Magma chamber stratification of the 1815 Tambora caldera-forming eruption

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Abstract

The eruption of the Tambora volcano in 1815 was initiated by two precursory Plinian falls and formed two generations of pyroclastic density current (PDC) deposits. In this study, we found slight changes in phenocrysts (modal mineralogy, content, and size), bulk-rock and feldspar microlite chemical compositions, and bubble and microlite number densities through the stratigraphic position. Plinian fall units are characterized by a lower phenocryst abundance (avg. of 5.1%), smaller phenocryst size (avg. of 0.06 mm²), and higher silica content (bulk-rock, 58–58.5 wt.%). The PDC deposits are characterized by a relatively higher crystal abundance (avg. of 12.1%), larger phenocryst sizes (avg. of 0.13 mm²), and lower silica content (bulk-rock, 56.7–57.9 wt.%). Therefore, the deposit stratigraphy and analyses suggest that phenocryst stratification in the magma chamber was established prior to the 1815 eruption and was thus responsible for yielding a slight contrast in bulk compositions. Feldspar microlite moves toward slightly more albitic compositions from Plinian falls to the PDCs, suggesting a slight decrease in the initial melt temperature from the upper to the lower magma chamber portion. Because the Plinian eruptions extracted the hottest magma, the degree of supercooling became low, and consequently yielded microlite-poor juveniles. By contrast, the PDCs experienced a larger degree of supercooling because the temperature was relatively low, thus yielding microlite-rich juveniles. Moreover, such temperature stratification coupled with the evidence of homogeneous melt composition (58.5–58.9 wt.% SiO₂) and the minor evidence of crystal mush (at most 27%) might suggest that the Tambora case is still in the early stage of magmatic evolution under cooling from the surrounding rocks.

Keywords Tambora · Caldera-forming eruption · Magma chamber stratification · Bubble number density · Microlite number density

Introduction

Caldera-forming eruptions are associated with large and long-lived magma reservoirs (Cashman and Giordano 2014; Bachmann and Huber 2016). They may occur over days to weeks and are often associated with the formation of an

extensive pyroclastic density current (PDC) that spreads in all directions (Sigurdsson and Carey 1989). As a result, caldera-forming eruptions are a threat to civilization and have caused tens of thousands of fatalities in the past, such as the volcanic eruption of Samalas in 1257 (Lavigne et al. 2013; Vidal et al. 2015), the eruption of Tambora in 1815 (Self et al. 1984; Foden 1986; Sigurdsson and Carey 1989; Gertisser et al. 2012), and the eruption of Krakatau in 1883 (Dahren et al. 2012; Gardner et al. 2013). It is therefore important to use detailed textural and geochemical studies to reconstruct the magmatic systems and eruption mechanisms associated with caldera-forming eruptions.

Many researchers have conducted intensive studies on some famous caldera-forming eruptions, such as the 39.8 ka Campi Flegrei eruption, the 12 ka Laacher See eruption, and the eruption of Vesuvius in 79 AD. Such studies have demonstrated that fieldwork, detailed stratigraphy, systematic sampling, grain size distribution, componentry, chemical characteristics, and textural observation are important tools for understanding the important

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Fig. 1 (A) The location of the Tambora volcano is highlighted by red square. (B) Sampling locations of this study. Symbols indicate the juvenile types as shown in (C). (D) Grain size (-6ϕ to 8ϕ) distribution and componentry results in terms of pumice, scoria, free crystal, and lithic

aspects of caldera-forming eruptions (Rosi et al. 1996, 1999; Ginibre et al. 2004; Gurioli et al. 2005). The eruption of Tambora in 1815 is considered to be one of the most explosive eruptions in history and has attracted global interest, leading to a better understanding of the stratigraphy, tephrochronology, petrogenesis, volume estimation, and climatic impact of volcanic eruptions (Self et al. 1984; Foden 1986; Sigurdsson and Carey 1989; Oppenheimer 2003; Self et al. 2004; Gertisser et al. 2012; Kandlbauer and Sparks 2014). However, several constraints on magma chamber dynamics and their correlation with the textures of juvenile materials remain poorly understood. In particular, the linkage between magma compositions and quantitative descriptions of the phenocrysts, vesicles, and microlites preserved in eruption products has not been demonstrated. The objective of this study is to advance our understanding of the 1815 Tambora caldera-forming eruption by revealing relations between the temporal changes in stratigraphy and the variation in the phenocrysts, magma compositions, vesicles, and microlite textures associated with the 1815 eruption.

In this study, we report the results of fieldwork conducted on the southern to northern flanks of the Tambora volcano. Samples were collected systematically as a function of the stratigraphic position. Such systematic sampling is important for identifying possible correlations between the stratigraphy of the eruption deposit and the construction of the magma chamber prior to eruption. Herein, we provide a detailed description of the individual stratigraphic members. Subsequently, we report the chemical compositions of plagioclase crystals (phenocrysts and microlites), and the results of quantitative analyses for crystal content, bubble and microlite number densities (BND and MND), and the magma decompression rate (dP/dt), with a focus on a comparison between Plinian fall and PDC deposits originating from the 1815 eruption. The chemical compositions of the plagioclase crystals and groundmass glass, crystal content, crystal size, and modal mineralogy were used to understand the condition of the pre-eruptive magma chamber. The BND, MND, and magma decompression rate were used to evaluate the role played by

Fig. 2 (A, B) Outcrop image of LOC 1A and (C, D, E) LOC 1B, which consists of paleosol, F-1, F-2, F-3, F-4, and PDC-1. (F, G) LOC 3 has a similar stratigraphic succession to LOC 1A and B, but the grain size is typically smaller. (H) Outcrop image of PDC-2 in the southern part and (I,

J) in the northern part of Tambora. (K) The complete stratigraphy succession of the 1815 eruption exposed in the western caldera wall. LB lag-breccia, BT brown-tuff formation

conduit processes in the generation of the texture of juvenile materials. We examine how pre-eruptive conditions in the

magma chamber play an important role in controlling the generation of the final texture of the pyroclasts.

Fig. 2 (continued)

The 1815 eruption

The Tambora volcano is located on the Sanggar Peninsula of Sumbawa Island (Fig. 1A), approximately 180 km above the active subduction zone of the Indo-Australian plate. The catastrophic 1815 eruption of the Tambora volcano occurred after a long period of inactivity when magma reached the most evolved composition of trachyandesite to tephriphonolite (Self et al. 1984; Foden 1986; Sigurdsson and Carey 1989; Oppenheimer 2003; Gertisser et al. 2012). The two most explosive events of the Plinian eruption occurred on April 5 and April 10 (each event was initiated by a phreatomagmatic eruption) and formed an eruption column height of 33 and 43 km above sea level, respectively (Sigurdsson and Carey 1989; Oppenheimer 2003). On the evening of April 10, the buoyant Plinian eruptive column suddenly shifted to a highly destructive PDC (Self et al. 1984; Sigurdsson and Carey 1989; Oppenheimer 2003). Sigurdsson and Carey (1989) distinguished such PDC deposits as crudely

stratified pyroclastic surges and massive pyroclastic flows. The pyroclastic airfall deposits reached distances of up to 1800 km and an approximate thickness of 0.1 cm (Kandlbauer and Sparks 2014). The ejection of 53–58 Tg of SO_2 and 93–118 Tg of sulfate aerosols into the stratosphere is likely responsible for generating the event called “the year without a summer” (Self et al. 2004). Foden (1986) and Gertisser et al. (2012) demonstrated that the 1815 magma reservoir was stored approximately between 1.5 and 7.5 km below the surface and contained 2.5 to 5.9 wt.% H_2O . The volume of the 1815 eruption has been estimated to be between 30 and 50 km^3 DRE (Self et al. 1984; Sigurdsson and Carey 1989; Kandlbauer and Sparks 2014).

Stratigraphy

The deposits consist of four early fall products (F-1, F-2, F-3, F-4), pyroclastic surges (S-1), and pyroclastic flows (PF-1). Based on color and vesicle characteristics, we identified four

Fig. 3 Stratigraphy correlation of the 1815 eruption. Yellow stars in the stratigraphic column correspond to the sampling positions

main types of juvenile products as follows: pale brown pumice, dark brown pumice, scoria, and bombs (Fig. 1C). Pale brown pumice is characterized by abundant small vesicles, whereas dark brown pumice is characterized by a relatively large vesicle size compared to pale brown pumice. The scoria and bombs are black. Scoria has vesicles that are similar to dark brown pumice. Bombs are typically dense, poorly vesiculated, and display distinctive surface textures (e.g., breadcrumb and cauliflower). Here, we used an abbreviation similar to that used in previous studies on fall deposits, such as F-2 and F-4 (Sigurdsson and Carey 1989). However, we use the abbreviation of PDC instead of the pyroclastic surge and pyroclastic flow used in a previous study (Sigurdsson and Carey 1989). Because PDC-1 and PDC-2 are typically pumice-rich and scoria-rich, we labeled these units pumice-rich PDC-1 and scoria-rich PDC-2, respectively.

We focused on relatively distal locations along the southern to the northern part of the Tambora volcano. LOC 1A and B are located at altitudes of 540 and 600 m, and approximately 18 and 19 km northwest from the central part of the caldera (Fig. 1B). LOC 1A has an approximately 2 m succession consisting of the following sequence: paleosol, first phreatomagmatic fall (F-1), first Plinian fall (F-2), second phreatomagmatic fall (F-3), second

Plinian fall (F-4), and pumice-rich pyroclastic density current (PDC-1) (Fig. 2A, B). We also observed a weak 20-cm thick, laminated layer at the lowermost portion of PDC-1 (Fig. 2A). LOC 1B has an approximately 6-m succession, which is relatively similar to the stratigraphy of LOC 1A. However, we identified a third phreatomagmatic fall deposit (F-5) just below PDC-1 in LOC 1B (Fig. 2C). We also observed a thicker deposit of PDC-1 in LOC 1B that includes several concentrated layers of pumice (Fig. 3D, E) and large (approximately 20 cm) carbonized wood (Fig. 2D). Plinian fall deposits in both locations are typically lithic-rich with reverse and normal graded structures. We also observed that most pumice in the fall and PDC-1 deposits were pale and dark brown, respectively. The grain size varies from 0.02 to 2.3 cm in the first Plinian fall and varies from 0.02 to 2.7 cm in the second Plinian fall. The first Plinian fall contains a smaller fraction of lithics than the second Plinian fall. PDC-1 is typically rich in pumice and lithic clasts. Most of the pumice fragments are dark brown and vary from 0.01 to 40 cm in size, whereas the matrices are typically composed of coarse ash. Larger pumice fragments are present mostly in the pumice-concentrated layers.

LOC 3 represents the most distal facies, approximately 25 km from the central part of the caldera. At LOC 3, the total

Fig. 4 Mineral assemblages and representative petrographic images of Tambora 1815 juvenile products with respect to stratigraphic position in the stratigraphic column. Yellow star symbol corresponds to sampling

thickness of the 1815 deposit was approximately 2.5 m and is overlain by multiple sequences of lahar deposits (Fig. 2G). Compared to the previous locations, LOC 3 preserves a finer grain size of pumice fragments and matrix ash (Fig. 2F), such that the pumice grain size varies from <0.02 to 1 cm in the first Plinian fall and from <0.02 to 1.2 cm in the second Plinian fall (Fig. 2F). Pumice from both Plinian falls is similar to LOC 1,

position. Pie charts illustrate the mineral assemblages. Plinian falls are typically plagioclase dominated, pyroxene-poor, and olivine-free. While PDCs are pyroxene-rich with rare-olivine

which is typically pale brown. The pumice fragments of PDC-1 are mostly dark brown, reaching up to 5 cm, and the matrix is typically composed of fine ash.

We also found scoria-rich PDC-2 in the southern (LOC 4) and northern (LOC 2) areas of the Tambora volcano, with thicknesses varying from 4 m in the south to 12 m in the north. In the south, PDC-2 lacks pumice and is rich in bombs with a

Table 1 Summary of petrography analysis of 1815 eruption products

Sample	Phases	Mode, vol. %	Crystal content (%)	Average crystal size (mm ²)	Zoning	Crystal aggregate (%)	Fraction of crystal aggregate within crystal content (%)
F2-5 <i>n</i> =3	pl	0.2	0.2	0.01	normal, non-zoned	0	0
	cpx	-			-		
	bt	-			-		
	opq	-			-		
F2-4 <i>n</i> =3	pl	4.3	5.1	0.08	normal, non-zoned	0.6	11.3
	cpx	0.1			non-zoned		
	bt	0.6			non-zoned		
	opq	0.1			non-zoned		
F2-3 <i>n</i> =3	pl	5.2	5.2	0.05	normal, non-zoned	0	0
	cpx	-			-		
	bt	-			-		
	opq	-			-		
F2-2 <i>n</i> =3	pl	4.8	5.2	0.05	normal, non-zoned	0.8	14.7
	cpx	0.2			non-zoned		
	bt	0.1			non-zoned		
	opq	0.1			non-zoned		
F2-1 <i>n</i> =3	pl	3.7	3.9	0.04	normal, non-zoned	0	0
	cpx	-			-		
	bt	-			-		
	opq	-			-		
F4-3 <i>n</i> =3	pl	5.5	6.2	0.08	normal, non-zoned	0.8	13.2
	cpx	0.4			non-zoned		
	bt	0.2			non-zoned		
	opq	0.1			non-zoned		
F4-2 <i>n</i> =3	pl	6.7	7.9	0.12	normal, non-zoned	1.3	15.9
	cpx	0.6			non-zoned		
	bt	0.4			non-zoned		
	opq	0.2			non-zoned		
F4-1 <i>n</i> =3	pl	4.8	7.1	0.08	normal, non-zoned	1.1	15.9
	cpx	0.7			non-zoned		
	bt	1.1			non-zoned		
	opq	0.5			non-zoned		
PDC1-10 <i>n</i> =3	pl	3.7	5.7	0.08	normal, non-zoned	1.2	21.8
	cpx	1.3			non-zoned		
	bt	0.3			non-zoned		
	opq	0.4			non-zoned		
PDC1-9 <i>n</i> =4	pl	8.6	10	0.09	normal, non-zoned	3.3	33.3
	cpx	0.7			normal, non-zoned		
	bt	0.6			non-zoned		
	opq	0.1			non-zoned		
PDC1-8 <i>n</i> =3	pl	10.3	11.5	0.13	normal, non-zoned	2.0	17.7
	cpx	0.4			non-zoned		
	bt	0.5			non-zoned		
	opq	0.3			non-zoned		

Table 1 (continued)

Sample	Phases	Mode, vol. %	Crystal content (%)	Average crystal size (mm ²)	Zoning	Crystal aggregate (%)	Fraction of crystal aggregate within crystal content (%)
PDC1-7 <i>n</i> =3	pl	6.2	8.7	0.13	normal, non-zoned, oscillatory	2.1	23.8
	cpx	1.6			normal, non-zoned		
	bt	0.5			non-zoned		
	opq	0.4			non-zoned		
PDC1-6 <i>n</i> =3	pl	10.5	11.7	0.18	normal, non-zoned	2.9	24.6
	cpx	0.5			normal, non-zoned		
	bt	0.4			non-zoned		
	opq	0.3			non-zoned		
PDC1-5 <i>n</i> =3	pl	7.7	9.0	0.10	normal, non-zoned	2.3	25.8
	cpx	0.8			normal, non-zoned		
	bt	0.3			non-zoned		
	opq	0.2			non-zoned		
PDC1-4 <i>n</i> =4	pl	9.4	12.2	0.10	normal, non-zoned	3.1	25.7
	cpx	2.2			normal, non-zoned		
	bt	0.5			non-zoned		
	opq	0.1			non-zoned		
PDC1-3 <i>n</i> =3	pl	8.6	10.5	0.09	normal, non-zoned	1.6	15.5
	cpx	1.6			normal, non-zoned		
	bt	0.2			non-zoned		
	ol	0.1			non-zoned		
PDC1-2 <i>n</i> =3	opq	0.1	10	0.10	non-zoned	1.7	17.0
	pl	8.9			normal, non-zoned		
	cpx	0.8			non-zoned		
	bt	0.2			non-zoned		
PDC1-1 <i>n</i> =3	opq	0.1	15.9	0.18	non-zoned	5.1	31.8
	pl	14.2			normal, non-zoned, oscillatory		
	cpx	1.1			normal, non-zoned		
	bt	0.4			normal, non-zoned		
PDC2-B <i>n</i> =3	ol	0.1			non-zoned		
	opq	0.1	13	0.17	non-zoned	1.9	18.1
	pl	10.1			normal, non-zoned, oscillatory		
	cpx	2.5			normal, non-zoned		
PDC2-S <i>n</i> =4	bt	0.3			non-zoned		
	opq	0.1	13.1	0.16	non-zoned	3.2	24.4
	pl	10.3			normal, non-zoned, oscillatory		
	cpx	2.4			normal, non-zoned		
PDC2-P <i>n</i> =4	bt	0.3			normal, non-zoned		
	opq	0.1	26.6	0.18	non-zoned	5.4	20.3
	pl	18.6			normal, non-zoned, oscillatory		
	cpx	5.2			normal, non-zoned		
	bt	1.5			normal, non-zoned		
	ol	0.4			non-zoned		
	opq	0.9			non-zoned		

Table 2 Representative whole rock chemical analysis result of Tambora samples

Sample Age	Pumice F 2-5 1815 AD	Pumice F 2-4 1815 AD	Pumice F 2-3 1815 AD	Pumice F 2-2 1815 AD	Pumice F 2-1 1815 AD	Pumice F 4-2 1815 AD	Pumice PDC 1-7 1815 AD	Pumice PDC 1-4 1815 AD	Pumice PDC 1-2 1815 AD	Scoria PDC 2S 1815 AD	Bomb PDC 2B 1815 AD	Pumice PDC 2P 1815 AD	Lava 1 Pre-1815 AD	Lava 2 Pre-1815 AD
wt%														
SiO ₂	58.01	57.98	58.15	58.27	58.46	58.32	57.91	57.69	57.93	57.65	57.75	56.73	52.71	54.23
TiO ₂	0.69	0.69	0.68	0.68	0.67	0.67	0.70	0.71	0.70	0.70	0.70	0.76	0.90	0.68
Al ₂ O ₃	18.66	18.60	18.57	18.53	18.49	18.55	18.62	18.60	18.60	18.58	18.56	18.60	19.90	20.29
MnO	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.14
MgO	2.12	2.13	2.05	2.02	2.02	2.03	2.15	2.24	2.12	2.14	2.15	2.51	3.43	2.03
CaO	3.96	4.05	3.91	3.82	3.70	3.75	4.07	4.29	4.08	4.18	4.10	4.74	10.71	7.87
Na ₂ O	4.35	4.41	4.44	4.48	4.54	4.53	4.34	4.31	4.42	4.59	4.52	4.27	2.93	4.76
K ₂ O	5.75	5.73	5.84	5.89	5.93	5.98	5.71	5.59	5.72	5.84	5.87	5.34	2.67	4.22
P ₂ O ₅	0.33	0.33	0.33	0.32	0.31	0.31	0.34	0.34	0.33	0.34	0.35	0.37	0.35	0.37
Fe ₂ O ₃	5.47	5.43	5.38	5.30	5.24	5.23	5.55	5.62	5.47	5.35	5.35	5.79	9.54	6.14
Total	99.56	99.56	99.57	99.56	99.57	99.58	99.61	99.59	99.58	99.39	99.57	99.41	102.64	100
F	0.05	0.05	0.04	0.06	0.05	0.04	0.00	0.03	0.03	0.04	0.03	0.07	0.06	0.04
Cl	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.08	0.09	0.27	0.10	0.22	0.06	0.07
ppm														
Sc	8.33	7.67	8.13	7.49	7.87	7.13	7.83	9.10	9.48	8.83	8.69	10.52	21.23	13.32
V	113.46	110.26	110.71	105.69	105.61	101.89	116.79	121.24	114.12	112.10	110.92	141.62	281.18	159.68
Cr	7.29	7.03	6.25	7.28	6.37	5.36	7.99	9.85	9.66	14.92	8.04	12.17	18.15	12.77
Cu	31.88	25.03	25.04	22.42	19.98	21.85	28.65	31.64	25.92	27.09	26.63	46.56	77.85	142.11
Zn	100.42	99.26	99.56	101.05	99.20	98.26	98.63	99.11	100.19	97.58	98.07	97.41	73.77	71.89
Rb	141.82	141.59	144.54	146.10	146.66	145.26	140.67	138.37	141.28	144.12	143.55	141.56	51.29	128.62
Sr	972.46	972.67	954.63	952.56	934.09	934.52	980.03	1013.96	983.91	978.34	977.45	1048.40	987.67	1346.51
Y	25.82	26.08	26.32	26.52	26.63	26.30	25.84	25.76	25.88	25.94	25.99	25.71	18.32	22.25
Zr	203.37	202.33	203.25	204.48	205.71	203.91	203.12	200.81	202.44	201.85	202.85	195.44	128.16	176.99
Nb	9.81	9.76	9.75	9.82	9.97	9.78	9.94	9.37	9.68	9.53	9.87	9.17	5.00	7.51
Ba	1245.32	1234.83	1236.07	1229.96	1235.63	1235.68	1229.74	1236.91	1238.72	1237.09	1255.75	1215.23	638.11	1003.36
Ce	115.26	116.60	120.60	121.68	116.28	119.77	118.71	123.38	118.97	111.45	116.58	120.88	87.53	101.99
Hf	3.88	3.89	3.90	3.86	3.88	3.83	3.88	3.92	3.89	3.91	3.86	3.98	4.40	4.06
Ta	3.01	2.72	2.71	3.18	3.07	2.92	2.84	2.62	3.46	2.93	2.58	3.01	1.76	1.79
Pb	16.94	17.36	17.74	17.22	17.83	17.00	16.26	16.11	17.10	15.27	15.97	16.10	9.13	13.96
Th	14.11	14.38	14.41	13.61	14.45	14.61	14.14	13.40	13.92	14.00	14.06	12.44	6.90	10.22
U	2.09	2.21	2.00	2.54	2.66	1.95	1.89	1.59	1.94	2.11	2.43	2.08	0.11	0.95

lower fraction of lithic content compared to the northern area (Fig. 2H). The size of scoria fragments ranges from 3 to 20 cm, whereas the bombs varies from 40 to 80 cm (Fig.

2H). In the north, the PDC-2 unit contains fragments of scoria, dark brown pumice, and bombs that vary in size from 3 to 30 cm. However, the abundance of dark brown pumice,

◀ **Fig. 5** (A) Representative major and trace elements of Tambora eruptive products. Analyses from this study are shown by symbols with blue, red, and black filled color. Gray colored symbols correspond to data from previous researchers. All analyses are normalized to 100 wt.% free of volatiles. (B) Ternary feldspar diagram of the Plinian falls (upper left), PDC-1 (upper right), and PDC-2 (bottom left). The plagioclase phenocrysts of PDCs are slightly more calcic compared to the Plinian falls. Slight change in microlite compositions can be observed from the Plinian falls to the PDCs and correspond to the difference in initial temperature (bottom right; indicated by yellow arrow). Isothermal sections of the dry ternary feldspar solvus follow the diagram used by Gertisser et al. (2012), which was calculated using SOLVCALC (Wen and Nekvasil 1994). Chemical data of plagioclase phenocrysts and microlites from Gertisser et al. (2012) are plotted in light gray and dark gray area, respectively

scoria, and bomb fragments is relatively poor and more lithic-rich compared to the southern deposits. Various types of lithics with a maximum size exceeding 40 cm were observed. The most representative outcrop to correlate the 1815 stratigraphic succession is preserved in the caldera wall, which became the backbone of our study (Fig. 3).

Analytical method

Sampling, grain size, and componentry

All of the sampled outcrops were fresh and not weathered, and lahar deposits were avoided (Fig. 3). We collected samples from Plinian falls and PDCs, including sublayers. To avoid contamination from the upper layers (i.e., sampling from the bottom might result in collapse of the upper layer), sampling was started from the top to the bottom layer of the deposit and the surface of the sampled layer was cleaned before sampling. However, we could not perform such systematic sampling for the entire PDC deposit because of its significant thickness and limited accessibility. The collected samples were sieved using -6ϕ to 8ϕ sieves. Subsequently, we conducted qualitative and quantitative analyses (weight percentage) of the grains within the -6ϕ to 4ϕ sieve size for componentry.

Petrography

A total of 65 thin sections of the most dominant juvenile types from each unit were observed under an optical microscope. The thin sections for Plinian units are represented by pale brown pumice, whereas those for PDC-1 and PDC-2 are represented by dark brown pumice, and scoria and bombs, respectively (Fig. 4). From each sampling location, we studied three to four different thin sections.

The crystals were divided into phenocrysts ($>100\ \mu\text{m}$) and microlites ($<100\ \mu\text{m}$). Phenocrysts are presumed to have formed in the magma chamber prior to eruption, whereas microlites formed in the conduit during the eruption. We measured the crystal size and crystal content of the phenocrysts

through a stratigraphic section to obtain the information on the magma chamber prior to eruption. The variation in crystal size is represented by the function of the average crystal size, which can be expressed by the following equation:

$$\bar{A} = \sum_{n=1}^N An / N \quad (1)$$

where $\bar{A}$ is the average crystal size in mm^2 obtained from the image processing program, $\sum_{n=1}^N An$ is a summation of the crystal size An in mm^2 , and N is the number of crystals in each thin section.

The crystal percentage (vesicle-free) is expressed by the following equation:

$$\varphi_c(\%) = \left(\sum_{n=1}^N Ac / (A_t - \sum_{n=1}^N Av) \right) \times 100 \quad (2)$$

where $\varphi_c(\%)$ is the crystal percentage, $\sum_{n=1}^N Ac$ is the summation of the crystal area Ac , $\sum_{n=1}^N Av$ is the summation of the vesicle area Av (a parameter that provides the bulk vesicularity), and A_t is the area of the sample in the thin section.

Chemical analyses

Bulk-rock chemical compositions were collected using a RIGAKU ZSX Primus IV at the Petrology and Volcanology Laboratory of Kyushu University. We selected 14 samples of a juvenile material from F-2, F-4, PDC-1, PDC-2, and lithic clasts of lava prior to the 1815 eruption. Juvenile fragments were manually picked from whole rock samples using tweezers to avoid lithic fragment contamination. The size of the selected juvenile clasts varied from 4 to $>32\ \text{mm}$. Subsequently, these juvenile and lithic samples were crushed into powder using a ball mill machine and pressed using a briquet press machine to make pellets. Our experimental policy required 100 g of homogenized powder of the juvenile materials such that each pellet consisted of 5–6 g of powder. Finally, the pressed samples were analyzed to determine the major and trace element compositions.

Mineral compositions of plagioclase crystals (phenocryst and microlite) and groundmass glass were obtained using a JEOL JXA 8530-F field emission electron microprobe (EPMA) at the Faculty of Science of Kyushu University. Point analyses were conducted using a focused beam current with a diameter of $3\ \mu\text{m}$ and an accelerating voltage of 15 kV. We analyzed 832 plagioclase phenocrysts and 70 plagioclase microlites from all eruptive units. For phenocrysts, we used the average anorthite values measured at three locations (close distance between each point, similar zones) in the core and rim of each plagioclase. The microlites are represented by only their core compositions because the thickness of the rim was too small for analysis (i.e., less than the beam diameter).

Table 3 Mean major element of the glass compositions of the 1815 products

Sample	Pumice Plinian falls 1815 AD (F2, F4) <i>n</i> =46	Pumice PDC 1 1815 AD <i>n</i> =24	Scoria + Bomb PDC 2 1815 AD <i>n</i> =38
wt%			
SiO ₂	58.89	58.51	58.84
TiO ₂	0.54	0.59	0.54
Al ₂ O ₃	19.59	19.91	20.08
MnO	0.20	0.20	0.21
MgO	1.43	1.59	1.42
CaO	3.37	3.53	2.70
Na ₂ O	5.23	5.04	5.32
K ₂ O	6.32	6.11	6.74
FeO	4.38	4.43	4.11
Total	99.9	99.99	99.99

Textural analyses

Textural analyses of vesicles and microlites from the thin sections of 1815 samples were conducted using images captured by a HITACHI TN3030 Plus scanning electron microscope (SEM) at the Petrology and Volcanology Laboratory of Kyushu University. The selected clasts were from 4 mm to >32 mm in size. We used the rim of the bombs to obtain the most representative area for a textural analysis. Vesicles and microlites from five different images at a 500x image magnification were manually digitized using Corel Draw X7 and such information of number and size was obtained from the image processing program (Image J). Next, the values from five different images were averaged to obtain BND and MND. For BND, we selected an area with the most homogeneous bubble size and made necessary corrections for the coalesced bubbles (de-coalesced) to obtain the most representative value (Klug and Cashman 1994). We used the following simple equations to obtain the BND and MND values:

$$BND (N_v) = (N_{av}/d_v) \times (1/1-\phi_{vi}) \quad (3)$$

$$MND (N_v) = (N_{am}/d_m) \quad (4)$$

where N_{av} is number density of bubbles per unit thin section area, d_v is average bubble size, ϕ_{vi} is the vesicularity of the groundmass, N_{am} is the number density of microlites per unit area, and d_m is average size of the microlite.

Results

Grain size distribution and componentry

Our stratigraphic descriptions are in good agreement with our componentry results (Fig. 1D). In particular, the first Plinian

fall was found to contain less coarser materials and lithic fragments in comparison to the second Plinian fall. The grain size of the PDC deposits varied widely (from ash to block) and included a substantial amount of lithic content.

Petrography

In general, juvenile materials of the 1815 eruption include plagioclase as its major phenocryst phase, followed by clinopyroxene, biotite, and Fe-Ti oxides with rare olivine. However, we found that Plinian falls and PDCs of the 1815 eruption exhibit stark differences in modal mineralogy and texture (Fig. 4). Plinian falls typically contain small phenocrysts (average of 0.06 mm²), crystal-poor (0.2%–7.9%), poor in aggregates (0%–1.3%, or 0%–15.9% within phenocryst), dominated by plagioclase crystals, and are free of olivine. By contrast, PDCs 1 and 2 typically contain large phenocrysts (average of 0.13 mm²), are crystal-rich (5.7%–13.6% for PDC-1 and 13%–26.6% for PDC-2), have more aggregates (1.2%–5.4%, or 15.5%–33.3% within phenocryst), and are rich in clinopyroxene, with some olivine (Fig. 4). Crystals that are zoned, non-zoned, or display sieve textures (glass inclusions) occur in all juvenile types, whereas oscillatory zoning is limited only in PDCs (Fig. 4, Table 1). In the groundmass, all eruptive units have similar mineralogy, which consists of plagioclase, anorthoclase, pyroxene, and Fe-Ti oxides. A summary of the petrographic observations is presented in Table 1.

Compositions of bulk rock, matrix glass, and plagioclase crystals

The major and trace element compositions of the representative samples are listed in Table 2. The juvenile samples from the 1815 eruption display a narrow range of silica content (bulk-rock, 56.7%–58.5% SiO₂), and the magmatic products of the pre-1815 lithics are less evolved (Fig. 5A). The abundance of Na₂O and K₂O shows a weak positive correlation with SiO₂, whereas CaO, Fe₂O₃, Sr, V, and Cr show a weak negative correlation with SiO₂ (Fig. 5A). Compared to the bulk-rock composition, the abundance of silica in the groundmass glass is slightly higher and is typically homogeneous (average of 58.5%–58.9 wt.%) (Fig. 5A). The compositions of the representative glasses are listed in Table 3.

The anorthite content of the plagioclase phenocrysts varies from An₄₀ to An₉₅ (Figs. 5B and 6). The average core anorthite composition of Plinian falls are slightly more evolved than the PDCs (An_{68–69} and An_{69–76}, respectively), whereas those of rim composition have similar value (An_{65–69} for both units). Plagioclase phenocrysts of Plinian falls show uniform composition (non-zoned), some normal zoning, and a lack of oscillatory zoning. Conversely, plagioclase phenocrysts of PDCs contain abundant normal zoning, with only some crystals displaying oscillatory or no zoning (Fig. 7A–D). We

Fig. 6 Slight change in bulk silica compositions, average crystal size, crystal content, and plagioclase phenocryst compositions as a function of stratigraphic height. Note that a significant increase in crystallinity yields a slight drop in bulk silica compositions (denoted by purple). The x-axes of histograms have the same bin size (interval of 5). The y-axis is

frequency in percentage (%). Blue and yellow bars correspond to compositions of core and rim, respectively. Note the slight change in the average core anorthite content through the stratigraphic positions (indicated by the blue star symbol)

found that the plagioclase microlite compositions vary from An₁₁ to An₄₅ and move toward more albitic compositions from the Plinian falls to the PDCs (Fig. 5B).

Quantitative description of textural analyses

Our data show a transition from Plinian falls to PDCs in terms of bubbles and microlites. In the Plinian falls, the BND values

range from $3.1 \times 10^{14} \text{ m}^{-3}$ to $4.2 \times 10^{14} \text{ m}^{-3}$, whereas the MND values range from $5.5 \times 10^{14} \text{ m}^{-3}$ to $7.7 \times 10^{14} \text{ m}^{-3}$. The BND value decreases and the MND value increases from Plinian toward the PDC units. In particular, dark brown pumice and scoria from PDCs have similar BND values (1.5×10^{14} – $2.1 \times 10^{14} \text{ m}^{-3}$) and high MND values (4.3×10^{15} – $7 \times 10^{15} \text{ m}^{-3}$ and 6.8×10^{15} – $9 \times 10^{15} \text{ m}^{-3}$, respectively). However, bombs have the lowest BND value (6.6×10^{13} –

Fig. 7 Representative BSE images of (A) non-zoned plagioclase, (B) zoned plagioclase, (C) aggregates of zoned plagioclase, and (D) oscillatory zoning texture. Blue and red dots correspond to measurement point of core and rim, respectively

$8.3 \times 10^{13} \text{ m}^{-3}$) and the highest MND value ($1 \times 10^{16} - 1.3 \times 10^{16} \text{ m}^{-3}$). Representative data for the BND and MND are listed in Table 4.

Magma decompression rate

We calculate the magma decompression rate similar to Toramaru (2006) as follows:

$$|dP/dt| = a \cdot D \cdot \sigma^2 \cdot P_W^{-1/3} \cdot T^{-1/2} \cdot N_V^{2/3} \quad (5)$$

where a is a constant (1×10^{15}), D is the diffusivity of water in a silicate magma (m^2/s), σ is the interfacial tension (N/m), P_W is the initial saturation pressure (Pa), T is the temperature (K), and N_V is the BND. The initial temperature, saturation pressure, and water concentration of the Tambora 1815 magma were estimated to be approximately 950 °C, 200 MPa, and 5 wt.%, respectively (Gertisser et al. 2012). Based on equations from Zhang and Behrens (2000) and Bagdassarov et al. (2000), the water diffusivity of Tambora magma was estimated to be approximately $5 \times 10^{-11} \text{ m}^2/\text{s}$ and the interfacial tension was approximately 0.124 N/m (for $T = 1342^\circ \text{K}$, $P_W = 200 \text{ MPa}$, and $\text{H}_2\text{O} = 5 \text{ wt.}\%$). It should be noted that the equation given by Bagdassarov et al. (2000) is for a rhyolitic melt composition; however, Tambora is alkali-trachyandesite in composition. Therefore, our calculations of interfacial tension generated an unrealistically high value that exceeds the limit of homogeneous nucleation. Based on an experiment, Gardner (2012) demonstrated that the interfacial

tension value for hydrous phonolitic melt (57.3 wt.% SiO_2 , 9.88 wt.% Na_2O , 5.44 wt.% K_2O) at 1423.5 K is between 0.065 and 0.068 N/m. Because Gardner (2012) provided a more realistic value, we used a moderate value (0.0665 N/m) of interfacial tension. Our calculation suggests that Plinian falls were generated from a higher magma decompression rate compared to PDCs, varying from 26 to 32 MPa/s and from 9 to 20 MPa/s, respectively.

Discussion

Magma chamber stratification

Although the products of the 1815 eruption show the homogeneity in melt compositions, the bulk compositions of Plinian falls and PDCs show slight differences in K_2O , Na_2O , CaO , and Fe_2O_3 , and some compatible elements (Sr, V, Cr) (Fig. 5A). We suggest that the observed difference in bulk composition is likely controlled by phenocryst variations (modal mineralogy and crystal content) in the pre-eruptive magma chamber. In particular, Plinian falls are typically pyroxene-poor, olivine-free, and have relatively small crystal sizes (average of 0.06 mm^2) and low crystal content (average of 5.1%). By contrast, PDCs contain higher amounts of pyroxene with rare olivine and a larger crystal size (average of 0.13 mm^2), and higher crystal content (average of 12.1%). Thus, the upper portion of the magma chamber tends to be more evolved in bulk compositions because of the relatively

Table 4 Textural parameter of Tambora 1815 samples

Sample	Bulk vesicularity (%)	Measured area for BND (mm ²)	Na	Average bubble diameter (mm)	Nv, BND (m ⁻³)	Estimated $\frac{dP}{dt}$ (Mpa s ⁻¹)	Measured area for MND (mm ²), non-corrected	Total number of microlites (from five measurement area)	Microlite volume fraction (%)	Average microlite diameter (μm)	Nv, MND (m ⁻³)
F 2-5 (pale brown pum.)	55	0.04	2817	0.0144	3.9×10^{14}	31	0.03	163	0.1	1.1	7.7×10^{14}
F 2-4 (pale brown pum.)	54	0.04	2592	0.0136	3.1×10^{14}	26	0.03	119	0.07	1	5.5×10^{14}
F 2-3 (pale brown pum.)	57	0.04	2483	0.0181	3.7×10^{14}	30	0.03	104	0.2	0.9	5.2×10^{14}
F 2-2 (pale brown pum.)	58	0.04	2775	0.0138	3.9×10^{14}	31	0.03	267	0.2	0.9	7.6×10^{15}
F 2-1 (pale brown pum.)	60	0.04	2817	0.0178	3.4×10^{14}	28	0.03	110	0.1	1	5.9×10^{14}
F 4-3 (pale brown pum.)	59	0.04	2842	0.0145	3.7×10^{14}	30	0.03	112	0.4	1.1	5.9×10^{14}
F 4-2 (pale brown pum.)	60	0.04	2767	0.0163	4.1×10^{14}	32	0.03	141	0.4	0.9	7.5×10^{14}
F 4-1 (pale brown pum.)	55	0.04	3012	0.0126	4.2×10^{14}	32	0.03	277	0.5	1	7.4×10^{15}
PDC 1-10 (dark brown pum.)	60	0.04	1625	0.0200	1.7×10^{14}	18	0.03	864	3.1	1.7	4.3×10^{15}
PDC 1-8 (dark brown pum.)	54	0.04	1583	0.0180	1.5×10^{14}	16	0.03	1083	5.7	1.9	7×10^{15}
PDC 1-1 (dark brown pum.)	57	0.04	1608	0.0193	1.6×10^{14}	17	0.03	684	2.1	1.4	5×10^{15}
PDC 2S-3 (scoria)	53	0.04	1600	0.0195	1.6×10^{14}	20	0.03	1824	2.5	2.2	6.8×10^{15}
PDC 2S-2 (scoria)	57	0.04	1650	0.0205	1.8×10^{14}	18	0.03	1581	18.6	2.5	8.2×10^{15}
PDC 2S-1 (scoria)	62	0.04	1817	0.0196	2.1×10^{14}	17	0.03	1542	8.3	2.1	9×10^{15}
PDC 2B-3 (bomb)	38	0.04	917	0.0247	6.6×10^{13}	9.4	0.03	3000	24.5	2.4	1.3×10^{16}
PDC 2B-2 (bomb)	41	0.04	11,058	0.0232	8.3×10^{13}	11	0.03	2202	13.1	1.9	1.1×10^{16}
PDC 2B-1 (bomb)	40	0.04	1041	0.0214	7.9×10^{13}	11	0.03	2775	11.3	1.8	1×10^{16}

Fig. 8 Representative BSE images of vesicles and microlites of Tambora 1815 juvenile product (pale brown pumice, dark brown pumice, scoria, and bombs). Large bubbles correspond to preexisting bubbles (pb) formed in magma chamber. Microlite is highlighted by red arrow. Yellow dot represents measurement point for glass compositions

low population of phenocrysts. In contrast, the lower portion of the magma chamber, which is inferred to have sourced the PDCs, is relatively phenocryst-rich with a higher abundance of mafic minerals; therefore, its bulk composition has the lower silica content (Fig. 6). This case is unusual because it provides evidence of phenocryst stratification in a nearly chemically homogeneous melt, whereas typical cases of zoned-magma chambers often involve striking variations in both phenocryst content and melt composition (i.e., from rhyolite to basalt or phonolite to basanite) (Fridrich and Mahood 1987; Bacon and Druitt 1988; Kaneko et al. 2007; Ginibre et al. 2004).

The evidence of homogeneous melt composition coupled with the absence of crystal mush (the most phenocryst-rich magma is approximately 27% crystallinity with at most 5.1% fraction of aggregate, whereas the typical evidence for crystal mush involves >50% crystallinity with ubiquitous crystal aggregate (e.g., Huber et al. 2012; Troch et al. 2017)) (Table 1) and the indication of hotter and lower temperature at the upper and lower parts of the chamber (obtained from feldspar microlite compositions, Fig. 6B) might suggest that the Tambora case is still in the early stage of magmatic evolution under cooling from the surrounding rocks (see Fig. 8 of Gutiérrez and Parada 2010). As numerically modeled by Gutiérrez and Parada (2010), during this early stage of magmatic evolution, temperature (together with crystal settling process that is simultaneously occurring) play an important role in controlling the melt migration and distribution via density variation, as the hotter melt rises to the upper part and the colder melt sinks to the lower part of the chamber (McBirney

2007; Gutiérrez and Parada 2010). The convection model is supported by the low viscosity behavior of trachyandesitic magma (at most 10^4 Pa s) (Takeuchi 2011) and the small phenocrysts are the product of a later crystallization stage or less-cooling condition at higher temperature at the upper portion of the magma chamber. Therefore, we suggest that crystal-settling and thermal convection are the likely processes responsible for the magma chamber stratification.

We also hypothesize that magma mixing might have taken place during the differentiation process by self-mixing and is possibly indicated by the occurrence of sieve textures (glass inclusion) in plagioclase phenocrysts. This is in agreement with Gertisser et al. (2012), who proposed that magma recharge-mixing played a role in forming the trachyandesitic-phonolitic magma erupted in 1815. However, we cannot comment on the extent of magma mixing given that we do not observe reverse zoning in plagioclase or mixing of different magma compositions in hand sample or outcrop (Sigurdsson and Sparks 1980; Yanagi et al. 1991).

In addition, Gertisser et al. (2012) also stated that there is no variation in the H_2O content recorded in the melt inclusions of calcic and sodic plagioclase phenocrysts. This might indicate a homogeneity of the water content in all 1815 magma. Furthermore, such information becomes extremely important for us to explain why PDCs contain a higher microlite number density (MND) (Fig. 9) with slightly more albitic compositions than Plinian falls. This is explained in the following section.

Fig. 9 Correlations between (A) MND and microlite crystallinity, (B) BND and MND, and (c) BND and bulk vesicularity. See text for discussion

Role of initial melt temperature in the generation of 1815 juvenile materials

In general, in decompression–vesiculation induced crystallization, the condition (pressure) of microlite crystallization and microlite compositions depends on the initial melt temperature, water content, and melt compositions (Cashman and Blundy 2000; Couch et al. 2003; Toramaru 2019). In the case of the same melt compositions as the present case, the initial melt temperature and water content control those. In this case, under the equilibrium vesiculation (we do not assume a disequilibrium process because of the high freedom and there is no unique solution for the chemical compositions of microlites (e.g., Martel and Schmidt 2003)) where the water content follows the solubility relation, the pressure of microlite crystallization is linked to the water content, which affects the increase in liquidus and the degree of supercooling. The microlite nucleation is controlled by the supercooling, that is, difference between liquidus and melt temperature.

Therefore, microlite crystallization may occur at the higher pressure (i.e., higher water content) for the colder magma and vice versa. Furthermore, the effect of water content and temperature on anorthite content of plagioclase microlites are similar, where higher water content and higher temperature results in higher anorthite content and vice versa (Couch et al. 2003). This mechanism to control the microlite characteristics is independent of the initial water content as long as the equilibrium vesiculation occurs. Therefore, we conclude that the melt temperature is the primary control of microlite crystallization. The fact that water content seems to be a constant as mentioned above does not deny this idea; or even if there is a difference in water content, microlite crystallization can still occur by following the water solubility curve. We interpret that the slight albite-rich feldspar microlite core compositions in the PDC-2 (the lower portion of the magma chamber) (Fig. 5B) is caused by a lower initial melt temperature.

Correlations between textural properties

We find the correlations between microlites and bubbles to be positive between MND and microlite crystallinity (Fig. 9A) (Martel and Poussineau 2007; Miwa et al. 2009) and negative between BND with MND (Fig. 9B). These correlations are interesting because, as we know, the BND and MND are the index of magma ascent rate and water exsolution rate (Toramaru 2006; Toramaru et al. 2008); hence, their correlation under the homogeneous nucleation condition should be positive at least theoretically. However, in this case, the correlation between BND and MND is negative, and the difference of MND values between the Plinian falls and PDCs are high (up to 1.5. order of magnitude within a slight decrease of BND). We think that the negative correlation between BND and MND is likely controlled by the difference in initial melt temperature, as the colder magma (PDCs) facilitates microlite crystallization at higher pressure (larger supercooling) than the hotter magma (Plinian falls). Moreover, the fact that the decompression rate (and BND) values between Plinian and PDCs are not significantly different imply that, in this case, decompression rate has no role in microlite crystallization. In addition, the increase in microlite crystallinity plays a role for the slight increase of K₂O content in the melt, especially for the PDC-2 samples (Fig. 5A). Namely an increase in the sodium-rich feldspar crystallization instead of potassium-rich feldspar causes the relative increase of K₂O concentration in the melt. However, there is no significant change in terms of silica content in the melt because crystallization of silicic minerals (i.e., quartz) did not occur, making the melt compositions of the 1815 eruption considerably homogeneous.

It is more difficult to explain the correlation of bubbles with bulk vesicularity (Fig. 9C) because the bulk vesicularity strongly depends on the quench pressure and magma viscosity, which limits the bubble expansion. It is inferred that

Fig. 10 (A) Pre-eruptive state of the 1815 magma chamber. Yellow and blue square represents the schematic images in the magma column in the chamber, which are the sources of the Plinian falls and the PDCs, respectively. Higher temperature melts are shown by light gray color and lower

temperature melts are shown by dark gray color. Note the changes in phenocryst abundance, microlite crystallization, and temperature of the uprising magmas from the Plinian phases (B, and C) to the PDCs (D)

bubbles in the Plinian falls that erupted at high temperature may expand more due to the low microlite content (i.e., temperature and microlite content controlling the effective magma viscosity) (Fig. 8). In addition, there is also the measurement effect because our vesicularity is bulk, including the phenocrysts and larger bubbles (preexisting bubbles) (Fig. 8).

Reconstruction of the 1815 eruption

Before the 1815 eruption, the magma chamber underwent slight phenocryst and temperature stratification in a nearly uniform trachyandesitic magma that was rich in preexisting bubbles (Fig. 8). Such preexisting bubbles (mostly) rise to the upper portion (Fig. 10A) and cause an overpressure (Parmigiani et al. 2016) that leads to the formation of fractures. The rising magma through the developed fractures will interact with the groundwater and trigger a phreatomagmatic eruption (phase F-1). This vent-opening phase is responsible for causing a sudden decompression, allowing magma to erupt explosively in

a Plinian fashion (F-2) because of the high magma decompression rate (26–31 MPa/s) (Fig. 10B). The estimated eruption column height and magma discharge rate during the first Plinian eruption were approximately 33 km and 1.1×10^8 kg/s, respectively (Sigurdsson and Carey 1989). The lower amount of lithic clasts in the first Plinian layer (Fig. 10) indicates that conduit erosion was sufficiently ineffective to generate a column collapse. After 2.1 h, the Plinian eruption was terminated and the eruption shifted to the second phreatomagmatic eruption (F-3) from April 5 to 10 (Sigurdsson and Carey 1989). During the second Plinian eruption (F-4) on April 10, magma experienced the highest magma decompression rate (30–32 MPa/s). As a result, the second Plinian yields a higher eruption column height and magma discharge rate compared to the first Plinian, exceeding 43 km and 2.8×10^8 kg/s, respectively (Sigurdsson and Carey 1989). Both Plinian eruptions were sourced from the upper portion of the magma chamber (approximately 1.6 km^3 DRE, Sigurdsson and Carey 1989), thus giving

rise to the extraction of the hottest magma and yields phenocryst- and microlite-poor juvenile material (Fig. 10B, C). The evidence of a high lithic content in the second Plinian layer (Fig. 1D) indicates that conduit erosion by the second Plinian eruption was stronger than that in the first Plinian. Consequently, intensive conduit erosion caused an enlargement of the conduit. When the conduit radius increases, column collapse occurs and generates PDCs (Fig. 10D) (Wilson et al. 1980; Woods and Wohletz 1991). During this stage, the magma discharge rate increased to 5×10^8 kg/s (Sigurdsson and Carey 1989), and the magma decompression rate decreased to 9–21 MPa/s for the next 30 h (Self et al. 1984). Because PDCs were sourced from the lower portion of the magma chamber with the coldest temperature, the resultant juvenile material are typically phenocryst and microlite rich (Fig. 10D).

Conclusions

The mineral assemblages, bulk rock compositions, crystal content, crystal size, and feldspar microlite compositions show that slight stratification of phenocryst and temperature in the magma chamber occurred before the eruption. This slight stratification of phenocryst and temperature in the magma chamber is responsible for yielding the slight variations in bulk-rock compositions and juvenile types in the top of the deposits. Furthermore, such temperature stratification plays an important role for controlling the syn-eruptive processes (i.e., degree of supercooling) and is ultimately responsible for determining the changes in the juvenile types (from dominantly pale brown pumice in Plinian falls to the dark brown pumice up to scoria in PDCs) in the top of the deposits.

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