ABSTRACT

THE EVOLUTION OF A CHEMICALLY ZONED MAGMA CHAMBER: THE 1707 ERUPTION OF FUJI VOLCANO, JAPAN

By Shizuko Watanabe

The eruptive history of Fuji Volcano has been dominated by basaltic volcanism. However, the 1707 eruption of Fuji Volcano resulted in a chemically zoned volcanic deposit ranging from basalt to dacite. Explosive, silicic eruptions such as this have occurred only twice in the history of Fuji Volcano, in 2800BP and in 1707AD, and the processes responsible for generating siliceous magmas beneath Fuji Volcano are not well understood.

Detailed petrographic, major and trace element, and Sr, Nd, Pb, and Os isotope studies were done in order to constrain the petrogenetic processes that formed a chemically zoned magma chamber prior to the 1707 eruption of Fuji Volcano. Kawamoto (1992) proposed that liquid immiscibility played a major role; however, this study shows that the chemical variations in the Fuji 1707 deposit can be explained by fractional crystallization of the observed mineral assemblage accompanied by minor crustal contamination.

THE EVOLUTION OF A CHEMICALLY ZONED MAGMA CHAMBER:
THE 1707 ERUPTION OF FUJI VOLCANO, JAPAN

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Chapter 1

Introduction to Fuji Volcano

Introduction

Fuji Volcano is a nearly symmetrical stratovolcano, which often appears in pictures and photographs as a symbol of Japan. It is the highest mountain in Japan with an elevation of 3776 m and a volume of approximately 1400 km³, located in south-central Honshu. It has been active since its formation in the late Pleistocene. The eruptive history has been dominated by basaltic volcanism; however, silicic eruptions have occurred twice, in 2800BP and in 1707AD. Although Fuji Volcano has been dormant for almost 300 years since the last eruption in 1707, increased seismic activity has been detected since the fall of 2000 (Asahi, 2001). Japan has one of the highest population densities and highest number of active volcanoes in the world; therefore, understanding the history of the volcanoes as well as the eruptive mechanism is essential to facilitate public safety in case of a large explosive eruption. The purpose of this chapter is to provide the geologic background to Fuji Volcano.

Geological setting of Fuji Volcano

Japan is an island arc located in the western Pacific, which is characterized by a linear chain of islands with frequent earthquakes and volcanic eruptions. Japan was originally part of the Eurasian continental margin (Ogawa et al., 1997) and the present day arc system began in the Neogene (Taira, 2001). The four main islands of Japan are, from north to south respectively Hokkaido, Honshu, Shikoku and Kyushu. The complicated arc-trench system of Japan is related to the Kuril, Honshu, Ryukyu, and Izu-Bonin (Ogasawara) arcs (fig. 1; Aramaki and Ui, 1982; Taira, 2001), and involves five plates: the Eurasia, Amur, Okhotsk, Pacific, and Philippine Sea plates. Fossa Magna is a weakly developed collisional zone (Ogawa et al., 1997) formed during the Miocene (Tsuya, 1968; Machida, 1981) located in central Honshu (fig. 1). Many quaternary volcanoes (fig. 2) are associated with Fossa Magna, and comprise the Fuji Volcanic Zone named after Fuji Volcano (Tsuya, 1968; Machida, 1981). Fuji Volcano is located in

Fig. 1. Arc-trench system of Japan (modified after Aramaki and Ui, 1982). Three chains of island arcs (North-east Honshu-Kurile, Izu-Bonin-Mariana, and Southwest Honshu-Ryukyu) meet in central Honshu (Aramake and Ui, 1982). The shaded area shown is the Fossa Magna. Fuji Volcano (■) is located in south Fossa Magna.

Fig. 2. Quaternary volcanoes (●) and their relationship to the volcanic front (modified after Aramaki and Ui, 1982). Fuji Volcano is shown in red square (■).

south-central Honshu, in the southern part of Fossa Magna, lying over the westward-subducting Pacific Plate at 130 to 150 km depth and possibly influenced by the northward-subducting Philippine Sea plate at 30 km depth (Arculus et al., 1991).

History of Fuji Volcano

Fuji Volcano is a relatively young volcano, which was formed around 80,000 years ago. Three distinct volcanoes including Komitake, Older Fuji, and Younger Fuji have been recognized, and make up the main edifice of the present Fuji Volcano (fig. 3). The basement rocks belong to the Misaka and Fujikawa groups, which consist of Tertiary marine sediments and volcanic rocks such as green tuffs (Miyaji, 1990). Due to the complicated growth history of the volcano, several classifications have been made. Tsuya (1940) of the former Earthquake Research Institute, who studied Fuji Volcano the most extensively in the past, divided the post-Komitake eruptions into Older Fuji and Younger Fuji based on the geology and petrology of lava flow and mudflow deposits. Miyaji (1988, 1990) recognized six periods of mudflow deposits within Older Fuji Volcano, while he subdivided Younger Fuji Volcano into six periods based on tephrostratigraphy combined with Carbon-14 dating. Machida (1967) divided the history of Fuji Volcano into three periods of Old Fuji I (80,000-10,000BP), Old Fuji II (10,000-8,000BP), and Younger Fuji (5,000BP-Present) based on tephrochronology. Fuji Volcano experienced quiescence between Old Fuji II and Younger Fuji, as evidenced by the formation of the Fuji Black humic ash layer between Old Fuji II and Younger Fuji (Machida, 1967).

Fig. 3. Cartoon diagram of Fuji Volcano (modified after Aihara, 2003). Three distinct volcanoes including Komitake, Older Fuji, and Younger Fuji make up the main edifice of the present Fuji Volcano.

Komitake Volcano

The oldest part of Fuji Volcano, Komitake Volcano (fig. 3) is located in the northern flank. A detailed history of Komitake Volcano is poorly understood due to the fact that it is buried underneath Older Fuji Volcano and Younger Fuji Volcano. Komitake Volcano is an andesitic stratovolcano with a summit crater. The elevation is approximately 2300 m.

The petrology and chemistry of the volcano are known to be similar to Ashitaka Volcano, a late Pleistocene volcano located to the south of Fuji Volcano, which mainly erupted olivine-bearing two pyroxene-andesite (Tsuya, 1968; Machida, 1981).

Older Fuji Volcano

Older Fuji volcano (fig. 3) began its activity on the southern flank of Komitake Volcano around 80,000 years ago. The degree of erosion as well as the length of the quiescent period after Komitake Volcano are still poorly understood. Tsuya (1968) differentiated Older Fuji Volcano from Younger Fuji Volcano based on several field observations. These observations include: 1) evidence of fault movement during or shortly after the eruption of Older Fuji; 2) all of the Older Fuji eruptive materials lie underneath those of Younger Fuji Volcano; 3) cemented mud flow deposits of Older Fuji are distinguishable from the porous deposits of Younger Fuji.

Eruptive products of Old Fuji Volcano include abundant tephra as well as basaltic lava and pyroclastic flows (Machida and Moriyama, 1968). These eruptions triggered large-scale mudflows later named Old Fuji mudflows. Miyaji (1990) recognized six different periods of mudflows: 1) 50,000 BP and older; 2) 50,000 BP; 3) 50,000-19,000 BP; 4) 19,000-18,000 BP; 5) 16,000 BP; and 5) 15,000-12,000 BP. The mudflow deposits are composed of several different types of basalts including aphyric basalt, olivine basalt, augite-olivine-basalt, hypersthene-augite-olivine-basalt, and hypersthene-olivine-basalt as well as xenolith rock fragments (Tsuya, 1968; Machida, 1981).

Younger Fuji Volcano

The eruptions of Younger Fuji Volcano (fig. 3) started near or on top of Older Fuji Volcano's summit around 11,000 years ago. Tsuya (1968) divided the eruptive materials of Younger Fuji Volcano into three units: 1) older ejecta; 2) middle ejecta; and 3) younger ejecta. This classification is based on the stratigraphy and petrology of the deposits. The deposits of

older ejecta are at the base of volcano, while the middle ejecta are deposited in mid-slope between 1,500 to 2,400 m. The deposits of younger ejecta are found in the upper-slope from 2,400 m and above.

The older ejecta consist of olivine basalts containing 5 to 10 millimeter plagioclase (anorthite or bytownite) phenocrysts. Nearly aphyric or pyroxene phenocryst-bearing basaltic lava flows are also observed. The middle ejecta comprise basaltic lava sheets interbedded with thin layers of pyroclastic deposits. They are aphyric to porphyritic basalts with olivine or olivine and augite and/or hypersthene pyroxenes. The younger ejecta are basaltic lava flows and pyroclastic deposits.

Tephrochronology

Machida (1967) introduced a tephrochronological classification based on the distribution of tephra combined with Carbon-14 dating. The east-facing gentle marginal slopes of Fuji volcano have been intensively studied due to the suitable thickness and exposure of the lower horizons. His classification (table 1) is different from that of Tsuya (1940) and consists of three eruptive periods: 1) Older Fuji I (80,000 to 10,000BP); 2) Older Fuji II (10,000 to 8,000BP); and 3) Younger Fuji (5,000BP to the present). Miyaji (1988, 1990) also used tephrochronology and ^{14}C dating, and based on more detailed description of the eruptive materials of Younger Fuji Volcano, subdivided the deposits into six periods: 1) 11,000 to 8,000 BP; 2) 8,000 to 4,500 BP; 3) 4,500 to 3,000 BP; 4) 3,000 to 2,000 BP; 5) 2,000 to 243 BP; and 6) 243BP (1707AD).

Older Fuji I is characterized by predominantly tephra producing eruptions, whereas Older Fuji II is characterized by lava flows (Machida, 1967). During 11,000 to 8,000 BP, the volcano erupted approximately 39.2 km^3 of pahoehoe and aa from the summit and the northwest-southeast aligned craters (Miyaji 1988,1990). Approximately 83 % of the total 48 km^3 of Younger Fuji eruptive products was erupted within this short period of 3,000 years. Fuji Volcano experienced quiescence between Old Fuji II and Younger Fuji, during which time Fuji black humic ash formed (Machida, 1967). Miyaji's (1988,1990) studies also confirmed the quiescent period since from 8,000 BP to 4,500 BP the volcano only erupted sporadically and produced only a relatively small amount of tephra and lava. Between 4,500 BP and 3,000 BP the volcano erupted approximately 0.3 km^3 of basaltic tephra and 3.0 km^3 of basaltic lava. A relatively large Plinian eruption took place during 3,000 BP to 2,000 BP, depositing approximately 2.6 km^3 of

tephra. During 2,000 BP to 234 BP, Strombolian and sub-Plinian style eruptions deposited pyroclastic falls (0.6 km^3) and lava flows (0.6 km^3). During the 1707 AD eruption, the volcano produced primarily Plinian eruptions.

Table 1. Summary of Machida's (1967) classification of Fuji Volcano eruptive products (modified after Machida, 1967).

Age	Stage	Division	Volcanic product	Development
Holocene	Younger Fuji	Younger Fuji tephra	Pyroclastic fall Pyroclastic flow Lava flow (all moderate)	Construction of the younger stratovolcano
	Quiescence	Fuji black humic soil	Pyroclastic flow (small)	Slight erosion on the flank and deposition at the foot
Pleistocene	Older Fuji II		Lava flow (large) Pyroclastic fall (small) Pyroclastic flow (small)	Construction of lava-rich volcano
	Older Fuji I	Older Fuji tepra	Pyroclastic fall (large) Pyroclastic flow (moderate) Lava flow (moderate)	Erosion and local dislocation by faulting at the foot Construction of Old Fuji, relatively lava - poor endogenous stratovolcano

The Fuji tephra classification (table 1) of Machida (1967) includes: 1) Older Fuji tephra; 2) the Fuji black humic ash; and 3) Younger Fuji tephra (Machida, 1967). The older Fuji tephra is weathered scoriaceous lapilli beds with a thickness of 100 m. It was deposited by repeated pyroclastic falls and surface weathering over a long period of time. Since there is no evidence of significant erosion, such as unconformities or buried soil bands, the explosive activity must have been relatively continuous without any significant periods of quiescence. Carbon-14 ages between 25,000 and 10,000 BP confirm that those are the eruptive materials from Older Fuji I.

The Fuji black humic ash overlies the Older Fuji tephra. This layer contains a soil profile, which was formed by weathering during the quiescent period. The thickness of the soil layer gradually increases toward the top of the volcano, while the layer gradually changes to laminated ash beds with altitude, resembling tephra fall. Carbon-14 ages of approximately 10,000 to 5,500 BP indicate that it was formed between Older Fuji II and Younger Fuji.

The Younger Fuji tephra is composed of dark brownish scoriaceous lapilli beds several meters thick. The top layer is the most recent deposit of the 1707AD eruption. The ash distribution pattern over the area is similar to that of Older Fuji tephra indicating long-term stable wind directions.

Geochemistry

Most of the eruptive materials (99%; Togashi et al., 1991) from Fuji Volcano are basaltic, primarily medium-K series and high-Al basalt (Togashi and Yamazaki, 1990). The major element oxides including SiO₂, MgO, and Al₂O₃ fall within restricted ranges of 48-52 wt%, 4-6.5 wt%, and 16-18 wt% respectively (Arculus et al., 1991). The Al₂O₃ versus SiO₂ diagram (fig. 4) illustrates the limited major element oxide compositions of Fuji Volcano. In addition, Sr and Nd isotopic compositions from Fuji Volcano fall within very limited ranges, from 0.70335-0.70355 and 0.51300-0.51305 respectively (Togashi, 1990). These values are typical of depleted island arc basalts (Togashi, 1990).

Eruptive materials from throughout the history of Fuji Volcano have been analyzed for whole rock major and trace elements (Togashi, 1990; Togashi and Yamazaki, 1990; Takahashi et al., 1991). Takahashi et al. (1991) subdivided the eruptive products from Older Fuji into four different periods (table 2): 1) >66,000BP; 2) 66,000-50,000BP; 3) 50,000-22,000BP; and 4) 22,000-10,000BP. Togashi (1990) subdivided the Older Fuji eruptive products into three different periods, slightly different from Takahashi et al. (1991): 1) 80,000-50,000BP; 2) 50,000-20,000BP; and 3) 20,000-10,000BP.

Most of the eruptive materials from Older Fuji (80,000-20,000BP) display relatively higher Al₂O₃ (18-20 wt%; Togashi, 1990; fig. 4) than the eruptive materials from Younger Fuji. Early (>66,000BP) and Late (22,000-10,000BP) Older Fuji eruptive products have similar FeO*/MgO ratios, while Middle (66,000-22,000BP) Older Fuji eruptive products display relatively lower FeO*/MgO ratios compared to the rest of the eruptive materials from Fuji Volcano (table 2). In general, Older Fuji eruptive products are depleted in incompatible elements including Rb, K, Ba, Y, Zr, and Ti (table 2; Togashi, 1990; Takahashi et al., 1991) compared to Younger Fuji eruptive products.

Fig. 4. Al_2O_3 versus SiO_2 (modified after Togashi, 1990). The eruptive materials of Fuji Volcano display a limited range in major element concentrations.

Eruptive products from Younger Fuji have been subdivided into four different periods (Takahashi et al., 1991): 1) 10,000-8,000BP; 2) 8,000-3,000BP; 3) 3,000-2,000BP; and 4) <2,000BP. Eruptive materials from Younger Fuji show relatively constant FeO^*/MgO ratios throughout (fig. 5; table 2), while other major and trace elements display slightly different trends during different eruptive periods. Overall, major and trace element variations from Younger Fuji can be divided into two compositional stages: 1) 10,000-8,000BP (Early) and <2,000BP (Late); and 2) 8,000-3,000 and 3,000-2,000BP (Middle). The Early Younger Fuji eruptive products are distinct from Middle (8,000-2,000BP) Younger Fuji eruptive products, but show overall similarity to the Late Younger Fuji eruptive products with the exception of higher TiO_2 and lower Al_2O_3 (table 2; fig. 5).

Table 2. Whole rock chemistry of Fuji Volcano (modified after Takahashi et al., 1991). Ages in years before present (BP). Major element oxides in weight %; trace element concentrations in ppm.

Older Fuji				
Age (BP)	>66,000	66,000- 50,000	50,000- 22,000	22,000- 10,000
SiO₂	52.0-54.7	49.5-54.0	49.7-53.5	49.5-52.5
Al₂O₃	19.0-19.9	17.4-18.2	17.1-19.0	15.8-19.0
K₂O	0.43-0.56	0.38-0.70	0.32-0.67	0.37-0.83
TiO₂	0.90-1.08	1.00-1.33	1.00-1.44	1.08-1.69
P₂O₅	0.12-0.17	0.12-0.25	0.10-0.21	0.13-0.33
Rb	4.85-6.67	3.18-11.06	4.39-8.33	3.79-10.15
Ba	117-148	110-186	121-214	128-221
Zr	45.8-61.6	40.5-71.5	45.8-82.6	41.6-90.5
Y	17.3-18.4	16.4-28.7	17.9-26.4	20.0-31.2
FeO*/MgO	1.95-2.65	1.45-2.08	1.52-2.12	1.72-2.68
Younger Fuji				
Age (BP)	10,000- 8,000	8,000- 3,000	3,000- 2,000	<2,000
SiO₂	49.0-51.5	49.2-51.7	49.8-52.9	50.6-52.7
Al₂O₃	15.5-17.9	18.9-21.5	17.4-19.2	16.0-18.4
K₂O	0.60-0.90	0.27-0.52	0.48-0.69	0.60-0.85
TiO₂	1.40-1.79	1.04-1.33	1.08-1.44	1.23-1.63
P₂O₅	0.25-0.36	0.14-0.24	0.18-0.27	0.23-0.33
Rb	8.26-17.24	4.83-10.69	8.28-12.31	9.66-15.86
Ba	198-279	113-179	165-202	179-255
Zr	76.3-106.3	47.4-55.6	62.2-87.4	69.6-106
Y	26.3-33.6	19.2-25.6	20.9-27.6	20.4-31.2
FeO*/MgO	1.83-2.91	1.69-2.43	1.56-2.80	1.68-2.62

Major elements including K_2O , TiO_2 , and P_2O_5 are higher (fig. 5), while trace elements including Rb, Ba, Y, and Zr are enriched (fig. 6) in Early and Late Younger compared to the Middle Younger Fuji eruptive products. Eruptive materials from the Middle Younger Fuji display higher Al_2O_3 , lower K_2O , TiO_2 , and P_2O_5 (fig. 5), and are depleted in Rb, Ba, Y, and Ba (fig. 6) compared to the Early and Late Younger Fuji eruptive materials (table 2; fig. 5); eruptive materials from 8,000-3,000BP in particular show distinct major and trace element variations.

Fig. 5. Major element oxides versus FeO^*/MgO for Younger Fuji samples (modified after Takahashi et al., 1991).

It is worth noting that the style of eruptive activities that took place during the Present-2,000 BP stage are distinct from those of other stages, with eruptions mainly on the flanks of the volcano and varying crater elevations from 3,000 to 1,000 m. FeO^*/MgO ratios (fig. 7) show an inverse correlation with crater elevation; however, incompatible elements (fig. 7) do not show such a correlation.

8,000-10,000BP ($\square$); 3,000-8,000BP ($\blacksquare$); 2,000-3,000BP ($\circ$); Present-2,000BP ($+$)

Fig. 6. Trace element concentrations versus FeO*/MgO for Younger Fuji samples (modified after Takahashi et al., 1991).

After 1,150BP ($\circ$); 2,000-1,150BP ($\bullet$)

Fig. 7. The relationship between altitude of eruptive centers and whole rock chemistry (modified after Takahashi et al., 1991).

Historic eruptions

The precise number of historic eruptions remains debatable; however, estimated numbers range from thirteen to eighteen based on geological evidence and written accounts. Thirteen eruptions documented in the geologic record include 781, 800, 826, 864, 870, 937, 999, 1033, 1083, 1511, 1560, 1700, and 1707 AD (Tsuya, 1955). Ten eruptions including 781, 800, 864, 937, 999, 1033, 1083, 1435, 1511, and 1707 AD have reliable written documentation, while six additional eruptions, including 826, 870, 952, 993, 1017, and 1427 AD, have less reliable written documentation (Koyama, 1998). Some older documents describing the eruptions before 780 AD have been found; however, there is no other conclusive evidence of those eruptions (Koyama, 1998). Despite the uncertainty in the precise number of eruptions, the fact that Fuji erupted frequently in historic times is clearly indicated.

Table 3. Summary of historical eruptions of Fuji Volcano (data compiled from Tsuya, 1955, 1968; Machida, 1981; Koyama, 1998). Black circle (●) indicates the presence of evidence.

Year (AD)	Geological record	Written documents		Remarks
		Reliable	Less reliable	
781	●	●		Tephra eruption
800	●	●		Major eruption
826	●		●	Minor eruption
864	●	●		Major eruption
870	●		●	Minor eruption
937	●	●		Flank eruption
952			●	
993			●	
999	●	●		Flank eruption
1017			●	
1033	●	●		Effusive eruption
1083	●	●		Minor eruption
1427			●	
1435		●		
1511	●	●		Minor eruption
1560	●			Minor eruption
1700	●			Minor eruption
1707	●	●		Major eruption

Written documents and geological studies agree that the 864AD and 1707AD eruptions were the most devastating eruptions in the history of Fuji Volcano. The 864AD eruption was a northwest flank eruption that occurred in June, which erupted a large quantity of augite and hypersthene phenocryst-bearing two-pyroxene-olivine-basalt (Tsuya, 1968; Machida, 1981). The lava flows were primarily blocky scoriaceous with some ropy and pahoehoe regions (Tsuya, 1968; Machida 1981). This eruption was unique with the formation of three volcanic lakes, Shoji-ko, Sai-ko, and Motosu-ko, and resulted in the largest number of written records next to the 1707 eruption (Koyama, 1998).

The 1707 eruption

The 1707 eruption is known as the Hoei Eruption among the Japanese, named after the era in which the eruption took place. It was one of the most explosive eruptions in the history of Fuji Volcano. Two major earthquakes preceded the eruption. The first earthquake, magnitude 8.2 (Aihara, 2003), occurred at Sagami Bay off the southern tip of the Boso Peninsula on December 31, 1703 (Tsuya, 1955). The second earthquake, magnitude 8.4 (Aihara, 2003), occurred in the sea off the southern tip of the Kii Peninsula on October 28, 1707 (Tsuya, 1955). The Weather Observatory (2001) summarized the detailed dates and times of volcanic phenomena that took place prior to and during the 1707 eruption (table 2). Around the beginning of December, Fuji Volcano experienced a series of seismic precursors. For the first several days, only small tremors were detected in the flank of the volcano. On December 15, much bigger tremors began, with a frequency of more than ten a day. From the evening of the 15th to the early morning of the 16th, more than thirty tremors occurred on the southern flank and more than fifty tremors occurred on the northeastern flank, destroying some houses on the eastern flank.

Around 8 o'clock in the morning of the 16th, a black plume shot up into the sky (Miyaji, 1993). The main eruption that ejected pumice and ash followed two hours later. These eruptive materials melted the snow and caused avalanches. By the evening, the ejecta became more mafic and changed to scoria. The volcano became violent and emitted large numbers of volcanic bombs. Later studies showing that people left their belongings in their houses indicates that this eruption occurred suddenly and that people could not take their belongings out of the house in time (Miyaji, 1993). On the 17th, the eruption became less violent and eventually pulsed. A cryptic dome later named Hoei-san was formed about halfway up the southeast side of the

volcano. This cryptodome consists of Older Fuji tephra; therefore, Older Fuji must have been pushed up to the surface by Hoei magma (Miyaji, 1993). On the 18th, a new crater was opened and started erupting. This phase of the eruption was sporadic and lasted until the end of the year. On the last day in 1707, the eruption once again became violent and ejected many volcanic bombs. The eruption ended with the end of the year.

Table 4. Description of the 1707 eruption (modified after the Weather Observatory, 2001).

Date	Time since the initial eruption	Events	Volcanic phenomena
Early Dec.	10 days prior	Small tremors	Small tremors occurred
15 –16 Dec.	1 day prior	Tremors	Residents felt several tremors
16 Dec, AM	few hours prior	Strong tremors	Large-scale tremors occurred
16 Dec, 10 AM-12PM	eruption	Black plume shoots up	Pumice eruption began at craters 2 and 3
16 Dec, PM	few hours later	Tephra falls in Edo	Westward wind blew the ash to Edo
16 Dec. evening	half day later	Frequent lightning, tephra changes to black	Frequent volcanic lightning occurred Ejecta changed from pumice to scoria
16 Dec. night	half day later	Houses burn	Pumice fell on nearby village and burnt houses
17 Dec.	1 day later	Tremors & lightning pulse	Eruption from craters 2 and 3 stops
17-18 Dec.	1-2 days later	-	Hoei-san was formed
18-20 Dec.	2-4 days later	Tephra falls in Edo	Sporadic eruption from crater 1 began (some were large-scale)
21-27 Dec.	5-11 days later	Tephra falls decrease in Edo	Eruption scale became smaller
27-31 Dec.	11-15 days later	Volcanic bombs	Volcanic bombs were erupted
1 Jan. 1708	16 days later	Eruption ends	Eruption ended

Three craters were formed in this series of eruptions (fig. 8). They are aligned and called the first, second and third crater from the highest to lowest elevation: 2300-3100, 2300-2500, and 2100-2300 m, respectively (Miyaji, 1993). The first crater is the biggest with a length of 1500 meters and a width of 750 meters. Since the second and third craters are eroded by the first crater, the eruption of the 16th and 17th took place in the second and third craters. The eruption on the 18th and all subsequent eruptions took place in the first crater (Miyaji, 1993).

Fig. 8. Cartoon diagram of three craters of the 1707 eruption (modified after Aihara, 2003). These craters are aligned and called the first, second, and third crater from the highest to lowest elevation.

Stratigraphy of the 1707AD deposit

Miyaji (1984, 1993) subdivided the 1707 deposit into four fall units, Ho-I, -II, -III, and – IV from the bottom to the top, based on grain size and constituents (fig. 9). The deposit is sandwiched in between soil layers, indicative of a continuous single eruptive event. No evidence of erosion is observed within the deposit.

The Ho-I unit is further subdivided into two fall units, Ho-Ia, and Ib. The Ho-Ia unit consists of mainly gray-white, porous, and coarse grained pumice, whereas the Ho-Ib unit is constituted of gray, porous, fine grained pumice, unique black-and-white banded pumice, red scoria and rock fragments. Ho-II is composed of black-gray, less porous and coarse-grained scoria, fine-grained red scoria, and rock fragments. Ho-III comprises mainly gray-black, porous, and coarse-grained scoria, red scoria, and rock fragments. Ho-IV consists of both coarse and fine-grained black scoria and rock fragments. Red scoria is not present in this fall unit.

Miyaji (1993) compiled historical documents and the distribution of tephra in order to determine the precise time of deposition of each fall unit. In Edo, the present-day Tokyo, a pumice ash-fall started around 1:00PM and changed to scoria by the evening. Therefore, silicic

Ho-I materials were deposited over 4 to 6 hours. The distribution of the Ho-II unit indicates that materials were erupted from the second and third craters, suggesting that eruptions on the 16th and 17th formed the Ho-II unit. The thickness of the Ho-III unit indicates that it was formed around the 18th to 20th when the eruption was still active. Ho-IV was formed in the last stage of sporadic small-scale eruptions during the 21st to 31st.

Fig. 9. Stratigraphy of the Fuji 1707 deposit (modified after Miyaji, 1984). The deposit has been subdivided into four fall units based on grain size and constituents (Miyaji, 1984).

The distribution of tephra

Based on Miyaji's (1993) research, the Weather Observatory (2001) constructed a table summarizing the eruptive volume, approximate date, approximate time and rate of eruption for each unit (table 3). The rate of eruption decreased rapidly with time. Within the first five days, Fuji Volcano produced more than half of the total erupted tephra (Weather Observatory, 2001). Most of the tephra were deposited in a fan-shaped lobe, eastward from the craters (Tsuya, 1968; Machida, 1981).

Table 5. Summary of the Fuji 1707 tephra eruption (modified after the Weather Observatory, 2001).

Fall unit	Amount (km ³)	App*. Date	App*. Time (h)	Rate (10 ⁻³ km ³ /h)
Ho-I Ia Ib	0.048 (0.032) (0.016)	16 Dec. 2001, 10:00-16:00	6	8.00
Ho-II	0.120	16 Dec. 2001,16:00-17 Dec. 2001, 24:00	32	3.75
Ho-III	0.176	18 Dec. 2001, 00:00- 20 Dec. 2001, 24:00	72	2.44
Ho-IV	0.332	21 Dec. 2001, 00:00-27 Dec. 2001, 24:00	192	1.73
Total	0.676	16 Dec. 2001,10:00- 27 Dec. 2001, 24:00	302	2.24

*Approximate

Major and trace element variations

Nakamura et al. (1986) divided the 1707 deposit into 17 fall units based on visual differences (fig. 10) and analyzed whole rock major and trace element concentrations (table 6 and 7). Their samples were collected from Midono, Shizuoka, 14 km east of the craters at an outcrop that no longer exists. The thickness of the deposit is 270 centimeters. In comparison with Miyaji's fall units, the fall units 1, 2-5, 6-14, and 15-17 correspond to Ho-I, Ho-II, Ho-III, and Ho-IV respectively.

The fall units 1 and 2 clearly show a change in SiO₂, from dacite to andesite (fig. 11). The other oxide concentrations also show a shift from felsic to intermediate with an increase in Fe₂O₃, CaO, MgO, MnO, P₂O₅, and TiO₂ and a decrease in Na₂O and K₂O. From fall unit 5 to 6, there was also a rapid change in chemistry, shifting to more mafic in composition. Although the

rapid change from 1 to 2 is visible in hand samples, the rapid change from 5 to 6 is only observable with whole rock chemistry. However, despite the visible differences in grain size and color in fall units 6 to 17, they are similar in whole rock chemistry.

Fig. 10. Stratigraphy of the Fuji 1707 deposit (modified after Nakamura et al., 1986). The deposit has been subdivided into 17 fall units by Nakamura et al. (1986). The correlation between Miyaji's (1984) classification is shown on the left.

The trace element data show some unexpected trends (fig. 12) given the major element variations. Sr, Pb, and Zn stay relatively constant throughout the fall units, whereas Li and Rb decrease by a factor of 2 from bottom to top. Co, Ni and Cu increase by a factor of 2 to 5 from bottom to top. Considering the significant change in SiO₂ from 64 (wt%) to 49 (wt%), it is surprising that no variations in Sr, Pb, and Zn are observed.

Summary

Fuji Volcano has been active throughout its history, and has produced a large volume of primarily basaltic deposits. Although several different detailed classification schemes have been developed for the stratigraphy of the Fuji deposits, three major eruptive cycles are widely recognized. These eruptive cycles are associated with three stratovolcanoes including the present-day Fuji Volcano (Younger Fuji) and its predecessors Older Fuji and Komitake.

Two eruptions in Fuji's recent history (past 3,000 years) were characterized by unusually silicic magmas, including andesite and dacite. The more recent of these two eruptions (1707AD), is the most recent eruption of Fuji Volcano. The stratigraphy of the resulting deposit has been thoroughly studied, and, combined with historic accounts, detailed reconstructions of this unusually explosive Fuji eruption have been developed. However, only preliminary petrographic and geochemical data are available for this deposit, and the processes responsible for formation of the evolved magmas are thus not well constrained. The purpose of the current study, the results of which are discussed in Chapter 3, is to develop a better understanding of the processes that resulted in the formation of the silicic magmas erupted in the explosive 1707AD eruption of Fuji Volcano.

Table 6. Major element concentrations (wt%) of the Fuji 1707 deposit (modified after Nakamura et al., 1986).

Fall Unit	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Total
1	64.09	0.92	16.51	6.72	0.14	1.96	5.15	3.67	2.04	0.25	101.45
2	58.00	1.20	16.95	9.24	0.18	3.11	6.89	3.46	1.37	0.33	100.73
3	56.88	1.26	17.11	9.96	0.19	3.50	7.34	3.33	1.24	0.33	101.14
4	55.87	1.29	16.91	10.27	0.19	3.73	7.48	3.29	1.19	0.33	100.55
5	55.35	1.32	16.92	10.63	0.20	3.89	7.78	3.22	1.13	0.32	100.76
6	52.23	1.34	16.93	11.96	0.20	5.04	9.31	2.64	0.79	0.26	100.70
7	51.93	1.35	16.95	12.01	0.20	5.19	9.42	2.59	0.76	0.26	100.66
8	51.24	1.36	17.08	12.14	0.20	5.16	9.48	2.50	0.72	0.26	100.14
9	51.10	1.36	17.07	12.25	0.20	5.15	9.51	2.51	0.71	0.26	100.12
10	51.66	1.35	17.06	11.98	0.19	5.19	9.64	2.59	0.74	0.26	100.66
11	51.51	1.35	17.00	12.01	0.19	5.32	9.65	2.59	0.73	0.26	100.61
12	51.64	1.36	17.06	12.09	0.19	5.14	9.58	2.59	0.74	0.26	100.65
13	51.46	1.38	17.03	12.36	0.19	5.16	9.52	2.57	0.74	0.26	100.67
14	51.21	1.37	16.87	12.17	0.19	5.04	9.55	2.56	0.74	0.26	99.96
15	51.19	1.37	16.85	12.03	0.19	5.14	9.51	2.54	0.73	0.26	99.81
16	51.24	1.39	16.96	12.16	0.19	5.31	9.52	2.60	0.73	0.27	100.37
17	49.84	1.38	16.60	12.19	0.19	5.15	9.46	2.58	0.73	0.26	98.38

Table 7. Trace element concentrations (ppm) of the Fuji 1707 deposit (modified after Nakamura et al., 1986).

Fall Unit	Li	Co	Ni	Cu	Zn	Rb	Sr	Pb
1	15	22	6	69	97	34	315	13
2	12	27	8	73	113	25	353	13
3	11	30	9	76	113	22	344	12
4	11	30	9	82	116	23	347	13
5	10	31	18	98	117	19	349	13
6	7	40	25	150	114	13	334	12
7	7	39	27	156	114	13	328	12
8	6	40	28	164	115	12	330	12
9	6	39	26	153	115	12	331	11
10	6	39	27	154	114	12	330	11
11	6	39	38	151	113	10	323	11
12	6	39	27	157	114	11	330	11
13	7	40	29	160	115	12	311	14
14	7	36	22	165	115	12	325	11
15	6	41	27	166	116	12	319	13
16	7	40	24	164	117	12	321	12
17	7	39	26	160	115	11	338	12

Fig. 11. Major element variations (wt%) of the Fuji 1707 deposit (modified after Nakamura et al., 1986). Numbers shown are the fall units (fig. 10; table 6).

Fig. 12. Trace element variations (ppm) of the Fuji 1707 deposit (modified after Nakamura et al., 1986). Numbers shown are the fall units (fig. 10; table 7).

Chapter 2

Introduction to chemically zoned magma chambers

Introduction

Most deposits from large, explosive volcanic eruptions are chemically zoned, representing chemical gradients in the magma chamber feeding the volcano. As magma evolves, it becomes more silicic and volatile-rich towards the top of the chamber, becoming potentially explosive. Chemically zoned volcanic deposits typically display a decrease in SiO_2 and increase in phenocryst contents from the bottom to the top. The processes generating such complex chambers are still being debated due to the complexity of interrelated variables and the wide ranges in volumes, compositions, and life spans of magmatic systems (Wörner and Schmincke, 1984). Based on previous studies, commonly proposed petrogenetic processes include crystal fractionation, magma mixing/replenishment, and crustal assimilation.

Closed system magma chamber evolution

Closed system crystal fractionation is recognized as one of the important processes that gradually forms chemical gradients within magma chambers (Michael, 1983). With a drop of magma temperature, crystals are formed and removed, changing the chemical composition of the residual liquid. Several parameters including starting bulk composition, temperature, and pressure control the process. Fractional crystallization can explain the systematic change in major and trace element concentrations in many chemically zoned eruptions (Wörner and Schmincke, 1984).

Expected geochemical features of magmas that have experienced closed system fractional crystallization include: 1) kinks in trends on element-element diagrams; 2) enrichment in rare earth elements (REE) and other incompatible trace elements with increasing SiO_2 ; and 3) constant isotopic compositions. Due to the fact that fractionating minerals typically change with time as a magma evolves, element-element diagrams display kinks in the trends. REEs show relative enrichment from least evolved to most evolved rocks due to their incompatibility in the major mineral phases. Isotopic compositions stay constant throughout the closed system crystal fractionation process unless magma residence times are sufficiently long that radioactive decay

modifies the isotopic compositions.

Open system magma chamber evolution

Potential open system magma chamber processes include magma mixing/replenishment (Wörner and Wright, 1984), crustal assimilation (Duffield et al., 1995), and volatile transfer (Bailey, 1970), processes invoked to explain isotopic heterogeneity in the magma chambers. Isotopic signatures can provide important information regarding the sources and petrogenesis of magmas. Due to the fact that magmas experience disequilibrium conditions during open system processes, petrographic features such as reverse zoning, resorbed and sieved textures, and disequilibrium mineral assemblages may also record the open system processes. Simple binary mixing results in linear trends between two end members on element-element diagrams.

Due to their high temperatures, magmas can melt and incorporate country rock. Since crust usually has radiogenic isotopic signatures, magmas that have assimilated crust can display radiogenic isotopic signatures compared to uncontaminated magma. In some cases, crustal assimilation cannot be detected by common isotope systems such as Sr, Nd, and Pb, used to evaluate igneous processes. Since there is sometimes not a large difference in isotopic compositions between the magma and the potential contaminants (crust or sediment) in those isotope systems, trace amounts of assimilation can be accommodated with no change in Sr, Nd, or Pb isotope signatures. In such cases, Os isotopes can serve as a very sensitive tracer for crustal assimilation, due to the extreme difference in $^{187}\text{Os}/^{188}\text{Os}$ ratios between the mantle and crust.

Case Studies

Several well-studied chemically zoned volcanic deposits, including the Bishop Tuff in eastern California, the Laacher See in Eifel Germany, and the Bandelier Tuff in the Jemez Volcanic Field, New Mexico, provide insight into processes operating in the evolution of large silicic magmatic systems. Although the precise petrogenetic processes operating in these magmatic systems have long been debated, comprehensive evaluation of petrography, major and trace element concentrations, REE abundances, and isotopic compositions have led to some constraints on the petrogenetic processes.

Bishop Tuff

The Bishop Tuff, a high-silica rhyolitic ash flow tuff is probably the most thoroughly studied chemically zoned volcanic deposit in the world. It was deposited by a catastrophic eruption at 740,000 BC, from Glass Mountain in eastern California.

The Bishop Tuff is a typical zoned deposit, representing the inverse of chemical gradients in the magma chamber. Concentrations of Sr, Ba, Eu, Mg, Fe and light REE increase while Nb, Cs, Rb, Sb, Sn, W, and heavy REE decrease from the bottom to the top of the deposit (Michael, 1983). Mineral phases observed are Ti-Fe oxide, quartz, sanidine, oligoclase, biotite, ilmenite, titanomagnetite, zircon, apatite, and allanite (Hildreth, 1979).

Hildreth (1979) originally proposed that thermogravitational diffusion, a liquid-state fractionation process, generated the highly fractionated Bishop Tuff because it was difficult to explain the trace element variations. However, later studies (Michael, 1983; Cameron, 1984; Knesel and Davidson, 1997) revealed that crystal fractionation can in fact explain the observed elemental concentrations when fractionation of trace mineral phases was accounted for. Fractional crystallization has become the widely accepted process to explain the generation of the chemically zoned Bishop Tuff.

However, the explanation for the isotopic heterogeneity in the Bishop Tuff is still being debated in terms of in-situ radioactive decay versus crustal assimilation. Christensen and Halliday (1996) proposed that the rhyolitic magma experienced rapid differentiation in a long-lived magma chamber based on apparent Rb-Sr isochrons and Nd isotopic compositions of melt inclusions. However, Duffield et al. (1995) argued that contamination by roof-rocks at the top of the magma chamber was responsible for the chemical as well as isotopic gradients in the Bishop Tuff. A recent study by Ried and Coath (2000) found evidence against a long-lived (>100 ka) magma chamber based on U-Pb ages of zircons, providing additional constraints on the role of open system processes in the Bishop Tuff.

Laacher See

The Laacher See Volcano is located in the East Eifel, West Germany. An eruption in 11,000 BC resulted in a zoned deposit ranging from highly evolved phonolite to mafic phonolite. Three teptra units are recognized including the LLST, MLST, and ULST (Lower, Middle, Upper Laacher See Tephra; Wörner and Schmincke, 1984). Major and trace element concentrations

measured by Wörner and Schmincke (1984) and Wörner et al. (1983) show variations throughout the deposit, displaying systematic correlations with stratigraphic position. LLST are alkali-rich and enriched in incompatible trace elements. Sr concentrations decrease and Nd concentrations increase rapidly toward the top of LLST. The boundary between LLST and MLST shows relatively steeper chemical gradients, shifting from differentiated phonolite (LLST) to mafic phonolite (MLST). Major and trace element concentrations of MLST are intermediate between LLST and ULST. ULST show the highest MgO and Sr and lowest Na₂O, Cl, Nd and Th concentrations. The phenocryst phases observed were feldspars, hauyne, amphibole, clinopyroxene, sphene, apatite, mica, olivine, nepheline, zircon, and cancrinite (Wörner and Schmincke, 1984). The chemical variations can be explained by crystal fractionation of the observed mineral phases.

The role of open system processes is evidenced by Sr and Nd isotopic compositions (Wörner et al., 1985). Sr isotopes show significant variations with ⁸⁷Sr/⁸⁶Sr ranging from 0.70477-0.70486 (ULST) to 0.71122 (LLST), while Nd isotopes display a relatively narrow range in ¹⁴³Nd/¹⁴⁴Nd, from 0.51261 (ULST) to 0.51268 (LLST). These isotopic gradients are attributed to variable degrees of assimilation of crustal materials including Devonian sedimentary rocks or metamorphic basement (Wörner et al., 1985). Abundant inclusions of mid-crustal rocks are found in East Eifel basanite scoria cones, providing additional evidence for crustal assimilation.

In order to explain petrographic observations and chemical and isotopic signatures of the final pyroclastic products erupted from Laacher See Volcano, Wörner and Wright (1984) proposed that magma mixing played an important role. Disequilibrium textures such as resorption of phonolitic phenocrysts as well as reverse zoning were observed. Linear relationships between two end members on element-element diagrams as well as variations in Sr isotopic compositions could be attributed to magma mixing. However, Wörner et al. (1985) argued that the above observations are better explained by crustal assimilation, because the Sr isotopic signatures are very radiogenic, indicative of crustal assimilation. Summarizing all studies, Wörner et al. (1985) proposed that the chemically zoned magma chamber underneath the Laacher See Volcano evolved through crystal fractionation with variable degrees of crustal assimilation.

Bandelier Tuff

The Bandelier Tuff is a zoned deposit from Valles Caldera, in the Jemez Volcanic Field, New Mexico. The Bandelier Tuff consists of two rhyolitic ignimbrite members: the Otowi Member (lower) and Tshirege Member (upper). The Otowi Member and Tshirege Member were erupted at 1.45 Ma and 1.12 Ma respectively. The Otowi Member and Tshirege Member both consist of ignimbrite underlain by pumice fall (Wolff et al., 1999).

Major element concentrations in the Otowi Member display very little variation throughout the deposit, while trace element concentrations show significant variations; incompatible trace elements show three- to five-fold variations (Kuentz, 1986; Spell et al., 1990; Wolff et al., 1999). Incompatible elements display linear correlations between early and late erupted pumice clasts in most element-element diagrams (Wolff et al., 1999). Elemental variations can be explained by crystal fractionation of the observed phenocryst assemblage including sanidine + quartz + clinopyroxene + magnetite + zircon + allanite $\pm$ fayalite $\pm$ chevkinite $\pm$ britholite (Wolff and Ramos, 2003).

The role of open system processes has been vigorously debated for years. In order to evaluate the cause of isotopic heterogeneity as well as residence time of the magma chamber, Sr isotopes (Wolff et al., 1999), and Pb isotopes (Wolff and Ramos, 2003) were utilized.

Although both in-situ radioactive decay and assimilation can modify the Sr isotopic compositions of magma, these processes display different trends as illustrated in the $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ diagram (fig. 1; Wolff et al., 1999). Sr isotope compositions of feldspars, glasses, and quartz phenocrysts with glass inclusions are consistent with contamination by Proterozoic basement with higher Sr contents and radiogenic Sr isotopic compositions (Wolff et al., 1999). Quartz with glass inclusions display more radiogenic Sr signatures compared to other phases. These quartz grains were interpreted to have formed at the edge of the magma chamber, and incorporated melts of the amphibolite country rock (Wolff et al., 1999).

Pb isotope variations provide additional insights into the ongoing debate on in-situ radioactive decay versus assimilation to explain the isotopic heterogeneity. A positive correlation between $^{87}\text{Sr}/^{86}\text{Sr}_i$ and $^{87}\text{Rb}/^{86}\text{Sr}$ displayed by feldspars from the Bandelier can be explained by either a long-lived magmatic system or crustal assimilation (Wolff and Ramos, 2003); however, variable Pb isotope signatures in these feldspars cannot be explained by radioactive decay at the U/Pb ratios of the host magma (fig. 2; Wolff and Ramos, 2003).

Therefore, in-situ radioactive decay in a closed-system magma chamber can be eliminated as a possible explanation of the isotopic heterogeneity in the Bandelier Tuff.

Fig. 1. Schematic variations of $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ for three different processes that form distinct trends (modified after Wolff et al., 1999).

Fig. 2. Pb isotope ratios of single sanidines for Guaje unit A, Upper Guaje, and ignimbrite, and glomerocrysts from the Bandelier Tuff (modified after Wolff and Ramos, 2003). The Pb isotope variations could not be produced by in situ radioactive decay over reasonable time periods given the low U/Pb ratios of the host magma (Wolff and Ramos, 2003). Pb isotope ratios provide the most conclusive evidence for open system processes, arguing against a long-lived magma chamber.

Fuji Volcano

Although the eruptive history of Fuji Volcano has been dominated by basaltic volcanism, explosive silicic eruptions have occurred twice in the history of Fuji Volcano, in 2800BC and in 1707AD. These silicic eruptions are rare at Fuji Volcano and have been relatively little studied; hence, the petrogenetic processes that formed the chemical gradients in the magma chamber are poorly understood. Kawamoto (1990, 1991a, 1991b, 1992) proposed that liquid immiscibility played a major role in the formation of the 1707 chemically zoned magma chamber. However, his publications are limited to four abstracts; therefore, the detailed results of his studies are unavailable.

Fig. 3. TiO_2 and P_2O_5 versus SiO_2 for the Fuji 1707 samples (modified after Kawamoto, 1990). Both TiO_2 and P_2O_5 decrease with increase in SiO_2 from andesite to dacite.

Kawamoto (1990) analyzed twenty-seven pumice and scoria samples for petrography and whole rock major and trace element concentrations. Major element concentrations show substantial variations in SiO_2 ranging from ~51 to ~68 wt% (fig. 3). Most of his samples are andesite and dacite, and a few are basalt. A compositional gap exists between basalt and dacite, from ~51 to ~56 wt%. Both TiO_2 and P_2O_5 decrease with increasing SiO_2 from andesite to dacite. Based on the fact that the andesites do not fall on a mixing line between basalt and dacite, Kawamoto (1990) argued that magma mixing could be eliminated as a potential process for producing the chemical zonation.

The 1707 eruptive materials were characterized into two types, basalt and andesite-dacite, based on petrography. The predominant phenocrysts in the basalts are olivine (2 vol%) and feldspar (4 vol%), while the andesites and dacites are aphyric with less than 1 vol% feldspar, orthopyroxene, and clinopyroxene. Chemical compositions of pyroxene were reported to be identical in andesite and dacite, evidence used against the magmas being related by fractional crystallization or magma mixing (Kawamoto, 1990).

Chemical data, including a decrease in TiO_2 and P_2O_5 with increasing SiO_2 also were used to argue against crystal fractionation, because Kawamoto (1990) was unable to explain the chemical trends by fractionation of the observed mineral assemblage. Because neither hypersthene nor augite incorporate significant TiO_2 and P_2O_5 into their structures, it was inferred that these oxides should have increased with increasing SiO_2 , if crystal fractionation of the observed mineral assemblage occurred (Kawamoto, 1990).

Fig. 4. Proposed liquid immiscibility model for the Fuji 1707 deposit (modified after Kawamoto, 1990, 1991a). According to Kawamoto's model, the 1707 magma chamber evolved through four main stages: 1) the presence of a homogeneous magma chamber; 2) the occurrence of liquid immiscibility; 3) the formation of a chemically zoned magma chamber; and 4) intrusion of basaltic magma.

According to Kawamoto (1990, 1991a, 1991b, 1992), liquid immiscibility can best explain the apparent limited trace element variations, limited temperature variations, identical phenocryst compositions, and identical Sr isotopic compositions. Based on the above observations, Kawamoto proposed that the 1707 magma chamber evolved through four main stages (fig. 4): 1) the presence of a homogeneous magma chamber; 2) separation into two liquids (andesite and dacite) by liquid immiscibility with a drop of the temperature; 3) density separation of andesitic and dacitic magmas to form a chemically zoned magma chamber; and 4) intrusion of

basaltic magma into a pre-existing magma chamber. His model supposedly can explain the common features in calc-alkaline rocks (Kawamoto, 1991a, 1991b, 1992) including: 1) constant FeO/MgO ratios during differentiation; 2) mafic inclusions in felsic rocks; and 3) compositional gaps between andesite and dacite/rhyolite.

Liquid immiscibility

Silicate liquid immiscibility is a process of magmatic differentiation, in which magma separates into two compositionally distinct immiscible silicate melts. Although liquid immiscibility was hypothesized by early experimental petrologists, (Greig, 1927; Bowen, 1928), it fell into disrepute until 1970 when the indisputable evidence of silicate liquid immiscibility was found in Apollo 11 lunar mare basalts (Roedder and Weiblen, 1970). Soon after the finding of the striking evidence in lunar mare basalts, Roedder and Weiblen (1971) and De (1974) found evidence for silicate liquid immiscibility in terrestrial basalts. Philpotts (1979) and Philpotts and Doyle (1983) observed evidence for silicate liquid immiscibility in Connecticut flood basalts and provided additional insights of silicate liquid immiscibility experimentally. Later, Fujii et al. (1980) observed a number of volcanic rocks from widely distributed localities in the Japanese Islands and also found evidence for silicate liquid immiscibility.

Silicate liquid immiscibility is common in high-alumina basalts and tholeiitic basalts as well as in andesites (Fujii et al., 1980). Although less common, evidence for silicate liquid immiscibility has also been found in some alkalic rocks (Fujii et al., 1980). The presence of dark-colored spherical globules within light-colored glass matrix (fig. 5) under a microscope provides petrographic evidence of liquid immiscibility. Some dark-colored globules are not spherical (Fujii et al., 1980) but rather elliptical, irregular blob or dumbbell shaped (De, 1974). De (1974) interpreted the non-spherical globules as spherical globules that were deformed while attaching to plagioclase crystals. In rare cases, light-colored glass globules are observed in dark-colored glass (Fujii et al., 1980). The average diameter of dark-colored (high-iron) globules is 0.004 mm, typically ranging from 0.002 to 0.010 mm (De, 1974). Although silicate liquid immiscibility occurs during late stage crystallization (a few percent of remaining liquid; Philpotts, 1979), in rare cases, occurrences of liquid immiscibility in early stages of crystallization have been reported (Roedder and Weiblen, 1971; De, 1974).

Fig. 5. Petrographic evidence for liquid immiscibility in an olivine basalt from the 864AD deposit of Fuji Volcano (modified after Fujii et al., 1980). These dark-colored globules in light-colored glass matrix are interpreted as evidence for liquid immiscibility.

De (1974) reported occurrences of silicate liquid immiscibility in basalts from the Deccan Traps, India. Through his detailed observations, he found evidence for silicate liquid immiscibility in approximately 50 percent of the total of 150 specimens. These specimens represent the Lower, Middle, and Upper stratigraphic subdivisions of the Deccan Traps and come from western India including Kutch, Bombay, and Daulatabad and Central India including Jabalpur, Nagpur, and Chindwara. These basalts are tholeiitic in composition and contain phenocrysts of olivine and plagioclase in a groundmass of plagioclase and augite.

Petrographic evidence showed the presence of dark-colored globules of glass in lighter colored glass. In addition, Cu-Fe sulfide globules were found in some specimens. High-silica glass globules in a base of darker colored high-iron glass such as those found in the lunar basalts were absent in the samples from Deccan Traps. The presence of Cu-Fe sulfides is interpreted as coexistence of both silicate/sulfide liquid immiscibility and high-iron/high-silica liquid immiscibility (De, 1974). While the high-silica and high-iron glasses of the lunar mare basalts have compositions equivalent to potassic granite and pyroxenite respectively, the residual glasses in the Deccan Trap basalts commonly have compositions corresponding to a mixture of augite and magnetite.

Silicate liquid immiscibility in a tholeiitic basalt flow in Connecticut was also reported (Philpotts, 1979; Philpotts and Doyle, 1983). In their studies, petrographic and textural characteristics of basalt were interpreted as evidence for silicate liquid immiscibility. Investigations were made on three well-exposed basaltic units ranging from highly amygdaloidal and altered, massive and fresh, and pillowed and slightly altered, from the lowest to the uppermost respectively. Evidence of silicate liquid immiscibility is present in the middle basalt. The presence of liquid immiscibility was difficult to observe under regular transmitted light, but it became apparent under reflected light. Both dark spheres and light spheres were found in respective glasses. The whole-rock chemistry of the host rocks shows that the basalts are common quartz normative tholeiites with a high ferrous/ferric iron ratio. The only unusual feature is the low potassium concentration with only half of typical theoleiitic potassium concentration.

In addition to the textural evidence, several experiments were conducted in order to constrain the factors that control liquid immiscibility (Philpotts, 1979; Philpotts and Doyle, 1983). These experiments documented that oxygen fugacity as well as oxidation state of the magma play a critical role in determining whether or not the liquid encounters the liquid immiscibility field.

Widespread occurrence of silicate liquid immiscibility has also been reported in Japanese volcanic rocks (Fujii et al., 1980), including samples from Fuji Volcano, Hakone Volcano, and Oshima Volcano. Olivine basalt and high-alumina basalt lava flows of Fuji Volcano display the most conspicuous evidence for liquid immiscibility, with dark-colored globules up to 0.010 mm in diameter. The composition of the dark globules was determined by electron probe microanalysis, which revealed that they have higher concentrations of P_2O_5 compared to those of lunar mare basalts. It is speculated that the concentration of Fe_2O_3 is also higher, but Fe concentrations could not be determined by microprobe analysis (Fujii et al., 1980).

Since the discovery of the indisputable evidence for silicate liquid immiscibility in lunar mare basalts, silicate liquid immiscibility has been widely recognized as a process of magmatic differentiation. The occurrence of two immiscible silicate melts in high-alumina basalts and tholeiitic basalts may be expected at low pressure during the late stages of crystallization. However, there are no examples of magma chamber scale silicate liquid immiscibility in Japanese or other volcanoes, arguing against silicate liquid immiscibility as a significant process

in magma chamber-scale magmatic differentiation.

Objectives of this study

This study involves a comprehensive evaluation of the 1707 magma chamber of Fuji Volcano, Japan. Although Kawamoto (1990, 1991a, 1991b, 1992) proposed that liquid immiscibility played a major role in the formation of the 1707 chemically zoned magma chamber, liquid immiscibility has not previously been identified as a magma chamber-scale differentiation process in silicic systems. Because his work appears only in four abstract publications, the detailed results of his study are unavailable for further assessment. Therefore, in order to better constrain the petrogenetic processes that occurred prior to the 1707 eruption, petrographic and microprobe data, major and trace element data, and Sr, Nd, Pb, and Os isotopic data were obtained. Table 1 summarizes the three most common petrogenetic processes that have been identified in the formation of chemically zoned magma chambers (Cameron, 1984; Wörner and Schmincke, 1984; Duffield et al., 1995; Christensen and Halliday, 1996; Wolff and Ramos, 2003) and the petrogenetic tracers that can be used to identify these processes.

Table 1. Summary of petrogenetic processes and indicators.

Processes	Petrography & Microprobe	Major and trace elements systematics	Isotope (Sr, Nd, Pb and Os) systematics
Closed system crystal fractionation		*Variations in mineral compositions with change in magma compositions *Systematic variations with kinks in element-element plots *Constant ratios of highly incompatible elements	*Constant
Magma mixing	*Mingling, *Reverse zoning *Disequilibrium textures	*Linear correlations on all element-element plots	*Hyperbolic mixing trends on isotope-isotope plots
Crustal assimilation	*Disequilibrium textures	*Modified ratios of highly incompatible elements	*More radiogenic signatures in evolved magmas

Chapter 3

The Evolution of a Chemically Zoned Magma Chamber: the 1707 eruption of Fuji Volcano, Japan

Abstract

Detailed petrographic, major and trace element and Sr, Nd, Pb, and Os isotope studies were done in order to constrain the petrogenetic processes that formed a chemically zoned magma chamber prior to the 1707 eruption of Fuji Volcano. The eruptive history of Fuji Volcano has been dominated by basaltic volcanism. However, the 1707 eruption of Fuji Volcano resulted in a chemically zoned volcanic deposit ranging from basalt to dacite. The processes responsible for generating these siliceous magmas beneath Fuji Volcano are not well understood although it has been proposed that liquid immiscibility played a major role (Kawamoto, 1992). Other possible explanations for a chemically zoned volcanic deposit include magma mixing, crystal fractionation, and crustal assimilation. Our observed petrography, chemical variations and isotopic compositions can be explained by crystal fractionation accompanied by minor crustal assimilation.

Introduction

Most deposits from large, explosive volcanic eruptions are chemically zoned, representing chemical gradients in the magma chamber feeding the volcano. Magma generally becomes more silicic and volatile-rich towards the top of the chamber as it evolves, becoming potentially explosive. However, the processes generating such complex magma chambers are still being debated. Although closed system crystal fractionation can often produce observed chemical gradients (Michael, 1983; Wörner and Schmincke, 1984; Widom et al., 1992), several recent detailed geochemical and isotopic studies have revealed the importance of open system processes such as magma mixing and crustal assimilation in some cases (Wörner and Wright, 1984; Duffield et al., 1995; Wolff and Ramos, 2003).

The most recent eruption of Fuji Volcano, in 1707AD, resulted in a chemically zoned volcanic deposit ranging from basalt to dacite. Such silicic eruptions have only occurred twice in the history of Fuji Volcano, in 2800BP and in 1707AD; most eruptions throughout Fuji's history have been dominated by basalt. Despite the unusual nature of the 1707 eruption, relatively little work has been done to unravel the processes that led to the silicic, chemically zoned deposit. Kawamoto (1992) proposed that liquid immiscibility was an important process in generating a chemically zoned magma chamber prior to the 1707 eruption. Kawamoto's model was based on four main observations of the 1707 deposit: 1) limited trace element abundance variations; 2) limited temperature variations; 3) identical phenocryst compositions; and 4) identical Sr isotopic compositions throughout the deposit. However, Kawamoto's work has been published only in abstract form, hence detailed assessment of the liquid immiscibility model is precluded. Although liquid immiscibility may be common in some mafic magmas such as high-alumina and tholeiitic basaltic magmas, it is only known to occur on a small scale (Fujii et al., 1980). Liquid immiscibility has not previously been recognized as a process that could produce magma chamber-scale chemical gradients or silicic magmas such as andesite and dacite.

The purpose of this study was to carefully evaluate the role of potential petrogenetic processes including crystal fractionation, magma mixing, crustal assimilation, and liquid immiscibility, in the development of the chemically zoned 1707 deposit of Fuji Volcano. Petrographic, major and trace element, and Sr, Nd, Pb, and Os isotope data have been combined to develop a detailed petrogenetic model of the evolution of the Fuji 1707 magmatic system.

Geological setting/ background

Fuji Volcano is located in south-central Honshu, Japan, lying over the westward subducting Pacific plate, and possibly influenced by the Phillipine Sea plate (fig.1). The main edifice of Fuji Volcano consists of three volcanoes including Komitake, Older Fuji, and Younger Fuji (oldest to youngest, respectively). The oldest known deposits from Fuji formed approximately 80,000 years ago (Takahashi et al., 1991). Fuji volcano has been very active throughout its history, and erupted predominantly medium K series and high alumina basalts with SiO₂ ranging from 48 to 52 wt% (Togashi, 1990). Sr and Nd isotope signatures in Fuji eruptive products display a relatively narrow range with ⁸⁷Sr/⁸⁶Sr varying from 0.70335 to 0.70355 and ¹⁴³Nd/¹⁴⁴Nd varying from 0.51300 to 0.51305 (Togashi, 1990). These values fall within the range of typical depleted island arc basalts (Togashi, 1990).

The most recent eruption in 1707AD was one of the most explosive eruptions in the history of Fuji Volcano and resulted in a chemically zoned volcanic deposit ranging from basalt to dacite. The Plinian eruption associated with the large-scale pumice, scoria, and volcanic tephra lasted for a little over two weeks, forming three craters on the SW flank of the volcano. These craters are called first, second, and third craters from the top to the bottom. Silicic eruptions took place at the second and third craters, and later basaltic eruptions took place at the first crater. The total eruptive volume was 0.7km³ (Weather Observatory, 2001).

Analytical techniques

A comprehensive suite of samples for this study was collected from two localities. The 1707 deposit has previously been subdivided into four fall units, Ho I to Ho IV from the bottom to the top, based on differences in grain size and constituents (Miyaji, 1984). The first sampling locality, 4.5 km SE of the craters (35°19'883"N, 138°47'831"E) exhibits an incomplete eruption record, with an absence of Ho III and Ho IV, but greater thicknesses of Ho I (22 cm) and Ho II (100cm). The second sampling site, 4 km NE of the craters (35°21'703"N, 138°47'035"E), displays a complete depositional section of the 1707 eruption (fig. 2). The thicknesses of the representative units at this site are 18, 19, 115, 125 cm.

Major and trace element concentrations of whole rocks were obtained by DCP-AES (Direct-Current Plasma Atomic Emission Spectrometry) at Miami University. For the rare earth element (REE) measurements, bulk REE separations were done by ion exchange with AG50W-

X8 cation exchange resin, followed by cleanup columns with AG50W-X12 resin. Sample preparation and measurement methods are described in Katoh et al. (1999). Mineral chemistry was obtained by electron microprobe analysis (Cameca SX 50) at Indiana University, Bloomington, IN.

Chemical separations and Sr, Nd, Pb, and Os isotopic measurements were done at Miami University. For Sr, Nd, and Pb analyses, approximately 0.1 g of whole rock powder was dissolved in concentrated HF-HNO₃ and taken up in HBr. Procedures for the separation are described in Walker et al. (1989) with the exception of Sm-Nd separations. EiChrom Ln-Spec resin was used for Sm-Nd separations following methods similar to Pin and Zalduegui (1997).

Sample digestion for Re-Os chemistry followed the Carius tube method described in Shirey and Walker (1995). Separation of Re and Os were done using an aqua regia-CCl₄ solvent extraction technique (Cohen and Waters, 1996). The Os was back-extracted into HBr and purified using a microdistillation procedure (Roy-Barman, 1993; Roy-Barman and Allegre, 1995).

All isotopic compositions were measured by thermal ionization mass spectrometry using a Finnigan Triton. Strontium isotopic ratios were corrected for fractionation using $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. Eighty measurements of NBS 987 gave an average of $^{87}\text{Sr}/^{86}\text{Sr} = 0.71024 \pm 1$ (2σ). Neodymium isotopic ratios were corrected for fractionation using $^{144}\text{Nd}/^{146}\text{Nd} = 0.7219$. Fifty measurements of La Jolla gave an average of $^{143}\text{Nd}/^{144}\text{Nd} = 0.51185 \pm 1$ (2σ). $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ were corrected for fractionation by 0.1%, 0.1%, and 0.125% per amu respectively. Errors on $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ measurements were ± 0.006 , ± 0.008 , and ± 0.026 respectively. Internal measurement errors for osmium isotopic ratios ranged from 0.5-2%. Blank correction errors ranged from 0.6-10% based on blanks of 1.0 ± 0.1 pg and $^{187}\text{Os}/^{188}\text{Os} = 0.174$. Replicate analyses (JF-15 and JF-15R) agreed within 3.6% for $^{187}\text{Os}/^{188}\text{Os}$.

Results

Chemical compositions of thirty-seven juvenile and twelve crustal xenolith samples were analyzed using DCP-AES (table 1). JF-20 and JF-21 are heterogeneous with mixture of pumices and scorias. In order to get meaningful whole rock chemistry, JF-20 was separated into dark (JF-20D) and light (JF-20L) fractions, while JF-21 was separated into dark (JF-21D), light (JF-21L), very light (JF-21V), and banded (JF-21B) fractions. All juvenile samples are medium-K

subalkaline basalt, andesite, and dacite respectively (fig. 3).

Major elements

The juvenile samples from the 1707 deposit shows substantial major element variations. SiO_2 ranges from 50 to 67 wt% (basalt to dacite); however, there is a compositional gap between 52 to 57 wt%. TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , MgO , CaO , and P_2O_5 decrease with increasing SiO_2 , while Na_2O and K_2O increase with increasing SiO_2 . Concentrations of Al_2O_3 , Fe_2O_3 , MgO , CaO , Na_2O , and K_2O vary by ~2 to 7.5 wt%, while TiO_2 , MnO , and P_2O_5 vary by less than 1 wt% (table 1).

Systematic variations are observed between elemental abundance and relative depth in the deposit (fig. 4). The deposit generally becomes more mafic from the bottom to the top, although as noted above, same samples from individual stratigraphic horizons are heterogeneous. Although the deposit is predominantly basaltic, andesite and dacite are still distinguishable at the bottom of the deposit based on their color.

Trace elements

Trace elements analyzed by DCP-AES include Ni, Zn, Cu, Cr, Zr, V, Sc, Ba, Sr, Rb, and Y (table 1). All trace elements vary in concentration throughout the deposit. Ni, Zn, Cu, Cr, V, Sc, and Sr decrease with increasing SiO_2 , while Zr, Ba, Rb, and Y increase with increasing SiO_2 . Trace element concentration variations range in magnitude from <3-fold (Zn, Zr, Ba, Y, Sr) to >5-fold (Ni, Cr, Rb). Intermediate (3-5 fold) variations are observed for Cu, V, and Sc.

REE abundances increase from basalt to andesite. There is slight increase in light REE (LREE) and no increase in heavy REE (HREE) from andesite to dacite and from the least evolved to more evolved dacite (table 2; fig. 5). Negative Eu anomalies are absent in the basalts, but increase in magnitude from andesite to more evolved dacite.

Petrography

The scorias and pumices from the 1707 deposit share some common characteristics: all are vesicular and porphyritic/vitrophyric. From the least evolved to more evolved samples, the amount of glassy matrix increases and the percentage of phenocrysts decreases from <10 to <2 vol% (basalt to dacite). Some plagioclase and olivine phenocrysts show evidence of

disequilibrium including sieved and resorbed textures (fig. 6).

Basalt

Basalts are the most abundant rock type in the 1707 deposit. Some basalts display diktytaxitic texture. Predominant mineral phases in basalt are plagioclase, olivine, augite and Fe-Ti oxide. Two characteristic matrices observed are dark black quenched matrix and light dark highly vesicular matrix. Relatively larger phenocrysts (<2000um) are often associated with the former matrix, while relatively smaller phenocrysts (<750um) are observed in the latter matrix.

Andesite

The andesitic pumices display a variety of colors from white to dark, and some are banded. Phenocryst phases observed in andesite are plagioclase, hypersthene, augite, and Fe-Ti oxide. Although light and dark bands are visible in pumices under the microscope, there is no significant change in the mineral phases in different colored bands.

Dacite

The dacitic pumices display a variety of colors from white to light-gray, and some are banded. Although phenocryst phases in the dacites are the same as those in the andesite, the dacites have fewer phenocrysts.

Crustal xenoliths

Two types of crustal xenoliths are found at the sampling sites, including basalts and gabbros. The basalts are gray in color and some are vesicular. The phenocryst assemblage in the basalt xenoliths is the same as that in the 1707 basalts. The size of the phenocrysts in the basalt xenoliths are relatively larger (<2000um), and euhedral to subhedral phenocrysts are ubiquitous. The gabbro xenoliths are very light in color, and consist primarily of anorthite (up to 90 vol%) with very few mafic phases.

Mineral chemistry

Representative samples spanning the range of chemical compositions in the 1707 deposit were analyzed for mineral chemistry by electron microprobe. Average chemical compositions of representative feldspar, olivine, hypersthene, augite, and Fe-Ti oxide are summarized in table 3.

Plagioclases are the most abundant phenocrysts throughout the deposit, and range in size from 50 μ m (microphenocrysts) to 2000 μ m. Plagioclases in basalt show a wide range in compositions varying from An₇₀₋₉₀ (fig. 7). In general, the plagioclases decrease in Ca content from core to rim (An₈₀₋₇₆). Plagioclase phenocrysts in andesite range from An₈₅₋₉₀, while microphenocrysts are much more sodic (An₅₉₋₆₃). Anorthite values typically decrease from core to rim. Plagioclase phenocrysts in dacite are much less calcic than those in andesite, ranging from An₆₉₋₅₃.

Olivine is present only in basalt. Olivine is Mg-rich, ranging from Fo₇₀₋₇₉. In the more evolved samples (andesite and dacite), olivine is absent and hypersthene is ubiquitous. The hypersthene are very distinct with green to light brown pleochroism. Augite is a common phase throughout the 1707 deposit. There is no significant difference in chemical compositions of hypersthene and augite from andesite to dacite. Fe-Ti oxides are ubiquitous throughout the 1707 deposit.

Isotope variations

Sr, Nd, and Pb isotopic compositions of six representative juvenile and six crustal xenolith (basalt and gabbro) samples are summarized in table 4. All samples show identical isotopic signatures with $^{87}\text{Sr}/^{86}\text{Sr} = 0.70341 \pm 1$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51304 \pm 1$, falling within the typical range for eruptive products from Fuji Volcano. All samples also show nearly identical Pb signatures with $^{206}\text{Pb}/^{204}\text{Pb} = 18.27 \pm 3$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.51 \pm 1$, and $^{208}\text{Pb}/^{204}\text{Pb} = 38.19 \pm 4$. The only Pb isotopic value reported from Fuji Volcano previously is that for JB 3 (Geological Survey of Japan reference). JB 3 is a basalt from the 864 AD eruption, and is similar to the 1707 samples with $^{206}\text{Pb}/^{204}\text{Pb} = 18.27 \pm 3$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.52 \pm 1$, and $^{208}\text{Pb}/^{204}\text{Pb} = 38.18 \pm 5$ (Geological Survey of Japan reference).

All samples fall just outside of the Pacific MORB field on $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ diagrams, while they fall within the Pacific MORB field on the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (fig. 8).

The Os isotopic compositions of five representative juvenile samples were also measured (table 5). The two basalts show the least radiogenic signatures, with $^{187}\text{Os}/^{188}\text{Os}$ ranging from 0.165 to 0.174 (fig. 9). These values are slightly elevated compared to mantle-derived basalts from mid-ocean ridges and ocean islands (0.13-0.15; Widom, 1997; Shirey and Walker, 1998) but fall within the range of other arc basalts and sub-arc mantle xenoliths (0.13-0.17; Widom et al., 2003a; Widom et al., 2003b). The more evolved samples (one andesite and two dacites) display radiogenic signatures with $^{187}\text{Os}/^{188}\text{Os}$ ranging from 0.26 to 0.39.

Discussion

Major and trace element variations

Major and trace elements show substantial variations throughout the deposit (fig. 3, 4). SiO_2 ranges from 50 to 67 wt%. Other major elements vary by ~2 to 7.5 wt%, with the exception of TiO_2 , MnO , and P_2O_5 , which vary by less than 1 wt%. Most of the trace elements show >3-fold variations, although Zn, Zr, and Ba display <3-fold variations.

The decrease in TiO_2 , Al_2O_3 , Fe_2O_3 , MnO , MgO , CaO , and P_2O_5 with increasing SiO_2 is suggestive of crystal fractionation (fig. 10). Elemental variation diagrams show non-linear trends over entire SiO_2 range, indicating changes in fractionating mineral assemblage (fig. 10). These non-linear trends argue against magma mixing as a dominant process in generating the zonation in the 1707 magma chamber; magma mixing would result in linear trends on all element-element variation diagrams. Furthermore, the REE abundances of the andesites do not fall between those of basalt and dacite, indicating that andesite was not formed by mixing of basalt and dacite (fig. 5).

Least squares major element modeling was done using Igpet 2000. The whole-rock and mineral chemistry used in the major element modeling are summarized in tables 1 and 2 respectively. Closed system crystal fractionation was modeled in three steps, in accordance with the apparent changes in fractionating mineral assemblages suggested by the element-element variation diagrams (fig. 10), from basalt to andesite, andesite to dacite, and less evolved to more evolved dacite. The results of the major element modeling are summarized in tables 6 and 7.

From basalt to andesite, 45.8% fractionation of plagioclase (55.6%), augite (22.2%), olivine (12.1%), and Fe-Ti oxide (10.1%) reproduce the observed chemical compositions of andesite. From andesite to dacite, 26.9% fractionation occurred with a slightly different

fractionating mineral assemblage including plagioclase (60.4%), hypersthene (21.7%), Fe-Ti oxide (10.7%), augite (6.6%), and apatite (0.6%). From the least evolved to most evolved dacite, 17.2% fractionation took place with the same fractionating assemblage, but different relative proportions: plagioclase (67.4%), hypersthene (13.7%), augite (8.8%), Fe-Ti oxide (9.1%), and apatite (1%). Figure 10 illustrates variations in MgO, CaO, and FeO versus SiO₂ along with the major element modeling results. The high degree of fit between observed and modeled liquid compositions are indicated by the extremely low sum of squares of residual (0.100, 0.329, and 0.073 for the three respective fractionation steps).

The major element modeling results are consistent with the observed mineral assemblages, as described previously. Although there is no definitive evidence of apatite in thin section, a few very small (~50 μ m) crystals resembling apatite were observed petrographically and by cathodoluminescence microscopy in andesite and dacite. These crystals have high relief and pseudo-hexagonal shape with apparent zoning and bright orange-red luminescence. Additional evidence for the presence of apatite is provided by major element variations and REE abundance data. P₂O₅ decreases with increasing SiO₂ from andesite to dacite, indicating that some mineral phase is pulling phosphorous out of the system, consistent with apatite fractionation (fig. 11). The slight increase in LREE and lack of increase in HREE from andesite to dacite and least- to most-evolved dacite (fig. 5) further supports the fractionation of trace amounts of apatite, because apatite strongly incorporates REE, particularly the middle REE. The chemical composition for apatite used in the major element modeling was based on average values of 475 samples from apatite in felsic-intermediate rocks not associated with ore-deposits (F-NA OD; Piccoli and Candela, 2002).

The relative mineral proportions and percentages of crystal fractionation obtained from the major element modeling with Iqpet 2000 were applied to trace element modeling (table 8). The mineral-liquid K_d values were taken from Wilson (1989) and the GERM website. The trace element modeling results are consistent with the major element modeling, with variations in trace element concentrations consistent with changing bulk D values for each fractionation step. As fractionation proceeds, the bulk D values of Sr change from 1.03 to 1.56 to 2.02 (fig. 12). Ba and Zr are both incompatible elements, increasing with increasing SiO₂ (fig. 12). Bulk D values of Ba change from 0.15 to 0.46 to 0.63, while bulk D values of Zr change from 0.04 to 0.55 to 0.75 as fractionation proceeds. The modeling results also are consistent with observed REE

variations (fig. 13). All models reproduce observed liquid compositions within analytical error or just outside of the error, further indicating the consistency of the major and trace element modeling. The major and trace element modeling results suggest that crystal fractionation played a major role in the development of chemical zonation in the Fuji 1707 magma chamber.

Sr-Nd-Pb isotope systematics

All juvenile and xenolith samples show identical isotopic signatures with $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.70341 ± 1 and $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.51304 ± 1 , falling within the typical range for eruptive products from Fuji Volcano (Togashi, 1990). All samples also show nearly identical Pb isotopic signatures with $^{206}\text{Pb}/^{204}\text{Pb}$ of 18.27 ± 3 , $^{207}\text{Pb}/^{204}\text{Pb}$ of 15.51 ± 1 , and $^{208}\text{Pb}/^{204}\text{Pb}$ of 38.19 ± 4 , which are very similar to the 864AD basalt. These observations are consistent with the Fuji 1707 chemical zonation being produced by closed system crystal fractionation. The mixing curve between the juvenile samples and typical Japanese upper crust (fig. 8) indicates that $\leq 0.2\%$ assimilation of typical Japanese upper crust could be accommodated.

Evidence for open system behavior

Although closed system crystal fractionation can explain the observed major and trace element variations, petrographic features such as sieved/resorbed plagioclase and olivine phenocrysts (fig. 6) can result from open system process such as crustal contamination, that may have taken place prior to the 1707 eruption.

In order to better constrain the potential role of crustal assimilation, the Os isotope system was utilized. Os isotopes are generally much more sensitive to crustal assimilation than are the Sr, Nd, and Pb systems, due to the extreme difference in $^{187}\text{Os}/^{188}\text{Os}$ between mantle and crust (Widom et al., 1996; Shirey and Walker, 1998; Widom et al., 1999). The $^{187}\text{Os}/^{188}\text{Os}$ signature of the andesite and dacites (0.26-0.39) are distinctly more radiogenic than those of the basalts (0.165-0.174), indicating that crustal assimilation played a role in the formation of the chemically zoned magma chamber. Mixing calculations for assimilation of typical upper crust ($^{187}\text{Os}/^{188}\text{Os}$ of ~ 1.0 ; Peucker-Ehrenbrink and Jahn, 2001) by Fuji magmas indicate the radiogenic signatures observed in the more evolved Fuji samples can be explained by $\leq 0.2\%$ of crustal assimilation (fig. 9), which would not cause significant variation in Sr and Nd isotopes. The basement rocks beneath Fuji (Tertiary marine sediments and green volcanic tuffs; Miyaji,

1990) are likely assimilants, but chemical and isotopic compositions are unavailable.

Proposed model

The petrography, major and trace element systematics, and Sr, Nd, Pb, and Os isotope results suggest that the 1707 chemically zoned magma chamber of Fuji Volcano evolved through three main stages including: 1) fractionation of parental magma; 2) formation of an evolved, chemically zoned magma chamber; and 3) basaltic intrusion into the magma chamber. Figure 14 illustrates each stage as a cartoon diagram.

Stage I: fractionation of parental magma

Because Fuji has erupted almost exclusively basaltic magmas throughout its history, it is inferred that the first stage involved crystal fractionation of basaltic parental magma. The compositional gap between basalt and andesite may indicate that the 1707 basalt is not directly related to the 1707 andesite and dacite. However, it is assumed that the chemical composition of basaltic parental magma that fractionated to form the silicic magmas was similar to the 1707 basalt, based on the chemical similarity of basalts erupted from Fuji Volcano throughout its history (Takahashi et al., 1991). Major, trace, and REE modeling are consistent with 45.8 % crystal fractionation of plagioclase (55.6%), augite (22.2%), olivine (12.1%), and Fe-Ti oxide (10.1%) from basalt to andesite (fig. 10).

Stage II: formation of a chemically zoned magma chamber

During the second stage, crystal fractionation continued within the magma chamber; 26.9% fractionation of andesite with a crystallizing assemblage of plagioclase (60.4%), hypersthene (21.7%), Fe-Ti oxide (10.7%), augite (6.6%), and apatite (0.6%) produced the least evolved dacitic magma. Further crystal fractionation formed more evolved dacitic magma; 17.2% fractionation of plagioclase (67.4%), hypersthene (13.7%), augite (8.8%), Fe-Ti oxide (9.1%), and apatite (1%) can explain the chemical compositions of the most evolved samples from the 1707 deposit. During stage II, a minor amount of crustal assimilation occurred. The assimilation is consistent with the sieved/resorbed textures in some plagioclase and olivine phenocrysts, which indicate that the crystals experienced chemical disequilibrium during their evolution. In addition, minor crustal assimilation is required to explain the elevated $^{187}\text{Os}/^{188}\text{Os}$

signatures in the more evolved samples. The small amount of assimilation would not be expected to significantly affect the major and trace elements or the Sr, Nd, and Pb isotope signatures.

Stage III: basaltic intrusion into the magma chamber

The compositional gap between basalt and andesite suggests that during the final stage, basaltic magma was intruded into a pre-existing silicic magma chamber consisting of andesite and dacite. The basaltic intrusion may have triggered the explosive 1707 eruption. Based on the relative erupted volumes, the amount of basaltic magma intruded into the magma chamber was likely much greater than the volume of evolved magma in the magma chamber. These three stages can explain the observed chemically zoned volcanic deposit of 1707 eruption of Fuji Volcano.

Liquid Immiscibility

Although liquid immiscibility is a common small-scale process in basaltic magmas (Fujii et al., 1980), it has not been recognized as a magma chamber-scale process in silicic magmas. All major, trace, and REE modeling for the Fuji 1707 deposit show consistency with crystal fractionation as the dominant petrogenetic process. Although Kawamoto (1990, 1991a, 1991b, 1992) proposed that the observed trace element abundances and a decrease in P_2O_5 from andesite to dacite cannot be explained by crystal fractionation, our study shows that both trace element variations and P_2O_5 abundances are consistent with crystal fractionation. The fact the crystal fractionation is a common process, and that it can easily explain the observed chemical gradients in the magma chamber, lead us to favor this model. We find that there is no need to resort to unusual processes such as liquid immiscibility to explain the 1707 chemically zoned magma chamber.

Conclusions

Major and trace element concentrations show substantial variations throughout the Fuji 1707 deposit. Decreasing abundances of MgO, CaO, Al_2O_3 , and P_2O_5 with increasing SiO_2 and non-linear trends on elemental variations are best explained by crystal fractionation of the observed phenocryst assemblage. Trace element models are consistent with the major element

modeling. Chondrite-normalized REE abundances show expected enrichment from basalt to andesite and display strong negative Eu anomalies that increase from andesite to dacite, further supporting magma evolution by crystal fractionation. Sr, Nd, and Pb isotopic compositions show identical signatures throughout the deposit. Although these observations are all consistent with closed system crystal fractionation, a minor amount of crustal contamination is required by the radiogenic Os isotope signatures in the more evolved samples.

Our combined geochemical, isotopic, and petrographic observations suggest that the 1707 magma chamber evolved through three main stages including: 1) fractionation of parental magma; 2) formation of a chemically zoned magma chamber by crystal fractionation and minor crustal assimilation; and 3) basaltic intrusion into the magma chamber. Basalt intrusion during the final stage likely catalyzed the explosive eruption leading to progressive emptying of the dacitic and andesitic chamber magmas and the volumetrically substantial eruption of the youngest basaltic materials.

Fig. 1. Plate tectonic framework of Japan (modified after Taira, 2001). Fuji Volcano is located in south-central Honshu, lying over the westward-subducting Pacific Plate and possibly influenced by the Philippine Sea plate (Arculus et. al., 1991).

Fig. 2. A photograph of sampling site (35°21'703"N, 138°47'035"E). This sampling site displays a complete depositional section of the 1707 eruption. The total thickness of the deposit is 277 cm.

Fig. 3. Alkali versus silica classification diagram (after LeBas et al., 1986). The samples span a wide range of SiO_2 from 50 to 67 wt % (basalt to dacite); however, there is a compositional gap between 52 to 57 wt% SiO_2 .

Fig. 4. Variation in CaO (wt %) and Sr concentration (ppm) versus relative depth in the 1707 deposit at the second sampling site (fig. 2). JF-20 (Ho II) and JF-21 (Ho I) are heterogeneous with mixture of pumices and scorias. JF-20 was separated into light (JF-20L) and dark (JF-20D) fractions, while JF-21 was separated into very light (JF-21V), light (JF-21L), banded (JF-21B), and dark (JF-21D) fractions. The vertical axis is Miyaji's (1984) classification.

Rock/Chondrites

Fig. 5. Chondrite-normalized REE patterns for the Fuji 1707 deposit. The samples show relative enrichment from basalt to andesite. A slight increase in LREE and no increase in HREE from andesite to evolved dacite are observed. Chondrite abundances are from Nakamura (1974). (LED: least evolved dacite; MED: most evolved dacite)

Fig. 6. Petrographic evidence of disequilibrium textures in plagioclase and olivine phenocrysts (10X magnification with crossed polarized light).

Fig. 8. Sr, Nd, and Pb isotopic compositions of representative juvenile and xenolith samples from the 1707 deposit. Also shown are fields for Pacific MORB (Ito et. al., 1987), Pacific sediments (BenOthman et. al., 1989), and Japanese upper crust (Geological Survey of Japan; Rezanov et al., 1999; Kondo et al., 2000). The square field ($^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ diagram) includes all volcanic products from Fuji Volcano (Togashi, 1990). The mixing curve shown on the $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ diagram represents bulk mixing between juvenile basalts and typical Japanese upper crust. Tick marks on mixing curve represent fraction of crust assimilated.

Fig. 8 (cont).

Fig. 9. Os isotopic compositions of representative samples from the Fuji 1707 deposit. Also shown are fields for MORB (Widom et al., 1997; Shirey and Walker, 1998), arc basalts (Widom et al., 2003a; Widom et al., 2003b), and upper crust (Peucker-Ehrenbrink and Jahn, 2001). The radiogenic signatures observed in the andesite and dacites can be explained by $\leq 0.2\%$ of crustal assimilation. Labels on data points indicate percentage of crustal assimilation required to produce the observed radiogenic signatures assuming the uncontaminated magmas had 0.3 ppt Os and $^{187}\text{Os}/^{188}\text{Os} = 0.170$, and the crustal assimilant had 50 ppt Os and $^{187}\text{Os}/^{188}\text{Os} = 1.06$, similar to estimates for average upper continental crust (Esser and Turekian, 1993; Peucker-Ehrenbrink and Jahn, 2001).

Fig. 10. Major element variation diagrams with modeling results. Arrows are schematic representations of modeled fractionation steps. The sum of squares of residuals were 0.100, 0.329, and 0.073 for the respective modeling steps. (Calc. Parent: Calculated parent; Obs. Parent: Observed parent; Calc. Daughter: Calculated daughter)

Fig. 11. P₂O₅ versus SiO₂ for the Fuji 1707 deposit. P₂O₅ decreases with increasing SiO₂ from andesite to dacite, consistent with apatite fractionation. The higher P₂O₅ contents of the andesites compared to the basalts indicate the absence of apatite fractionation during this stage.

Fig. 12. Trace element variation diagrams with modeling results. Trace element modeling is consistent with major element modeling. Arrows are schematic representations of modeled fractionation steps.

Fig. 13. Chondrite-normalized REE patterns with modeling results. The mineral assemblages and the relative proportions obtained from major element modeling were used for REE modeling. The error bars represent 13% analytical errors based on estimated agreement of standards and linearity of regression lines. Chondrite values are from Nakamura (1974). (Obs.: Observed; Calc: Calculated; LED: least evolved dacite; MED: most evolved dacite)

Fig. 14. Cartoon diagram of the evolution of the 1707 chemically zoned magma chamber. Stage 1: fractionation of parental basaltic magma to form evolved magma. Stage 2: formation of a chemically zoned magma chamber by crystal fractionation with minor crustal assimilation. Stage 3: basaltic intrusion into the magma chamber, possibly triggering eruption (Blake, 1984).

Table 1. Major and trace concentrations of juvenile and xenolith samples from the 1707 Fuji deposit. Major elements are in weight %. Trace elements are in ppm.

Sampling site 1									
Sample	JF-2a	JF-2b	JF-2c	JF-2d	JF-2e	JF-2f	JF-3a	JF-3b	JF-3c
units	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II
SiO₂	58.14	57.25	57.04	59.83	57.80	64.05	61.47	60.63	62.05
TiO₂	1.13	1.07	1.10	0.96	1.07	0.83	0.91	1.01	0.90
Al₂O₃	16.80	16.29	16.57	15.93	16.49	15.92	15.87	16.34	16.05
Fe₂O₃	8.88	8.33	8.53	7.42	8.35	6.12	6.93	7.91	6.72
MnO	0.17	0.16	0.17	0.15	0.16	0.13	0.14	0.15	0.13
MgO	3.17	2.94	3.03	2.48	2.95	1.89	2.25	2.67	2.19
CaO	6.79	6.40	6.58	5.80	6.37	5.07	5.42	6.12	5.34
Na₂O	3.61	3.55	3.58	3.64	3.56	3.77	3.65	3.64	3.68
K₂O	1.39	1.40	1.36	1.60	1.41	2.10	1.80	1.65	1.88
P₂O₅	0.33*	0.47	0.44	0.29*	0.31*	0.25*	0.41	0.16	0.39
LOI	-0.03	-0.08	-0.12	0.05	0.02	0.23	0.02	0.00	0.23
Total	100.39	97.79	98.27	98.13	98.50	100.35	98.86	100.29	99.56
Ni	12	8	0	7	13	5	5	7	4
Zn	102	111	108	86	87	85	79	100	86
Cu	69	68	74	69	71	57	65	70	61
Cr	0	6	3	0	9	0	7	1	5
Zr	142	145	137	145	138	166	154	151	153
V	207	193	193	180	205	124	166	184	141
Sc	23	24	24	19	23	15	19	21	18
Ba	375	368	360	403	364	1594	422	401	404
Sr	391	364	374	341	367	316	325	362	322
Rb	34	21	24	47	46	53	38	24	35
Y	35	34	33	32	33	32	34	34	33

*Measured using standard addition method.

Table 1 (cont).

Sampling site 1									
	JF-3d	JF-3e	JF-3f	JF-4a	JF-4b	JF-4d	JF-4e	JF-4f	JF-5a
	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho II	Ho I
SiO₂	59.84	59.82	63.48	59.47	59.67	60.62	64.55	62.71	62.73
TiO₂	1.03	0.99	0.82	1.03	1.01	0.98	0.77	0.83	0.88
Al₂O₃	16.37	16.19	15.79	16.40	16.42	16.17	15.46	15.76	15.77
Fe₂O₃	8.04	7.58	6.23	8.20	7.74	7.46	5.69	6.35	6.67
MnO	0.16	0.15	0.13	0.16	0.15	0.14	0.12	0.13	0.14
MgO	2.73	2.54	1.97	2.87	2.62	2.48	1.76	1.98	2.11
CaO	6.31	5.88	5.10	6.46	5.99	5.82	4.76	5.19	5.50
Na₂O	3.66	3.64	3.75	3.66	3.64	3.66	3.75	3.80	3.79
K₂O	1.59	1.59	2.14	1.51	1.57	1.64	2.27	2.06	1.94
P₂O₅	0.29*	0.29*	0.31	0.28	0.44	0.50	0.21*	0.29	0.32
LOI	-0.11	-0.03	0.23	-0.01	0.05	0.01	0.53	0.34	-0.06
Total	99.90	98.63	99.95	100.03	99.31	99.47	99.88	99.44	99.77
Ni	7	1	11	14	5	3	3	6	4
Zn	89	88	83	101	95	85	66	67	76
Cu	71	74	54	76	74	74	51	60	61
Cr	0	2	0	2	2	1	1	0	2
Zr	148	147	166	140	145	151	166	158	161
V	185	184	136	185	183	169	139	132	157
Sc	21	21	16	22	23	22	13	16	16
Ba	397	395	495	386	392	404	449	430	423
Sr	370	347	316	373	354	345	298	320	332
Rb	40	32	43	24	41	45	50	35	52
Y	33	33	34	34	33	35	33	33	32

Table 1 (cont).

Sampling site 1			
	JF-5b	JF-5c	JF-5d
	Ho I	Ho I	Ho I
SiO₂	61.90	66.51	66.75
TiO₂	0.92	0.68	0.69
Al₂O₃	16.04	15.14	15.26
Fe₂O₃	6.82	4.87	4.84
MnO	0.13	0.11	0.10
MgO	2.16	1.43	1.41
CaO	5.41	4.26	4.13
Na₂O	3.70	3.81	3.71
K₂O	1.78	2.52	2.41
P₂O₅	0.28*	0.20	0.19*
LOI	-0.01	0.31	0.35
Total	99.13	99.83	99.85
Ni	1	5	0
Zn	96	63	60
Cu	75	42	52
Cr	5	3	8
Zr	159	164	178
V	158	101	101
Sc	21	11	16
Ba	436	448	472
Sr	329	273	265
Rb	28	43	45
Y	35	29	36

Table 1 (cont).

Sampling site 2

	JF-10	JF-11	JF-12	JF-13	JF-14	JF-15	JF-16	JF-17	JF-18
	Ho IV	Ho IV	Ho IV	Ho IV	Ho IV	Ho IV	Ho IV	Ho III	Ho III
SiO₂	50.41	51.43	50.50	50.78	50.61	51.61	50.96	51.01	50.87
TiO₂	1.35	1.36	1.35	1.34	1.33	1.34	1.33	1.34	1.33
Al₂O₃	17.06	17.11	16.80	16.76	17.14	17.10	16.89	17.14	16.93
Fe₂O₃	11.92	11.87	11.98	11.96	11.90	11.84	11.91	11.91	11.76
MnO	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19	0.19
MgO	5.26	5.23	5.30	5.28	5.30	5.26	5.36	5.37	5.26
CaO	9.59	9.60	9.59	9.51	9.51	9.49	9.63	9.76	9.67
Na₂O	2.70	2.68	2.63	2.62	2.65	2.83	2.61	2.65	2.67
K₂O	0.73	0.74	0.73	0.72	0.93	0.67	0.72	0.72	0.73
P₂O₅	0.23*	0.29	0.23*	0.35	0.17	0.22*	0.24	0.23*	0.36
LOI	-0.21	-0.27	-0.18	-0.24	-0.26	-0.23	-0.23	-0.20	-0.29
Total	99.22	100.24	99.10	99.27	99.46	100.33	99.62	100.12	99.49
Ni	26	24	32	24	29	22	26	19	21
Zn	101	110	100	105	112	93	89	104	108
Cu	164	169	179	175	176	168	174	178	171
Cr	40	39	39	41	48	43	43	42	42
Zr	75	79	77	76	78	73	73	76	73
V	395	403	429	399	426	406	396	375	423
Sc	36	38	39	38	40	37	36	36	39
Ba	197	219	213	218	214	210	219	214	213
Sr	389	394	387	384	385	391	386	392	389
Rb	8	22	0	28	0	2	7	0	19
Y	23	26	26	26	26	24	25	24	24

*Measured using standard addition method.

Table 1 (cont).

Sampling site 2

	JF-19	JF-20L	JF-20D	JF-21V	JF-21L	JF-21D	JF-21B
	Ho III	Ho II	Ho II	Ho I	Ho I	Ho I	Ho I
SiO₂	51.52	60.43	58.43	65.26	62.37	59.48	61.96
TiO₂	1.33	1.03	1.12	0.75	0.94	1.04	0.95
Al₂O₃	17.16	16.47	16.61	15.47	16.21	16.41	16.20
Fe₂O₃	11.59	7.80	8.72	5.61	7.02	8.11	7.13
MnO	0.19	0.15	0.17	0.12	0.14	0.16	0.14
MgO	5.12	2.64	3.09	1.71	2.26	2.82	2.31
CaO	9.54	6.10	6.76	4.69	5.66	6.42	5.70
Na₂O	2.79	3.67	3.55	3.75	3.87	3.60	3.74
K₂O	0.80	1.61	1.30	2.29	1.77	1.44	1.78
P₂O₅	0.24*	0.30*	0.48	0.23*	0.28*	0.39	0.39
LOI	-0.22	0.15	-0.03	0.39	0.10	-0.01	0.13
Total	100.05	100.37	100.20	100.28	100.06	99.85	100.43
Ni	19	9	12	4	5	10	5
Zn	86	92	99	71	95	90	105
Cu	161	75	81	54	69	79	66
Cr	38	5	8	0	0	1	0
Zr	80	145	131	172	152	137	148
V	371	176	198	109	143	186	155
Sc	36	21	24	14	19	22	18
Ba	231	383	364	461	406	371	410
Sr	393	363	383	292	342	365	345
Rb	11	51	37	45	51	28	34
Y	26	31	31	34	33	31	33

Table 1 (cont).**Xenoliths: basalt**

	JF-6a	JF-6b	JF-6c	JF-6d	JF-7a	JF-8a	JF-9b
SiO₂	49.68	51.42	51.10	51.15	50.49	50.08	50.92
TiO₂	1.37	0.98	1.37	0.96	1.31	1.34	0.92
Al₂O₃	16.89	18.65	17.14	18.43	17.04	16.92	17.72
Fe₂O₃	12.60	10.18	12.16	10.67	11.80	12.25	10.52
MnO	0.19	0.15	0.19	0.17	0.18	0.19	0.17
MgO	5.94	5.51	5.06	5.43	6.33	6.14	6.10
CaO	9.87	9.37	9.40	9.60	9.80	9.99	10.28
Na₂O	2.37	2.85	2.91	2.80	2.51	2.40	2.56
K₂O	0.60	0.51	0.68	0.47	0.59	0.54	0.37
P₂O₅	0.26	0.24	0.31	0.12	0.28	0.24	0.29
LOI	-0.39	-0.28	-0.30	-0.30	-0.29	-0.28	-0.11
Total	99.39	99.60	100.02	99.50	100.05	99.79	99.73
Ni	49	48	26	25	57	36	38
Zn	102	82	119	90	98	107	95
Cu	199	114	147	84	170	196	94
Cr	57	55	30	31	110	99	81
Zr	71	48	71	39	76	71	41
V	409	314	385	367	386	399	336
Sc	38	32	32	31	41	43	35
Ba	205	160	208	150	196	202	133
Sr	363	397	410	396	363	354	359
Rb	5	17	13	10	11	18	5
Y	25	18	26	18	27	27	17

Table 1 (cont).

Xenoliths: gabbro					
	JF-1	JF-4c	JF-7b	JF-8b	JF-9a
SiO₂	48.40	44.90	46.68	46.80	46.94
TiO₂	1.11	0.19	0.23	1.14	0.57
Al₂O₃	16.35	23.19	23.17	18.97	25.92
Fe₂O₃	12.14	7.67	6.24	12.70	5.65
MnO	0.18	0.12	0.10	0.16	0.08
MgO	7.81	8.82	7.60	7.69	2.81
CaO	10.23	13.29	14.74	11.44	14.64
Na₂O	2.22	1.00	0.95	1.75	1.77
K₂O	0.13	0.05	0.03	0.18	0.09
P₂O₅	0.21	0.12	0.10	0.13	0.12
LOI	-0.34	-0.27	-0.12	-0.35	-0.07
Total	98.45	99.08	99.72	100.61	98.51
Ni	64	83	77	64	13
Zn	102	133	36	89	45
Cu	40	15	12	64	53
Cr	110	153	276	137	84
Zr	13	6	8	22	5
V	461	64	100	459	220
Sc	41	19	28	36	17
Ba	81	37	39	81	73
Sr	382	463	452	419	590
Rb	0	3	14	35	7
Y	14	6	7	13	6

Table 2. REE signatures of representative juvenile samples. Concentrations are in ppm.

	Basalt		Andesite		Dacite	
	JF-10	JF-15	JF-2a	JF-3b	JF-3f	JF-5c
La	7.0	6.1	12.1	12.9	13.0	14.1
Ce	18.9	15.7	33.5	35.4	36.0	37.9
Nd	13.5	11.9	20.8	21.5	21.0	21.9
Sm	3.8	3.5	5.4	5.5	5.4	5.5
Eu	1.2	1.1	1.6	1.5	1.3	1.2
Gd	4.0	3.1	6.0	5.8	5.6	5.7
Dy	4.2	4.1	5.5	5.8	5.5	6.0
Er	2.2	1.9	3.4	3.4	3.3	3.5
Yb	2.3	2.1	3.3	3.5	3.4	3.6
Lu	0.3	0.3	0.5	0.5	0.5	0.5

Table 3. Average chemical compositions of phenocrysts. Major elements are in wt%.
(Plg: Plagioclase; CPX: Clinopyroxene; OPX: Orthopyroxene; Fe-Ti: Fe-Ti oxide)

	Basalt		Andesite			Dacite		
	Plag	Olivine	Plag	CPX	OPX	Fe-Ti	Plag	OPX
SiO₂	49.73	38.00	48.92	51.38	52.40	0.78	55.62	52.08
TiO₂	-	-	-	0.41	0.17	9.90	-	0.20
Cr₂O₃	-	0.05	-	0.01	0.07	0.12	-	0.06
Al₂O₃	30.74	-	31.73	1.71	1.49	2.31	27.23	1.54
MgO	-	39.17	-	13.29	23.76	1.85	-	23.47
Fe₂O₃	-	0.00	-	1.07	2.62	48.32	-	2.15
CaO	15.17	0.21	15.55	20.56	1.23	0.07	10.74	1.33
MnO	0.01	0.31	0.01	0.46	0.79	0.45	0.03	0.72
FeO	0.90	23.01	0.91	10.06	18.08	37.47	0.84	18.16
Na₂O	2.61	-	2.44	0.27	-	-	4.00	0.01
K₂O	0.10	-	0.09	-	-	-	0.49	-
NiO	-	0.12	-	-	-	-	-	-
Total	99.26	100.85	99.65	99.23	100.61	101.93	98.95	99.71

Table 4. Sr, Nd, and Pb isotopic compositions of representative juvenile and xenolith samples. Strontium isotopic ratios were corrected for fractionation using $^{86}\text{Sr}/^{88}\text{Sr}=0.1194$. NBS 987 gave an average of $^{87}\text{Sr}/^{86}\text{Sr}=0.710241\pm 1$ (n=80). Neodymium isotopic ratios were corrected for fractionation using $^{144}\text{Nd}/^{146}\text{Nd}=0.7219$. La Jolla gave an average of $^{144}\text{Nd}/^{143}\text{Nd}=0.51185\pm 1$ (n=50). $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ were corrected for fractionation by 0.1%, 0.1%, and 0.125% per amu respectively. Errors on $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ measurements were ± 0.006 , ± 0.008 , and ± 0.026 respectively.

	Basalt		Andesite		Dacite	
	JF-10	JF-15	JF-2a	JF-3b	JF-3f	Jf-5c
$^{87}\text{Sr}/^{86}\text{Sr}$	0.7034	0.7034	0.7034	0.7034	0.7034	0.7034
$^{144}\text{Nd}/^{143}\text{Nd}$	0.51304	0.51304	0.51304	0.51304	0.51304	0.51304
$^{206}\text{Pb}/^{204}\text{Pb}$	18.284	18.284	18.269	18.257	18.250	18.241
$^{207}\text{Pb}/^{204}\text{Pb}$	15.511	15.513	15.510	15.502	15.504	15.502
$^{208}\text{Pb}/^{204}\text{Pb}$	38.208	38.214	38.200	38.166	38.168	38.157

	Xenolith: basalt		Xenolith: gabbro	
	JF-6b	JF-6c	JF-7b	Jf-8b
$^{87}\text{Sr}/^{86}\text{Sr}$	0.7034	0.7034	0.7034	0.7034
$^{144}\text{Nd}/^{143}\text{Nd}$	0.51304	0.51305	0.51304	0.51304
$^{206}\text{Pb}/^{204}\text{Pb}$	18.277	18.263		
$^{207}\text{Pb}/^{204}\text{Pb}$	15.519	15.512		
$^{208}\text{Pb}/^{204}\text{Pb}$	38.228	38.181		

Table 5. Os isotopic compositions of representative juvenile samples. Internal measurement errors ranged from 0.5-2%. Blank correction errors (shown in the data table) ranged from 0.6-10% based on blanks of 1.0 ± 0.1 pg and $^{187}\text{Os}/^{188}\text{Os} = 0.174$. Replicate analyses (JF-15 and JF-15R) agreed within 3.6% for $^{187}\text{Os}/^{188}\text{Os}$.

	Basalt			Andesite	Dacite	
	JF-10	JF-15	JF-15R	JF-3b	JF-3f	Jf-5c
Os ppb	0.0024	0.0015	0.0025	0.0003	0.0003	0.0003
$^{187}\text{Os}/^{188}\text{Os}$	0.174 ± 1	0.165 ± 1	0.171 ± 1	0.26 ± 1	0.39 ± 4	0.30 ± 2

Table 6. The calculated fractionating mineral assemblages with the relative mineral proportions. (pl: plagioclase, ol: olivine; cpx: clinopyroxene, opx: orthopyroxene; Ti-Fe: Ti-Fe oxide; ap: apatite; FC%: percentage of fractional crystallization obtained by major element modeling)

Basalt		Andesite		Dacite	
pl	55.6%	pl	60.4%	pl	67.4%
cpx	22.2%	opx	21.7%	opx	13.7%
ol	12.1%	mt	10.7%	mt	9.1%
mt	10.1%	cpx	6.6%	cpx	8.8%
		ap	0.6%	ap	1.0%
%FC	45.8%	% FC	26.9%	% FC	17.2%

Table 7. The results of major element modeling. Major elements are in wt%. Total Fe as Fe_2O_3 (from DCP-AES) was converted to FeO. (Pobs: Observed parent; Pcalc: Calculated parent; D: Daughter)

	Basalt				Andesite				Dacite			
	Pobs	Pcalc	D		Pobs	Pcalc	D		Pobs	Pcalc	D	
SiO_2	51.58	51.82	58.46		58.46	58.77	64.06		64.06	64.29	67.33	
TiO_2	1.36	1.13	1.14		1.14	0.92	0.83		0.83	0.75	0.7	
Al_2O_3	17.22	17.31	16.89		16.89	17	15.93		15.93	16.03	15.39	
FeO	10.81	10.87	8.04		8.04	8.09	5.66		5.66	5.69	4.39	
MnO	0.19	0.18	0.17		0.17	0.16	0.13		0.13	0.12	0.1	
MgO	5.34	5.34	3.19		3.19	3.13	1.99		1.99	1.96	1.42	
CaO	9.71	9.72	6.83		6.83	6.83	5.15		5.15	5.15	4.17	
Na_2O	2.72	2.66	3.63		3.63	3.16	3.78		3.78	3.57	3.74	
K_2O	0.76	0.78	1.4		1.4	1.59	2.16		2.16	2.07	2.43	
P_2O_5	0.31	0.14	0.25		0.25	0.3	0.31		0.31	0.35	0.33	
Total	100.00	99.95	100.00		100.00	99.96	100.00		100.00	99.98	100.00	

Table 8. Results of trace element modeling. Bulk D values are calculated using this formula: $Bulk\ D = \sum X_a * (Kd)_a$ where a=mineral phase; X=fraction of mineral in total crystallizing assemblage; Kd=solid-liquid partition coefficient for the mineral. Kd values are taken from Wilson (1989) and GERM website. (pl: plagioclase, ol: olivine; cpx; clinopyroxene, opx: orthopyroxene; Ti-Fe: Ti-Fe oxide; ap: apatite; FC%: percentage of fractional crystallization obtained by major element modeling)

Basalt							Andesite						
Kd	pl	cpx	ol	mt	Bulk D		Kd	pl	opx	cpx	mt	ap	Bulk D
Zr	0.05	0.003	0.05	0.02	0.04		Zr	0.45	0.04	0.8	2	0.636	0.55
Ba	0.23	0.0515	0.03	0.025	0.15		Ba	0.67	0.015	0.1	0.4	0.05	0.46
Sr	1.8	0.06	0.01	0.1	1.03		Sr	2.52	0.005	0.3	0.11	1.3	1.56
La	0.022	0.002	0.0004	0.015	0.01		La	0.4	0.031	0.05	0.45	14.5	0.39
Ce	0.016	0.017	0.007	0.016	0.02		Ce	0.2214	0.040	0.0838	0.42	21.1	0.32
Nd	0.014	0.06	0.0046	0.026	0.02		Nd	0.1487	0.028	0.1828	0.55	32.8	0.36
Sm	0.5	0.2	0.03	0.2	0.35		Sm	0.8	0.050	0.4	0.55	46	0.86
Eu	0.3	1	0.03	0.3	0.42		Eu	1.326	0.080	3	0.5	25.5	1.22
Gd	0.0066	0.6	0.1	0.018	0.15		Gd	0.0665	0.039	0.5826	0.3	43.9	0.38
Dy	0.314	1.18	0.007	0.3	0.47		Dy	0.0498	0.076	0.7739	0.58	34.8	0.37
Er	0.041	0.5	0.2	0.01	0.16		Er	-	-	-	-	-	-
Yb	0.06	1.05	0.04	0.5	0.32		Yb	0.06	0.600	0.95	0.47	15.4	0.37
Lu	0.037	1.12	0.018	0.023	0.27		Lu	0.0389	0.323	0.6654	0.44	13.8	0.27
FC%	55.6%	22.2%	12.1%	10.1%	100.0%		FC%	60.4%	21.7%	6.6%	10.7%	0.6%	100.0%

Table 8 (cont).

Dacite						
Kd	pl	opx	mt	cpx	ap	Bulk D
Zr	0.55	0.0333	3.9	0.1841	0.636	0.75
Ba	0.88	0.02	0.028	0.4	0.05	0.63
Sr	2.93	0.1	0.077	0.11	1.3	2.02
La	-	-	-	-	-	-
Ce	0.347	0.0163	0.057	0.044	21	0.46
Nd	0.29	0.0162	0.55	0.1661	52	0.78
Sm	0.153	0.0174	0.074	0.4574	27	0.42
Eu	1.6	0.89	0.062	0.4108	14.5	1.39
Gd	0.129	0.027	0.3	0.7028	21.7	0.40
Dy	0.25	0.0412	0.58	0.7759	25	0.55
Er	-	-	-	-	-	-
Yb	0.302	0.1149	0.11	0.64	9.4	0.38
Lu	0.29	0.154	0.14	0.683	30.2	0.59
FC%	67.40%	13.70%	9.10%	8.80%	1.00%	100.00%

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Appendix 1. The Fuji 1707 Sample Processing Procedures.

	HC	US bath	JC	DM	SB	R-Type	Thin section
JF-1	n	15m x 2	metal	n	BM	x:gabbro	-
JF-2a	pl	15m x 4	y	n	y	andesite	polished
JF-2b	pl	15m X 4	y	n	y	andesite	-
JF-2c	pl	15m X 4	y	n	y	andesite	-
JF-2d	pl	15m X 4	y	n	y	andesite	-
JF-2e	pl	15m X 4	y	n	y	andesite	-
JF-2f	pl	15m x 4	y	n	y	dacite	polished
JF-3a	pl	15m X 4	y	n	y	andesite	-
JF-3b	pl	15m x 4	y	n	y	andesite	polished
JF-3c	pl	15m X 4	y	n	y	andesite	-
JF-3d	pl	15m x 4	y	n	y	andesite	unpolished
JF-3e	pl	15m X 4	y	n	y	andesite	-
JF-3f	pl	15m x 4	y	n	y	dacite	polished
JF-4a	pl	15m x 4	y	n	y	andesite	unpolished
JF-4b	pl	15m x 4	y	n	y	andesite	-
JF-4c	n	15m x 2	metal	n	BM	x:gabbro	-
JF-4d	pl	15m x 4	y	n	y	andesite	-
JF-4e	pl	15m x 4	y	n	y	dacite	polished
JF-4f	pl	15m x 4	y	n	y	andesite	unpolished
JF-5a	pl	15m x 4	y	n	y	andesite	unpolished
JF-5b	pl	15m x 4	y	n	y	andesite	-
JF-5c	pl	15m x 4	y	n	y	dacite	polished
JF-5d	pl	15m x 4	y	n	y	dacite	-
JF-6a	sp	15m x 2	metal	y x 1	y	x:basalt	-
JF-6b	sp	15m x 2	metal	y x 1	y	x:basalt	-
JF-6c	pl	15m x 2	metal	y x 2	y	x:basalt	unpolished
JF-6d	pl	15m x 2	metal	y x 2	y	x:basalt	unpolished
JF-7a	n	15m x 2	metal	y x 1	y	x:basalt	-
JF-7b	n	15m x 2	metal	y x 1	y	x:gabbro	-
JF-8a	n	15m x 2	metal	y x 1	y	x:basalt	-
JF-8b	n	15m x 2	metal	n	BM	x:gabbro	-
JF-9a	pl	15m x 2	metal	y x 2	y	x:gabbro	unpolished
JF-9b	pl	15m x 2	metal	y x 2	y	x:basalt	unpolished
JF-10	n	15m x 4	y	n	y	basalt	polished
JF-11	n	15m x 4	y	n	y	basalt	unpolished
JF-12	n	15m x 4	y	n	y	basalt	polished
JF-13	n	15m x 4	y	n	y	basalt	unpolished
JF-14	n	15m x 4	y	n	y	basalt	unpolished
JF-15	n	15m x 4	y	n	y	basalt	polished
JF-16	n	15m x 4	y	n	y	basalt	unpolished
JF-17	n	15m x 4	y	n	y	basalt	polished

	HC	US bath	JC	DM	SB	R-Type	Thin section
JF-18	n	15m x 4	y	n	y	basalt	unpolished
JF-19	n	15m x 4	hc (pl)	n	y	basalt	polished
JF-20L	n	15m x 4	hc (pl+pa)	n	y	andesite	polished
JF-20D	n	15m x 4	hc (pl+pa)	n	y	andesite	polished
JF-21V	n	15m x 4	y	n	y	dacite	polished
JF-21L	n	15m x 4	y	n	y	andesite	unpolished
JF-21D	n	15m x 4	y	n	y	andesite	unpolished
JF-21B	n	15m x 4	y	n	y	andesite	polished

HC: hand crush using a rock hammer; SP: steel splitter; US bath: ultrasonic bath; JC: alumina jaw crusher; DM: alumina discmill; SB: alumina shatter box; BM: alumina ball mill; pl: plastic; pa: paper; y: yes; n: no; x: xenolith

Appendix 2. Major (wt%) and trace (ppm) element variations of the Fuji 1707 deposit.

Appendix. 3. Mineral chemistry of the Fuji 1707 deposit. Major elements are in weight %.

Olivine	JF17	JF17	JF17	JF10	JF10	JF10	JF10	JF10	JF10
	basalt	basalt	basalt	basalt	basalt	basalt	basalt	basalt	basalt
SiO₂	38.40	37.71	38.48	38.68	38.57	38.61	37.63	37.19	
Cr₂O₃	0.00	0.04	0.01	0.07	0.08	0.03	0.03	0.05	
MgO	39.35	36.32	38.76	40.47	41.99	40.27	38.38	36.04	
CaO	0.20	0.18	0.24	0.22	0.15	0.21	0.21	0.26	
MnO	0.30	0.42	0.39	0.30	0.21	0.29	0.32	0.40	
FeO	22.57	26.78	23.75	20.26	19.42	22.03	23.51	27.08	
NiO	0.10	0.10	0.08	0.12	0.24	0.13	0.02	0.08	
total	100.91	101.54	101.71	100.12	100.65	101.57	100.09	101.09	
mo:	0.27	0.25	0.33	0.31	0.21	0.28	0.29	0.36	
fo:	75.62	70.59	74.35	78.06	79.38	76.49	74.38	70.28	
fa:	24.33	29.2	25.56	21.93	20.59	23.47	25.55	29.63	
li:	0.1	0.1	0.08	0.13	0.24	0.13	0.02	0.08	
te:	0.33	0.46	0.42	0.33	0.23	0.32	0.35	0.44	
cations									
Si	0.9917	0.9882	0.9914	0.9962	0.9852	0.9882	0.9858	0.9820	
Cr	0.0000	0.0009	0.0003	0.0015	0.0016	0.0006	0.0007	0.0010	
Mg	1.5151	1.4190	1.4885	1.5538	1.5989	1.5364	1.4991	1.4185	
Ca	0.0055	0.0050	0.0065	0.0061	0.0041	0.0057	0.0058	0.0072	
Mn	0.0066	0.0092	0.0084	0.0065	0.0046	0.0064	0.0071	0.0089	
Fe	0.4874	0.5870	0.5117	0.4365	0.4148	0.4715	0.5150	0.5980	
Ni	0.0020	0.0021	0.0017	0.0026	0.0049	0.0027	0.0004	0.0017	
total	3.0083	3.0114	3.0085	3.0032	3.0141	3.0115	3.0139	3.0173	

Orthopyroxene				Clinopyroxene				Oxide	
	JF-20D	JF-3f	JF-3f	JF-20L	JF-20L	JF-20L	JF-20L	JF-20L	and
	and	dacite	dacite	and	and	and			
SiO ₂	52.40	49.72	53.25	51.46	51.99	50.70		SiO ₂	0.08
TiO ₂	0.17	0.24	0.19	0.34	0.30	0.58		TiO ₂	9.90
Al ₂ O ₃	1.49	2.28	1.21	1.79	1.26	2.09		Al ₂ O ₃	2.31
Cr ₂ O ₃	0.07	0.07	0.04	0.02	0.00	0.01		V ₂ O ₃	0.68
Fe ₂ O ₃	2.62	4.11	1.22	1.14	0.40	1.68		Cr ₂ O ₃	0.12
MgO	23.76	22.54	23.99	14.09	13.17	12.62		Fe ₂ O ₃	48.32
CaO	1.23	1.48	1.30	20.90	20.41	20.37		MgO	1.85
MnO	0.79	0.73	0.70	0.38	0.51	0.48		CaO	0.07
FeO	18.08	16.84	18.71	8.38	11.08	10.72		MnO	0.45
Na ₂ O	0.00	0.00	0.00	0.26	0.26	0.29		FeO	37.47
total	100.61	98.01	100.60	98.76	99.38	99.55		total	101.24
wo:	2.5	3.18	2.6	44.14	42.72	43.64			
en:	67.44	67.35	66.98	41.41	38.34	37.62			
fs:	30.06	29.47	30.42	14.45	18.94	18.74			
cations									
Si	1.9259	1.8827	1.9514	1.9434	1.9673	1.9232			
Ti	0.0046	0.0069	0.0052	0.0096	0.0087	0.0167			
Al	0.0645	0.1017	0.0523	0.0797	0.0562	0.0933			
Cr	0.002	0.0021	0.0012	0.0006	0	0.0004			
Fe	0.0725	0.1171	0.0336	0.0325	0.0113	0.048			
Mg	1.302	1.2726	1.3103	0.7932	0.7428	0.7135			
Ca	0.0483	0.0601	0.0509	0.8455	0.8275	0.8277			
Mn	0.0246	0.0235	0.0217	0.0122	0.0163	0.0154			
Fe	0.5556	0.5334	0.5734	0.2646	0.3506	0.3401			
Na	0	0	0.0001	0.0188	0.0194	0.0216			
total	4	4.0001	4.0001	4.0001	4.0001	3.9999			

And: Andesite

Feldspar

	JF-17 basalt	JF-17 basalt	JF-19 basalt	JF-19 basalt	JF-19 basalt	JF-19 basalt	JF-19 basalt	JF-10 basalt	JF-10 basalt	JF-10 basalt	JF-10 basalt	JF-10 basalt	JF-20D basalt	JF-20L basalt	JF-20L basalt	JF-20L basalt
SiO ₂	46.41	46.80	49.40	50.12	50.90	50.33	49.91	49.79	51.02	44.91	49.09	54.55	45.64	45.15		
Al ₂ O ₃	33.75	34.02	30.43	30.83	29.30	30.25	30.87	31.40	30.41	34.62	31.46	27.62	34.00	34.98		
CaO	17.52	17.84	15.09	15.10	13.88	14.85	14.35	14.32	14.14	18.37	15.63	11.60	18.19	18.42		
MnO	0.03	0.04	0.01	0.00	0.05	0.03	0.03	0.02	0.03	0.00	0.00	0.01	0.02	0.00		
FeO	0.88	0.78	0.98	0.98	1.37	0.85	1.01	0.98	0.84	0.90	0.77	1.12	0.88	0.70		
Na ₂ O	1.38	1.36	2.59	2.57	3.12	2.87	3.22	3.02	3.15	1.05	2.41	4.34	1.20	1.10		
K ₂ O	0.09	0.10	0.08	0.13	0.15	0.10	0.15	0.14	0.20	0.08	0.08	0.17	0.06	0.04		
total	100.06	100.93	98.58	99.73	98.77	99.28	99.54	99.67	99.80	99.93	99.45	99.41	100.00	100.39		
ab:	12.44	12.03	23.6	23.37	28.67	25.76	28.63	27.39	28.38	9.35	21.74	39.96	10.67	9.73		
or:	0.53	0.56	0.46	0.76	0.89	0.61	0.88	0.81	1.21	0.48	0.5	1.02	0.33	0.21		
an:	87.04	87.41	79.95	75.87	70.44	73.63	70.49	71.79	70.41	90.17	77.77	59.02	89	90.06		
cs:	0	0	0	0	0	0	0	0	0	0	0	0	0	0		
cations																
Si	2.141	2.1403	2.2957	2.3001	2.3568	2.3188	2.2965	2.2853	2.334	2.0825	2.2624	2.485	2.1126	2.0812		
Al	1.8351	1.8338	1.6667	1.6675	1.5991	1.6426	1.6742	1.6986	1.6398	1.8918	1.709	1.4828	1.8549	1.9003		
Ca	0.866	0.8741	0.7511	0.7426	0.6885	0.7332	0.7075	0.7045	0.6931	0.9126	0.7718	0.566	0.9018	0.9099		
Mn	0.001	0.0017	0.0003	0	0.0018	0.0013	0.0012	0.0008	0.0013	0	0	0.0005	0.0008	0		
Fe	0.0338	0.0299	0.0382	0.0378	0.053	0.0326	0.0388	0.0378	0.0321	0.035	0.0295	0.0428	0.0342	0.0269		
Na	0.1237	0.1203	0.2334	0.2287	0.2803	0.2565	0.2874	0.2688	0.2794	0.0946	0.2157	0.3833	0.1081	0.0983		
K	0.0052	0.0056	0.0045	0.0075	0.0087	0.0061	0.0089	0.008	0.0119	0.0049	0.0049	0.0098	0.0033	0.0021		
total	5.0058	5.0057	4.9899	4.9842	4.9882	4.9911	5.0145	5.0038	4.9916	5.0214	4.9933	4.9702	5.0157	5.0187		

And: Andesite

Feldspar		JF-20L	JF-20L	JF-3f	JF-4e	JF-4e	JF-4e
		and	and	dacite	dacite	dacite	dacite
SiO₂	46.34	52.92	56.79	54.72	53.94	57.05	
Al₂O₃	33.60	28.45	26.29	28.18	28.65	25.82	
CaO	17.19	12.38	10.65	10.35	10.80	11.14	
MnO	0.00	0.00	0.03	0.01	0.05	0.02	
FeO	0.86	0.99	1.12	0.63	0.39	1.21	
Na₂O	1.71	3.84	4.42	4.79	4.75	2.06	
K₂O	0.03	0.16	0.38	0.31	0.27	0.99	
total	99.72	98.73	99.66	98.99	98.85	98.30	
ab:	15.24	35.58	41.88	44.67	43.62	23.22	
or:	0.16	0.99	2.35	1.93	1.63	7.36	
an:	84.6	63.43	55.77	53.4	54.76	69.42	
cs:	0	0	0	0	0	0	
cations							
Si	2.1442	2.4324	2.5676	2.4912	2.4622	2.6029	
Al	1.8323	1.5411	1.4008	1.5119	1.5417	1.3883	
Ca	0.8523	0.6096	0.5158	0.5051	0.528	0.5447	
Mn	0	0	0.0012	0.0005	0.002	0.0008	
Fe	0.0332	0.0381	0.0422	0.0238	0.0149	0.0463	
Na	0.1535	0.342	0.3874	0.4224	0.4206	0.1822	
K	0.0016	0.0095	0.0217	0.0182	0.0157	0.0578	
total	5.0171	4.9727	4.9367	4.9731	4.9851	4.823	

Appendix 4. Photomicrographs of the Fuji 1707 deposit. Photographs are taken either with plane polarized light (PP) or crossed polarized light (XP) with 4X, 10X or 20X magnification.

Basalt

Olivine and oxide (PP/10X)

Augite and plagioclase (PP/20X)

Resorbed olivine (PP/10X)

Euhedral Plagioclase (PP/10X)

Euhedral olivine (PP/10X)

Resorbed Plagioclase (PP/10X)

Andesite

Hypersthene (PP/20X)

Plagioclase (PP/10X)

Hypersthene (PP/20X)

Plagioclase (PP/20X)

Augite (PP/20X)

Hypersthene and oxide (PP/4X)

Dacite

Hypersthene (PP/10X)

Augite (PP/10X)

Plagioclase (PP/10X)

Plagioclase phenocrysts and microphenocrysts (XP/4X)

Hypersthene (PP/20X)

Hypersthene and oxide (PP/10X)