

1 Revision 2

2 **Impact of fluorine on the thermal stability of phlogopite**

3 JIAQI SUN¹, YAN YANG^{1*}, JANNICK INGRIN², ZHONGPING WANG³, and QUNKE
4 XIA¹

5 ¹Key Laboratory of Geoscience Big Data and Deep Resource of Zhejiang Province,
6 School of Earth Sciences, Zhejiang University, Hangzhou, China

7 ²Univ.Lille, CNRS, INRA, ENSCL, UMR 8207, UMET, Unité Matériaux et
8 Transformations, F 59000 Lille, France

9 ³Department of Physics, University of Science and Technology of China, Hefei,
10 China

11 *Corresponding author: Yan Yang (yanyang2005@zju.edu.cn)

12 **ABSTRACT**

13 Knowledge of volatile cycling is vital to understand evolution of the planet and
14 the life it supports. Although it is gradually accepted that the mantle is a vast store
15 house of H₂O and other volatiles, the impact of co-existing volatiles on the thermal
16 stabilities of OH and the lattice of the host mineral is still poorly understood.
17 Phlogopite is one of the few hydrous minerals capable of carrying both water and
18 halogens into the mantle. Previous observations from both experiments and textural
19 relationships in natural samples have indicated that F-rich phlogopite can be stable
20 under ultra-high-temperature conditions. Here, the impact of F on the thermal stability
21 of phlogopite was investigated via XRD, Raman, and IR spectroscopy from room
22 temperature to 1000 to 1200 °C. Based on the experimental results from a F-poor and
23 a F-rich natural phlogopites, we show that about 4 wt% F can increase the breakdown
24 temperature of phlogopite by 100 °C under ambient pressure. The impact mechanism
25 mainly involves preventing OH and lattice softening² at high temperatures. This study
26 reveals the links between F and the behavior of OH and phlogopite lattice, which is
27 important for constraining volatile cycling as well as F's role in the physical and
28 chemical properties of the upper mantle.

29 **Keywords:** Fluorine, water, thermal stability, phlogopite, mantle

30

INTRODUCTION

31 Volatiles (e.g., hydrogen, nitrogen, carbon, and halogens) are not only present on
32 the surface of the Earth, but also in the deep Earth. Volatile exchange between the
33 surface and mantle plays a critical role in the evolution and habitability of the Earth.
34 Although a large number of studies have investigated volatile cycling in the deep Earth,
35 it is still unclear how these volatiles are transported. Volatile transport in the deep Earth
36 mainly depends on the thermal stabilities of the host minerals (van Keken et al. 2011).
37 Most hydrous minerals are not only important water carriers, but carry also other
38 volatiles such as nitrogen and the halogens (Williams et al. 1992; Bebout 1997; Hall
39 1999; Sadofsky and Bebout 2000; Yokochi et al. 2009; John et al. 2011; Palya 2011;
40 Kendrick et al. 2013; Halama et al. 2014; Pagé et al. 2016; Grützner et al. 2017a). Few
41 studies have paid attention to the possible links between the volatiles when evaluating
42 volatile transport by the host minerals. It has been noticed that fluorine (F) substituting
43 for hydroxyl (OH) can extend the pressure-temperature stability field of the hydrous
44 minerals such as amphibole, talc, and clinohumite (Foley 1991; Chibisov 2011;
45 Grützner et al. 2017a). Recent studies have also pointed out the intimate links between
46 nitrogen and hydrogen diffusion in phengite (Yang et al. 2017; Liu et al. 2019),
47 providing new insights into water release in the Earth's interior. Therefore, exploring
48 the effects of specific volatiles on water release from the host hydrous minerals is
49 greatly needed in order to understand the associated cycling of volatiles and H₂O.

50 Phlogopite is a sheet silicate frequently found in mantle nodules brought to the
51 Earth's surface mainly by alkaline basalt and kimberlite lavas (Frey and Prinz 1978;
52 Erlank et al. 1987; Fritschle et al. 2013). Fluorine has been experimentally proven to be
53 compatible in phlogopite, with a F partition coefficient between phlogopite and melt
54 larger than 1 (Flemetakis et al. 2021). Thus, phlogopite acts as an important reservoir
55 and carrier of water and F in the upper mantle despite the fact that it is volumetrically
56 insignificant when compared to olivine and pyroxenes (Frost 2006). Many studies have

57 focused on the *P-T* stability field of phlogopite in different systems. Most importantly,
58 observations from both experiments and textural relationships in natural samples have
59 shown that F-rich phlogopite can be stable up to ultra-high-temperature conditions
60 (Peterson et al. 1991; Dooley and Patiño Douce 1996; Motoyoshi and Hensen 2001).
61 However, the impact of F on the stability of phlogopite has never been explored further
62 and many unknowns still exist. For instance, what is the relationship between the
63 proportion of F (F/(F+OH) ratio) in phlogopite and the breakdown temperature down to
64 the atomic level? In addition, it has been reported that dehydrogenation could happen at
65 low temperatures, such as 500 °C, before the breakdown of phlogopite (Chon et al.
66 2006; Ventruti et al. 2009; Zema et al. 2010). But it remains unclear whether F has an
67 impact on the dehydrogenation of phlogopite. These unknowns limit a full
68 understanding of volatile release from phlogopite.

69 In this study, we systematically investigate the effect of F on the thermal stability
70 and dehydrogenation of phlogopite, using high temperature Raman spectroscopy and
71 X-ray diffraction (XRD) to monitor the breakdown processes of the lattice, and high
72 temperature Fourier transform infrared (FTIR) spectroscopy to reveal the behavior of
73 OH with increasing temperature.

74 MATERIALS AND METHODS

75 Samples

76 The two starting samples are phlogopite crystals from two different localities
77 described in Li et al. (2017). The F-rich phlogopite is a 10 × 5 mm, golden-brown
78 crystal from Franklin, New Jersey, United States. The F-poor phlogopite is a 70 × 50
79 mm, reddish-brown crystal from Badakhshan, Afghanistan. The chemical
80 compositions were obtained using a shimadzu electron probe microanalyzer (EPMA
81 1720) at 15 kV and 10 nA beam with a 5 μm beam diameter. Counting time was 10 s on
82 the peak and 10 s on the background for all elements except F, which was 60 s on peak
83 and 30 s on background. Four to seven points were measured on each phlogopite grain.

84 The average chemical compositions of the two samples are listed in Table 1. Structure
85 formula was calculated on the basis of 22 cation charges, assuming an anion framework
86 of 10 oxygens and 2(OH+F+Cl), and all Fe as Fe²⁺ (Dymek 1983; Rieder et al. 1998).
87 This results in the following formula
88 $(K_{0.963}Na_{0.015})(Mg_{2.677}Fe_{0.166}Al_{0.059}Mn_{0.002}Ti_{0.015}Cr_{0.009}Ni_{0.003})(Si_{3.067}Al_{0.933})O_{10}[F_{0.905}C$
89 $l_{0.016}OH_{1.079}]$ for the F-rich phlogopite ($X_F = 0.45$), and
90 $(K_{0.911}Na_{0.047})(Mg_{2.723}Fe_{0.026}Al_{0.111}Mn_{0.001}Ti_{0.062}Cr_{0.002}Ni_{0.001})(Si_{2.955}Al_{1.045})O_{10}[F_{0.053}C$
91 $l_{0.002}OH_{1.945}]$ for the F-poor phlogopite ($X_F = 0.026$).

92 **X-ray diffraction spectroscopy (XRD)**

93 To monitor the breakdown processes of the two samples, high temperature X-ray
94 diffraction measurements were carried out on a PANalytical X'Pert PRO Multipurpose
95 Diffractometer (MPD), powered by a Philips PW3040 generator and equipped with an
96 X'Celerator detector using CuK α radiation ($\lambda=1.5406 \text{ \AA}$). The measurements were
97 operated at 40 kV and 40 mA with 2 θ ranging from 7 to 70°. Both samples were finely
98 pulverized in an agate mortar for over 3 h and then passed through a 74 μm sieve.
99 Powders were mounted on a platinum resistive heating stage and placed in an evacuated
100 chamber. For each run, the temperature was increased from room temperature to
101 1100 °C at 100 °C intervals and with a heating rate of 10 °C/min. The temperature was
102 monitored by a TCU1000N (2604 unit) controller with a temperature error of less than
103 1 °C. The XRD spectra were collected at each temperature. Each 100 °C temperature
104 step consisted of a 15 min dwell time and a 30 min detection time. All data were
105 analyzed using the Jade 6 software.

106 **Raman spectroscopy**

107 To further characterize phlogopite breakdown processes, high temperature Raman
108 spectra of the phlogopites in the frequency range 50 to 1200 cm^{-1} were collected using a
109 LABRAM-HR spectrometer with an 1800 grooves/mm grating. The sample was
110 excited by a 532 nm green light from a Spectra Physics Ar ion laser for a total time of

111 100 s for each data point. Single-crystal silicon was used as the reference spectrum.
112 High temperatures were produced by a Linkam TS1500 heating stage, equipped with a
113 resistance heater and a S type thermocouple. Temperatures were determined with an
114 uncertainty of less than 1 °C. The powder samples (grain size < 74 µm) were heated
115 while continuously purged by Ar and held for 30 min at the desired temperatures of 900,
116 1000, 1100, and 1200 °C. Raman spectra were measured on the samples after they have
117 been quenched to room temperature. In order to investigate the evolution of lattice
118 vibrations as a function of increasing temperature, temperature-dependent Raman
119 spectra, from room temperature to 1000 °C, were measured on crystal samples in the
120 heating stage. The heating rate was set to be 10 °C/min, and an idle time of 5 min was
121 applied to each experimental temperature.

122 **Fourier transform infrared (FTIR) spectroscopy**

123 To reveal the behavior of OH and dehydrogenation in the two samples at high
124 temperatures, thin cleavage flakes perpendicular to the (001) plane with a thickness of
125 50 to 130 µm were used for the temperature-dependent FTIR measurements. Polarized
126 and unpolarized FTIR spectra in the frequency range 1000-4000 cm⁻¹ were collected
127 with an IR beam direction perpendicular to the (001) plane (i.e., to the layers) using a
128 Nicolet iS50 FTIR spectrometer coupled with a Continuum microscope. A KBr
129 beam-splitter and a liquid-nitrogen cooled MCT-A detector were applied. A total of
130 128 scans were accumulated for each spectrum at a 4 cm⁻¹ resolution. The aperture size
131 was set to 50 × 50 µm.

132 High temperatures were produced using an Instec HS1300 heating stage with CaF₂
133 windows, equipped with a resistance heater and a S type thermocouple. The phlogopite
134 grains were placed on a sapphire plate in the heating stage continuously purged with N₂
135 gas. The sample temperature was determined with an uncertainty of less than ±1 °C.
136 The FTIR measurements were performed from room temperature to 100 °C and then
137 1000 °C using an interval of 100 °C. Samples were heated at a rate of 10 °C/min and

138 held at the desired temperature for 5 min. Background spectra and sample spectra were
139 collected on the same selected area, respectively, for each grain and each temperature.

140 **RESULTS**

141 **Stability of F-rich and F-poor phlogopites at high temperatures**

142 Studies on the stabilities of phlogopite in complex systems have long been carried
143 out (Frost 2006 and references therein). They proposed that the stability field and
144 decomposition products depend on the initial chemical compositions of the systems. In
145 this study, the focus is on the role of F during the breakdown of individual phlogopite
146 crystals.

147 X-ray diffraction patterns at elevated temperatures are shown in Figure 1. With
148 increasing temperature, the intensities of the original diffraction peaks (005) and (003)
149 of the phlogopite decrease, whereas the intensities of the diffraction peaks for forsterite,
150 such as (222), (031) and (120), grow concomitantly (Figs. 1c-d). Both phlogopite
151 samples began to decompose at 1000°C, as indicated by the appearance of the forsterite
152 diffraction peaks. Moreover, most of the phlogopite diffraction peaks disappear at
153 1100 °C for the F-poor phlogopite, whereas the sharp diffraction peaks (005) and (003)
154 persist for the F-rich phlogopite, indicating the different degrees of breakdown for the
155 two samples. Comparison of the XRD patterns from the two samples before and after
156 the run shows more clearly the different degrees of breakdown (Fig. 1e). These X-ray
157 diffraction results demonstrate that breakdown starts at 1000 °C for both samples, but
158 the F-poor phlogopite completely decomposes at 1100°C, whereas the F-rich
159 phlogopite has only partly decomposed at the same temperature. The breakdown
160 process of the two phlogopites was further characterized by Raman spectra from the
161 quenched powder samples after annealing at different temperatures (Fig. 2). In
162 accordance with observations from the XRD results, the Raman spectra show that both
163 phlogopites start to decompose at 1000 °C as illustrated by the appearance of the
164 characteristic peaks of forsterite at 825 and 857 cm⁻¹. But the breakdown degrees were

165 different between the two samples. After annealing at 1100 °C, the F-poor phlogopite
166 completely decomposed as indicated by the disappearance of the initial characteristic
167 peaks of phlogopite, while there still exist remnants of these peaks for the F-rich
168 phlogopite. By 1200 °C, the F-rich phlogopite had completely decomposed, according
169 to the Raman spectra in Figure 2, indicating that the breakdown process is faster for the
170 F-poor phlogopite than for the F-rich one.

171 **Lattice of the F-rich and F-poor phlogopites at various temperatures**

172 The thermal behavior of the crystal lattice for the two samples were investigated
173 using temperature-dependent Raman spectra in the range of 50-1200 cm^{-1} up to 1000°C
174 (Fig. 3). Raman modes of the two phlogopite samples are similar to those observed in
175 previous studies (Loh 1973; Clemens et al. 1987; Tlili et al. 1989; McKeown et al. 1999;
176 Tutti et al. 2000), but minor differences exist in band positions of some modes (Fig. 3a).
177 The predominant modes are 193, 281, 327, 358, and 682 cm^{-1} for the F-poor phlogopite,
178 and 193, 277, 320, and 682 cm^{-1} for the F-rich phlogopite. Modes at 193, 277 to 281,
179 320 to 327, and 682 cm^{-1} appear to correspond to the same modes in both phlogopites.
180 The intensity of the 358 cm^{-1} mode is much stronger for the F-poor phlogopite than the
181 F-rich one, which is in agreement with the fact that it has only been reported in the
182 synthetic pure OH phlogopite (Clemens et al. 1987). Modes at frequencies less than 350
183 cm^{-1} are generally classified as the lattice modes where long-range T_4O_{10} sheet motions
184 can mix with M2 and (F, OH) displacements. Modes between 350 and 850 cm^{-1} have
185 bond stretch and bend motions with the T_4O_{10} sheets and M2 octahedra. The mode at
186 682 cm^{-1} is assigned to O3-T-O1 band, which reflects motion along the $z(c)$ direction in
187 the structure (McKeown et al. 1999).

188 No new band in the spectra appears nor disappears at elevated temperatures (Figs.
189 3b-c). Most lattice modes shift continuously toward lower wavenumbers without
190 apparent turning points, except for some small accidents around 400 to 500 °C for the
191 193 and 682 cm^{-1} modes (Figs. 3d-h). These small breaks do not support any structural

192 phase transitions occurring during the heating process, although phlogopite has been
193 found to undergo a structural phase transition at temperatures of around 400 °C (Tutti
194 and Lazor 2008). Similarly, Zhang et al. (2016) found that the phonon modes of
195 phlogopite all exhibited gradual changes with increasing temperature except for the
196 Al-O and Si-O stretching vibrations, which show a discontinuity near 327 °C. This was
197 suggested by the authors to be related to the alteration of the OH environment. The
198 small accidents around 400 to 500 °C for the 193 and 682 cm⁻¹ modes observed in this
199 study could be related to dehydrogenation processes, which is evidenced by the
200 behavior of OH at elevated temperatures (Fig. 4).

201 **Behavior of OH in F-rich and F-poor phlogopites as a function of temperature**

202 Figures 4a-b show the temperature-dependent unpolarized IR spectra of OH
203 vibrations in the 3500 to 3750 cm⁻¹ range up to 1000 °C for the F-rich and F-poor
204 phlogopites. For both samples, the absorptions of OH stretching bands recorded
205 through IR beam perpendicular to the (001) plane have isotropic properties, thus, the
206 spectra with polarizer rotating to different angles are similar (Fig. 4c). Some
207 differences in OH wavenumbers between the two samples can nevertheless be observed.
208 For the F-poor phlogopite, there are mainly three OH bands at 3714, 3671, and 3621
209 cm⁻¹. In contrast, three OH bands at 3707, 3665 and 3599 cm⁻¹ are shown in the IR
210 spectra of the F-rich phlogopite. Vedder (1964) divided the OH bands in phlogopite
211 into three groups: type N band (normal) corresponding to OH bonding to three
212 octahedrally coordinated divalent cations, type I band (impurity) related to OH bonding
213 to two divalent and one trivalent cations, and type V band (vacancy) due to an OH
214 adjacent to an octahedral vacancy. All previous studies assign the bands at 3714 and
215 3707 cm⁻¹ to the type N OH (Vedder and Wilkins 1969; Zhang et al. 2016). Bands at
216 3671 and 3665 cm⁻¹ were assigned to the type I OH, relating to Mg₂(Al, Fe³⁺)-OH
217 species (Robert and Kodama 1988; Ventruti et al. 2009; Zhang et al. 2016). The bands
218 at 3621 and 3599 cm⁻¹ are generally associated with the type V OH, corresponding to a

219 vacancy with two divalent ions in the octahedral sheet and a vacancy with one divalent
220 ion and one trivalent ion, respectively (Vedder 1964). The relative intensity of the type
221 N OH to the other two groups is larger for the F-poor phlogopite than the F-rich
222 phlogopite, which is consistent with the conclusion that F prefers to substitute for the
223 type N OH based on analyses of synthetic F-rich phlogopites (Papin et al. 1997).

224 Since the IR spectra exhibit isotropic features independent of the polarizer
225 orientation in (001) plane, we used only unpolarized IR spectroscopy to study the OH in
226 phlogopite at elevated temperatures. Comparison between the IR spectra of the F-poor
227 phlogopite before heating and after quenching shows no evident decrease of the
228 absorbance for the type N OH (3714 and 3707 cm^{-1}), and a small decrease for the type I
229 OH (3671 and 3665 cm^{-1}), while the type V OH (3621 and 3599 cm^{-1}) almost
230 disappears. These variations in the absorbances suggest dehydrogenation, especially
231 for the type V OH (Figs. 4a-b). We could not tell from the spectra whether
232 dehydrogenation happened at elevated temperatures for the F-rich phlogopite, because
233 the spectrum of the quenched sample from $1000\text{ }^{\circ}\text{C}$ is seriously distorted. But
234 dehydrogenation could be inferred from the evolution of the OH band wavenumbers
235 with temperature. The three OH band groups shift to lower wavenumbers with
236 increasing temperature, but with discontinuities at 400 to $500\text{ }^{\circ}\text{C}$ for the type I OH and
237 the type V OH bands (Fig. 4d). The abrupt shift in the temperature-induced OH
238 wavenumber for type I OH and type V OH, observed around 400 to $500\text{ }^{\circ}\text{C}$, occurs in
239 the temperature range where dehydrogenation is generally observed in phlogopite (e.g.,
240 Vedder and Wilkins 1969; Tutti et al. 2000; Zhang et al. 2016). Thus, both phlogopites
241 experienced dehydrogenation of type I OH and type V OH at around 400 to $500\text{ }^{\circ}\text{C}$
242 during heating, further confirming the IR spectra before heating and after quenching,
243 although we cannot confirm it with the spectrum from the quenched F-rich
244 phlogopite.

245

DISCUSSION

246 **The effects of F on the thermal stability of phlogopite**

247 It has already been documented that F may markedly extend the thermal stability
248 of phlogopite. For example, the experimental study by Hensen and Osanai (1994)
249 demonstrated that synthetic phlogopite with about 6 wt% F decomposed at 1045 °C and
250 9 kbar. Moreover, Dooley and Patiño Douce (1996) found that the synthetic phlogopite
251 with 8.7 wt% F can be stable at 1263 °C at 10 kbar. Accordingly, F-rich phlogopite was
252 reported as a stable constituent in ultra-high-temperature rocks (Motoyoshi and Hensen
253 2001). To reveal the relationship between F and the breakdown temperature of
254 phlogopite, we plotted the breakdown temperature to proportion of F (F/(F+OH) ratio)
255 in Figure 5 based on the data from this study and previous studies. It shows that
256 addition of F can increase the breakdown temperature, confirming the higher stabilities
257 of F-rich phlogopites previously observed. The F-poor and F-rich phlogopite samples
258 start to decompose at 1000 °C under ambient pressure. But the breakdown process is
259 faster for the F-poor phlogopite compared to the F-rich phlogopite, since the complete
260 breakdown temperature is at least 100 °C lower for the F-poor phlogopite compared
261 with the F-rich phlogopite.

262 It should be noted that in addition to F, the chemical compositions of these two
263 phlogopites differ. Previous studies have suggested that Ti can stabilize biotite both at
264 high *P-T* and at low H₂O activity (Henry et al. 2005; Harlov et al. 2006; Hansen and
265 Harlov 2007). As basically the Mg endmember of the biotite series, phlogopite should
266 also be stabilized by Ti. To check the effect of Ti on the thermal stability of phlogopite,
267 we plotted the Ti contents vs. the breakdown temperatures of the samples from Figure 5
268 in the supplementary material (Fig. S1). Indeed, a high Ti content can increase the
269 thermal stability of phlogopite as well. Although the F-poor phlogopite contains more
270 Ti than the F-rich phlogopite, F content still plays an important role in phlogopite
271 stability as shown in Figure 5. Therefore, the actual effect of F on phlogopite could be
272 even stronger if the Ti content was the same in both phlogopite samples. On the other

273 hand, pressure effect on the mineral stability is also important. To check the effect of F
274 on the breakdown temperature of phlogopite under different pressures, compiled data
275 previously reported for phlogopite and other hydrous minerals at high pressure is
276 plotted in Figure 5. It is shown that F generally plays an important role on the thermal
277 stabilities of the host minerals at the same pressure. For phlogopite, the breakdown
278 temperature increases with the increasing F content no matter whether at ambient
279 pressure or at 7 to 15 kbar. Moreover, the breakdown temperatures of phlogopites with
280 the same amount of F are higher at ambient pressure than at high pressures. Considering
281 the impact of Ti on the phlogopite samples with higher Ti contents used in experiments
282 at 7 to 15 kbar, the breakdown temperatures may be even lower than those shown in
283 Figure 5. Thus, it seems that pressure plays a negative role in the thermal stability of
284 phlogopite, which is the reverse of F effect. However, this pressure effect for
285 phlogopite is not applicable to other minerals. For example, for clinohumite, with a F
286 fraction of 0.5, can be stable up to 1100 °C at 30 kbar (Fig. 5), and up to 1600 °C at 10
287 GPa as inferred from Grützner et al. (2017).

288 **The effects of F on the breakdown process at high temperatures**

289 With increasing temperature, most Raman modes shift toward lower
290 wavenumbers (Figs. 3d-h), indicating a temperature-induced softening of the
291 phlogopite lattice for the two samples. So far, no one has reported the effect of F on the
292 lattice vibrations of phlogopite at high temperatures. To investigate this, Figure 6a
293 compares the temperature-induced shifts of the main Raman modes of the F-poor
294 phlogopite and the F-rich one. It is seen that F has an impact on the
295 temperature-induced shifts of the modes around 320 and 682 cm^{-1} . The shifted
296 amplitudes of these modes are drastically reduced for the F-rich sample. This suggests
297 that incorporation of F retards the lattice softening at high temperatures.

298 Fluorine appears to have an impact on the thermal behavior of OH in phlogopite. It
299 can be observed that wavenumbers of the three groups OH bands in the F-rich

300 phlogopite are generally lower than those in the F-poor phlogopite (Fig. 4d).
301 Relationship between the OH wavenumbers for the three groups of OH bands and Al
302 content has been reported by Robert and Kodama (1988). According to them, micas
303 with higher Al contents show lower OH wavenumbers for the type N and I OH bands,
304 which is in contrast with our study that the F-poor phlogopite with a higher Al content
305 has a higher OH wavenumber for all the types of OH bands. Therefore, the Al content
306 alone cannot explain the differences in OH wavenumbers between the two samples.
307 Another factor may be the higher Fe content in the F-rich phlogopite which will
308 decrease the OH wavenumber (Vedder 1964). In addition, the impact of F on the OH
309 wavenumber has been reported for mica, amphibole, and talc (Robert et al. 1993, 1999;
310 Rywak and Burlitch 1996), which is not difficult to understand considering the stronger
311 electronegativity of F than oxygen (Pauling 1932). Thus, it may be expected that F
312 substitution for OH in phlogopite will strengthen the hydrogen bonding (O-H \cdots F),
313 thereby reducing the OH wavenumber (Libowitzky 1999). This has been seen in the
314 F-induced OH band shift to lower wavenumbers in apatite (e.g., Tacker 2004). To
315 explore the correlation between F incorporation and hydrogen bonding strength, we
316 plot wavenumbers of the type N OH against the F content represented by their X_F
317 values in Figure 6b, based on the data from this study and a compilation of data from
318 literature. Although the data are somewhat scattered, mainly due to different chemical
319 compositions besides F, the negative relationship is obvious. Consequently, with
320 increasing F substitution, the strength of hydrogen bonding in phlogopite will be
321 enhanced. This strengthened hydrogen bonding may influence the thermal stability of
322 OH, which is seen in the different thermal behavior of OH for the F-poor phlogopite vs.
323 F-rich phlogopite. With increasing temperature, the OH bands all become broader, and
324 shift towards lower wavenumbers, suggesting temperature-induced OH softening (Figs.
325 4a-b). There also exist some differences in the amplitudes of the temperature-induced
326 wavenumber shifts of the OH bands between the two samples. The shift amplitudes are
327 generally larger for the F-poor phlogopite than for the F-rich one (Fig. 6c). In other

328 words, despite the fact that a higher temperature induces OH softening for all the
329 phlogopite samples, phlogopite will experience less softening with addition of F. Thus,
330 dehydrogenation may be more difficult for the F-rich phlogopite, and this needs to be
331 verified by dehydrogenation kinetics studies in future.

332 From the vibrational spectra, the isobaric Grüneisen parameters of the lattice and
333 OH modes in the two samples are calculated via the equation $\gamma_{iP} = -\frac{1}{\alpha v_i} \left(\frac{\partial v_i}{\partial T} \right)_P$, where
334 γ_{iP} is isobaric Grüneisen parameter of a mode (v_i), and α is the thermal expansion
335 coefficient. Since there has been no report of the effects of F on the thermal expansion
336 coefficient of phlogopite, we here refer to the thermal expansion coefficient of $6.26 \times$
337 10^{-5} K^{-1} from Tutti et al. (2000). The calculated isobaric Grüneisen parameters of the
338 two samples are listed in Tables 2 and 3. The parameters are generally larger for the
339 F-poor phlogopite than the F-rich phlogopite, indicating that the OH and lattice of the
340 F-poor phlogopite are more sensitive to temperature. Therefore, our study reveals that
341 both OH and the lattice of F-rich phlogopite have a higher thermal stability, compared
342 with those of the F-poor phlogopite. Hydrogen diffusion from the lattice has been
343 suggested to control the stability of muscovite (Gaines and Vedder 1964). The impact
344 of F on the lattice and the OH are thus probably responsible for the different breakdown
345 processes of the F-rich and F-poor phlogopites at high temperatures. To date, one study
346 has reported on the high-pressure Raman spectra of phlogopite (Williams et al. 2012).
347 They reported that both OH and lattice stiffened with increasing pressure, whereas the
348 impact of F is unknown. Considering that pressure is another important parameter,
349 future experimental work should focus on high pressures to explore the effects of F on
350 the behavior of OH and the lattice of the phlogopite.

351 **IMPLICATIONS**

352 Because of the size similarity between F and OH, it is generally supposed that
353 the incorporation of F into both nominally hydrous and anhydrous silicate minerals is
354 mainly through the substitution for OH (e.g., Robert et al. 1999; Crépeisson et al. 2014;

355 Roberge et al. 2015; Beyer et al. 2016; Grützner et al. 2017b; Yoshino and Jaseem
356 2018; Hughes and Pawley 2019). This substitution inevitably causes variations in the
357 structure and property of the phlogopite. The data from this study demonstrates that
358 increasing the F substitution in phlogopite can prevent O-H bonding weakening and
359 lattice softening at high temperatures, and consequently stabilize the phlogopite,
360 which has an impact on the stability of phlogopite at high temperatures.

361 It is well known that the mantle is a major water reservoir in the Earth because of
362 the storage of hydrogen defects in nominally anhydrous mantle minerals such as
363 olivine (e.g., Bell and Rossman 1992; Pearson et al. 2014). Recent studies have also
364 indicated that the mantle has acted as a reservoir for F (e.g., Beyer et al. 2016;
365 Grützner et al. 2018; Yoshino and Jaseem 2018). A close association between F and
366 OH has been demonstrated via their coupled incorporation mechanism in olivine
367 (Crépeisson et al. 2014). To date, only one study has reported on the impact of F on
368 nominally anhydrous mantle minerals, mainly focusing on F storage and phase
369 stability (Grützner et al. 2018). Phlogopite is a F-rich and hydrous key mineral in the
370 upper mantle. Experimental studies have suggested that F concentrates preferentially
371 in phlogopite compared to other minerals (Edgar and Charbonneau 1991). The F
372 contents in mantle phlogopites are mostly from 0 to ~5 wt% and some are up to
373 ~8-9.5 wt% (Li et al. 2016). Figure 5 indicates that $X_F > 0.5$ can increase the
374 breakdown temperature of phlogopite to over 100 °C under ambient pressure.
375 Considering the high-pressure conditions in the upper mantle and the impact of other
376 elements, such as Ti, the specific breakdown temperature for phlogopite would be
377 affected. For example, a phlogopite with 8 wt% F ($X_F = 0.9$) is expected to be stable
378 up to 1350 °C under ambient pressure, while it is expected to be stable up to a
379 temperature lower than 1300°C at 7-15 kbar (Fig. 5). Indeed, high-Ti phlogopite
380 containing 8 wt% F has been found in an ultra-high-temperature (UHT) metamorphic
381 terrane (980-1120 °C, below 9 kbar) (Motoyoshi and Hensen 2001) suggesting that
382 F-rich samples can be stable at higher temperatures at ambient pressure than at a

383 higher pressure. Consequently, the increased stability and hydrogen bonding of
384 F-bearing phlogopite makes it an excellent candidate for the storage and transport of
385 both H₂O and the halogens in the upper mantle. In the lithospheric mantle, F can be
386 stored both in hydrous minerals and nominally anhydrous minerals. Although
387 amphibole, phlogopite, and apatite are accessory mantle phases, they are dominant F
388 carriers because F is likely a compatible element in these minerals (Klemme and
389 Stalder 2018). Furthermore, the transport of F in the mantle is dependent on the
390 thermal stabilities of F-bearing minerals. Previous studies showed that phlogopite can
391 be stable up to 1100 °C under 6 GPa, amphibole (K-richterite) can be stable up to
392 1300 °C, and apatite can be stable up to 1100 °C under 6 to 8 GPa (Konzett and
393 Ulmer 1999; Konzett et al. 2012). Since F plays an important role in stabilizing
394 phlogopite, amphibole, and apatite (e.g., Frost 2006), F-bearing phlogopite,
395 amphibole, and apatite act as important repositories of F in the mantle, dependent on
396 their distributions in some regional zones.

397 The effect of F on the hydrogen bonding may have some implications on the
398 dehydrogenation and electrical conductivity of phlogopite. Although it has been
399 reported that dehydrogenation will occur before phlogopite breaks down (e.g., Vedder
400 and Wilkins 1969; Tutti et al. 2000; Zhang et al. 2016), our study shows that the
401 addition of F may make dehydrogenation more difficult. Furthermore, Li et al. (2017)
402 has reported that increasing the F content can significantly enhance the electrical
403 conductivity of phlogopite. Our study shows that F incorporation can prevent O-H
404 bonds from weakening at high temperatures, which may impede hydrogen mobility.
405 Therefore, the contribution of hydrogen conductivity should decrease with increasing
406 F content in phlogopite.

407

408

ACKNOWLEDGMENTS

409 This work is supported by the National Natural Science Foundation of China

410 (41972038), the Zhejiang Province Natural Science Foundation of China
411 (LY18D020001), and the Fundamental Research Funds for the Central Universities
412 (K20210168). Can Rao and Suwen Qiu are thanked for the analysis of F using EPMA.

413 **REFERENCES**

414 Bebout, G.E. (1997) Nitrogen isotope tracers of high-temperature fluid-rock
415 interactions: Case study of the Catalina Schist, California. *Earth and Planetary*
416 *Science Letters*, 151(1-2), 77-90.

417 Bell, D.R., and Rossman, G.R. (1992) Water in Earth's Mantle: The Role of Nominally
418 Anhydrous Minerals. *Science*, 255, 1391-1397.

419 Beyer, C., Klemme, S., Grützner, T., Ireland, T.R., Magee, C.W., and Frost, D.J. (2016)
420 Fluorine partitioning between eclogitic garnet, clinopyroxene, and melt at upper
421 mantle conditions. *Chemical Geology*, 437, 88-97.

422 Chibisov, A.N. (2011) Effect of Fluorine Additions on the Stability of $Mg_3Si_4O_{10}(OH)_2$:
423 Computer Simulation. *Glass Physics and Chemistry*, 37(4), 441-444.

424 Chon, C.M., Lee, C.K., Song, Y., and Kim, S.A. (2006) Structural changes and
425 oxidation of ferroan phlogopite with increasing temperature: in situ neutron
426 powder diffraction and Fourier transform infrared spectroscopy. *Physics and*
427 *Chemistry of Minerals*, 33(5), 289-299.

428 Clemens, J.D., Circone, S., Navrotsky, A., McMillan, P.F., Smith, B.K., and Wall, V.J.
429 (1987) Phlogopite: High temperature solution calorimetry, thermodynamic
430 properties, Al-Si and stacking disorder, and phase equilibria. *Geochimica et*
431 *Cosmochimica Acta*, 51(9), 2569-2578.

432 Crépisson, C., Blanchard, M., Bureau, H., Sanloup, C., Withers, A.C., Khodja, H.,
433 Surblé, S., Raepsaet, C., Béneut, K., Leroy, C., Giura, P., and Balan, E. (2014)
434 Clumped fluoride-hydroxyl defects in forsterite: Implications for the upper-mantle.
435 *Earth and Planetary Science Letters*, 390, 287-295.

436 Dooley, D.F., and Patño Douce, A.E. (1996) Fluid-absent melting of F-rich phlogopite
437 + rutile + quartz. *American Mineralogist*, 81(1-2), 202-212.

438 Dymek, R.F. (1983) Titanium, aluminum and interlayer cation substitutions in biotite
439 from high-grade gneisses, West Greenland. *American Mineralogist*, 68(9-10),
440 880-899.

441 Edgar, A.D., and Charbonneau, H.E. (1991) Fluorine-bearing phases in lamproites.
442 *Mineralogy and Petrology*, 44(1-2), 125-149.

- 443 Erlank, A.J., Waters, F.G., Hawkesworth, C.J., Haggerty, S.E., Allsopp, H.L., Rickard,
444 R.S., and Menzies, M.A. (1987) Evidence for mantle metasomatism in peridotite
445 nodules from the Kimberley pipes, South Africa. In *Mantle Metasomatism*;
446 Menzies, M.A., Hawkesworth, C.J., Eds.; Academic Press: London, UK; pp. 221–
447 311.
- 448 Flemetakis, S., Klemme, S., Stracke, A., Genske, F., Berndt, J., and Rohrbach, A. (2021)
449 Constraining the presence of amphibole and mica in metasomatized mantle
450 sources through halogen partitioning experiments. *Lithos*, 380-381, 1-14.
- 451 Foley, S. (1991) High-pressure stability of the fluor- and hydroxy-endmembers of
452 pargasite and K-richterite. *Geochimica et Cosmochimica Acta*, 55(9), 2689-2694.
- 453 Frey, F.A., and Prinz, M. (1978) Ultramafic inclusions from San Carlos, Arizona:
454 Petrologic and geochemical data bearing on their petrogenesis. *Earth and*
455 *Planetary Science Letters*, 38(1), 129-176.
- 456 Fritschle, T., Prelević, D., Foley, S.F., and Jacob, D.E. (2013) Petrological
457 characterization of the mantle source of Mediterranean lamproites: Indications
458 from major and trace elements of phlogopite. *Chemical Geology*, 353, 267-279.
- 459 Frost, D.J. (2006) The Stability of Hydrous Mantle Phases. *Reviews in Mineralogy and*
460 *Geochemistry*, 62(1), 243-271.
- 461 Gaines, G.L., and Vedder, W. (1964) Dehydroxylation of muscovite. *Nature*, 201, 495.
- 462 Grützner, T., Klemme, S., Rohrbach, A., Gervasoni, F., and Berndt, J. (2017a) The role
463 of F-clinohumite in volatile recycling processes in subduction zones. *Geology*,
464 45(5), 443-446.
- 465 Grützner, T., Kohn, S.C., Bromiley, D.W., Rohrbach, A., Berndt, J., and Klemme, S.
466 (2017b) The storage capacity of fluorine in olivine and pyroxene under upper
467 mantle conditions. *Geochimica et Cosmochimica Acta*, 208, 160-170.
- 468 Grützner, T., Klemme, S., Rohrbach, A., Gervasoni, F., and Berndt, J. (2018) The effect
469 of fluorine on the stability of wadsleyite: implications for the nature and depths of
470 the transition zone in the Earth's mantle. *Earth and Planetary Science Letters*, 482,
471 236-244.
- 472 Halama, R., Bebout, G.E., John, T., and Scambelluri, M. (2014) Nitrogen recycling in
473 subducted mantle rocks and implications for the global nitrogen cycle.
474 *International Journal of Earth Sciences*, 103(7), 2081-2099.
- 475 Hall, A. (1999) Ammonium in granites and its petrogenetic significance. *Earth-Science*
476 *Reviews*, 45(3), 145-165.

- 477 Hammouda, T., Pichavant, M., Barbey, P., and Brearley, A. (1995) Synthesis of
478 fluorophlogopite single crystals. Applications to experimental studies. European
479 Journal of Mineralogy, 7(6), 1381-1387.
- 480 Hansen, E.C., and Harlov, D.E. (2007) Whole-rock, phosphate, and silicate
481 compositional trends across an amphibolite- to granulite-facies transition, Tamil
482 Nadu, India. Journal of Petrology, 48(9), 1641-1680.
- 483 Harlov, D.E., Johansson, L., Kerkhof, A.V.D., and Förster, H.-J. (2006) The role of
484 advective fluid flow and diffusion during localized, solid-state dehydration:
485 Söndrum Stenhuggeriet, Halmstad, SW Sweden. Journal of Petrology, 47(1),
486 3-33.
- 487 Henry, D.J., Guidotti, C.V., and Thomson, J.A. (2005) The Ti-saturation surface for
488 low-to-medium pressure metapelitic biotites: implications for geothermometry
489 and Ti-substitution mechanisms. American Mineralogist, 90, 316-328.
- 490 Hensen, B.J., and Osanai, Y. (1994) Experimental Study of Dehydration Melting of
491 F-Bearing Biotite in Model Pelitic Compositions. Mineralogical Magazine,
492 58A(1), 410-411.
- 493 Holloway, J.R., and Ford, C.E. (1975) Fluid-absent melting of the fluoro-hydroxy
494 amphibole pargasite to 35 Kbar. Earth and Planetary Science Letters, 25, 44-48.
- 495 Hughes, L., and Pawley, A. (2019) Fluorine partitioning between humite-group
496 minerals and aqueous fluids: implications for volatile storage in the upper mantle.
497 Contributions to Mineralogy and Petrology, 174, 78, 1-18.
- 498 John, T., Scambelluri, M., Frische, M., Barnes, J.D., and Bach, W. (2011) Dehydration
499 of subducting serpentinite: Implications for halogen mobility in subduction zones
500 and the deep halogen cycle. Earth and Planetary Science Letters, 308(1-2), 65-76.
- 501 Kendrick, M.A., Honda, M., Pettke, T., Scambelluri, M., Phillips, D., and Giuliani, A.
502 (2013) Subduction zone fluxes of halogens and noble gases in seafloor and forearc
503 serpentinites. Earth and Planetary Science Letters, 365, 86-96.
- 504 Klemme, S., and Stalder, R. (2018) Halogens in the Earth's Mantle: What We Know
505 and What We Don't. In: Harlov D., Aranovich L. (eds) The Role of Halogens in
506 Terrestrial and Extraterrestrial Geochemical Processes. Springer Geochemistry.
507 Springer, Cham.
- 508 Konzett, J., and Ulmer, P. (1999) The stability of hydrous potassic phases in lherzolitic
509 mantle-an experimental study to 9.5 GPa in simplified and natural bulk
510 compositions. Journal of Petrology, 40, 629-652.
- 511 Konzett, J., Rhede, D., and Frost, D.J. (2012) The high PT stability of apatite and Cl
512 partitioning between apatite and hydrous potassic phases in peridotite: an

- 513 experimental study to 19 GPa with implications for the transport of P, Cl and K in
514 the upper mantle. *Contributions to Mineralogy and Petrology*, 163, 277-296.
- 515 Lacalamita, M., Schingaro, E., Scordari, F., Ventruti, G., Fabbrizio, A., and Pedrazzi, G.
516 (2011) Substitution mechanisms and implications for the estimate of water
517 fugacity for Ti-rich phlogopite from Mt. Vulture, Potenza, Italy. *American*
518 *Mineralogist*, 96(8-9), 1381-1391.
- 519 Li, Y., Jiang, H., and Yang, X. (2017) Fluorine follows water: Effect on electrical
520 conductivity of silicate minerals by experimental constraints from phlogopite.
521 *Geochimica et Cosmochimica Acta*, 217, 16-27.
- 522 Li, Y., Yang, X., Yu, J.-H., and Cai, Y.-F. (2016) Unusually high electrical
523 conductivity of phlogopite: the possible role of fluorine and geophysical
524 implications. *Contributions to Mineralogy and Petrology*, 171(4), 37.
- 525 Libowitzky, E. (1999) Correlation of O-H stretching frequencies and O-H···O
526 hydrogen bond lengths in minerals. *Monatshefte für Chemie*, 130, 1047-1059.
- 527 Liu, W., Yang, Y., Busigny, V., and Xia, Q.-K. (2019) Intimate link between
528 ammonium loss of phengite and the deep Earth's water cycle. *Earth and Planetary*
529 *Science Letters*, 513, 95-102.
- 530 Loh, E. (1973) Optical vibrations in sheet silicates. *Journal of Physics C: Solid State*
531 *Physics*, 6(6), 1091-1104.
- 532 McKeown, D.A., Bell, M.I., and Etz, E.S. (1999) Raman spectra and vibrational
533 analysis of the trioctahedral mica phlogopite. *American Mineralogist*, 84(5-6),
534 970-976.
- 535 Motoyoshi, Y., and Hensen, B.J. (2001) F-rich phlogopite stability in
536 ultra-high-temperature metapelites from the Napier Complex, East Antarctica.
537 *American Mineralogist*, 86, 1404-1413.
- 538 Oba, T. (1990) Experimental study on the tremolite-pargasite join at variable
539 temperatures under 10 kbar. *Journal of Earth System Science*, 99(1), 81-90.
- 540 Ogorodova, L.P., Kiseleva, I.A., Mel'chakova, L.V., and Vladykin, N.V. (2009)
541 Thermodynamic properties of natural tetraferriphlogopite. *Geochemistry*
542 *International*, 47(11), 1137-1140.
- 543 Pagé, L., Hattori, K., de Hoog, J.C.M., and Okay, A.I. (2016) Halogen (F, Cl, Br, I)
544 behaviour in subducting slabs: A study of lawsonite blueschists in western Turkey.
545 *Earth and Planetary Science Letters*, 442, 133-142.
- 546 Palya, A.P. (2011) Storage and mobility of organic nitrogen and carbon in the
547 continental crust: Evidence from partially melted metasedimentary rocks, Mt.
548 Stafford, Australia. *Chemical Geology*, 281(3), 211-226.

- 549 Papin, A., Sergent, J., and Robert, J.L. (1997) Intersite OH-F distribution in an Al-rich
550 synthetic phlogopite. *European Journal of Mineralogy*, 9, 501-508.
- 551 Pauling, L. (1932) The nature of the chemical bond. IV. The energy of single bonds and
552 the relative electronegativity of atoms. *Journal of the American Chemical Society*,
553 54, 3570-3582.
- 554 Pearson, D.G., Brenker, F.E., Nestola, F., McNeill, J., Nasdala, L., Hutchison, M.T.,
555 Matveev, S., Mather, K., Silversmit, G., Schmitz, S., Vekemans, B., and Vincze, L.
556 (2014) Hydrous mantle transition zone indicated by ringwoodite included within
557 diamond. *Nature*, 507, 221-224.
- 558 Peterson, J.W., Chacko, T., and Kuehner, S.M. (1991) The effects of fluorine on the
559 vapor-absent melting of phlogopite + quartz; implications for deep-crustal
560 processes. *American Mineralogist*, 76, 470-476.
- 561 Piccinini, M., Cibin, G., Marcelli, A., Ventura, G.D., Bellatreccia, F., and Mottana, A.
562 (2006) Synchrotron radiation FT-IR micro-spectroscopy of fluorophlogopite in
563 the O-H stretching region. *Vibrational Spectroscopy*, 42(1), 59-62.
- 564 Prost, R., and Laperche, V. (1990) Far-infrared study of potassium in micas. *Clays and
565 Clay Minerals*, 38(4), 351-355.
- 566 Rieder, M., Cavazzini, G., D'yakonov, Y.S., Frank-Kamenetskii, V.A., Gottardi, G.,
567 Guggenheim, S., Koval', P.W., Moller, G., Neiva, A.M.R., Radoslovich, E.W.,
568 Robert, J.L., Sassi, F.P., Takeda, H., Weiss, Z., and Wones, D.R. (1998)
569 Nomenclature of the micas. *Clays and Clay Minerals*, 46, 586-595.
- 570 Rimsaite, J. (1970) Structural formulae of oxidized and hydroxyl-deficient micas and
571 decomposition of the hydroxyl group. *Contributions to Mineralogy and Petrology*,
572 25(3), 225-240.
- 573 Rimsaite, J. (1972) DTA, TG, IR and isotopic analyses and properties of phlogopite,
574 biotite muscovite and lepidolite in temperature range of metamorphic reactions. In:
575 H.G. Wiedemann (Eds.), *Thermal Analysis* (Vol. 3, pp. 683-695). Basel:
576 Birkhäuser.
- 577 Roberge, M., Bureau, H., Bolfan-Casanova, N., Frost, D.J., Raepsaet, C., Surble, S.,
578 Khodja, H., Auzende, A., and Fiquet, G. (2015) Is the transition zone a deep
579 reservoir for fluorine? *Earth and Planetary Science Letters*, 429, 25-32.
- 580 Robert, J.L., Beny, J.M., Ventura, G., and Hardy, M. (1993) Fluorine in micas:
581 crystal-chemical control of the OH-F distribution between trioctahedral and
582 dioctahedral sites. *European Journal of Mineralogy*, 5, 7-18.
- 583 Robert, J.L., and Kodama, H. (1988) Generalization of the correlation between
584 hydroxyl-stretching wavenumbers and composition of micas in the system

- 585 K₂O-MgO-Al₂O₃-SiO₂-H₂O: a single model for trioctahedral and dioctahedral
586 micas. American Journal of Science, 288(A), 196-212.
- 587 Robert, J.L., Ventura, G.D., and Hawthorne, F.C. (1999) Near-infrared study of
588 short-range disorder of OH and F in monoclinic amphiboles. American
589 Mineralogist, 84, 86-91.
- 590 Rywak, A., and Burlitch, J. (1996) The crystal chemistry and thermal stability of
591 sol-gel prepared fluoride-substituted talc. Physics and Chemistry of Minerals, 23,
592 418-431.
- 593 Sadofsky, S.J., and Bebout, G.E. (2000) Ammonium partitioning and nitrogen-isotope
594 fractionation among coexisting micas during high-temperature fluid-rock
595 interactions: examples from the New England Appalachians. Geochimica et
596 Cosmochimica Acta, 64(16), 2835-2849.
- 597 Scordari, F., Ventruti, G., Sabato, A., Bellatreccia, F., Ventura, G., and Pedrazzi, G.
598 (2006) Ti-rich phlogopite from Mt. Vulture (Potenza, Italy) investigated by a
599 multianalytical approach: substitutional mechanisms and orientation of the OH
600 dipoles. European Journal of Mineralogy, 18(3), 379-391.
- 601 Tacker, R.C. (2004) Hydroxyl ordering in igneous apatite. American Mineralogist, 89,
602 1411-1421.
- 603 Tlili, A., Smith, D.C., Beny, J.-M., and Boyer, H. (1989) A Raman Microprobe Study
604 of Natural Micas. Mineralogical Magazine, 53, 165-179.
- 605 Tutti, F., Dubrovinsky, L.S., and Nygren, M. (2000) High-temperature study and
606 thermal expansion of phlogopite. Physics and Chemistry of Minerals, 27,
607 599-603.
- 608 Tutti, F., and Lazor, P. (2008) Temperature-induced phase transition in phlogopite
609 revealed by Raman spectroscopy. Journal of Physics and Chemistry of Solids,
610 69(10), 2535-2539.
- 611 van Keken, P.E., Hacker, B.R., Syracuse, E.M., and Abers, G.A. (2011) Subduction
612 factory: 4. Depth-dependent flux of H₂O from subducting slabs worldwide.
613 Journal of Geophysical Research, 116(B1).
- 614 Vedder, W. (1964) Correlations between infrared spectrum and chemical composition
615 of mica. American Mineralogist, 49, 736-768.
- 616 Vedder, W., and Wilkins, R.W.T. (1969) Dehydroxylation and rehydroxylation,
617 oxidation and reduction of micas. American Mineralogist, 54, 482-509.
- 618 Ventruti, G., Levy, D., Pavese, A., Scordari, F., and Suard, E. (2009) High-temperature
619 treatment, hydrogen behaviour and cation partitioning of a Fe-Ti bearing volcanic

- 620 phlogopite by in situ neutron powder diffraction and FTIR spectroscopy.
621 European Journal of Mineralogy, 21(2), 385-396.
- 622 Weiss, M. (1997) Clinohumites, a field and experimental study. Ph.D. thesis. Zürich,
623 ETH Zürich, 1-168.
- 624 Williams, L., Wilcoxon, B.R., Ferrell, R., and Sassen, R. (1992) Diagenesis of
625 ammonium during hydrocarbon maturation and migration, Wilcox Group,
626 Louisiana, U.S.A. Applied Geochemistry, 7, 123-134.
- 627 Williams, Q., Knittle, E., Scott, H. P., and Liu, Z. (2012) The high-pressure behavior of
628 micas: vibrational spectra of muscovite, biotite, and phlogopite to 30 GPa.
629 American Mineralogist, 97(1), 241-252.
- 630 Xu, J.S., and Shen, G.F. (2005) Mineralogical study on fluorophlogopite from the Bayan
631 Obo ore deposit. Acta Mineralogica Sinica, 25(3), 213-216 (in Chinese).
- 632 Yang, Y., Busigny, V., Wang, Z.P., and Xia, Q.K. (2017) The fate of ammonium in
633 phengite at high temperature. American Mineralogist, 102(11), 2244-2253.
- 634 Yoshino, T., and Jaseem, V. (2018) Fluorine solubility in bridgmanite: A potential
635 fluorine reservoir in the Earth's mantle. Earth and Planetary Science Letters, 504,
636 106-114.
- 637 Yokochi, R., Marty, B., Chazot, G., and Burnard, P. (2009) Nitrogen in peridotite
638 xenoliths: Lithophile behavior and magmatic isotope fractionation. Geochimica et
639 Cosmochimica Acta, 73(16), 4843-4861.
- 640 Zema, M., Ventruti, G., Lacalamita, M., and Scordari, F. (2010) Kinetics of
641 Fe-oxidation/deprotonation process in Fe-rich phlogopite under isothermal
642 conditions. American Mineralogist, 95(10), 1458-1466.
- 643 Zhang, M., Tarantino, S.C., Su, W., Lou, X., Ren, X., Salje, E.K.H., Carpenter, M.A.,
644 and Redfern, S.A.T. (2016) Optical phonons, OH vibrations, and structural
645 modifications of phlogopite at high temperatures: An in-situ infrared
646 spectroscopic study. American Mineralogist, 101(8), 1873-1883.
- 647

648 **Table 1.** Chemical compositions of the two phlogopite samples (in wt%, dash = below 0.01 wt%).

	F-poor phlogopite (from Badakhshan)						F-rich phlogopite (from Franklin)								
	Phl1	Phl2	Phl3	Phl4	Average	SD	Phl5	Phl6	Phl7	Phl8	Phl9	phl10	phl11	Average	SD
SiO ₂	42.39	43.23	42.67	42.31	42.65	0.42	43.16	43.68	43.51	43.69	43.87	43.41	43.52	43.55	0.23
TiO ₂	1.15	1.13	1.24	1.20	1.18	0.05	0.33	0.42	0.18	0.25	0.24	0.23	0.21	0.27	0.08
Al ₂ O ₃	14.31	14.18	14.14	14.03	14.16	0.11	12.16	11.84	11.87	12.00	12.13	11.85	11.77	11.95	0.15
FeO	0.48	0.47	0.41	0.48	0.46	0.04	2.80	2.47	2.70	2.73	2.86	2.98	3.17	2.82	0.22
MnO	-	0.04	-	0.06	0.02	0.03	0.01	0.04	0.01	0.09	0.03	-	0.05	0.03	0.03
MgO	26.62	26.34	26.02	26.48	26.37	0.26	25.77	25.58	25.88	25.59	25.06	25.36	25.29	25.50	0.29
CaO	-	0.01	-	-	-	-	-	-	-	-	-	-	-	-	-
Na ₂ O	0.30	0.39	0.36	0.35	0.35	0.03	0.07	0.09	0.14	0.10	0.12	0.12	0.15	0.11	0.03
K ₂ O	10.36	10.27	10.22	10.41	10.31	0.09	10.76	10.77	10.70	10.63	10.71	10.69	10.76	10.72	0.05
F	0.23	0.26	0.24	0.24	0.24	0.01	3.98	4.10	4.10	4.01	3.97	4.03	4.24	4.06	0.09
Cl	-	0.03	0.04	0.02	0.02	0.02	0.12	0.12	0.13	0.14	0.16	0.17	0.14	0.14	0.02
Cr ₂ O ₃	0.03	-	0.08	0.04	0.04	0.03	0.20	0.14	0.13	0.15	0.16	0.14	0.16	0.15	0.02
NiO	-	-	-	0.05	0.01	0.03	0.07	-	0.05	0.04	0.05	0.08	-	0.04	0.03
H ₂ O*	4.22	4.23	4.19	4.19	4.21		2.34	2.28	2.28	2.33	2.34	2.29	2.20	2.30	
Subtotal	100.08	100.57	99.60	99.85	100.03		101.77	101.52	101.66	101.75	101.70	101.36	101.65	101.63	
O=F,Cl	0.10	0.12	0.11	0.11	0.11		1.70	1.75	1.76	1.72	1.71	1.74	1.81	1.74	
Total	99.99	100.45	99.49	99.74	99.92		100.06	99.77	99.90	100.03	99.99	99.62	99.84	99.89	

Note: *H₂O and OH estimated on the basis of 22 cation charges, assuming 10 oxygen and 2(OH+F+Cl) (Rieder et al. 1998).

649

650

651

652 **Table 2.** Evolutions of wavenumbers of the lattice modes (cm^{-1}) with temperature and the isobaric Grüneisen parameters. The data in italics are the
 653 uncertainties.

Temperature (°C)	F-poor phlogopite Lattice mode (cm^{-1})					F-rich phlogopite Lattice mode (cm^{-1})			
	20	192.8 <i>0.29</i>	281.3 <i>0.39</i>	326.9 <i>0.73</i>	357.7 <i>0.42</i>	681.6 <i>0.28</i>	192.6 <i>0.20</i>	276.6 <i>0.22</i>	320.4 <i>0.70</i>
100	191.7 <i>0.20</i>	280.8 <i>0.29</i>	325.3 <i>0.26</i>	356.4 <i>0.08</i>	680.6 <i>0.27</i>	191.3 <i>0.12</i>	275.2 <i>0.33</i>	320.1 <i>0.35</i>	681.2 <i>0.31</i>
200	190.5 <i>0.20</i>	279.9 <i>0.17</i>	323.3 <i>0.26</i>	354.4 <i>0.30</i>	679.7 <i>0.01</i>	190.0 <i>0.02</i>	274.5 <i>0.21</i>	319.1 <i>0.53</i>	680.3 <i>0.67</i>
300	189.0 <i>0.06</i>	278.5 <i>0.28</i>	321.2 <i>0.62</i>	352.7 <i>0.21</i>	678.2 <i>0.24</i>	188.9 <i>0.09</i>	273.3 <i>0.65</i>	318.3 <i>0.10</i>	679.3 <i>0.69</i>
400	188.2 <i>0.23</i>	277.7 <i>0.44</i>	319.6 <i>0.21</i>	349.9 <i>0.13</i>	676.9 <i>0.36</i>	187.3 <i>0.35</i>	272.0 <i>0.52</i>	317.0 <i>0.73</i>	678.7 <i>0.23</i>
500	186.5 <i>0.26</i>	276.3 <i>0.07</i>	316.8 <i>0.39</i>	347.9 <i>0.10</i>	674.3 <i>0.42</i>	185.5 <i>0.32</i>	270.8 <i>0.27</i>	315.8 <i>0.37</i>	676.9 <i>0.40</i>
600	185.4 <i>0.21</i>	275.5 <i>0.34</i>	314.4 <i>0.76</i>	346.0 <i>0.19</i>	673.3 <i>0.21</i>	184.7 <i>0.29</i>	270.0 <i>0.39</i>	314.9 <i>0.48</i>	675.5 <i>0.07</i>
700	184.5 <i>0.24</i>	274.1 <i>0.20</i>	311.9 <i>0.23</i>	344.0 <i>0.05</i>	671.8 <i>0.21</i>	183.9 <i>0.20</i>	269.0 <i>0.66</i>	314.2 <i>0.37</i>	674.5 <i>0.04</i>
800	183.2 <i>0.39</i>	273.5 <i>0.22</i>	308.6 <i>0.54</i>	342.3 <i>0.02</i>	670.1 <i>0.38</i>	183.1 <i>0.19</i>	267.7 <i>0.42</i>	312.2 <i>0.57</i>	672.8 <i>0.40</i>
900	181.8 <i>0.19</i>	272.4 <i>0.71</i>	306.0 <i>0.22</i>	340.6 <i>0.13</i>	668.6 <i>0.26</i>	182.0 <i>0.50</i>	266.4 <i>0.51</i>	310.8 <i>1.02</i>	671.2 <i>0.09</i>
1000	180.6 <i>0.20</i>	271.3 <i>0.91</i>	304.0 <i>0.60</i>	338.7 <i>0.07</i>	667.5 <i>0.22</i>	180.7 <i>0.09</i>	265.7 <i>0.03</i>	309.6 <i>0.48</i>	669.2 <i>1.25</i>
Wavenumber shift with temperature ($\text{cm}^{-1}/^\circ\text{C}$)	-0.0128 (20-400°C)	-0.0108 (20-1000°C)	-0.0242 (20-1000°C)	-0.0197 (20-1000°C)	-0.0123 (20-400°C)	-0.0125 (20-400°C)	-0.0109 (20-1000°C)	-0.0114 (20-1000°C)	-0.0097 (20-400°C)
Isobaric Grüneisen parameters	1.061	0.613	1.179	0.880	0.288	1.039	0.630	0.567	0.227

654

655 **Table 3.** Evolutions of wavenumbers of the three groups of OH bands (cm^{-1}) with temperature and the isobaric Grüneisen parameters of the OH modes. The
 656 data in italic are the uncertainties.

Temperature ($^{\circ}\text{C}$)	F-poor phlogopite OH wavenumber (cm^{-1})			F-rich phlogopite OH wavenumber (cm^{-1})		
	20	3714.2 <i>0.18</i>	3671.3 <i>0.25</i>	3621.7 <i>0.20</i>	3706.8 <i>0.22</i>	3665.5 <i>0.51</i>
100	3711.5 <i>0.26</i>	3670.6 <i>1.65</i>	3620.8 <i>0.11</i>	3704.4 <i>0.08</i>	3664.6 <i>0.87</i>	3598.0 <i>0.91</i>
200	3708.7 <i>0.14</i>	3670.0 <i>0.69</i>	3619.9 <i>0.08</i>	3701.7 <i>0.12</i>	3664.1 <i>0.75</i>	3597.5 <i>0.33</i>
300	3705.1 <i>0.18</i>	3668.0 <i>0.31</i>	3618.6 <i>0.16</i>	3698.8 <i>0.14</i>	3663.0 <i>0.46</i>	3596.0 <i>0.88</i>
400	3701.6 <i>0.12</i>	3666.4 <i>0.75</i>	3616.6 <i>0.68</i>	3695.2 <i>0.16</i>	3661.8 <i>1.52</i>	3596.0 <i>0.60</i>
500	3698.5 <i>0.04</i>	3662.7 <i>1.41</i>	3614.4 <i>1.79</i>	3691.6 <i>0.28</i>	3659.3 <i>1.92</i>	3594.8 <i>0.56</i>
600	3694.8 <i>0.17</i>	3658.9 <i>2.25</i>	3607.6 <i>4.06</i>	3687.9 <i>0.33</i>	3658.1 <i>1.64</i>	3583.9 <i>0.55</i>
700	3691.6 <i>0.23</i>	3653.2 <i>1.74</i>	3594.4 <i>2.91</i>	3685.6 <i>0.16</i>	3655.2 <i>2.37</i>	3580.4 <i>1.42</i>
800	3688.2 <i>0.34</i>	3650.7 <i>0.35</i>	3586.5 <i>1.53</i>	3682.8 <i>0.27</i>		3581.4 <i>2.19</i>
900	3684.6 <i>0.48</i>			3680.9 <i>0.16</i>		
1000	3681.0 <i>0.49</i>					
Wavenumber shift with temperature ($\text{cm}^{-1}/^{\circ}\text{C}$)	-0.0336 (20-1000 $^{\circ}\text{C}$)	-0.0119 (20-400 $^{\circ}\text{C}$)	-0.0111 (20-500 $^{\circ}\text{C}$)	-0.0304 (20-900 $^{\circ}\text{C}$)	-0.0089 (20-400 $^{\circ}\text{C}$)	-0.0089 (20-500 $^{\circ}\text{C}$)
Isobaric Grüneisen parameters	0.145	0.052	0.049	0.131	0.039	0.040

657

658 **Figure captions:**

659 Figure 1. (a-b) X-ray diffraction patterns of phlogopite powders at elevated
660 temperatures (“RT” represents room temperature). (c-d) Enlarged regions showing the
661 breakdown processes. (e) Comparison of phlogopite samples before and after the
662 heating and cooling runs.

663 Figure 2. Comparisons of Raman spectra of lattice vibrations between the initial and
664 quenched powder samples from 900, 1000, 1100, and 1200 °C, respectively. (a-d)
665 F-poor phlogopite; (e-h) F-rich phlogopite. The red arrows indicate the appearance of
666 the characteristic modes of forsterite at 825 and 857 cm⁻¹.

667 Figure 3. (a) Raman spectra of phlogopites at ambient conditions. (b) Raman spectra
668 at various temperatures from 20 to 1000 °C of the F-poor phlogopite. (c) Raman
669 spectra at various temperatures from 20 to 1000 °C of the F-rich phlogopite. (d-h)
670 Evolution of Raman shifts of lattice modes of the phlogopites (data are in Table 2).

671 Figure 4. (a) Unpolarized IR spectra of OH stretching vibrations from the F-poor
672 phlogopite at elevated temperatures and quenched from 1000 °C. (b) Unpolarized IR
673 spectra of OH stretching vibrations of F-rich phlogopite at various temperatures and
674 quenched from 1000 °C. The spectrum at 1000 °C of the F-rich phlogopite was
675 corrupt due to serious interference, is not presented here. (c) Unpolarized and
676 polarized IR spectra of OH stretching vibrations at ambient conditions, with the

677 polarizer rotated to 0° and 90° in (001) plane. (d) Wavenumbers of the three groups
678 from unpolarized OH bands with temperature (data from Table 3).

679 Figure 5. Breakdown temperature of phlogopite and other hydrous minerals (pargasite,
680 K-richterite, and clinohumite) as function of F content at different pressures. The blue
681 field represents the phlogopite stability field. Data are from this study and the
682 literature. Phlogopite at ambient pressure: Rimsaite (1970, 1972), Prost and Laperche
683 (1990), Hammouda et al. (1995), Ogorodova et al. (2009). Phlogopite at high
684 pressures: Hensen and Osanai (1994), Dooley and Patiño Douce (1996). Pargasite:
685 Holloway and Ford (1975), Oba (1990). K-richterite: Foley (1991). Clinohumite:
686 Weiss (1997). The open square represents the poorly constrained breakdown
687 temperature of fluorphlogopite in Hammouda et al. (1995). It is probably higher than
688 1403°C because it completely melts without breakdown at this temperature. The red
689 and blue filled squares represent F-poor and F-rich phlogopites in this study,
690 respectively.

691 Figure 6. (a) Temperature-induced wavenumber shifts of the lattice modes. (b)
692 Wavenumbers of the type N OH at ambient conditions versus X_F . Two data are from
693 this study, and the others are from previous studies (Vedder 1964; Prost and Laperche
694 1990; Xu and Shen 2005; Chon et al. 2006; Piccinini et al. 2006; Scordari et al. 2006;
695 Ventruti et al. 2009; Lacalamita et al. 2011; Zhang et al. 2016). (c)
696 Temperature-induced wavenumber shifts of the OH bands. Since there exist

- 697 discontinuities in the wavenumber evolutions of some lattice modes and OH bands, the
698 dotted squares indicate the temperature range of the linear fit.

Figure 1

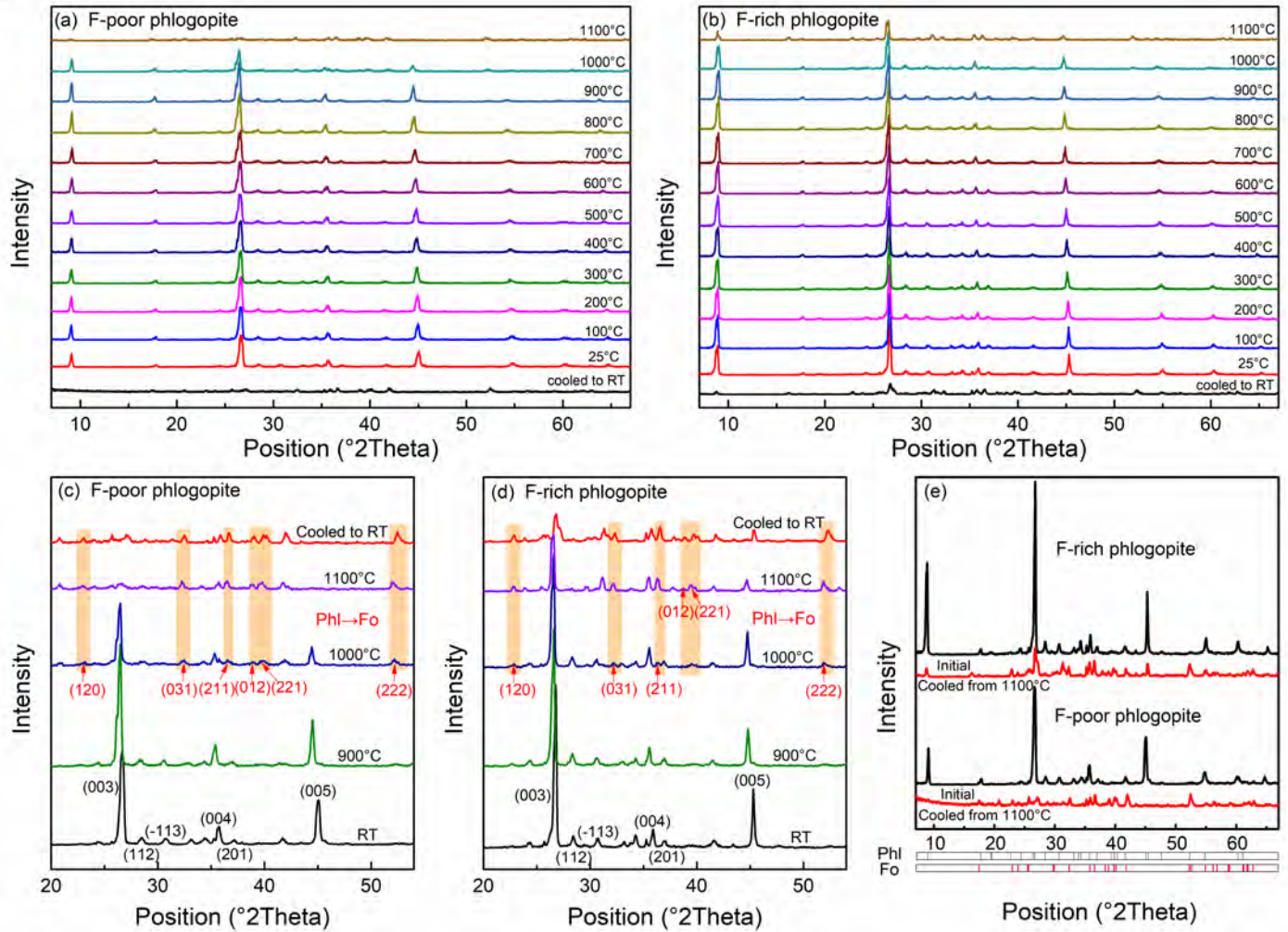


Figure 2

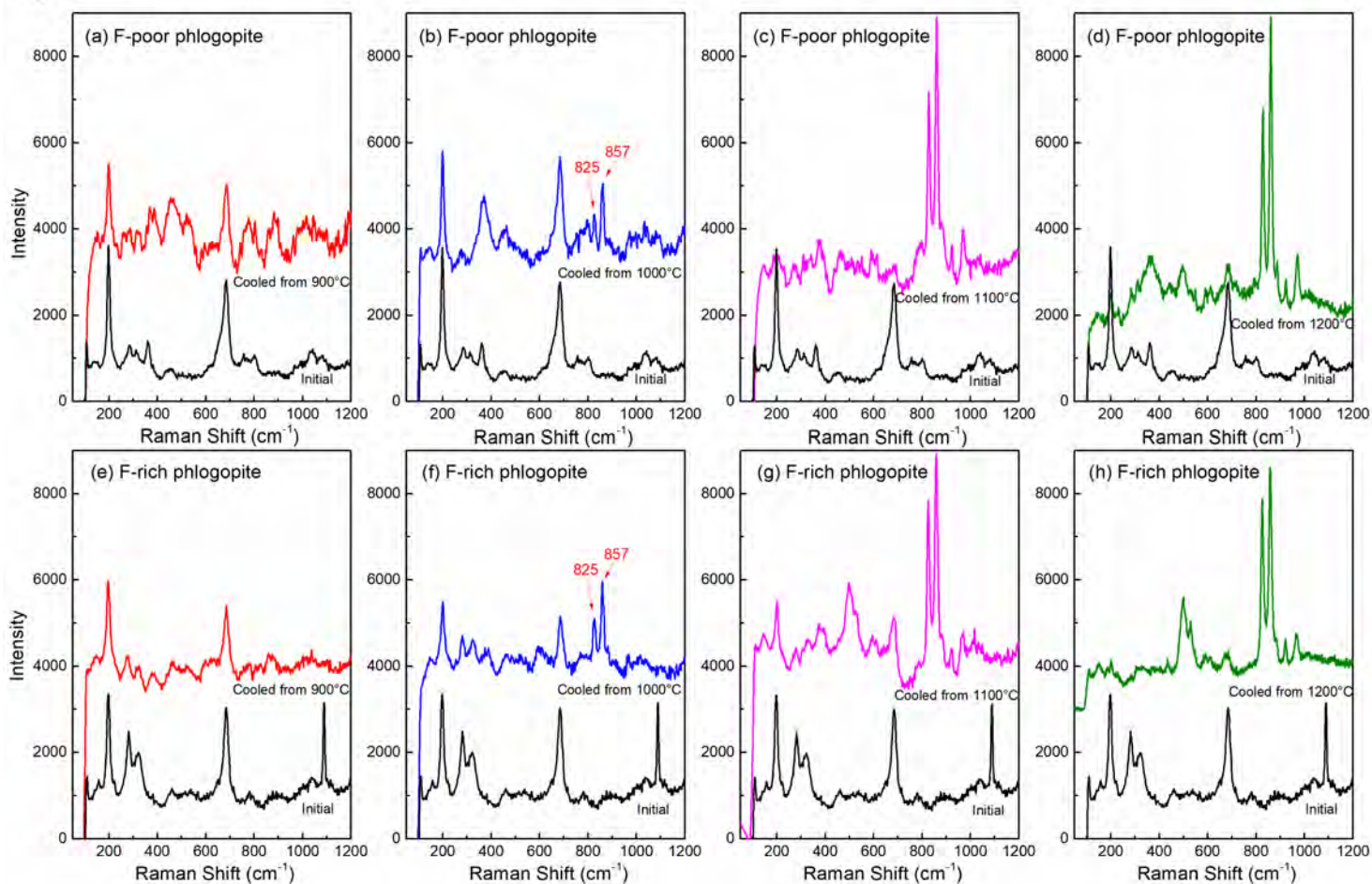


Figure 3

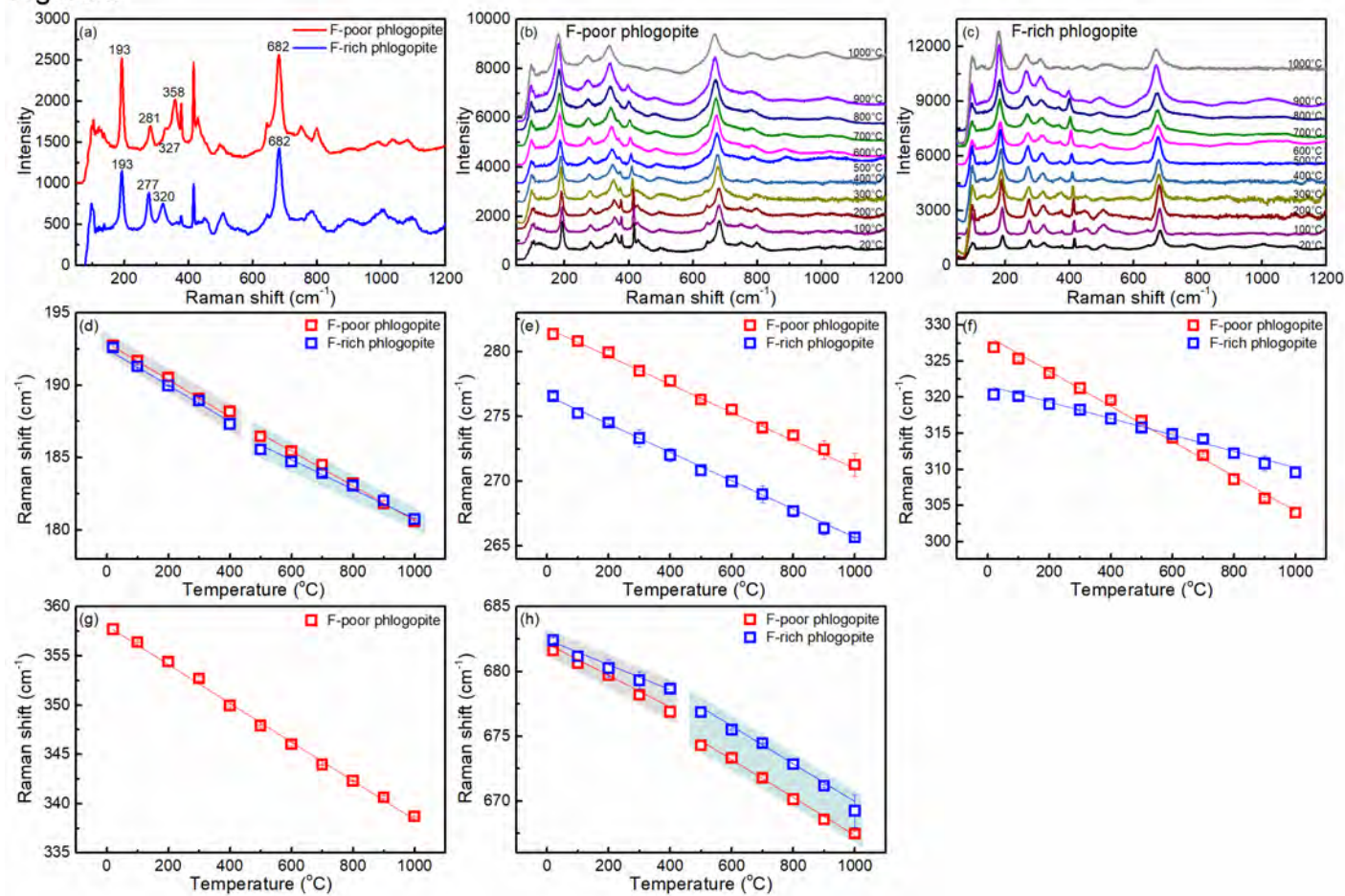


Figure 4

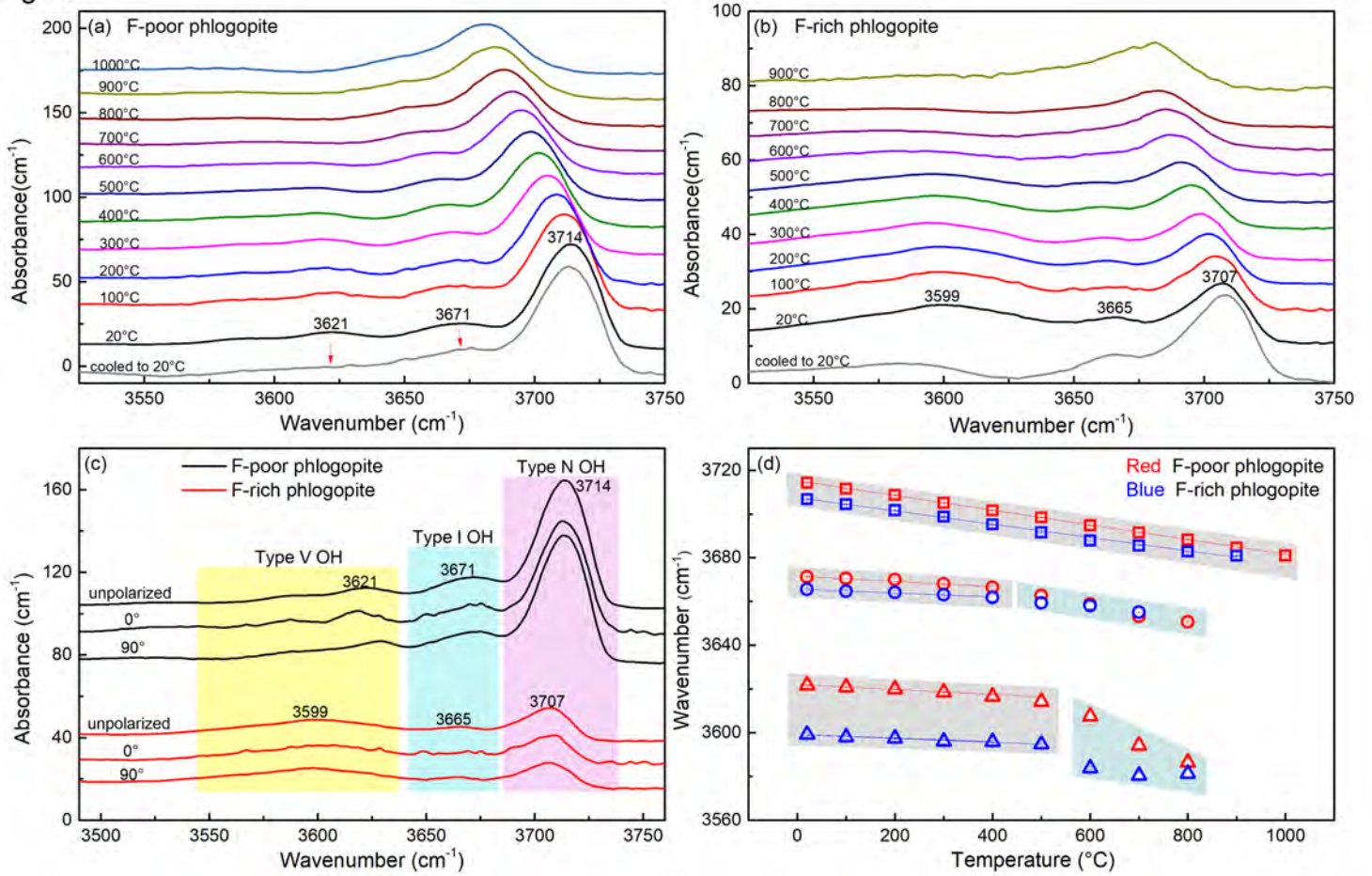


Figure 5

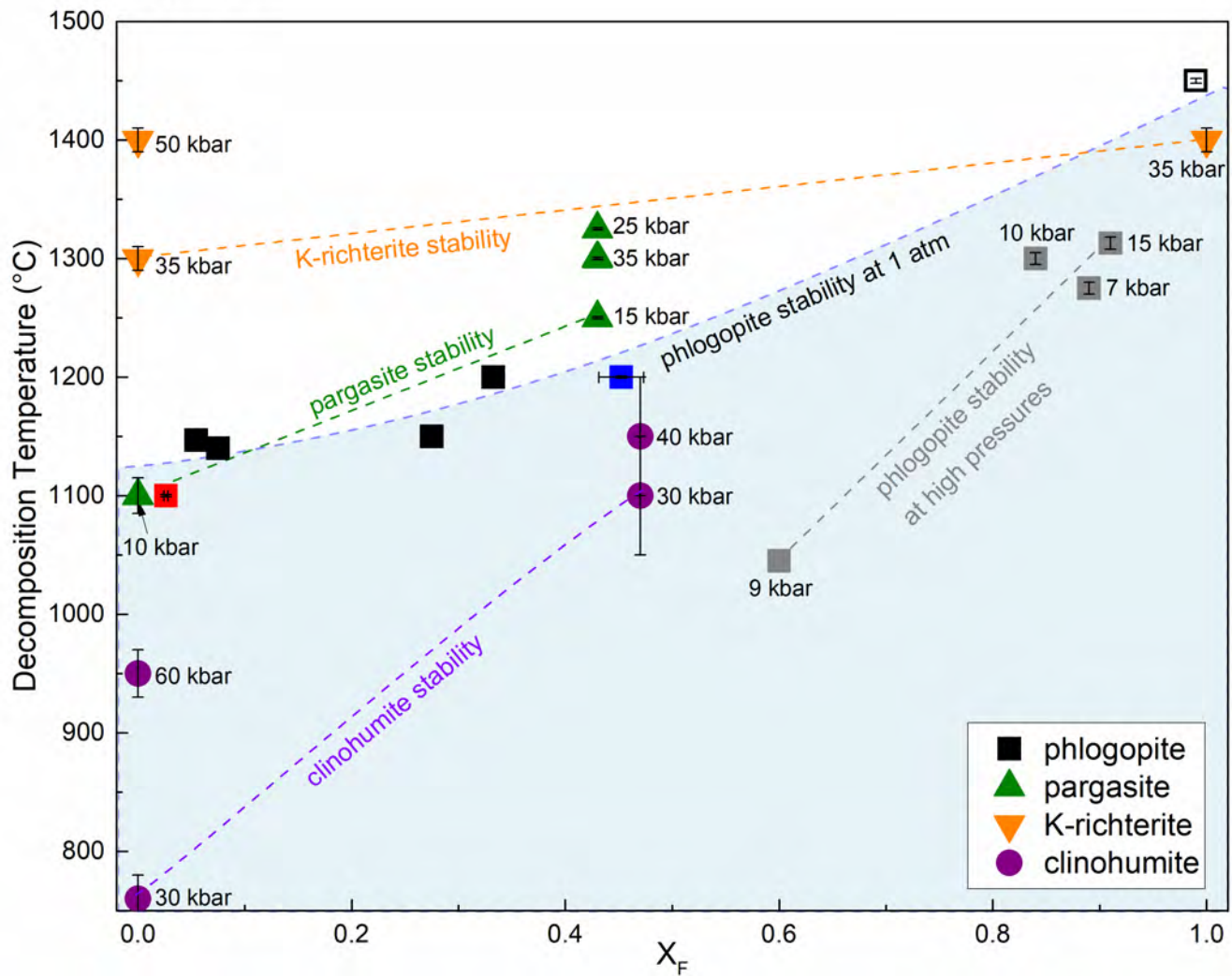


Figure 6

