Major Ore Deposits in an*
583 *Evolving Orogen*. Geological Society, London, Special Publications, **204**, 13-37.

584 Butler, R. W. H. 1984. Structural evolution of the Moine thrust belt between Loch More and
585 Glendhu, Sutherland. *Scottish Journal of Geology*, **20**, 161-179.

586 Claesson, S., Huhma, H., Kinny, P. D. & Williams, I. S. 1993. Svecofennian detrital zircon
587 ages – implications for the Precambrian evolution of the Baltic Shield. *Precambrian*
588 *Research*, **68**, 231-240.

589 Eilu, P., Sorjonen-Ward, P., Nurmi, P. & Niiranen, T. 2003. A review of gold mineralization
590 styles in Finland. *Economic Geology*, **98**, 1329-1353.

591 Gaál G. & Gorbatshev R. 1987. An outline of the Precambrian evolution of the Baltic
592 Shield. In: Gaál G. & Gorbatshev R. (eds) *Precambrian Geology and Evolution of the*
593 *Central Baltic Shield*. Precambrian Research, **35**, 15-52.

594 Huhma, H., Claesson, S., Kinny, P. D. & Williams, I. S. 1991. The growth of early
595 Proterozoic crust: new evidence from Svecofennian detrital zircons. *Terra Nova*, **3**, 175-
596 179.

597 Kähkönen, Y. 1987. Geochemistry and tectonomagmatic affinities of the metavolcanic
598 rocks of the early Proterozoic Tampere Schist Belt, southern Finland. *Precambrian*
599 *Research*, **35**, 295-311.

600 Kähkönen, Y. 1989. Geochemistry and petrology of the metavolcanic rocks of the early
601 Proterozoic Tampere Schist Belt, southern Finland. *Geological Survey of Finland, Bulletin*,
602 **345**, 104 p.

603 Kähkönen, Y. 1994. Shoshonitic and high-K metavolcanic rocks in the southern parts of
604 the Paleoproterozoic Tampere Schist Belt, southern Finland: evidence for an evolved arc-
605 type setting. *Geological Survey of Finland, Special Paper*, **19**, 101-115.

606 Kähkönen, Y. 1999. Stratigraphy of the central parts of the Palaeoproterozoic Tampere
607 schist belt, southern Finland: review and revision. *Bulletin on the Geological Society of*
608 *Finland*, **71**, 13-29.

609 Kähkönen, Y. 2005. Svecofennian supracrustal rocks. In: Lehtinen, M., Nurmi, P. & Rämö,
610 T. (eds) *The Precambrian Bedrock of Finland – Key to the evolution of the Fennoscandian*
611 *Shield*. Elsevier B.V., Amsterdam, 343-406.

612 Kähkönen, Y., Huhma, H. 2012. Revised U-Pb zircon ages of supracrustal rocks of the
613 Palaeoproterozoic Tampere Schist Belt, southern Finland. In: Kukkonen, I. *et al.* (eds)
614 *Lithosphere 2012, Programme and extended abstracts*. Institute of Seismology, University
615 of Helsinki, Report S-56, 132 p.

616 Kähkönen, Y., Huhma, H. & Aro, K. 1989. U-Pb zircon ages and Rb-Sr whole-rock isotope
617 studies of early Proterozoic volcanic and plutonic rocks near Tampere, southern Finland.
618 *Precambrian Research*, **45**, 27-43.

619 Kähkönen, Y., Huhma, H. & Mänttari, I. 2004. TIMS and SIMS U-Pb zircon ages and Rb-
620 Sr whole-rock isotope studies of early Proterozoic volcanic rocks near Tampere, southern

- 621 Finland. In: Mansfield, J. (ed) *The 26th Nordic Geological Winter Meeting, Abstract*
622 *Volume*. GFF, **126**, 25.
- 623 Kähkönen, Y. & Nironen, M. 1994. Supracrustal rocks around the Paleoproterozoic Haveri
624 Au-Cu deposit, southern Finland: evolution from a spreading center to a volcanic arc
625 environment. *Geological Survey of Finland, Special Paper*, **19**, 141–159.
- 626 Kilpeläinen, T., Korikovsky, S., Korsman, K. & Nironen, M. 1994. Tectono-metamorphic
627 evolution in the Tampere-Vammala area. *Geological Survey of Finland, Guide*, **37**, 27-34.
- 628 Korsman, K., Koistinen, T., Kohonen, J., Wennerström, M., Ekdahl, E., Honkamo, M.,
629 Idman, H. & Pekkala, Y. (eds) 1997. *Bedrock Map of Finland 1:1000000*. Geological
630 Survey of Finland.
- 631 Lahtinen, R. & Huhma, H. 1997. Isotopic and geochemical constraints on the evolution of
632 the 1.93-1.79 Ga Svecofennian crust and mantle in Finland. *Precambrian Research*, **82**,
633 13-34.
- 634 Lahtinen, R., Huhma, H., Kähkönen, Y. & Mänttari, I. 2009. Paleoproterozoic sediment
635 recycling during multiphase orogenic evolution in Fennoscandia, the Tampere and
636 Pirkanmaa belts, Finland. *Precambrian Research*, **174**, 310-336.
- 637 Lahtinen R., Korja A. & Nironen M. 2005. Palaeoproterozoic tectonic evolution. In:
638 Lehtinen, M., Nurmi, P. A. & Rämö, O. T. (eds) *Precambrian Geology of Finland – Key to*
639 *the Evolution of the Fennoscandian Shield*. Elsevier B.V., Amsterdam, 481-532.
- 640 Law, R. D. 1987. Heterogeneous deformation and quartz crystallographic transitions:
641 natural examples from the Moine Thrust zone at the Stack of Glencoul, northern Assynt.
642 *Journal of Structural Geology*, **9**, 819-833.
- 643 Law, R. D., Mainprice, D., Casey, M., Lloyd, G. E., Knipe, R. J., Cook, B. & Thigpen, J. R.
644 2010. Moine Thrust zone mylonites at the Stack of Glencoul: I – microstructures, strain
645 and influence of recrystallization on quartz crystal fabric development. In: Law, R. D.,
646 Butler, R. W. H., Holdsworth, R. E., Krabbendam, M., Strachand, R. A. (eds) *Continental*
647 *Tectonics and Mountain Building: The Legacy of Peach and Horne*. Geological Society of
648 London, Special Publications, **335**, 543-577.

- 649 Lentz, D. R. 1998. Petrogenetic evolution of felsic volcanic sequences associated with
650 Phanerozoic volcanic-hosted massive sulphide systems: the role of extensional
651 geodynamics. *Ore Geology Reviews*, **12**, 289-327.
- 652 Mäkelä, K. 1980. Geochemistry and origin of Haveri and Kiipu, Proterozoic strata-bound
653 volcanogenic gold-copper and zinc mineralizations from southwestern Finland. *Geological
654 Survey of Finland, Bulletin*, **310**. 79 p.
- 655 Nguyen, P. T., Cox, S. F., Harris, L. B. & Powell, C. McA. 1998. Fault-valve behaviour in
656 optimally oriented shear zones: an example at the Revenge gold mine, Kambalda,
657 Western Australia. *Journal of Structural Geology*, **20**, 1625-1640.
- 658 Nironen, M. 1989a. The Tampere schist belt: structural style within an early Proterozoic
659 volcanic arc system in southern Finland. *Precambrian Research*, **43**, 23-40.
- 660 Nironen, M. 1989b. Emplacement and structural setting of granitoids in the early
661 Proterozoic Tampere and Savo schist belts, Finland - implications for contrasting crustal
662 evolution. *Geological Survey of Finland, Bulletin*, **346**, 83 p.
- 663 Nironen, M. 1994. Structural control and (re)mobilization of the extinct Haveri Au-Cu
664 deposit, southern Finland. *Bulletin on the Geological Society of Finland*, **66**, 39-44.
- 665 Nironen, M., Elliott, B. A. & Rämö, O. T. 2000. 1.88-1.87 Ga post-kinematic intrusions of
666 the Central Finland Granitoid Complex: a shift from C-type to A-type magmatism during
667 lithospheric convergence. *Lithos*, **53**, 37-58.
- 668 Ohmoto, H. 1996. Formation of volcanogenic massive sulphide deposits: the Kuroko
669 perspective. *Ore Geology Reviews*, **10**, 135-177.
- 670 Ojakangas, R. W. 1986. An early Proterozoic metagraywacke-slate turbidite sequence: the
671 Tampere schist belt, southwestern Finland. *Bulletin of the Geological Society of Finland*,
672 **58**, 241-261.
- 673 Patchett, J. & Kouvo, O. 1986. Origin of continental crust of 1.9-1.7 Ga age: Nd isotopes
674 and U-Pb zircon ages in the Svecokarelian terrain of South Finland. *Contributions to
675 Mineralogy and Petrology*, **92**, 1-12.

676 Piessens, K., Muchez, Ph., Dewaele S., Boyce A., De Vos W., Sintubin, M., Debacker, T.
677 N., Burke, E. A. J. & Viaene, W. 2002. Fluid flow, alteration and polysulphide
678 mineralisation associated with a low-angle reverse shear zone in the Lower Palaeozoic of
679 the Anglo-Brabant fold belt, Belgium. *Tectonophysics*, **348**, 73– 92.

680 Poutiainen, M. & Grönholm, P. 1996. Hydrothermal fluid evolution of the Palaeoproterozoic
681 Kutemajärvi gold telluride deposit, Southwest Finland. *Economic Geology*, **91**, 1335–1353.

682 Rutland, R. W. R., Williams, I. S. & Korsman, K. 2004. Pre-1.91 Ga deformation and
683 metamorphism in the Palaeoproterozoic Vammala Migmatite Belt, southern Finland, and
684 implications for Svecofennian tectonics. *Bulletin of the Geological Society of Finland*, **76**,
685 93-140.

686 Seitsaari, J. 1951. The schist belt northeast of Tampere in Finland. *Bulletin de la*
687 *Commission Géologique de Finlande*, **153**, 121 p.

688 Sibson, R. H., Robert, F. & Poulsen, K.H. 1988. High-angle reverse faults, fluid pressure
689 cycling and mesothermal gold– quartz deposits. *Geology*, **16**, 551–555.

690 Simonen, A. 1953. Stratigraphy and sedimentation of the Svecofennidic early Archaean
691 supracrustal rocks in southwestern Finland. *Bulletin de la Commission Géologique de*
692 *Finlande*, **160**, 64 p.

693 Teagle, D. A. H., Norris, R. J. & Craw, D. 1990. Structural controls on gold-bearing quartz
694 mineralization in a duplex thrust system, Hyde– Macreas Shear zone, Otago schist, New
695 Zealand. *Economic Geology*, **85**, 1711– 1719.

696 Weihed, P., Arndt, N., Billström, K., Duchesne, J.-C., Eilu, P., Martinsson, O., Papunen, H.
697 & Lahtinen, R. 2005. Precambrian geodynamics and ore formation: the Fennoscandian
698 Shield. *Ore Geology Reviews*, **27**, 273-322.

699 Williams, I. S., Rutland, R. W. R. & Kousa, J. 2008. A regional 1.92 Ga tectonothermal
700 episode in Ostrobothnia, Finland: Implications for models of Svecofennian accretion.
701 *Precambrian Research*, **165**, 15-36.

702

703 **Figures**

704 **Fig. 1.** a) Simplified geological map of the Fennoscandian shield. Key: **1** Phanerozoic
705 sedimentary cover <0.57 Ga; **2** Caledonian rocks 0.6-0.4 Ga; **3** Sveconorwegian rocks
706 1.25-0.9 Ga; **4** Anorogenic rapakivi granites 1.65-1.4 Ga; **5** Late-Svecofennian granites
707 and migmatites 1.85-1.77 Ga; **6** Early Svecofennian pre- and synorogenic magmatic rocks
708 1.95-1.85 Ga; **7** Early Svecofennian supracrustal gneisses and migmatites 2.0-1.85 Ga; **8**
709 Early Palaeoproterozoic metasupracrustal rocks 2.5-1.9 Ga; **9** Archaean 3.2-2.5 Ga;
710 Modified from Korsman et al. (1997). b) Generalised geological map of the Tampere area.
711 Key: H = Haveri Cu-Au deposit; K = Kutemajärvi gold deposit; M = Myllyniemi; T =
712 Tohloppi; Ka = Karppi; P = Pohjala, TSZ = Tampere shear zone. The location of Fig. 2 is
713 shown with a rectangle. Modified from Kähkönen (1994, 1999), grid coordinates YKJ.

714 **Fig. 2.** Overall structure of the eastern part of the central TSB (see Fig. 1 for location and
715 for key). U-Pb zircon age data are indicated (in billions of years, see text for references).
716 Figure compiled and modified from Nironen (1989a; the cross section, individual units not
717 shown) and Kähkönen (1994, 1999; map section). Width of the view is c. 16 km, length c.
718 28 km. The location of the study area is shown with a rectangle.

719 **Fig. 3.** Representative photos of the various rock types in the subunits described in the
720 text. The VU subunits VU1-VU6 are shown in Figs. 3a-f, the SCU in Figs. 3g-l. a) VU1,
721 stratified felsic volcanic sandstone; b) VU2, stratified intermediate volcanoclastic sandstone
722 and conglomerate, erosion surface between the fine and the coarse layers indicates
723 southwards top-of-strata (stippled line); c) VU3, felsic plagioclase porphyry; d) VU4, ur-pl
724 porphyry; e) VU5, sandstone within the 'southern' volcanoclastics, subvertical section, view
725 toward W; this exposure is located north of the subunit VU6 and shows southward
726 younging strata; f) VU6, mafic and felsic tuffs; note the tight folds (F_1) with moderately (30-
727 40°) eastwards plunging fold axes; g-j) SCU rocks: g) stratified, in part graded sandstones-
728 microconglomerates, top of strata to the north (graded bedding); h) sandstones (in part
729 graded) with mica-rich mudstone layer, top-of-strata to north; note the well preserved soft-
730 sediment structures in the sandy layer (left of centre of the photo), and the NE-SW striking
731 S_2 schistosity crenulating the S_1 in the mudstone (right of the centre); i) non-sulphidized
732 mudstone; j) sulphide-bearing mudstone/schist; k-l) photomicrograph of the sulphide-
733 bearing schist, in plane-polarised light (k), and in cross-polarised light (l). The schistosity is

734 defined by brown mica. Note the abundant opaque minerals associated with the mica, and
735 the secondary fractures mineralised with red haematite and opaques. The opaques are
736 presumably mostly pyrite and chalcopyrite, observable in some hand samples. The non-
737 sulphidized schists do not contain abundant opaque minerals in thin section.

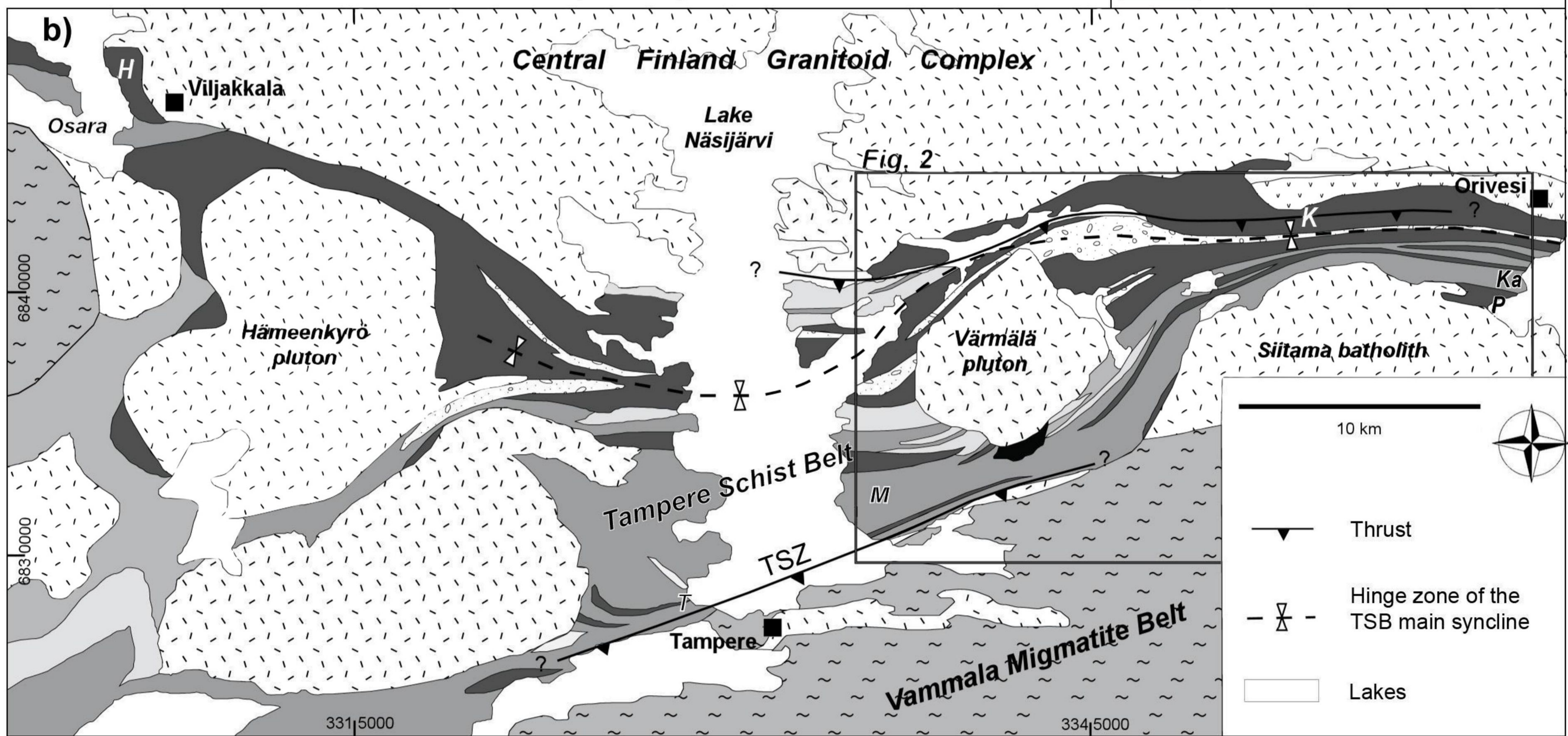
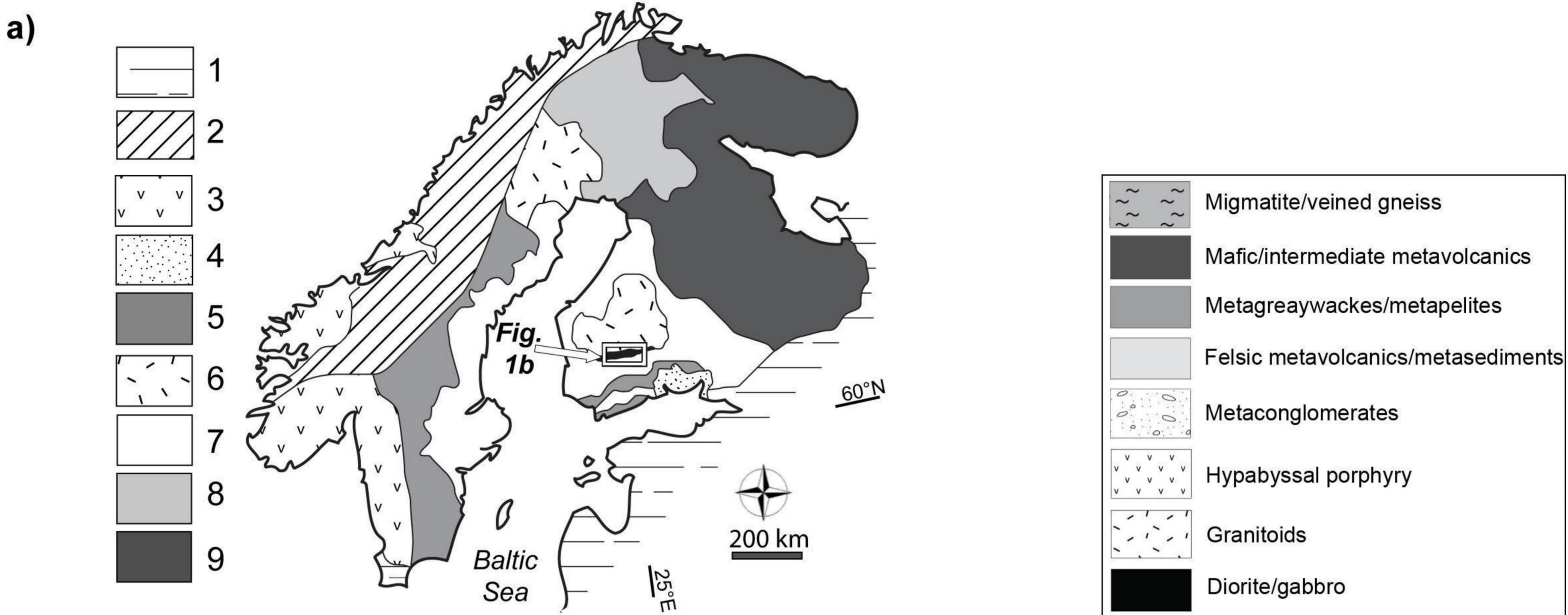
738 **Fig. 4.** Geological map of the study area, with the interpreted fault/shear zones. See text
739 for age references and discussion. The bedding (S_0) symbols are omitted for clarity; the
740 bedding is normally subparallel to S_1 schistosity. See also Fig. 5.

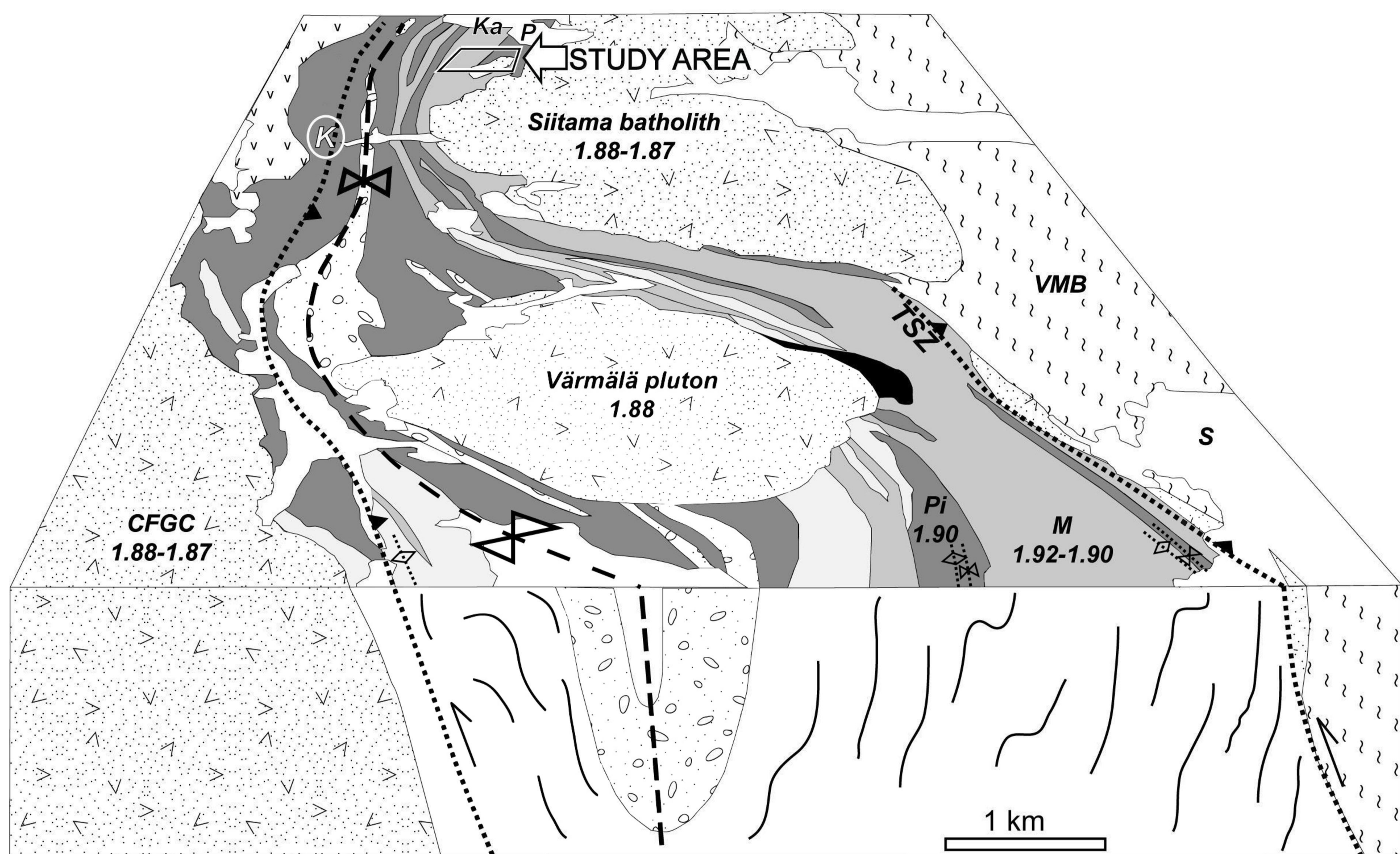
741 **Fig. 5.** A 3D block diagram presenting the interpreted structural geometry of the study
742 area. The legend is as in Fig. 4. The volcanic unit (VU) forms the “Pohjala syncline”,
743 defined by the opposite younging directions of the VU units. The VU is separated from the
744 siliciclastic unit (SCU) by a fault, interpreted as an early (D_1) thrust fault (see text and Fig.
745 7). The second, curved, right-lateral strike slip fault offsetting the siliciclastic subunits is
746 interpreted to have formed during a later deformation phase (D_2).

747 **Fig. 6.** Conceptual diagram of the interpreted depositional environment and the
748 succession for the metasiliciclastic and metavolcanic rocks of the SCU and VU,
749 respectively. The study area essentially represents a submarine (turbidite) fan close to
750 submarine and/or island arc volcanism.

751 **Fig. 7.** Conceptual tectonic model for the (southern) Tampere basin, not to scale. The
752 general crustal-scale tectonic evolution (from continental rift to a continental slope in Fig.
753 7a, to a subduction setting in Fig. 7b, and finally to a collisional setting in Fig. 7d) is not
754 shown. The internal architecture of the SCU and VU illustrated in Fig. 6 are omitted for
755 clarity. See text for a more detailed discussion. a) Deposition of the SCU sediments the
756 SCU units are deposited on a rift marginal basin (turbidite fans on a continental slope). b)
757 Subduction and associated volcanism initiates, and leads to basin inversion and a creation
758 of an (intra-arc) basin at c. 1.89 Ga. The volcanic activity leads to the deposition of the VU
759 (note that the entire volcanic package in the TSB is much thicker than inferred in the
760 figure). Some of the extensional faults are reactivated as (blind) reverse faults, leading to
761 open folding of the volcanic and turbiditic units along the southern margin of the basin. c) A
762 southward-propagating thrust, rooting layer-parallel within the SCU, develops to
763 accommodate further shortening of the TSB. Imbricate faults may form at this point within

764 the SCU (omitted from the model). Hydrothermal fluids sourcing from the base of the basin
765 infill (that possibly contains Haveri-type VMS mineralisations) may start channelling into
766 the thrust zone at this stage. Note that the northernmost thrust and the associated
767 synclines (outside the study area) are hypothetical, the thrust only inferred due to the
768 presence of sulphide schist horizons north of the study area. Within the Vammala basin,
769 another thrust with an opposite thrust direction (S-side-up) develops at the southern
770 margin of the basin (or possibly already during the previous stage). d) The continued
771 shortening leads to a tightening of the folds and in the rotation of the thrust faults into an
772 upright position. At depth, the thrusts may be folded together with the rock units that host
773 them. S_1 starts to form, trapping the sulphides of the first mineralisation phase. At the end
774 of this phase, further reverse movements along the old thrust planes are unlikely, but
775 strike-slip deformation can be expected. Fracture zones develop in the palaeothrust
776 planes, forming a trap to the secondary mineralisation within the fractures (see text). e)
777 Basin closure and S_1 is finalised, and the Siitama batholith intrudes along the interpreted
778 continuation of the TSZ.





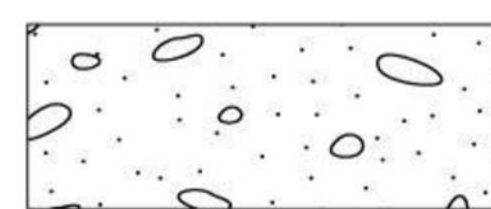
Metagreywackes and metapelites (turbidites)



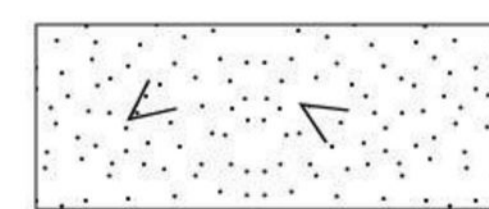
Mafic and intermediate metavolcanic rocks



Felsic metavolcanics and metasediments



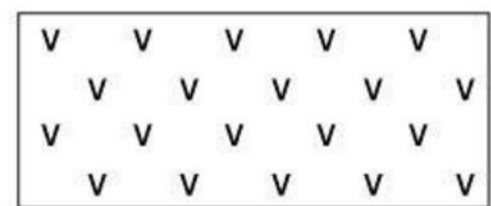
Metaconglomerates



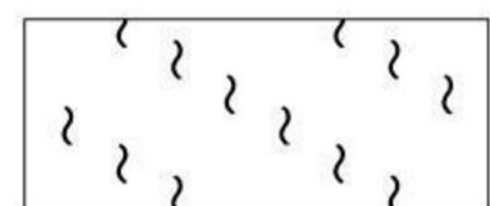
Granitoids



Diorite/gabbro



Hypabyssal porphyry



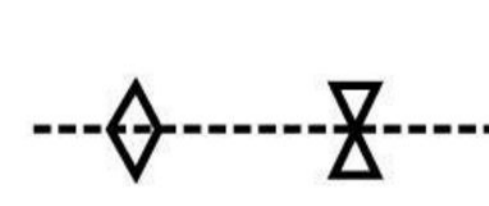
Migmatites and veined gneisses



Lake



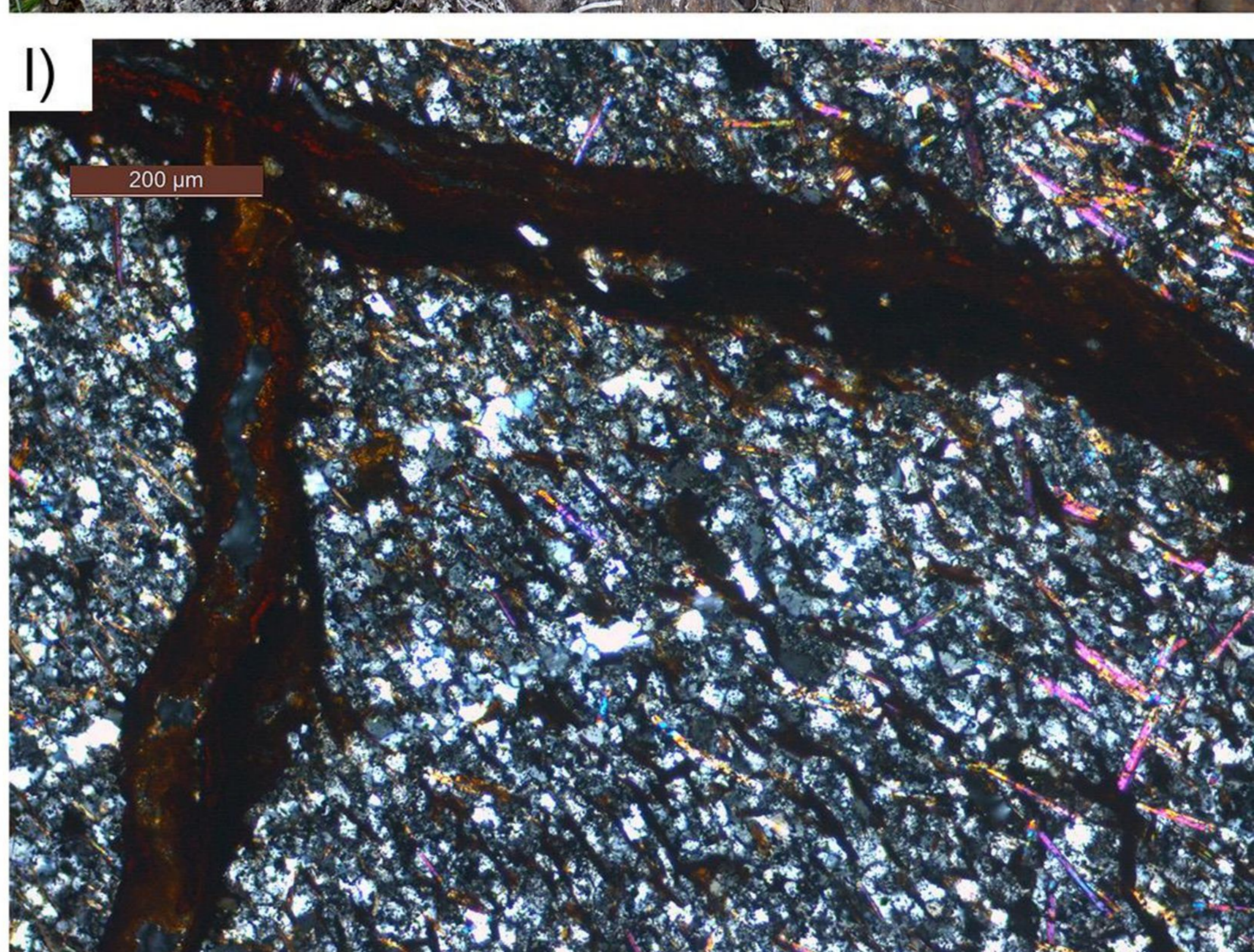
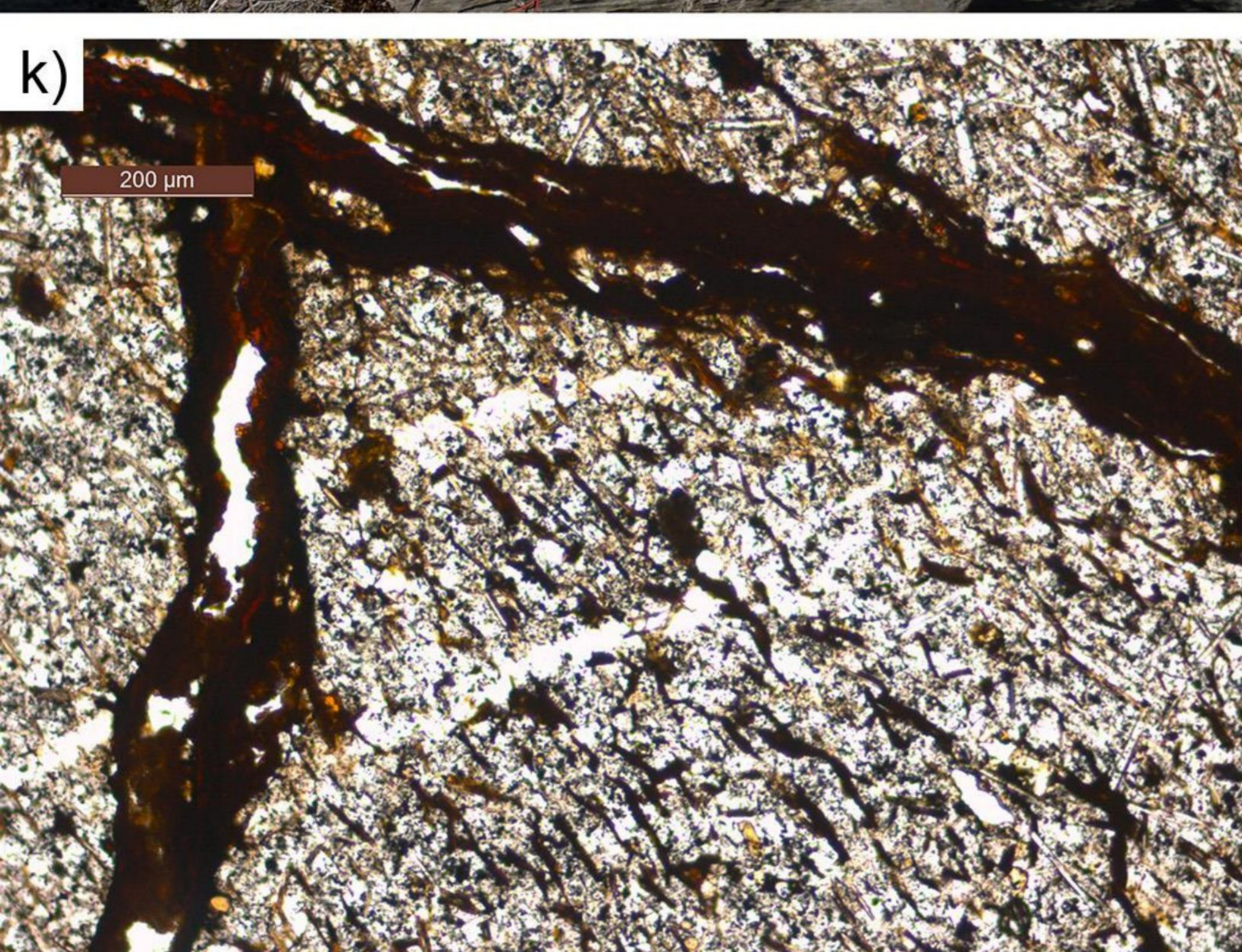
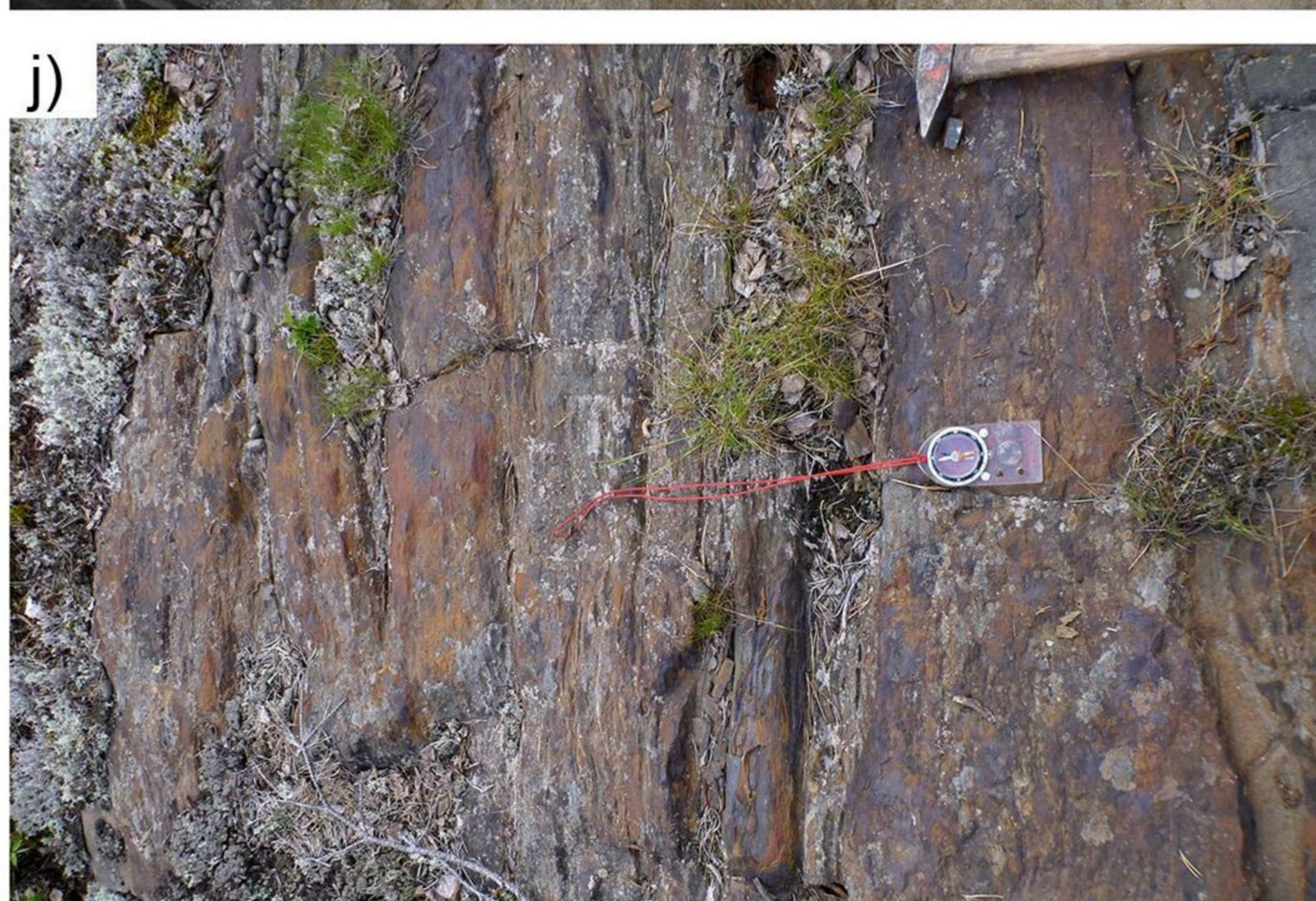
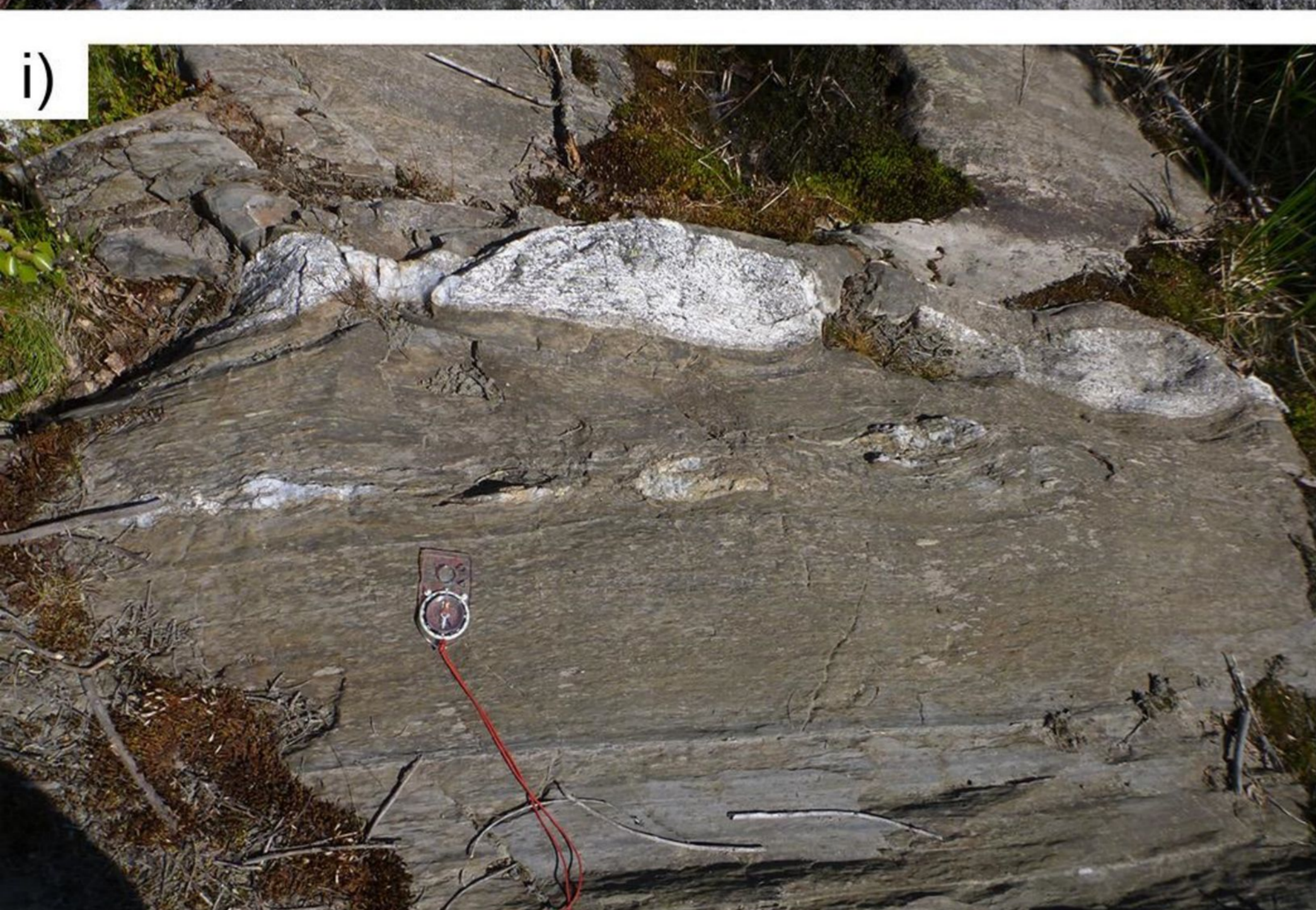
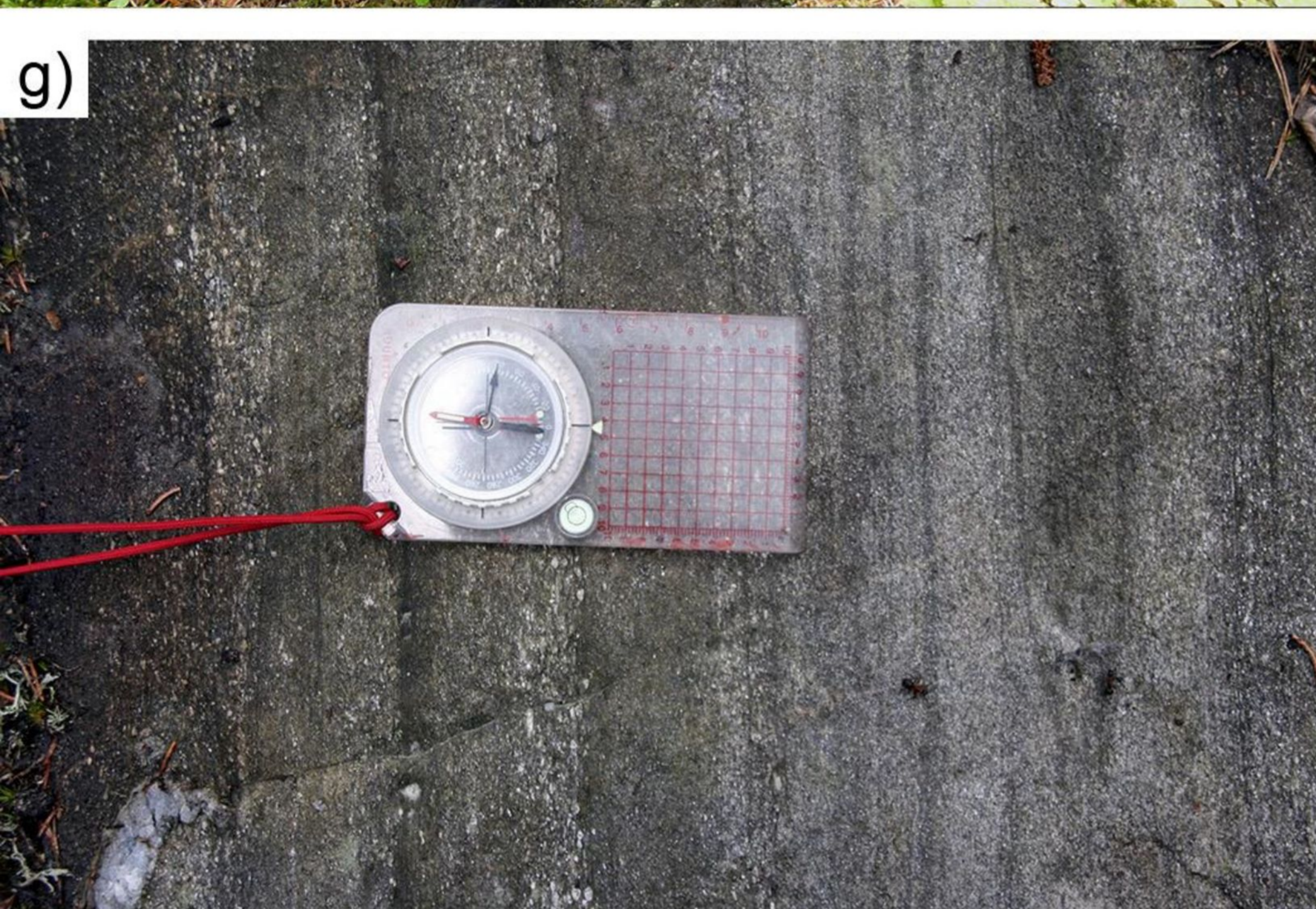
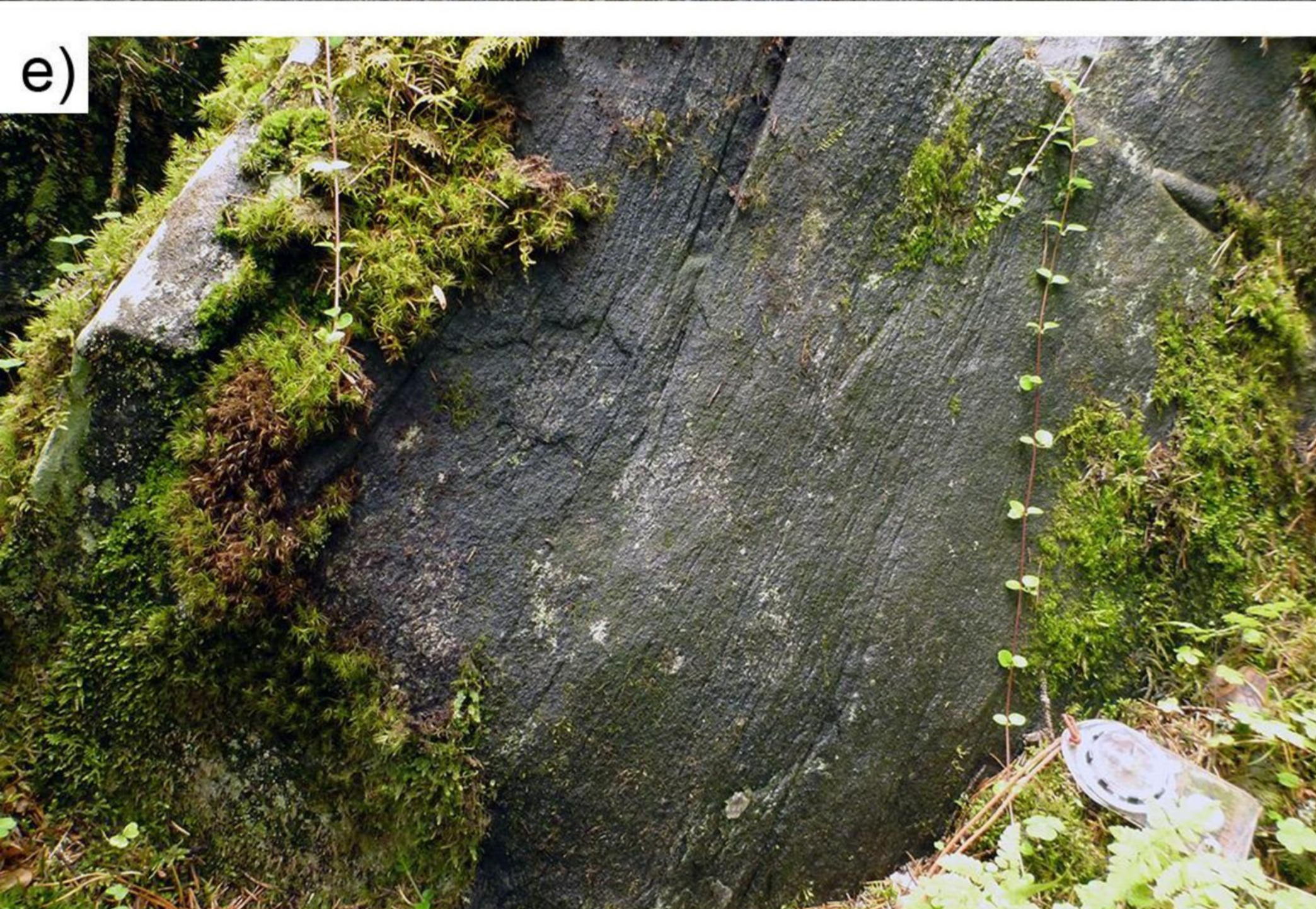
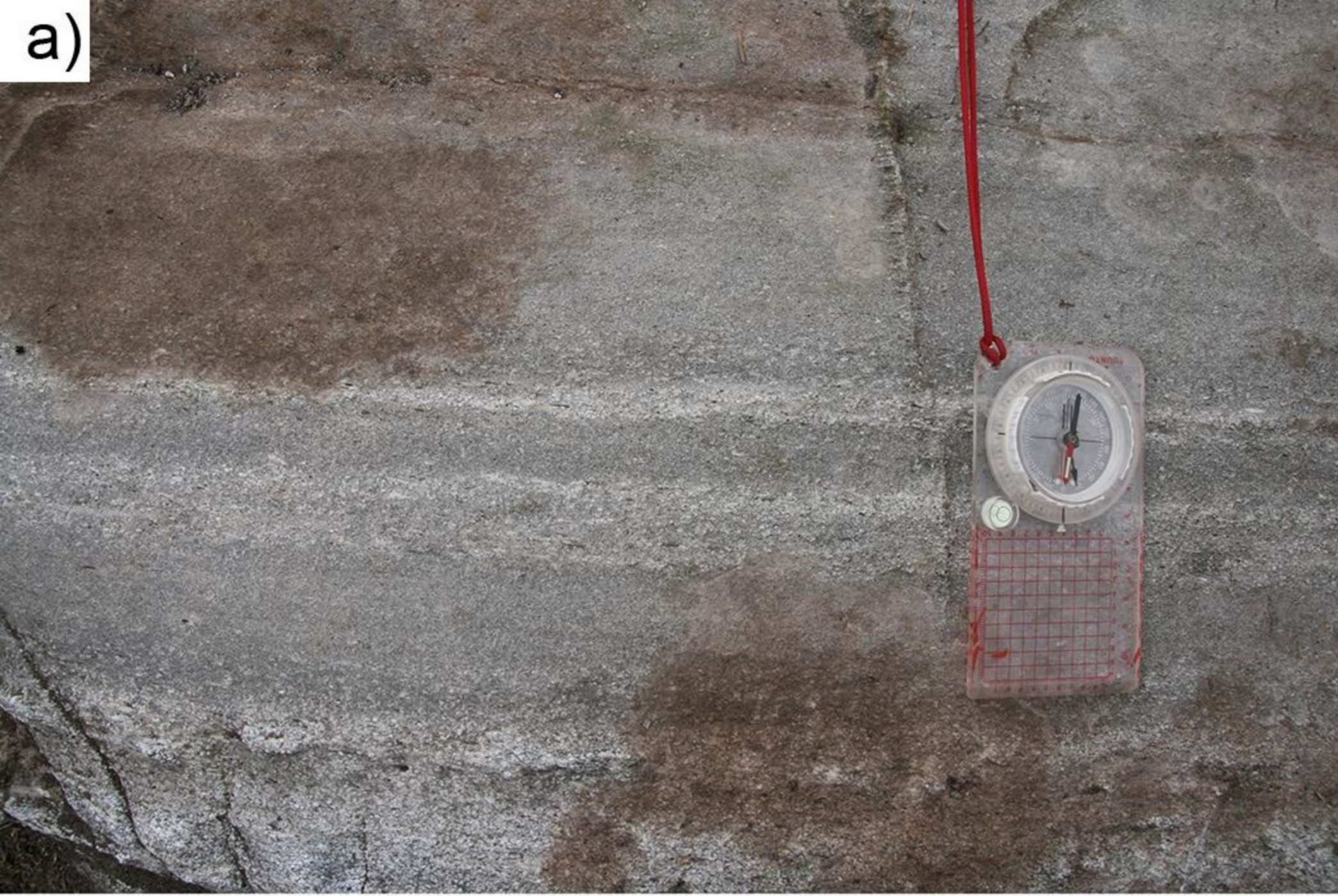
Main syncline

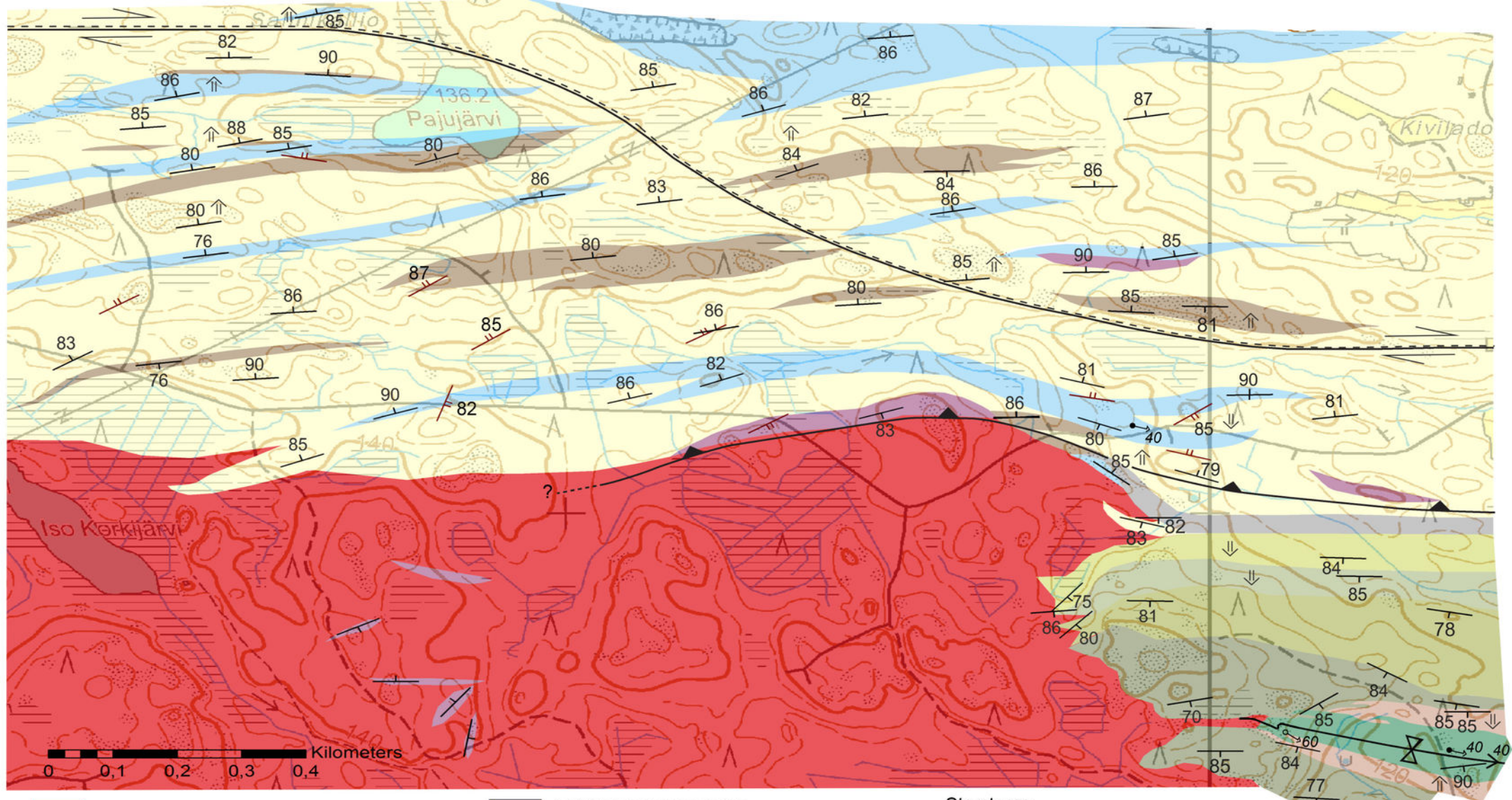


Minor anticline/syncline





Thrust zone













 Siitama batholith, c. 1.88-1.87 Ga





 Lapillistones and coarse tuffs interbedded with fine ash } Felsic volcanoclastics (VU1)
 Fine ash

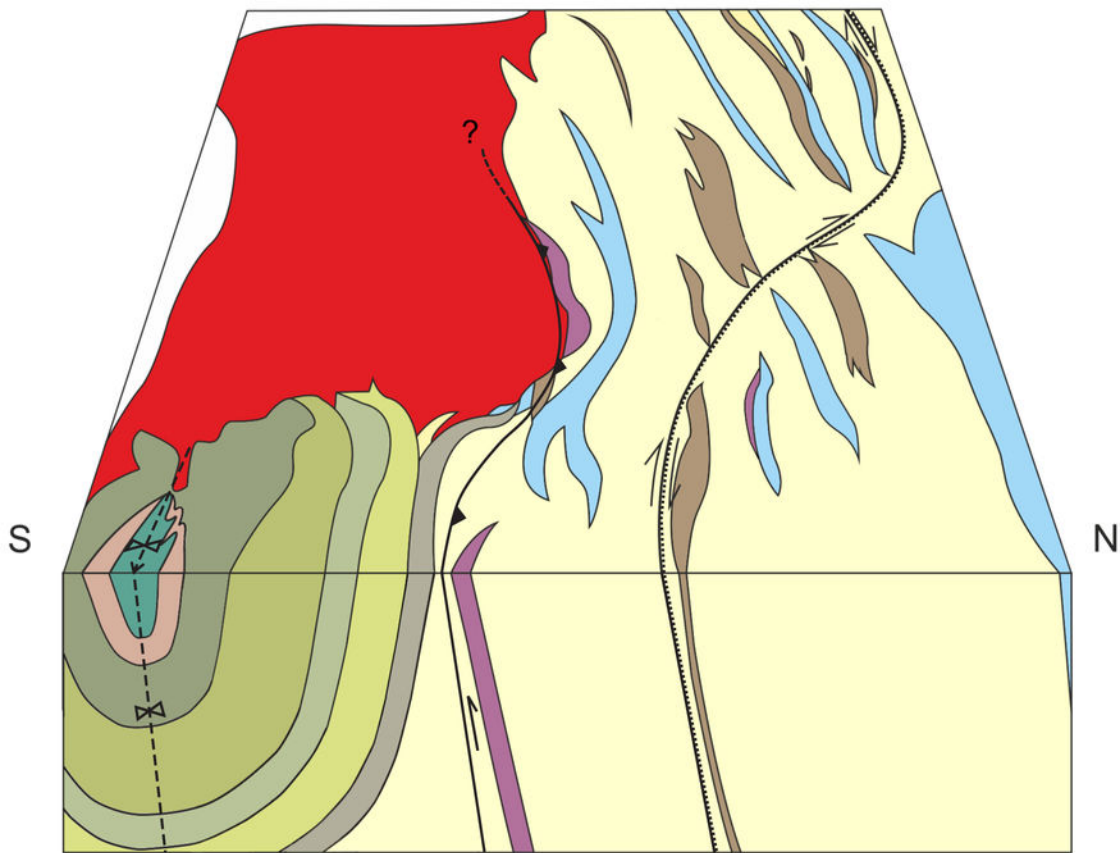
Structures

-  Top of strata
-  D1 thrust fault
-  Tilted F1 syncline with plunge
-  F1 fold, with axial plunge
-  S1 schistosity, with dip
-  D2 strike-slip fault
-  F2 fold, with axial plunge
-  S2 schistosity, with dip

VU subunits VU1-VU6 (oldest to youngest) c. 1.89 Ga

-  Mafic and felsic tuffs (VU6)
-  "Southern volcanoclastics" (VU5)
-  Uralite ± plagioclase porphyrys (VU4)
-  Felsic plagioclase porphyry (VU3)
-  Intermediate volcanoclastics (VU2)

- SCU rocks, c. 1.92-1.90 Ga**
-  Conglomerates
 -  Sandstones
 -  Schists
 -  Sulphide-bearing schists

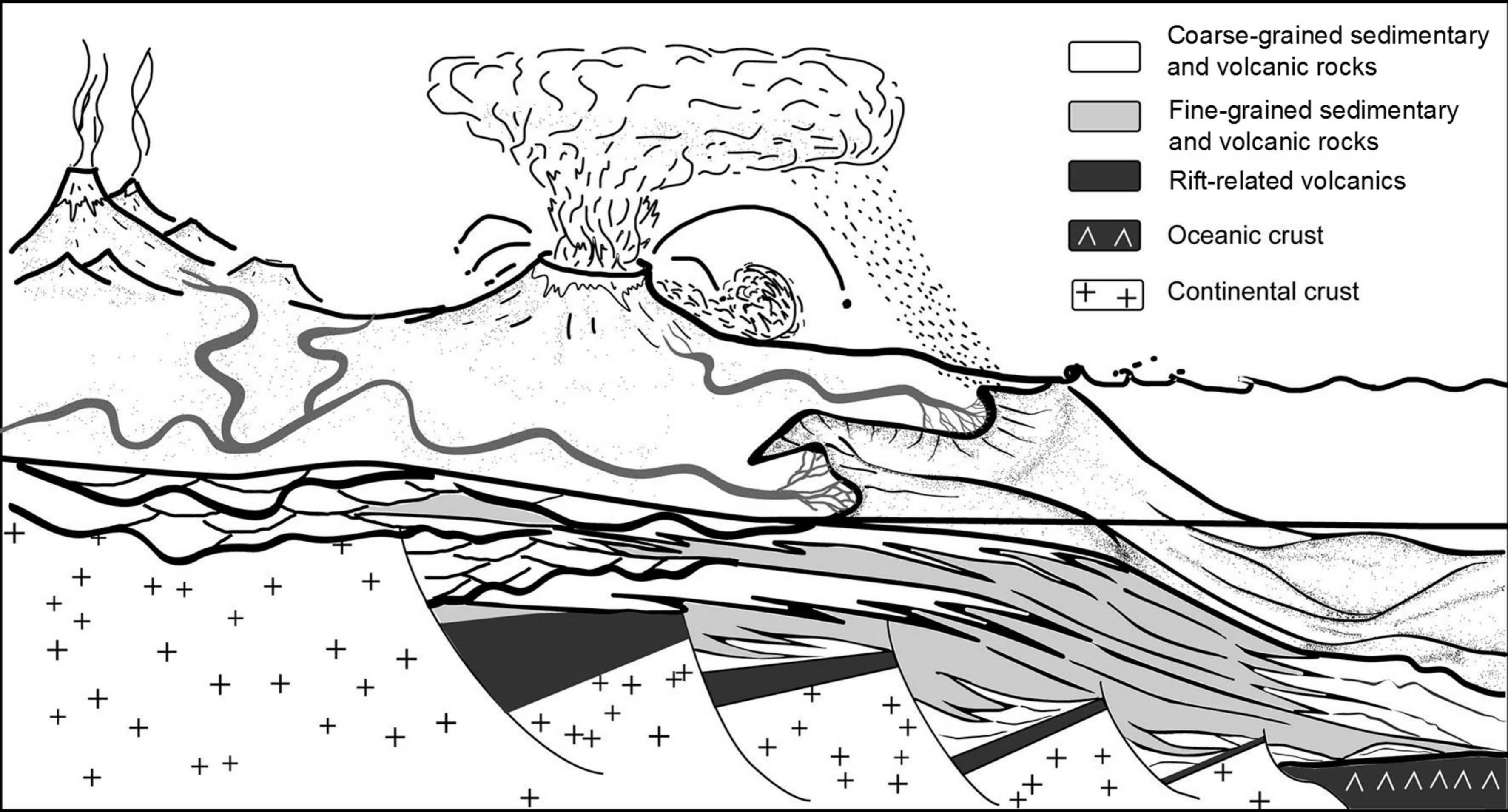


D1 thrust fault

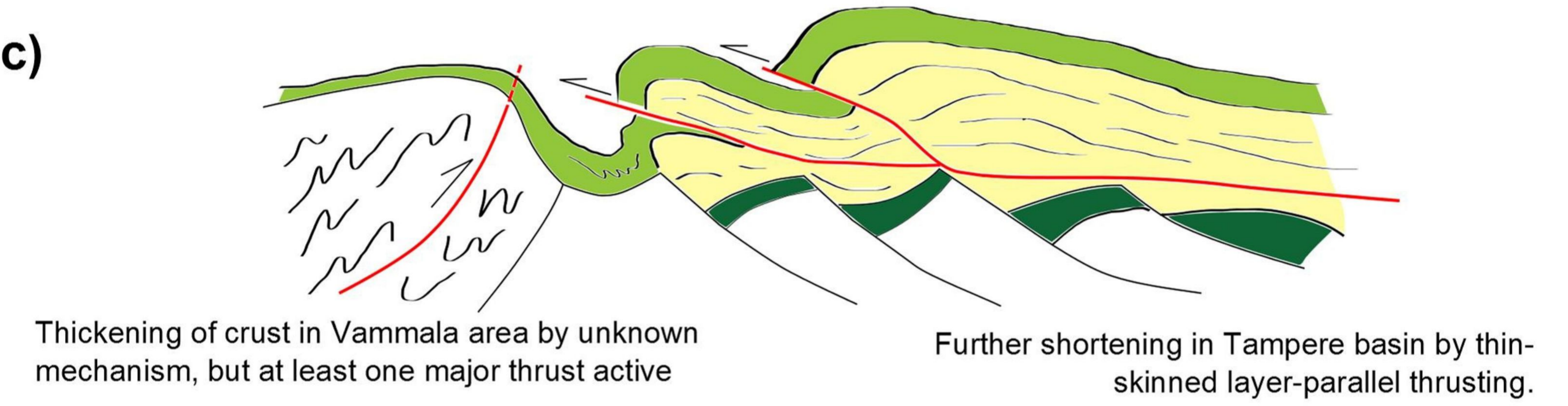
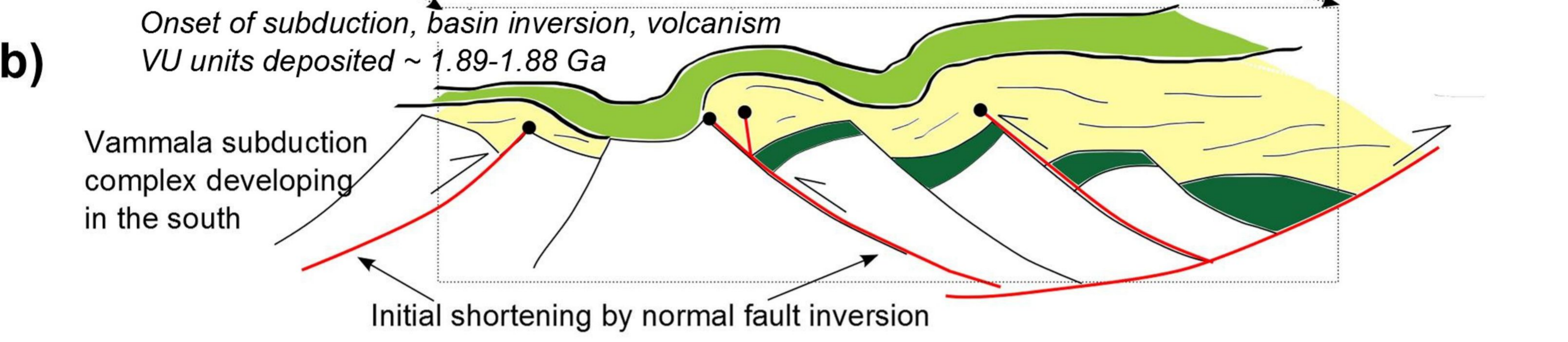
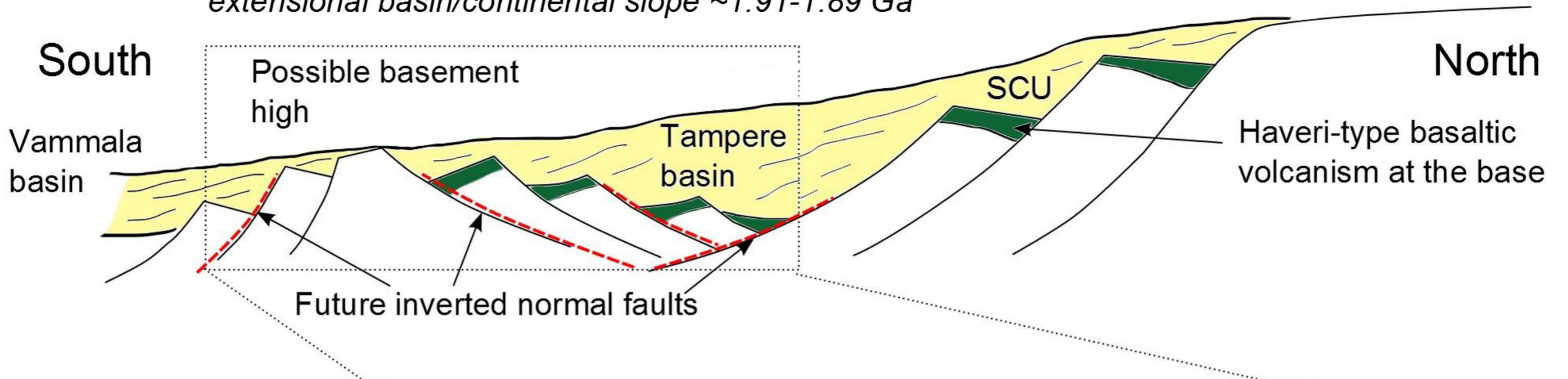
D2 strike-slip fault (implied)

Tilted F1 syncline

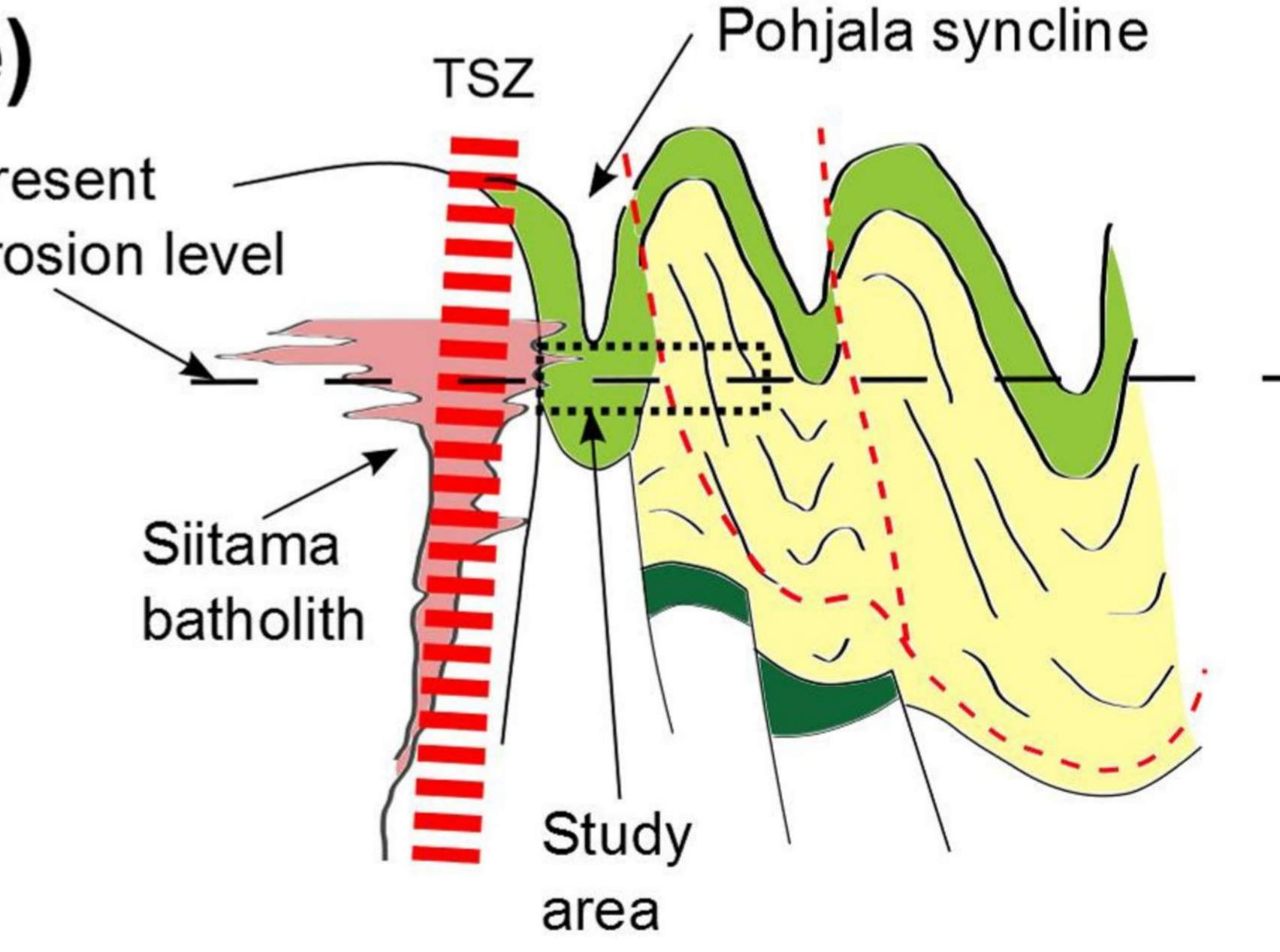
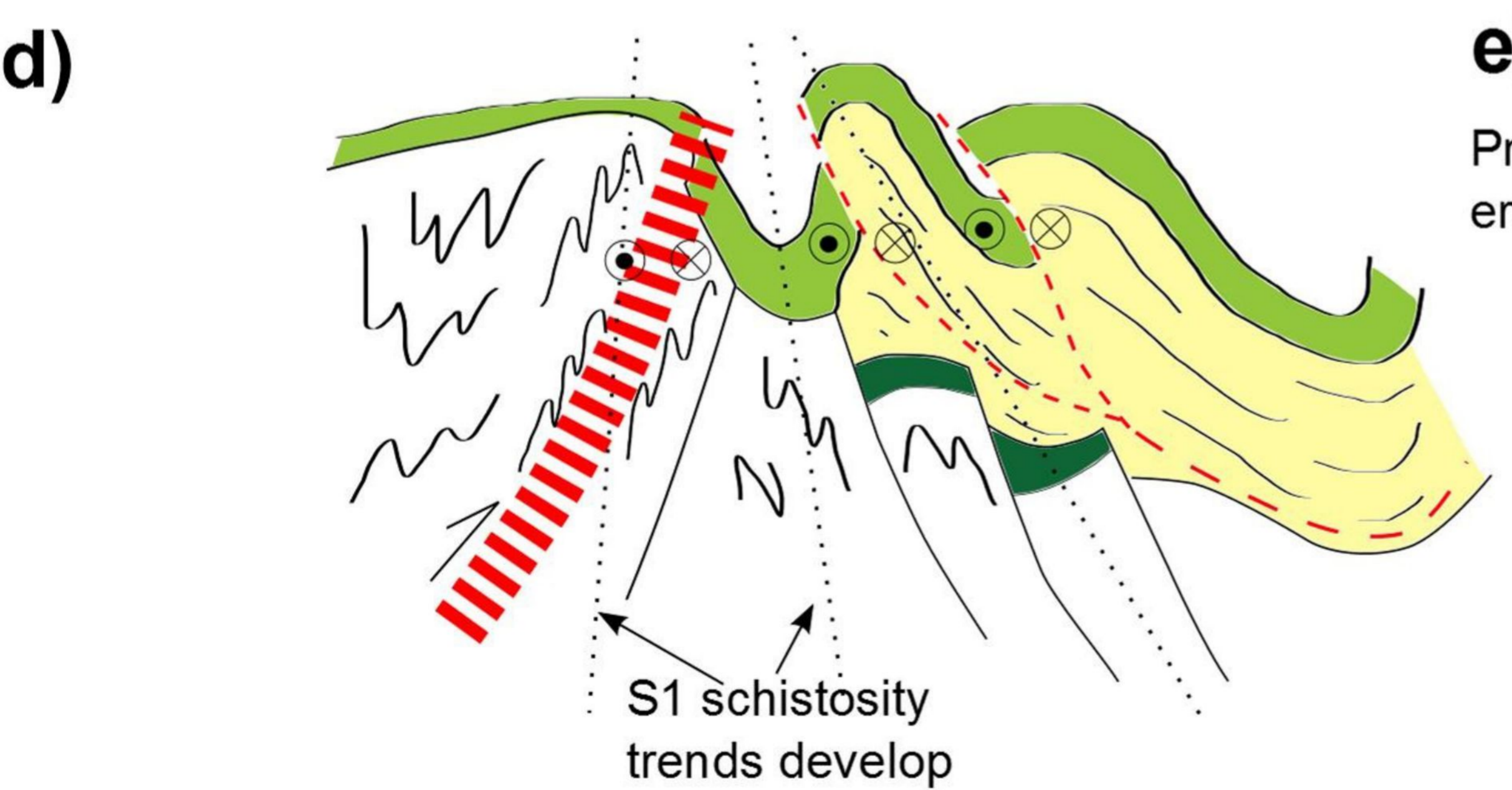
200 m



a) Continental rift/slope stage: deposition of SCU in an extensional basin/continental slope ~1.91-1.89 Ga



Basin closure stage ~1.88-.187 Ga



Tightening of folds, rotation of faults; strike-slip shearing and faulting along earlier thrust zones provides channels for mineralising fluids

Final basin closure, intrusion of Siitama batholith.

