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1	Magmatic reactivation of the Campi Flegrei volcanic										
2	system: insights from the Baia–Fondi di Baia eruption										
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26 Abstract

27 The Baia–Fondi di Baia was a multi-stage, small-scale eruption which occurred in the 28 western part of the Campi Flegrei caldera at 9525-9696 BP, marking the onset of Epoch 2 29 of post-Neapolitan Yellow Tuff volcanism. The eruption was characterized by a complex series of events related to two distinct eruptive episodes (Baia and Fondi di Baia) separated 30 31 by a short time interval, and each characterized by several eruptive phases. Mineralogical, 32 geochemical (major, and trace elements on whole-rocks, major and volatile elements on 33 matrix glasses and melt inclusions) and Sr-isotope characterization of the tephra material 34 sampled along the entire sequence was carried out in order to constrain magmatic evolution and dynamics of the feeding system. Three main compositional groups were 35 36 identified in matrix glasses and interpreted as representative of different magma bodies: i) 37 a trachyte (SiO₂: 60.3-64.7 wt.%), which is volumetrically predominant; ii) a tephriphonolite-latite (SiO₂: 55.1–57.9 wt.%); and iii) an intermediate magma group 38 39 between phonolite and trachyte compositions. This wide compositional heterogeneity contrasts with the narrow variability recognized in the bulk-rock compositions, which are 40 41 all trachytic. Mineral, melt inclusions and Sr-isotope data suggest that the trachytic magma 42 possibly derived from the Campanian Ignimbrite reservoir located at 6-9 km depth. 43 Volatile content in matrix glass indicates a storage depth of at least 6 km for the 44 tephriphonolite-latitic magma. The intermediate magma is interpreted as being derived 45 from a remelting and assimilation process of a partially crystallized trachytic body (crystal mush) by the hotter tephriphonolite-latitic magma. As the tephriphonolite-latite was 46 47 erupted together with the trachyte from the beginning of the eruption, we suggest that the 48 ascent of this magma played a fundamental role in triggering the eruption. Upwards through the tephra sequence, we observed a progressive increase of the tephriphonolite-49

latitic and intermediate phonolite-trachytic components. The presence of banded clasts
 characterized by different compositions is also indicative of syn-eruptive mingling during
 the final phases of the eruption.

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54 Introduction

Forecasting of future eruptive activity within calderas is a challenging goal in volcanology 55 56 and an important element in volcanic emergency plans (Newhall and Dzurisin 1988; 57 Sparks 2003; Gottsmann and Martì 2008). The study of caldera-forming eruptions has seen 58 tremendous progress in the last decade, in particular regarding related fields of pyroclastic 59 flow analysis (Gurioli et al. 2005; Esposti Ongaro et al. 2008; Sulpizio et al. 2014), caldera geology (Cole et al. 2005; Quick et al. 2009; De Natale et al. 2016) and caldera dynamics 60 61 (Roggensack 1996; Acocella 2007; Di Giuseppe et al. 2017). However, post-caldera 62 volcanism still remains under-investigated and is an inherently complex topic (Orsi et al. 63 2004; Bevilacqua et al. 2015, 2017; Neri et al. 2015). Most calderas have, in fact, been 64 associated with a range of activity styles from highly explosive to effusive with variable magmatic compositions and from widespread vents that are difficult to reconcile with any 65 regular spatio-temporal pattern (e.g., Bosworth et al. 2003; Acocella 2007; Cashman and 66 67 Giordano 2014).

The Campi Flegrei caldera (CFc) is an active volcanic system situated in the western part of the densely-populated area of the Bay of Naples (Southern Italy). Over the last few decades, the Campi Flegrei magmatic system has been showing signs of unrest (Barberi et al. 1984, 1989; D'Auria et al 2015), with episodes of ground uplift (Troise et al. 2007; Del Gaudio et al. 2010; Walter et al. 2014) and significant increase in hydrothermal degassing (Chiodini et al. 2015, 2016; Moretti et al. 2018) accompanied by minor, very shallow, seismicity (Orsi et al. 1999; Saccorotti et al. 2007; D'Auria et al. 2011). Since 2005, after a period of subsidence, a new minor inflation event has started (Orsi et al. 2004; De Natale et
al. 2006; Chiodini et al. 2016). The vicinity of the city of Naples (with ~1 million of
inhabitants in the metropolitan area alone) and a population of 1.5 million living within the
caldera, ranks the Campi Flegrei as one of the highest-risk volcanic areas in the world
(Alberico et al. 2002; Orsi et al. 2004; Selva et al. 2012; Bevilacqua et al. 2015; Neri et al.
2015).

In its past, the Campi Flegrei caldera has been characterized by various periods of volcanic activity. The two major events of the Campanian Ignimbrite (CI) and the Neapolitan Yellow Tuff (NYT) occurred at 39.8 ka BP (Giaccio et al. 2017) and 15 ka BP (Deino et al. 2004), respectively. These two events mostly shaped the present morphology of the partly submerged, nested caldera (Rosi and Sbrana 1987; Orsi et al. 2004).

The volcanic activity after the NYT eruption was characterized by a large number of 86 87 eruptive events of variable styles and intensities (e.g., Di Vito et al. 1999; Orsi et al. 2004; 88 Isaia et al. 2009). About 70 eruptions occurred during this period, mainly clustered during 89 three time intervals of activity (Epochs), separated by periods of quiescence of variable 90 duration, these being: Epoch 1 (15.0–10.6 ka BP), Epoch 2 (9.6–9.1 ka BP), and Epoch 3 91 (5.5–3.5 ka BP) (Di Vito et al. 1999; Isaia et al. 2009; Smith et al. 2011). This volcanic 92 activity was dominated by low- to medium-magnitude eruptions, never exceeding 0.1 km³ 93 of dense rock equivalent volume (Di Renzo et al. 2011), with vent positions which varied 94 during the three epochs of activity (Rosi and Sbrana 1987; Di Vito et al. 1999; Orsi et al. 95 2004; Bevilacqua et al. 2015).

Magmas feeding the post-NYT volcanic activity span a wide compositional range, varying from shoshonite to trachyte and phonolite, defining a slightly silica-undersaturated trend, with some enrichment in potassium (D'Antonio et al. 1999; Di Renzo et al. 2011). Most eruptions, which were located in the central-eastern sector of the caldera, show latitic to trachytic compositions (Smith et al. 2011). The less frequent eruptions of the western
sector, on the other hand, are generally characterized by more evolved compositions,
including trachyte and trachytic phonolite (Smith et al. 2011), and involve minor erupted
volumes (Orsi et al. 2004, 2009).

104 Over the past few years, tephro-stratigraphic, geochemical, geochronological, and isotopic 105 studies have helped in developing a model of the general structure and evolution of the 106 CFc magmatic system (Di Renzo et al. 2011; Mormone et al. 2011). Some studies have 107 been specifically dedicated to understanding the active magmatic system tapped during the 108 two largest eruptions (CI and NYT; Marianelli et al. 2006; Pabst et al. 2007) or to the 109 dynamics of smaller eruptions fed by magmas with different compositions (Cannatelli et al. 110 2007; Mangiacapra et al. 2008; Tonarini et al. 2009; Arienzo et al. 2009; Mormone et al. 111 2011; Di Vito et al. 2011; Fourmentraux et al. 2012; Pistolesi et al. 2016). However, due to 112 the complexity of the post-caldera eruptive activity, key eruptions which marked the 113 reactivation of the magmatic system after periods of prolonged quiescence still lack a 114 detailed investigation. Such studies may, however, provide fundamental insights into the 115 state of the magmatic system before its reactivation (such as magma residence times and 116 recharge events) and on the duration of expected unrest phases preceding an eruption (i.e., 117 timing of magma ascent).

The Baia–Fondi di Baia eruption (9525–9696 BP) was a small-scale, multi-stage event that marked the reactivation of the CFc volcanic system after a 1000 year-long quiescence and the onset of Epoch 2 (Di Vito et al. 1999; Smith et al. 2011). Here we present a new study on this key eruption which aims for a complete geochemical characterization of the eruptive sequences by integrating whole rock compositions, major element, volatile element (H₂O, Cl, F) and Sr-isotope data on matrix glasses and melt inclusions with mineral chemistry. Together, these data provide insights into the plumbing system of the eruption, and into the relationships between distinct magma batches at different depths feeding the eruption. The presented dataset and its interpretation complements the previous work recently published by Pistolesi et al. (2017) which reconstructed the tephra stratigraphy and the physical volcanology of the eruption. Starting from that reconstruction, and by sharing the same tephra sampling, this work extends the knowledge of the Baia– Fondi di Baia eruption to the magma dynamics and to the different melts which fed the explosive event.

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33 The Baia–Fondi di Baia eruption

134 The Baia–Fondi di Baia eruption was recently described in detail by Pistolesi et al. (2017), 135 who interpreted the pyroclastic sequence as the result of two distinct eruptive episodes separated by a short time interval, each in turn characterized by different eruptive phases. 136 137 The two vents of the eruption are located along a N-S lineament (Fig. 1) on the western 138 side of the Pozzuoli Bay; the largest crater (the present Baia harbour) is currently partly 139 invaded by the sea (Fig. 1). In previous studies, the eruptive centers were initially 140 interpreted to result from coeval eruptions of Baia and Fondi di Baia (Rosi and Sbrana, 141 1987). However, following the subdivision of the post-15 ka eruptions into three epochs 142 (Di Vito et al. 1999; Orsi et al. 2004; Isaia et al. 2009), they were ascribed to either a single 143 event (Fondi di Baia eruption, Di Vito et al. 1999; D'Antonio et al. 1999; Orsi et al. 2004; 144 Smith et al. 2011; Fedele et al. 2011) or distinguished as two events – the Baia and Fondi 145 di Baia eruptions (Di Renzo et al. 2011). The two tephra sequences of Baia and Fondi di 146 Baia (FdB) events overlie a paleosoil and are characterized by coarse-grained, fallout 147 pumice-bearing layers, embedded in fine-grained, cross-bedded pyroclastic density current 148 (PDC) deposits (Pistolesi et al. 2017).

Based on sedimentological features, Pistolesi et al. (2017) divide the Baia eruptive 149 succession into three main units (I to III), corresponding to different phases of the event 150 151 (Fig. 2). The Baia succession forming the proximal sequence consists of a tuff ring 152 structure with a basal vent-opening breccia (Unit I) which includes pumices, scoriae, 153 bombs and blocks of yellow tuff embedded in a yellowish muddy matrix and shows clear 154 evidence of hydrothermal alteration. The overlying Unit II forms the main part of the 155 sequence and consists of stacked PDC deposits with laterally-continuous coarse fallout 156 layers of light grey pumice clasts containing ballistic blocks and bombs. Yellowish-colored 157 material indicates that the lower part of this unit was influenced by the coda of the 158 preceding opening phase, which has been interpreted to have vented in a littoral environment (Pistolesi et al. 2017). During this phase, a 17 km-high column formed, 159 160 resulting in the emplacement of a pumice fallout deposit which covers an area >60 km² with a volume of 0.06±0.008 km³. In the upper part of Unit II, PDCs are predominant and 161 162 consist of cross-bedded flow deposits with dense ballistic bombs of up to several dm-size. 163 Unit III consists of fine-grained, lithic-rich, laminated ash beds with cross-bedding structures, which represent deposits from low-density PDCs (Fig. 2). 164

The proximal deposits of the Fondi di Baia eruptive succession outcrop along the rim of 165 166 the southern vent (Fig. 1b). According to Pistolesi et al. (2017), they are separated from the 167 fall deposits of Unit III of the Baia sequence by a mm-thick oxidation layer which marks a 168 short stasis. The lowermost deposit of Fondi di Baia consists of a coarse vent-opening 169 breccia (Unit IV) composed of blocks of tuff and altered rocks (Fig. 2). This is overlain by 170 the main part of the Fondi di Baia sequence (Unit V), which consists of stratified layers of 171 fine ash and pumice clasts with intercalated obsidian clasts and obsidian bombs. Overall, 172 Unit V deposits are related to the emplacement of PDCs (Fig. 2) which are characterized by lower energy. These deposits show generally higher lithic and obsidian contents thanthe Baia deposits (Pistolesi et al. 2017).

175 For this study, fourteen samples from the dataset of Pistolesi et al. (2017) were investigated 176 in further detail. Samples were collected at proximal key sections (for detailed location 177 refer to Table 1 from Pistolesi et al. 2017) and include bulk tephra, selected pumiceous 178 bombs and crystal-rich layers, representative of both PDC and fallout deposits (Fig. 2). The 179 selection was carried out based on representatives of the different stratigraphic units from 180 both the eruptive successions of Baia (Units II, III) and FdB (Unit V) and in order to 181 further investigate the large range of groundmass glass compositions already described by 182 Pistolesi et al. (2017). Figure 2 illustrates the stratigraphic relationships and positions of 183 the samples discussed in this study. Units I and IV (the vent-opening breccias of Baia and 184 Fondi di Baia, respectively) contain juvenile material, which is, however, too altered to be 185 analyzed. Respective pumice samples are completely transformed due to interaction with 186 seawater and hydrothermal alteration, and were thus not considered for chemical analyses.

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188 Methods

189 Bulk rock analyses were performed on unaltered pumices of both Baia (BFB10, BFB11, 190 BFB19) and FdB (BFB12, BFB36, BFB38) eruptive successions (Fig. 2). Glass and 191 mineral compositions were determined on juvenile clasts selected by hand-picking of 374 192 single clasts from the 1-2 mm and 1-0.5 mm fractions under a stereomicroscope. The 193 dataset is representative of both Baia (BFB5, BFB7, BFB8, BFB9, BFB10, BFB11, BFB13, 194 BFB17, BFB19) and FdB (BFB12, BFB36, BFB38). Crystals were extracted by gently 195 crushing pumice clasts or directly from bulk samples of deposits. Samples for melt 196 inclusion analyses consist in 42 hand-picked phenocrysts from the 1 mm (feldspar) and

0.250 mm (clinopyroxene) grain size fractions, taken from samples BFB5 (Unit II) and
BFB20 (Unit III).

199 Major and trace element analyses were performed on homogeneous vesicular pumice clasts 200 at Australian Laboratory Services (ALS) in Seville (Spain), using ICP-AES and ICP-MS, 201 respectively. Major elements analyses on matrix glasses were carried out using a Zeiss 202 EVO MA 10 scanning electron microscope (SEM) equipped with an Oxford Si(Li) energy-203 dispersive X-Ray detector (EDS) at the Istituto Nazionale di Geofisica e Vulcanologia 204 (INGV) in Pisa (Italy). Analytical conditions were a 15 kV acceleration voltage and a 205 probe current of ~300 pA. The measurements were carried out at a working distance of 8.5 206 mm with a 100 s live time in raster mode. Mineral phases were analyzed using the same 207 SEM-EDS equipment with identical analytical conditions. Zoning profiles on feldspar and 208 pyroxene crystals were calculated with various step sizes, generally $<20 \mu m$. Fifty-eight 209 analyses were also performed on melt inclusions (MIs) hosted in clinopyroxenes, and 59 210 on MIs hosted in feldspars.

Selected clasts and melt inclusions were subsequently analyzed using a CAMECA SX five electron microprobe at Camparis, Jussieu, France. Analytical conditions were a 6 nA probe current and a 15 µm spot size for major elements with counting times of 10 s for Na, Si and Mn, 15 s for Al, Ca, Ti, 20 s for Mg and Fe, and 30 s for K. We checked that the datasets obtained with SEM and EMPA on groundmass glasses were coherent with each other (Fig. S1), both being normalized on an anhydrous basis.

217 Minor elements were measured likewise with the CAMECA SX five electron microprobe 218 (Camparis Jussieu, France), analyzing the same set of groundmass glasses and melt 219 inclusions used for major elements. Analytical conditions consisted in a 30 nA beam 220 current, a 15 µm spot size and measuring times of 120 s for F, P, Cl and S, and 5 s for high sensitivity Mg. The detection limits for Cl, S, P, F and Mg were estimated to be 85, 60, 100,
500 and 320 ppm, respectively.

Analyses of ⁸⁷Sr/⁸⁶Sr were performed on a total number of 21 juvenile clasts from samples 223 224 BFB5 (Baia) and BFB12 (Fondi di Baia), from the 1-2 mm and 1-0.5 mm size fractions. 225 Glasses were selected using a binocular microscope to collect both vesicular and dense 226 clasts of different colour. To obtain exactly paired elemental and isotope data, each 227 selected clast was split into two portions: one for major element analyses and one for 228 ⁸⁷Sr/⁸⁶Sr measurement. Major element analyses were performed by EMPA at the Consiglio 229 Nazionale delle Ricerche-Istituto di Geoscienze e Georisorse (CNR-IGG) of Florence 230 (Italy) using a JEOL-JXA 8600 electron microprobe, with 15 kV accelerating potential and 231 a 10 nA beam current. Peak and background counting times of 40 s and 20 s, respectively, 232 were used. A defocused 5–10 mm beam and a lower counting time was used for Na (10 s) 233 to avoid alkali loss. Strontium isotope ratios were determined using a Thermal Ionisation 234 Mass Spectrometer (ThermoFinnigan Triton-Ti). The entire procedure was carried out at 235 the Radiogenic Isotope Laboratory of the Department of Earth Sciences (University of 236 Florence, Italy).

237 Water contents were obtained for 31 melt inclusions using a Micro-Raman Horiba Jobin 238 Yvon LABRAM HR800 with a green HeNe laser (wavelength 532 nm) equipped with an 239 OLYMPUS optical microscope system (Objectives $10\times$, $20\times$, $50\times$, $100\times$) at the University 240 of Roma Tre (Rome, Italy). Laser power for sample excitation was 60 mW with filter 241 reduction of 1/30; both the wide-aperture slit and opening of the confocal hole were 500 μ m. Samples were analyzed for 20 seconds with six repetitions. A spatial resolution of 1-2 242 243 μ um² was obtained using a confocal system with the 100× objective. All the methodologies 244 and the analytical conditions are further detailed in the Supplementary Material.

245

246 **Results**

247 Petrography and texture of Baia and Fondi di Baia juvenile clasts

248 Juvenile clasts from both Baia and FdB samples contain <5 vol.% of phenocrysts, which 249 consist of 250 µm- to 2 mm-sized feldspar and minor, ~250 µm-sized clinopyroxene, with 250 scarce biotite, apatite and spinel. Two main types of juvenile pyroclasts (vesicular and 251 dense clasts) can be distinguished throughout all grain size classes. Vesicular fragments 252 range from highly to moderately vesicular, following the classification of Houghton and Wilson (1989). Vesicles are generally of heterogeneous size (0.01-2.10 mm) and shapes, 253 254 and often coalescent. Dense clasts are grey to black and are characterized by smaller 255 vesicles (average diameter = 0.14 mm) and a glassy groundmass. Minor dense clasts 256 contain microlites (<100 µm), generally distributed randomly or grouped in clusters. The 257 FdB sequence also contains obsidian clasts, which are more abundant in the finer fraction 258 (<2 mm). Some clasts are characterized by colour banding, mostly corresponding to 259 variations in vesicularity and microlite content and size. Darker bands are characterized by 260 fewer vesicles and contain more microlites. In some cases, however, banding corresponds 261 to a chemical variability (as described below).

262

263 Whole rocks and mineral analysis

Bulk-rock compositions span a narrow range in the trachytic field of total alkali-silica (TAS) diagram (Le Maitre 1989), between 62.1 and 65 wt.% of SiO₂ and total alkali between 10.5 and 13.5 wt.%, where FdB samples have slightly higher alkali contents (Fig. 3, Table S1). The high concentrations in incompatible trace elements such as Nb (85-97 ppm) and Zr (605-787 ppm), and the very low Ba (0.7-138 ppm) and Sr (20-100 ppm) contents, place the pumice samples of Baia and FdB among the most evolved bulk rocks of

- Campi Flegrei (cf. Smith et al. 2011). The lowest values of incompatible elements, and thehighest Ba and Sr contents, are found in Fondi di Baia samples (Table S1).
- 272 Based on compositions, four different groups of feldspar phenocrysts (P_1 , P_2 , F_1 and F_2) can be defined (Fig. 4). Plagioclases P_1 are bytownites (An₆₂₋₈₃), whereas P_2 include 273 274 andesine and labradorite (An₃₀₋₅₅). Similarly, alkali feldspars indicate two distinct 275 populations. The first (F_1) is characterized by Or_{75-90} and An_3 , where Ba-rich (up to 2.4) 276 wt.%) rims and internal zoning with high Ba contents are common in F₁ sanidines. The 277 second population (F₂) shows lower Or contents (Or₃₅₋₇₄) and An₂₋₂₂ values (Fig. 4). An-278 rich plagioclase phenocrysts are extremely rare (group P₁; Fig. 4) and mainly form resorbed cores in the F₁ sanidine phenocrysts. They are also present as inclusions hosted in 279 280 sanidine. Conversely, most feldspars show normal compositional zoning with An₃₀₋₅₅ 281 and esine cores (group P_2 ; Fig. 4) and F_2 -type sanidine rims. Crystals with a sanidine core 282 and an andesine rim are less common. Microlite compositions match well those of 283 phenocrysts from group F_2 (Or₃₅₋₆₀, An₆₋₁₈), suggesting that this group represents the last-284 grown crystals.

285 Clinopyroxenes have diopsidic compositions (En₃₄₋₄₉, Fs₄₋₁₈, Wo₄₅₋₄₉). They are 286 characterized by normal zoning with Mg-rich cores and Fe-rich rims. They are also, 287 generally, finer-grained (0.5-0.25 mm) with respect to the feldspars.

288

289 Groundmass glasses and melt inclusions

290 *Major elements*

291 The compositions of Baia and FdB matrix glasses range from tephriphonolite-latite (SiO₂ \sim

292 55-58 wt.%) to phonolite and trachyte (SiO₂ ~ 60-65 wt.%). The glass data define three

293 main compositional groups (A, B and C; Fig. 5):

Group A plots in the tephriphonolite-latite domain of the TAS diagram, and is the least
evolved of all the Baia and FdB products. It displays a limited variability in alkali content,
with K₂O ranging from 7.4 to 8.8 wt.% (Fig. 5a) and Na₂O from 3.0 to 3.5 wt.% (Fig. 5b).
CaO and MgO concentrations are the highest of the whole glass dataset (>4.2 wt.% and
>1.5 wt.%, respectively; Fig. 5c, d), whereas SiO₂ contents are the lowest, ranging between
54.9 and 58 wt.%.

300 - Group B has a trachytic composition, with SiO₂ between 60.3 and 64.7 wt.% and the 301 lowest MgO, CaO and K₂O contents and highest Na₂O contents. This glass population 302 shares the same evolved composition as the whole rock analyses (cf. Figs. 3 and 5), in 303 agreement with the low crystal content of the group B pumices. It is worth noting that the 304 trachytic magma represents the main volume of the products erupted by the Baia and FdB 305 events.

Group C has an intermediate composition (phonolite at the boundary with trachyte field
of the TAS diagram) and all contents fall between those of Groups A and B (Fig. 5b-d),
except for K₂O (Fig. 5a). In particular, this group is well defined in the K₂O-SiO₂ diagram,
having silica contents between 58 and 63 wt.% and high K₂O contents, which
distinguished this group from the lower abundances of Group B. Interestingly,
compositionally banded clasts from Unit V of the FdB eruptive sequence display
compositions of the end-members A and C as well as hybrid A-C compositions.

Although no strong correlation results between matrix glass composition and juvenile clast texture (Figs. 3, 5), with vesicular and dense clasts being associated with the whole chemical variability, a large number of dense clasts are characterized by less evolved compositions. This results from an increase in the proportion of tephriphonolite-latite (Group A) and intermediate phonolite-trachytic (Group C) components between Units II and V, the latter being richer in dense and juvenile obsidian clasts than all other units. The 319 absence of an univocal correlation between texture and composition does not allow 320 quantification of the different magma volumes. However, based on the small size and 321 amount of dense clasts observed in the deposits, and on the results of whole-rock data 322 which only show the trachytic component, we infer that melt batch B represents at least 90 323 vol.% of the erupted magma.

324 Melt inclusions selected for analysis are enclosed in phenocrysts of Unit II (BFB5) and 325 Unit III (BFB20) of the Baia sequence (Table 2). Matrix glass enclosing the crystals has a 326 trachytic composition (i.e., it is associated with melt batch B). No suitable MI-hosting 327 crystals were found in samples of the FdB deposits. MIs are hosted both in F₁ (35 MIs) and F₂-type (5 MIs) sanidines, and in diopsidic clinopyroxenes (11 MIs). These are all \geq 40 µm 328 329 in size. In both sanidine and pyroxene crystals, MI compositions cover a restricted interval with an average SiO₂ content of 59.6±0.7 wt.% and a total alkali content of 12-13 wt.%. In 330 331 the TAS diagram they plot at the border between the phonolite and trachytic domains.

332

333 Sr-isotope dataStrontium isotope ratios are highly variable (0.70745-0.70778; Fig. 6), in 334 agreement with the high isotopic variability already shown for the last 10 ka of eruptive 335 activity at Campi Flegrei (Di Renzo et al. 2011). However, previous data on Baia-FdB 336 whole rocks shows a smaller variability (0.70753-0.70768; Civetta et al. 1991; Di Renzo et al. 2011). Ratios of ⁸⁷Sr/⁸⁶Sr are higher in melt batch B glasses (0.70759-0.70778), with 337 338 Baia samples having systematically higher values than FdB samples. Ratios of ⁸⁷Sr/⁸⁶Sr for 339 melt batch A glasses have the lowest values, ranging between 0.70745 and 0.70750; with 340 melt batch C clasts displaying slightly higher values (0.70747-0.70753). Most literature 341 values for whole rock isotopic ratios (as given by Civetta et al. 1991; D'Antonio et al. 342 1999: Di Renzo et al. 2011) are comparable to melt batch B, with some data having 343 intermediate values between Baia and FdB (Fig. 6).

344

345 *Cl and F*

346 Glass analyses show that each individual clast is compositionally homogenous with low standard deviation values (Table 1). The highest Cl and F concentrations of 0.82-0.91 wt.% 347 348 and 0.35-0.36 wt.% respectively, typify melt batch B clasts, without distinction between 349 Baia and FdB samples (Fig. 7a, b). In melt batch A clasts, on the other hand, Cl (0.49-0.52 350 wt.%) and F (0.16-0.18 wt.%) concentrations have a narrow range (80 analyzed spots on nine distinct clasts), and they average at 0.50±0.01 wt.% and 0.175±0.010 wt.%, 351 352 respectively (Fig. 7a, b). 353 Melt inclusions display a wide range of halogen concentrations (up to 0.70 wt.% Cl and 354 0.23 wt.% F), with the highest contents found in MIs of F₂ feldspars (Fig. 8). They also 355 registered the lowest Cl (0.36-0.42 wt.%) and F (0.09-0.14 wt.%) when compared to the

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356

358 Water

matrix glass data (Table 2).

Water content measured in 31 MIs hosted in both feldspar and clinopyroxene show highly variable values. They range from 1.1 to 4.2 wt.%, but most are between 1.0 and 2.2 wt.% and do not show clear correlations with other elements (Fig. 8). They are typical of the water contents of CFc magmas (MIs) which cover a large interval from 1 to 6.9 wt.% (Marianelli et al. 2006; Fourmentraux et al. 2012). The highest water content for a CFc magma was found in clinopyroxene-hosted MIs of the less evolved magmas of Fondo Riccio (Cannatelli et al. 2007).

366 Water versus F and Cl plots also show two sub-parallel trends. MIs of F_2 feldspars, having 367 the highest Cl (>0.5 wt.%) and F (>0.18 wt.%) contents, have a slightly positive linear 368 trend with water (Fig. 8b, c). Instead, MIs of F_1 feldspars have constant Cl and F contents 369 with increasing H₂O.

370

371 **Discussion**

372 Magmatic end-members and their relationships

373 The Baia-Fondi di Baia eruption offers insights into the processes characterizing small, 374 post-caldera eruptions occurring at calderic systems. This explosive event, in particular, 375 appear to be fed by multiple magma batches located at different depths, and including 376 interaction prior and during the eruption. To aid in our discussion of the possible 377 relationships among the described different magma batches, all data considered here are 378 summarized in Table 3. In summary, whole-rock data have the composition of batch B, 379 whereas glass from single clasts displays a much wider compositional spectrum (Figs. 3 380 and 5). Groundmass glasses thus provide a more complete overview of the compositional 381 heterogeneities of the magmatic system. Glass compositions recognized in the Baia-FdB 382 eruption range from tephriphonolite-latite to phonolite and trachyte compositions that are 383 comparable with the whole compositional spectrum of the Neapolitan Yellow Tuff and 384 post-NYT magmas (Orsi et al. 1995; D'Antonio et al. 1999; De Vita et al. 1999; Tonarini 385 et al. 2009; Di Renzo et al. 2011).

Volumetrically, the magma erupted during the Baia and FdB eruptive episodes had a dominantly trachytic composition representative of batch B compositions (Table 3). This magma was more Sr-radiogenic than melt batches A and C, and Sr isotopes are higher in Baia than in FdB glasses. The F₂-type feldspar phenocrysts and micro-phenocrysts of melt batch B magma cover the same compositional range of microlites, thus the F_2 - P_2 feldspars represent pre-eruption mineral phases in equilibrium with melt batch B. Most melt inclusions are hosted in F₁-type sanidine, which are not in equilibrium with melt batch B. These crystals may have been mechanically entrained during magma ascent so that their melt inclusions record a previous phase of the magmatic history. The repeated zoning with Ba-rich bands suggests that F_1 crystals are related to a complex history of refilling from less evolved melts and crystallization in a deeper reservoir (following, for example, Slaby et al. 2008; Hildreth and Wilson 2007; Ginibre et al. 2004; Streck 2008). They probably represent an earlier crystal mush. A few MIs, mainly hosted in F_2 crystals, have a more evolved composition close to that of melt batch B (Table 2).

The tephriphonolite-latitic magma with the lowest Sr-isotope ratios (batch A) is apparent only in the glass matrix of mostly aphyric clasts. Although present, this melt is rare in the Baia products, but is ubiquitous in those of the following FdB sequence. Hence, this magma composition was tapped mostly during the second part of the eruption (Phase V in Figure 1). Finally, phonolite-trachytic magma (melt batch C) was found only as matrix glass, mainly in the uppermost deposits of the FdB event, and thus at the end of the eruption.

407 Identifying and understanding the interactions among these distinct magmatic end-408 members are crucial in defining the magma dynamics associated with this eruption 409 sequence. The relationships between these end-member compositions are best defined and 410 followed using the SiO₂-K₂O diagram (Fig. 5a) in which the three groups (A, B and C) are 411 well separated. In this diagram, we plot changes in melt composition resulting from the 412 possible magmatic processes of fractional crystallization, melting and assimilation. We 413 simulate these processes using rhyolite-MELTS (Gualda et al. 2012) in the temperature 414 range 1100-900°C with fO₂=QFM and a pressure range of 200-50 MPa; using melt batches 415 A, B, C as starting compositions (Fig. 9). The rhyolite-MELTS software package 416 calculates equilibrium states of coexisting solids and liquids at given temperatures, 417 pressures, compositions, and oxygen fugacities from thermodynamic models of the

relevant phases (Gualda et al. 2012). The complete dataset produced by the rhyoliteMELTS simulations is reported in Tables S5-S8 of the Supplementary Material.

- 420
- 421 Crystal fractionation and crustal contamination

422 Starting from the least evolved melt (batch A) at $T=1060^{\circ}C$, simple crystal fractionation 423 produces a significant increase of K₂O with respect to silica in the melt. This is due to the 424 precipitation of solid phases as spinel, diopside and plagioclase (An₇₄₋₄₂) and minor biotite. 425 After crystallization of about 20% at T=980°C, the melt composition moves well away 426 from that of both the Group B and C compositions (Fig. 9, curve α). We stop the simulation at T=920°C at which point the residual liquid represents 71% of the mixture 427 428 (solid phases representing the remaining 29%) and is distant from the natural trend defined 429 by our data, thus being unrealistic. The increase in the values of Sr-isotope ratios between 430 batch A and batches B-C (Fig. 6), however, may suggest a more complex process of crystal 431 fractionation associated with limestone contamination. This process could be responsible 432 for the measured decrease in K₂O content between batches A and C (black arrow in Fig. 9). 433 However, a high assimilation degree is required if we are to match the K₂O content of melt 434 batch C (Fig. 9), resulting in an unlikely increase in CaO content (9.5 wt.%) (Table S9). 435 Fractional crystallization in the temperature range 990 – 900°C of batch B melt produces only evolved glasses of batch C, after precipitating variable quantities (19 wt.% at 436 437 T=930°C to 27 wt.% at T=900°C) of solid phases comprising plagioclase (An₃₅₋₂₁), spinel 438 and anorthose (Ab₄₄₋₄₃-Or₅₂₋₅₀) (Fig. 9, curve β). Furthermore, our measurements show that 439 melt batch B has a higher isotope ratios compared to batch C (Fig. 6), and this is not 440 compatible with possible assimilation due to high-Sr content crustal rocks.

441 Fractional crystallization starting from C compositions (Fig. 9, curve γ) produces 442 precipitation of plagioclase (An₃₅₋₂₇) and spinel at T=960°C, and anorthose (Ab₄₀₋₃₈-Or₅₅₋₅₈)

- at T=920°C. Resulting melts reproduce most of the variability observed within batch C.
 Thus, processes of phase segregation, even when possibly associated with crustal
 contamination, cannot justify the compositional differences between batches A, B and C.
- 446

447 Crystal mush rejuvenation

448 Forni et al. (2016) suggests that, in the case of the Campanian Ignimbrite, processes of 449 rejuvenation and mafic recharge in an existing magmatic reservoir led to the formation of a 450 high-crystallinity region, from which volatile-rich and crystal-poor melts were extracted. 451 To assess possible processes of rejuvenation of early-emplaced magmatic reservoirs, we simulated the remelting and assimilation of a highly-crystallized magma body using 452 453 rhyolite-MELTS. We assume that this process is heat-balanced (isenthalpic), and includes 454 assimilation and crystal fractionation (AFC) (De Paolo 1981). Isenthalpic AFC considers 455 the case where a change in enthalpy of the assimilating magma is counterbalanced by a 456 change in enthalpy of the assimilated crystal mush, which normally is at T \ll T_{liquidus}. As 457 stated by Reiners et al. (1995), assumptions of this model include: 1) no heat loss to the 458 portion of the magma body not assimilated, and 2) contamination is by assimilation of 459 melts of country rock. In our case, we expect times of reaction and mixing between 460 assimilating and assimilated melts faster than times of heat conduction away from the 461 interface between the melt and crystal mush. This implies that the first assumption is reasonable, though in more realistic view we should take into account influence of 462 463 conductive heat loss, diffusion, and reaction kinetics. Therefore, we consider that a melt with composition A at T=1100°C assimilated a crystal mush obtained by cooling of a melt 464 465 of B composition down to temperatures of 900°C, which corresponds to a temperature for 466 which the mineral assemblage is consistent with the observed paragenesis of magma B. This crystal mush is in equilibrium at T=900°C and includes 22% melt, 1% diopside, 3% 467

468 spinel, 55% anorthose (Or_{54} -An₄) and 15% plagioclase (An₂₃), as obtained by rhyolite-469 MELTS simulations. In this process, the assimilating melt mixes progressively with the 470 partial melt produced by the dissolution of the crystal mush.

471 In the K₂O-SiO₂ diagram (Fig. 9, curve $\alpha\beta$), the melt composition resulting from crystal 472 mush assimilation follows a trend that reproduces a large part of the variability observed 473 between the glass compositions of the Group A and C melts. This trend is controlled by the 474 mixing of the melt phases and by the crystallization of clinopyroxene, feldspar (An $_{61-29}$) 475 and spinel. At a modelled temperature of 964°C, when 60% of the mush has been 476 assimilated, anorthose starts to precipitate determining a K₂O depletion of the melt. This 477 trend never extends to B compositions, even if an unrealistic mass percentage of 478 assimilation is assumed (>86 wt.%).

479 The Sr isotope ratios of Group C, being intermediate between those of the A and B compositions, seem to confirm this process. In fact, quantitative modeling of ⁸⁷Sr/⁸⁶Sr (and 480 481 Sr content) variations indicates that the compositions of batch C can be obtained by a mixing process between a magma with the characteristics of batch A (⁸⁷Sr/⁸⁶Sr of 0.70745. 482 Sr 1000 ppm) and a component derived from partial melting of a crystal mush (⁸⁷Sr/⁸⁶Sr 483 484 0.70771 and Sr 300 ppm; Fig. 6). This is consistent with values of crystal mush assimilation of approximately 45% at a SiO₂ content of 60% (Fig. 9). On the other hand, 485 only 5% mixing of the evolved batch B (⁸⁷Sr/⁸⁶Sr: 0.70771; Sr: 25 ppm) with batch C 486 487 (⁸⁷Sr/⁸⁶Sr: 0.70750; Sr: 700 ppm) (black arrow in Fig. 6) explains the isotope variability 488 between Baia and FdB ratios within Group B compositions.

To perform these isotopic calculations, we assigned to batch A the lowest ⁸⁷Sr/⁸⁶Sr value of all the measurements for this group and the Sr content found in Campi Flegrei rocks having CaO similar to that of melt batch A (around 6 wt.%; Fig. S2). Furthermore, being the crystal mush derived from batch B crystallization, we assigned the average Sr-isotopic ratio of the batch B glasses from Baia (*c.m.* in Fig. 6) to its partial melts. This is based on
the assumption that radiogenic isotopes do not change between solid and liquid phases
during crystallization and melting processes. Finally, we use partition coefficients of ~2-3
(following Fedele et al. 2015) to calculate theoretical Sr contents in silicate liquids in
equilibrium with observed feldspar compositions. This yields a value of 300 ppm for the
Sr-content of partial melts.

499

500 *Mingling and mixing processes*

501 Compositionally banded clasts from Unit V of the FdB sequence indicate mingling. Clast 502 BFB36-d5 (Fig. 10b) highlights interaction between batch A and the silica-rich magmas of 503 Group C. Clast BFB36-d7 (Fig. 10c), on the other hand shows interaction within batch C, 504 involving different silica contents. Banded clasts thus indicate that compositionally 505 different magmas interacted syn-eruptively, suggesting that the timescale of this interaction 506 was too short to allow extensive chemical diffusion and hence magma mixing. They were 507 possibly located in isolated portions of this complex plumbing system, which resulted from 508 the combination of processes of re-melting, crystal mush assimilation and multiple inputs 509 of less-evolved magma.

510

In summary, the observed compositional variability can be explained by the combination of multiple processes. While crystal fractionation can be invoked for the internal variability within the three magma groups, the melt of composition B seems to have no parental relations with the other two magmas. The generation of end-member C can be related to a process of assimilation, where a B-related crystal mush is assimilated to various degrees in a high-temperature melt of composition A. Finally, mixing occurred when small amounts of newly-formed melt C started to mix with FdB residual liquids of batch B. 518

519 Volatiles and source depths of magma batches

Analysis of volatiles may place constraints on pressures of magma saturation at equilibrium (Anderson et al. 1989; Papale 2006), which, in combination with the study of melt inclusions can be used to estimate minimum depths and pressures of melt entrapment (Marianelli et al. 2006; Fourmentraux et al. 2012; Audétat and Lowenstern 2014).

524 Volatiles, especially chlorine, have been previously studied for the compositional range of 525 magmas erupted in the Campi Flegrei caldera. This includes analyses of the trachytes of 526 the Campanian Ignimbrite (Signorelli et al. 1999; Marianelli et al. 2006) and of the Averno 2 eruption (Fourmentraux et al. 2012), and the less evolved shoshonitic-latitic magmas of 527 528 the Agnano-Monte Spina (Arienzo et al. 2010), Nisida (Arienzo et al. 2016), Fondo Riccio 529 and Minopoli 1 and 2 (Cannatelli et al. 2007; Mangiacapra et al. 2008) and Solchiaro 530 eruptions (Mormone et al. 2016). A compilation of these literature data for Cl and F, where 531 F is considered as the most incompatible element, is reported in Figure 11. These data 532 highlight that our Cl concentrations (from 0.34 to 0.91 wt.%) cover the full range of Cl 533 concentrations found in the trachytic/phonolitic glasses of the Campi Flegrei products 534 younger than 15 ka.

535 Chlorine solubility strongly depends on melt composition, the initial Cl/H_2O ratio and the 536 sulphur content of the silicate melts, all of which determine the partitioning of Cl into a 537 hydrosaline high-Cl liquid (brine) and/or into a low-Cl, H₂O-rich vapour phase at pressures 538 ≤200 MPa (Webster 2004; Webster and Botcharnikov 2011). In turn, Cl saturation of H₂O-539 rich silicic melts leads to a flat chlorine distribution during isobaric crystallization and Cl 540 exsolution (Signorelli and Carroll 2000; 2002; Balcone-Boissard et al. 2016). Following 541 this line of reasoning, Balcone-Boissard et al. (2016) proposed the use of Cl contents in 542 matrix glasses of pumice clasts as a geobarometer for determining magma storage

pressures (and depths). This approach implies no Cl loss during eruption (due to very fast 543 544 decompression), low Cl diffusivity, and rapid cooling during emplacement, as is true for 545 the glassy matrices of pumice clasts produced during plinian eruptions (Balcone-Boissard 546 et al. 2016; Vidal et al. 2016). Applying these assumptions, we use this geobarometer to 547 determine the storage depths of magmas before the Baia-FdB eruptions. This application, 548 however, involves the existing dataset on chlorine solubility in hydrous trachytic and 549 phonolitic melts of the CFc which has been experimentally determined by Signorelli and 550 Carroll (2000; 2002).

551 Batch A has a low concentration of Cl and F (Fig. 11). It has to be noted that, during a process of crystal fractionation, Cl and F would behave as incompatible elements (thus 552 553 having a nearly constant Cl/F ratio), with apatite having a minor impact (Fourmentraux et 554 al. 2012). Data points that depart from this trend to form a plateau at fixed pressure (depth) 555 imply isobaric Cl exsolution. Following this line of reasoning, batch A – with an average 556 Cl concentration of 0.5 wt.% (Fig. 11a) – would have been saturated at a pressure of 160 557 MPa, equivalent to a depth of ~ 6 km (calculated with a density of 2700 kg m⁻³, Fig. 11). 558 The higher Cl and F concentrations of batch B instead suggest that it possibly equilibrated 559 at lower pressure. In fact, batch B data plot on the plateau that is delineated by the Cl 560 concentrations of the Averno 2 trachytes as defined by Fourmentraux et al. (2012; Fig. 561 11b). According to Signorelli and Carroll (2002), 0.9 wt.% of Cl dissolved in a trachytic 562 melt with 2.5 wt.% H₂O at 850 °C corresponds to a saturation pressure of ~25 MPa. This, if we use an average density of 2300 kg m⁻³ for the surrounding rocks (following 563 Montanaro et al. 2016), suggests a depth of ~1 km before eruption so this batch is 564 565 associated with the final stage of crystallization immediately before the eruption. The Cl 566 and F contents measured in C matrix glasses are similar to those of magma A, also 567 suggesting a comparable storage depth (Fig. 11).

568 Melt inclusions (MI) provide additional information (Fig. 11a, b); Most MIs (hosted in F₁-569 type crystals) have volatile concentrations that are variable in a narrow range, between 570 0.20 and 0.50 wt.% Cl, and between 0.05 and 0.14 wt.% F. Such halogen concentrations 571 are significantly lower than those recorded in the groundmass glass, and are very close to 572 those measured in MIs and matrix glasses of the Campanian Ignimbrite trachytes by 573 Signorelli et al. (1999) and Marianelli et al. (2006) (Fig. 11b), as well as to those measured 574 in younger magmas emplaced during the Fondo Riccio eruption (9.5 ka BP) in the CFc 575 north-western sector by Cannatelli et al. (2007). We thus suggest that F_1 crystals were 576 related to pre-existing, partially crystallized reservoir(s) and were subsequently entrained and recycled by batch B. The lower Sr isotope ratios of Baia whole-rock compared to those 577 578 obtained from the matrix glass analyses (Fig. 6) also suggest that F₁ crystals were possibly 579 characterized by low Sr isotopes. For comparison, CI melts and crystals are characterized by lower ⁸⁷Sr/⁸⁶Sr contents relative to Baia batch B compositions (Fig. S3 and Arienzo et al. 580 581 2009 for CI isotope values of crystals). Batch B values, in turn, are comparable to those of 582 Fondo Riccio magmas (0.70773-0.70779; Di Renzo et al. 2011). Consequently, we relate 583 the lower Sr isotope values of whole rock analyses to the presence of recycled F₁ crystals. 584 This leads us to prefer the CI reservoir as the source of the F₁ crystals. MIs hosted in F₂-585 type feldspars, on the other hand, show higher (up to 0.7 wt.%) Cl and (up to 0.23 wt.%) F 586 concentrations (Fig. 11a, b), in equilibrium with a pressure lower than that recorded by F_1 587 crystals. We propose that these crystals formed during ascent of batch B to a depth of 588 about 1 km, as indicated by Cl contents in B groundmass clasts (Fig. 11b).

In Campi Flegrei trachyandesitic-trachytic magmas water appears to be the dominant volatile species, as inferred from MIs which commonly display concentrations of 1.0-3.5 wt.% (e.g., Fourmentraux et al. 2012), but may have values in excess of >5 wt.% (Marianelli et al. 2006). Most H₂O contents of Baia–Fondi di Baia MIs, mainly measured 593 in F₁ crystals, vary between 1.1 and 2.2 wt.%, and do not exceed 4.2 wt.%, suggesting that 594 the trapped melt was undersaturated at the pressure of 160 MPa, as indicated by the Cl 595 plateau (where water solubility is about 6 wt.%; Di Matteo et al. 2004). Also, these values 596 are comparable or lower than the amounts of H₂O dissolved in the most primitive 597 tephritic/latitic melts at Campi Flegrei which span a range from 2.7 w% (Mormone et al. 598 2011) to 3.5-6.9 wt.% (Cannatelli et al. 2007). This suggests that the melt was in 599 equilibrium with, and/or buffered by, a CO₂-H₂O rich gas phase at the time of entrapment, 600 similar to the present situation which is characterized by a high CO_2 flux (Chiodini et al. 601 2016). Melt buffering by a CO₂-rich phase (and eventual re-equilibration of MIs) also 602 explains why H₂O is not correlated with the other elements.

603

604 Magma ascent dynamics

The Baia–Fondi di Baia eruption was fed by three distinct magma batches, which interacted prior to and during the event (Fig. 12a). Batch B represents the dominant mass erupted during the eruption, with batches A and C magmas being largely subordinate. Moreover, batch C was mainly erupted during the FdB episode.

609 During its ascent, batch B entrained F₁-type feldspars inherited from a crystal mush at the 610 level of the Campanian Ignimbrite and/or Fondo Riccio reservoirs (located at depths of 6-9 611 km following Marianelli et al. 2006; Cannatelli et al. 2007; Mangiacapra et al. 2008). This 612 implies that this trachytic magma rose from a depth range of 6-9 km, which coincides with 613 the partially melted zone identified as being around a depth of 7.5 km by seismic 614 tomography (Zollo et al. 2008). This is also in agreement with melt inclusions in apatite 615 which suggest the Campanian Ignimbrite as a possible contaminant for the evolved 616 trachytes of the Baia and FdB eruptions (Stock et al. 2018). Similarly evolved compositions are typical of the small, but rarer, eruptions that occurred in the western part 617

618 of the caldera, indicating a possible structural control during magma ascent (Di Vito et al. 619 2010; Smith et al. 2011; Fourmentraux at al. 2012). Our data show that during its rise, 620 batch B crystallized F_2 -type feldspars which accordingly recorded higher F and Cl contents 621 and led to saturation conditions in batch B. This was achieved at ~1 km, as shown by Cl 622 concentration in matrix glasses (Fig. 12b).

623 Batch A could have been originally stored at a depth similar to that of B. However, if Cl 624 and F concentrations in matrix glasses of batch A are related to final quenching upon 625 ascent, this storage depth can be considered as a minimum value. As batches A and B 626 coexist since the first stages of the eruption as pure compositional end-members, we suggest that the eruption trigger was the ascent of batch A, which entered the storage zone 627 628 of batch B. Here, A overpressurized B, but was also partially quenched by B. Batches A 629 and B were thus initially erupted together and, simultaneously, began to interact at depth 630 (Fig. 12b), where the time span ascent of A and eruption onset was long enough to allow 631 interaction between the A and B end-members. This interaction among melt batches A and 632 B can also explain the origin of C compositions as shown by modeled assimilation process (Fig. 9; $\alpha\beta$ trend). 633

A trachytic melt (B) thus fed both the Baia and FdB episodes, with increasing contents of A and C during the FdB phase (Fig. 12c). The occurrence of banded clasts, the presence of matrix glass of intermediate compositions, and Sr data suggest that, during the final FdB phase, A and C magmas interacted through mixing and mingling processes. As a result, B and C mixed, leading to isotopic variability within Group B during the final phases of the Baia–FdB eruptions.

640

641 **Conclusions**

642 Geochemical and mineralogical characterization of the Baia-Fondi di Baia eruption was carried out to constrain magmatic evolution and dynamics of the feeding system. Whereas 643 644 bulk-rock compositions span a narrow compositional range within the trachytic field, 645 matrix glasses reveal a wider heterogeneity. By combining the geochemical data from 646 matrix glasses, crystals and melt inclusions sampled through the eruptive sequence, we 647 found that three magma batches (A, B, and C) were involved in the eruption. A trachytic 648 magma (batch B) fed the two phases of Baia–Fondi di Baia eruption and originated within 649 the Campanian Ignimbrite reservoir at a depth of 6-9 km. We suggest that the ascent of a 650 less evolved, aphyric, tephriphonolite-latitic batch (A) overpressurized batch B triggering 651 the eruption. Most of the feldspars (and hosted melt inclusions), which were embedded 652 during the ascent of batch B, record a pre-existing, crystallized trachytic magma not 653 directly connected with the Baia-Fondi di Baia magmas, as already testified for other 654 eruptions at Campi Flegrei (see, for example, Forni et al. 2016). We thus suggest that a 655 remelting process triggered by batch A of a partially crystallized fraction of the B batch 656 was the most likely origin of melt batch C. Once formed, batch C also interacted with A to 657 form banded clasts and intermediate compositions.

658 The trachytic composition of batch B, as characterized by high silica contents coupled with 659 low potassium is rather unusual for Campi Flegrei compositions after the Campanian 660 Ignimbrite eruption (D'Antonio et al. 1999). In fact, a magma with similar composition 661 represents a large volume fraction of the Campanian Ignimbrite (Civetta et al. 1997; Forni 662 et al. 2016), and has also been described in the Averno 2 eruption (Di Vito et al. 2010; 663 Fourmentraux et al. 2012). The Averno 2 eruption involved comparably small erupted 664 volumes (0.07 km³; Di Vito et al. 2010) and was characterized by similarly evolved 665 compositions (Fourmentraux et al. 2012). Likewise, it is located in the northwestern part of the CFc (Di Vito et al. 2010; Fourmentraux et al. 2012). The sporadic occurrence of this 666

trachytic end-member in the post-Neapolitan Yellow Tuff activity can be related to a real 667 668 infrequence of this composition or to the geochemical dataset of Campi Flegrei magmas 669 which is largely focused on whole-rock data. In this latter case, single-clast analyses 670 completed here, coupled with volatile content analyses, proved capable of showing 671 chemical variability and activation of a compositionally-heterogeneous system. Based on 672 the proposed magma dynamics, we speculate that, after a period of rest, the ascent of a less 673 evolved magma can reactivate magma batches residing at the depth of the Campanian 674 Ignimbrite reservoir, which in turn can interact with a complex polybaric plumbing system. 675 This is particularly relevant for the western sector of the caldera, where eruptions are less 676 frequent and generally characterized by more evolved compositions, and where structural 677 features possibly play a major role. This geochemical study carried out with detailed 678 single-clast analyses and by means of variable analytical techniques shed light on the 679 complexity of the Campi Flegrei plumbing system. This, coupled with further 680 investigations (e.g. diffusion profiles in matrix glasses and crystals) could also help 681 estimating more precisely the timescales of its dynamics, providing an important tool for 682 volcanic emergency plans at Campi Flegrei and more generally at calderas.

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- 935 936
- 937
- 938939 FIGURE CAPTIONS
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Figure 1. a) Overview of the Campi Flegrei caldera with the main locations and eruption
vents cited in the text. b) Google Earth image showing the inferred vents for the Baia and
Fondi di Baia eruptive sequences (with dashed lines. Map data: SIO, NOAA, U.S. Navy,
NGA, GEBCO, TerraMetrics, ©2018 Google).

945

Figure 2. Stratigraphic sequence of the Baia–Fondi di Baia eruption. Stratigraphic heights
of the analyzed samples are also indicated. Modified after Pistolesi et al. (2017).

948

949 Figure 3. Total alkali vs. silica diagram (TAS) after Le Maitre et al. (1989) showing 950 matrix glass compositions for Baia (gold symbols) and Fondi di Baia (FdB, red symbols) 951 eruptive sequences. Whole rock compositions are also reported (light blue: Baia samples, 952 dark blue: FdB samples). Circles: dense clasts, diamonds: vesicular clasts. Compositions 953 are recalculated on anhydrous basis. TB: trachybasalt. Literature data for Neapolitan 954 Yellow Tuff (dark grevish field) and post-15 ka volcanic rocks (light grev field) from Orsi 955 et al. (1995), D'Antonio et al. (1999), De Vita et al. (1999), Tonarini et al. (2009) and Di 956 Renzo et al. (2011).

957

Figure 4. Ternary diagram of feldspars in various samples of the Baia (gold symbols) and the Fondi di Baia eruptive sequence (FdB, red symbols). Squares: phenocrysts; triangles: microlites. F_X and P_X mark groups of distinct feldspar compositions. Insets illustrate the different relationships among feldspars, microlites and matrix glass.

962

Figure 5. Variation diagrams of (a) K₂O vs SiO₂, (b) Na₂O vs MgO, (c) CaO vs Si₂O and
(d) CaO vs MgO for Baia and Fondi di Baia matrix glasses and whole rocks. The letters A,
B and C indicate three distinct compositional groups. Compositions are recalculated on an
anhydrous basis.

967

968 Figure 6. Variation diagram of Sr isotopes versus SiO₂ for matrix glasses of single, dense 969 or vesicular, clasts from Baia (BFB5) and Fondi di Baia (FdB, BFB12). Whole rock (WR) 970 and matrix glass data from literature are also reported for comparison (glass data from 971 literature are reported as lines because they lack of silica composition). SiO₂ is reported on 972 a water-free basis. The mixing line between a melt batch A (⁸⁷Sr/⁸⁶Sr: 0.70745; Sr: 1000 ppm) and a melt (⁸⁷Sr/⁸⁶Sr: 0.70771; Sr: 300 ppm) formed by remelting a crystal mush 973 974 derived from the crystallization of melt B magmas is also reported. Isotope ratio of the 975 crystal mush is calculated as the average value of melt B from Baia. The numbers along 976 the line indicate % of the crystal mush-derived melt involved in the process. We derived Sr 977 content based on partition coefficients (Ds/1) of Fedele et al. (2015) for less evolved 978 trachytic magmas, and assuming the presence of a small amount of Ca-plagioclase and 979 sanidine (with Sr > 1000 ppm) in the crystal mush ($D_{s/l}$ for feldspars ~2-3).

980

Figure 7. Variation diagrams of (a) Cl vs SiO₂ in groundmass glasses and (b) K₂O vs F in
glasses and melt inclusions (MI) from Baia (golden circles) and Fondi di Baia (FdB, red

983 circles) eruptive sequences. Literature data (open circles) for groundmasses are from Smith 984 et al. (2011). Melt inclusions in clinopyroxene (cpx; represented by gold squares) and 985 feldspar crystals (fsp; gold triangles for F_1 -feldspars and gold diamonds for F_2 -feldspars) 986 are also reported.

987

Figure 8. Variation diagrams showing (a) MgO vs. H_2O , (b) F vs. H_2O and (c) Cl vs. H_2O in melt inclusions hosted in F₁-feldspars (F₁-fsp; triangles), F₂-feldspars (F₂-fsp; diamonds) and clinopyroxenes (squares). Standard deviations (±1 σ) are indicated as error bars.

991

992 Figure 9. K₂O vs SiO₂ variation diagram with measured groundmass glasses (grey dots 993 and grey diamonds represent dense and vesicular matrix glass, respectively) and curves 994 showing different melt evolution paths obtained by the rhyolite-MELTS code (Gualda et al. 995 2012) for temperature range 1100-900°C and fO_2 =QFM. For simplicity, we report only 996 results obtained at P=100 MPa among the considered pressure range (50-200 MPa). Curves 997 α , β and γ represent melt evolution by crystal fractionation starting from less evolved 998 compositions of each group A, B and C, respectively. Numbers represent percent of melt. 999 Black arrow indicates K₂O decrease assuming 15% of associated limestone assimilation. 1000 Curve $\alpha\beta$ represents melt evolution obtained by isoenthalpic assimilation of a crystal mush 1001 with B composition at 900°C by melt with the composition of A. Numbers in brackets 1002 report the mass percentage of assimilation $[B/(A+B) \times 100]$.

1003

Figure 10. a) Variation diagram K₂O-SiO₂ showing the compositions of the glasses in the banded clasts BFB36 d5 (b) and BFB36 d7 (c). Matrix glass data for Baia (diamonds) and Fondi di Baia (FdB circles) are also reported in grey. Images of clasts are taken in backscattered electron mode. The grey scales are proportional to the mean atomic number of the analyzed area. Brighter and darker areas thus reflect less and more evolved glasscompositions, respectively.

1010

1011 Figure 11. Variation diagrams for a) Cl vs. MgO and b) Cl vs. F in glasses and melt 1012 inclusions from Baia and Fondi di Baia eruptive sequences. Symbols represent averaged 1013 compositions of individual clasts, with 1σ error. Feldspar-hosted melt inclusions are 1014 reported as gold triangles (F₁-type feldspar: F₁-fsp) and gold diamonds (F₂-type feldspar: 1015 F_2 -fsp) whereas clinopyroxene-hosted melt inclusions are represented by gold squares (MI 1016 in cpx). Literature data for melt inclusions are 1: Cannatelli et al. (2007) for Fondo Riccio; 1017 2: Marianelli et al. (2006) and Signorelli et al. (1999) for the Campanian Ignimbrite (CI); 1018 3: Fourmentraux et al. (2012) for Averno 2; 4: Arienzo et al. (2010) for Agnano Monte 1019 Spina (AMS). Data for juvenile glasses are reported for the CI and post-Neapolitan Yellow 1020 Tuff eruptions (NYT, 6: Smith et al. (2011)). Literature data for juvenile glass are reported 1021 for the CI (5: Signorelli et al. (1999)) and the post-NYT eruptions (6: Smith et al. (2011)). 1022 b) Cl vs. F for Baia dense matrix glasses (gold circles), and Fondi di Baia dense matrix 1023 glasses (red circles). Standard deviations (1σ are reported as error bars). The grey line 1024 illustrates melt evolution when driven by fractional crystallization (FC), assuming Cl and F 1025 as incompatible elements. The two horizontal red lines mark constant Cl compositions 1026 suggesting that the magma has reached saturation with the Cl-rich phase. Corresponding pressures are taken from experimentally determined Cl solubility studies on hydrous 1027 trachytes and phonolites (Signorelli and Carroll 2002) and are indicated next to the 1028 1029 respective saturation plateau.

1030

Figure 12. Sketch of the main pre-eruptive and syn-eruptive phases of the Baia–Fondi diBaia eruption. (a) Inferred locations of the inferred A and B batches. (b) Ascent of the

1033	trachytic (B) and tephriphonolite-latitic (A) batches from 6-9 km feeding the first part of
1034	the eruption, involving F_1 crystals from a mush and crystallization F_2 feldspars grown
1035	during ascent. Interaction between batches A and B to form melt C and initial withdrawal
1036	of the three compositions. (c) Vent shift, and onset of the FdB eruptive sequence. Ejection
1037	of batches A, B and C, including banded clasts with intermediate A-C compositions
1038	(shown as back-scattered image in the inset, where the banded clast compositions are only
1039	as example).

Table 1. Major element and volatile average compositions of matrix glasses. Compositions
1042 are given as averages with one standard deviation, as indicated beneath composition.
1043 MgO_{trace} was analyzed as a trace element, together with other trace elements (F, P₂O₅, Cl);
1044 n: number of points that were analyzed and averaged.

Table 2. Point analyses of melt inclusions hosted in feldspar and clinopyroxene crystals.1047Raw data are reported here, prior to normalization and averaging. MI refers to the analyzed1048melt inclusion, whereas MI_n refers to the corresponding analysis. The last row reports1049information on the mineral which enclosed the melt inclusion.

Table 3. Summary of the main characteristics of melt batches, crystals and melt inclusions
1052 in Baia–Fondi di Baia eruption.































Sample	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Total	Sample	F	P ₂ O ₅	Cl	MgO _{trace}
BFB5_vt1	56.33	0.72	18.25	5.93	0.14	1.95	5.36	3.33	7.99	95.23	BFB5_vt1	0.186	0.418	0.514	1.89
<i>n</i> =13	0.46	0.07	0.14	0.21	0.06	0.10	0.25	0.08	0.18		n=9	0.009	0.023	0.007	0.09
BFB5_vt2	56.46	0.68	18.28	5.72	0.13	1.82	5.21	3.45	8.24	95.56	BFB5_vt2	0.179	0.379	0.510	1.721
<i>n</i> =6	0.39	0.07	0.09	0.08	0.04	0.02	0.13	0.11	0.06		n=9	0.019	0.018	0.010	0.056
BFB12_vt1	61.77	0.46	18.38	3.04	0.13	0.47	2.30	4.63	8.83	97.48	BFB12_vt1	0.198	0.077	0.649	0.477
<i>n</i> =11	0.49	0.04	0.21	0.12	0.04	0.02	0.06	0.07	0.09		<i>n</i> =15	0.021	0.007	0.020	0.027
BFB12_vt2	56.64	0.65	18.40	5.78	0.14	1.69	5.03	3.41	8.26	95.75	BFB12_vt2	0.177	0.366	0.506	1.61
<i>n</i> =47	0.37	0.05	0.15	0.13	0.05	0.03	0.13	0.06	0.07		<i>n</i> =10	0.022	0.016	0.009	0.05
BFB12_vt3	59.19	0.51	18.50	4.24	0.13	1.00	3.47	3.60	9.36	96.02	BFB12_vt3	0.158	0.210	0.521	0.950
n=20	0.51	0.05	0.16	0.16	0.04	0.08	0.17	0.10	0.14		<i>n</i> =10	0.025	0.019	0.056	0.087
BFB12_vt4	62.23	0.41	18.14	3.26	0.21	0.26	1.97	6.10	7.43	96.99	BFB12_vt4	0.354	0.038	0.911	0.263
n=20	0.33	0.04	0.17	0.09	0.07	0.01	0.07	0.07	0.06		n=9	0.015	0.005	0.014	0.018
BFB12_vt5	57.56	0.57	18.47	5.27	0.15	1.50	4.59	3.30	8.59	96.57	BFB12_vt5	0.158	0.324	0.489	1.44
<i>n</i> =12	0.38	0.04	0.13	0.16	0.05	0.04	0.13	0.04	0.11		n=9	0.015	0.010	0.009	0.06
BFB20_vt1	60.25	0.45	18.20	4.19	0.14	0.86	2.92	4.01	8.98	97.35	BFB20_vt1	0.178	0.166	0.626	0.82
<i>n</i> =17	0.40	0.06	0.14	0.18	0.05	0.07	0.17	0.08	0.08		<i>n</i> =10	0.014	0.021	0.034	0.10
BFB20_vt2	62.22	0.42	18.06	3.24	0.21	0.26	1.99	6.17	7.43	96.99	BFB20_vt2	0.362	0.041	0.818	0.25
<i>n</i> =47	0.58	0.05	0.23	0.09	0.06	0.02	0.07	0.11	0.07		n=9	0.019	0.005	0.054	0.01
BFB36_vt1	57.45	0.60	18.44	5.35	0.13	1.59	4.72	3.46	8.26	96.52	BFB36_vt1	0.184	0.332	0.519	1.53
<i>n</i> =27	0.63	0.05	0.18	0.11	0.06	0.04	0.11	0.07	0.07		n=10	0.015	0.021	0.007	0.08
BFB36_vt2	56.88	0.66	18.34	5.61	0.14	1.80	5.05	3.37	8.15	96.45	BFB36_vt2	0.174	0.372	0.493	1.66
<i>n</i> =12	0.26	0.05	0.20	0.14	0.05	0.04	0.13	0.08	0.08		n=10	0.017	0.010	0.012	0.05
BFB36_vt3	56.23	0.67	18.29	5.96	0.15	2.03	5.44	3.34	7.89	95.69	BFB36_vt3	0.182	0.415	0.504	1.86
n=20	0.56	0.04	0.22	0.13	0.04	0.04	0.12	0.08	0.07		<i>n</i> =10	0.014	0.013	0.005	0.04
BFB36_vt4	56.89	0.66	18.38	5.54	0.14	1.69	4.93	3.42	8.35	95.65	BFB36_vt4	0.160	0.346	0.492	1.56
<i>n</i> =10	0.38	0.08	0.23	0.16	0.03	0.08	0.20	0.07	0.13		<i>n</i> =10	0.020	0.014	0.020	0.10

 Table 1. Major element and volatile average compositions of matrix glasses

I able	2. PC	oint an	aryses	or me	It inclu	usions	nostee	1 in re	laspar	and cl	inopy	roxene	e crysta	ais. Ka	iw data	a is rep	portea	nere, j	prior to	o norn	ializat	ion an	a aver	aging.		
MI	2	2	5	5	5	5	6	7	7	9	9	9	9	10	10	11	12	12	12	24	24	26	28	28	28	29
	MI1	MI2	MI3	MI4	MI2	MI1	MI3	MI2	MI1	MI2	MI3	MI4	MI1	MI2	MI1	MI1	MI1	MI8	MI7	MI3	MI2	MI1	MI2	MI1	MI3	MI1
Sample									BFE	3 5 - felds	par									BFB 20 – feldspar						
SiO ₂	59.75	60.90	59.15	59.68	59.41	60.44	60.94	60.45	60.36	60.44	59.33	59.69	57.11	59.78	59.29	59.77	61.31	59.60	59.48	59.49	60.17	59.38	59.81	59.35	59.09	59.10
TiO ₂	0.33	0.32	0.40	0.35	0.32	0.34	0.23	0.28	0.38	0.43	0.38	0.42	0.35	0.39	0.35	0.42	0.34	0.32	0.31	0.28	0.31	0.28	0.31	0.33	0.39	0.45
Al ₂ O ₃	17.23	18.04	17.10	17.29	17.27	17.96	17.67	17.77	18.01	17.62	17.78	16.94	16.59	17.57	17.57	17.12	17.61	17.28	17.52	17.60	17.61	17.22	17.42	17.48	17.34	17.65
FeO	2.11	2.15	3.16	3.03	2.93	2.95	2.70	2.28	2.25	2.85	3.02	2.92	2.89	2.98	2.96	2.95	2.75	2.75	2.64	3.01	3.04	2.92	2.86	3.08	3.14	2.99
MnO	0.06	0.14	0.17	0.12	0.13	0.12	0.10	0.10	0.12	0.08	0.10	0.06	0.12	0.14	0.13	0.11	0.18	0.08	0.13	0.08	0.02	0.09	0.12	0.12	0.14	0.12
MgO	0.27	0.29	0.50	0.51	0.50	0.51	0.47	0.30	0.33	0.45	0.61	0.49	0.56	0.58	0.52	0.58	0.38	0.51	0.44	0.64	0.50	0.58	0.56	0.59	0.56	0.60
CaO	1.32	1.40	2.23	2.27	2.15	2.36	2.10	1.37	1.37	2.27	2.36	2.26	2.23	2.22	2.26	2.48	2.18	2.07	2.12	2.12	2.10	2.32	2.28	2.26	2.36	2.29
Na₂O	4.61	4.89	3.31	3.70	3.53	3.64	3.31	5.20	5.36	3.55	3.46	3.51	3.43	3.75	3.98	2.94	3.43	3.44	3.40	3.65	3.46	3.52	3.81	3.85	3.39	3.25
K₂O	8.17	8.56	8.91	8.69	9.14	8.63	9.39	7.91	7.97	9.08	9.15	9.22	8.60	8.70	8.47	9.74	9.08	9.11	9.17	8.92	9.21	9.07	8.55	8.41	8.91	9.21
P ₂ O ₅	0.051	0.088	0.057		0.081	0.057		0.046	0.038	0.081	0.133	0.065	0.125	0.093	0.066	0.053	0.032	0.038	0.019	0.088	0.106	0.117	0.116	0.122	0.136	0.085
Cl	0.384	0.509	0.468		0.382	0.433		0.583	0.591	0.443	0.417	0.405	0.375	0.388	0.374	0.261	0.519	0.308	0.433	0.362	0.343	0.372	0.366	0.365	0.393	0.397
F	0.128	0.179	0.106		0.079	0.132		0.225	0.215	0.116	0.140	0.123	0.143	0.136	0.119	0.053	0.129	0.076	0.113	0.094	0.107	0.129	0.102	0.124	0.133	0.109
H₂O	2.86	1.17		1.97		2.00		1.40	1.60				3.70		1.19	1.43	1.55			1.73	1.78	1.91			1.66	1.49
Tot	97.27	98.64	95.53	97.62	95.91	99.57	96.90	97.91	98.58	97.41	96.88	96.10	96.22	96.73	97.27	97.90	99.48	95.58	95.78	98.07	98.74	97.91	96.29	96.08	97.63	97.73
Туре	F ₂	F ₂	F1	F1	F1	F1	F1	F ₂	F ₂	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1	F1

Table 2. Point analyses of melt inclusions hosted in feldspar and clinopyroxene crystals. Raw data is reported here, prior to normalization and averaging.

MI	29	29	30	30	30	31	31	33	33	34	34	35	38	38	10	10	12	17	17	18	15	16	28	29	29	30
	MI2	MI3	MI1	MI3	MI2	MI2	MI3	MI2	MI1	MI1	MI2	MI1	MI1	MI2	MI1	MI2	MI1	MI2	MI1	MI1	MI1	MI1	MI1	MI1	MI2	MI1
Sample				BFB	20 - felds	par										BFB 20) – clinop	roxene					BFB 5 – c	linopyrox	ene	
SiO ₂	59.57	59.53	58.70	58.66	59.08	59.31	59.65	60.35	59.13	58.71	59.65	59.19	59.25	59.97	60.08	59.55	59.34	59.41	59.81	58.84	59.65		58.98	59.91	61.85	59.20
TiO ₂	0.31	0.34	0.40	0.44	0.40	0.35	0.40	0.47	0.41	0.32	0.36	0.40	0.29	0.33	0.34	0.35	0.47	0.54	0.50	0.42	0.28		0.42	0.39	0.17	0.44
Al ₂ O ₃	17.57	17.30	17.04	17.15	17.12	17.01	17.06	17.41	17.63	17.46	17.38	17.39	17.59	17.97	18.11	17.81	18.65	18.82	16.91	17.93	17.58		17.78	17.21	17.85	17.03
		2.46				2.00					2.4.6					• • • •		2.05	a a r		a <i>cc</i>					~ • •
FeO	3.15	3.16	3.00	2.86	3.01	3.06	3.04	2.74	2.42	3.17	3.16	2.75	2.80	2.79	2.79	2.93	2.87	3.06	2.95	2.98	2.66		2.93	2.88	2.48	3.44
	0.11	0.42	0.40	0.47	0.02	0.00	0.00	0.00	0.4.0	0.40	0.42	0.20	0.02	0.42	0.00	0.40	0.44	0.25	0.40	0.24	0.02		0.1.1	0.44	0.05	0.4.4
MINO	0.11	0.12	0.10	0.17	0.02	0.09	0.09	0.09	0.18	0.10	0.13	0.20	0.03	0.13	0.09	0.10	0.11	0.25	0.18	0.24	0.02		0.14	0.11	0.05	0.14
M-0	0.62	0.60	0.62	0.65	0.62	0.61	0 5 9	0.42	0.26	0.66	0.62	0.25	0.52	0.40	0.61	0.64	0.75	0.10	0.97	0.55	1 22		1 01	0.51	0.44	0.59
INIGO	0.02	0.00	0.05	0.05	0.05	0.01	0.56	0.42	0.50	0.00	0.02	0.55	0.52	0.49	0.01	0.04	0.75	0.19	0.87	0.55	1.22		1.01	0.51	0.44	0.56
0e0	2 3/	2 30	2 38	2 2 2	2 3 7	2.26	2 1 3	2 09	1 95	2 / 5	2 30	1 77	2 25	2.07	2 10	2 1 2	2 /0	1 73	2 90	2 27	3 81		3 10	2 16	2.05	2 20
cuo	2.54	2.50	2.50	2.55	2.57	2.20	2.15	2.05	1.55	2.45	2.50	1.77	2.25	2.07	2.15	2.15	2.45	1.75	2.50	2.27	5.01		5.15	2.10	2.05	2.25
Na ₂ O	3.26	3.34	3.38	3.12	3.08	3.73	3.80	4.22	3.98	3.33	3.38	5.29	3.47	3.55	3.18	3.45	3.24	4.68	3.83	4.69	2.77		4.61	3.40	3.17	3.25
	0.20	0.01	0.00	0.11	5100	0170	5.00		0.00	0.00	0.00	0.20	0117	0.00	0.10	0110	0.2		0.00		,			5110	0.127	0.20
K ₂ O	9.25	9.18	8.92	9.34	9.20	8.92	8.86	8.34	9.26	9.36	8.89	7.79	9.13	9.21	9.44	9.33	9.38	6.13	8.40	7.22	6.80		7.49	9.04	9.60	9.04
														•												
P ₂ O ₅	0.056	0.054	0.052	0.096	0.077		0.025		0.078	0.129	0.105	0.067	0.087	0.064	0.153	0.131	0.134							0.010	0.095	0.009
CI	0.353	0.341	0.367	0.359	0.342		0.364		0.652	0.366	0.364	0.697	0.407	0.373	0.396	0.412	0.392							0.427	0.418	0.418
F	0.115	0.102	0.143	0.125	0.130		0.104		0.218	0.104	0.113	0.231	0.123	0.094	0.132	0.114	0.132							0.141	0.138	0.138
H ₂ O		1.64	2.00		2.43		1.93	1.51	1.98	1.69		2.49		1.51	2.07			1.11	1.29	4.18	2.16	3.19	1.39			2.31
Tot	96.71	98.00	97.12	95.31	97.89	95.34	98.03	97.64	98.25	97.85	96.44	98.60	95.95	98.55	99.57	96.96	97.95	95.91	97.64	99.31	96.96	3.19	97.93	96.19	98.30	98.28
Туре	F1	F1	F1	F1	F1	F1	F1	F ₂	F ₂	F1	F1	F ₂	F1	F1												

Melt batches	Presence along eruptive sequence	Occurrence	TAS classification	Major elements wt%	Sr isotopes					
Α	Rare in Baia, increase in FdB	Groundmass glass	Tephriphonolite- latite	SiO ₂ =54.9-58 K ₂ O=7.4- 8.8 CaO>4.2 MgO>1.5	⁸⁷ Sr/ ⁸⁶ Sr 0.70745-0.70750					
В	Baia and FdB	Whole rock, groundmass glass	Trachyte	Trachyte $SiO_2=60.3-64.7 \text{ NaO} < 7$ $K_2O=6.5-8 \text{ MgO}=0.1-$ ⁸⁷ Sr 0.6 CaO=1.8-2.6						
С	Rare in Baia, increase in FdB	Groundmass glass, melt inclusions	Trachyte-phonolite	SiO ₂ =58-63 K ₂ O=8.1-9.6 CaO=2.7-4.2 MgO=0.8-4	⁸⁷ Sr/ ⁸⁶ Sr 0.70747-0.70753					
Feldspars	Presence along eruptive sequence	Composition		Zoning						
P1	Baia and FdB	An62-83, Or2-6	None. Resor	bed cores or inclusions in F	1 crystals					
P2	Baia and FdB	An30-55, Or6-11		Core in F ₂ feldspars						
F1	Baia and FdB	Or75-90, An3	Ze	oning due to Ba variability						
F2	Baia and FdB	Or35-74, An2-22	Ι	Phenocysts and microlites						
Melt inclusions	TAS classification	Major elements wt%)	Volatiles wt%						
F2-hosted MI (7)	Trachyte	SiO ₂ =57.1-60.9 K ₂ O=8. CaO=1.9-2.4 MgO=0.3-	3-9.7 •0.6	Cl=0.3-0.6 F=0.1-0.2 H ₂ O=1						
F1-hosted MI (33)	Trachyte	SiO ₂ =59.1-60.9 K ₂ O=7. CaO=1.3-1.7 MgO=0.2-	.7-8.5 -0.3 Cl=0.2-0.6 F=0.05-0.2 H ₂ O=1.1-3.7							

Table 3. Summary of the main characteristics of melt batches, crystals and melt inclusions in Baia-Fondi di Baia eruption