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Silicic magmatism in the upper crust and relationships to caldera-forming eruptions: insights from the Takidani magmatic system in Central Japan

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Silicic magmatism in the upper crust and relationships to
caldera-forming eruptions: Insights from the Takidani
magmatic system in Central Japan

THÈSE

présentée à la Faculté des sciences de l'Université de Genève
pour obtenir le grade de Docteur ès Sciences, mention sciences de la Terre

par
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de
Aachen (Allemagne)

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Thèse de Madame Eva HARTUNG

intitulée :

**«Silicic Magmatism in the Upper Crust and Relationships to
Caldera-forming Eruptions: Insights from the Takidani
Magmatic System in Central Japan»**

La Faculté des sciences, sur le préavis de Monsieur L. CARICCHI, professeur associé et directeur de thèse (Département des sciences de la Terre), Monsieur S. WALLIS, professeur (School of science, University of Tokyo, Japan), Monsieur O. MÜNTENER, professeur (Institut des sciences de la Terre, Université de Lausanne, Suisse), Monsieur O. BACHMANN, professeur (Institut für Geochemie und Petrologie, Eidgenössische Technische Hochschule Zürich, Schweiz), Monsieur G. GUALDA, professeur (Department of Earth and Environmental Sciences, Vanderbilt University, Nashville, U.S.A.), autorise l'impression de la présente thèse, sans exprimer d'opinion sur les propositions qui y sont énoncées.

Genève, le 16 novembre 2018

Thèse - 5282 -

Le Doyen

N.B. - La thèse doit porter la déclaration précédente et remplir les conditions énumérées dans les "Informations relatives aux thèses de doctorat à l'Université de Genève".

Abstract

During the course of this PhD study, a detailed petrological investigation of the Takidani pluton and related volcanic deposits was performed. The Takidani pluton in Central Japan is one of the youngest exposed plutons in the world. It presents a sub-volcanic magmatic reservoir that is associated with two voluminous caldera-forming eruptions, including the Nyukawa pyroclastic flow deposit (PFD) and Chayano–Ebisutoge pyroclastic deposit (PD).

The Takidani Pluton is vertically zoned and comprises at least six distinct lithological units consisting of granodiorites and granites. Textural and geochemical evidence for melt extraction from a crystal-rich magma is preserved in the upper section of the pluton. Thermodynamic simulations and thermal modelling suggest that magma had sufficient time to segregate before the solidification of the pluton, supporting an origin of granitic magma by fractional crystallisation. Geochemical analyses of the Chayano–Ebisutoge PD indicate that the generation of rhyolitic melts was dominated by crustal melting processes. This implies that erupted rhyolites cannot represent the extracted melts of the Takidani Pluton and that the caldera-forming eruptions of the Chayano–Ebisutoge are not genetically related to the Takidani Pluton. Petrogenetic links, however, exist between the Takidani Pluton and the dacitic Nyukawa PFD supporting conceptual models of crystal mush remobilisation.

New major and trace element analyses of whole-rock and minerals provide insights into the magmatic plumbing systems of large caldera-forming eruptions in Central Japan. The results of this study show that both, fractional crystallisation and crustal melting, play an important role in the generation of silicic magma. The relationship between the plutonic and volcanic rocks remains to be further investigated.

Résumé

Au cours de cette étude de doctorat, une étude pétrologique détaillée du pluton de Takidani et des dépôts volcaniques associés a été réalisée. Le pluton de Takidani au centre du Japon est l'un des plus jeunes plutons exposés au monde. Il présente un réservoir magmatique sub-volcanique associé à deux éruptions volumineuses formant des calderas, à savoir le dépôt d'écoulement pyroclastique de Nyukawa et le dépôt pyroclastique de Chayano-Ebisutoge.

Le pluton de Takidani est zoné verticalement et comprend au moins six unités lithologiques distinctes composées de granodiorites et de granites. Les preuves texturales et géochimiques de l'extraction de melt d'un magma riche en cristaux sont préservées dans la partie supérieure du pluton. Des simulations thermodynamiques et des modélisations thermiques suggèrent que le magma a eu suffisamment de temps pour se séparer avant la solidification du pluton, ce qui justifie une origine de magma granitique dominée par de la cristallisation fractionnée. L'analyse géochimique du dépôt pyroclastique de Chayano-Ebisutoge indique que la génération des melts rhyolitiques a été engendrée par des processus de fusion crustale. Cela implique que la rhyolite issue de l'éruption ne peut pas présenter de melt extrait du pluton de Takidani et que l'éruption formant la caldera n'est pas génétiquement liée au pluton de Takidani. Des liens pétrogénétiques existent entre le pluton de Takidani et le dépôt dacitique de Nyukawa, supportant des modèles conceptuels de remobilisation de la bouillie de cristaux.

De nouvelles analyses d'éléments majeurs et traces sur roches totales et minéraux permettent de mieux comprendre les systèmes de plomberie magmatiques des grandes éruptions à l'origine de la formation de caldera dans le centre du Japon. Les résultats de cette étude montrent que la cristallisation fractionnée et la fusion de la croûte jouent un rôle important pour la production du magma siliceux. La relation existant entre les roches plutoniques et volcaniques reste à approfondir.

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List of Abbreviations

a	Years (date)
A	Ampere (SI unit)
Ab	Albite
AFC	Assimilation and fractional crystallisation
Al	Aluminium
am	Amphibole
AMS	Anisotropy of Magnetic Susceptibility
a.m.u.	Atomic mass unit
An	Anorthite
ap	Apatite
a.p.f.u.	Atoms per formula unit
Ar	Argon
As	Arsenic
B	Boron
Ba	Barium
Br	Bromine
BS	Bulk susceptibility
BSE	Backscattered electrons
bt	Biotite
C	Carbon
°C	Degree Celsius (SI unit)
Ca	Calcium
Ce	Cerium
Ch	Chayano Tuff
CIPW	Cation norm after Cross, Iddings, Pirrson and Washington
Cl	Chlorine
chl	Chlorite
CL	Cathodoluminescence
ccm	Cubic meter
cm	Centimeter
cm ²	Square centimeter
cm ³	Cubic centimeter
Co	Cobalt
cpx	Clinopyroxene
Cr	Chromium
Cs	Caesium
Cu	Copper
Dy	Dysprosium
DRE	Dense rock equivalent
Eb	Ebisutoge PD
EDS	Electron dispersive spectrometer
EDX	Electron dispersive X-rays
En	Enstatite

ep	Epidote
EPMA	Electron probe micro-analyser
Er	Erbium
Eu	Europium
F	Fluorine
F	Fluorine
FC	Fractional crystallisation
Fe	Iron
fGT	Fine-grained biotite granite (lithological unit of TK)
fGT2	Fine-grained porphyritic granite (lithological unit of TK)
fs	Feldspar
Fs	Ferrosilite
Ga	Gallium
Gd	Gadolinium
Ge	Germanium
GDT	Equigranular hbl-bt-granodiorite (lithological unit of TK)
GRT	Equigranular granodiorite (lithological unit of TK)
h	Hours
H	Hydrogen
hbl	Hornblende
He	Helium
Hf	Hafnium
Hg	Mercury
HF	Hydrofluoric acid
HMR	Hida Mountain Range
HNO ₃	Nitric acid
Ho	Holmium
Hz	Hertz (SI unit)
H ₂ O	Water
ICPMS	Inductively coupled plasma mass spectrometry
ID-TIMS	Isotope dilution thermal ionization mass spectrometry
ilm	Ilmenite
K	Potassium
K	Kelvin (SI unit)
ka	Thousand years (date)
kbar	Kilobars
keV	Kilo electron volt
kfs	Alkali feldspar
Kr	Krypton
kV	Kilo volt
kyr	Thousand years (duration)
La	Lanthanum
LA	Laser ablation
Li	Lithium
Li ₂ B ₄ O ₇	Lithium tetraborate
Lu	Lutetium
m	Meter
m ²	Square meter
m ³	Cubic meter
Ma	Million years (date)
mag	Magnetite

Mg	Magnesium
Mg#	Magnesium number
mGT	Marginal granodiorite (lithological unit of TK)
mm	Millimeter
Mn	Manganese
Mo	Molybdenum
MPa	Mega Pascal
Myr	Million years (duration)
μm	Micrometer
Na	Sodium
nA	Nano ampere
Nb	Niobium
Nd	Neodymium
Ni	Nickel
nm	Nanometer
NNO	Nickel-nickel-oxygen
Nyu	Nyukawa PFD
O	Oxygen
Or	Orthoclase
opx	Orthopyroxene
P	Phosphorus
<i>P</i>	Pressure
Pa	Pascal (SI unit)
Pb	Lead
PD	Pyroclastic deposit
PDF	Pyroclastic flow deposit
pGT	Porphyritic granodiorite and granite (lithological unit of TK)
plg	Plagioclase
Pm	Promethium
ppm	Parts per million
Pr	Praseodymium
QEMSCAN	Quantitative evaluation of minerals by scanning electron microscopy
qtz	Quartz
Rb	Rubidium
REE	Rare Earth elements
s	Seconds (SI unit)
S	Sulfur
Sc	Scandium
Se	Selenium
SE	Secondary electrons
SEM	Scanning electron microscopy
Si	Silica
SI	International system of Units (Système international d'unités)
SIMS	Secondary ion mass spectrometry
Sm	Samarium
sp	Spinel
SPO	Shape preferred orientation
Sr	Strontium
<i>T</i>	Temperature
Ta	Tantalum
Tb	Terbium

Ti	Titanium
Th	Thorium
TK	Takidani Pluton
Tl	Thallium
Tm	Thulium
ttn	Titanite
U	Uranium
V	Vanadium
V	Volt (SI unit)
W	Tungsten
W	Watt (SI unit)
WDS	Wavelength dispersive spectrometer
Wo	Wollastonite
XRF	X-ray fluorescence
Y	Yttrium
Yb	Ytterbium
yr	Year (duration)
Zn	Zinc
zr	Zircon
Zr	Zirconium

Physical Constants

Acceleration of gravity	$g = 9.81 \text{ ms}^{-2}$
Pi (Mathematical constant)	$\pi = 3.14159$

List of Symbols

A	Area	m^2
c	Specific heat	$\text{J kg}^{-1} \text{K}^{-1}$
J	Joule (Energy)	$\text{kg m}^2 \text{s}^{-2}$
k	Thermal conductivity	$\text{W m}^{-1} \text{K}^{-1}$
K	Permeability coefficient	
L	Latent heat of crystallisation	$\text{J kg}^{-1} \text{K}^{-1}$
N	Newton (Force)	kg m s^{-2}
Pa	Pascal (Pressure)	$\text{kg m}^{-1} \text{s}^{-2}$
r	Radius	m
t	Time	s
T	Temperature	K
Th	Thickness	m
V	Volume	m^3
z	Depth	m
μ	Viscosity	Pa s
ϕ	Porosity	m^2
ρ	Density	kg m^{-3}
κ	Permeability	m^2
κ	Magnetic susceptibility	SI
η	Melt viscosity	Pa s

Chapter 1

Introduction

1.1 General Introduction

Silicic magmatism produces some of the largest volcanic eruptions on Earth (Miller & Wark 2008). The products of these voluminous eruptions are found in different geological settings around the world and can total up to tens to thousands of cubic kilometres of pyroclastic deposits. Voluminous eruptions are generally associated with caldera volcanoes, which are identified as large topographic basins that form as result of subsidence into partially emptied magma reservoirs (Figure 1.1). The size of calderas can vary widely and yield information on the volume of the associated eruption. Some of the largest known eruptions include the Fish Canyon Tuff (5000 km³, Lipman et al. 1997), the Young Toba Tuff (2800 km³, Rose & Chesner 1987), the Lund Tuff (2600 km³, Maughan et al. 2002) and the Huckleberry Ridge Tuff (2450 km³, Christiansen 1984). Caldera-forming eruptions produce caldera fill deposits comprising thick welded ignimbrite sheets that are rapidly deposited during caldera collapse, as well as breccias and volcanoclastic lake sediments (Figure 1.1). Associated deposits outside the caldera walls include voluminous ignimbrite sheets and widespread ash fall deposits. Two types of deposits are typically associated with caldera-forming eruptions: (1) crystal-rich dacites (i.e. Fish Canyon Tuff; Lipman et al. 1997) commonly referred to as monotonous intermediates (Hildreth 1981), and crystal-poor rhyolites (i.e. Bishop Tuff; Hildreth 1981).

Over the last decades, our conceptual understanding of magma reservoirs associated with large caldera-forming events has changed from a big-tank full of melts to reservoirs filled by magma with different degrees of crystallinity (i.e. crystal mush, Cashman & Giordano 2014, Cooper 2017, Bachmann & Bergantz 2004, 2008, Hildreth & Wilson 2007). In our current understanding, monotonous intermediates present crystal-rich magmas that were rapidly mobilised by thermal reactivation (Bachmann & Bergantz 2004, Huber et al. 2010). The generation of rhyolitic (or granitic) melts is associated with various processes including (1) melting of crustal lithologies (Sisson et al. 2005) and (2) liquid-crystal segregation of crystallising magma (Bowen 1928, Bachmann & Bergantz 2008, McKenzie 1984). In both cases, substantial amounts of hot mafic melt are required due to thermal limitations on crustal melting (Dufek &

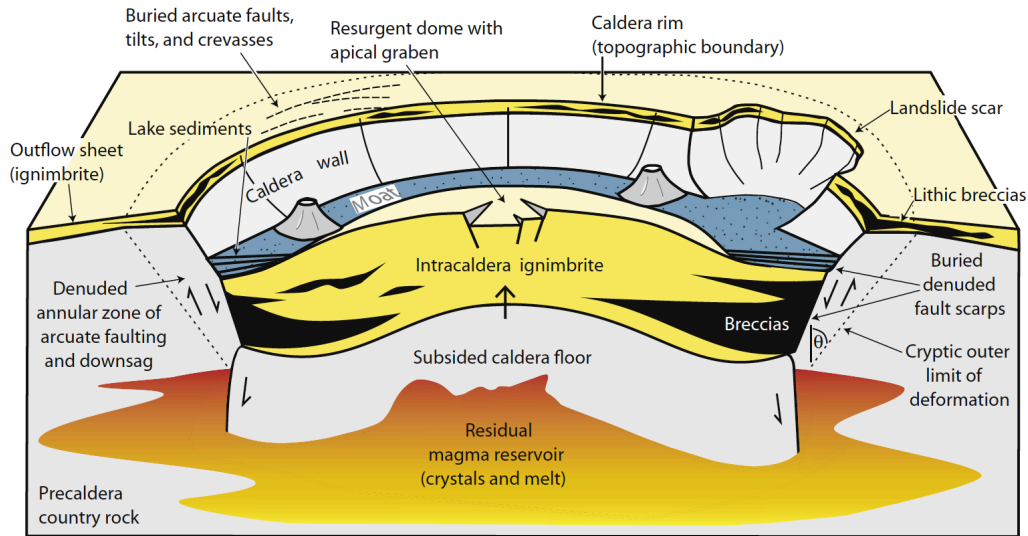


FIGURE 1.1: General features of a resurgent explosive caldera from Branney & Acocella (2015) showing the caldera rim, subsided caldera floor overlain by an intracaldera infill. Outer caldera deposits are indicated as ignimbrite sheets.

Bergantz 2005) and magma accumulation and reservoir growth (Schöpa & Annen 2013, Blundy & Annen 2016).

Geochronological studies show that intrusive bodies grow over long timescales of hundred thousands to millions of years (Annen et al. 2015), while volcanic activity occurs repeatedly on short timescales. The chemical similarities of volcanic and plutonic rocks has led to many questions about the relationship between intrusive and extrusive rocks and architecture of the Earth's crust (Glazner et al. 2018, 2015, 2008, Lundstrom & Glazner 2016). Are volcanic and plutonic rocks genetically related? Do plutons represent frozen magma reservoirs that once fed volcanic eruptions? How are large silicic magmas and eruptions generated? What is the origin of granitic and rhyolitic melts? This research project aims to answers some of these questions.

1.2 Aim of this study

The goal of this study is to gain a deeper understanding of the physical processes that lead to the emplacement and chemical evolution of magmatic reservoirs in the Earth's crust. This thesis focuses on the Takidani Pluton in Central Japan, one of the youngest exposed plutons in the world (Harayama 1992). The Takidani Pluton is thought to present the shallow magmatic reservoir that fed two large caldera-forming eruptions in the Quarternary (Kimura & Nagahashi 2007); thus, providing a unique location to study the architecture of magmatic plumbing systems beneath caldera-forming volcanoes and to investigate the relationship between plutonic and volcanic rocks.

During the first part of this doctoral study I worked on the processes of residual melt extraction observed within the Takidani Pluton. During the second part of my studies I focused on measurements that target the connection between the Takidani pluton and coexisting eruptive products.

1.3 Geological overview

The formation of the island arcs of Japan is dominated by subduction of the Pacific and Philippine Sea plates, which started more than 450 Myr ago (Jahn 2010, Wakita 2013, Taira et al. 2016 and references therein). Igneous activity related to the subduction is recorded in Cambrian granites (520 Ma), the Hikami granite (412 Ma) and in the metagranites and granites in the Hida Belt during the Paleozoic, Cenozoic and Mesozoic (330–180 Ma, Wakita 2018, 2013 and references therein). During the Cretaceous and Paleogene, extensive igneous activity is recorded across the Japanese islands including the Ryoke Granite (100–70 Ma, Nakajima 1994, Morioka et al. 2000), the San-yo and San-in Granites (110–37 Ma, Wakita 2018), the Abukuma Granite (100–90 Ma, Shibata & Ishihara 1979) and Kitakami Granite (120–110 Ma, Kawano & Ueda 1965). The youngest plutonism is recorded in the Quaternary and include the Kurabegawa Granite (0.9–2.3 Ma; Ito et al. 2017) and Takidani Pluton (1.6 Ma; Ito et al. 2017) in the Japan Alps in Central Japan.

The Japan Alps are located at the triple junction of the Northeastern Japan Arc, the Southwest Japan Arc and the Izu Bonin Arc. They are the largest mountain range in Japan and comprise the Hida Mountain Range (HMR), a typical alpine mountain range that stretches along 150 km in N-S direction. Hotaka-dake (3190 m) is the highest mountain in this alpine range and the third highest mountain in Japan, after Mount Fuji (3776 m) and Kita-dake (3193 m). Tectonic uplift of HMR started during the late Tertiary (late Miocene) and continues until present. The uplift during the Quaternary is estimated up to 5 mm per year, and is the highest rate recorded in Japan. The mountain range is bound by the Itoigawa-Shizuoka tectonic line to the east and fault zones at the western periphery of the Fossa Magna.

The Norikura Volcanic Chain is an active volcanic chain that sits on the crest of the North Japan Alps (Figure 1.2). Activity began in the southern half of the volcanic chain, at about 2.7 Ma and include the Ueno Basalts, the Nomugi-toge Volcanic Rocks, and the Hida Volcanic Rocks. The Nyukawa Pyroclastic Flow Deposit (PFD; 1.76 Ma) and Chayano–Ebisutoge Pyroclastic Deposits (PD; 1.75 Ma) are part of the felsic volcanic activity of the Hida Volcanic Rocks. The Ueno Basalts (Kimura & Yoshida 1999, Kimura et al. 2002) and the andesitic Nomugi-Toge Volcanic Rocks (Kimura & Yoshida 1999, Kimura et al. 2002) are regarded as contemporaneous with the Hida volcanic activity (Kimura & Yoshida 1999), and continued until 1.5 Ma (Kimura & Yoshida 1999). After about 300 thousand years of repose, six new eruption centres

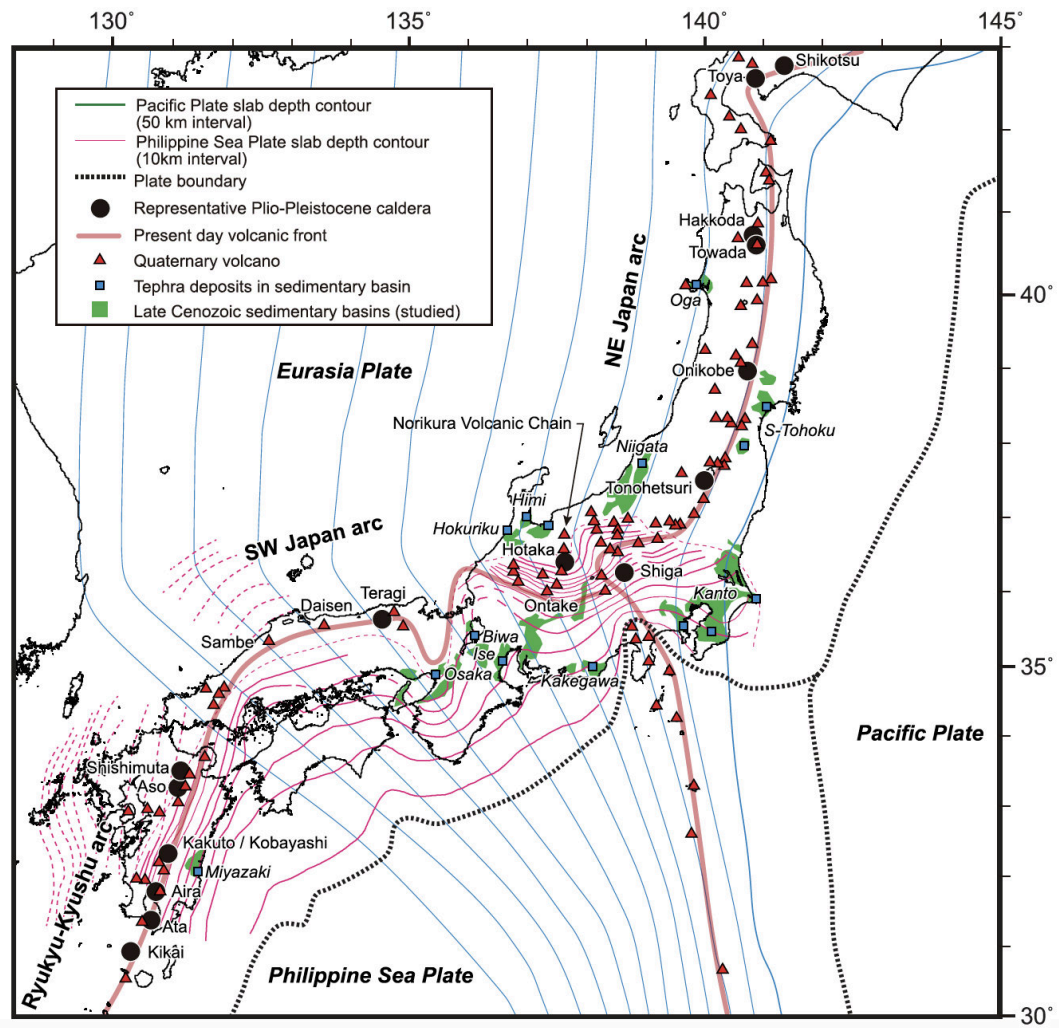


FIGURE 1.2: Locations of Quaternary volcanoes and Plio-Pleistocene calderas in Japan. Contours of slab depths of subducted Pacific plate and the Philippine Sea plate (from Kimura et al. 2015).

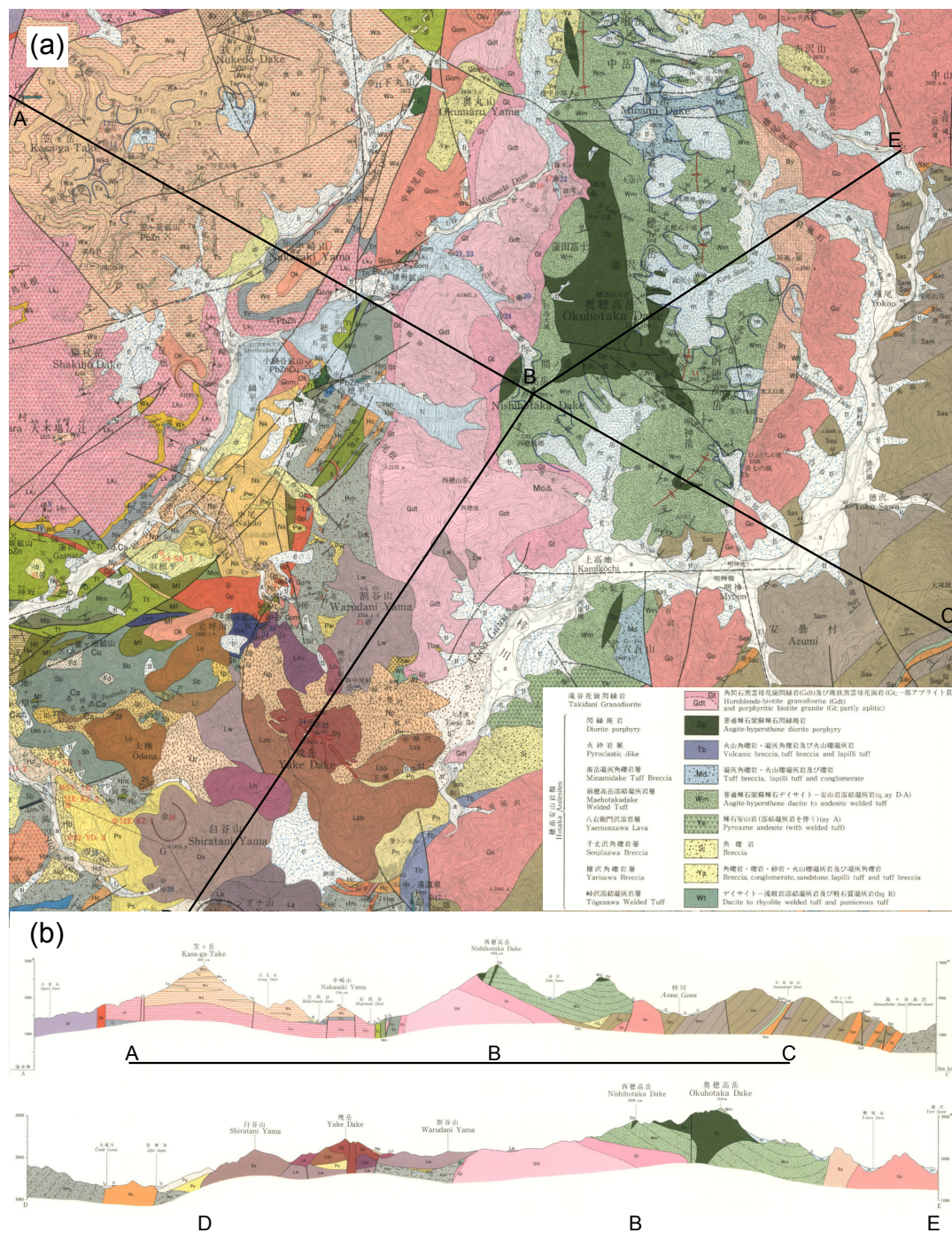


FIGURE 1.3: (a) Geological map of the Hida Mountain Range by Harayama (1990). (b) Cross-sections of the geological map showing eastward tilting of the Takidani-Hotaka plutonic-volcanic complex

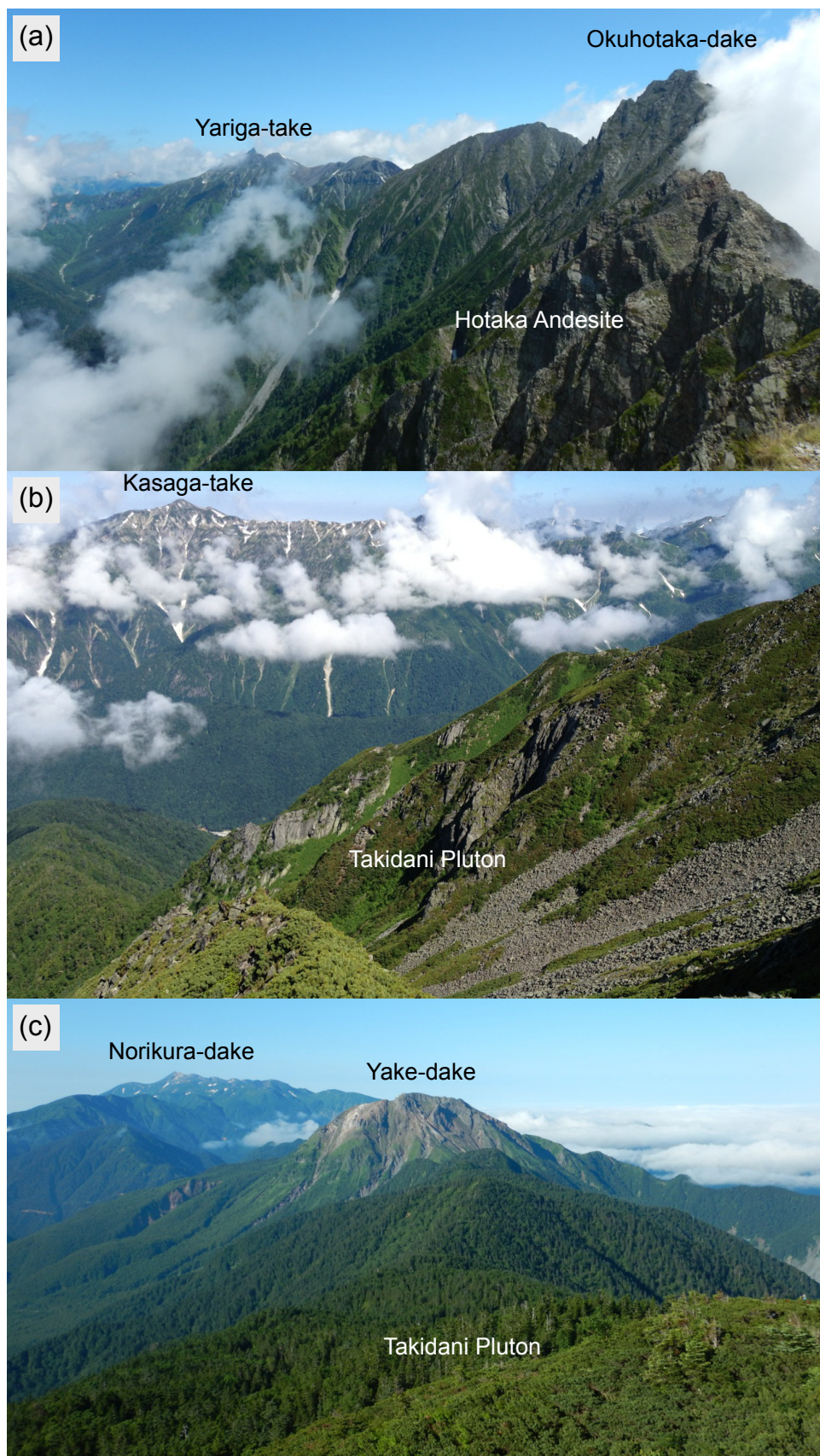


FIGURE 1.4: (a) Northwards view on Hotaka Andesite from the top of Nishihotaka-dake; (b) Eastwards view on Kasaga-take and Takidani Pluton; (c) Southwards view on Yake-dake and Norikura-Dake

became active. The volcanic centres are distributed over the entire North Japan Alps and comprise Ontake-dake, Norikura-dake, Yake-dake, Washiba-Kumono-taira, Tateyama, and Shirouma-Oike volcanoes, from south to north, respectively. Most of these volcanoes have records of eruption within the past 300 kyr and Ontake-dake, Yake-dake, and Tateyama have been active historically (Kimura & Yoshida 1999) including a phreatic explosion at Mount Ontake, which took place on 27 September 2014 and killed 63 people.

The Takidani pluton (TK) is a young felsic intrusion (1.6 Ma, Ito et al. 2017) located within the active Norikura Volcanic Chain between Yake-dake and Hotake-dake (Figure 1.3, 1.4). TK comprises at least six distinct units (Figure 1.5–1.7) that vary from fine-to-medium-grained biotite-granite at the base to medium-grained hbl-bt-granodiorites in the middle section of the pluton (Figure 1.7; Harayama 1992, Harayama et al. 2003). The upper section of the pluton consists of fine and medium-grained porphyritic granodiorites and granites (Figure 1.7). Along its eastern margin TK forms a plutonic-volcanic complex with the Hotaka Andesite (2.4 Ma; Figure 1.3, Harayama 1990, Yamada et al. 1985), a 3 km thick volcanic complex of the Late Pleistocene. The main unit of HA, a 1500 m thick andesitic to dacitic welded tuff sheet, was emplaced as single unit inside a volcanic-tectonic graben (i.e. caldera) elongated in a north-south direction. The Takidani Pluton was emplaced at the base of Hotaka welded tuff sheet and its intrusion led to a 1500 m thick contact aureole of hornblende hornfels and albite-epidote-hornfels (Harayama 1990). The Takidani-Hotaka complex was tilted eastwards (10–30°) during uplift (Figure 1.3b). Simultaneous erosion rates led to a nearly 2 km vertical exposure of the Takidani magmatic reservoir (Bando et al. 2003, Harayama et al. 2003).

1.4 Field campaign

Field work was carried during a two week long mapping and sampling campaign in July 2014. Field measurements and rock samples were taken along several transects that run from the structural base of the pluton to the roof contact with the Hotaka Andesite. Most transects run in direction NW-SE. A total of 110 samples were collected for textural and geochemical analysis including 50 oriented samples for a structural study of the rock fabrics. The results of this part of the study are being processed and will be presented in the appendix to this thesis. In addition, 102 measurements of bulk rock susceptibility were carried out in the field using a hand-held magnetometer. Sampling of the volcanic units including the Nyukawa PFD and Ebistuge PD was carried out during one day in the Takayama basin.

The field campaign was carried out in collaboration with Professor Simon Wallis of the University of Tokyo in Japan, who organised sampling permits, and Professor Satoru Harayama of Shinshu University in Japan, who has guided us to key location

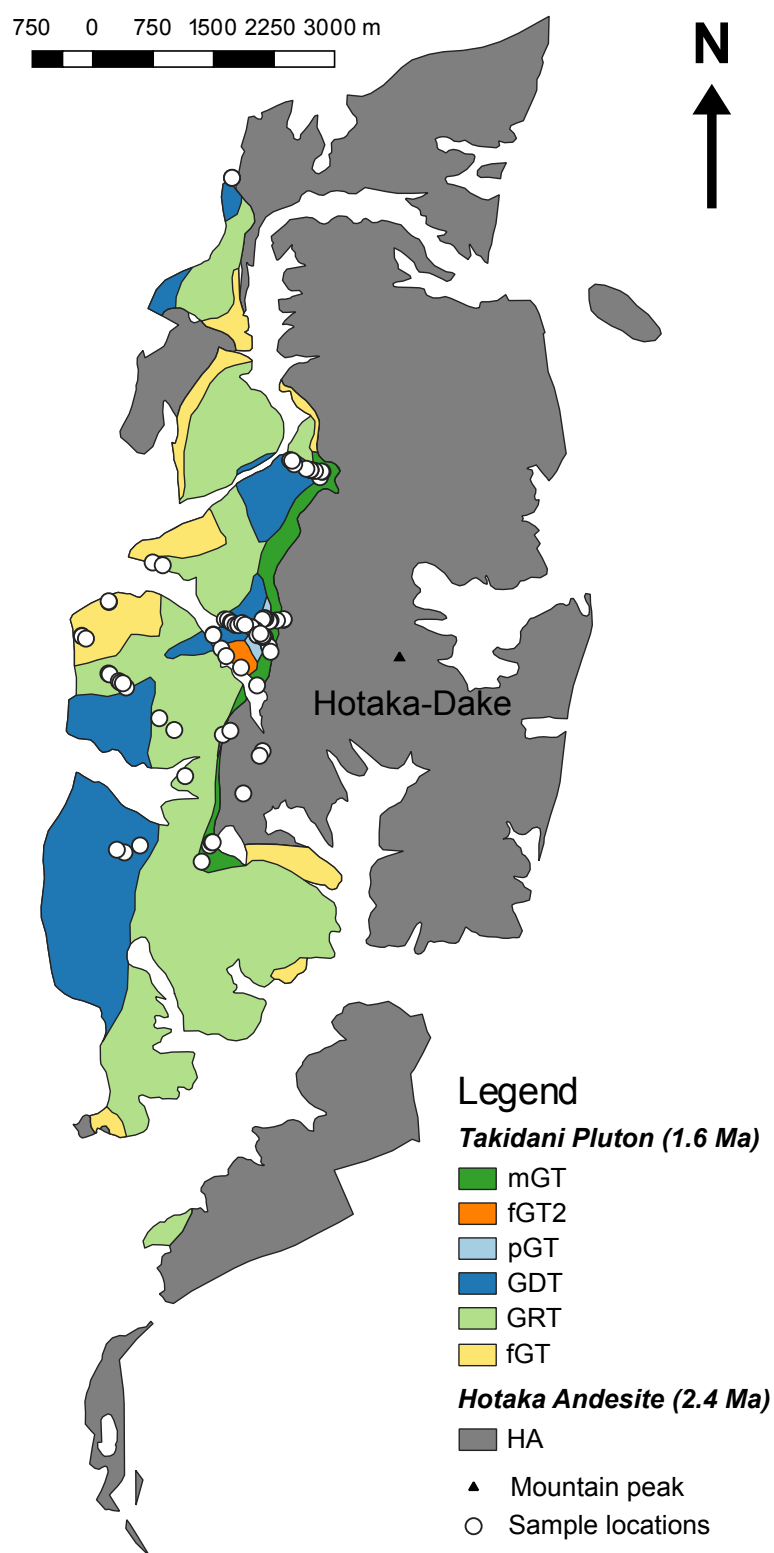


FIGURE 1.5: Lithological units and sample locations of the Takidani Pluton in the Northern Japan Alps; mGT – marginal granodiorite; fGT2 – fine-grained porphyritic granite; pGT – porphyritic granodiorite; GDT – equigranular Hbl-granodiorite; GRT – granodiorite; fGT – biotite granite; HA – Hotaka Andesite.

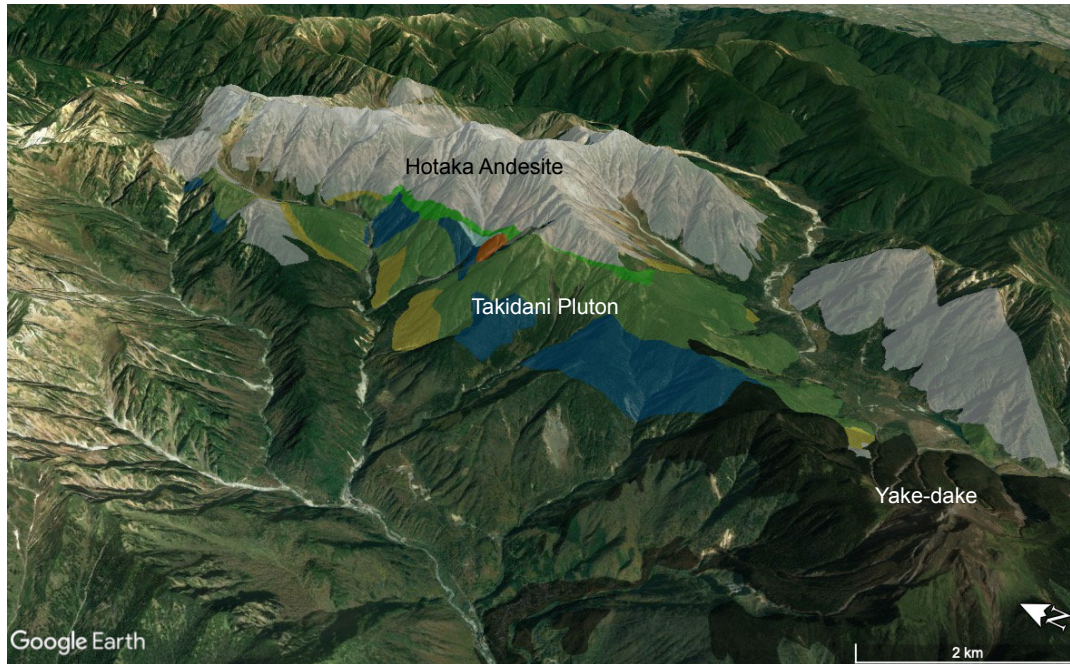


FIGURE 1.6: Overlay of the geological map of Figure 1.5 on Google Earth.

in the field area and shared his knowledge of the regional geology from the last 30 years. Dr David Floess was involved during the initial stages of this project, assisting with field work and rock sampling.

1.5 Organisation of the thesis

The PhD thesis is divided into six main chapters. A comprehensive appendix is included at the end of this thesis that includes additional data that was acquired during the period of this PhD but is not discussed here. Chapters 3 and 4 present manuscripts that are already published or currently under review in peer-reviewed journals.

Chapter 2 provides a general overview of the different methods used in the course of this PhD project. In addition to the general introduction of methods, Chapters 3 to 5 are arranged as scientific articles and thus provide the instrument specifications and the analytical conditions, as well as information on data processing.

Chapter 3 is a published article in the *Journal of Petrology* (Hartung et al. 2017) and provides a detailed petrological study of the upper section of the Takidani Pluton. The results presented in this study suggest that the textural and chemical variations observed in the upper section of the Takidani Pluton are the result of melt extraction from a rheologically locked crystal framework, which is commonly associated with

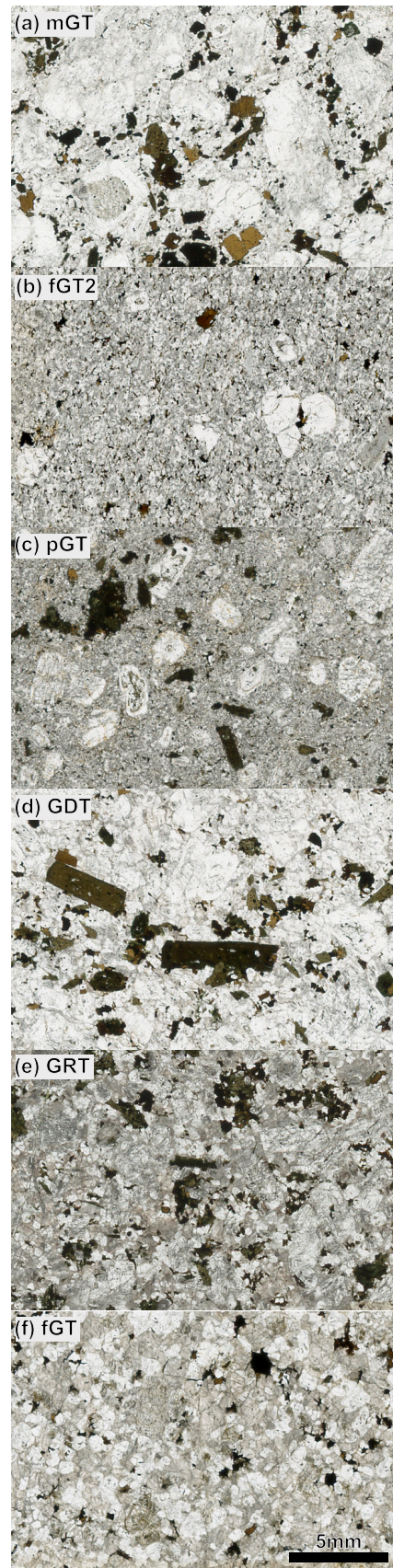


FIGURE 1.7: Lithological units of the Takidani Pluton from the roof to the base of the pluton. (a) phenocryst supported porphyritic marginal granodiorite (mGT), (b) fine-grained porphyritic granite (fGT2), (c) matrix supported porphyritic granodiorite (pGT), (d) medium-grained equigranular Hbl-Bt granodiorite (GDT), (e) equigranular granodiorite (GRT), (f) fine-grained biotite granite (fGT).

the generation of crystal-poor rhyolite. This chapter provides the foundation for Chapters 4 and 5.

Chapter 4 is currently under review in *Earth and Planetary Science Letters*. This chapter investigates the rheological and temporal evolution of crystallising hydrous (i.e. ≥ 1 wt.% H₂O) dacitic magma by combining thermodynamic calculations (i.e. rhyolite-MELTS; Gualda et al. 2012) and thermal modelling. This study is based on the findings of Chapter 3 and assesses the role of initial H₂O content and latent heat of crystallisation on the efficiency and timescales of melt extraction.

Chapter 5 provides a comprehensive petrological investigation of the main units constituting the Takidani pluton and enclaves they contain, and associated volcanic deposits to establish a deeper understanding of the architecture of the magmatic plumbing system beneath caldera-forming volcanoes. This study investigates whether genetic relationships exist between the volcanic and plutonic rocks by using major and trace element chemistry of whole-rock, glass and mineral analysis. The aim of this chapter is to resolve whether the rhyolites of the Chayano–Ebisutoge PD are extracted melts of the Takidani magmatic system.

Chapter 6 provides general conclusions and an outlook of future research to be conducted and describes how the already collected structural data will be integrated in this study.

Chapter 2

Analytical methods

2.1 Overview

This chapter describes the different analytical methods used throughout this project. Each section provides an overview of the basis of the method and method specific sample preparation. Chapters 3, 4 and 5 record instrument specifications, analytical conditions and data processing protocols. Measurement parameters are recorded in Appendix A.

2.2 X-ray Fluorescence (XRF)

X-ray Fluorescence (XRF) is a non-destructive analytical technique that is used to determine major and trace element compositions of rock samples by measuring secondary X-ray emission. The analytical technique is widely used and described in detail by several authors (Jenkins 1995, 2000, Beckhoff et al. 2007)

In general, XRF is a secondary X-ray emission that is generated when a specimen is bombarded with a primary X-ray beam (or gamma beam). The wavelength of the secondary X-ray is separated by Bragg diffraction from a single crystal and detected with a wavelength-dispersive spectrometer (WDS). The wavelength of the secondary X-ray is characteristic for each element. The intensity of the fluorescent X-rays relates to the concentration of the abundant element and is detected with an energy-dispersive spectrometer (EDS) detector. XRF analysis are relative measurements that require conversions using suitable mathematical calibrations and correction formulae to account for matrix effects.

The primary X-ray beam is produced in a X-ray tube when electrons (or other high-energy charged particles) are first emitted from a tungsten filament, lanthanum hexaboride crystal cathode or field emission electron source and then bundled and accelerated by a positively biased anode onto a metal target (i.e. rhodium). The spectrum of primary X-rays depends on the energy and monochromatisation (Jenkins 1995) of the electron beam and on the atomic number and crystalline structure of the scattering material.

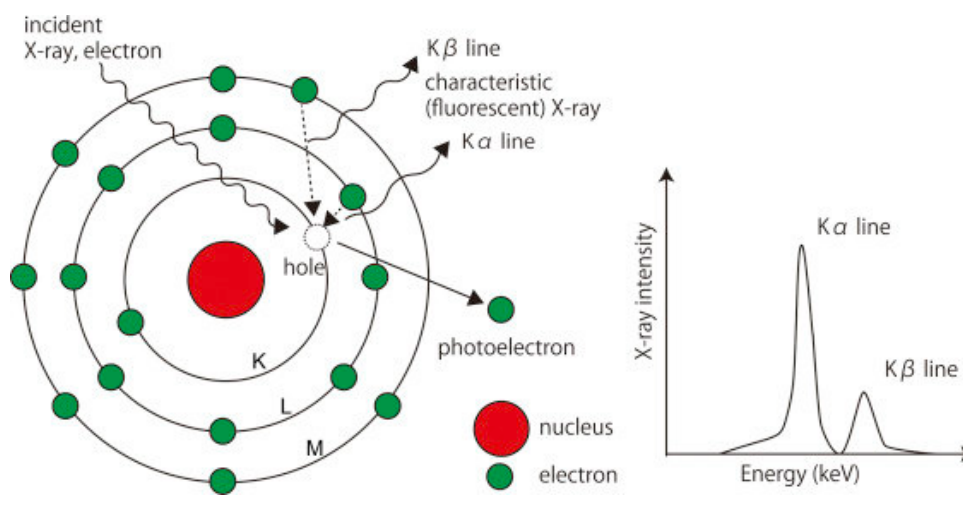


FIGURE 2.1: The mechanism of characteristic X-ray emission from Uo et al. (2015)

2.2.1 Characteristic X-rays

Primary or characteristic X-rays (Röntgen 1895) are generated when orbital electrons of atoms move from an outer to an inner atomic shell. This transition takes place when an electron from an inner shell (i.e. K shell) is ejected by electron bombardment or photon absorption (Figure 2.1). As a result, the atom becomes ionised (or charged) and the vacant space in the atomic shell is filled with an electron from an outer shell, which itself is replaced by an electron from outer orbits or free electrons. X-ray emission has characteristic energy levels for each element and possible electron exchange. In addition to the emission of primary X-rays, continuous X-ray radiation is produced.

2.2.2 Preparation and analysis

Fused glass beads are used for major element analysis. Samples are first prepared into homogeneous, fine-grained powders. Rocks are cut into small chips using a diamond coated saw and then crushed into sand to fine gravel using a hydraulic press at the University of Geneva. Subsequently, samples are ground into fine powders using an agate grinding mill. Each sample is processed for 20 min to ensure grain size of less than 10 μm . To minimise contamination, each equipment is cleaned with water and ethanol after each step. Powders are dried at 60°C for 48 h to ensure loss of volatiles. Finally, 1.2 g of rock powder is mixed with 6 g of lithium tetraborate and melted at 1150°C in a rotating platinum stage and subsequently quenched into glass beads.

2.3 Scanning Electron Microscope (SEM)

The scanning electron microscope (SEM) uses a focused electron beam to analyse the topography and composition (i.e. chemical zoning) of materials. A range of

signals and images is produced by the interaction of the electron beam with atoms at various depths of the sample material. The SEM is commonly equipped with EDS detectors to analyse the emission of secondary electrons, back-scattered electrons, cathodoluminescence and characteristic X-rays (Figure 2.2).

Secondary electrons are emitted by atoms that are excited near the surface region of a sample (Figure 2.2). They are commonly used for high-resolution imaging of the surface topography and morphology. The brightness of secondary electron images corresponds to the amount of emitted electrons, which is highest at steep surfaces and edges (i.e. edge-effect). Secondary electron images have a three-dimensional appearance.

Back-scattered electrons (BSE) are high-energy electrons that scatter and escape from the sample. Heavy atoms are stronger scattered than light atoms. BSE are used to obtain information on chemical differences of the material.

Cathodoluminescence (CL) is the emission of light from a material when atoms near the surface are excited by electrons. CL is commonly used to image trace element zoning in minerals (i.e. quartz, zircon, apatite) and synthetic materials.

2.4 Quantitative Evaluation of Minerals by Scanning Electron Microscopy (QEMSCAN)

QEMSCAN is an automated solution in quantitative mineral analysis of natural rocks and synthetic materials produced by FEI. It comprises a SEM with energy-dispersive X-ray spectrometers and a software for automated analysis. The surface of a material is mapped with an electron beam using pre-defined raster scan. The software uses the X-ray count rates together with BSE images to identify and quantify abundant mineral phases over the scanned surface area.

2.5 Electron Probe Micro Analyser (EPMA)

The electron probe micro-analyser (EPMA) is a non-destructive method used for *in-situ* chemical analysis of solid material (i.e. mineral and glass analyses). The analytical technique is related to the electron beam technique described above. However, instead of measuring secondary X-ray emission, the EPMA detects the wavelengths of primary X-rays that are emitted from the sample during electron bombardment using WDS detectors.

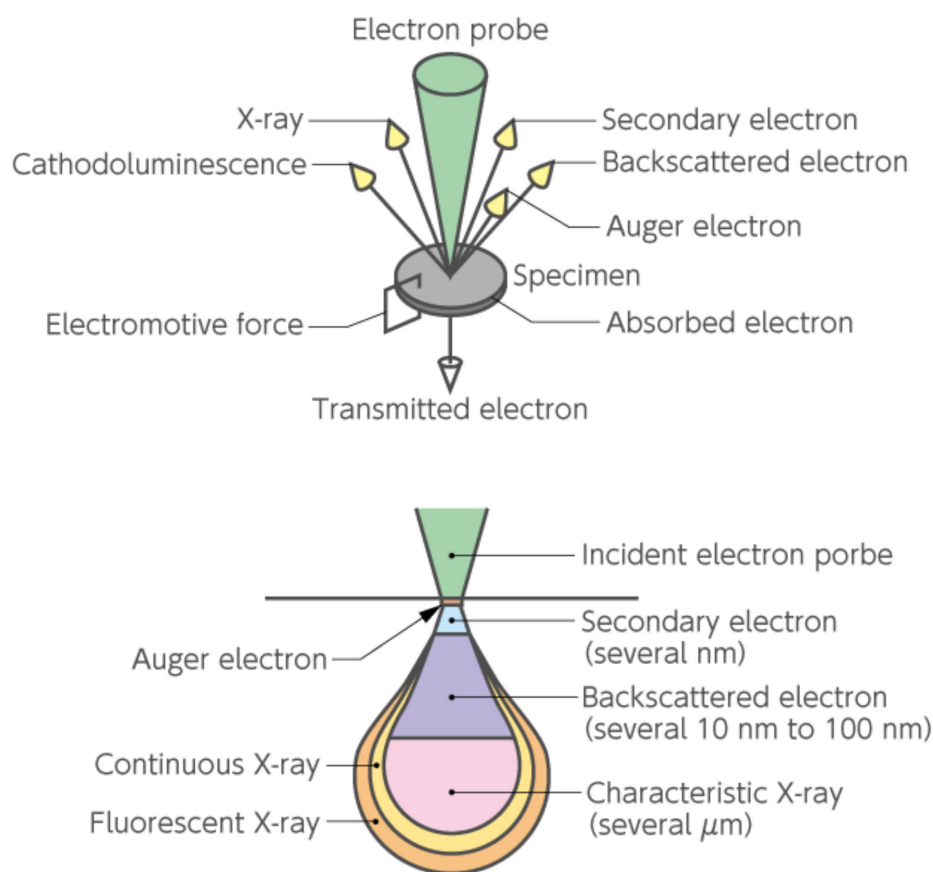


FIGURE 2.2: Signals produced from the interaction of the focussed electron beam with atoms at various depths of the material include secondary electrons, back-scattered electrons, cathodoluminescence, characteristic X-rays (from JEOL)

2.5.1 Preparation and analysis

Mineral and glass analysis are performed on $>30\ \mu\text{m}$ thick uncovered polished thin sections or epoxy mounts. A thin carbon coat is applied for conductivity. Light elements with a atomic number below 4 cannot be detected by the EPMA (i.e. H, He, Li). A calibration of the different elements is necessary using classified standards of known composition. Some elements produce X-rays with overlapping peak position, in which cases peak corrections need to be performed (e.g. REE). Accelerating voltage, electron beam and beam diameter need to be adjusted depending on the analysed mineral phase. Cation proportions and mineral formulae need to be calculated after chemical analysis. Ferric-ferrous ratios cannot be determined by EPMA and require evaluation by other means. Thin sections and crystal mounts were prepared by Jean-Marie Boccard at the University of Geneva.

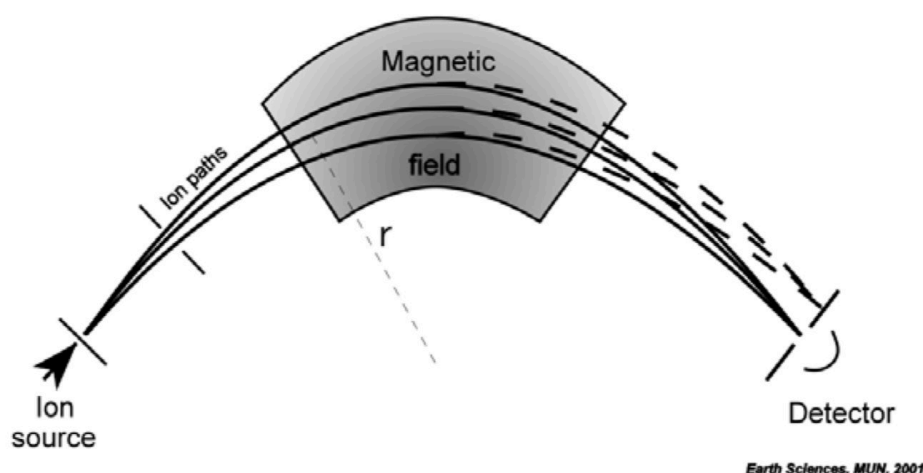


FIGURE 2.3: Schematic illustration of mass spectrometry (from Long-gerich 2008)

2.6 Inductively Coupled Plasma Mass Spectrometry (ICPMS)

Mass spectrometry (MS) is a micro-analytical technique that determines the elemental and isotopic composition of solids, liquids or gases. The sample is bombarded with an electron or ion beam. Positively charged ions are ejected from the sample and channelled and accelerated by an electric field to negative charged plates. A magnetic field is used to deflect and separate accelerated ions based on their masses. Ions of the same mass will undergo the same amount of deflection. Ions of a particular mass-to-charge ratio are detected either by electron multipliers or Faraday collectors. Signal are converted into concentrations through calibration with internal and external reference materials.

Inductively-coupled plasma mass spectrometry (ICPMS) combines a high-temperature ionisation source (i.e. plasma) with a mass spectrometer. ICPMS is commonly used in geochemistry for trace element detection and to measure relative abundance of different isotopes (i.e. U-Pb; Sylvester 2008). A sample is introduced into an Argon plasma torch that runs at atmospheric pressures. The plasma partly or completely vaporises and ionises the sample. A quadrupole mass filter is applied to extract the ions of the plasma into a mass spectrometer, which further separates and quantifies the cations under vacuum.

The laser ablation (LA) technique is used to vaporise and analyse the surface of solid materials. The sample surface is irradiated with a focused laser beam inside the ablation chamber. The ablated material is introduced as vapour (or plasma) into the argon flow of the ICPMS.

2.7 Anisotropy of Magnetic Susceptibility (AMS)

Analysis of Anisotropy of Magnetic Susceptibility (AMS) is a rapid method to determine the magnetic fabric of rocks. The physical and theoretical background of the magnetic anisotropy of rocks is described in Tarling & Hrouda (1993). In general, magnetic susceptibility is a measure for the magnetisation of a material when an external magnetic field is applied. The total (bulk) susceptibility of a material reflects the amount of magnetic substances present. The anisotropy of the magnetic susceptibility indicates the orientation of the magnetic substances in space.

The bulk susceptibility of a rock and its anisotropy is determined by the mineral content of the rock sample and the magnetic properties of the minerals. In rock samples two kinds of mineral species are often present: non-magnetic and magnetic minerals. Non-magnetic minerals (diamagnetic and paramagnetic) do not maintain a magnetic field without an externally applied field. Magnetic minerals (ferromagnetic, antiferromagnetic and ferrimagnetic), on the other hand, retain a distinct magnetisation after an external magnetic field is removed. Ferromagnetic minerals have very strong positive susceptibilities. Depending on the size of the ferromagnetic crystal, the magnetisation can either be uniformly distributed throughout the grain (i.e. isotropic susceptibility) or aligned along preferred crystallographic axes in single, double or multi domains (i.e. anisotropic susceptibility; Tarling & Hrouda 1993). If ferromagnetic minerals (e.g. magnetite, hematite) are abundant (>0.1 vol%), their high susceptibilities and anisotropy will dominate the magnetic behaviour of the rock. If little or no magnetic minerals are present, paramagnetic minerals (e.g. olivine, amphibole, pyroxene, biotite, mica) can be used as principal indicators for magnetic fabrics. The magnetic susceptibility of paramagnetic minerals is positive and two to three order higher than the weak and negative susceptibilities of diamagnetic minerals (i.e. quartz, feldspars).

The anisotropic magnetic susceptibility (AMS) is measured on oriented cylindrical specimens using a AGICO Kappabridge. Specimens are analysed on three perpendicular planes. During each plane measurement, the sample is slowly rotated across one axis and analysed on 15 different positions. The bulk susceptibility is measured in the end to compute the susceptibility tensors using the AGICO software. The three principal susceptibilities (K1, K2, K3) are used to calculate the intensity and shape of the magnetic fabric ellipsoid (Figure 2.4).

2.7.1 Sampling and orientation

Taking precise orientation measurements of samples in the field is a prerequisite to determine the directional magnetic properties. Hand samples are marked *in-situ* in the field along two orientations with reference to the ground. After orientation the sample is removed from the outcrop using a hammer and chisel. In the laboratory,

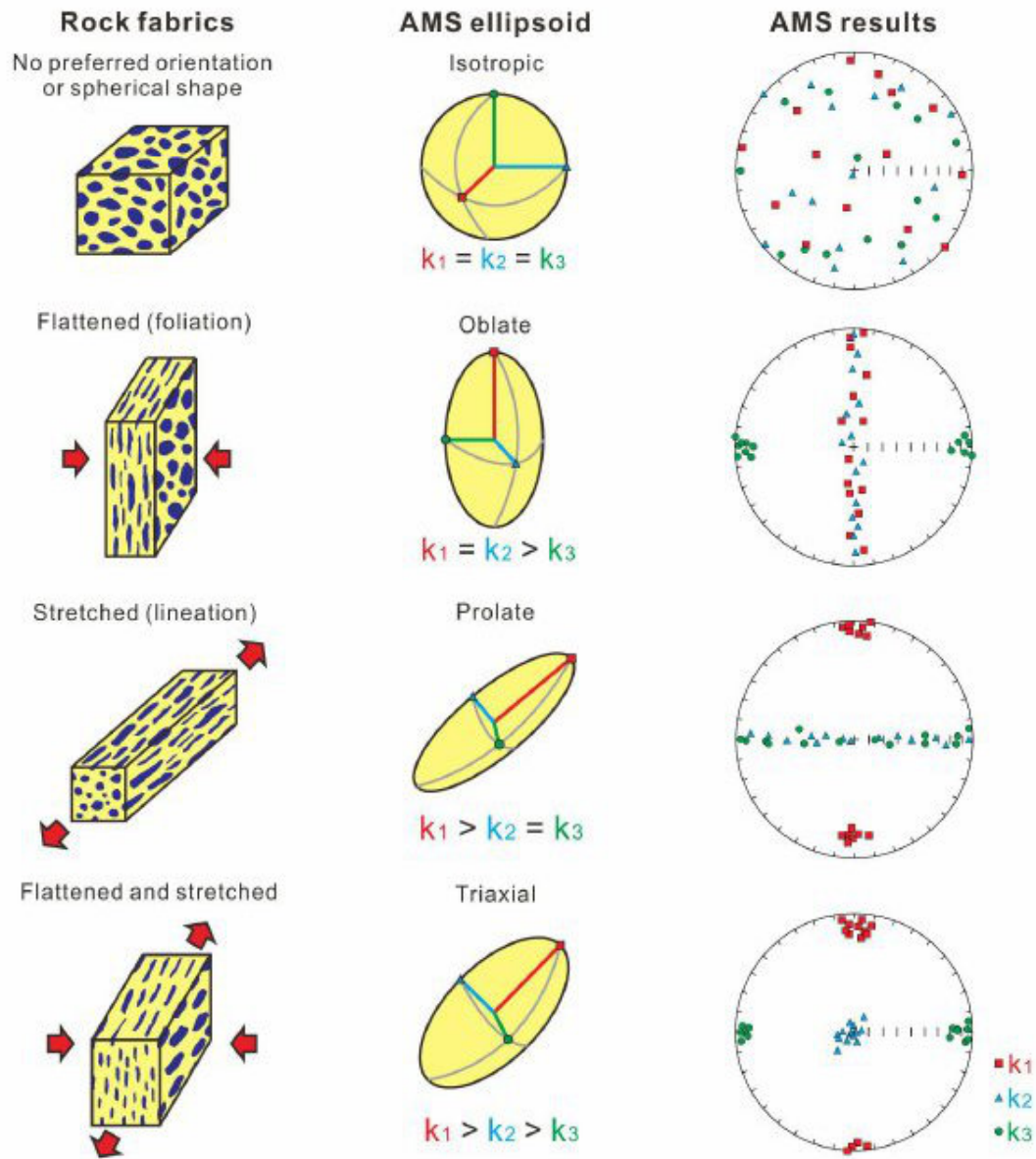


FIGURE 2.4: Principles of AMS method for determining rock fabrics after Tarling & Hrouda (1993). Rock fabrics (left column) that correspond to principles axis of AMS ellipsoids (centre column) and AMS measurements (right column). Modified from Kim et al. (2015).

cylindrical cores with a diameter of 25.4 mm are drilled from each samples. After drilling, cored samples are re-oriented using an orientation table that is build with a slit in the tube to mark the new orientation. Cores are sliced into 22 mm long pieces. Residual metal stains are removed with a polishing machine.

Chapter 3

Evidence for Residual Melt Extraction in the Takidani Pluton

3.1 Abstract

The Takidani pluton represents one of a few locations where melt extraction from a crystal mush is preserved in the natural rock record, making it an extremely good case study for investigating the generation of evolved melt reservoirs in the upper crust. Located in the Japan Alps, the Takidani pluton shows a clear vertical zonation consisting of granite and granodiorite in the lower and middle section, a fine-grained porphyritic granitic unit in the upper section and a marginal granodiorite at the roof contact with the host-rock. We present a detailed petrographic and geochemical study using samples collected along a section that traverses the entire vertical section of the pluton. No sharp contacts are found between units. Instead, gradual changes in rock fabric and mineralogy are observed between the lower granodiorite and overlying porphyritic unit. Major and trace element bulk-rock compositions show sigmoidal variations from the bottom to top of the pluton. Incompatible elements and silica contents increase roofwards within the porphyritic unit. Plagioclase chemistry reveals three main crystal populations (P1, P2 and P3) with Fe contents increasing towards the base of the pluton. Comparison with existing crystallisation experiments, thermobarometry and hygrometry indicate that the magmas were emplaced at around 200 MPa, 850–900°C and bulk water contents of 3–4 wt.%. Whole-rock major and trace element analyses together with mineral chemistry and textural observations suggest that the fine-grained porphyritic unit was extracted from the underlying granodiorite at temperatures between 800 and 740°C and crystallinities of 45–65 wt.%. Radiogenic isotopes indicate only minor assimilation (2–6 wt.%) and support melt evolution through crystal fractionation. The fine-grained matrix of the porphyritic unit may have been the result of pressure quenching associated with a volcanic eruption.

3.2 Introduction

The segregation of residual melt from partially crystallised and rheologically locked magmas (Marsh 1981, Brophy 1991, Vigneresse et al. 1996) is a process that can lead to

the formation of crystal-poor and potentially eruptible pockets of magma in the upper crust (Bachmann & Bergantz 2004, 2008, Hildreth 2004, Dufek & Bachmann 2010, Deering et al. 2011). A variety of petrographic and geochemical observations (Hildreth & Wilson 2007, Barboni & Schoene 2014, Cashman & Giordano 2014, Ellis et al. 2014, Wotzlaw et al. 2014, Christopher et al. 2015) and numerical analysis (Bachmann & Bergantz 2004, Dufek & Bachmann 2010, Solano et al. 2012) have illustrated that melt extraction is a common process in volcanic sequences. Nevertheless, evidence for large-scale melt extraction and accumulation is rarely found in intrusive magma bodies (Coleman et al. 2012). This may reflect the ephemeral nature of such melt-rich magma lenses, as suggested by thermal modelling (Annen 2009, Annen et al. 2015).

Segregated melt may be difficult to recognise in the field if it crystallises at similar rates to the surrounding magma and generates holocrystalline rock textures. The thermal structure and evolution of melt reservoirs, together with phase petrology and especially the relationships between temperature and crystallinity, control the timespan a magma spends at conditions favourable for melt segregation (Marsh 1981, Dufek & Bachmann 2010, Melekhova et al. 2013, Caricchi & Blundy 2015*b*). The low duration of this period results in a relatively low probability of finding evidence for melt segregation in the intrusive record. A final but nonetheless important point is that plutons are the accumulated product of numerous incremental injections of magma (Glazner et al. 2004, de Saint-Blanquat et al. 2006, Menand et al. 2015) and this periodic replenishing of hot magma tends to obliterate earlier fabrics and chemical heterogeneities (Paterson et al. 1998, Zak et al. 2007).

The Takidani pluton (TK) is an extremely good candidate to study the extraction of residual melt from magmatic mushes because of its chemical stratification and the gradual transition from holocrystalline textures at the intrusion's base and centre to porphyritic textures towards the roof contact (Harayama 1992, Bando et al. 2003). Here we present the results of whole-rock major and trace element and isotopic analyses for samples collected along a complete section along the Shiradashi-zawa valley, in combination with the mineral chemistry of various minerals at different levels within the intrusion. The main goal of this study is to investigate the dominant processes responsible for the chemical zoning and evolution in the upper part of the TK.

3.3 The Takidani Pluton

3.3.1 Geology

The pluton is located in the active Norikura Volcanic Chain between Yake Lake and Hotaka Lake (Figure 3.1) and covers an area of $\sim 21 \text{ km}^2$ (Harayama 1992).

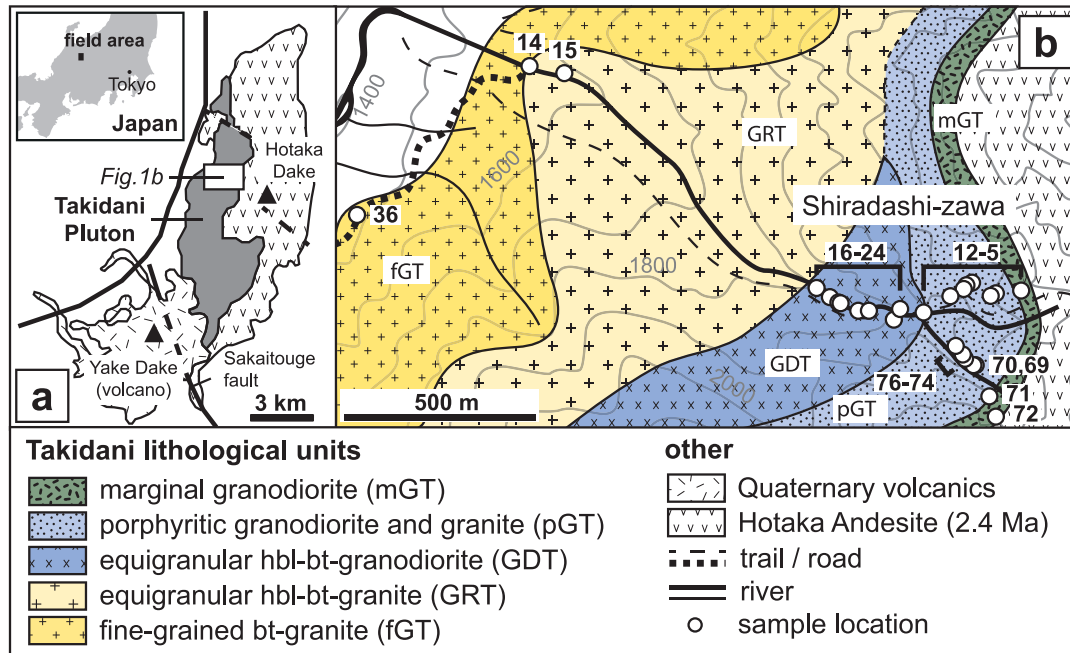


FIGURE 3.1: (a) Location of the Takidani Pluton (TK) in central Honshu, modified after Harayama et al. (2003). (b) Geological map of the Shiradashi-zawa valley including sample locations and topographic contour lines.

Along its western margin, the TK is tectonically juxtaposed with Mesozoic basement and Paleogene granitic rocks. On its eastern margin it borders the Late Pliocene Hotaka Andesite (2.4 Ma; Yamada et al. 1985, Harayama 1990) with a magmatic contact formed during the intrusion of the felsic melts forming the TK. To the south the TK is unconformably overlain by the Quaternary volcanic deposits of Mount Yake. The TK was rapidly uplifted ($5.1\text{--}5.9\text{ mm yr}^{-1}$; Bando et al. 2003) and tilted eastwards resulting in a nearly vertical contact on its western side and a shallow eastwards dipping contact of about 20° on its eastern side (Harayama 1994). The vertical thickness of the outcropping portion of the TK is estimated to be between 1200 and 1800 m (Bando et al. 2003, Harayama et al. 2003).

The pluton is vertically zoned varying from fine-grained biotite (bt)-granite (fGT) and medium-grained hornblende (hbl)-bt-granite (GRT) in the lower section of the pluton, to medium-grained hbl-bt-granodiorite (GDT) in the middle section (Figure 3.1; Harayama 1992, Harayama et al. 2003). In the upper section, GDT marks the transition into fine-grained porphyritic granodiorite and granite (pGT). The marginal unit at the roof of the pluton is a coarse-grained porphyritic granodiorite (mGT; Bando & Tsuchiya 2000, Bando et al. 2003, Harayama et al. 2003). Rocks of the fGT are dominantly found along the basal contact (Bando & Tsuchiya 2000, Bando et al. 2003), but a similar lithology has been identified near the roof contact in the southern part of the pluton (Harayama et al. 2003).

3.3.2 Geochronology

U–Pb zircon isochron ages range between 1.36 and 1.54 Ma (± 0.23 Ma; Sano et al. 2002) whereas zircon fission-track ages (effective closure temperature $240^{\circ}\text{C} \pm 30^{\circ}\text{C}$; Bernet & Garver 2005) yield an age of 0.80 Ma (± 0.02 Ma; Harayama 1990, 1992, 1994). K–Ar ages of hornblende, biotite and alkali feldspar vary from 1.43 Ma (± 0.12 Ma) to 1.94 Ma (± 0.21 Ma), from 1.07 Ma (± 0.04 Ma) to 1.27 Ma (± 0.06 Ma) and from 1.07 Ma (± 0.06 Ma) to 1.23 Ma (± 0.06 Ma), respectively (Harayama 1992, 1994). Most K–Ar mineral ages are consistent with the zircon crystallisation ages. The significance of the older hornblende K–Ar ages is less clear but they may reflect the presence of extraneous ^{40}Ar (Harayama et al. 2003).

Studies constraining the temperature–time path of the TK have suggested a two-stage cooling history with an initial rapid rate of about $150^{\circ}\text{C kyr}^{-1}$ followed by slower cooling rates of about $40^{\circ}\text{C kyr}^{-1}$ (Harayama 1992, 1994, Kano et al. 2000, Tsuchiya & Fujino 2000, Sano et al. 2002). The presence of a magmatic contact with the Hotaka Andesite and the results of thermochronological studies show that the TK magmatic reservoir was emplaced after the Hotaka Andesite (Harayama 1992, 1994), but before eruption of the 1.76 Ma (± 0.17 Ma) Nyukawa Pyroclastic Flow Deposit (PFD; Harayama et al. 1991) and the 10 kyr younger Chayano–Ebisutoge Pyroclastic Deposits (PD; Nagahashi 1995, Nagahashi et al. 1996, 2000, Kataoka et al. 2001, Kimura & Nagahashi 2007).

Previous work has predominantly focused on the emplacement and cooling history of the TK. In this study, we focus on the magmatic processes that can be identified based on petrographic, geochemical and structural investigation of the TK. This pluton offers an extremely rare opportunity to examine the petrogenetic processes within a young and well-exposed zoned crustal magmatic body.

3.4 Sampling strategy and field observations

The TK lies within a rugged mountain terrain in the Chubu Sangaku National Park. It is exposed at elevations between about 1400 and 2500 m, especially along steep valleys (i.e. Shiradashi-zawa, Yanagani-dani, Takidani). In many places, large parts of the outcrops are covered under scree fields. Mountain slopes are generally about 40° and change into nearly vertical rock faces at higher elevations.

During our field campaign in 2014 we collected a total of 101 samples along four NW–SE-trending river valleys in the northern section of the TK and along mountain ridges in the southern part of the intrusion, to characterise the lithofacies and variability observed in rock textures and mineralogy from the base (west) to the roof (east) of the intrusion. The exposure along the Shiradashi-zawa valley provides a complete cross-section of the entire pluton where a gradual transition from the

equigranular to fine-grained porphyritic units (i.e. GDT to pGT) is observed towards the roof (Figure 3.1). In this study we present petrographic and geochemical data for 25 rock samples collected along or close to (e.g. location 36; Figure 3.1) the Shiradashi-zawa valley. This cross-section represents not only the most representative transect across the TK but is also critical to understanding the processes responsible for the transitions in rock fabric and mineralogy seen from GDT to pGT in the upper part of the intrusion.

3.5 Analytical Methods

3.5.1 Whole-rock analyses

Bulk-rock major and trace element analyses were performed for 25 samples collected along the Shiradashi-zawa valley (Figure 3.1; Table ??, ??). Neodymium, Sr and Pb isotopes were analysed for 12 of these samples. Major elements were determined by X-ray fluorescence (XRF) using a PANalytical AXIOS MAX with a rhodium anode tube at 4 W at the University of Lausanne. SY-2, JCH-1, NIM-N and NIM-G standards were used for quality control. Trace element concentrations were measured using glass beads by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) using a Quadrupole Agilent 7700 spectrometer interfaced to a GeoLas 200M 193 nm excimer ablation system at the University of Lausanne. The system was operating at 1420 W with a laser repetition rate of 10 Hz and a 120 μm spot size. Each sample was measured three times for 60 s. The background was measured for 90 s before the three ablations. Trace element concentrations were recalculated using CaO contents from XRF analyses as an internal standard with reference to the NIST SRM 612 external standard. Data reduction was performed using the LAMTrace software (Jackson et al. 1992, Longrich et al. 1996, Jackson 2008a,b).

Radiogenic isotopes of Sr, Nd and Pb were analysed at the University of Geneva using a Thermo Neptune PLUS Multi-Collector ICPMS system following the methods described by (Chiaradia et al. 2011). Isotopic ratio were corrected for internal fractionation using $^{88}\text{Sr}/^{86}\text{Sr} = 8.375209$ for the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ for the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio and $^{203}\text{Tl}/^{205}\text{Tl} = 0.418922$ for the three Pb ratios (a Tl standard was added to the solution). SRM987 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710248$, long-term external reproducibility: 10 ppm), JNdi-1 ($^{143}\text{Nd}/^{144}\text{Nd} = 0.512115$; Tanaka et al. 2000; long-term external reproducibility: 10 ppm), and SRM 981 (Pb-isotopes; Baker et al. 2004; long-term external reproducibility of 0.0048% for $^{206}\text{Pb}/^{204}\text{Pb}$, 0.0049% for $^{207}\text{Pb}/^{204}\text{Pb}$ and 0.0062% for $^{208}\text{Pb}/^{204}\text{Pb}$) were used as external standards. Owing to a systematic difference between measured and accepted standard ratios Sr, Nd and Pb isotope ratios were further corrected for external fractionation by a value of -0.039, +0.047 and +0.5 a.m.u., respectively. Mass interferences at 84 (^{84}Kr), 86 (^{86}Kr) and 87 (^{87}Rb) were corrected for by monitoring ^{83}Kr and ^{85}Rb . The interference of ^{144}Sm on ^{144}Nd

was monitored on ^{147}Sm and corrected with a value of 0.206700 ($^{144}\text{Sm}/^{147}\text{Sm}$); the interference of ^{204}Hg on ^{204}Pb was corrected by monitoring ^{202}Hg .

3.5.2 Mineral chemistry

Mineral analyses were performed with a JEOL 8200 Superprobe electron probe microanalyser (EPMA) at the University of Lausanne. The microprobe is equipped with a five-channel wavelength-dispersive spectroscopy system (WDS) and was operated at an accelerating voltage of 15 kV and a beam current of 15 nA. A defocused beam of 2 μm diameter was used for feldspar, amphibole and biotite analyses. Oxides were analysed with a beam diameter of 1 μm . Analyses with a maximum 1 σ error of 1% were retained and additionally checked by calculating the appropriate mineral formula.

Representative mineral analyses of plagioclase and amphibole are listed in Tables ??–??. Alkali feldspar, biotite and oxide data are provided in Tables ??–??. All data and operating conditions are reported in Tables A.1–A.3.

3.5.3 Quantitative evaluation of minerals by scanning electron microscopy (QEMSCAN)

Six thin sections were analysed by automated mineral analysis and textural imaging using an FEI QEMSCAN[®] Quanta 650 F at the University of Geneva. The system is equipped with two Bruker QUANTAX light-element energy-dispersive X-ray spectrometers. Analyses were conducted at high vacuum, accelerating voltage of 25 kV, and a probe current of 10 nA on carbon-coated polished thin sections. Field Image operating mode (Pirrie et al. 2004) was used for analyses. Two-hundred and twenty-one fields were measured per sample, with 1500 mm per field, and a point spacing of 5 mm. Data processing was performed using the iDiscover[®] software package to produce high-quality spatially resolved and fully quantified mineralogical maps. The modal abundances of mineral phases above 0.1 area% are listed in Table ??.

Plagioclase analyses from EPMA sessions were used to assign Anorthite contents to Ca intensities obtained from the EDX spectra during QEMSCAN analyses. To obtain Anorthite distribution maps of plagioclase, other Ca-bearing phases, such as amphibole and apatite, were filtered out from the obtained Ca maps. Amphibole grains were grouped and subtracted from the Ca distribution maps based on Fe intensities. Apatite grains were grouped and removed based on their high Ca intensity. Four plagioclase groups of different composition were then created using bins of 20 mol% Anorthite (i.e. An_{10–30}, An_{30–50}, An_{50–70}, An_{70–90}). The area percentages of plagioclase bins are listed in Table ?? It should be noted that small differences (up to 8%) exist in three samples between the total area percentage of plagioclase from the QEMSCAN maps and plagioclase obtained from Ca-distribution maps.

3.6 Petrography

3.6.1 Fine-grained bt-granite (fGT)

This unit crops out at the basal contact of the pluton and is markedly fine-grained (<1 mm; Figure 3.2). Plagioclase is normally zoned and occasionally contains resorbed cores. Quartz shows undulose extinction and a mosaic texture. Alkali feldspar is interstitial and perthitic. Amphibole is entirely altered to biotite and iron oxides. Zircon is usually found within plagioclase and quartz; apatite is commonly included in quartz. Overall, the alteration of fGT is more profound with higher degrees of chloritisation and oxidation.

3.6.2 Equigranular hbl–bt-granite (GRT)

GRT has a characteristic medium-grained equigranular texture (<5 mm; Figure 3.2). Plagioclase crystals (<5 mm) are oscillatory zoned and surrounded by darker rims. Some crystals contain resorbed or patchy zoned cores. Amphibole (<2 mm) forms euhedral to subhedral grains that locally include euhedral plagioclase or magnetite. In the lower section of GRT, amphibole is almost entirely replaced by biotite and iron oxides, similar to fGT. Biotite (<1 mm) occurs as brown to pale brown flakes and becomes reddish brown to pale brown towards the base of GRT with increased alteration to chlorite and iron oxides. Quartz and alkali feldspar are typically subhedral to anhedral with undulose extinction and fine perthitic exsolution, respectively. Zircon commonly appears as inclusions in biotite and quartz.

3.6.3 Equigranular hbl–bt-granodiorite (GDT)

The basal contact between GDT and GRT is covered by boulders and blocks and is not visible in the Shiradashi-zawa valley. The top contact with pGT can be followed in the field and is not sharp but gradual. GDT is equigranular, medium- to coarse-grained (<10 mm; Figure 3.2) and has similar textural characteristics throughout the TK. GDT contains plagioclase (43–49 area%), quartz (21–26 area%), alkali feldspar (14–15 area%), biotite (7–8 area%) and amphibole (5–6 area%; Table ??). Plagioclase (<10 mm) is typically euhedral and contains either a resorbed calcic or patchy zoned core, an oscillatory-zoned mantle and a darker rim. Rarely, plagioclase has overgrown partly resorbed mineral aggregates. Amphibole (<4 mm) is euhedral to subhedral, often patchy zoned and may contain various mineral inclusions (i.e. iron oxides, biotite, apatite, amphibole). Biotite (<1 mm) occurs as brown to pale brown subhedral flakes. In more altered samples, biotite is dark brown in plane-polarised light. In the lower section of GDT quartz shows undulose extinction and graphic intergrowth with alkali feldspar. Apatite and zircon are mainly included in biotite and amphibole crystals. Vermicular chlorite is sporadically present and is usually associated with quartz crystals. Allanite was found in two samples.

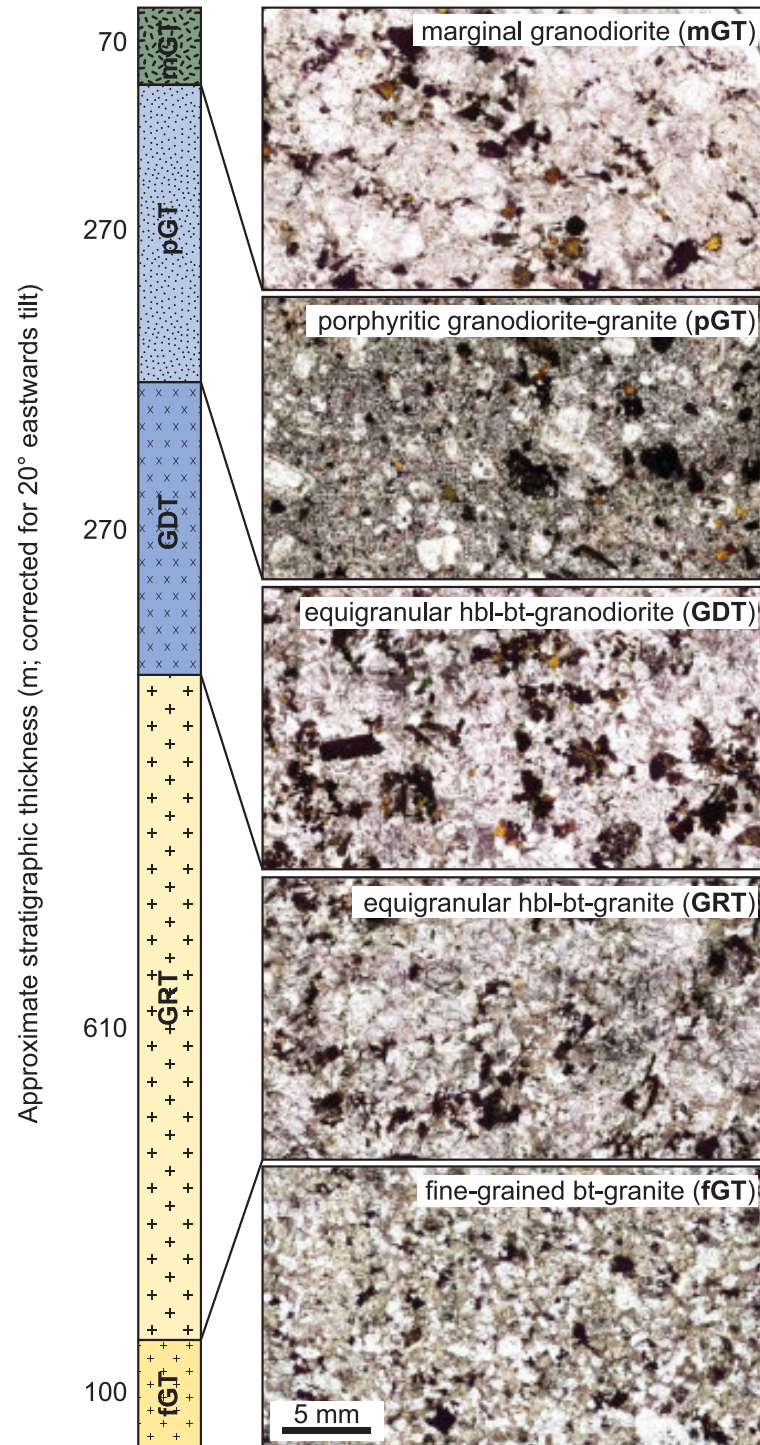


FIGURE 3.2: Rock textures of five TK units as observed in thin section, indicating the corresponding stratigraphic locations. The basal granite (fGT) has a fine-grained equigranular texture, whereas the overlying granite (GRT) and granodiorite (GDT) are medium-grained. The porphyritic granodiorite and granite (pGT) in the upper section of the TK is distinctly porphyritic and dominated by a fine-grained groundmass. The marginal granodiorite (mGT) at the roof contact is also porphyritic, but phenocryst-dominated. Stratigraphic heights have been corrected for pluton tilting and show the estimated thickness of the various units along the Shiradashi-zawa valley.

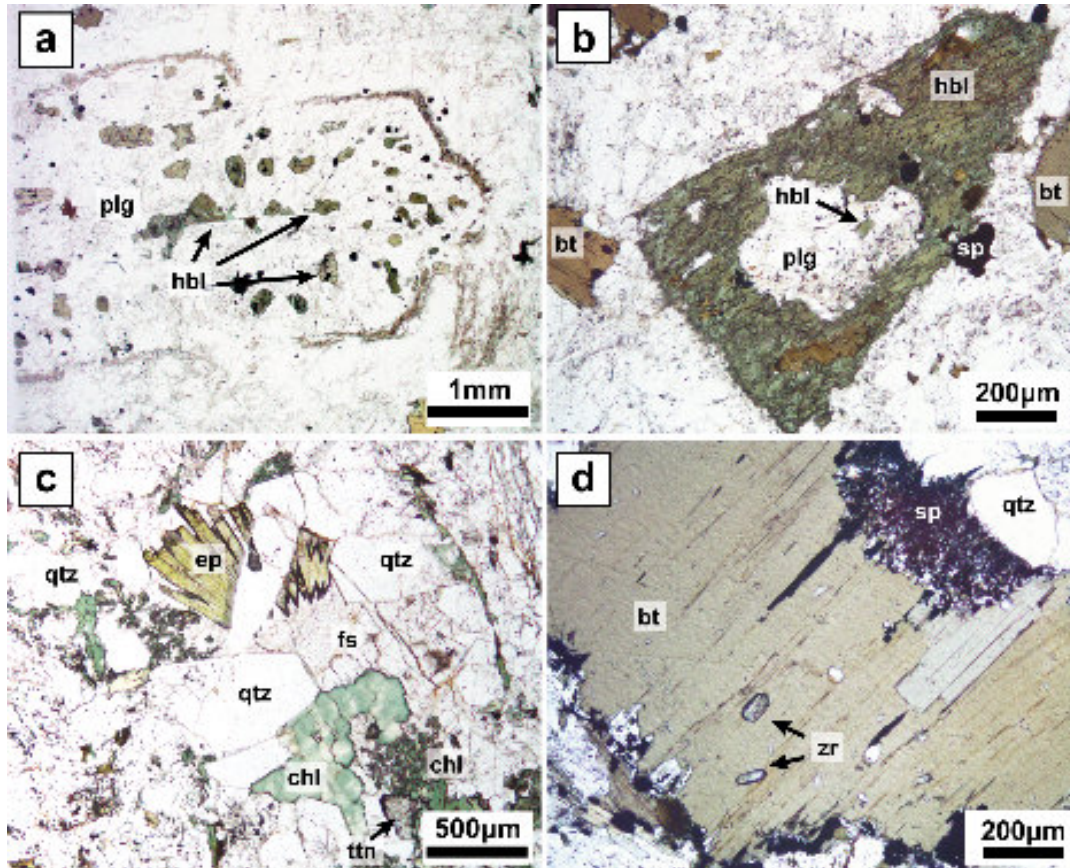


FIGURE 3.3: Petrography of the TK rocks: (a) rare plagioclase phenocrysts containing numerous mineral inclusions including amphibole; (b) resorbed plagioclase inclusion in amphibole phenocryst in pGT; (c) miarolitic cavities in pGT containing quartz, alkali feldspar, epidote, and spherulitic and vermicular chlorite; (d) crystallisation of Fe-Ti oxides along biotite rims in mGT.

3.6.4 Porphyritic granodiorite and granite (pGT)

This unit lies structurally between GDT and mGT and is distinctly porphyritic with a predominantly fine-grained matrix (<1 mm) and plagioclase and amphibole phenocrysts (<8 mm, Figure 3.2). The modal proportions of the main components vary throughout the unit with the amount of plagioclase decreasing from about 43 to 37 area% upwards (Table ??). Concurrently, quartz and alkali feldspar abundances increase from about 27 to 31 area% and from 16 to 22 area%, respectively. Amphibole (3–4 area%), biotite (6–8 area%), magnetite and apatite decrease towards the upper part of the pGT (Table ??). The matrix mainly comprises alkali feldspar, quartz and plagioclase, becoming very fine-grained (20–50 μm) towards the upper contact with the mGT.

Euhedral plagioclase phenocrysts also contain resorbed or patchy zoned cores and an oscillatory-zoned mantle. Overgrowth rims become more volumetric towards the base of the pGT. Occasionally, plagioclase cores contain large amounts of mineral

inclusions (Figure 3.3a). Quartz occurs as 1–2 mm long euhedral to subhedral grains and as small crystals in the groundmass, whereas alkali feldspar is abundant only in the groundmass. Subhedral biotite (<1 mm) and sub- to euhedral amphibole (<4 mm) crystals commonly incorporate iron oxides and in the case of amphibole also biotite. Sometimes, amphibole contains resorbed plagioclase (Figure 3.3b). Zircon and apatite are present as accessory phases included in biotite and amphibole and within the groundmass. Micrographic intergrowth of quartz and alkali feldspar and skeletal growth of biotite and amphibole is observed near the contact with mGT. Some biotites in samples close to mGT are rimmed by iron oxides. Vermicular chlorite and epidote are observed within a miarolitic texture (Figure 3.3c). Traces of orthopyroxene are included in amphibole cores.

3.6.5 Marginal granodiorite (mGT)

The marginal granodiorite is located at the roof of the intrusion and it is porphyritic with predominantly plagioclase phenocrysts (<10 mm) and a fine- to medium-grained matrix (<5 mm; Figure 3.2). The main components are plagioclase (47 area%), quartz (24 area%), alkali feldspar (16 area%) as well as amphibole (6 area%) and biotite (6 area%; Table ??). Plagioclase phenocrysts (<10 mm) are oscillatory zoned and overgrown by a rim of different composition. Few crystals contain resorbed cores. Quartz shows weak to strong undulose extinction and occurs as large subhedral grains (<5 mm) and as small interstitial grains (<1 mm) in the matrix. Alkali feldspar occurs only in the matrix and perthites are visible in only a few cases. Amphiboles (<3 mm) are subhedral to euhedral, predominantly patchy zoned and occasionally occur as oikocrysts including biotite, iron oxides and apatite. Biotite crystals (<2 mm) occur in brown to pale brown subhedral flakes and show alteration to chlorite and iron oxides. Many biotite grains are surrounded by small iron oxides (Figure 3.3d). Apatite is common and is mainly included in biotite and amphibole. Zircon is predominantly found as inclusions in biotite (Figure 3.3d), but also occurs in amphibole and quartz. Titanite and rutile are present in relatively small amounts (<0.2 area%; Table ??).

3.7 Bulk-rock geochemistry

3.7.1 Major and trace elements

The rocks of the TK are calc-alkaline, high in potassium and predominantly metaluminous (Figure 3.4). GDT represents the least evolved unit of the TK with a silica content of 65–67 wt.% SiO₂ (anhydrous). GRT and mGT are slightly more evolved (67 and 68 wt.% SiO₂). The pGT unit has a more evolved composition with up to 71 wt.% SiO₂ in the Shiradashi-zawa valley. fGT is the most silicic unit of the TK (73 wt.% SiO₂).

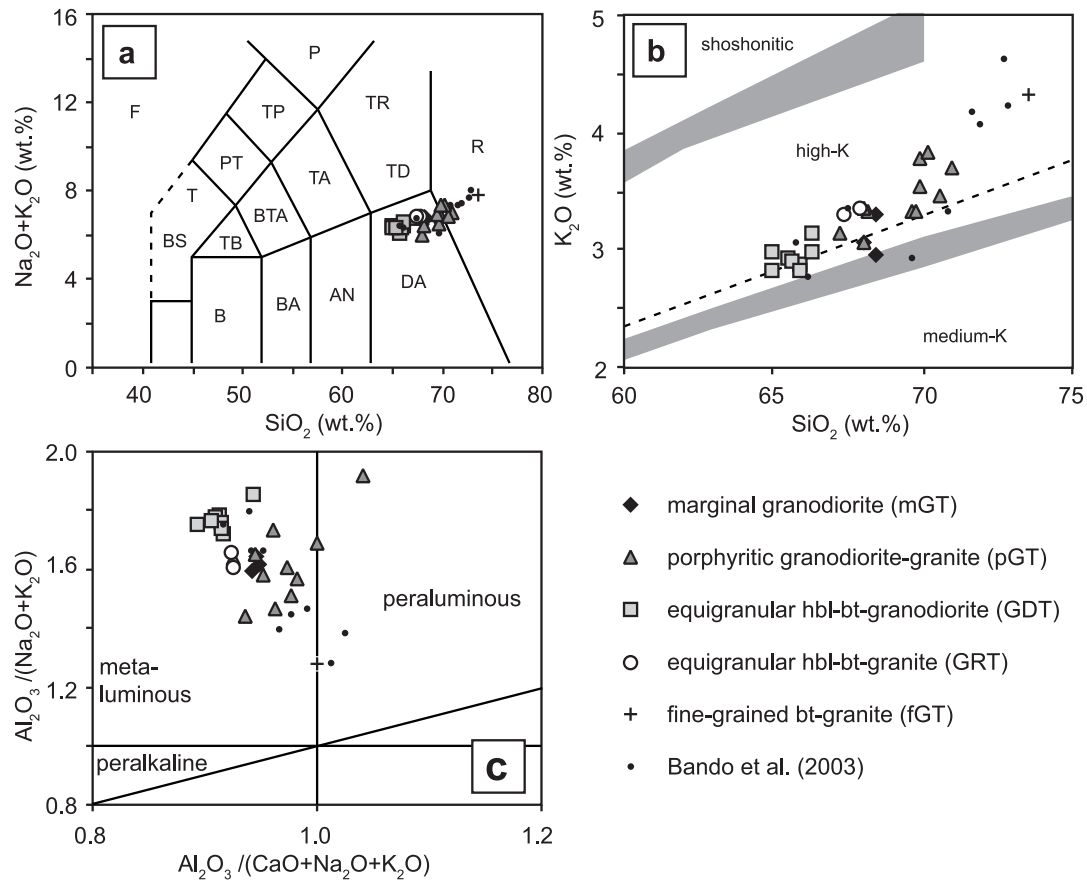


FIGURE 3.4: Chemical classification of samples from the TK after Le Maitre (1989). The rocks from Shiradashi-zawa are plotted together with the rock compositions of Bando et al. (2003). The TK samples are predominantly calc-alkaline (a), high-K (b), metaluminous rocks (c) of dacitic (i.e. granodioritic) to rhyolitic (i.e. granitic) composition.

Incompatible elements such as K_2O noticeably increase with SiO_2 content (Figure 3.4b). Striking, however, is the sigmoidal variation of major and trace element concentrations from the bottom to the roof of the intrusion (Figure 3.5; Tables ??–??). The concentration of SiO_2 , K_2O , Rb, and Th first decrease above the base, up to about 1800 m then increase to about 2050 m, and finally decrease towards the roof of the pluton at an elevation of about 2200 m (Figure 3.5). All other major and minor elements show opposite trends. It is noticeable that the concentrations of all elements are similar for GRT and mGT units.

3.7.2 Isotope geochemistry

All isotope values are corrected for fractionation and an age of 1.5 Ma (Sano et al. 2002). The TK shows an isotopic signature characteristic for the Hida Mountain Granites with $^{87}Sr/^{86}Sr$ of 0.707061–0.707517 and $^{143}Nd/^{144}Nd$ of 0.512583–0.512462 (Figure 3.6a; Tables ??–??). The Sr-isotope ratios of the TK generally increase towards the roof, similar to observations from the Bushveld intrusion (Sharpe 1985). The Sr, Nd and Pb isotopic compositions of GDT and pGT plot along continuous trends between the more primitive GRT and more radiogenic mGT (Figure 3.6a,b; Tables ??–??). A more substantial spread is observed in the $^{206}Pb/^{204}Pb$ (18.391–18.404) and $^{207}Pb/^{204}Pb$ (15.613–15.625) isotopic ratios.

3.8 Mineral Chemistry

3.8.1 Plagioclase

Plagioclase is the main mineral phase of the TK. Most phenocrysts contain resorbed cores of different size, texture and chemistry (Figures 3.7–3.9; Table ??), with an oscillatory-zoned mantle of andesine composition (An_{30-40}), and sodic rims of different thickness, with Anorthite contents decreasing down to about 20 mol%. Anorthite and Fe contents of the oscillatory-zoned mantle were used to separate plagioclase into three main populations (Figure 3.7, 3.8). Population 1 (P1) is characterised by reverse zoning with Anorthite content increasing progressively from about 30 to 40 mol% and a rapid drop to 20 mol% in the outer rim. An-rich cores (>60 mol%) are sporadically observed in this population (Figure 3.7a). Iron contents in plagioclase vary only slightly between 0.025 and 0.03 mol% and co-vary with Anorthite content. The Anorthite contents of populations 2 (P2) and 3 (P3) are generally higher than those of P1 and typically vary between 40 and 50 mol% (Figure 3.7b,c). P2 and P3 are separated on the basis of the lower FeO content of the plagioclase in P2 (0.035–0.045 mol%) than in P3 (0.045–0.055 mol%).

QEMSCAN and EPMA analyses (42 traverses) were used to investigate the distribution of plagioclase populations across the different units (Figure 3.9; Tables ??–??). Anorthite distribution maps can be used to distinguish P1 from P2 and P3 (Figure 3.9).

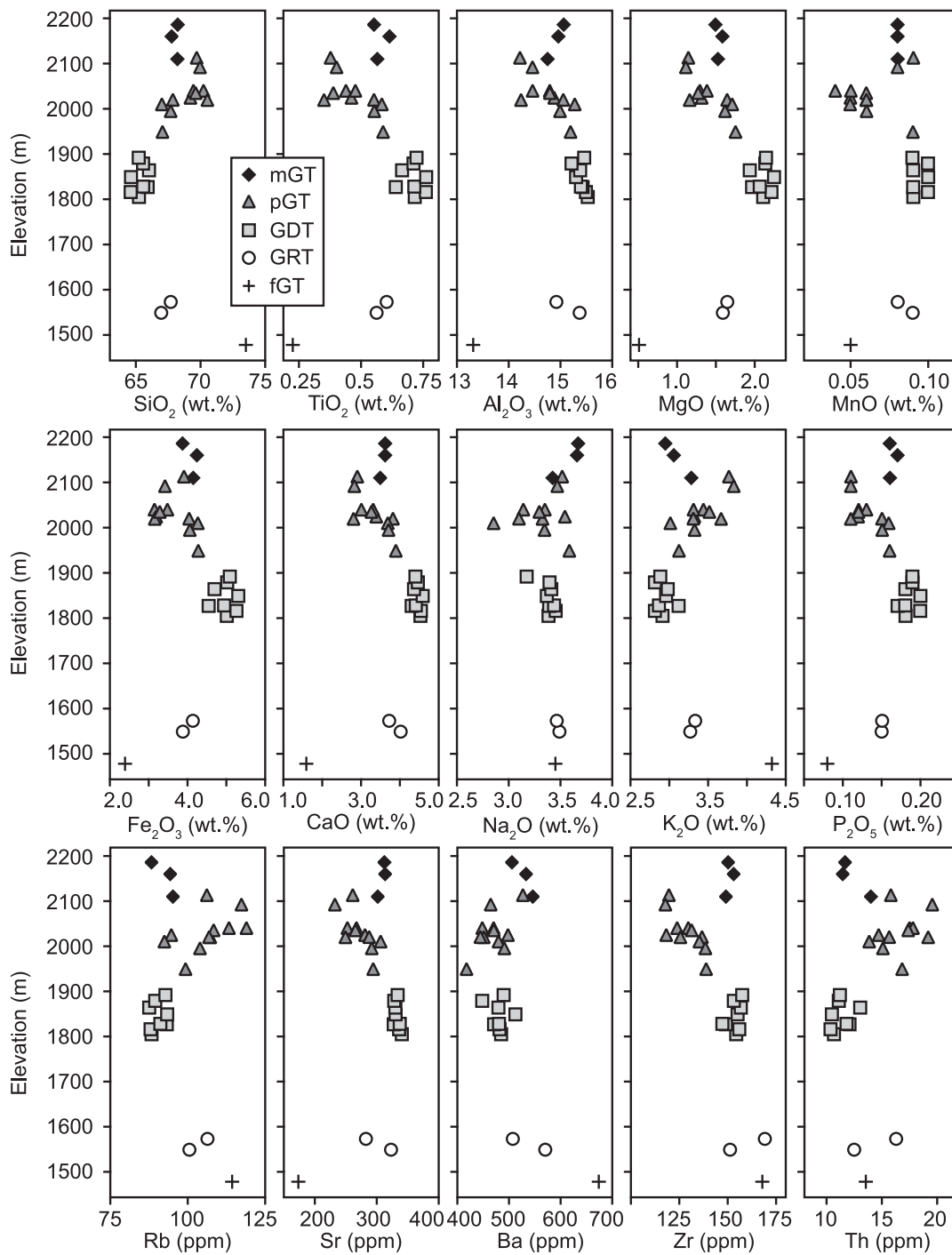


FIGURE 3.5: Chemical variations from the base to the roof of the intrusion along the Shiradashi-zawa valley. Samples from the lower section (below 1900m) were collected along the river. The upper part of this section (above 1900m) was sampled along a hiking trail. Sigmoidal variations are observed from the bottom to the top of the pluton. No outcrops were observed between sample locations 15 (GRT) and 16 (GDT; see Figure 3.1). Sample 36 was collected near the road, south of Shiradashi-zawa valley. Data points are projected onto a bottom-to-top profile against the topographic altitude. No correction for the tilting of the pluton has been applied.

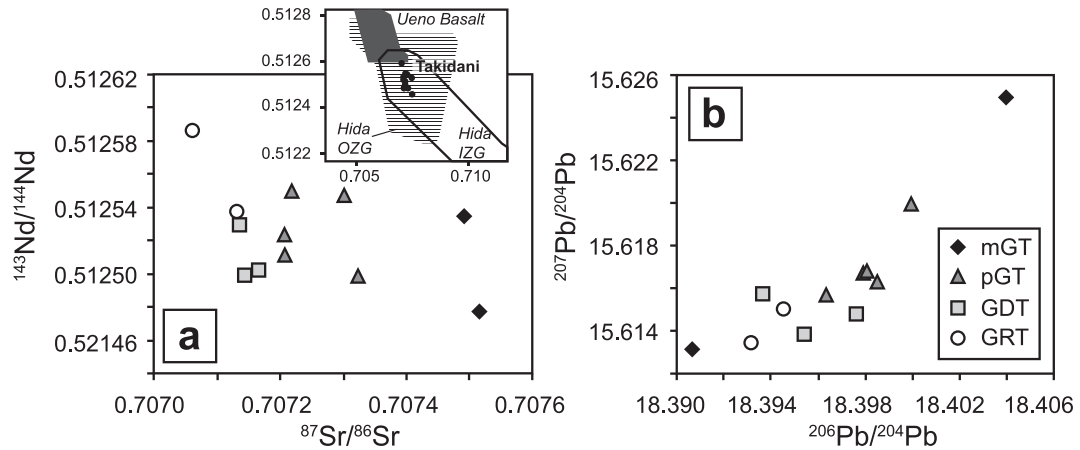


FIGURE 3.6: Variation of $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ for whole-rock samples from the TK. Sr-isotope compositions become more radiogenic roofwards from GRT to mGT. The samples have a typical signature of Hida outer zone granites (OZG) and inner zone granites (IZG; Kimura & Nagahashi 2007). (b) Variation of $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ for whole-rock samples from the TK. Pb isotope compositions generally increase roofwards from GRT to mGT. The samples of the mGT unit show a large spread in Pb isotope ratios. All values are corrected for an age of 1.5 Ma.

However, P2 and P3 are indistinguishable in An-maps and can be distinguished only through quantitative analyses (e.g. FeO content). No An-rich cores (Figure 3.9) are observed in mGT plagioclase, whereas GDT and pGT plagioclase contains up to 2 area% (Figure 3.9; Table ??). QEMSCAN and EPMA data both show that mGT contains only P1. EPMA analyses reveal that P2 is common in the GDT and pGT units (70–91%). P3 is the dominant population in GRT. Importantly, P1 is found in pGT (24%) but does not occur in either GDT or GRT. P3 is also present to small amounts in pGT (6%) and GDT (9%) units (Table ??).

3.8.2 Amphibole

We recalculated all amphibole compositions based on 23 oxygens, using the formula of Holland & Blundy (1994, Table ??). To obtain $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratios, an average factor (f_{av}) was computed after Holland & Blundy (1994), who extended the criteria for elemental site occupancy of Robinson et al. (1982) and Spear & Kimball (1984). Based on the calculated stoichiometry, all amphiboles can be classified as magnesio-hornblendes (Figure 3.10a; Leake et al. 1997), with the exception of a few high-Si crystals, which classify as actinolite.

Amphibole shows moderate compositional variation across the Takidani pluton with two distinct populations (Figure 3.10b). Based on their aluminium contents, we classify these populations into low-Al ($\text{Al}_2\text{O}_3 < 5.8$ wt.%; $\text{Al}_{total} < 1.0$ a.p.f.u.; Table ??) and high-Al amphibole-types ($\text{Al}_2\text{O}_3 > 5.8$ wt.%; $\text{Al}_{total} > 1.0$ a.p.f.u.; Table ??). Most

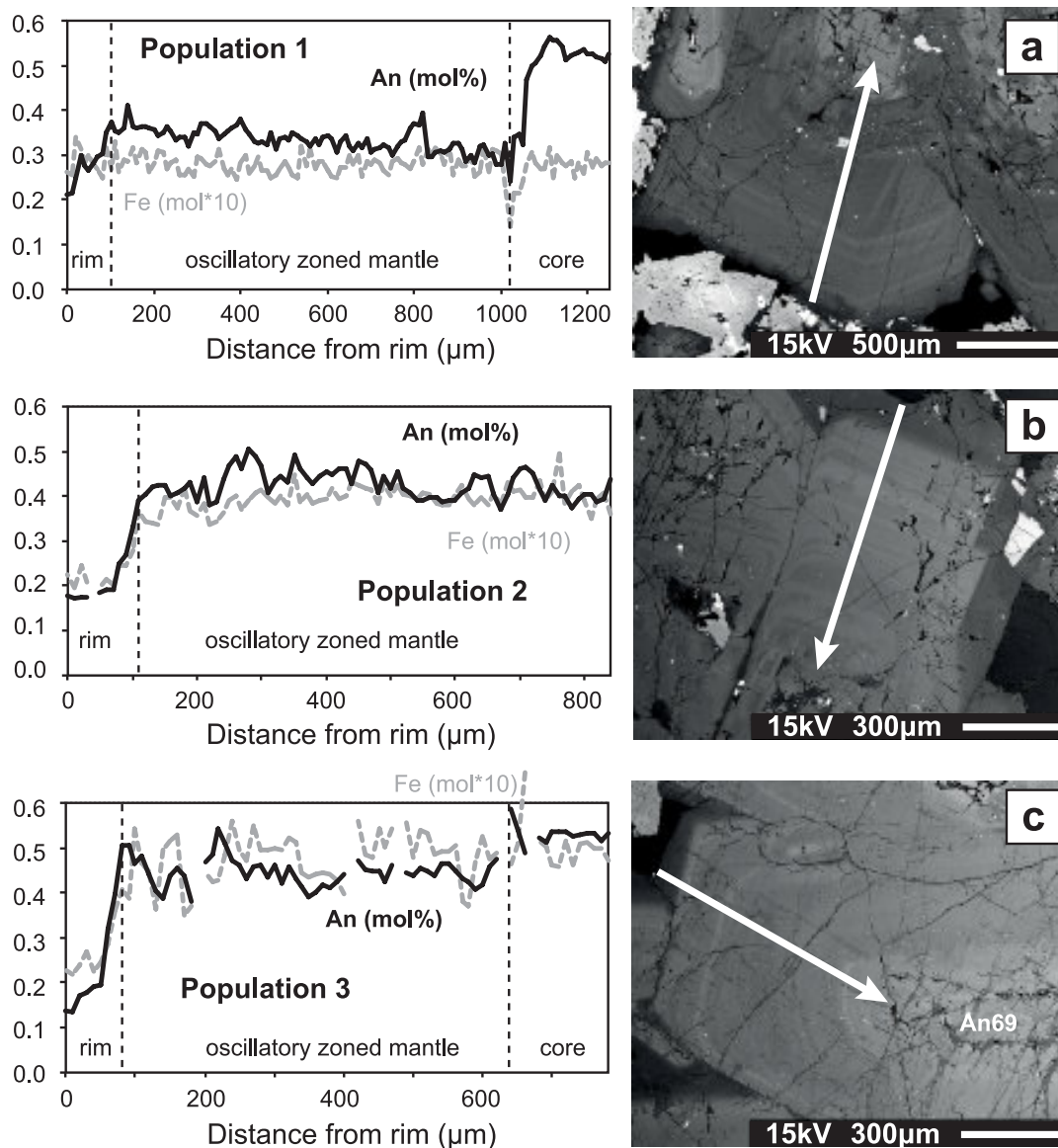


FIGURE 3.7: Electron microprobe profiles (10 μm steps) of examples of the three plagioclase populations. Each population is defined by the Anorthite and iron content of the oscillatory-zoned mantle of each crystal. Population 1 is reversely zoned from about 30 to 40 mol% Anorthite (a). Populations 2 and 3 have similar An contents, varying between about 40 and 50 mol%, but distinctively different iron concentrations (b, c)

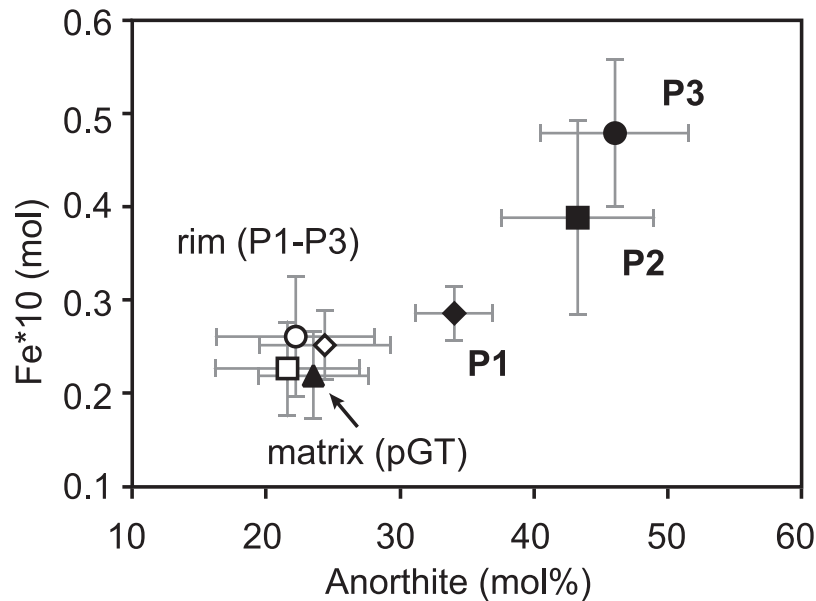


FIGURE 3.8: Variations of Fe and An contents of plagioclase populations (i.e. oscillatory-zoned mantle) from 42 plagioclase traverses of the TK. Analyses of rim plagioclase (white fill) and matrix plagioclase of pGT (black triangle) have similar chemistry. Errors are given as 1 σ

crystals display chemically distinct patchy zones (Figure 3.11a,b); very few crystals have rounded amphibole cores or contain amphibole grains with Al_2O_3 content above 8 wt.% (Figure 3.10c,d). Back-scattered electron (BSE) images of amphibole grains show that brighter zones are higher in Al than darker ones (Figure 3.11b).

3.8.3 Biotite

Biotite is invariably interstitial and anhedral (with the exception of pGT), suggesting its late appearance in the crystallisation sequence. Chemical compositions vary only slightly and without any obvious trends between the units (Table ??). However, concentrations of Mn (Figure 3.12a) and Cl (Figure 3.12b) plotted against biotite magnesium-number ($\text{Mg}\#$) seem to outline different populations. The concentration of Mn in biotite tends to increase towards the roof of the intrusion (from GRT to pGT), whereas the concentration of Cl in biotite generally decreases roofwards from 0.48 to 0.35 wt.%.

3.8.4 Oxides

Titano-magnetite is the dominant oxide phase in Takidani samples (Tables ??–??). Ilmenite is also present and is usually associated with magnetite, forming exsolution lamellae or single grains. The compositions of the Fe–Ti oxides are generally homogeneous across the pluton (Tables ??–??); however, a relatively minor increase of MnO content in ilmenite follows the same increasing trend observed for MnO content in biotite and low-Al amphiboles.

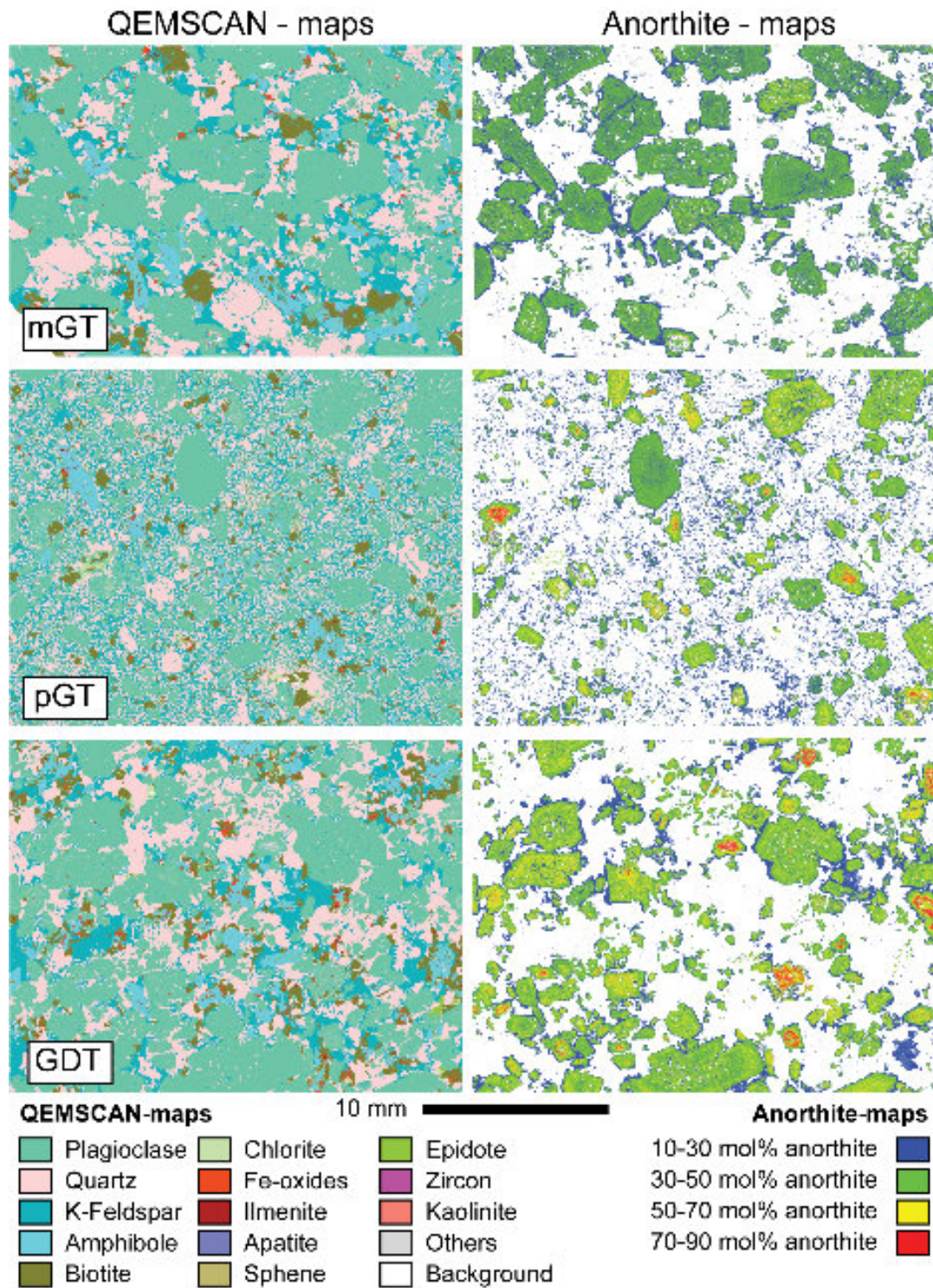


FIGURE 3.9: QEMSCAN and Anorthite maps of the GDT, pGT and mGT units. Anorthite maps are calibrated with electron microprobe data to define Anorthite bins of 20 mol%

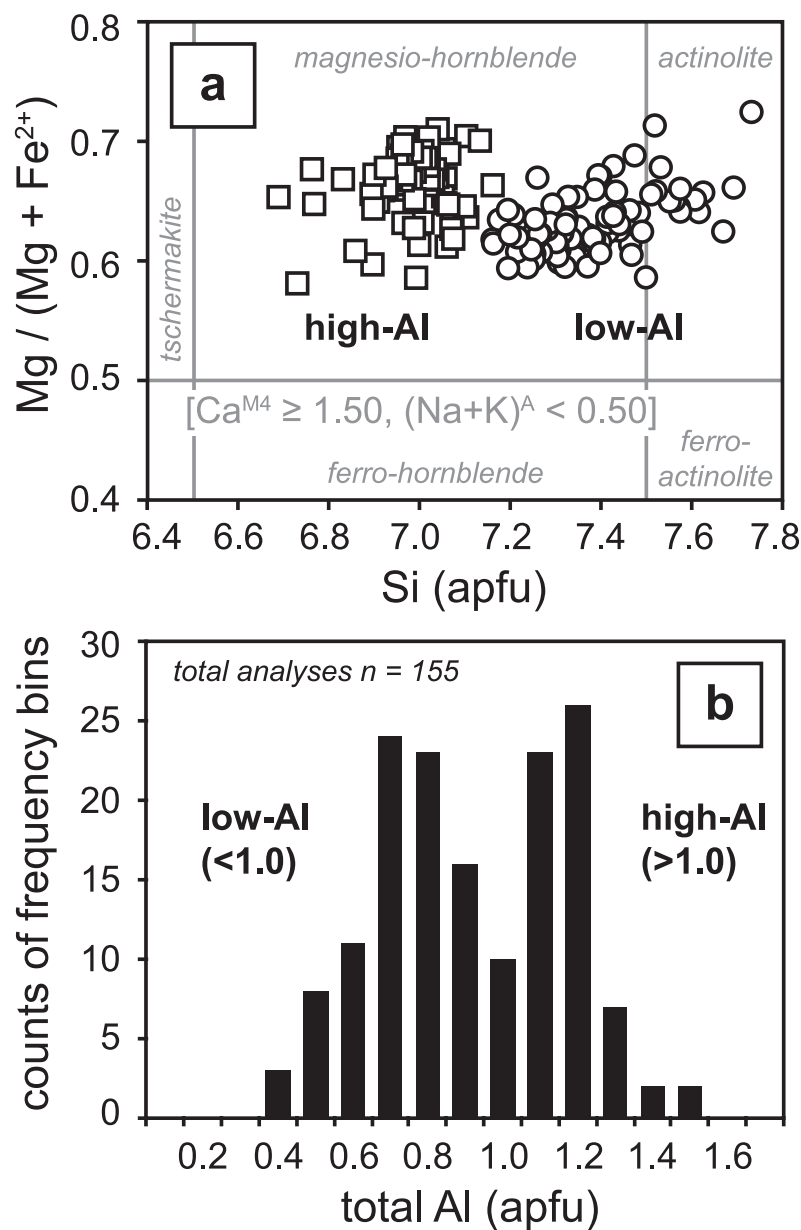


FIGURE 3.10: Amphibole classification (a) and subdivision of amphiboles based on total Al a.p.f.u. (b). Stoichiometry is recalculated after Holland & Blundy (1994) and plotted using the classification scheme of Leake et al. (1997). Most amphiboles classify as magnesio-hornblende with the exception of a few amphibole rims, which fall into the actinolite field. The total Al (in a.p.f.u.) values show a bimodal distribution with low-Al (<1.0 total Al a.p.f.u.; Table ??) and high-Al (>1.0 total Al a.p.f.u.; Table ??) amphibole populations

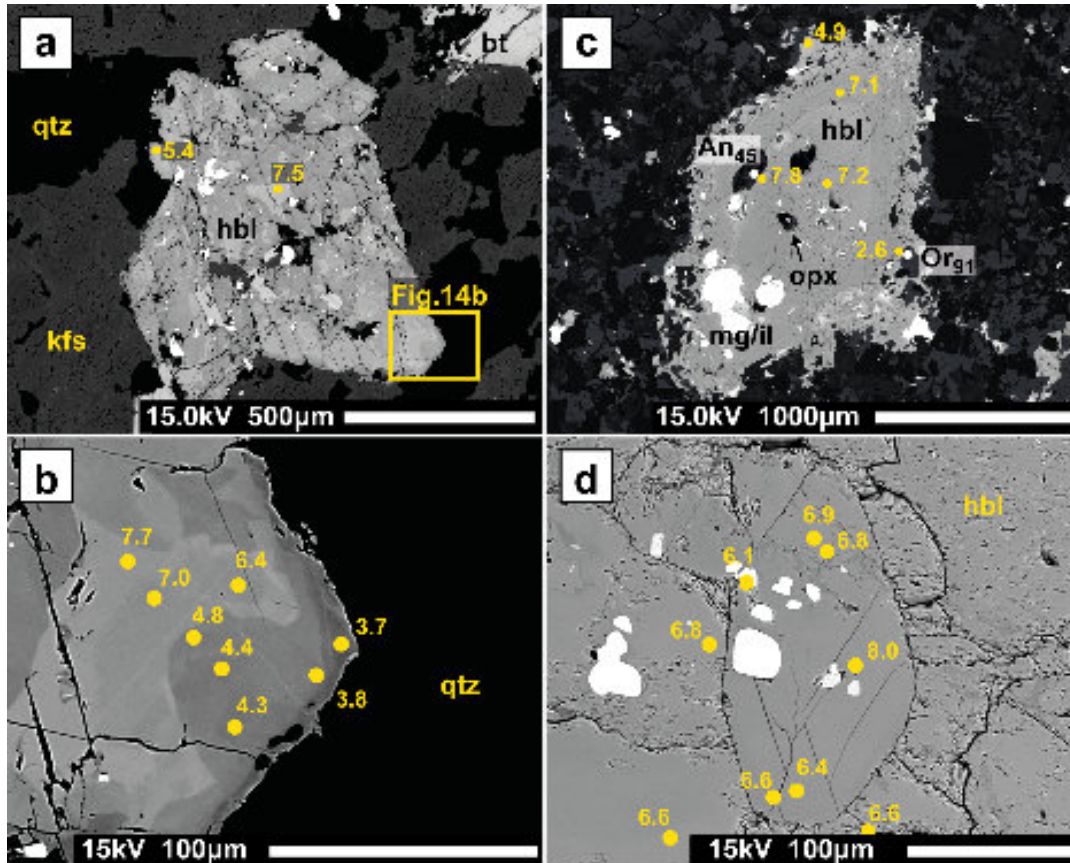


FIGURE 3.11: Back-scattered electron (BSE) images of selected amphibole grains including Al_2O_3 contents (in wt.%) of electron microprobe spot analyses. (a) Patchy zoned amphibole in mGT. (b) Patchy zoning in amphibole with Al_2O_3 varying from 7.7 wt.% in the bright zone to 3.7 wt.% in the darker coloured rim. (c) Amphibole in pGT contains plagioclase with 45 mol% Anorthite along a rounded crystal core. The interstitially grown crystal rim contains less than 5 wt.% Al_2O_3 . (d) Amphibole inclusion with Al_2O_3 content up to 8 wt.%.

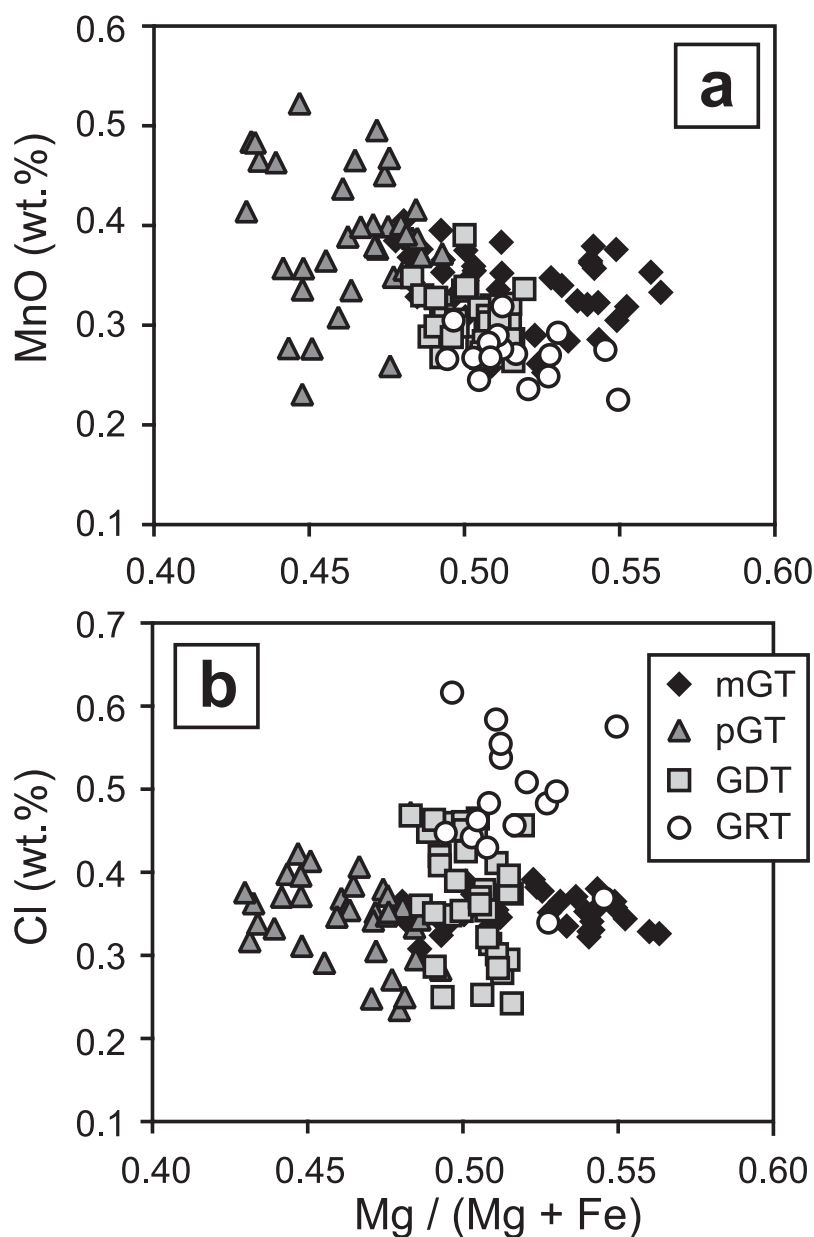


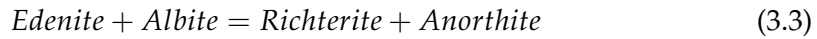
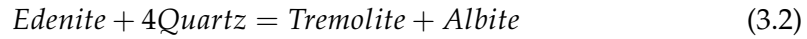
FIGURE 3.12: (a) Variation of MnO vs Mg/(Mg+Fe) for biotite from TK lithological units. Biotite grains from the fine-grained pGT unit have generally lower magnesium contents than in all other units, indicating that they crystallised from potentially more evolved melts. (b) Variation of Cl vs Mg/(Mg+Fe) for biotite from the TK. Cl contents can be used to separate biotite of GRT and mGT into two distinct groups. The highest Cl contents in biotite are found in GRT at the base of the TK.

3.8.5 Barometry and thermometry

The Al-in-hornblende geobarometer of Anderson & Smith (1995) is calibrated using data from Johnson & Rutherford (1989) and Schmidt (1992), and can be used to estimate the crystallisation pressure, P (in kbar), of hornblende, based on the total aluminium content (in atoms per formula unit, or a.p.f.u.), provided that the crystallisation temperature, T ($^{\circ}\text{C}$), is independently known:

$$P(\pm 0.6\text{kbar}) = 4.76Al - 3.01 - \left(\frac{T[^{\circ}\text{C}] - 675}{85(0.530Al + 0.005294(T[^{\circ}\text{C}] - 675))} \right) \quad (3.1)$$

Temperature estimates can be derived from the pressure-dependent plagioclase-hornblende thermometers of Blundy & Holland (1990; BH90) and Holland & Blundy (1994; HB94-A and HB94-B) using the following two equilibria:



The first equilibrium reaction is used for BH90 and HB94-A thermometers (equation 3.2) and requires silica-saturation. BH90 is based on a more restricted experimental dataset of igneous rock suites, considering only ideal interactions in amphibole. The HB94 model is extended to a wider dataset of igneous and metamorphic rocks to account for non-ideal interactions in amphiboles and includes an improved hornblende formulation to estimate ferric iron contents, optimised for the HB94 thermometers. A comparison of different thermometers (Anderson 1996, Bachmann & Dungan 2002, Pamukcu et al. 2015) shows that the Edenite–Richterite thermometer of HB94 (equation 3.3) in combination with the Al-in-hornblende barometer of Anderson & Smith (1995) produces the most reliable temperature solutions for plutonic and volcanic rocks at or near the granitic to tonalitic solidi (Anderson et al. 2008).

Fe-rich hornblendes may not be suitable to use in geobarometry, depending on the amount of ferric iron. Anderson & Smith (1995) noted that $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratios that are lower than 0.25 for amphiboles recalculated after Leake et al. (1997) should not be used for their Al-in-hornblende barometer. Because Holland & Blundy (1994) used a different method to calculate $\text{Fe}^{3+}/\text{Fe}^{2+}$, we first recalculated the stoichiometric formula for amphiboles using the method of Leake et al. (1997) and excluded analyses with $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratios lower than 0.25; then we recalculated the cation occupancies in amphibole after Holland & Blundy (1994) to apply their thermometer.

3.9 Discussion

3.9.1 Reconstructing the plumbing system of Takidani

A comparison between natural samples and experiments performed at different pressures, temperatures and water activities has the potential to estimate ranges of crystallisation depth and temperature, therefore providing information about the architecture of magmatic plumbing systems.

Previous estimates of emplacement depth and temperature for the TK were based on the Al-in-hornblende geobarometer of Anderson & Smith (1995) and the thermometer of Blundy & Holland (1990). The derived average emplacement pressure was about 250 MPa (± 60 MPa) with slightly lower estimates for the emplacement pressure in the Shiradashi-zawa valley (222 MPa; Bando et al. 2003). Calculations performed using samples collected in different portions of the pluton indicate an increase of emplacement pressure from 165 MPa at the base to 250 MPa at the roof of the intrusion. The progressive decrease of emplacement depth from about 9.5 km to 6.5 km has been previously explained by unroofing during emplacement of the pluton by underaccretion (Bando et al. 2003). Bando et al. (2003) estimated the magma temperature during intrusion at 710°C. In this study we reassess the emplacement conditions of the TK using the high spatial resolution sampling performed in our study and applying the geothermometers of Holland & Blundy (1994). We further compare the whole-rock chemistry and mineral chemistry of our samples with phase equilibria experiments performed at various pressures, temperatures and water contents with starting material of composition similar to the rocks we have studied (Whitney 1975, 1988, Sisson & Grove 1993, Scaillet & Evans 1999, Costa et al. 2004, Holtz et al. 2005, Arce et al. 2013, Caricchi & Blundy 2015b).

Experimental studies have shown that temperature and pressure have an effect on amphibole compositions as a result of atomic exchange reactions (Spear 1981, Johnson & Rutherford 1989, Blundy & Holland 1990, Thomas & Ernst 1990, Schmidt 1992). An increase in pressure leads to the replacement of Mg by Al in the octahedral M-sites (Al-Tschermak substitution; $\text{Si}^T + \text{Mg}^{M1-M3} = \text{Al}^T + \text{Al}^{M1-M3}$), whereas an increase in temperature causes Ti to replace Mn in the octahedral M-sites (Ti-Tschermak substitution; $2\text{Si}^T + \text{Mn}^{M1-M3} = 2\text{Al}^T + \text{Ti}^{M1-M3}$) and causes Na and K to occupy the A-sites [edenite exchange; $\text{Si}^T + \square^{A1} = \text{Al}^T + (\text{Na}+\text{K})^A$]. In all three cases, Al substitutes for Si in the tetrahedral T-sites.

The TK hornblendes show a range of Al^T between 0.3 and 1.4 a.p.f.u. that correlates positively with the edenite $[(\text{Na}+\text{K})^A]$; Figure 3.13a], Ti-Tschermak (Ti^{M1-M3} ; Figure 3.13b) and Al-Tschermak (Al^{M1-M3} ; Figure 3.13c) components throughout the various units constituting the TK. Edenite and Ti-Tschermak exchange reactions are estimated from regression slopes (Shane & Smith 2013) and account for about 21%

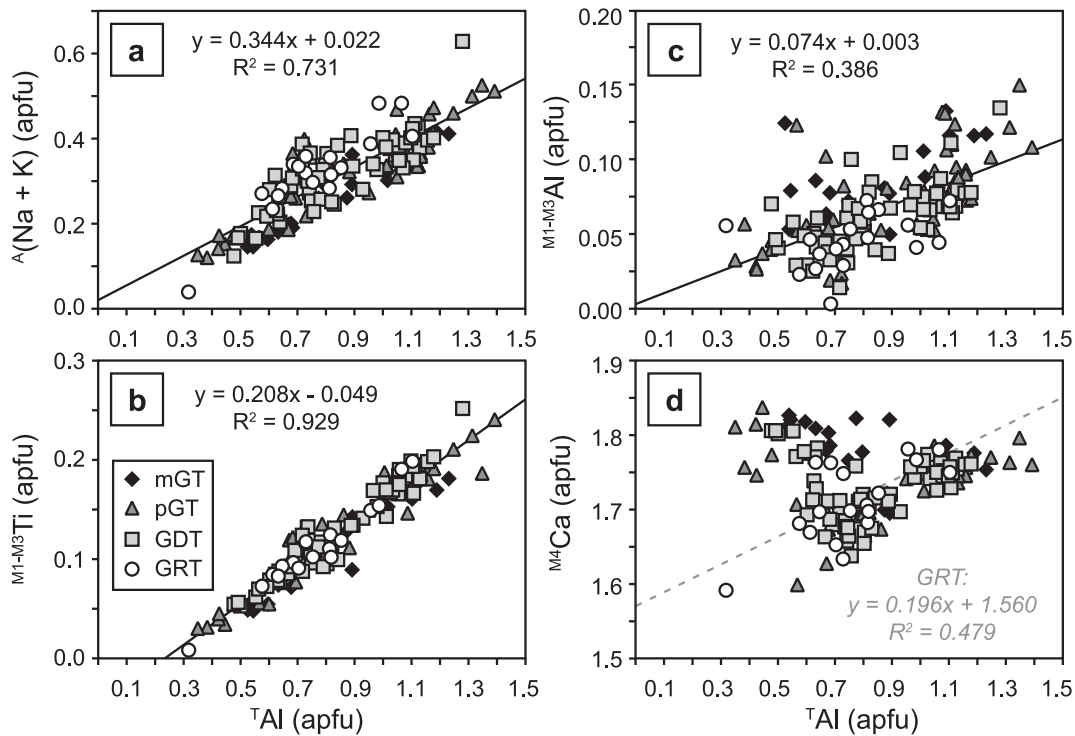


FIGURE 3.13: Amphibole substitutions: (a) temperature-dependent Ti-Tschermak exchange; (b) temperature-dependent edenite exchange; (c) pressure-dependent Al-Tschermak exchange; (d) plagioclase-dependent Ca exchange. Slopes are calculated using all the data with the exception of the Ca-exchange reaction, where the slope is calculated based on the analyses of GRT only.

and 34% respectively of the total Al variability, whereas the Al-exchange reaction accounts for only about 7%. The plagioclase exchange reaction on the larger M4 site ($\text{Si}^T + \text{Na}^{M4} = \text{Al}^T + \text{Ca}^{M4}$) also appears to play a significant role in the Takidani amphiboles (Figure 3.13d). The rather poor correlation between Ca^{M4} and Al^{IV} is probably caused by differences in the An–Ab components of the co-crystallising plagioclase. Whereas the majority of amphibole analyses above 0.65 Al (a.p.f.u.) show a positive correlation between Ca^{M4} and Al^{IV} , the correlation is negative (–0.094) for amphiboles from mGT (0.5–1.2 Al a.p.f.u.). Such negative correlation, together with the reverse zoning in P1 of mGT, was possibly caused by the input of mafic magma.

Plagioclase–hornblende pairs

Subhedral plagioclase with Anorthite content of 40–45 mol% is occasionally included in high-Al amphiboles with Al_2O_3 close to or above 8 wt.%. Amphibole oikocrysts occasionally contain small plagioclase (An_{30}) or alkali feldspar inclusions. The majority of amphibole grains are associated with apatite, Fe–Ti oxides and biotite. Amphibole crystals also tend to be much smaller than plagioclase phenocrysts and therefore have most probably formed after the crystallisation of the plagioclase phenocrysts ($<\text{An}_{40}$; Figures 3.8, 3.9). A rapid decrease from An_{40} to An_{30} is observed in the

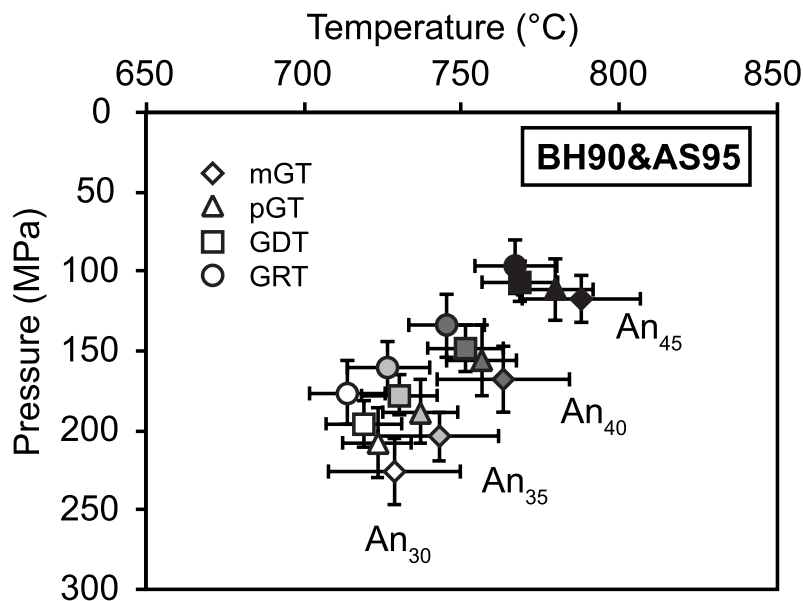


FIGURE 3.14: Mean temperature and pressure estimates with 1σ error bars for high-Al amphiboles using the Blundy & Holland (1990) thermometer and Anderson & Smith (1995) barometer. Results are shown for a range of An contents varying from 30 to 45 mol%.

plagioclase rims, indicating either chemical evolution of the residual melt during progressive cooling or a decrease in water activity (Riker et al. 2015) caused by the introduction of CO_2 from depth (Newman & Lowenstern 2002, Papale & Barbato 2006). All plagioclase phenocrysts share the same late-stage evolution of the outer rim chemistry, which suggests that the Na-rich overgrowth was formed when the magma achieved a mush state with a crystallinity above 40–50 vol% (Hildreth 1981, Marsh 1981).

Given the difficulty of finding enough amphibole–plagioclase pairs in direct contact to apply the Blundy & Holland (1990) and Holland & Blundy (1994) thermometers in a meaningful way, we considered a range of Anorthite contents of the plagioclase in equilibrium with amphibole of 30–45 mol%. These estimates are based on the composition of plagioclase inclusions found in amphiboles. Additionally, matrix plagioclase in pGT has a maximum Anorthite content of 30 mol%, suggesting that amphibole crystallised before the matrix plagioclase. Calculated mean temperatures for Shiradashi-zawa samples (Table ??) vary from $673^\circ\text{C} (\pm 28^\circ\text{C})$ to $700^\circ\text{C} (\pm 33^\circ\text{C})$ using Anorthite contents of 30 mol%. Mean values increase from temperatures of $715^\circ\text{C} (\pm 35^\circ\text{C})$ to $750^\circ\text{C} (\pm 42^\circ\text{C})$ using Anorthite contents of 45 mol%. Our pressure estimates are somewhat lower than those determined by Bando et al. (2003), with average pressures between 99 MPa (± 59 MPa) and 161 MPa (± 67 MPa) using 30 mol% Anorthite (Table ??), and between 63 MPa (± 33 MPa) and 92 MPa (± 32 MPa) using 45 mol% Anorthite (Table ??).

To constrain better the crystallisation conditions of the TK we considered the two main amphibole populations [low-Al (3.6–5.8 wt.% Al_2O_3) and high-Al (5.8–8.0 wt.% Al_2O_3)] individually. Overall, the average temperature ranges between 714°C ($\pm 12^\circ\text{C}$) and 788°C ($\pm 24^\circ\text{C}$) for high-Al amphiboles and between 654°C ($\pm 24^\circ\text{C}$) and 700°C ($\pm 18^\circ\text{C}$) for low-Al amphiboles (Figure 3.14; Table ??). Recalculated crystallisation pressures are also distinctively higher for high-Al versus low-Al amphibole populations (Table ??). Pressures determined for high-Al amphibole show a general increase of crystallisation pressure towards the roof from about 97 MPa (± 6 MPa) to 226 MPa (± 21 MPa), in general agreement with Bando et al. (2003). Calculations with low-Al amphibole yield relatively homogeneous crystallisation pressures of 52–82 MPa across the entire pluton (Table ??). Considering a range of An contents, our pressure estimates (P) correspond to a crystallisation depth, h [$h = P/(g\rho)$], of 3.5–8.5 km for high-Al amphibole, and of 2.0–3.0 km for low-Al amphibole, applying an average crustal density (ρ) of 2700 kg m⁻³.

Importantly, low-Al amphiboles are generally observed as patchy zones and may indicate magma oxidation and near- or sub-solidus alteration (Hammarstrom & Zen 1986, Kawakatsu & Yamaguchi 1987, Martin 2007). Therefore, both amphibole populations could have formed under near-isobaric conditions and the difference in Al may be related to the temperature and chemical variability of the residual melt from which they crystallised, more than pressure, as also suggested by the calculated exchange reactions (Figure 3.13; Spear 1981, Blundy & Holland 1990, Scaillet & Evans 1999, Bachmann & Dungan 2002, Costa et al. 2004).

Considering that pressure appears to have a secondary effect in controlling the chemical variability of amphibole, we consider isobaric crystallisation and recalculate temperatures at 200 MPa. Under this assumption high-Al amphiboles yield crystallisation temperatures of 766–810°C for HB94-A, 733–814°C for HB94-B and 711–776°C for BH90 (Figure 3.15; Table ??), temperatures significantly higher than those estimated by Bando et al. (2003). Temperature estimates for low-Al amphiboles range between 686 and 742°C for HB94-A, between 685 and 771°C for HB94-B and between 638 and 681°C for BH90 (Table ??). The results from the BH90 thermometer yield values lower than temperature estimates obtained from the semi-empirical Ti-in-hornblende thermometer of Otten (1984; Table ??), with the exception of amphiboles from mGT. The HB94-A and HB94-B thermometers produce temperature estimates that correlate with or are higher than the semi-empirical temperature calibrations (Otten 1984). This is not the case for the mGT unit, which appears to have crystallised under slightly different magmatic conditions. HB94-B produces a distribution of temperature estimates that is close to unimodal for the analysed sample (Figure 3.15) and shows peaks between 750 and 810°C for Anorthite 30 and 45 mol%, respectively.

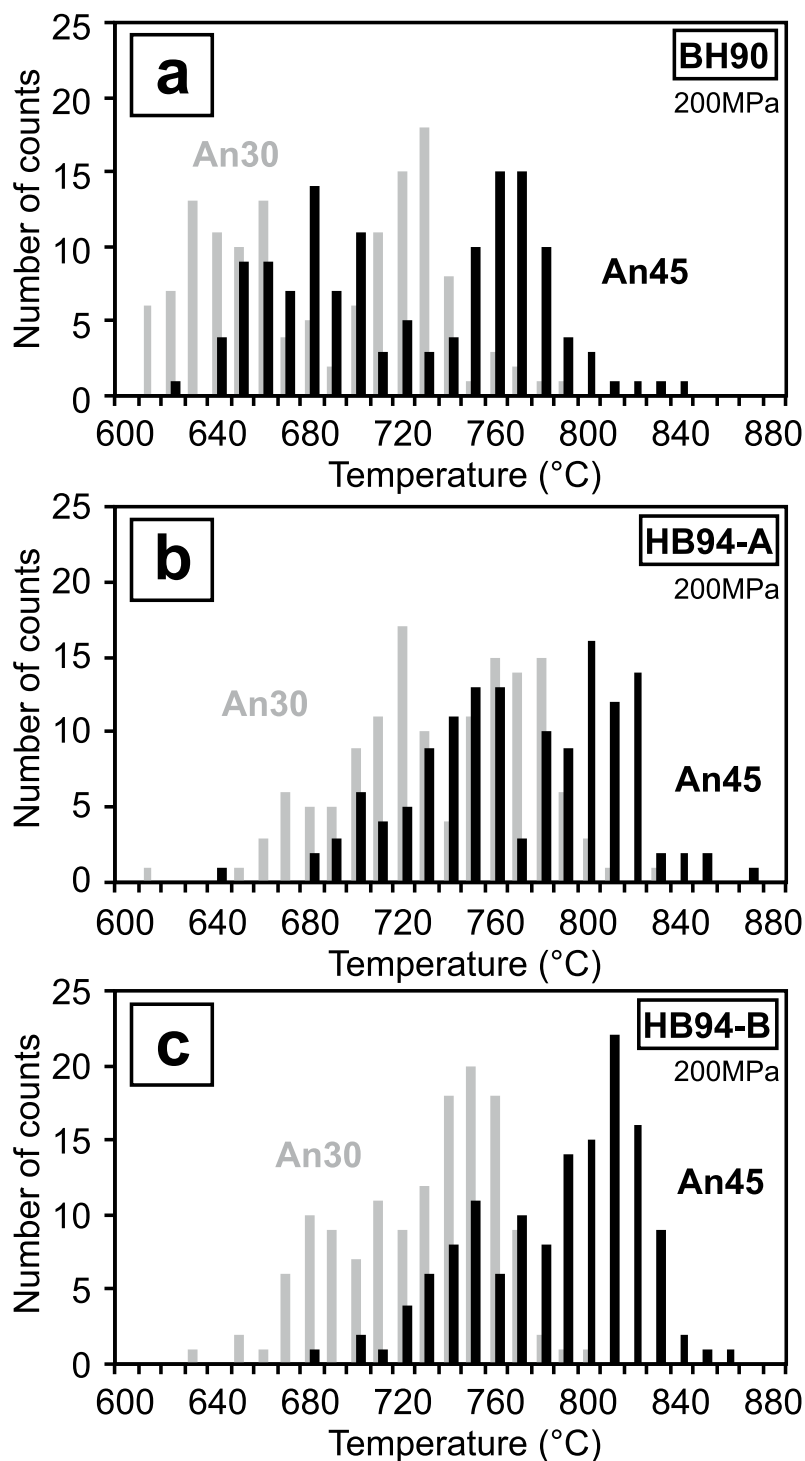


FIGURE 3.15: Temperature distribution for TK amphiboles at 200 MPa using the thermometers of Blundy & Holland [1990; BH90 (a)] and Holland & Blundy [1994; HB94-A (b) and HB94-B(c)]. Temperatures are calculated for An contents of 30 mol% (grey) and 45 mol% (black).

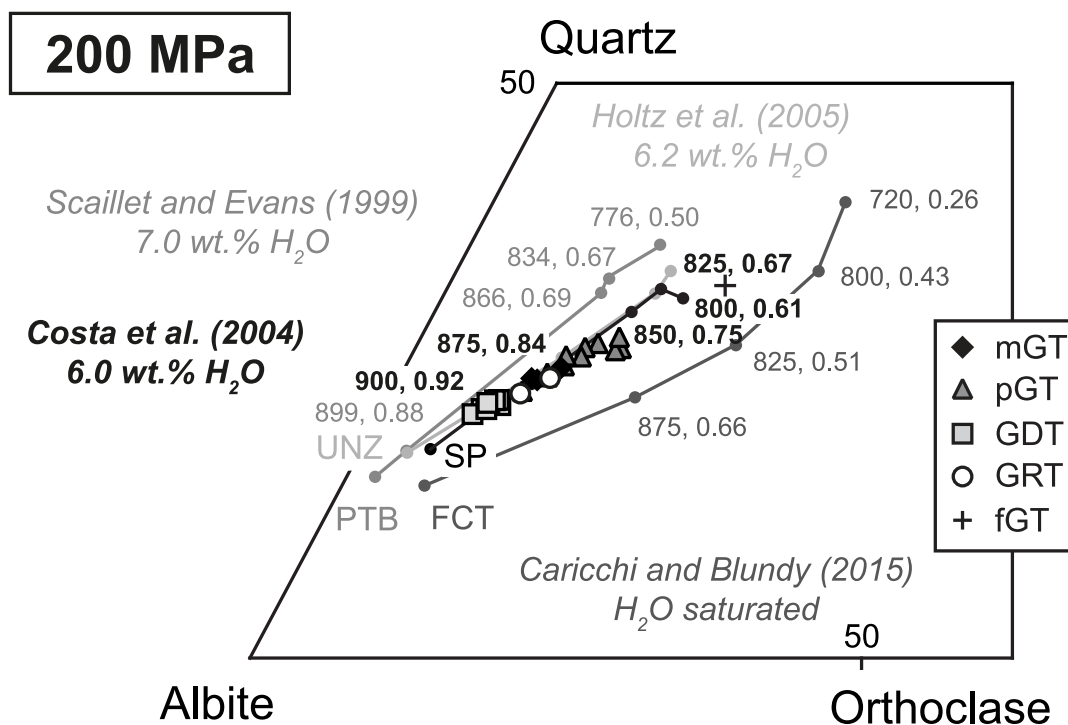


FIGURE 3.16: Compositions of TK bulk-rock and liquids from experimental studies projected into the ternary Qtz–Ab–Or system. The ternary projection is corrected for Anorthite following Blundy & Cashman (2001). The TK samples are consistent with the liquids of Costa et al. (2004) and Holtz et al. (2005). Temperatures are listed in degrees Celsius followed by the melt fraction after the comma. TK samples plot between 850 and 900°C. UNZ, Unzen; PTB, Pinatubo; SP, Volcán San Pedro; FCT, Fish Canyon Tuff.

Experimental studies

Chemical analyses of the TK were compared with several relevant experimental and petrological studies (Whitney 1975, 1988, Sisson & Grove 1993, Scaillet & Evans 1999, Costa et al. 2004, Holtz et al. 2005, Caricchi & Blundy 2015b) to assess further the P – T – H_2O conditions at and after emplacement. We projected the residual melts from various experimental studies, equilibrated at different temperatures at around 200 MPa, in the haplogranitic system together with the whole-rock compositions of the TK rocks by calculating their normative Qtz–Ab–Or abundances (i.e. CIPW; Figure 3.16). All calculations are corrected for the effect of Anorthite in the Qtz–Ab–Or ternary projection (Blundy & Cashman 2001). Good correlations of the TK samples are observed with experimental liquids (at or near water saturation) synthesised at 200 MPa using compositions from Unzen Volcano (Holtz et al. 2005) and Volcàn San Pedro (Costa et al. 2004). The TK samples exhibit compositional variability within this ternary system compatible with crystallisation within a temperature range of between 850 and 900°C, higher than the temperatures obtained by hornblende thermometry (Figure 3.15). The TK plagioclase grains span a range of compositions very similar to those obtained experimentally by (Costa et al. 2004). However, in the experiments hornblende contains more than 8 wt.% of Al_2O_3 and is the liquidus phase for water content above 4 wt.% (Figure 3.17; Costa et al. 2004). The mineral chemistry of hornblende from the experiments differs significantly from the high-Al amphibole in the TK. This, together with microscopic observations, suggests that the TK amphibole crystallised after plagioclase and potentially orthopyroxene (Figure 3.11c). This implies that at emplacement the magma was not water-saturated and must have contained less than 4 wt.% water (Newman & Lowenstern 2002). During cooling and crystallisation, the water content progressively increased until amphibole (>4 wt.%) and biotite saturation (>5.5 wt.%) was reached (Figure 3.17).

P – T – H_2O summary

Our detailed comparison of bulk-rock and mineral chemistry, available thermometers, barometers and experimental data allows us to infer the pressure of emplacement and the evolution of temperature and water content of the residual melt within the TK. The calculated crystallisation conditions suggest that different pulses of chemically distinct magmas (at least two main pulses) were emplaced at about 200 MPa and temperatures between 850 and 900°C with bulk water contents of approximately 3–4 wt.% (Figure 3.17; Costa et al. 2004). Plagioclase stabilised first (i.e. phenocrysts $>An_{50}$) followed by orthopyroxene. Amphiboles with Al_2O_3 contents above 8 wt.% started crystallizing at An contents of about 45 mol% and temperature of about 850°C (Figure 3.17). The sharp drop in An content observed in plagioclase rims to about 30 mol% indicates that the magma cooled fairly rapidly to temperatures below 800°C, at which point the water content of the residual melt was appropriate to crystallise biotite (>5.5 wt.%). Microscopic observations and temperature calculations (with $An < 40$) also

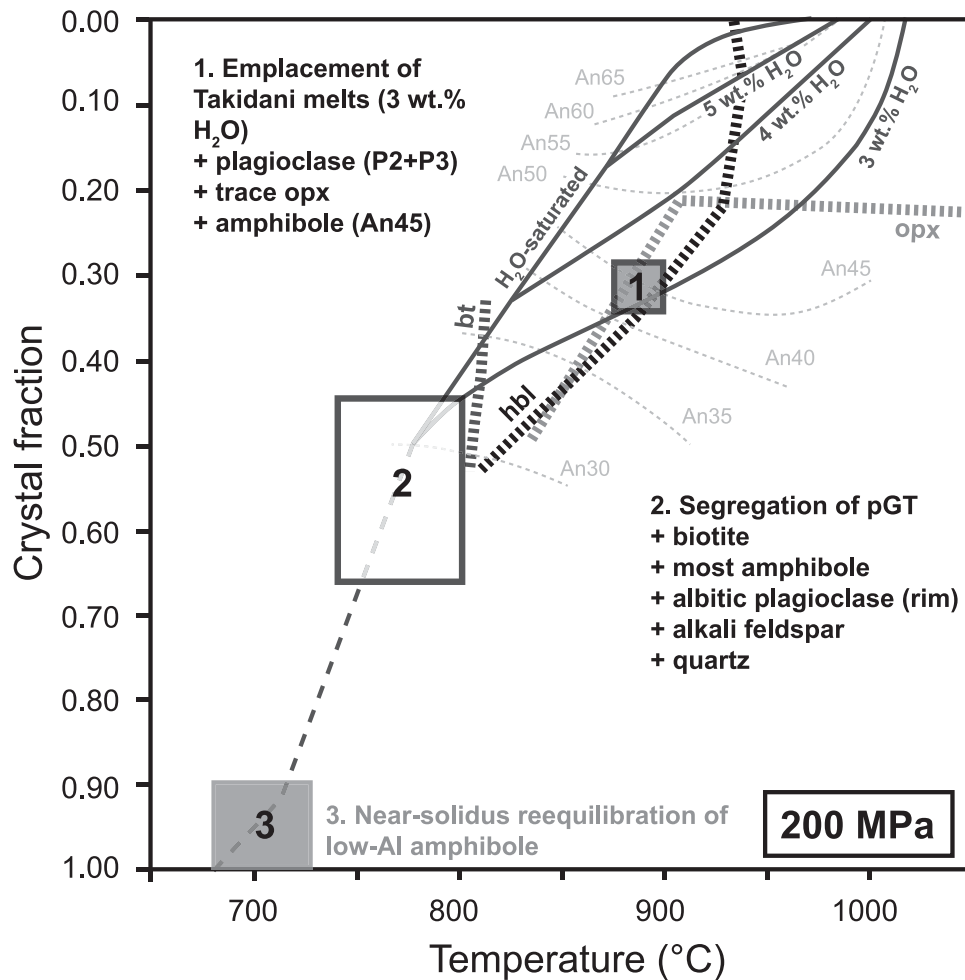


FIGURE 3.17: Magmatic evolution of TK plotted in a binary diagram of crystal fraction vs temperature. Crystal fraction, temperature, An contents and phase stability fields are derived from the experiments of Costa et al. (2004). The melt evolution is projected for bulkrock water contents of 3, 4, 5 and 6 wt.%. Liquidus temperatures are calculated using rhyolite-MELTS (Gualda et al. 2012, Ghiorso & Gualda 2015). Water saturation of rhyolitic melts is calculated to be about 6 wt.% using VolatileCalc (Newman & Lowenstern 2002).

show that most amphiboles record crystallisation temperatures below 800°C (Figure 3.15), suggesting that the TK magmas spent significantly more time below 800°C to crystallise amphibole than above 800°C. Considering the relatively small amount of crystallisation occurring between 900 and 800°C for a water-undersaturated magma with about 3 wt.% H₂O (or in equilibrium with a mixed H₂O–CO₂ fluid and a water activity of about 0.5; Riker et al. 2015), we can expect a relatively rapid temperature drop within this range owing to the limited thermal buffering provided by the latent heat of crystallisation (Huber et al. 2009, Caricchi & Blundy 2015b). In contrast, the rate of latent heat release produced by crystallisation increases below 800°C, prolonging the residence time of magma within this temperature range (Lee et al. 2015) and leading to the distribution of amphibole compositions (Figure 3.15). Moreover, at temperatures below 800°C magma crystallinities would have reached values in excess of 50 wt.% (Figure 3.17; Costa et al. 2004). At such high crystallinity the rheological locking of the system would account for the homogeneous composition of the plagioclase rims observed in the TK (Figures 3.8, 3.9).

3.9.2 From emplacement to eruption

Emplacement and growth of the intrusion

Many studies of sill emplacement and pluton growth have illustrated that large magma reservoirs are most probably constructed incrementally over ten thousand to millions of years (Bachl et al. 2001, Coleman et al. 2004, Glazner et al. 2004, de Saint-Blanquat et al. 2006, Kavanagh et al. 2006, Matzel et al. 2006, Lipman 2007, Pasquarè & Tibaldi 2007, Menand 2008, 2011, Michel et al. 2008, Annen 2009, 2011, Leuthold et al. 2012, Annen et al. 2015, Menand et al. 2015). The TK is petrochemically stratified, suggesting an incremental growth of the magma reservoir. The mGT unit has the highest ⁸⁷Sr/⁸⁶Sr within the TK; it is isotopically distinct from other units and has a peculiar mineral chemistry. Moreover, barometric calculations suggest that this unit was emplaced under the deepest conditions. Reverse zoned plagioclases in mGT indicate that the magma was reheated after emplacement and, therefore, we suggest that the mGT was the first unit emplaced at the roof of the intrusion. During subsequent magma injection, possibly at shallower depths, a certain amount of mingling must have occurred between the mGT magma and the under-accreted units (i.e. pGT and GDT). For instance, whereas no or only very few of the P2 plagioclase grains (characteristic of GDT and pGT units) are present within the mGT unit, some P1 phenocrysts are found within pGT. This suggests that mGT was still partially molten when the underlying units were emplaced.

Previous researchers have argued that the construction of the TK reservoir must have occurred within ~ 600 kyr after the deposition of the 2.4 Ma Hotaka Andesite (Harayama et al. 2003) but before the deposition of the genetically related Nyukawa PFD and Chayano–Ebisutoge Pyroclastic Deposits at about 1.76–1.75 Ma (Nagahashi

1995, Kataoka et al. 2001, Kimura & Nagahashi 2007). Zircon crystallisation ages of the TK ($1.54 \text{ Ma} \pm 0.23 \text{ Ma}$; Sano et al. 2002) suggest that magmatic conditions were maintained after the eruption of these voluminous volcanic units. Older K–Ar ages of amphibole (i.e. 1.9 Ma ; Harayama 1992) may therefore represent older and solidified parts of the TK that had cooled below 580°C (closure temperature of K–Ar in hornblende; Harayama 1992). This would imply that parts of the pluton had to be re-melted.

We argue that the TK was incrementally under-accreted to form one large coherent magma reservoir. The top mGT unit represents the first magma batch, which was still partially molten upon the intrusion of subsequent units (e.g. GDT and pGT). The granitic units at the base of the pluton may present a third and last period of magma injection.

Origin of pGT: evidence for internal differentiation and segregation of upper crustal residual melt

A unique feature of the pGT unit is the presence of features suggesting a two-stage cooling from an initially crystal-free or crystal-poor magma. All other magma bodies of the TK body show evidence for steady cooling from, presumably, crystal-rich magma upon intrusion. These differences suggest that a detailed study of the pGT may reveal important information about transport of melt in the TK pluton. As discussed below, a combination of field relations, mineral chemistry and mass-balance calculations provides strong evidence to suggest that pGT formed as a consequence of internal differentiation followed by segregation and accumulation of the interstitial melt as a discrete magma body. If correct, the GDT–pGT units present a rare example of upper crustal melt segregation from a stagnant crystal–liquid mush, which in turn lends credence to the multiple theoretical and numerical studies that have suggested such a process (Hildreth 1981, Marsh 1981, Brophy 1991, Srogi & Lutz 1996, Vigneresse et al. 1996, Bachmann & Bergantz 2004, 2008, Hildreth & Wilson 2007, Dufek & Bachmann 2010, Deering et al. 2011, Gelman et al. 2014, Lee & Morton 2015).

Three lines of evidence support an origin of the pGT through internal differentiation of the GDT, followed by melt extraction. First, and most significant, is the sigmoidal variation in both major and trace elements defined by the GRT, GDT, pGT and mGT shown in Figure 3.5. This is entirely consistent with the successive intrusion of three different bodies of granodioritic magma, which underwent in-situ crystallisation. For whatever reason, the first and third intrusions underwent complete crystallisation to form the mGT and GRT whereas the second intrusion underwent incomplete crystallisation, followed by melt extraction and segregation. The pGT unit represents the more evolved extracted melt whereas the GDT represents the seemingly less evolved cumulate remainder, which together lead to the sigmoidal pattern. Second, the gradual contact between the GDT and pGT is consistent with

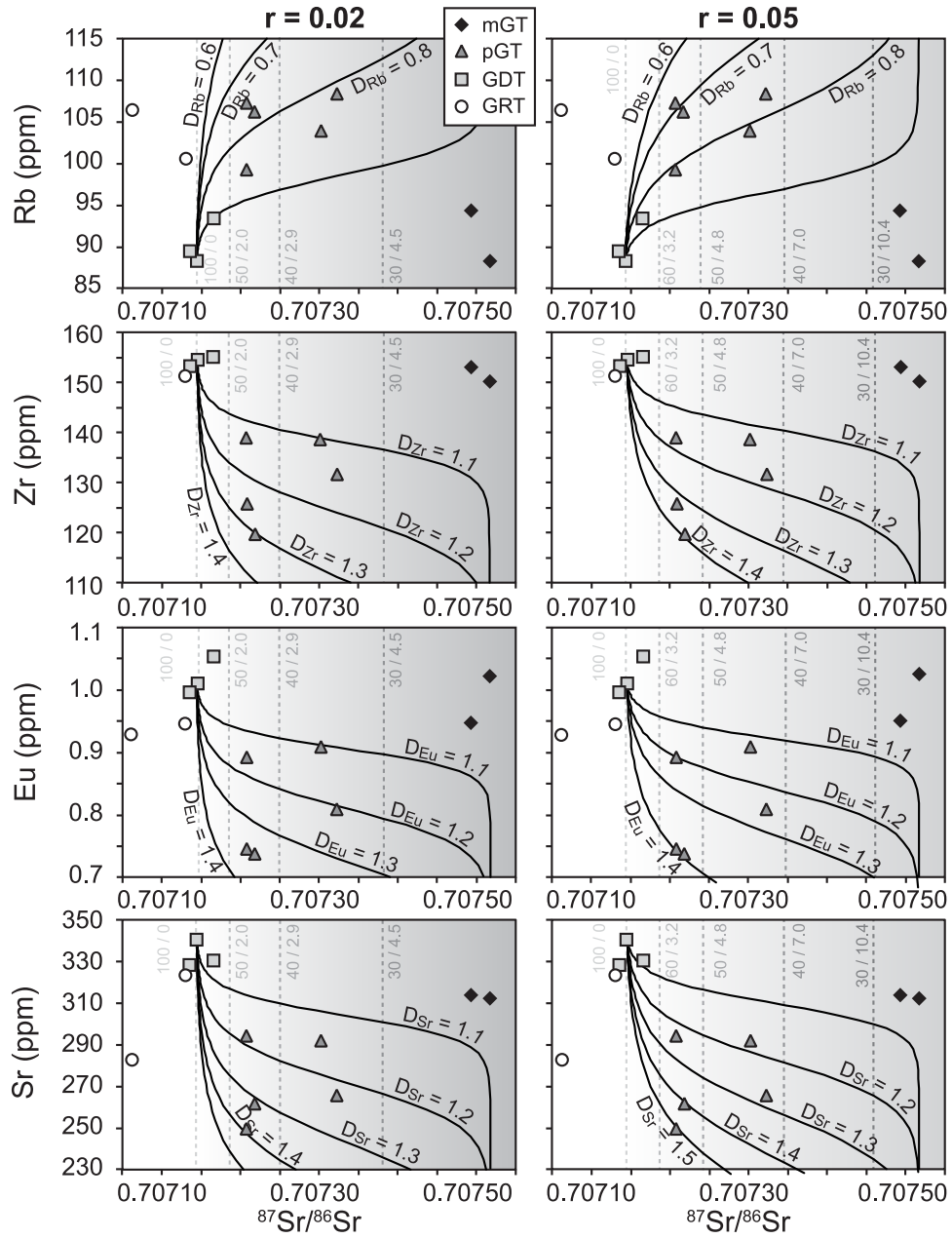


FIGURE 3.18: Assimilation and fractional crystallisation (AFC) modeling of TK melts after (DePaolo 1981). Trace element and isotopic trends of pGT are reproduced with 45–65% of crystal fractionation of GDT and 2–6% of assimilation of mGT.

extraction and segregation of an interstitial melt. Third, the fact that the plagioclase rims of the GDT are compositionally identical to the fine-grained matrix plagioclase of the pGT strongly suggests a genetic relation between the two.

Radiogenic isotope compositions and assimilation-fractional crystallisation models (AFC; DePaolo 1981) support fractionation-dominated differentiation (Figure 3.18). The isotopic and trace element variation of the GDT, pGT and mGT can be explained by 45–65% fractional crystallisation of GDT magma, accompanied by a small amount (2–6%) of incorporation of the overlying mGT. Given that P1 is the only plagioclase type in mGT, its presence in pGT is consistent with the assimilation of partially molten mGT required by the AFC calculations (Figure 3.18). The higher silica content and lower Zr concentration of the pGT (Figure 3.5), with respect to the GDT, suggest that the residual melt was extracted from the crystallizing granodioritic magma at a temperature near or lower than its zircon saturation temperature ($731^{\circ}\text{C} \pm 7^{\circ}\text{C}$; Watson & Harrison 1983). This temperature is a lower limit because the calculations of zircon saturation were performed considering the bulk-rock composition (Harrison et al. 2007).

Based upon a comparison with the experimental results of Costa et al. (2004), we can attempt to identify the conditions at which the extraction of the residual melt occurred. Partly resorbed bytownite cores in plagioclase (up to An_{90}) can be reconciled with the ascent of water-undersaturated magma containing a few plagioclase crystals formed at relatively high pressure (Sisson & Grove 1993, Nandedkar et al. 2014). These Anorthite-rich cores (<1 total area%; Table ??) change rapidly into andesine compositions (An_{50-40}) within the plagioclase domain characterised by oscillatory growth zones. In addition, the Anorthite content of plagioclase included in amphibole is never higher than 45 mol%, suggesting that amphibole becomes stable when An_{45} plagioclase crystallised. According to Costa et al. (2004) contemporaneous crystallisation of amphibole and An_{45} occurs at about 850°C for an H_2O content of 4 wt.% (Figure 3.17). Traces of orthopyroxene were found within amphibole phenocrysts. Orthopyroxene is stable at higher temperature than amphibole: at water contents of about 3 wt.% its presence indicates temperatures above 925°C . Altogether these lines of evidence suggest that after a period of crystallisation at relatively high pressure (700 MPa; Nandedkar et al. 2014), water-undersaturated melt was transported to shallower depths where the crystallisation of An_{45} plagioclase occurred contemporaneously with amphibole at $\sim 850^{\circ}\text{C}$ (Figures 3.15, 3.17).

Following the temperature–crystallinity diagram obtained from the experiments of Costa et al. (2004), the TK magmas would have reached water saturation at about 800°C and 50 wt.% crystals, after progressive cooling and the onset of amphibole crystallisation (Figure 3.17). The chemical variations between the GDT and pGT can be explained by crystal fractionation of up to 65 wt.% (Figure 3.18).

These lines of evidence suggest that the extraction of the residual melt from the GDT occurred at crystallinities between 45 and 65 wt.% and resulted in the formation of the pGT (Figures 3.18, 3.19), which is in agreement with the optimal extraction window identified by the thermomechanical models of Dufek & Bachmann (2010). The coincidence between the crystallinity at which melt extraction seem to have occurred and the achievement of water saturation suggests that the residual melt extraction may have been triggered by the exsolution of excess volatiles (Sisson & Bacon 1999, Fowler et al. 2015).

Pressure quenching and the porphyritic texture of pGT

One of the most striking textural observations of the various units of the TK is the gradual transition from a holocrystalline to a porphyritic texture seen from the top of GDT into pGT. The fine-grained matrix of the porphyritic unit is difficult to explain by temperature quenching given the gradual transition of the major and trace element geochemistry observed from GDT to pGT (Figure 3.5). An alternative explanation for the fine-grained nature of pGT is a sudden decompression that would result in the rapid exsolution of H₂O (Newman & Lowenstern 2002), an increase of liquidus temperature and consequent rapid crystallisation. Because the transition from holocrystalline to porphyritic is gradual, such decompression cannot be associated with the rise of magma toward the surface. A viable process producing a rapid pressure drop and triggering rapid crystallisation could be an eruption. The observation that the proportion of phenocrysts in the pGT varies across the unit suggests that the melt was mobile before it was quenched.

3.10 Conclusion and future work

Our detailed field and petrographic investigations, bulk-rock and mineral analyses of the TK along the Shiradashi-zawa valley lead to the following conclusions regarding its magmatic evolution.

1. The TK represents an evolved magma reservoir in the upper crust comprising five distinct units of granodioritic to granitic composition. The pluton was emplaced incrementally at about 9.5–6.5 km depth. The intrusion of successive magma pulses led to the formation of one large, interconnected magma body.

2. The chemical variations observed in the upper part of the TK are best explained in terms of fractional crystallisation acting together with segregation of the residual melt. The chemical and isotopic compositions of the pGT are best explained by its extraction during crystallisation of the GDT (45–65 wt.% crystals) after minor assimilation of 2–6 wt.% of the overlying mGT unit (Figure 3.19).

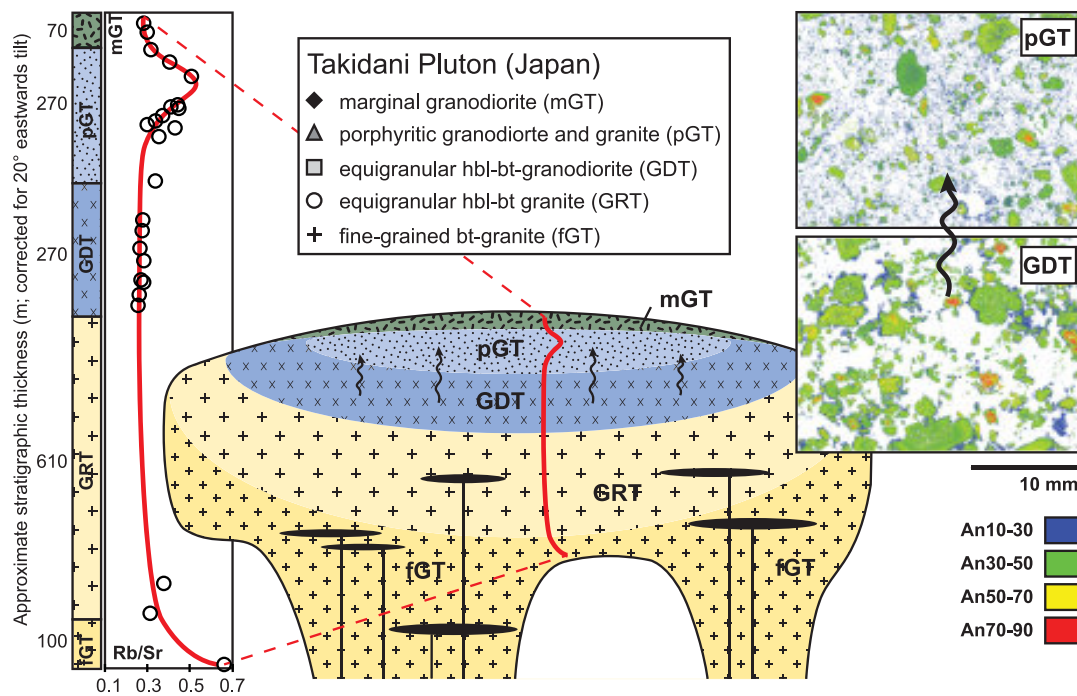


FIGURE 3.19: Cross-section of the TK including melt segregation evidence from the lower granodiorite (GDT) to the overlying porphyritic unit (pGT) illustrated through sigmoidal variations in whole-rock chemistry. GDT and pGT contain the same population of plagioclase phenocrysts (i.e. P2). Overgrowth rims of plagioclase in GDT have the same composition as matrix plagioclase in pGT, further supporting upwards migration of interstitial melt. This figure is based on the schematic illustration of a crystal mush from Hildreth & Wilson (2007)

3. Rapid crystallisation must have occurred to create the fine-grained porphyritic texture of the pGT near the roof of the TK. A volcanic eruption is a plausible mechanism that can lower the pressure, trigger volatile exsolution and cause rapid crystallisation of the remaining melt in the magma reservoir.

Our results suggest that chemical evolution towards the roof of the TK was controlled by melt extraction from crystallizing magma in the upper crust. The presence of a porphyritic unit might be evidence for volcanic activity, making the TK an exciting location to investigate the magmatic build-up to caldera-forming eruptions. A study of trace element variations (i.e. Sr and Eu) in single zoned feldspar grains is currently being undertaken with the aim of shedding more light on the Takidani magmatic system and its genetic association with the Nyukawa and Chayano–Ebisutoge deposits (Kimura & Nagahashi 2007).

3.11 Acknowledgments

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Chapter 4

The role of H₂O on the extraction of melt from crystallising magmas

4.1 Abstract

The segregation and accumulation of felsic melts, from crystallising crustal magma reservoirs, is essential for the chemical evolution of the crust and is a phenomenon preceding some of the largest eruptions on Earth. The physical properties of residual melt and magma and the time over which the conditions remain appropriate for melt extraction are important factors controlling the efficiency of melt extraction and the distribution of melt in magma reservoirs. Here we focus on the initial magma H₂O content as it affects both the physical properties of the residual melt and the timescales at which conditions remain appropriate for melt extraction during progressive magma crystallisation.

We use rhyolite-MELTS simulations to evaluate the physical evolution of crystallising granodioritic (or dacitic) hydrous magma (i.e. ≥ 1 wt.% H₂O) at shallow depth at 200 MPa. To constrain the solidification timescales of reservoirs containing magmas with initially different water content, we perform 2.5D thermal modelling. We combine these results with calculations of melt extraction velocity by compaction and hindered settling to identify the optimal conditions at which melt segregation occurs. These calculations suggest that hydrous felsic magmas that attain water saturation after 40 wt.% crystallisation (rheological locking point) are best suited for melt extraction. Once water-saturation is achieved, the rate of release of latent heat of crystallisation and with it the time magma spends within a given temperature interval increases while the viscosity of the residual liquid and crystal-liquid density contrast remain favourable for melt segregation. We first test our findings on the Takidani pluton (Japan) because it shows evidences of residual melt segregation from crystallising magma, and is associated with caldera-forming eruptions. We finally generalise our results to crustal magma reservoirs containing hydrous felsic magmas. Our results suggest that if segregation starts at rheological locking (i.e. crystallinity of 40–50 wt.%) upper crustal reservoirs of ≥ 100 km³ granodioritic (i.e. dacitic) magma with more than 2 wt.% H₂O can produce large melt-rich caps at the top of partially

crystallised magma reservoirs in few hundreds to few thousands of years. The formation of separate melt lenses could become more likely when segregation of melt starts at high crystallinities (i.e. >0.6).

Our results suggest that the initial water content of magma may play an important role in modulating the distribution of eruptible melt in upper crustal reservoirs. Reservoirs of felsic and water-poor magmas (<2 wt.%) tend to be associated with the formation of isolated melt pockets of crystal poor and eruptible magma, which could account for the often-observed geochemical heterogeneity of the products of large caldera forming eruptions in the Snake River Plane. The limited dimensions of these eruptible magma pockets make their detection by geophysical methods challenging.

4.2 Introduction

The accumulation and storage of viscous high-silica melt in shallow magmatic reservoirs is a process preceding some of the largest explosive eruptions on Earth. The distribution of crystal poor and eruptible lenses of magmas within highly crystallised magma reservoir cannot be resolved by geophysical methods for intrinsic limit of spatial resolution (Bedrosian et al. 2018, Huang et al. 2015). The timescales of extraction of viscous felsic melt from highly crystallised magmas are also only broadly constrained varying from months to thousands of years (Bachmann & Huber 2018, Druitt et al. 2012, Gualda et al. 2018, Huber et al. 2012, Wilson & Charlier 2016). Here we use a field example of extraction of felsic melt from crystallising magma in combination with thermal modelling to investigate the role of water content on the distribution and timescales of extraction of felsic melts from crystallising felsic magma.

Segregation of interstitial melt from a rheologically-locked partially-crystallised magma body is a potential mechanism for the accumulation of crystal-poor and eruptible rhyolite (Bachmann & Bergantz 2004, Dufek & Bachmann 2010, Hildreth 2004, 1981, Hildreth & Wilson 2007, Marsh 1981). Thermo-mechanical simulations suggest that the efficiency of melt extraction for common hydrous silicic magma compositions is highest at crystal contents between 50% and 70% (Dufek & Bachmann 2010). Moreover, these studies emphasise that the probability of extraction and the amount of interstitial melt segregated is not only controlled by the physical properties of residual melt and magma, but also by the time spent by magma at conditions best suited for melt extraction (Dufek & Bachmann 2010, Huber et al. 2009). This, in turn, is a function of the ratio between the rate of heat loss and the rate of release of latent heat of crystallisation during progressive cooling and solidification of magma reservoir, and evolution of the physical properties of the residual melt and magma with increasing crystallinity (Caricchi & Blundy 2015a, Huber et al. 2009, Lee & Morton 2015, Melekhova et al. 2013). The results of these studies permit to

draw some general conclusions about extraction of residual melt in felsic systems: i) Independently of the process leading to the extraction of residual melt in crystallising felsic magmas, the separation between residual melt and crystals occurs when magma is rheologically locked (i.e. crystal fraction >0.4 ; Dufek & Bachmann 2010, Huber et al. 2010, Marsh 1981); ii) The velocity of residual melt extraction is directly proportional to the ratio between the density difference of crystals and residual melt and the viscosity of the residual melt (Bachmann & Bergantz 2004, Dufek & Bachmann 2010); iii) The longer magma spends at conditions suitable for residual melt extraction, the larger is the amount of residual melt extracted (Dufek & Bachmann 2010, Huber et al. 2009).

Large-scale segregation of rhyolitic melts from highly crystallised magmas is commonly inferred to precede the eruption of rhyolitic magmas (Bachmann & Bergantz 2004, Deering et al. 2011, Hildreth & Wilson 2007), however, evidences for such segregation processes remains scarce or obscure in the intrusive record (Coleman et al. 2004, Gelman et al. 2014, Lee & Morton 2015, Vigneresse 2014). In recent years, various studies have targeted melt extraction processes in the geological record using radiogenic isotopes (Andersen et al. 2017), whole-rock and mineral chemistry (Barnes et al. 2017, Hartung et al. 2017), and rock fabrics (Holness et al. 2017). The Takidani pluton in Central Japan shows evidence of melt segregation processes and is used here as a case study to investigate the formation of bodies of crystal-poor and eruptible melt in the upper crust (Hartung et al. 2017). This pluton has been shown to present the source of dacitic and rhyolitic volcanic eruptions (Harayama 1992, Kimura & Nagahashi 2007, Nagahashi et al. 2000). The pluton is texturally zoned, with a gradual transition (over about 50 m) from equigranular granodiorite to porphyritic granite (Figure 4.1). Whole rock and mineral chemistry suggest that the porphyritic unit (pGT) was extracted from underlying granodiorite (GDT) once the residual melt fraction dropped to 40–50 wt.% (Figures 4.1, 4.2a; Hartung et al. 2017). The initial water content of the granodiorite associated with melt segregation is estimated between 3 and 4 wt.% H_2O based on mineral chemistry, crystallisation sequence and relative abundance of the mineral phases (Hartung et al. 2017).

In this study, we focus on the effects of the initial water content on the timescales the magma spends at crystallinities larger than about 40 wt.%, which are considered favourable for the extraction of residual melt (Bachmann & Bergantz 2004, Dufek & Bachmann 2010). We first calculate the evolution with temperature and crystallinity of the physical properties of magma and residual melt for granodioritic (i.e. dacitic) magmas with water contents (H_2O_i) between 1 and 6 wt.%. The velocity of melt extraction for crystallinities between 40 and 80 wt.% are estimated using hindered settling (Davis & Acrivos 1985) and compaction-driven segregation (McKenzie 1984). To constrain the maximum timescales available for segregation to occur we perform thermal modelling for reservoirs of different volumes and shapes. The aims of this

study are (1) to constrain the effect of H_2O_i content on the efficiency of melt extraction, (2) to identify the conditions that led to the extraction of residual melt from the Takidani pluton and estimate the timescales of this process, (3) to define the impact of H_2O_i content on the architecture of upper crustal magma reservoirs.

4.3 Material and methods

4.3.1 The Takidani Pluton: Evidences for melt segregation

In the following we provide a summary of the main results of a geochemical study previously performed on the Takidani pluton (Hartung et al. 2017), which provides the background and motivation for this study. The Takidani pluton is a well exposed and young pluton (1.6 Ma; Harayama 1992, Ito et al. 2017), located in the Central Japan Alps. The pluton is vertically exposed over 1800 m (Harayama et al. 2003) from a tectonic contact at the base to a magmatic roof contact with older volcanic rocks (i.e. Hotaka Andesite; Harayama 1994). Textural, chemical and isotopic evidence of large-scale melt segregation is observed in the upper part of the pluton (Figures 4.1, 4.2; Hartung et al. 2017). The rock textures of the Takidani pluton change from holocrystalline to progressively more porphyritic appearance from the base and centre to the roof of the intrusion. Whole-rock geochemistry shows that the rocks immediately below the porphyritic unit are depleted in incompatible elements, while the porphyritic unit is enriched in incompatible elements (Figure 4.1; Hartung et al. 2017). Data obtained through quantitative evaluation of minerals by QEMSCAN and electron microprobe analyses (EPMA) are used to determine the area percent and chemical composition of the matrix components (equivalent to residual melt composition) throughout the upper section of the Takidani pluton, where evidence for melt extraction is observed (Figure 4.1; Appendix C: Table C.1). These data show progressive enrichment in the residual melt components defined by quartz (Qtz), alkali feldspar (Kfs) and albite-rich plagioclase ($Plg < An_{30}$) from the equigranular granodiorite (GDT) to the porphyritic granite (pGT; Figure 4.1; Appendix C: Table C.1). The relative amounts of quartz (Qtz), albite (Ab), and orthoclase (Or) and therefore the melt compositions, however, do not change across the textural and chemical transition (Figure 4.1; Appendix C: Table C.2). This suggests that the residual melt had a chemical composition close to the granitic minimum after emplacement at approximately 200 MPa (Johannes & Holtz 1996) and that the extraction of the residual melt, now represented by the porphyritic unit, occurred once magma crystallised sufficiently for the residual melt to acquire a composition close to the granitic minimum (Figure 4.1). Mineral chemistry provides additional support for the extraction of residual melt once the magma was rheologically locked (Hartung et al. 2017).

Plagioclase phenocrysts in the granodioritic unit below the porphyritic have cores of different chemistry that are overgrown by a distinct rim of common composition,

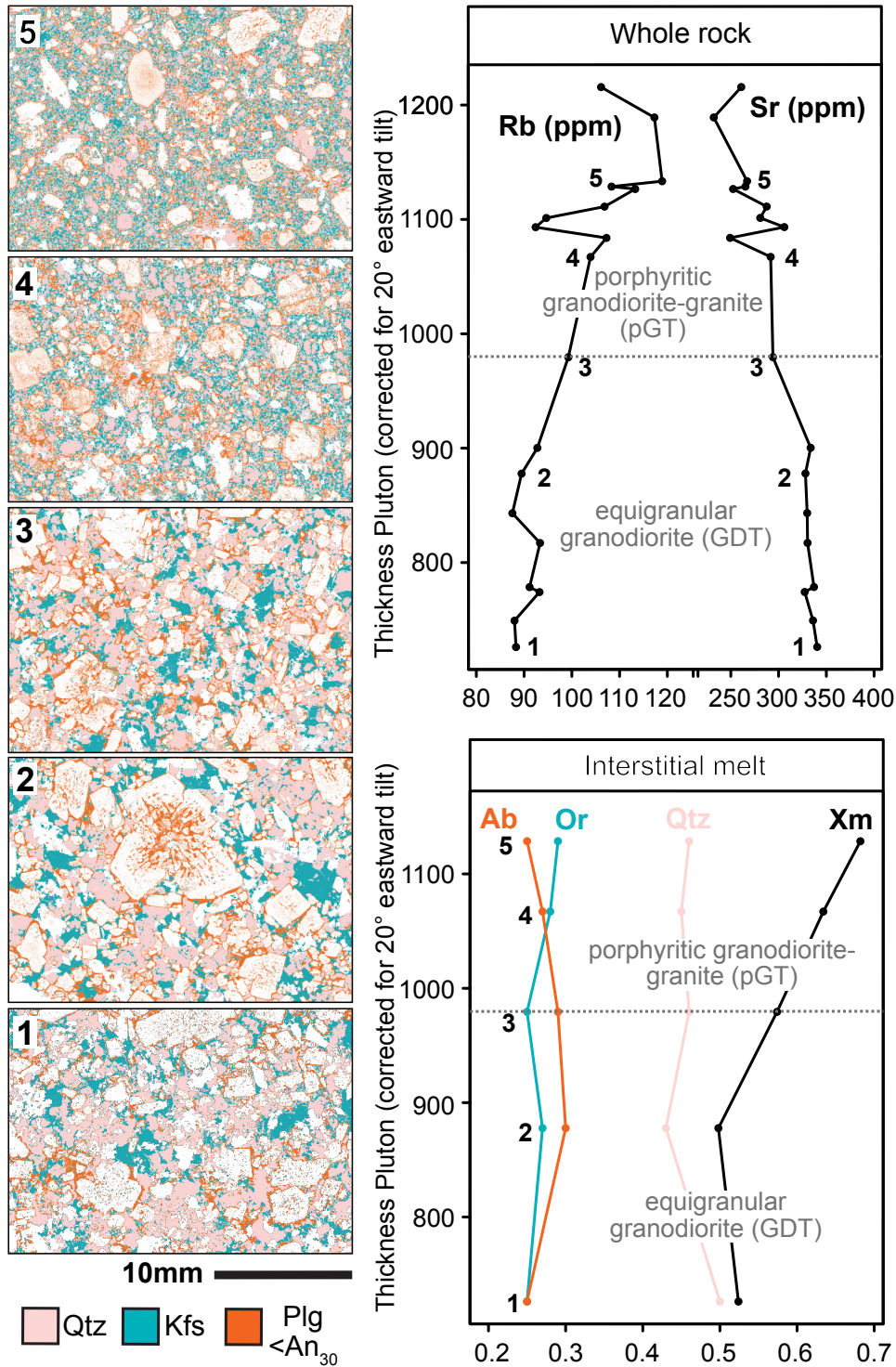


FIGURE 4.1: Evidence for melt segregation in the upper section of the Takidani Pluton. (left) QEMSCAN images collected from the granodiorite (1; GDT) to the porphyritic portion of the pluton (5; pGT). (top right) Variations of Rb and Sr whole-rock content from GDT to pGT. (bottom right) Interstitial residual melt variations represented by quartz (pink), orthoclase (green) and plagioclase with anorthite content <30 wt.% (orange). The fraction of interstitial residual melt (x_{melt}), gradually increases from the GDT unit towards the pGT unit. The content of quartz, orthoclase and albite, normalised to a fraction of 1, remain relative constant and suggest that the residual melt composition was buffered at the granitic minimum (Johannes & Holtz 1996).

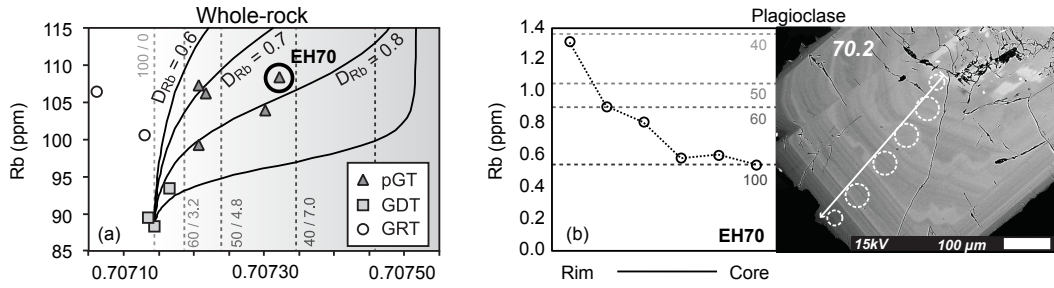


FIGURE 4.2: Chemical variability in whole rock (a) and mineral chemistry (b). (a) Assimilation and fractional crystallisation (AFC) models (Hartung et al. 2017) performed on whole rock analyses show that compositional diversity is dominantly produced by crystal fractionation. The grey dashed lines and numbers on the side show the amount of melt (i.e. 60 wt.%) and assimilation (i.e. 3.2 wt.%). The evolution of Rb concentration as function of $^{86}\text{Sr}/^{87}\text{Sr}$ is calculated for different bulk partition coefficients (D_{Rb}). (b) Concentrations of Rb in plagioclase phenocrysts (i.e. sample EH70) increase from 0.54 to 1.31 ppm from core to rim, respectively, and point towards a progressive enrichment of the incompatible element Rb in the melt phase through crystal fractionation. The grey dashed lines and number below indicate the amount of melt fraction.

which is consistent with the composition of plagioclase in the matrix of the porphyritic unit (Hartung et al. 2017). An increase in Rb (incompatible during crystallisation of the Takidani granodiorite; Hartung et al. 2017) in plagioclase phenocrysts increases by more than a factor of two from the core to the common outer rim. This suggests that once magma crystallised to approximately 40–50 wt.%, plagioclase acquired the same composition indicating rheological locking (Figure 4.2). Within this crystallinity range the residual melt was extracted leading to the formation of the porphyritic unit of the Takidani pluton that contains matrix plagioclase with the same composition of the plagioclase rims of the underlying granodioritic unit.

Inclusions of orthopyroxene in amphibole, amphibole thermometry, plagioclase composition, and comparison with the phase equilibria experiments of Costa et al. (2004), highlight that amphibole became stable once the magma achieved crystallinities of 40–50 wt.% (Hartung et al. 2017). The late appearance of amphibole at low temperatures and core to rim plagioclase chemistry indicate that the Takidani magma was not initially water-saturated (Costa et al. 2004) and contained initially between 3 and 4 wt.% H₂O.

On the basis of the collected evidences we conclude that the pGT unit represents a lens of residual melt extracted from the underlying GDT granodiorite. The extraction of residual melt from the crystallising magma occurred after the magma became

rheologically locked, at which point the residual magma had reached water saturation (Costa et al. 2004, Hartung et al. 2017). The evidence supporting initial water undersaturation of the magma and melt extraction for the Takidani pluton motivate us to investigate the role of initial water content on the extraction of felsic melt from crystallising magmas.

4.3.2 Rhyolite-MELTS simulations

Existing experimental data do not cover the entire range of temperature and water content required to trace the evolution of residual melt during cooling and crystallisation of magma in the upper crust (Costa et al. (2004), Holtz et al. (2005), Scaillet & Evans (1999)). Thus, we use rhyolite-MELTS (Gualda et al. 2012) to calculate the chemical and physical evolution of residual melt of dacitic (or granodioritic) magma from liquidus to near solidus temperature and over the entire range of initial water content between 1 wt.% and water saturation. Because granodiorites represent about 95 wt.% of the upper crust (Rudnick & Gao 2003) with a composition similar to the starting material of Costa et al. (2004), we use the latter for the rhyolite-MELTS calculations. While rhyolite-MELTS it currently not capable of correctly identifying the stability of hydrous phases such as amphibole and biotite, here we use rhyolite-MELTS, especially, to trace general evolution of residual melt chemistry and magma crystallinity. To test the performance of rhyolite-MELTS we compare the calculated residual melt compositions with residual melts produced experimentally by Costa et al. (2004) using the same starting composition. For all calculations, the confining pressure was fixed at 200 MPa, which are the conditions applied in the experiments of Costa et al. (2004) and comparable to the inferred emplacement depth of many granitic intrusions including the Takidani pluton (Hartung et al. 2017). The oxygen fugacity was initially set to the nickel-nickel oxide buffer (NNO) to calculate the liquidus temperature (for different H_2O_i), but remained unconstrained during progressive heat extraction. The evolution of the residual melt chemistry as function of temperature and initial water content calculated with rhyolite-MELTS is comparable to that obtained in the experiments of Costa et al. (2004; Figure 4.3). Importantly, rhyolite-MELTS can accurately reproduce the evolution of silica content of the residual melt, which together with H_2O plays a dominating role in controlling the viscosity of melts (Hess & Dingwell 1996). Additionally, the decrease of melt fraction with temperature for different initial water content follows paths that are comparable to those determined experimentally by Whitney (1988; Figure 4.4b), hence we consider the rhyolite-MELTS calculations appropriate to compare the evolution of physical properties of hydrous magmas during progressive crystallisation.

To quantify the time magma spends within different temperature interval we perform rhyolite-MELTS simulations by removing an equal amount of enthalpy (i.e. 1 J g^{-1}) at constant pressure from the system at each step (n) starting from the liquidus temperature down to a temperature (T) of about 740°C , which corresponds to a

residual melt fraction of about 0.1. If a fixed amount of enthalpy is extracted from the magma during solidification, which would be appropriate for quasi-eutectic magma compositions (Gualda et al. 2018), the number of modelling steps within a given temperature interval becomes proportional to the time spent by the magma within a given interval of temperature. This is important to be able to quantify the total time that magma spends within the temperature and crystallinity range at which the conditions are most conducive to melt extraction (Dufek & Bachmann 2010). Considering that the rate of heat loss will decrease during magma crystallisation because of the decreasing temperature difference between magma and host rocks, our approach increasingly underestimates the time spent within a given temperature range. This is especially the case at intermediate to high crystallinities, where segregation of interstitial melt takes place. The differences between this scenario and one more appropriate for non-eutectic magmas, for which the non-linear rate of heat release over time must be considered, is addressed by thermal modelling.

4.3.3 Thermal models

To provide constraints on the timescales of melt extraction, we use thermal modelling to compute the temporal evolution of temperature in instantaneously emplaced cylindrical magma bodies of various volumes, aspect ratios and initial water contents. We solved the two-dimensional axisymmetric formulation of the heat conduction equation, which can be written as:

$$\rho c \frac{\delta T}{\delta t} = \frac{1}{r} \frac{\delta}{\delta r} \left(r k \frac{\delta T}{\delta r} \right) + \frac{\delta}{\delta z} \left(k \frac{\delta T}{\delta z} \right) + \rho L \frac{\delta x_c}{\delta t} \quad (4.1)$$

where T is the temperature, t is the time, r is the radial coordinate relative to the symmetry axis, z is the depth, k is the thermal conductivity, L is the latent heat of crystallisation, ρ is the density, c is the specific heat and X_c is the fraction of crystals in the magma. For a list of the parameters used in the modelling the reader is referred to Table C.3. The calculations were performed on a numerical grid using an explicit finite difference discretisation of the above equation. The model considers latent heat of crystallisation, which was implemented using our rhyolite-MELTS results to parametrise the crystal fraction (X_c) versus temperature for the different initial water contents used in this study (Figure 4.5a). As the dependence of X_c on T is non-linear, the governing equation was solved using an iterative strategy. In all models, we integrated an initial geothermal gradient of 25°C km^{-1} and a temperature-dependent thermal conductivity (k) as described in Whittington et al. (2009) for average crust. The upper (i.e. surface) and lower boundary (i.e. 25–35 km) in the models was set to a fixed temperature determined by the initial geothermal gradient, while zero flux boundary conditions were imposed on the left and right side. Cooling due to circulation of hydrothermal fluids around the magma reservoir was neglected in

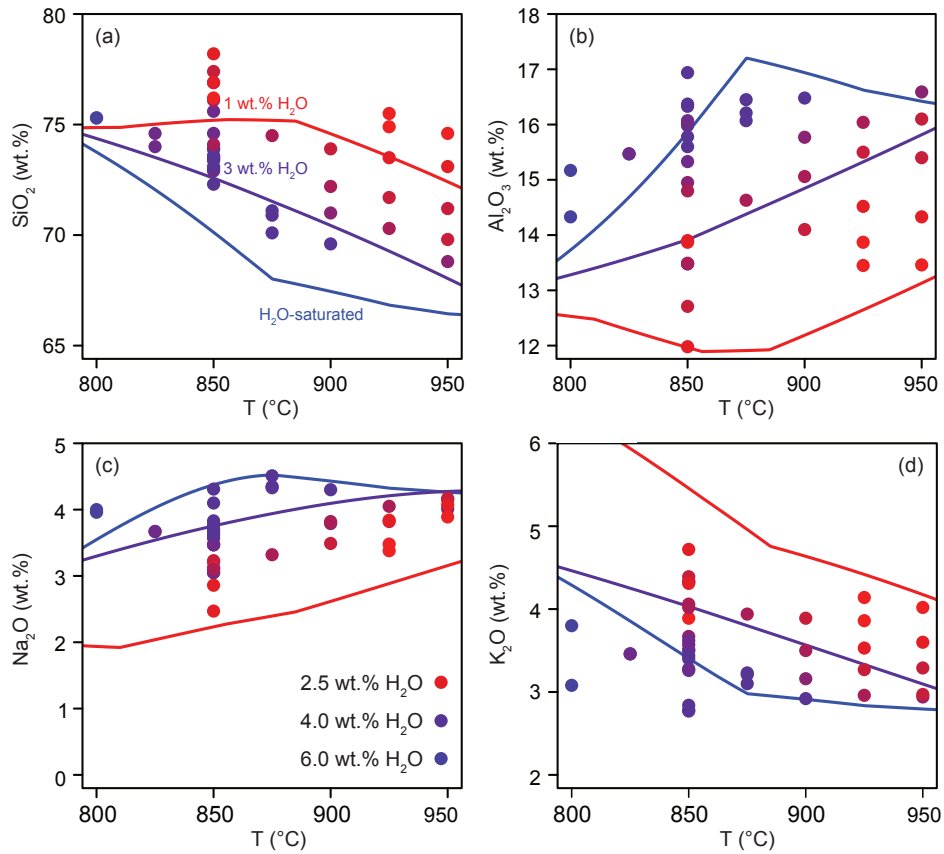


FIGURE 4.3: Comparison between rhyolite-MELTS simulations (lines) and the matrix glass compositions (circles) measured between 950°C and 800°C by Costa et al. (2004). The colour contouring indicates the initial water content of the starting material (Costa et al. 2004). Rhyolite-MELTS and experiments are in broad agreement and show the effect of initial water content on the chemical evolution (i.e. SiO₂, Al₂O₃, Na₂O, K₂O) of residual melt with decreasing temperature. No experimental data is available below 800°C.

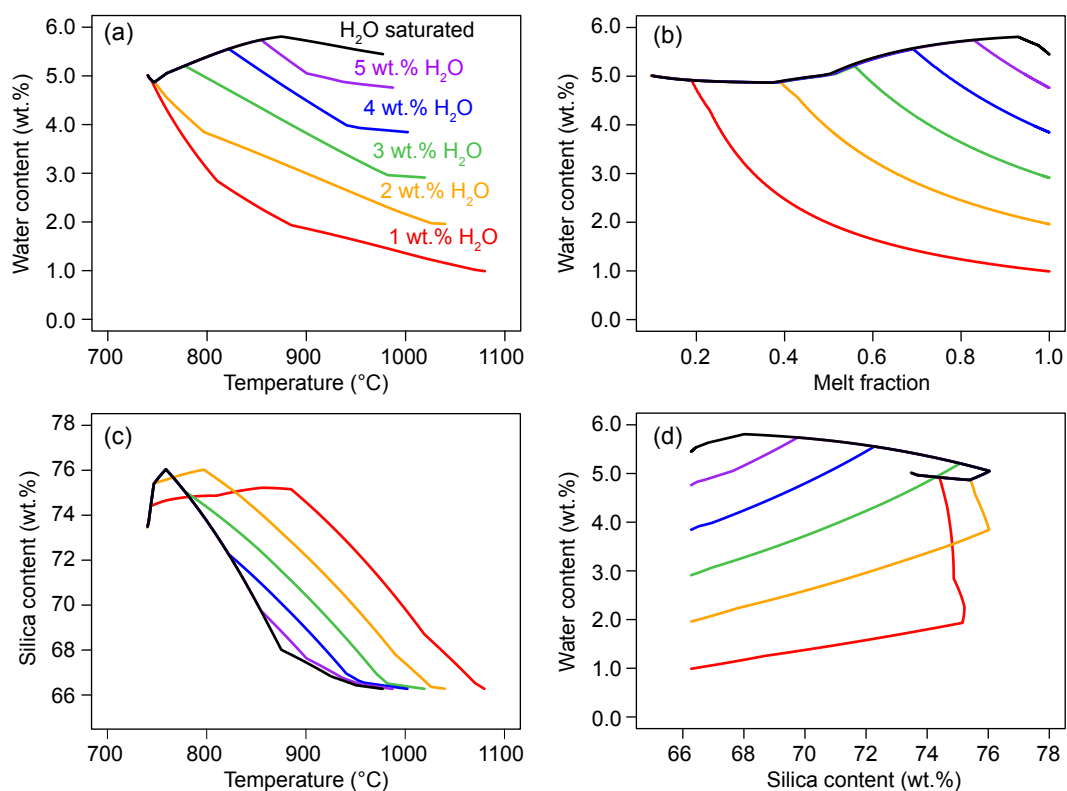


FIGURE 4.4: Results of rhyolite-MELTS simulations for dacitic magma with composition similar to the Takidani granodiorite and the starting material of Costa et al. (2004). (a) Evolution of the water content of the residual melt as function of temperature. (b) Relationship between water content of the residual melt and melt fraction. (c) Silica content of the residual melt versus temperature. (d) Water content versus silica content of the residual melt.

the modelling and the potential effects of this simplification are discussed in the following.

The setup consists of a 25 x 25 km domain for magma volumes of 100 and 1000 km³, and a 35 x 35 km domain for intrusion of 10.000 km³, using 400 x 400 numerical grid for all simulations (Figure C.1). In each model the magma body was intruded instantaneously at 10 km depth at its liquidus temperature and crystallised to near solidus conditions, as determined by the rhyolite-MELTS simulations. To test the effect of reservoir shapes on the solidification history, we ran models with intrusion aspect ratios of 2 and 10 for magma volumes of 100 and 1000 km³, while model pluton volumes of 10.000 km³ were only performed with an aspect ratio of 10, due to the large vertical extent of such reservoirs, which makes the application of a single melt fraction-temperature relation unreasonable. Finally, to compare melt segregation velocities to solidification timescales we tracked the temporal propagation of isotherms and melt fraction with depth in the centre of the intrusion.

4.4 Results

4.4.1 Thermal, chemical and physical evolution of dacitic magma

Water has an important effect on phase equilibria as it depresses liquidus temperatures and modifies the relationships between temperature, crystallinity, and the physical properties of magmas over a wide range of chemical compositions (e.g. viscosity and density of the residual melt; Blatter et al. 2013, Caricchi et al. 2007, Giordano et al. 2008, Lange 1994, Melekhova et al. 2013, Müntener & Ulmer 2018, Ulmer et al. 2018, Whitney 1988, Hess & Dingwell 1996). Crystallisation of magma leads to a non-linear increase of silica and H₂O content in the residual melt (Figure 4.4a-d). Once the residual melt becomes water saturated, *H₂O_i*-undersaturated magmas join the T-melt fraction trajectory of initially water-saturated magma (Figure 4.4b). The non-linear relationships between melt fraction and temperature, contributes to modulate both the timescales magma spends at different temperatures, and the temporal evolution of the physical properties of the residual melt during progressive magma crystallisation. In crystallising magma reservoirs, magma spends relatively more time at temperatures at which the rate of crystallisation (i.e. rate of latent heat release) is highest (Marsh 1981, Caricchi & Blundy 2015a). Thus, if the crystallisation rate increases once rheological locking is achieved, magma will spend relatively more time at these temperature conditions (Huber et al. 2009, Caricchi & Blundy 2015a). The time is even larger considering that the rate of heat release from magma reservoirs drops with cooling and progressive decrease of the thermal gradient between magma and host rocks.

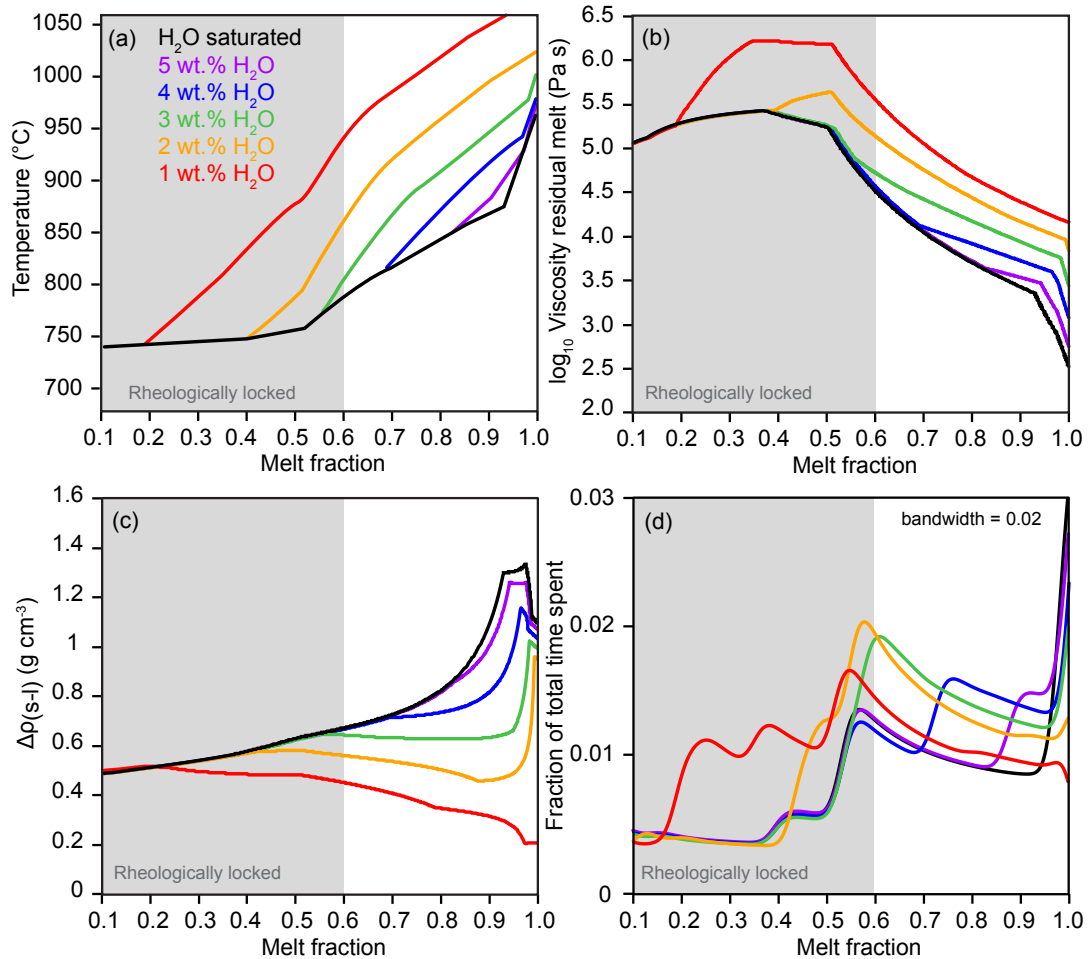


FIGURE 4.5: Physical melt fraction and physical properties of residual melt and magma calculated with rhyolite-MELTS. (a) Relationship between melt fraction and temperature. (b) Evolution of melt viscosity (logarithmic) as function of melt fraction. (c) Density difference between crystals and the residual melt as function of melt fraction. (d) Fraction of the total time spent by the magma within each interval of melt fraction. This is calculated by normalising the number of simulation step within a melt fraction unit of 0.02 to the total amount of steps between 1.0 and 0.1 melt fraction assuming constant heat loss.

Magmas with lower water contents will achieve rheological locking at higher temperatures with respect to more water-rich magmas, which will, in turn, impact viscosity, density and velocity of residual melt extraction (Figure 4.5b, c; Bachmann & Bergantz 2004). Our calculations show that regardless of the initial water content of the magma, residual melt viscosity increases down to melt fractions of 0.5 to 0.4 because of decreasing temperature and increasing silica content. At lower melt fractions (<0.4) and once volatile saturation is achieved, the viscosity of the residual melt remains relatively constant independent of the initial water content (Figure 4.5b). Relatively dry melts (H_2O_i 2 wt.%) reach a maximum in viscosity before joining the same trajectory of water-richer magmas at lower temperatures (Figure 4.5b). The contrast in density between the solid phase and residual melt spans a wide range of values near liquidus conditions (for different H_2O_i) but becomes less dependent on H_2O_i for melt fractions <0.6 (Figure 4.5c). The ratio of the difference in density between crystals and residual melt and the viscosity of the residual melt, which is directly related to the velocity of melt extraction (Bachmann & Bergantz 2004), generally increases with water content (Figure 4.5b, c).

Based on the physical properties of residual melt and magma, the melt extraction velocity is the fastest for H_2O_i -saturated magmas. However, the total time spent by magma at melt fractions <0.6 (i.e. rheologically locked conditions) is inversely proportional to the initial water content (Figure 4.5d). To assess the relative importance of the initial water content on the physical properties of magma and the timescales available for melt segregation to occur, we calculate the velocity of melt extraction by hindered settling (Bachmann & Bergantz 2004) and compaction (McKenzie 1984) for granodioritic-dacitic magma for melt fraction <0.6 . We notice that the formulations used to calculate the velocity of melt extraction are rather simplified, however, such order of magnitude estimates are important to compare the efficiency of residual melt for magmas with different H_2O_i . We do not consider the presence of excess magmatic fluids (i.e. H_2O , CO_2 , S) during melt extraction, which can have both positive and detrimental effects on the efficiency of melt extraction (Boudreau 2016, Caricchi et al. 2018, Parmigiani et al. 2016, Pistone et al. 2015, Sisson & Bacon 1999).

4.4.2 Velocities of melt segregation

Hindered settling and compaction-driven segregation describe the relative motion and separation of liquid (i.e. melt) and solid (i.e. crystal) in a two-phase system (i.e. crystallising magma). The process of particle settling in a monodisperse suspension is described by hindered settling (Equation 4.2). Compaction-driven segregation defines the process of compacting a porous crystalline mush and resulting melt expulsion (Equation 4.3). A detailed evaluation of hindered settling and compaction-driven segregation processes is presented in Bachmann & Bergantz (2004) and Lee & Morton (2015).

To obtain velocities for hindered settling (U_{hs}) and compaction-driven segregation (V_{comp}) for crystallising dacitic magma we use the physical properties obtained from rhyolite-MELTS simulations. We calculate segregation velocities following the equations provided by Bachmann & Bergantz (2004; and references therein):

$$U_{hs} = \frac{2r^2 g \Delta\rho}{9\mu} \frac{(1-c)^2}{(1+c^{\frac{1}{3}})^{\frac{5}{3}} c^{(1-c)}} \quad (4.2)$$

$$V_{comp} = \kappa(1-\phi)\Delta\rho \frac{g}{\mu\phi} \quad (4.3)$$

Where r is the crystal radius, g is gravitational acceleration (9.81 m s^{-1}), $\Delta\rho$ is the density difference between crystal and melt, μ is the viscosity, c is the crystal fraction, κ is the permeability and ϕ is the porosity. The permeability is calculated from the porosity, crystal radius and permeability coefficient (K) after McKenzie (1984):

$$\kappa = \frac{\phi^3 r^2}{K(1-\phi)^2} \quad (4.4)$$

We consider that porosity is equal to melt fraction and we use K values between 50 and 200 (Bachmann & Bergantz 2004, Pistone et al. 2015). Here we assume monodisperse suspensions of melt and crystals. As magmas contains crystals of different sizes, and this tends to reduce the permeability with respect to monodisperse systems (Bachmann & Bergantz 2004, Rust & Cashman 2011), the volumetric rates of melt extraction calculated here for each grain size are maximum estimates.

To calculate and compare segregation velocities for dacitic magmas with different initial water content, we first assume that no convection occurs and that magmas are rheologically locked at crystallinities $>40 \text{ wt.}\%$. Based on Equation 4.2, velocities for hindered settling vary between 3.6 and 0.06 m yr^{-1} for water-saturated magmas and melt fractions decreasing from 0.6 to 0.2 (Figure 4.6a) using a crystal radius of 3 mm . Velocities for under-saturated magmas are lower (i.e. $1 \text{ wt.}\% H_2O_i$; Figure 4.6a) and decrease from about 0.22 to 0.02 m yr^{-1} with increasing crystallinity. Segregation velocities for $H_2O_i \geq 3 \text{ wt.}\%$ increase in similar fashion as water saturated melts within the rheologically locked interval (Figure 4.5a, 4.6a). Melt segregation of undersaturated magmas with water content $<3 \text{ wt.}\%$ occurs on timescales up to one magnitude slower as they become water saturated at lower temperature and at higher crystallinities resulting in slower melt extraction velocities. The timescales of melt extraction processes strongly depend on the crystal size and vary by more than one order of magnitude when considering crystal radii of 1 to 5 mm (Figure 4.6b).

We also calculate the velocity of melt extraction considering compaction using a

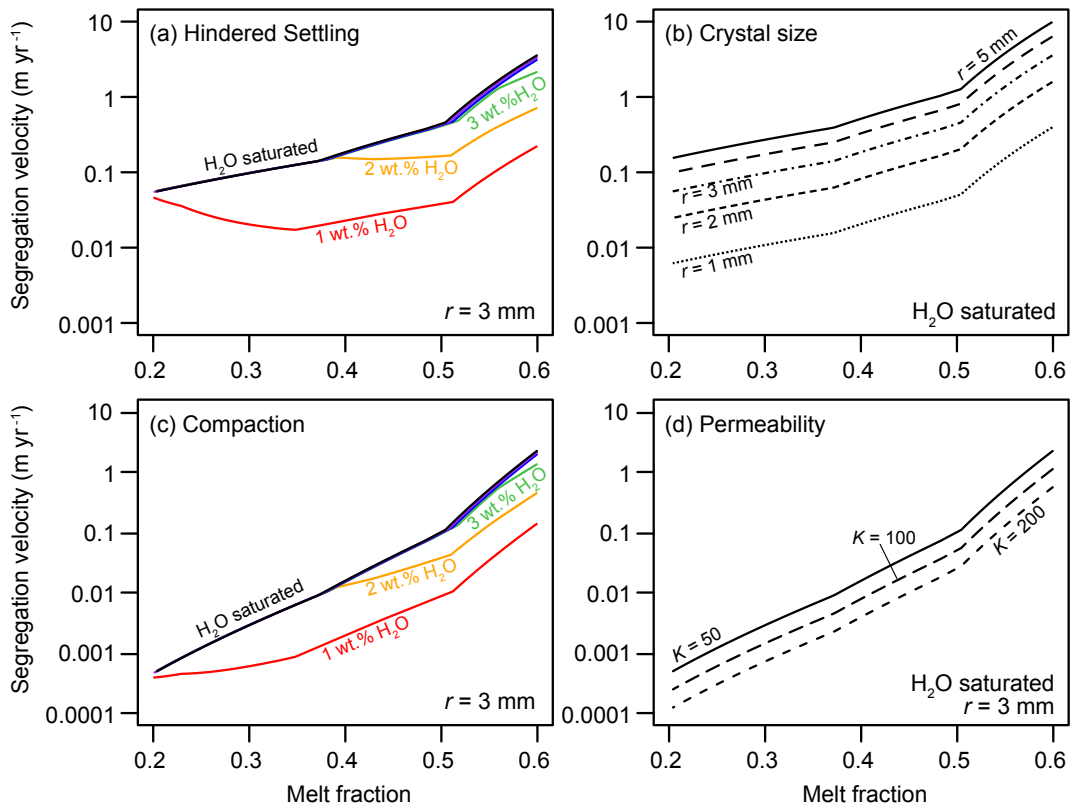


FIGURE 4.6: Velocities of melt segregation as function of the melt fraction for magma with different water content for hindered settling (a, b) and compaction-driven segregation (c, d). (a) Hindered settling velocity for different initial water contents (H_2O_i) and crystal radius of 3 mm. (b) Hindered settling velocity for water saturated magma containing crystals of different radii (r). (c) Velocity of compaction-driven segregation for different initial water content and magma with crystal of 3 mm radius. (d) Compaction-driven segregation velocity for water-saturated magma, calculated for different permeability coefficients (K) and crystal radius of 3 mm. Segregation velocities are calculated using the physical properties obtained from rhyolite-MELTS simulations.

permeability coefficient of 50 and a crystal radius of 3 mm and obtain values between 2.2 m yr^{-1} and $5.0 \times 10^{-4} \text{ m yr}^{-1}$ for water saturated magma and between $1.4 \times 10^{-1} \text{ m yr}^{-1}$ and $3.9 \times 10^{-4} \text{ m yr}^{-1}$ for water undersaturated magma. Segregation velocities decrease up to one order of magnitude when applying larger permeability coefficients (i.e. 100–200; Figure 4.6d). Velocities of melt extraction for compaction-driven segregation and hindered settling are similar at melt fractions between 0.6 and 0.5, however, compaction velocities decrease more rapidly with decreasing melt fraction than hindered settling (Bachmann & Bergantz 2004, Lee & Morton 2015). Below melt fractions of 0.2, the velocity of compaction driven segregation has been shown to increase relative to hindered settling (Lee & Morton 2015). Our results show that residual melt separates most effectively from highly crystallised dacitic magmas if the initial water concentration is equal or greater than 3 wt.%, independently of the extraction mechanisms considered (Figure 4.6). As phase equilibria and H_2O solubility are both affected by pressure, 3 wt.% H_2O_i is a threshold value appropriate for confining pressures of 200 MPa. The increase of H_2O solubility with pressure could increase the efficiency of melt-extraction at mid to deep crustal levels (i.e. 15–20 km depths), where potentially larger H_2O_i would result in lower melt viscosity and accelerate segregation processes. At such depths, the rate of heat release from magma reservoir is also lower, which would also increase the time available for segregation of residual melt.

4.4.3 Crystallisation timescales of hydrous felsic magma

The thermal modelling results allow us to constrain the crystallisation timescales of hydrous dacitic magmas at melt fraction between 0.6 and 0.2, where segregation is most efficient (Dufek & Bachmann 2010). We track the position of the isotherm corresponding to a given melt fraction along a vertical section through the middle of the intrusion. Each isotherm propagates from the base and the top of the intrusion toward its inner portions (Figure 4.7a). We define the “maximum segregation timescale” as the time difference between the moment the isotherms corresponding to melt fractions of 0.6 and 0.2 reach the centre of the intrusion (Figure 4.7b).

The thermal models show that magmas with low initial water contents (i.e. <3 wt.% H_2O_i) spend substantially more time at rheologically locked conditions than magmas that are initially water-saturated (Figure 4.8). For reservoir volumes of 100, 1000, and 10,000 km^3 , the maximum timescale for segregation increase from 13 kyr, 61 kyr and 274 kyr for water saturated magmas to 19 kyr, 84 kyr and 349 kyr, and 20 kyr, 94 kyr and 444 kyr for magmas with initial water contents of 3 and 2 wt.% H_2O_i , respectively. The timescales available for melt segregation increase by a factor of 4 to 5 with decreasing aspect ratios from 10 to 2 (Figure 4.8).

To estimate the maximum melt migration distances, we first calculate the average segregation velocity of hindered settling and compaction (Equation 4.2, 4.3) for each

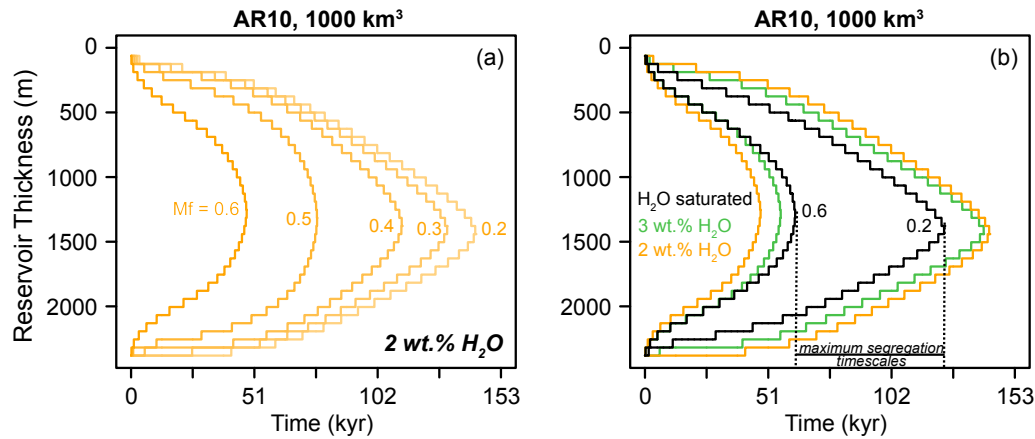


FIGURE 4.7: (a) Vertical position of the isotherms, corresponding to different melt fractions (M_f), along a vertical section through the centre of the intrusion as function of time. Reservoir thickness of 0 m presents the top. Location of isotherms are shown for a dacitic magma with initial water content of 2 wt.%. (b) Maximum segregation timescales for magma reservoirs with different initial water content. Isotherm curves are shown for a magma reservoir of 1000 km³ with an aspect ratio (AR) of 10. The Maximum segregation timescale increases with decreasing initial water content of the magma.

melt fraction interval of 0.1 between 0.6 and 0.2. Secondly, each of these values are multiplied by the time required by the corresponding isotherm to reach the core of the intrusion, once the melt fraction within the reservoir has reached 0.6 (Figures 4.7a, 4.9a,b). Because of the relatively high segregation velocities at melt fractions between 0.6 and 0.5, the residual melt can migrate large distances over the duration of isotherm propagation from 0.6 to 0.5 (Figures 4.6, 4.9a,b). For example, maximum segregation distances fall in the order of tens of kilometre for a magma reservoir with a volume of 1000 km³ (Figure 4.9a,b). The timescales for melt to travel to the roof are fastest for water saturated magma increasing from 600 to 2500 yr for water-saturated magmas and magmas with an initial water content of 2 wt.%, respectively (Figure 4.9b). These values decrease to about 300 to 1200 yr for reservoir volumes of 100 km³ and increase to about 1300 and 5400 yr for magma reservoirs with a volume of 10.000 km³ (Figure 4.9b). For a reservoir of 1000 km³, the time required for the extracted melt to reach the roof by compaction-driven segregation is twice as long and range from 1300 to 5700 yr for water saturated and magma with 2 wt.% H₂O_i, respectively.

4.5 Discussion

4.5.1 Segregation timescales of the Takidani Pluton

Magmas of the Takidani Pluton were initially H₂O-undersaturated with water contents of about 3 to 4 wt.% (Hartung et al. 2017, Costa et al. 2004). During crystallisation

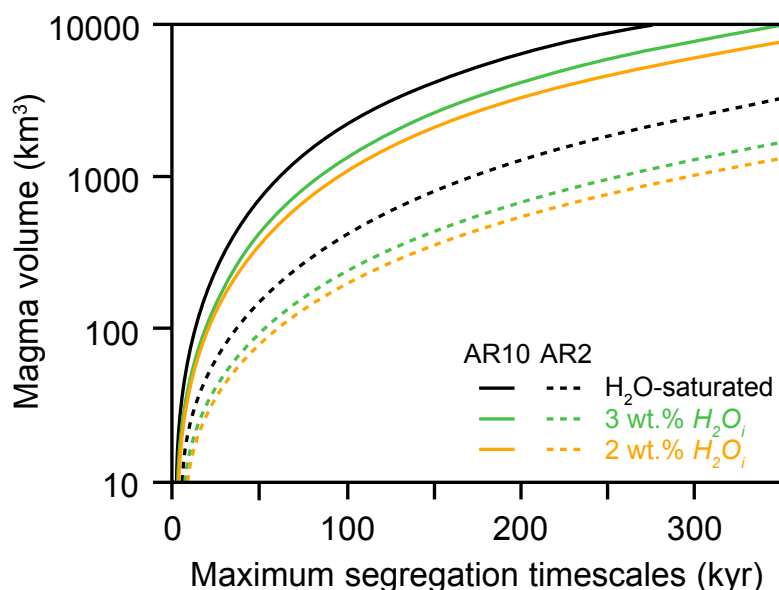


FIGURE 4.8: Maximum timescales available for segregation for different reservoir volumes, water contents and aspect ratios of the magma reservoir. Curves were obtained by fitting the thermal modelling results using 2nd and 3rd order polynomials (Table C.4).

and melt evolution the residual liquids would have reached volatile saturation at melt fractions between 0.70 (i.e. 4 wt.% H_2O_i) and 0.55 (i.e. 3 wt.% H_2O_i) and temperatures around 820°C to 780°C (Figure 4.5a,b). At this point an increase of crystallisation rate and release of latent heat of crystallisation would have resulted in near isothermal crystallisation. A potential indication of this process may come from amphibole thermometry, which suggest the crystallisation of this phase occurred almost entirely in between 800°C and 750°C (Hartung et al. 2017). This could support our hypothesis that magma was thermally buffered over long time periods by the release of latent heat of crystallisation.

Based on our rhyolite-MELTS simulations and thermal modelling we estimate the maximum and minimum timescales over which melt segregation took place in the Takidani Pluton. We consider that the column of melt (h) is equal to the thickness of the porphyritic unit (270 m; Figure 4.1, Hartung et al. 2017). Velocities of hindered settling segregation of magma with initial water content of 3 to 4 wt.% range from 3.1 m yr⁻¹ to about 0.06 m yr⁻¹ between 0.6 and 0.2 melt fraction for an intermediate crystal radius of 3 mm, which is considered appropriate for granitoids (Bachmann & Bergantz 2004, Lee & Morton 2015) including the Takidani pluton (Figure 4.1, Hartung et al. 2017). This would imply that the extraction of the melt producing the porphyritic unit of the Takidani pluton required between about 100 yr and 5000 yr. If we considered a smaller crystal size (i.e. 1 mm) the time required for segregation by hindered settling would increase by approximately one order of magnitude to 1000 yr

and 45 kyr. Velocities for compaction-driven segregation for the same water contents and crystal size (i.e. 3 mm) vary between 0.5 m yr^{-1} and 0.0001 m yr^{-1} leading to much slower extraction timescales of about 800 yr to 2 Myr.

The Takidani Pluton has an exposed horizontal extension of about 14 km and exposed vertical extension of about 2 km, which implies a minimum volume of 300 km^3 considering a cylindrical shape. The results of our thermal models show that the time available for segregation in the Takidani Pluton with magmas containing 3 to 4 wt.% H_2O_i varies between 40 and 120 kyr for aspect ratios of 10 and 2, respectively (Figure 4.8). The thermal modelling and velocity calculations we performed, show that compaction-driven segregation generally operates at timescales that are larger than estimated solidification timescales and therefore cannot explain the observed segregation in the Takidani Pluton. Hindered settling, on the other side, is a much faster process that can explain the observed melt segregation in the Takidani Pluton. Only in the eventuality that segregation would have started at melt fraction smaller than 0.3, the porphyritic unit of the Takidani pluton would not represent a melt-rich cap near the roof of the pluton. Considering that the base contact of the pluton is tectonic and therefore its volume could be considerably larger than 300 km^3 , our results show that sufficient time was available for melt segregation to occur and to form a melt cap via hindered settling (Figure 4.10) or processes that operate at similar timescales (i.e. gas filter pressing; Sisson & Bacon 1999). Moreover, evidence from quartz hosted fluid inclusion studies suggest that the Takidani Pluton developed a liquid-dominated hydrothermal systems at some point in the past (Bando et al. 2003, Sekine et al. 2001). Heat advection associated with hydrothermal circulation is not considered in our calculations but would have increased the rate of heat release depending on the vigour of fluid convection (Delaney et al. 1995). This directly decreases the time available for melt segregation (Dutrow et al. 2001), thus increasing the potential for melt to form isolated melt lenses instead of a large cap on top of magmatic reservoirs.

4.5.2 The control of H_2O_i on the extraction of residual melt from crystallising magmas

Geochemical and petrologic studies show that crystal poor rhyolites are sourced either from caps at the top of partially crystallised reservoirs (Bachmann & Bergantz 2004, Hildreth & Wilson 2007), or from the amalgamation of isolated melt pockets dispersed within a highly crystallised magma (Wotzlaw et al. 2014, Ellis et al. 2014). The processes responsible for the generation of reservoirs with such distinct architecture are not yet fully understood. The total distance over which melt migrates before the system cools to its solidus temperature, controls the final distribution of crystal-poor felsic melt in magmatic reservoirs. Thus, the maximum amount of rhyolitic melt that can potentially be accumulated depends on the initial water content and temperature of the magma, the volume and shape of the magma body and the temperature

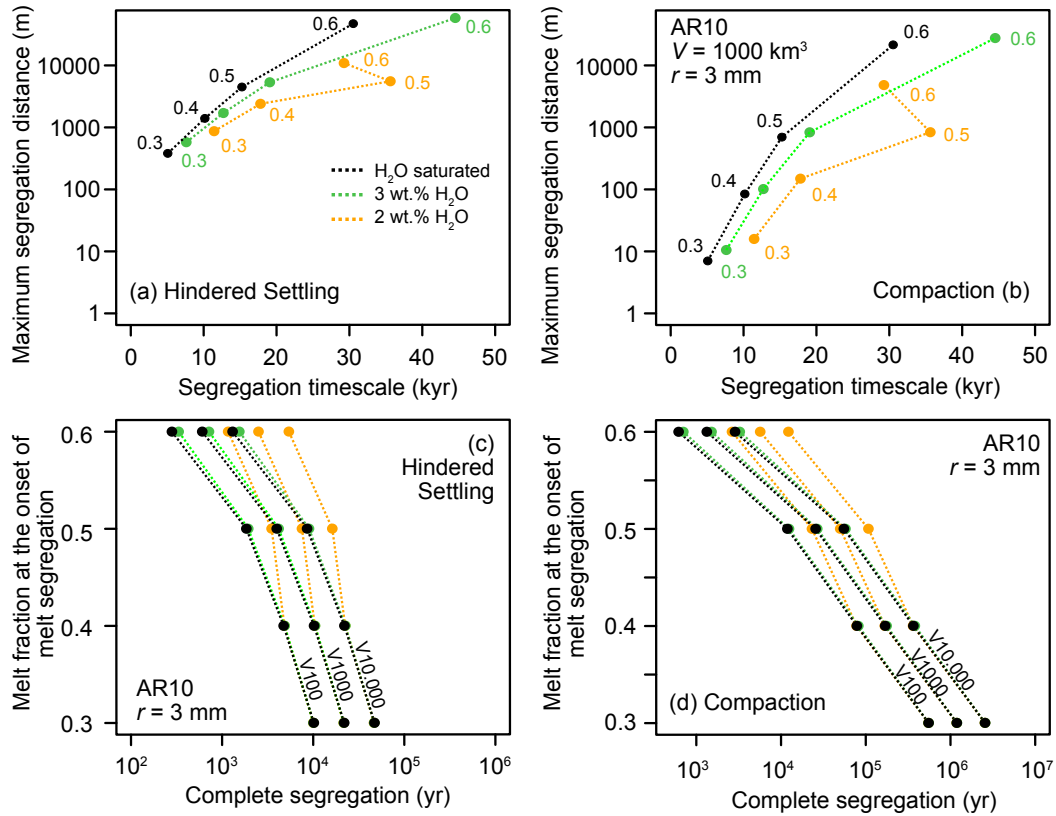


FIGURE 4.9: (a, b) Maximum segregation distance for a reservoir volume of 1000 km^3 and aspect ratio of 10 based on hindered settling and compaction velocities corresponding to an intermediate crystal radius (i.e. $r = 3 \text{ mm}$). (c, d) Time required by the residual melt (extracted by hindered settling and compaction) to reach the roof of the magma reservoir for reservoirs of 100, 1000 and $10,000 \text{ km}^3$. The y-axis presents the onset on melt segregation for intervals of 0.1 starting at 0.6, 0.5, 0.4 and 0.3.

difference of the intruding magma and the host rock (i.e. thermal gradient). We have illustrated that the initial amount of water dissolved in magmas affects (1) the physical properties and segregation velocities of residual melts (Figures 4.5, 4.6) and (2) the total amount of time spent within the rheologically locked temperature interval (Figures 4.7, 4.8; Caricchi & Blundy 2015a).

Our calculations show that crystal size is an important factor controlling segregation velocities of residual melt (i.e. Bachmann & Bergantz 2004), while the initial water content affect both the melt extraction velocity and the release of latent heat, which, in turn, modulates the time spent by the magma within different temperature intervals. Volume and shape of the magma reservoir affect the thermal evolution and timescales of reservoir solidification. In case of hindered settling, residual melts of magmas that contain at least 2 wt.% H_2O_i are extracted sufficiently fast to form melt rich caps at or near the roof of the magma reservoir (Figure 4.8) considering an intermediate crystal size of 3 mm. Magma reservoirs that have a minimum volume of 1000 km³ can facilitate complete melt segregation for smaller crystal sizes (Figure 4.10a). On the other hand, smaller magma bodies (i.e. 100 km³) with less than 2 wt.% H_2O_i are less likely to form any melt-rich body. Although magmas with 1 wt.% H_2O_i spent half of their solidification time within the rheological locking temperature window (Figure 4.5d), the high viscosity of the residual melt (i.e. low segregation velocity) does not favour the formation caps or melt-rich lenses of crystal poor rhyolite under the modelled conditions. The boundaries between caps and lenses, thus, may largely depend on the segregation velocity of the residual melt. Slower segregation of the residual melt, for instance through compaction or through hindered settling in magma with smaller crystals (<3 mm radius; Bachmann & Bergantz 2004), will decrease the probability of forming caps of crystal-poor rhyolitic melt and increase the probability of forming isolated melt-rich lenses.

The onset of melt segregation processes can play an important role in the formation of crystal-poor rhyolite (Figures 4.9, 4.10). The likelihood of forming a crystal-poor cap is very high when segregation processes start early within the rheologically locked crystallinity window, and segregation velocities remain elevated (Figures 4.6a, 4.10a). Unless magmas are extremely dry (1 wt.% H_2O_i), the initial water content of the magma does not seem to play a major role in controlling the architecture of the reservoir and melt distribution when melt extraction occurs at high melt fractions (i.e. 0.6; Figure 4.9a,b). The timescales of melt accumulation, however, are strongly dependent on the type of segregation process (i.e. hindered settling and compaction) and on the initial water content: melts containing higher amounts of water migrate faster (Figure 4.9a,b) because of their lower viscosities (Figure 4.5b). When the onset of melt segregation occurs at relatively low melt fractions (i.e. 0.3), the initial water content may influence whether caps or separate lenses are formed. Water undersaturated magmas may have a higher chance to form melt caps compared to

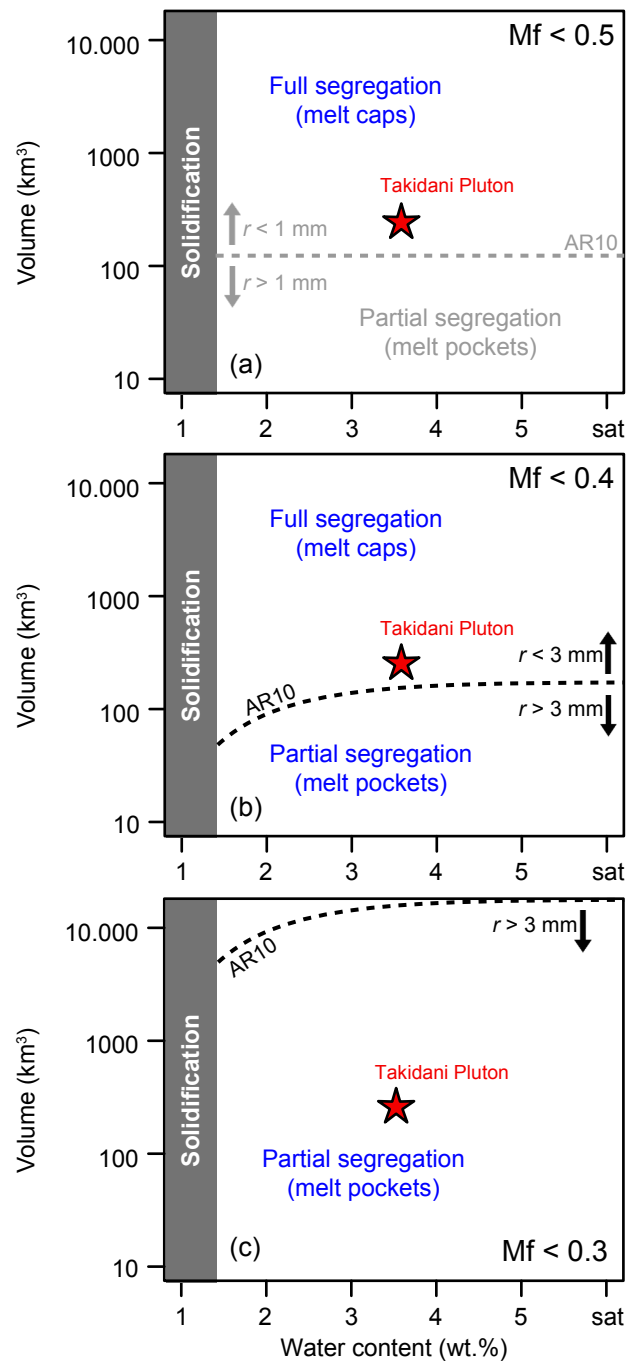


FIGURE 4.10: Schematic model for the formation of melt caps and melt pockets for different volumes and water contents for hindered settling. (a) Full segregation is expected in reservoir $>100\text{ km}^3$ for crystal radii $>1\text{ mm}$ for the case that segregation processes start at melt fraction of 0.5 or higher. (b) When melt segregation starts at $Mf < 0.4$ only reservoirs that are larger than 100 km^3 or that have low initial water contents are expected to form melt caps. (c) Partial melt segregation and the formation of crystal lenses are most likely to occur if segregation occurs at high crystallinity. Solidification of the reservoir without significant segregation is expected for magmas with water contents $\ll 2\text{ wt.}\%$. The Takidani Pluton most likely formed a coherent melt cap that may have been erupted.

magmas with higher water contents, because of their prolonged cooling timescales at rheologically locked conditions (Figure 4.10c). Overall, at low melt fraction timescales for melt segregation are short, which together with low segregation velocities decrease the capacity of these magmas to form thick melt rich bodies near the roof of the reservoir and the formation of isolated melt-rich lenses becomes more likely.

4.6 Conclusion

The interplay between magma water content, the viscosity of residual melt, the density contrast between residual melt and crystals, and the time spent within the rheologically locked crystallinity interval, favoured the extraction of residual melt from the Takidani pluton (Figure 4.5). Our time estimates for melt segregation indicate that the extraction of residual liquids in silicic reservoirs like the Takidani Pluton occurs over centuries and millennia which is consistent with similar studies (also see Bachmann & Huber 2018).

While the initial water content in magma has a significant control on the segregation timescales of magmatic reservoirs, the architecture of magmatic reservoirs is controlled by the magma crystallinity, crystal size and permeability at the onset of melt segregation processes. The results from our models suggest that hydrous felsic magma are prone to form large melt caps at the roof of a magma reservoir if segregation occurs early at intermediate melt fractions (e.g. 0.6–0.5). However, if segregation starts at low melt fractions (i.e. compaction-driven segregation), silica-rich melts may not be able to reach the roof of magma reservoirs because of their low segregation velocity and may form isolated melt lenses within highly crystallised magma. A similar reservoir configuration is likely for water-poor magmas, as residual melt segregation require longer timescales (Figure 4.7). The drainage of isolated crystal poor lenses of magma during eruption may account for the chemically heterogeneous nature of rhyolitic eruptions in relatively dry systems such as Yellowstone (Ellis et al. 2014, Wotzlav et al. 2014). We notice that our study focuses on felsic magmas characteristic of some of the large eruptions on Earth; the dependency of viscosity and density of magma and residual melt on chemistry does not warrant the extension of our conclusions to systems of significantly different chemistry.

The timescales of melt segregation vary by several orders depending on the crystal size and the mechanism that dominates the extraction process (Figure 4.6b). The extremely short timescales obtained from geochronology and geospeedometry (months to decades; Druitt et al. 2012, Gualda et al. 2018, Wilson & Charlier 2016), are difficult to explain by gravity-driven segregation even considering a large crystal size (>5 mm).

Despite limitations, our calculations provide a framework to deduce some of the additional factors that may influence the capacity of magmatic systems to contain lenses or caps of eruptible felsic melt. For example, large systems such as Yellowstone host well-developed hydrothermal systems, which increases the rate of heat extraction from magmatic systems with respect to what we consider here. This directly decreases the time available for melt segregation (Dutrow et al. 2001), thus increasing the potential for melt to form isolated melt lenses instead of large caps on top of magmatic reservoirs. Ultimately the complex feedbacks between magma chemistry and its physical properties and the thermal evolution of magmatic systems require a multidisciplinary approach (Gualda et al. 2018, Holness 2018, Huber & Parmigiani 2018) to identify the potential signs of an impending volcanic eruption. Our results suggest that for hydrous felsic magmatic systems the initial water content of magma play an important role on the extraction of residual by affecting both the evolution of the physical properties of magma with temperature and the time magma spends within different temperature intervals.

4.7 Acknowledgements

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Chapter 5

Relationships between the Takidani Pluton and caldera-forming eruptions

5.1 Abstract

The Takidani Pluton is thought to represent the exhumed magma reservoir that fed two voluminous volcanic eruptions including the Nyukawa Pyroclastic Flow Deposit and Chayano-Ebisutoge Pyroclastic Deposits. In this study, we use major and trace element chemistry of whole-rock, glass and minerals to (1) reconstruct the architecture of the magmatic plumbing system, and (2) to establish genetic links between the volcanic and plutonic rocks.

Whole-rock and glass compositions of the volcanic and plutonic rocks lie on parallel trends that range from low-silica dacite and granodiorite to high-silica rhyolites and granites. Trace element variations of the different units also follow similar trends indicating a genetic relationship between the volcanic and plutonic units. Trace element variations of the Takidani Pluton, Nyukawa PFD and Ebisutoge PD can generally be reproduced by crystal fractionation of plagioclase, orthopyroxene ($\pm$ clinopyroxene, amphibole), iron oxides, apatite and zircon from a source that corresponds to the Hotaka Andesite. The granitic units of the Takidani, on the other hand, require fractionation of amphibole and biotite and a mineral assemblage that corresponds to the granodiorites of the Takidani Pluton.

Plagioclase phenocrysts of the Takidani Pluton contain resorbed high-Anorthite (An_{90-70}) cores that are overgrown by an oscillatory zone mantle (An_{60-30}) and a sodic rim (An_{30-10}). Trace element chemistry and a sharp decrease in anorthite content in the plagioclase rims composition are consistent with crystallisation at the granitic minimum. Plagioclase crystals of the Nyukawa PFD crystals are oscillatory zoned (An_{60-38}) and show a general increase in FeO content towards the rim indicating replenishment of more mafic melt. The homogeneous trace element content of plagioclase rims reflect homogenisation prior to eruption. Ebisutoge PD contains

plagioclase crystals that commonly display resorption and overgrowth of sodic cores (An₃₀). The increase in An content and variations in trace elements are consistent with reservoir replenishment by hotter and more mafic magma. Crystals are normally zoned towards the rims (An_{30–28}). Plagioclase of the Takidani Pluton and Nyukawa PFD are indistinguishable in major, and trace elements suggesting crystallisation from a common source. Ebisutoge plagioclase, however, have Sr and Eu concentrations that are much higher and cannot be explained with crystallisation from a Takidani-type magma. However, Ebisutoge plagioclase contain xenocrystic cores that have major and trace element compositions comparable to the plagioclase in the Takidani Pluton and Nyukawa PFD suggesting interaction between these two magma types.

Our data suggests that the crystal-rich Nyukawa Pyroclastic Flow Deposit (>400 km³) is a reactivated portion of a deeper section of the Takidani magmatic reservoir. The 10 kyr younger Ebisutoge-Chayano crystal-poor rhyolite (>100 km³) is chemically distinct from Takidani magmas showing evidences for crustal melting. Our petrological investigations of the shallow magmatic system and erupted deposits suggest that the voluminous rhyolites are not extracted melts from the Takidani pluton. Instead the Chayano-Ebisutoge magmas were produced with a substantial contribution of crustal melting. The Nyukawa PFD, on the other hand, can be related to the Takidani Pluton.

5.2 Introduction

The relationship between volcanic and plutonic rocks has important implications for the evolution of magmas in the continental crust and generation of large silicic eruptions. Geochemical studies show that volcanic and plutonic rocks are essentially identical in composition, and add to a common view that plutons resemble frozen magma reservoirs that fed volcanic eruption (Lyell 1838). Crystal poor high-silica magmas (i.e. rhyolites) are commonly regarded as products of residual melt segregation from a crystallising magma reservoir (Bowen 1928, Brophy 1991, Hildreth 2004, Bachmann & Bergantz 2004). Plutons of intermediate composition (i.e. granodiorites) are considered in some cases as crystal residues or cumulates left behind after extraction and eruption of the interstitial rhyolitic melt (Bachmann & Bergantz 2004). Evidence for crystal-liquid segregation is drawn from volcanic sequences in systems where fractional crystallisation appears to control magma evolution, however, few cases of large melt extraction are found in the plutonic record (Bachl et al. 2001, Walker et al. 2015, Lee & Morton 2015, Schaen et al. 2017, Hartung et al. 2017). Moreover, a lack of coexisting plutonic and volcanic deposits makes it difficult to distinguish the extent of the relationship between plutonic and volcanic rocks.

In Japan, we have the unique opportunity to examine a young and active caldera-forming system where two types of volcanic deposits (i.e. high-silica ignimbrites and

monotonous magmas; Huber et al. 2012 and references therein) and a nearly 2 km thick section of its potential magma reservoir (i.e. Takidani Pluton; Harayama 1992) are exposed. The eruptive units include the Nyukawa Pyroclastic Flow Deposit (PFD) and the Chayano-Ebisutoge Pyroclastic Deposit (PD), some of the largest eruptions in Central Japan near the Plio-Pleistocene boundary ($1.75\text{--}1.76 \pm 0.17$ Ma; fission track; Harayama et al. 1991, Nagahashi et al. 2000). The volcanic units comprise pyroclastic material of about 500 km^3 DRE (Kimura & Nagahashi 2007, Oikawa 2003), and are preserved in stratigraphic order in the Takayama Basin (Figure 5.1a; Nagahashi 1995, Nagahashi et al. 2000). The deposits change from orthopyroxene-clinopyroxene-bearing crystal-rich dacite (Nyukawa PFD) to orthopyroxene-hornblende bearing crystal-poor rhyolites (Chayano-Ebisutoge PD) in 10 kyr (Kimura & Nagahashi 2007). The welded ignimbrites and co-ignimbrite ash fallout of the three deposits spatially overlap and distribute concentrically around what is supposed to be a common eruption centre, located about 40 km north-east Hotake-Dake (Figure 5.1b; Kataoka et al. 2001, Nagahashi et al. 2000), where the young Takidani Pluton is located.

Here we explore the relationships between the volcanic and plutonic units of the Takidani Pluton and Nyukawa PFD and Chayano-Ebisutoge PD using major and trace element chemistry of whole rock, glass and minerals. The fundamental questions that we address in this contribution include: (1) is Takidani Pluton the exhumed magma reservoir that fed the Nyukawa and Chayano-Ebisutoge eruptions? (2) How were the Nyukawa dacite the Chayano-Ebisutoge rhyolites generated? (3) Was the Chayano-Ebisutoge PD generated by melt extraction from the Takidani Pluton? Our study provides new insights into silicic magmatism and the generation of caldera-forming eruptions.

5.3 The Volcanic Units

5.3.1 The Nyukawa PFD

The Nyukawa Pyroclastic Flow Deposit (PFD) is a thick welded dacite flow unit exposed in the Takayama basin (Figure 5.2a) and Matsumoto basin, of up to 100 m thickness (Kimura & Nagahashi 2007, Nagahashi 1995, Nagahashi et al. 2000). The erupted volume has been estimated to exceed 400 km^3 DRE (Oikawa 2003). The chemical signature and spatial distribution of correlated tephra beds (Hotaka-Kd 39) point towards an eruption centre in the Hotaka area in the central Northern Japan Alps (Kataoka et al. 2001, Nagahashi et al. 2000). The age of the Nyukawa PFD is estimated at 1.76 ± 0.17 Ma using fission track dating (Harayama 1999, Nagahashi et al. 2000). The Nyukawa PFD is crystal-rich (about 30–40%) and contains plagioclase, orthopyroxene, clinopyroxene, iron oxides and trace amounts of quartz and hornblende. Zircon and apatite are common accessory phases.

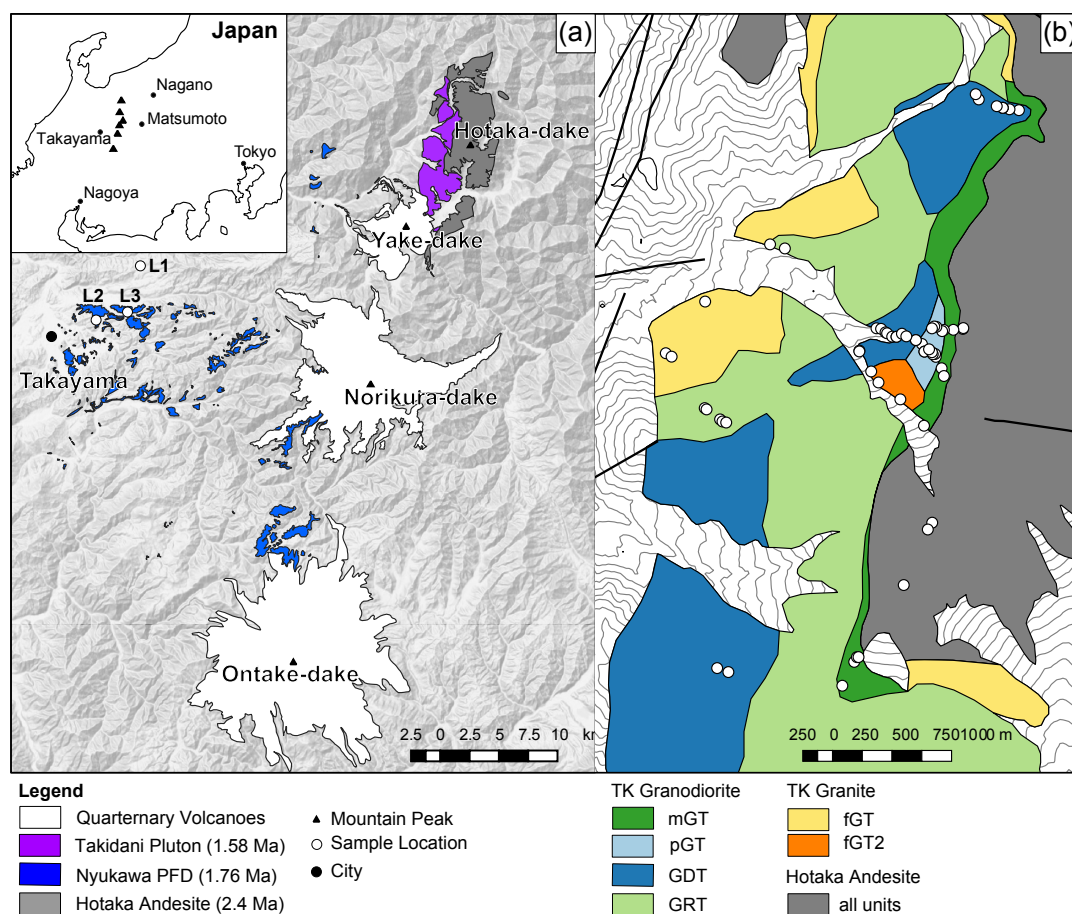


FIGURE 5.1: Location (a) and geological map (b) of the Takidani Pluton and associated volcanic deposits. Inlet shows overview location in Central Japan. Active volcanoes of the Norikura Volcanic Chain are highlighted in white. Sample locations are shown with white circles. Faults are indicated with black lines in (b). Grey lines indicate contours of 50 m.



FIGURE 5.2: Proximal deposits of the Nyukawa Pyroclastic Flow Deposit (a) and Chayano-Ebisutoge Pyroclastic Deposit in the Takayama Basin. (c) Leucocratic segregation with porphyritic plagioclase in granodioritic host rock observed on large boulders in the Shiradaashizawa valley. (d) Aplitic dike cross-cuts the granodiorite unit of the Takidani Pluton and embedded rounded enclave. (e) Mafic-felsic interaction in the Takidani Pluton and break-up of a mafic injection in the granodiorite, observed in a large boulder in the Shiradashi-zawa valley. (f) Intrusion of quartz veins and fracturing of the Hotaka Andesite welded-tuff at the roof of the Takidani Pluton.

5.3.2 The Chayano-Ebisutoge PD

The Chayano Tuff and Ebisutoge Pyroclastic Deposit (PD) are a sequence of rhyolitic voluminous ash fall and pyroclastic flow deposits (Figure 5.2b). The Chayano-Ebisutoge PD has an estimated eruptive volume of 380–490 km³ (Kimura & Nagahashi 2007, Nagahashi et al. 2000) with a DRE of >100 km³ (Kataoka et al. 2001). The outflow deposits are found in the Takayama Basin and Matsumoto area, and sit on top of the Nyukawa PDF (Kataoka et al. 2001, Nagahashi et al. 2000). The Chayano Tuff and Ebisutoge PD have been correlated with the Fukuda-Kd38 ash (Kataoka et al. 2001), a widespread marker bed distributed across Honshu Island. The stratigraphic location of the Fukuda tephra beds in sea sediments suggest that the rhyolites erupted only 10 kyr after the Nyukawa PFD at about 1.75 Ma (Harayama 1999, Nagahashi et al. 2000, Yoshikawa et al. 1996). The overlapping spatial distribution of pyroclastic deposits suggest that the Chayano–Ebisutoge PD was erupted from the same location of the Nyukawa PDF (Nagahashi et al. 2000).

The Chayano Tuff consists of pyroclastic fall (Chayano Tuff I) and non-welded flow deposits (Chayano I and II). The deposit is up to 4 m thick (Figure 5.2b) and associated with phreato-plinian activity (Kataoka et al. 2001). The Chayano Tuff contains plagioclase, iron oxides, hornblende, orthopyroxene and traces of clinopyroxene, biotite and zircon (Nagahashi et al. 2000). The Ebisutoge PD comprises four main units (Figure 5.2b; Nagahashi 1995, Nagahashi et al. 2000): Unit-A, a 20 cm thick pumice fall out containing small amounts of rock fragments; Unit-B, a up to 10 m thick unconsolidated flow unit containing pumice and rock fragments; Unit-C, a 10 to 20 cm thick pumice fall out containing small amounts of lithics; and Unit-D a >30 m thick welded ignimbrite containing fiamme (Nagahashi et al. 2000). The Ebisutoge PD is characterised by the abundance of plagioclase, iron oxides (magnetite and trace amounts of ilmenite), orthopyroxene and trace amounts of amphibole, biotite, clinopyroxene and quartz. Zircon is present as accessory phase. Detailed stratigraphic description and mineral fractions of the proximal and distal deposits are presented in Kataoka et al. (2001), Nagahashi (1995) and Nagahashi et al. (2000). Chemical and mineralogical data are presented in Kimura & Nagahashi (2007).

5.4 The Takidani Pluton

The Takidani pluton is one of the youngest exhumed intrusions on Earth (1.6 Ma $\pm$ 0.1, U-Pb ages of in-situ and detrital zircons; Ito et al. 2017, Harayama 1992). It is located between Yake-Dake and Hotaka-Dake in the Hida Mountain Range in the central Northern Japan Alps (Figure 5.1). The Takidani Pluton is emplaced at the base of the Hotaka Andesite, a co-genetic series of tuffs, lavas and porphyritic sills constituting a caldera fill sequence. High exhumation rates of the Central Alps and eastward tilting of the pluton lead to an estimated 2 km vertical exposure of this zoned intrusion (Harayama et al. 2003).

The Takidani pluton consists of at least six texturally and chemically distinct lithological units (Figure 5.1b), which contain plagioclase, quartz, alkali feldspar, hornblende and/or biotite and iron oxides in varying amounts (Bando et al. 2003, Hartung et al. 2017). Apatite (0.2–0.3%), titanite (0.1–0.3%) and zircon (0.02%) are common accessory minerals. A gradual transition in textures and chemistry from equigranular granodiorite (GDT) to more porphyritic granodiorites and granites (pGT) is observed in the upper part of the pluton. The simultaneous increase in the melt-matrix fraction of quartz and feldspars from 50 to 70 area% is interpreted to result from upwards melt segregation and accumulation (Figure 5.2c; Hartung et al. 2017, Chapter 4). The fine-grained nature of the segregated porphyritic granodiorite (pGT) near the roof of the reservoir may be indicative of rapid crystallisation of rhyolitic melt as result of eruption-induced decompression of the magma reservoir (Hartung et al. 2017). Mafic enclaves and aplitic dikes are present throughout the pluton (Figure 5.2d). Decimetric to metric scale enclaves are also observed in blocks at the base of the pluton (Figure 5.2e).

The Takidani Pluton is associated with two large caldera-forming events in the Late Pleistocene that produced the Nyukawa PFD and Chayano–Ebisutoge PD (1.76–1.75 Ma; Nagahashi et al. 1996, 2000). Chemical, temporal and field constraints indicate close relationships between the volcanic and plutonic deposits and suggest the Takidani Pluton as a potential magmatic reservoir that fed those volcanic eruptions (Figure 5.1; Harayama 1992, Kimura & Nagahashi 2007).

5.5 The Hotaka Andesite

The Hotaka Andesite (2.4 Ma) comprises 3000 m of deposit of andesitic to dacitic welded tuff sheets and subordinate lavas, collapse breccias, granophyric dikes and porphyritic sills (Harayama 1992, 1990). The different units were deposited inside a north-south oriented volcanic caldera. The Maehotakadake Welded Tuff (Wm) is a 1500 m welded tuff sheet that forms the main unit of the Hotaka Andesite (Figure 5.2f). It was intruded by porphyry dikes and contact metamorphosed to hornblende hornfels facies by the intrusion of the Takidani Pluton (Harayama 1990). The Hotaka Andesite contains high amounts of plagioclase, orthopyroxene, clinopyroxene and iron oxides (45–60 vol%, Harayama 1990).

5.6 Locations, materials and methods

Rock samples were collected from west-east trending river valleys called Takidani, Shirasashi-zawa, Nishiho-zawa and Yanagi-dani and from the Nishiho-track between the Nishihodakaguchi Station and mountain peak of Nishihotaka-dake (Figure 5.1b). The Nyukawa PFD and Ebisutoge PD were sampled at two different locations in the

Takayama Basin (Figure 5.1a). Unit B to D of the Ebisutoge PD were sampled from the outcrop shown in Figure 5.2b .

5.6.1 Whole-rock geochemistry

Whole rock major and trace element analyses were performed for a total of 77 samples from the Takidani Pluton (25 already published in Hartung et al. 2017), 4 samples from the Hotaka Andesite and for 9 samples of the volcanic units collected in the Takayama Basin (Figure 5.1; Tables D.1–D.6). Major elements were determined on glass beads by X-ray fluorescence (XRF) using a PANalytical AXIOS MAX with a Rhodium anode tube at 4 W at the University of Lausanne. SY-2, BHVO, JCH-1, NIM-N and NIM-G standards were used for quality control. Trace element concentrations were measured on the same glass beads by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) using a Quadrupole spectrometer Agilent 7700 interfaced to a GeoLas 200M 193 nm excimer ablation system at the University of Lausanne. The system was operating at 1420 W with a laser repetition rate of 10 Hz and a 120 μm spot size. Each sample was measured three times for 60 s. The background was measured for 90 s before the three ablations. Trace element concentrations were recalculated using CaO contents from XRF analyses and NIST SRM 612 as external standard. Data reduction was performed using the LAMTrace software package (Jackson 2008a,b, Jackson et al. 1992, Longerich et al. 1996).

5.6.2 Glass and mineral analyses

Major and minor element analyses of minerals and glasses were performed on 100 μm thick thin sections using a JEOL 8200 Superprobe electron probe micro-analyser (EPMA) at the University of Geneva. The microprobe is equipped with a five-channel wavelength-dispersive spectroscopy system (WDS) and was operated at an accelerating voltage of 15 keV and a beam current of 15 nA for plagioclase analysis, 20 keV and 15 keV for oxide analysis and 20 keV and 2 nA for glass analysis. A defocused beam of 2 μm diameter was used for plagioclase analysis. Oxides were analysed with a beam diameter of 1 μm . Glass was analysed in pumice, fiamme and matrix with a defocused beam of 10 to 20 μm . Analyses with a maximum error of 1% relative were retained and cross checked by recalculating the appropriate mineral formula. Glass and mineral analyses are provided in the Tables D.7–D.23. Analytical conditions are reported in Tables A.1–A.4.

Trace element concentration profiles were measured in 18 plagioclase crystals with LA-ICMPS using a sector-field spectrometer Element XR interfaced to a NewWave UP-193 ArF excimer ablation system at the University of Lausanne. The system was operating at 1400 W with a laser repetition rate of 20 Hz. The spot size ranged between 25 μm (i.e. high-An cores) and 100 μm for plagioclase analysis. Standards were analysed with a spot size of 75 μm . For each spot analysis, the background was first

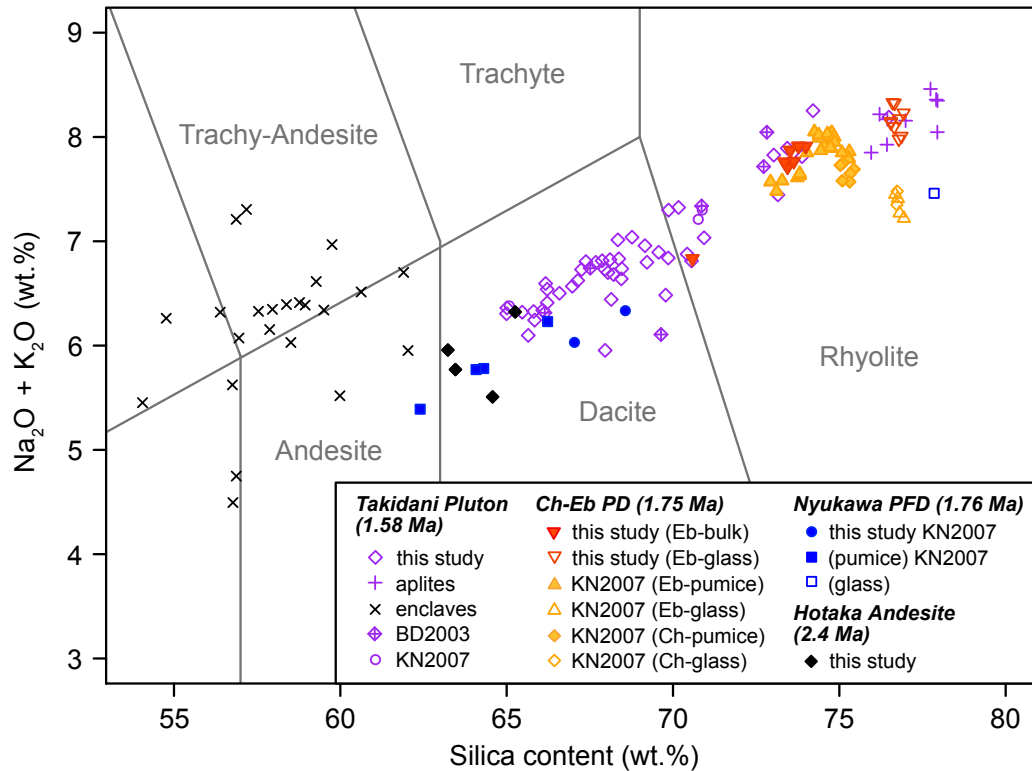


FIGURE 5.3: Total alkalis vs silica content of Takidani Pluton, Chayano–Ebisutoge Pyroclastic Deposit (PD), Nyukawa Pyroclastic Flow Deposit (PFD) and Hotaka Andesite in Central Japan. Diagram contains published data from Bando et al. (2003, BD2003) and Kimura & Nagahashi (2007, KN2007). Eb, Ebisutoge PD; Ch, Chayano Tuff.

measured for 30 s before ablation and peak measurement of 60 s each. Trace element concentrations were recalculated using CaO contents of plagioclase from EPMA analyses as internal standard and NIST SRM 612 as external standard. Data reduction was performed using the LAMTrace software package (Jackson 2008a,b, Jackson et al. 1992, Longerich et al. 1996). Trace element data are reported in Tables D.24–D.30.

5.7 Geochemistry

5.7.1 Whole-rock chemistry

Compositions of the different plutonic and volcanic units are presented in Table D.1 and Table D.7. Bulk rock compositions of the Takidani Pluton and coexisting volcanic deposits are sub-alkaline and range from dacitic to rhyolitic compositions (Figure 5.3). Mafic enclaves in the Takidani Pluton cover the range from alkaline to subalkaline with (basaltic) andesite and (basaltic) trachy-andesite compositions (Figure 5.3, Table D.5, D.6). The Hotaka Andesite and Nyukawa PFD form the relatively low silica end-members with silica contents of pumice and whole rock ranging between 63 to 65 wt.% SiO₂ and 62 to 68 wt.% SiO₂, respectively (Figure 5.5). The composition of

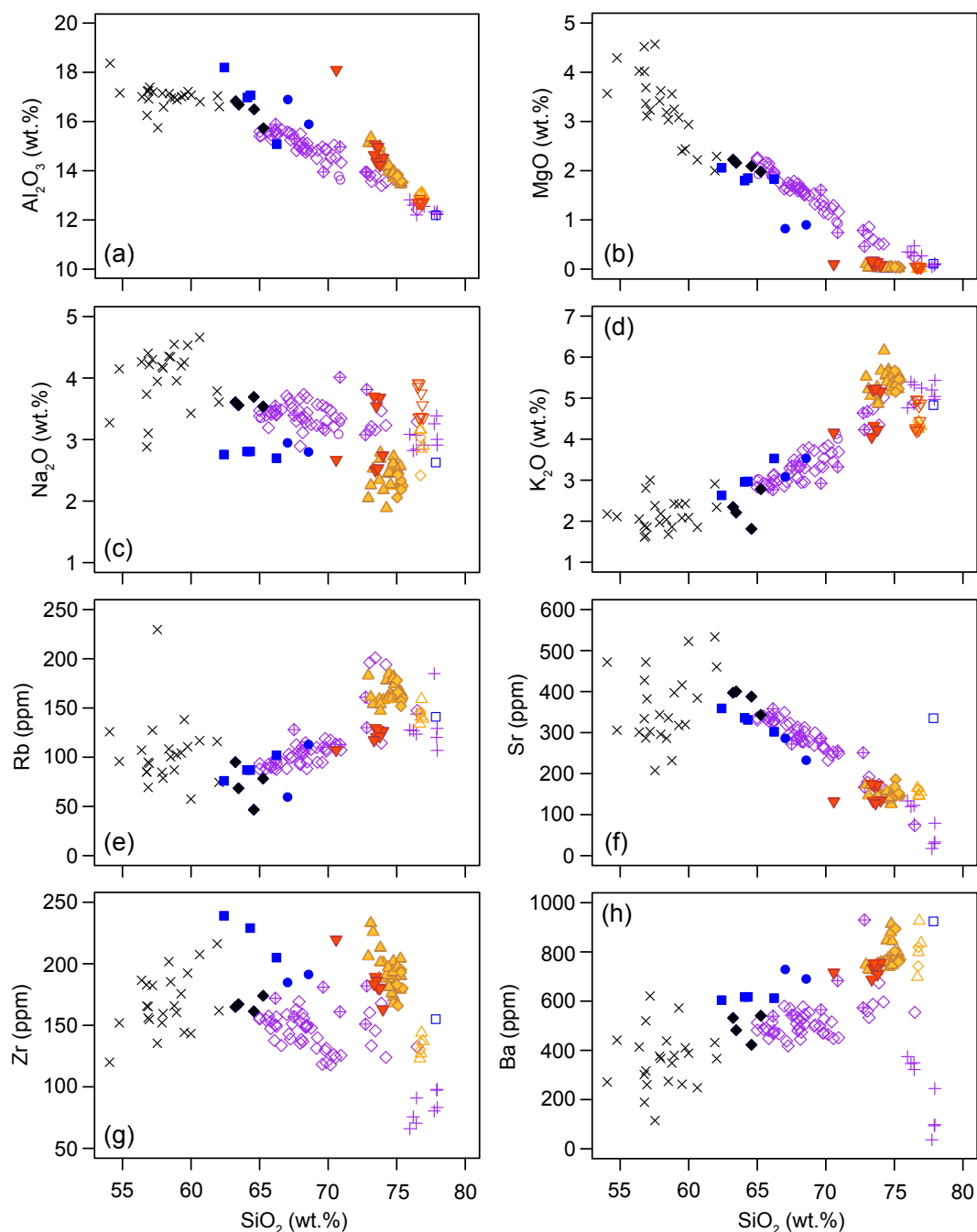


FIGURE 5.4: Major and trace element binary diagrams of the volcanic and plutonic deposits. All data are recalculated to anhydrous conditions. (a,b) Al_2O_3 and MgO content decrease with increasing silica content along liquid line of descent from enclave to high-silica aplites. (c) Variations in Na_2O content with silica content. Nyukawa PFD and Chayano-Ebsiutoge PD are depleted in Na_2O relative to compositions of the Takidani Pluton and Hotaka Andesite. (d) Incompatible behaviour and enrichment of K_2O with from andesite to rhyolite Liquid line of descent from Hotaka Andesite, Nyukawa PFD, Takidani Pluton to Ebsiutoge PD compositions. (e,f) Variations of Sr and Rb concentrations with increasing silica content. Andesitic enclaves are enriched in Rb content with respect to dacites of the Hotaka Andesite and Nyukawa PFD, and show large variations in Sr content. (g) Concentrations of Zr from andesite to rhyolite form M-shaped pattern with volcanic units being relatively enriched in Zr . (h) Increase in Ba content from low to high-silica compositions. Aplites are depleted in Ba content.

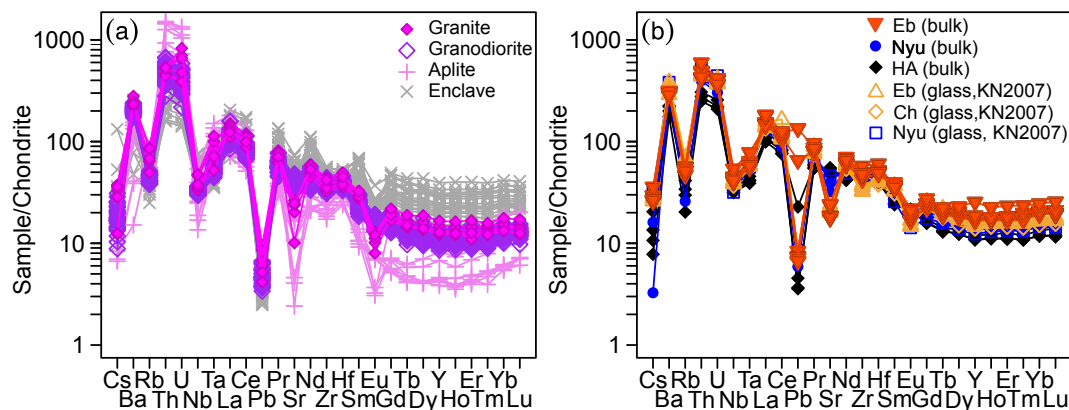


FIGURE 5.5: Spider diagrams of the Takidani Pluton (a) and volcanic units (b). Granodiorite and granites of the Takidani Pluton show similar trace element variations. Granites are slightly more enriched in heavy REE relative to granodiorite concentrations. Aplites are depleted in light and heavy REE, while enclaves in the Takidani Pluton are in enriched in REE. Trace element trends are comparable between different volcanic units and consistent with Takidani granodiorites and granites. Dacites and granodiorite trends are consistent and do not display a Sr or Eu anomaly. Rhyolite and granite trace element variations are also consistent showing negative Sr and Eu variations. All values are normalised to Chondrite concentration of McDonough & Sun (1995).

the Takidani Pluton varies from granodiorite to high-silica granite from about 64 to 78 wt.% SiO_2 . Whole rock analyses of the Ebisutoge PD (including Units B, C and D) show silica contents of about 73 to 74 wt.% and are consistent with published analyses of pumice (Figure 5.3; Kimura & Nagahashi 2007). Bulk rock analysis of Unit C (i.e. pumice fall out) displays a lower silica content of 71 wt.% SiO_2 . Takidani granites contain between 73 and 76 wt.% SiO_2 and overlap with glass and pumice of the Ebisutoge PD (Kimura & Nagahashi 2007). Aplitic dikes cross cutting the Takidani Pluton (Figure 5.2d) form a high-silica endmember of the intrusive portion of the system with composition similar to matrix glass of the Nyukawa PFD (Kimura & Nagahashi 2007). Glass analyses of the Chayano–Ebisutoge PD and Nyukawa PFD from Kimura & Nagahashi (2007) show lower Na_2O contents with respect to the Takidani pluton. Our new glass analyses for the Ebisutoge glass matrix (i.e. Units B to D), however, are consistent with compositions from the Takidani Pluton, which may suggest that previously reported matrix glass analyses were affected by Na_2O loss.

Spider diagrams overall show compositional similarities between plutonic (Figure 5.5a) and volcanic units (Figure 5.5b). The granites of the Takidani Pluton are consistent with granodiorites with exception of a negative Sr and Eu anomaly. Aplitic dikes are strongly depleted in REE. Enclaves are enriched in REE. Nyukawa PFD and Hotaka Andesite show similar compositional variability consistent with Takidani

granodiorites. Pumice and whole rock compositions of Chayano–Ebisutoge PD also have negative Sr and Eu anomalies, similar to Takidani granites. Major and trace element variations for selected elements show distinct trends for the volcanic and plutonic rocks (Figure 5.4). Concentration in major elements of Al_2O_3 and MgO generally decrease with increasing silica content from about 18 to 12 wt.% and 4.5 to 0.02 wt.%, respectively (Figure 5.4a,b). Pumice bulk-analyses of Chayano–Ebisutoge PD display lower MgO and higher Al_2O_3 with respect to Takidani compositions (Figure 5.4a,b). Parallel trends of Na_2O vs SiO_2 contents are observed for volcanic and plutonic rocks (Figure 5.4c). The composition of Ebisutoge PD splits into two groups: Unit B (i.e. unconsolidated pumice flow deposit) and Unit C (i.e. pumice fall outs) plot together with published pumice analysis from Kimura & Nagahashi (2007). Bulk analysis of Unit D (i.e. welded ignimbrite) display higher Na_2O and lower K_2O contents (Figure 5.4d) and are consistent with compositions of the Takidani Pluton. K_2O content is consistent between volcanic and plutonic units and generally increases from about 2.0 to 5.5 wt.% from andesite to rhyolite compositions. Concentrations of Rb are elevated in andesitic enclaves with respect to the dacitic composition of the Hotaka Andesite and Nyukawa PFD (Figure 5.4e). A sharp increase in Rb concentrations at about 73 wt.% SiO_2 is observed in the volcanic and plutonic rocks (i.e. granites) followed by a drop at about 75 wt.% SiO_2 (Figure 5.4e). A wide range of Sr content is observed in the mafic enclaves ranging from about 200 to 530 ppm (Figure 5.4f). A continuous decrease in Sr content from about 400 ppm is observed from the Hotaka Andesite to aplitic dikes in the Takidani Pluton. Matrix glass of the Nyukawa PFD contains significantly more Sr than the matrix of the Chayano–Ebisutoge PD (Figure 5.4f; Kimura & Nagahashi 2007). Zirconium concentrations are distributed over a M-shaped pattern showing different trends for volcanic and plutonic rocks (Figure 5.4g). Barium behaves incompatible in all units with concentrations reaching nearly 1000 ppm in the volcanic glasses. Aplitic dykes become progressively more depleted in Ba with increasing silica content consistent with fractionation of alkali feldspar and biotite (Figure 5.4h).

5.7.2 Matrix glass chemistry

Major and trace element analyses of the Nyukawa PFD and Chayano–Ebisutoge PD are reported in Kimura & Nagahashi (2007). Here, we present new major element data from glass analyses of the Ebisutoge PD. Glass composition of Ebisutoge PD are generally in agreement with published data from Kimura & Nagahashi (2007), with silica contents recorded between 76.6 and 76.9 wt.% SiO_2 (Tables D.7–D.16). Na_2O content varies between 3.4 and 3.9 wt.% with a mean value of 3.6 wt.%, which is about 0.7 wt.% higher than published data (Figure 5.4c). K_2O contents range between 4.2 and 5.0 wt.% with a mean value of 4.6 wt.% consistent with published data (Table D.2; Kimura & Nagahashi 2007). MgO contents of Ebisutoge glasses display a mean value of 0.03 wt.% and maximum concentrations of 0.15 wt.% (Table D.7). This is consistent with MgO concentrations of aplitic dikes but significantly lower than concentrations

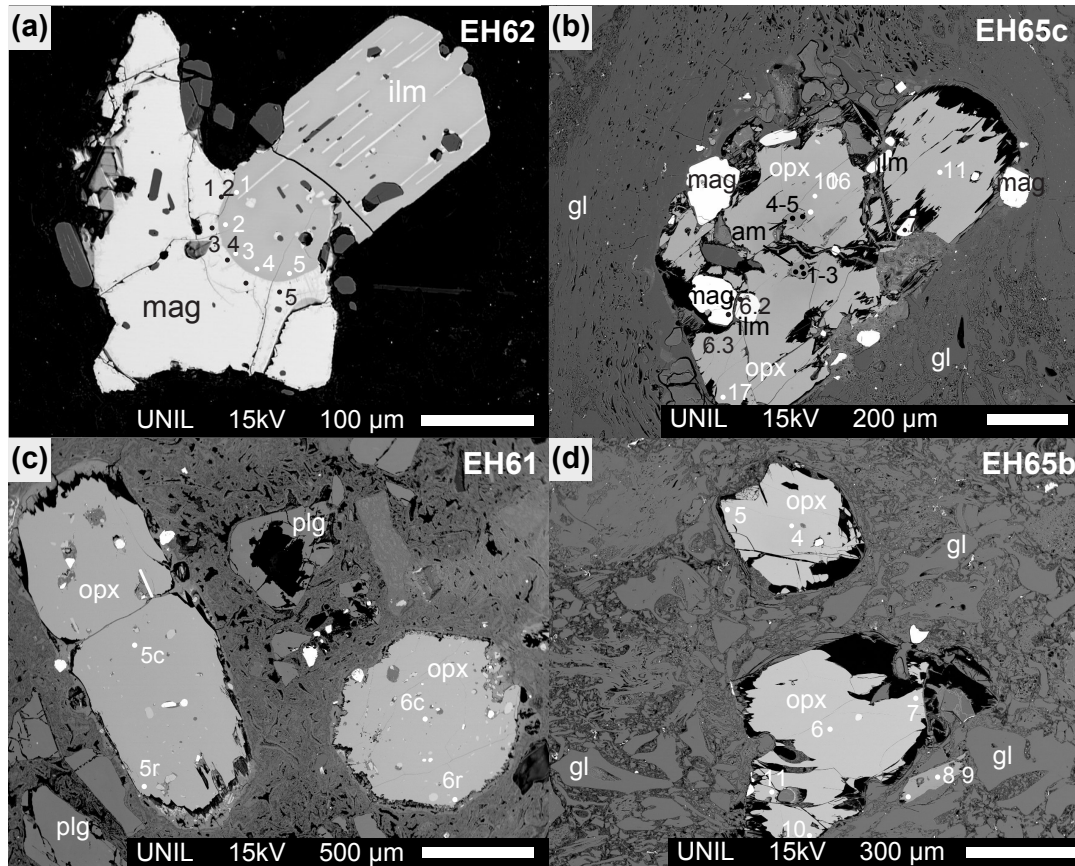


FIGURE 5.6: Back-scattered electron (BSE) images of magnetite-ilmenite pairs and orthopyroxene in the Nyukawa PFD and Ebisutoge PD. (a) Analysis of touching magnetite and ilmenite crystals in the Nyukawa PFD. (b) Glomerocryst of orthopyroxene, magnetite, ilmenite and zircon in the Ebisutoge PD. Orthopyroxene contains inclusions of ferric amphibole. (c) Large orthopyroxene phenocrysts in the crystal-rich Nyukawa PFD dacite. (d) Two types of orthopyroxene in the Ebisutoge PD. Large crystals are iron-rich orthopyroxene. Small crystals with numbers 8 and 9 present Mg-rich orthopyroxene. Numbers indicate number of compositional analysis. mag, magnetite; ilm, ilmenite; opx, orthopyroxene; am, amphibole; gl, glass.

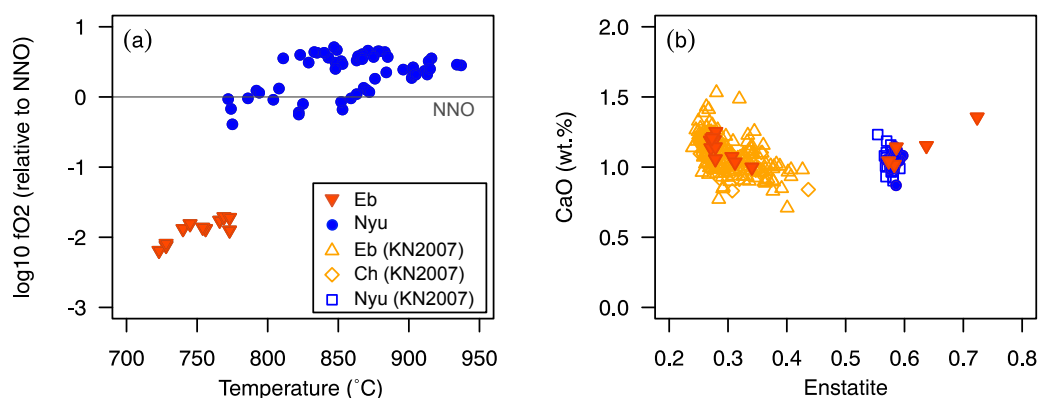


FIGURE 5.7: (a) Oxygen fugacity and temperature estimates of Ebisutoge PD and Nyukawa PFD from magnetite-ilmenite equilibrium using Ghiorso & Evans (2008). (b) Compositional variations of orthopyroxene in Ebisutoge PD and Nyukawa PFD from new and published analyses (Kimura & Nagahashi 2007).

in Takidani granites (Figure 5.4b). Trace element trends of Chayano and Ebisutoge glasses from Kimura & Nagahashi (2007) show negative Eu and Sr anomalies. Rare earth elements are enriched with respect to dacitic compositions of the Nyukawa PFD and Hotaka Andesite.

5.7.3 Fe-Ti oxides

Magnetite and ilmenite pairs were analysed from samples of the Nyukawa PFD and Ebisutoge PD (Figure 5.6a,b; Tables D.17–D.20). Oxide pairs are relatively abundant in the Nyukawa PFD but sparse in the Ebisutoge PD. Major element compositions are relatively homogeneous within Nyukawa PFD and within Ebisutoge PD, but show some chemical differences between the two volcanic units (Table D.17–D.20). Magnetite and ilmenite compositions in the Nyukawa PFD are generally characterised by lower TiO_2 contents and higher Fe_2O_3 to FeO ratios. Estimates from Fe-Ti exchange (Ghiorso & Evans 2008) yield average temperatures of $859^{\circ}C (\pm 41^{\circ}C)$ and $750^{\circ}C (\pm 19^{\circ}C)$ for Nyukawa PFD and Ebisutoge PD, respectively, consistent with temperature obtained in Kimura & Nagahashi (2007). Oxygen fugacity calculated from the oxides couples are slightly above the nickel-nickel-oxide buffer (NNO) for Nyukawa PFD (Figure 5.7a). Calculations for the Ebisutoge PD reveal lower fugacities of $\log_{10} -1.9$ relative to NNO (Figure 5.7a). Slow cooling of plutonic rocks and re-equilibration to lower temperatures prevents oxide pairs to be used as geothermometers in plutonic rocks (Carmichael & Nicholls 1967). Magnetite in the Takidani Pluton generally contains less TiO_2 and MnO with respect to Nyukawa and Ebisutoge deposits (Hartung et al. 2017).

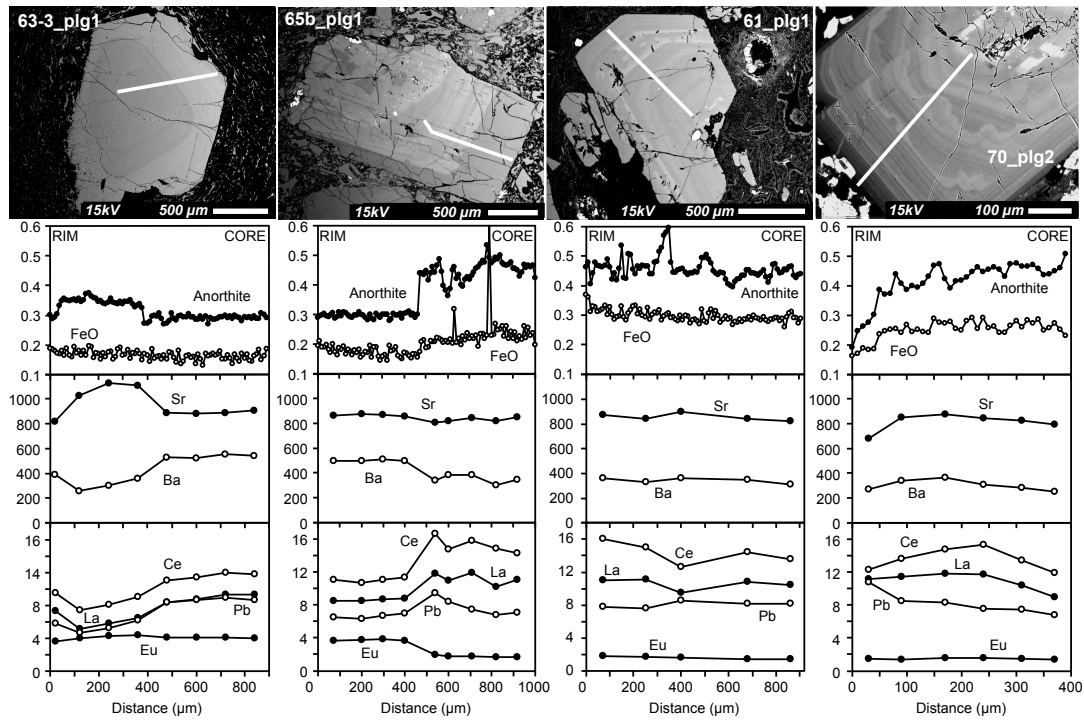


FIGURE 5.8: BSE images and chemical profiles of selected plagioclase crystals from the Ebisutoge PD (EH63.3 and EH65b), Nyukawa PFD (EH61) and Takidani Pluton (EH70). Anorthite and FeO content are from EPMA analysis. Trace element concentration are from LA-ICPMS.

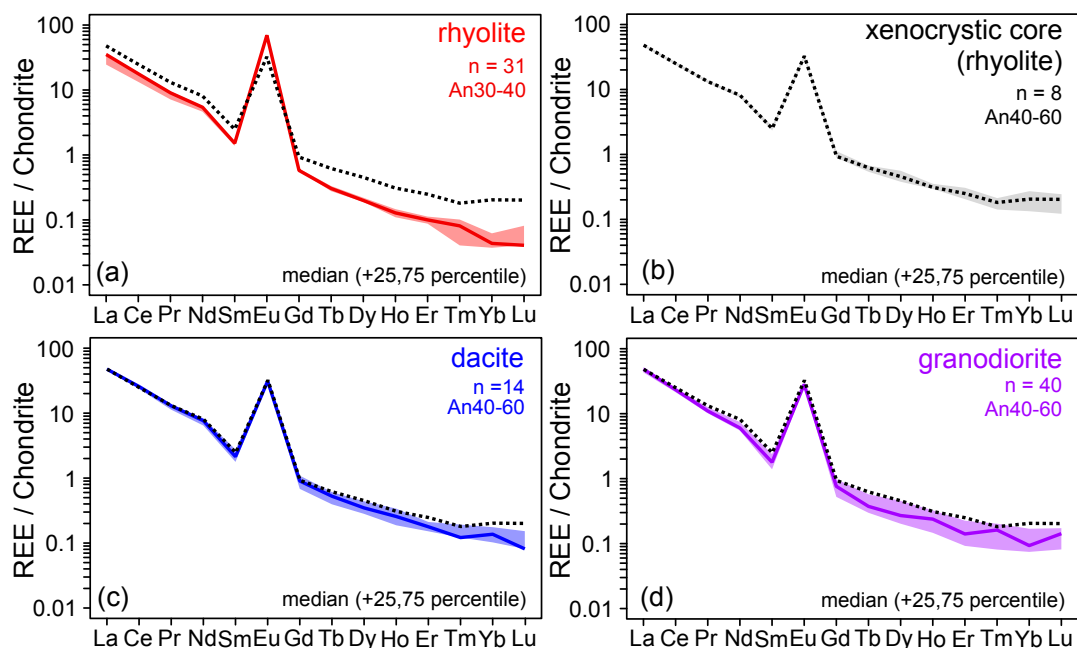


FIGURE 5.9: Rare Earth elements in plagioclase phenocrysts in the rhyolites of the Ebisutoge PD (a), xenocrystic cores in Ebisutoge rhyolite (b), the dacite of the Nyukawa PFD (c) and Takidani granodiorites (d). REE are normalised to Chondrite composition of McDonough & Sun (1995)

5.7.4 Orthopyroxene

Orthopyroxene is abundant in all volcanic units and usually occurs as subhedral crystals. We analysed several crystals in two samples of the Nyukawa PFD and two samples of the Ebisutoge PD (Unit D; Figure 5.6b-d; Tables D.21- D.22). Our results are consistent with compositions recorded in Kimura & Nagahashi (2007) and show that Nyukawa PFD typically contains Mg-rich pyroxene (En_{60-57}), whereas the Ebisutoge PD contains iron-rich orthopyroxene (En_{31-27}). Orthopyroxene in the Ebisutoge PD occasionally contains anhedral iron-rich amphibole inclusions (Figure 5.6b, Table D.23). A composition similar to this amphibole has been reported as clinopyroxene in Kimura & Nagahashi (2007) showing high Al_2O_3 content and low CaO content. Traces of Nyukawa-type orthopyroxene are found in both samples from the Ebisutoge PD (Figure 5.6d).

5.7.5 Plagioclase

Plagioclase is the dominant mineral phase in the Ebisutoge PD, Nyukawa PFD and Takidani Pluton. We analysed major and trace element concentrations along core-to-rim transects including nine plagioclase crystals from the Takidani Pluton, three crystals from the Nyukawa PFD and six crystals from the Ebisutoge PD. Back-scattered images and chemical profiles of representative crystals from the different units are

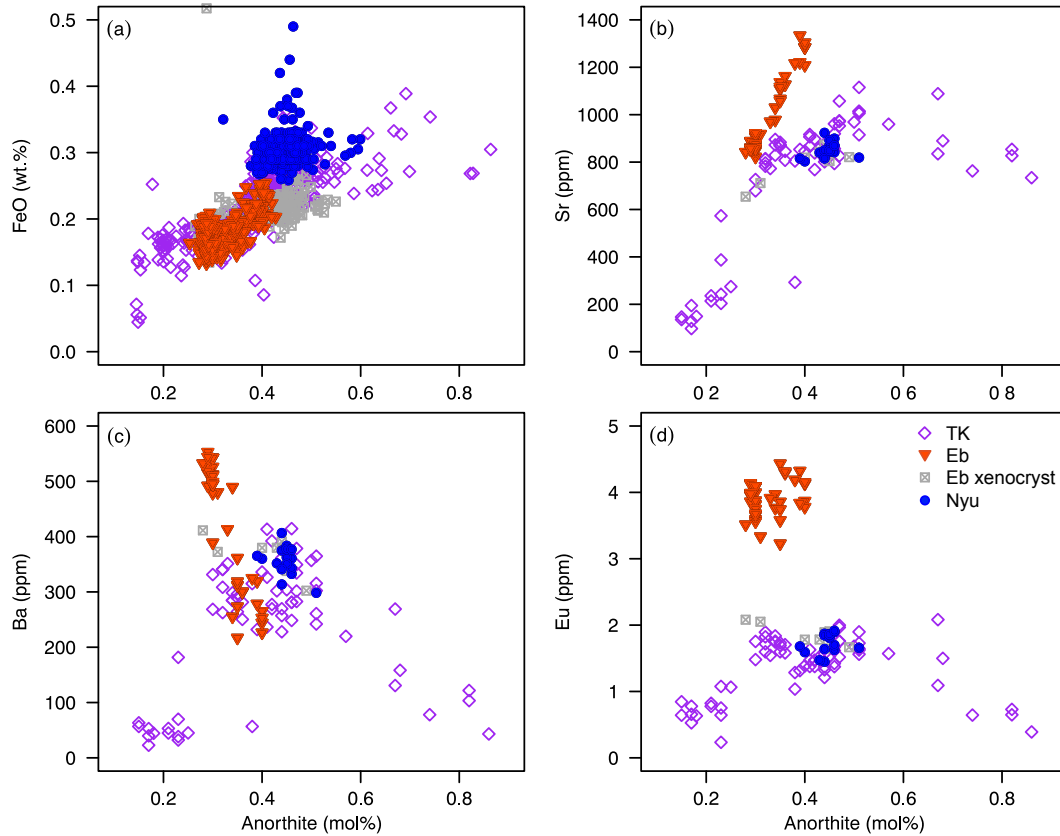


FIGURE 5.10: Compositional variability of 18 plagioclase phenocrysts from the Ebisutoge PD, Nyukawa PFD and Takidani Pluton. (a) FeO vs. anorthite content; (b) Two chemical trends are shown in Sr concentrations with high content in Ebisutoge plagioclase; (c) Rimward increase in Ba content in the Ebisutoge PD and decrease in the Takidani plagioclase; (d) Concentration in Eu in Ebisutoge plagioclase are twice as high as in Nyukawa and Takidani crystals.

shown in Figure 5.8. Rare Earth element trends are compared in Figure 9. BSE images of all analysed plagioclase are provided in Appendix D (Figures D.1–D.18).

Ebisutoge PD plagioclase are both euhedral and subhedral laths of up to 2 mm maximum length. In agreement with previously published values Anorthite content varies between 27 and 43 mol% (Kimura & Nagahashi 2007). Back-scattered electron (BSE) images and chemical analyses reveal that some of the crystals contain resorbed Albite-rich cores (An_{30}) that are overgrown by a more calcic growth band. Outer rims are typically sodium rich with Anorthite contents of 28–30 mol%. Finely spaced oscillatory zoning of 1 mol% Anorthite is common across all crystals. The compositional heterogeneity in major elements measured in core and overgrowth band is reflected in trace element variations (Figures 5.8–5.10). Concentrations of Sr increase with the increase in Anorthite from core to rim while Ba, La, Ce and Pb concentrations decrease. FeO content remains relatively unchanged (Figure 5.10a). Some plagioclase phenocrysts in the Ebisutoge PD contain xenocrystic cores that are texturally and

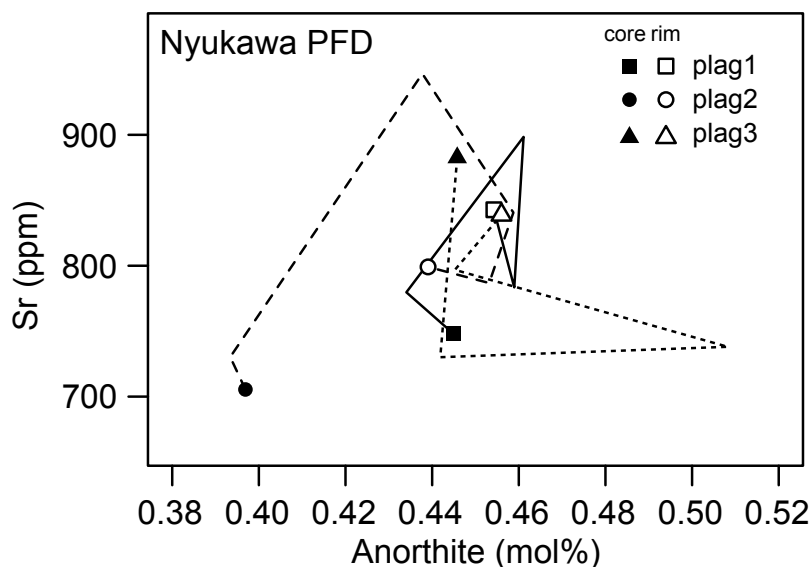


FIGURE 5.11: Sr-Anorthite trajectories for three plagioclase phenocrysts of Nyukawa PFD from core (black fill) to rim (white fill).

chemically distinct from the resorbed albitic cores (Figure 5.8). Xenocrystic cores are subhedral and have Anorthite contents of about 40 to 50 mol% that are distinctively higher than common Ebisutoge PD plagioclase composition (Figures 5.8–5.10).

The Nyukawa PFD contains subhedral plagioclase with Anorthite contents between 38 and 60 mol% (Figures 5.8–5.10). Crystals are characterised by oscillatory zoning between about 40 and 50 mol% Anorthite and occasional high-An zones of up to 60 mol% Anorthite. A rimward increase in anorthite content is observed, accompanied with an increase in FeO content from about 0.3 to 0.4 wt.%. All analysed plagioclase phenocrysts evolve to a common rim composition of about 40 to 45 mol% Anorthite (Figure 5.11). The Nyukawa plagioclase is texturally and chemically comparable to xenocrystic cores observed in Ebisutoge plagioclase.

Plagioclase phenocrysts in the Takidani Pluton range from An₉₀ to An₃₀ and are overgrown by a sodic rim (An_{30–15}), that varies in thickness in the different units. FeO content generally ranges between 0.4 and 0.2 wt.% and decreases to about 0.1 wt.% FeO in the outer rim (Figures 5.8, 5.10a). Rb concentrations are elevated in rim plagioclase (Figure 5.10b), whereas Sr, Ba, Pb and REE concentrations are relatively low (Figures 5.8–5.10). Hartung et al. (2017) have identified at least three distinct plagioclase populations in the Takidani Pluton. Population 1 comprises plagioclase mainly abundant in the mGT unit showing characteristic reverse zoning from 30 to 40 mol% Anorthite (Hartung et al. 2017). Crystals of population 2 and 3 are commonly constituted by resorbed cores of either cellular texture or high-An content (up to 90 mol%), that are overgrown by a distinct oscillatory zoned mantle with An-contents ranging generally between 40 and 55 mol%. Population 2 crystals are slightly less

anorthitic and have lower FeO content in average compared to population 3 crystals (Hartung et al. 2017). The major and trace element chemistry of Takidani plagioclase overlaps widely with plagioclase composition of the Nyukawa PFD and xenocrystic cores in the Ebisutoge PD (Figures 5.8–5.10).

5.8 Discussion

5.8.1 Origin of caldera-forming eruptions

The isotopic and geochemical signatures of the Chayano–Ebisutoge PD and Nyukawa PFD have been previously related to extensive crustal melting and mixing of end-members (Kimura & Nagahashi 2007) as part of mantle diapir-induced basaltic arc volcanism (Kimura & Yoshida 1999, Kimura et al. 2002). Considering the thermal and mechanical limitations on upper crustal melting (Dufek & Bergantz 2005) and mixing (Sparks & Marshall 1986) we re-evaluate this model using the existing data of Kimura & Nagahashi (2007) and our new petrological investigations.

Crystal-poor rhyolites

The depth of origin of the crystal-poor rhyolites is constrained between 5 and 10 km using phase relationships of the Chayano–Ebisutoge PD and experimental results from Rutherford & Devine (1996). Our pressure estimates from rhyolite-MELTS barometry using new Ebisutoge glass analyses (Figure 5.12a) are consistent with estimates of Kimura & Nagahashi (2007) and indicate a melt origin that corresponds to the emplacement level of the Takidani Pluton (Bando et al. 2003, Hartung et al. 2017). A decrease in magmatic temperatures from 860°C to 700°C from the Chayano Tuff to Unit D of the Ebisutoge Deposit, respectively, is recorded from magnetite-ilmenite thermometry in Kimura & Nagahashi (2007). Two-oxide analyses also reveal low magmatic temperatures of the Ebisutoge PD of 750°C using the calibration of Ghiorso & Evans (2008), and are comparable to those recorded by Kimura & Nagahashi (2007). The oxygen fugacity of Chayano–Ebisutoge rhyolites is estimated at 2 log below the NNO buffer (Figure 5.7a; Ghiorso & Evans 2008). Given such a low oxygen fugacity (i.e. below FMQ buffer) the magmatic assemblage of magnetite and ilmenite in the rhyolites is unusual. The partitioning of Eu between plagioclase and melt strongly depends on the oxygen fugacity (Wilke & Behrens 1999) and is reflected by the high Eu content recorded in Ebisutoge plagioclase, which is most likely the result of the reduced magmatic conditions of the rhyolites.

Melting of pelitic lithologies can lower the oxygen fugacity of magmas. Kimura & Nagahashi (2007) suggest that extensive melting of the pelitic gneisses and granites of the Hida belt produced the Chayano–Ebisutoge rhyolites. Evidences for melting is found in the mineral phases of the Ebisutoge PD. Anhedral amphibole inclusions exist in Ebisutoge orthopyroxene. These amphiboles have very high Fe-contents

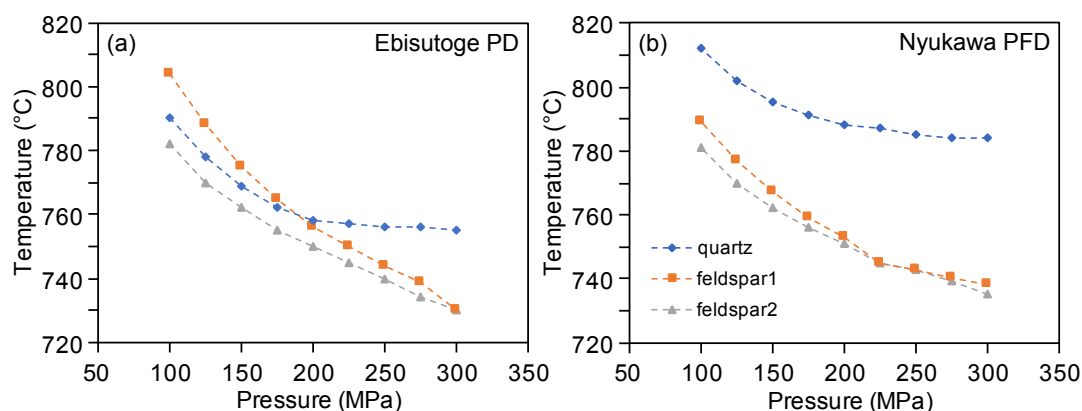


FIGURE 5.12: Phase equilibria of rhyolite glasses of the Ebisutoge PD (a) and Nyukawa PFD (b) showing stability fields of quartz and two feldspars using rhyolite-MELTS simulations (Gualda & Ghiorso 2014). Calculations for Nyukawa PFD are based on published data from Kimura & Nagahashi (2007).

(22.9–23.8 wt.%; Table D.23) and low Mg-numbers (0.40–0.36; Table D.23). Melting of Fe-rich amphibole, thus, is likely responsible for the Fe-rich contents of Ebisutoge orthopyroxene and the observed iron enrichment relative to magnesium in Ebisutoge glasses and pumices. Plagioclase in the Ebisutoge PD commonly contains resorbed Albite-rich cores that are overgrown by more Anorthite-rich plagioclase. The concentrations of Sr and Ba are respectively higher and lower in the overgrowth (Figure 5.8) indicating an injection of more mafic and less fractionated magma. This suggests that partial melting of pelitic rocks (e.g. amphibolite) produced or contributed to the generation of the Fe-rich Chayano–Ebisutoge rhyolites. The contemporaneous Ueno Basalt, which erupted in the southern part of the Norikura Volcanic Chain, thus, could have produced the Chayano–Ebisutoge PD. Kimura & Nagahashi (2007) estimated that a minimum volume of 300 km³ of Ueno basalt remained in the mid crust, which could have been sufficient to produce the estimated 100 km³ of rhyolite.

Overall, the rhyolitic magmas of the Chayano–Ebisutoge PD show little variation in major elements reflecting near eutectic conditions. However, a stratigraphic increase in K₂O and Rb contents from 4.20 to 4.96 wt.% and from 134 to 159 ppm (Kimura & Nagahashi 2007), respectively, indicates that magmatic differentiation must have acted on top of partial melting. A chemically zoned magma reservoir may have formed prior to eruption. Moreover, whole-rock analyses of Unit C display less evolved composition and together with andesitic and dacitic glass analyses of distal ash beds (Kataoka et al. 2001) indicate mafic magma replenishment prior the voluminous eruption of Unit D. This highlights that the Chayano–Ebisutoge magmas are more heterogeneous as previously stated and that rhyolitic melts of the Chayano–Ebisutoge PD are not pure crustal melts.

Crystal-rich dacite

The Nyukawa PFD (i.e. pumices and bulk analysis) and Hotaka Andesite are chemically and mineralogically similar. The Nyukawa PFD can be reproduced by 20–40% fractionation of a mineral assemblage of 60–70% plagioclase, 20–25% orthopyroxene ($\pm$ clinopyroxene) and 10–15% magnetite using partition coefficients for dacitic compositions. The composition of the rhyolitic melt of the Nyukawa PFD, however, cannot be produced from Nyukawa bulk and/or pumice compositions with the same mineral assemblage. To lower Y concentration from bulk concentrations of 22.3 ppm to 19.4 ppm in the melt, about 15% of amphibole fractionation is required, which is inconsistent with REE variations. Moreover, the observed increase in Sr content from 259 ppm (or 332 ppm in pumices) to 335 ppm cannot be explained with plagioclase fractionation. Melting of Sr bearing mineral phases (i.e. feldspars, biotite) could explain the observed increase in Sr in the melt. This suggests that the melt of the Nyukawa PFD has a different origin than the crystal cargo. Additional glass analysis would be required to test this hypothesis, especially in light of few existing analysis of glass composition of the Nyukawa PFD.

Plagioclase chemistry is controlled by the composition of the magma, temperature and water pressure (Housh & Luhr 1991, Hammer & Rutherford 2002). Plagioclase phenocrysts of the Nyukawa PFD show a rimward increase in Anorthite and FeO content indicating the arrival of a new and hotter magma. Furthermore, trace element compositions of plagioclase rims are homogeneous ((Figure 5.10) and support the process of heating and homogenisation of crystal mushes (Bachmann & Bergantz 2006, Huber et al. 2010).

5.8.2 Magmatic evolution of the Takidani Pluton

The samples of the Takidani Pluton show compositions that overlap with the low-silica dacite to high-silica rhyolite of the volcanic units (Figure 5.4). Similarities in major and trace element variations of the Takidani Pluton, Hotaka Andesite and Nyukawa PFD also point towards a common magma source in the mid to lower crust (e.g. Ueno Basalt, Kimura & Nagahashi 2007). The average composition of the Takidani granodiorite can be produced by crystal fractionation of plagioclase (50–60%), orthopyroxene (20%), clinopyroxene (15–20%), magnetite (5–10%) and apatite (1%), and corresponds to the mineral assemblage of the Hotaka Andesite. The granites of the Takidani Pluton can be produced by 50–60% fractionation of plagioclase (70–85%), amphibole (8%), biotite (4%), magnetite (3%) and trace amounts of apatite and zircon using an average partition coefficients for dacitic and rhyolitic compositions (i.e. Geochemical Earth Reference Model). The required mineral assemblage is consistent with Takidani granodiorites and suggests that Takidani granites are derived from Takidani granodiorites. Based on field evidence and isotopic signature (Harayama, personal communications) the fine-grained biotite granite at the base and roof of

the pluton present the oldest unit of the Takidani Pluton. However, if the granites are derived from the granodiorites, this means that the granodiorites were either remobilised from a deeper part of the pluton or that granodiorites were produced in-situ by the injection of more mafic melts.

Plagioclase of the Takidani Pluton have chemical characteristic that correlate with the Nyukawa PFD. Most crystal in the Takidani Pluton have textural and compositional features that are consistent with Nyukawa plagioclase and could indicate that both crystallised from compositionally similar or potentially identical magmas. The formation of sodium rich rims in the Takidani Pluton occurs when magmas become rheologically locked and dive into the granitic minimum. The different trajectories of rim compositions of the Takidani Pluton and Nyukawa PFD, thus, reflect the different fates of crystal-rich magmas. The textural and compositional similarities between the Takidani Pluton, the Nyukawa PFD and xenocrystic cores in Ebisutoge plagioclase show that a genetic link exists between the volcanic and plutonic deposits. Isotope analyse of the plagioclase phenocryst would be useful to confirm our hypothesis and potentially link individual growth zones between different deposits.

5.8.3 Volcanic-plutonic relationships

The magmatic dynamics of the Takidani-volcanic system are revealed in plagioclase growth history and major and trace element chemistry. The Takidani Pluton, Nyukawa PFD and Ebisutoge PD contain plagioclase crystals and xenocrysts that share the same textural and compositional features and provide evidence for a connected magmatic system.

The chemical signature of Ebisutoge plagioclase points towards a crustal origin as previously proposed by Kimura & Nagahashi (2007). However, it is unlikely that the Ebisutoge rhyolites present pure crustal melts, as they show crystal fractionation trends from the lower to upper units (Kimura & Nagahashi 2007) and evidences of magma replenishment. The emplacement of the Takidani Pluton likely occurred between about 100 MPa (i.e. equivalent to the total thickness of the Hotaka Andesite) and 200 MPa (i.e. pressure estimates from plagioclase-hornblende barometers, Hartung et al. 2017). Pressure estimates of the Nyukawa PFD are based on comparison of mineral assemblages between natural rocks and experiments (Kimura & Nagahashi 2007) and indicate reservoir depth of about 100 MPa. Pressure estimates could be higher if magmas were undersaturated. Rhyolite-MELTS barometer indicates that Ebisutoge melts equilibrated at pressures of about 200 MPa (Figure 5.12). The existence of plagioclase xenocrysts and Nyukawa-type orthopyroxene in the Ebisutoge PD support melt storage and equilibration at shallow levels prior to eruption.

The formation of the magmatic plumbing system and succession of events is shown in Figure 5.13. The Hotaka Andesite erupted at 2.4 Ma and created large

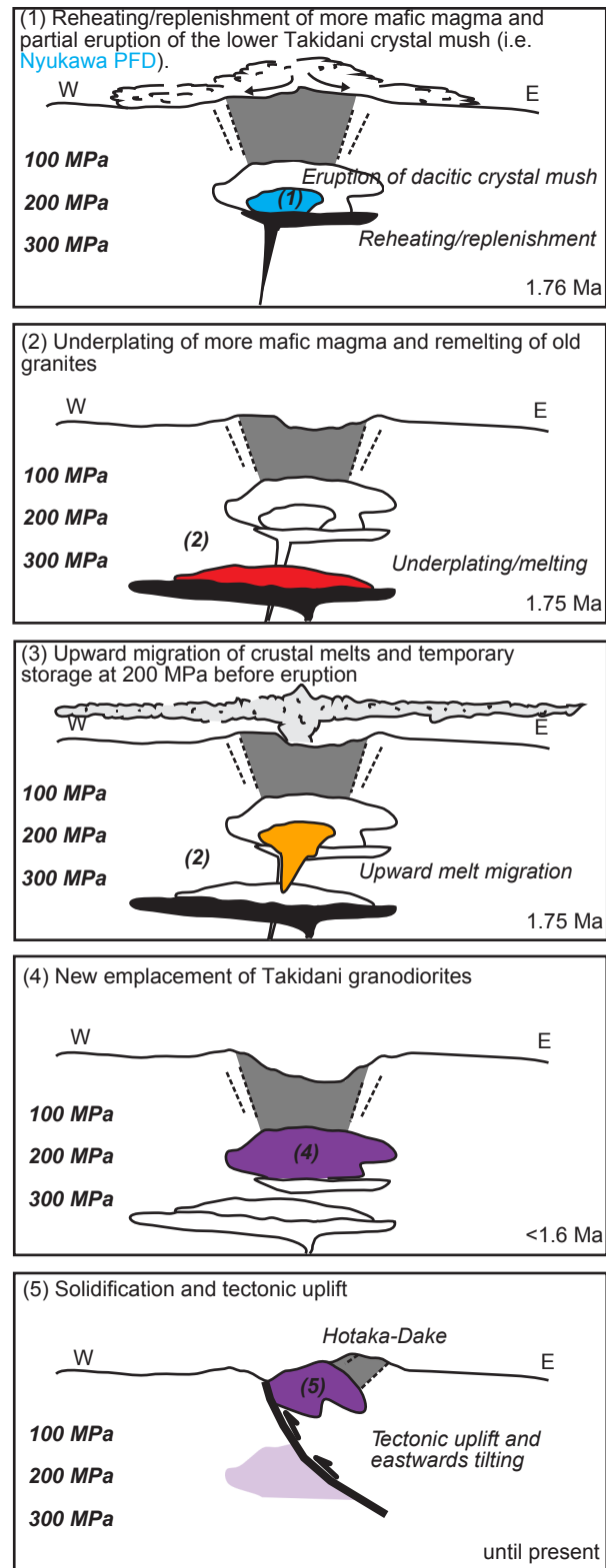


FIGURE 5.13: Model of the Takidani magmatic system. (1) Reactivation of crystal mush leads to the eruption of the Nyukawa PFD. (2) Melting of mid-crustal lithologies leads to rhyolite generation of the Ebisutoge and Chayano eruptions. (3) Rhyolite stagnated and equilibrates at shallow levels prior to eruption. Petrological cannibalism of plagioclase and pyroxene. (4) Emplacement of the Takidani Pluton and (5) tectonic uplift.

caldera infill deposits that were intruded by the Takidani granites. The Takidani granites could have been extracted or remobilised from a deeper part of the Takidani system. Injection of hot magma led to partial melting and thermal reactivation of an existing mush zone, and eruption of the Nyukawa PFD. Extensive melting of crustal lithologies (e.g. amphibolite) produced the Chayano–Ebisutoge rhyolites, which incorporated crystals from the overlying Takidani–Nyukawa magmatic system. A last influx of magma and melt lead to the final emplacement of Takidani granodiorites below the Hotaka Andesite to form a plutonic-volcanic complex.

5.9 Conclusions

Major and trace element chemistry of whole-rock and minerals provide new insights into the magmatic plumbing systems of large-caldera forming eruptions in Central Japan and genetic relationships of volcanic and plutonic systems. Our geochemical investigations suggest that the Chayano–Ebisutoge PD, Nyukawa PFD and Takidani Pluton originated from a common source at mid-crustal levels. While the Nyukawa PFD and Takidani Pluton are likely the result of crystal fractionation, the iron-rich rhyolites of the Chayano–Ebisutoge PD are dominantly derived from partial melting of crustal lithologies. This implies that the rhyolites cannot present extracted melts from the Takidani Pluton. The observed melt extraction in the Takidani Pluton may have been associated to smaller silicic eruptions that have not been identified yet.

Chapter 6

General conclusions and future directions

6.1 Summary

Petrological investigations of the Takidani Pluton and the related Nyukawa and Chayano–Ebisutoge pyroclastic deposits provide new insights into the magmatic evolution of the upper crust and relationships between plutonic and volcanic rocks. The Takidani Pluton is one of the youngest exposed plutons on Earth (Harayama 1992). It was constructed by multiple intrusions of granodioritic and granitic magmas with different mineralogical and geochemical characteristics. Evidence for residual melt extraction from a partially crystallised magma is observed in the upper section of the pluton (Chapter 3). Segregation timescales in the Takidani Pluton are estimated to range from centuries to millennia using thermodynamic and thermal modelling (Chapter 4).

Our new mineral and glass analysis support an origin of the rhyolite from crustal melting (Chapter 5), which has been previously suggested by Kimura & Nagahashi (2007). The extraction of residual melts in the Takidani Pluton did not produce the rhyolites of the Chayano–Ebisutoge PD as initially hypothesised (Chapter 3). Whether the Takidani Pluton was feeding silicic volcanic eruptions remains unclear. Mineral and whole rock chemistry, however, suggest that the Takidani Pluton and the Nyukawa PFD crystallised from a common source that corresponds to the Hotaka Andesite (Chapter 5). This implies that the Nyukawa PFD most likely represents a remobilised part of a larger magmatic reservoir (i.e. crystal mush) consistent with existing evidences and models of mush remobilisation (Bachmann & Dungan 2002, Degruyter & Huber 2014, Huber et al. 2011, 2009, Sliwinski et al. 2017, Parmigiani et al. 2014, Couch et al. 2001, Murphy et al. 2000, Burgisser & Bergantz 2011). Both, the Nyukawa PFD and Chayano–Ebisutoge PD record mafic replenishment prior to eruption, which may have been the trigger for the large volcanic eruptions (Pallister et al. 1992). Mafic enclaves exist in the Takidani Pluton providing additional evidence for the interaction of mafic and felsic magmas at depth. The extent of mafic injections and interaction in the Takidani Pluton was not constrained during the course of this

PhD study.

6.2 Ongoing and future research

This PhD study and new comprehensive dataset of textural, geochemical and isotopic analysis of whole-rock, glass and minerals of the Takidani Pluton, Nyukawa PFD, Ebisutoge PD and Hotaka Andesite provide a large foundation for future investigations of upper crustal magmatism in relation to silicic volcanism.

A detailed study on the textures, geochemistry and geochronology of the porphyritic unit is needed to better understand the evolution of textures leading to melt segregation and potentially volcanic unrest. If the fine-grained texture of the porphyritic unit was indeed caused by decompression crystallisation as result of a volcanic eruption, we should expect to find evidence of dendritic growth throughout the porphyritic unit and in the granodiorite sitting below.

To better constrain the thermal and temporal evolution of the Takidani magmatic system and maximum timescales of residual melt segregation in the Takidani pluton a detailed geochemical and geochronological study is currently being carried out in collaboration with Dr Federico Farina at the University of Geneva. The data sets are omitted from the thesis text, as work is still ongoing. As part of this project, I have separated zircons from two different units of the Takidani Pluton, the Nyukawa PFD and Ebisutoge PD and imaged over 300 zircon grains using cathodoluminescence (CL). Major and trace element analysis of 82 selected zircon grains were performed using the electron microprobe at the University of Geneva. In addition, oxygen isotopes of 80 selected zircon grains were measured with technical assistance from Cindy Luisier at the Swiss SIMS facility in Lausanne. U-Pb dating using ID-TIMS was performed by Dr Federico Farina at the University of Geneva. The first results are presented in Figure 6.1 showing 19 zircon ages of the granodiorite and porphyritic unit. The zircon ages of the two units overlap. Available data give an averaged age of 1.3 Ma, which is somewhat younger as the published data by Ito et al. (2017). A manuscript for publication is currently being prepared.

The Ebisutoge rhyolite contains plagioclase with euhedral xenocrystic cores that are texturally and chemically similar to Takidani and Nyukawa plagioclase phenocrysts. Diffusion chronometry on the plagioclase xenocrysts and overgrowth could provide additional information on timescales of migration of rhyolitic melt through shallow magmatic reservoirs. This study would require high-resolution analysis of plagioclase phenocrysts (i.e. SIMS, LA-ICPMS) coupled with diffusion modelling. The results of this study could be compared to crystallisation ages of quartz in the volcanic rocks (Gualda et al. 2018 and references therein).

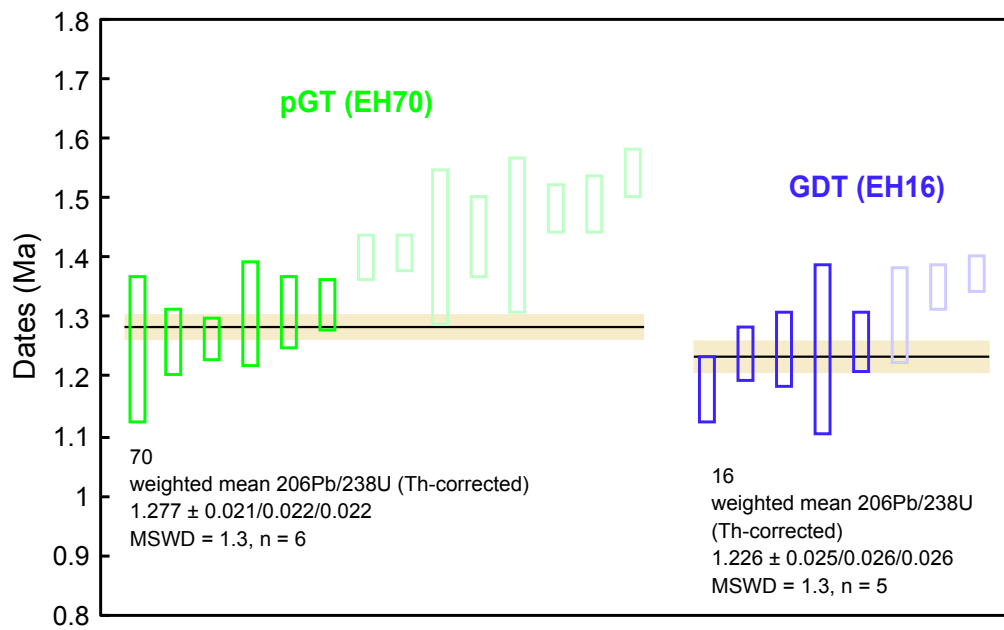


FIGURE 6.1: U-Pb ages of zircon grains from the granodiorite and porphyritic granodiorite in the Takidani Pluton. Ages were measured by Dr. Federico Farina using ID-TIMS at the University of Geneva. Ages are Th-corrected.

Detailed investigations on the chemical variability of glasses and minerals of the Chayano–Ebisutoge PD are required to further constrain the sequence of rhyolite eruptions and related magmatic processes. Additional glass analysis of Nyukawa PFD would be useful to better constrain the amount of partial melting of crystal mushes.

Additional whole-rock radiogenic isotopes of the Takidani granites (fGT and fGT2) and the Hotaka Andesite will be useful to better constrain the emplacement sequence of the intruded magmas and to test whether any genetic relationships exist between the granites and the erupted rhyolites. Isotope analysis of minerals (i.e. plagioclase) can also help to further constrain potential genetic relationships of the plutonic and volcanic rocks. More specifically, isotope analysis of Takidani and Nyukawa plagioclase and xenocrysts in Ebisutoge plagioclase should resolve any remaining doubts on their genetic relationship. Finally, new isotopic analysis of whole-rocks and minerals would be useful to better constrain the extent of melting and fractionation of the high-silica magmas.

To better understand the emplacement of felsic magma in the shallow crust and to identify the structures associated with magma emplacement I analysed the magnetic and magmatic fabrics of the Takidani Pluton and Hotaka Andesite (Appendix E). Measurements of anisotropy of magnetic susceptibility (AMS) were performed for 50

oriented samples. Magmatic fabrics were determined through direct measurements of mineral lineations in the field. The magnetic and magmatic fabrics of the Takdani will be used to determine the orientations of the principle axes of strain related to magma flow and emplacement (Archanjo et al. 2012, Bouchez 1997, Hutton 1988, Launeau & Cruden 1998, Paterson et al. 1998, Payacán et al. 2014, de Saint-Blanquat et al. 2001, Schöpa et al. 2015, Trubač et al. 2009, Vernon 2000). All datasets and a short summary of results are provided in Appendix E. A manuscript for publication is currently being prepared.

6.3 Concluding remarks

The Northern Japan Alps are an exceptional location to investigate the chemical, physical and temporal evolution of magmatic systems in the Earth's crust. The Takidani-Hotaka volcanic-plutonic complex provides a vertical cross-section of a young and potentially still active volcanic caldera including contact metamorphosed intracaldera fills (i.e. Hotaka Andesite) and a 2 km thick subvolcanic magmatic reservoir (i.e. Takidani Pluton). The complex is located within an active volcanic chain and also linked to two large caldera-eruptions of crystal-rich dacite and crystal-poor rhyolites. Additional work is required to better constrain melt segregation in shallow magmatic reservoirs and the evolution of caldera-forming systems. I am eager to see the outcomes of future research in this area and hope to be involved in many projects over the next years and decades.

Bibliography

- Allmendinger, R. W., Cardozo, N. & Fisher, D. M. (2011), *Structural geology algorithms: Vectors and tensors*, Cambridge University Press.
- Andersen, N. L., Singer, B. S., Jicha, B. R., Beard, B. L., Johnson, C. M. & Licciardi, J. M. (2017), 'Pleistocene to holocene growth of a large upper crustal rhyolitic magma reservoir beneath the active laguna del Maule Volcanic Field, Central Chile', *Journal of Petrology* **58**(1), 85–114.
- Anderson, J. L. (1996), 'Status of thermobarometry in granitic batholiths', *Geological Society of America Special Papers* **315**, 125–138.
- Anderson, J. L., Barth, a. P., Wooden, J. L. & Mazdab, F. (2008), 'Thermometers and Thermobarometers in Granitic Systems', *Reviews in Mineralogy and Geochemistry* **69**(Chapter 4), 121–142.
- Anderson, J. & Smith, D. (1995), 'The effects of temperature and f_{O_2} on the Al-in-hornblende barometer', *American Mineralogist* **80**, 549–559.
- Annen, C. (2009), 'From plutons to magma chambers: Thermal constraints on the accumulation of eruptible silicic magma in the upper crust', *Earth and Planetary Science Letters* **284**(3-4), 409–416.
- Annen, C. (2011), 'Implications of incremental emplacement of magma bodies for magma differentiation, thermal aureole dimensions and plutonism-volcanism relationships', *Tectonophysics* **500**(1-4), 3–10.
- Annen, C., Blundy, J. D., Leuthold, J. & Sparks, R. S. J. (2015), 'Construction and evolution of igneous bodies: Towards an integrated perspective of crustal magmatism', *Lithos* **230**(1 August), 206–221.
- Arce, J. L., Gardner, J. E. & Macías, J. L. (2013), 'Pre-eruptive conditions of dacitic magma erupted during the 21.7ka Plinian event at Nevado de Toluca volcano, Central Mexico', *Journal of Volcanology and Geothermal Research* **249**, 49–65.
- Archanjo, C. J., Campanha, G. A. C., Salazar, C. A. & Launeau, P. (2012), 'Using AMS combined with mineral shape preferred orientation analysis to understand the emplacement fabrics of the Apiaí gabbro-norite (Ribeira Belt, SE Brazil)', *International Journal of Earth Sciences* **101**(3), 731–745.

- Bachl, C. A., Miller, C. F., Miller, J. S. & Faulds, J. E. (2001), 'Construction of a pluton: Evidence from an exposed cross section of the Searchlight pluton, Eldorado Mountains, Nevada', *Geological Society of America Bulletin* **113**(9), 1213–1228.
- Bachmann, O. & Bergantz, G. (2004), 'On the Origin of Crystal-poor Rhyolites : Extracted from Batholithic Crystal Mushes', *Journal of Petrology* **45**(8), 1565–1582.
- Bachmann, O. & Bergantz, G. (2008), 'Rhyolites and their Source Mushes across Tectonic Settings', *Journal of Petrology* **49**(12), 2277–2285.
- Bachmann, O. & Bergantz, G. W. (2006), 'Gas percolation in upper-crustal silicic crystal mushes as a mechanism for upward heat advection and rejuvenation of near-solidus magma bodies', *Journal of Volcanology and Geothermal Research* **149**(1-2), 85–102.
- Bachmann, O. & Dungan, M. A. (2002), 'Temperature-induced Al-zoning in hornblendes of the Fish Canyon magma, Colorado', *American Mineralogist* **87**, 1062–1076.
- Bachmann, O. & Huber, C. (2018), 'The inner workings of crustal distillation columns; the physical mechanisms and rates controlling phase separation in silicic magma reservoirs', *Journal of Petrology* p. epy103.
- Baker, J., Peate, D., Waight, T. & Meyzen, C. (2004), 'Pb isotopic analysis of standards and samples using a 207Pb-204Pb double spike and thallium to correct for mass bias with a double-focusing MC-ICP-MS', *Chemical Geology* **211**(3-4), 275–303.
- Bando, M., Bignall, G., Sekine, K. & Tsuchiya, N. (2003), 'Petrography and uplift history of the Quaternary Takidani Granodiorite : could it have hosted a supercritical (HDR) geothermal reservoir?', *Journal of Volcanology and Geothermal Research* **120**, 215–234.
- Bando, M. & Tsuchiya, N. (2000), Petrologic characterization of the Quaternary volcanic-plutonic system: The Takidani Pluton and associated volcanic rocks in the Japan Alps, Japan, in 'Proceedings World Geothermal Congress 2000', Vol. May 28 - J, Kyushu-Tohoku, Japan, pp. 3637–3641.
- Barboni, M. & Schoene, B. (2014), 'Short eruption window revealed by absolute crystal growth rates in a granitic magma', *Nature Geoscience* **7**(June), 524–528.
- Barnes, C. G., Berry, R., Barnes, M. A. & Ernst, W. G. (2017), 'Trace element zoning in hornblende: Tracking and modeling the crystallization of a calc-alkaline arc pluton', *American Mineralogist* **102**(12), 2390–2405.
- Beckhoff, B., Kanngießer, B., Langhoff, N., Wedell, R. & Wolff, H. (2007), *Handbook of practical X-ray fluorescence analysis*, Springer Science & Business Media.
- Bedrosian, P. A., Peacock, J. R., Bowles-Martinez, E., Schultz, A. & Hill, G. J. (2018), 'Crustal inheritance and a top-down control on arc magmatism at Mount St Helens', *Nature Geoscience* **11**(11), 865–870.

- Bernet, M. & Garver, J. I. (2005), 'Fission-track Analysis of Detrital Zircon', *Reviews in Mineralogy and Geochemistry* **58**(1), 205–237.
- Bhattacharyya, D. (1966), 'Orientation of mineral lineation along the flow direction in rocks', *Tectonophysics* **3**(1), 29–33.
- Blatter, D. L., Sisson, T. W. & Hankins, W. B. (2013), 'Crystallization of oxidized, moderately hydrous arc basalt at mid- to lower-crustal pressures: Implications for andesite genesis', *Contributions to Mineralogy and Petrology* **166**(3), 861–886.
- Blundy, J. D. & Annen, C. J. (2016), 'Crustal magmatic systems from the perspective of heat transfer', *Elements* **12**(2), 115–120.
- Blundy, J. D. & Cashman, K. (2001), 'Ascent-driven crystallisation of dacite magmas at Mount St Helens, 1980–1986', *Contributions to Mineralogy and Petrology* **140**, 631–650.
- Blundy, J. D. & Holland, T. J. B. (1990), 'Calcic amphibole equilibria and a new amphibole-plagioclase geothermometer', *Contributions to Mineralogy and Petrology* **104**(2), 208–224.
- Bouchez, J. L. (1997), Granite is never isotropic: an introduction to ams studies of granitic rocks, in 'Granite: from segregation of melt to emplacement fabrics', Springer, pp. 95–112.
- Boudreau, A. (2016), 'Bubble migration in a compacting crystal-liquid mush', *Contributions to Mineralogy and Petrology* **171**(4), 32.
- Bowen, N. (1928), *The Evolution of Igneous Rocks*, Princeton University Press, Princeton.
- Branney, M. & Acocella, V. (2015), Chapter 16 - calderas, in H. Sigurdsson, ed., 'The Encyclopedia of Volcanoes (Second Edition)', second edition edn, Academic Press, Amsterdam, pp. 299–315.
- Brophy, J. G. (1991), 'Composition gaps, critical crystallinity, and fractional crystallization in orogenic (calc-alkaline) magmatic systems', *Contributions to Mineralogy and Petrology* **109**(2), 173–182.
- Burgisser, A. & Bergantz, G. W. (2011), 'A rapid mechanism to remobilize and homogenize highly crystalline magma bodies.', *Nature* **471**(7337), 212–215.
- Callahan, C. N. & Markley, M. J. (2003), 'A record of crustal-scale stress; igneous foliation and lineation in the mount waldo pluton, waldo county, maine', *Journal of Structural Geology* **25**(4), 541–555.
- Cardozo, N. & Allmendinger, R. W. (2013), 'Spherical projections with osxstereonet', *Computers & Geosciences* **51**, 193–205.
- Caricchi, L. & Blundy, J. (2015a), 'Experimental petrology of monotonous intermediate magmas', *Geological Society, London, Special Publications* **422**, SP422–9.

- Caricchi, L. & Blundy, J. (2015b), 'The temporal evolution of chemical and physical properties of magmatic systems', *Geological Society, London, Special Publications* **422**, SP422–11.
- Caricchi, L., Burlini, L., Ulmer, P., Gerya, T., Vassalli, M. & Papale, P. (2007), 'Non-newtonian rheology of crystal-bearing magmas and implications for magma ascent dynamics', *Earth and Planetary Science Letters* **264**(3–4), 402–419.
- Caricchi, L., Sheldrake, T. E. & Blundy, J. (2018), 'Modulation of magmatic processes by CO₂ flushing', *Earth and Planetary Science Letters* **491**, 160–171.
- Carmichael, I. S. E. & Nicholls, J. (1967), 'Iron-titanium oxides and oxygen fugacities in volcanic rocks', *Journal of Geophysical Research* **72**(18), 4665–4687.
- Cashman, K. V. & Giordano, G. (2014), 'Calderas and magma reservoirs', *Journal of Volcanology and Geothermal Research* **288**, 28–45.
- Chiaradia, M., Müntener, O. & Beate, B. (2011), 'Enriched Basaltic Andesites from Mid-crustal Fractional Crystallization, Recharge, and Assimilation (Pilavo Volcano, Western Cordillera of Ecuador)', *Journal of Petrology* **52**(6), 1107–1141.
- Christiansen, R. L. (1984), 'Yellowstone magmatic evolution: Its bearing on understanding large-volume explosive volcanism.', *Explosive Volcanism: Inception, Evolution, and Hazards*. pp. 84–95.
- Christopher, T. E., Blundy, J., Cashman, K., Cole, P., Edmonds, M., Smith, P. J., Sparks, R. S. J., Stinton, A., Christopher, T., Blundy, J., Cashman, K., Cole, P., Edmonds, M., Smith, P. J., Sparks, R. S. J. & Stinton, A. (2015), 'Crustal-scale degassing due to magma system destabilization and magma-gas decoupling at Soufriere Hills Volcano, Montserrat', *Geochemistry, Geophysics, Geosystems* **16**(9), 1–15.
- Coleman, D. S., Bartley, J. M., Glazner, A. F. & Pardue, M. J. (2012), 'Is chemical zonation in plutonic rocks driven by changes in source magma composition or shallow-crustal differentiation?', *Geosphere* **8**(6), 1568–1587.
- Coleman, D. S., Gray, W. & Glazner, A. F. (2004), 'Rethinking the emplacement and evolution of zoned plutons: Geochronologic evidence for incremental assembly of the Tuolumne Intrusive Suite, California', *Geology* **32**(5), 433–436.
- Cooper, K. M. (2017), 'What does a magma reservoir look like? the "crystal's-eye" view', *Elements* **13**(1), 23–28.
- Costa, F., Scaillet, B. & Pichavant, M. (2004), 'Petrological and Experimental Constraints on the Pre-eruption Conditions of Holocene Dacite from Volcan San Pedro (36°S, Chilean Andes) and the Importance of Sulphur in Silicic Subduction-related Magmas', *Journal of Petrology* **45**(4), 855–881.

- Couch, S., Sparks, R. S. J. & Carroll, M. R. (2001), 'Mineral disequilibrium in lavas explained by convective self-mixing in open magma chambers.', *Nature* **411**(6841), 1037–1039.
- Davis, R. H. & Acrivos, A. (1985), 'Sedimentation of noncolloidal particles at low reynolds numbers', *Annual Review of Fluid Mechanics* **17**(1), 91–118.
- de Saint-Blanquat, M., Habert, G., Horsman, E., Morgan, S. S., Tikoff, B., Launeau, P. & Gleizes, G. (2006), 'Mechanisms and duration of non-tectonically assisted magma emplacement in the upper crust: The Black Mesa pluton, Henry Mountains, Utah', *Tectonophysics* **428**(1-4), 1–31.
- de Saint-Blanquat, M., Law, R. D., Bouchez, J. L. & Morgan, S. S. (2001), 'Internal structure and emplacement of the Papoose Flat pluton: An integrated structural, petrographic, and magnetic susceptibility study', *Bulletin of the Geological Society of America* **113**(8), 976–995.
- Deering, C. D., Bachmann, O., Dufek, J. & Gravley, D. M. (2011), 'Rift-Related Transition from Andesite to Rhyolite Volcanism in the Taupo Volcanic Zone (New Zealand) Controlled by Crystal-melt Dynamics in Mush Zones with Variable Mineral Assemblages', *Journal of Petrology* **52**(11), 2243–2263.
- Degruyter, W. & Huber, C. (2014), 'A model for eruption frequency of upper crustal silicic magma chambers', *Earth and Planetary Science Letters* **403**, 117–130.
- Delaney, P. T., Sass, J. H., Duffield, W. A. & Kauahikaua, J. P. (1995), 'Temperature and response time of magma hydrothermal systems, with an example from Kilauea, Hawaii', *[Proceedings]* pp. 1239–1244.
- DePaolo, D. J. (1981), 'Trace element and isotopic effects of combined wallrock assimilation and fractional crystallization', *Earth and Planetary Science Letters* **53**(2), 189–202.
- Druitt, T. H., Costa, F., Deloule, E., Dungan, M. & Scaillet, B. (2012), 'Decadal to monthly timescales of magma transfer and reservoir growth at a caldera volcano.', *Nature* **482**(7383), 77–80.
- Dufek, J. & Bachmann, O. (2010), 'Quantum magmatism: Magmatic compositional gaps generated by melt-crystal dynamics', *Geology* **38**(8), 687–690.
- Dufek, J. & Bergantz, G. (2005), 'Lower Crustal Magma Genesis and Preservation : a Stochastic Framework for the Evaluation of Basalt–Crust Interaction', *Journal of Petrology* **46**(11), 2167–2195.
- Dutrow, B. L., Travis, B. J., Gable, C. W. & Henry, D. J. (2001), 'Coupled heat and silica transport associated with dike intrusion into sedimentary rock: effects on isotherm location and permeability evolution', *Geochimica et Cosmochimica Acta* **65**(21), 3749–3767.

- Ellis, B. S., Bachmann, O. & Wolff, J. a. (2014), 'Cumulate fragments in silicic ignimbrites: The case of the Snake River Plain', *Geology* **42**(5), 431–434.
- Fowler, A. C., Rust, A. C. & Vynnycky, M. (2015), 'The formation of vesicular cylinders in pahoehoe lava flows', *Geophysical and Astrophysical Fluid Dynamics* **109**(1), 39–61.
- Gelman, S. E., Deering, C. D., Bachmann, O., Huber, C. & Gutiérrez, F. J. (2014), 'Identifying the crystal graveyards remaining after large silicic eruptions', *Earth and Planetary Science Letters* **403**, 299–306.
- Ghiorso, M. S. & Evans, B. W. (2008), 'Thermodynamics of Rhombohedral Oxide Solid Solutions and a Revision of the Fe-Ti Two-Oxide Geothermometer and Oxygen-Barometer', *American Journal of Science* **308**(9), 957–1039.
- Ghiorso, M. S. & Gualda, G. a. R. (2015), 'An H₂O–CO₂ mixed fluid saturation model compatible with rhyolite-MELTS', *Contributions to Mineralogy and Petrology* **169**(6), 53.
- Giordano, D., Russell, J. K. & Dingwell, D. B. (2008), 'Viscosity of magmatic liquids: A model', *Earth and Planetary Science Letters* **271**(1–4), 123–134.
- Glazner, A. F., Bartley, J. M., Coleman, D. S., Gray, W. & Taylor, R. Z. (2004), 'Are plutons assembled over millions of years by amalgamation from small magma chambers?', *GSA Today* **14**(4), 1–8.
- Glazner, A. F., Coleman, D. S. & Bartley, J. M. (2008), 'The tenuous connection between high-silica rhyolites and granodiorite plutons', *Geology* **36**(2), 183–186.
- Glazner, A. F., Coleman, D. S. & Mills, R. D. (2015), The Volcanic-Plutonic Connection, in 'Advances in Volcanology', Springer International Publishing Switzerland, pp. 1–22.
- Glazner, A. F., Coleman, D. S. & Mills, R. D. (2018), The volcanic-plutonic connection, in 'Physical Geology of Shallow Magmatic Systems', Springer, pp. 61–82.
- Gualda, G. A., Gravley, D. M., Connor, M., Hollmann, B., Pamukcu, A. S., Bégué, F., Ghiorso, M. S. & Deering, C. D. (2018), 'Climbing the crustal ladder: Magma storage-depth evolution during a volcanic flare-up', *Science advances* **4**(10), eaap7567.
- Gualda, G. A. R. & Ghiorso, M. S. (2014), 'Phase-equilibrium geobarometers for silicic rocks based on rhyolite-MELTS. Part 1: Principles, procedures, and evaluation of the method', *Contributions to Mineralogy and Petrology* **168**, 1033.
- Gualda, G. A. R., Ghiorso, M. S., Lemons, R. V. & Carley, T. L. (2012), 'Rhyolite-MELTS: a Modified Calibration of MELTS Optimized for Silica-rich, Fluid-bearing Magmatic Systems', *Journal of Petrology* **53**, 875–890.
- Hammarstrom, J. M. & Zen, E.-A. (1986), 'Aluminum in hornblende: An empirical igneous geobarometer', *American Mineralogist* **71**(11–12), 1297–1313.

- Hammer, J. E. & Rutherford, M. J. (2002), 'An experimental study of the kinetics of decompression-induced crystallization in silicic melt', *Journal of Geophysical Research: Solid Earth* **107**(B1), ECV 8–1–ECV 8–24.
- Harayama, S. (1990), 'Geology of the Kamigochi district with geological sheet map at 1:50000'.
- Harayama, S. (1992), 'Geology Youngest exposed granitoid pluton on Earth : Cooling and rapid uplift of the Pliocene-Quaternary Takidani Granodiorite in the Japan Alps, central Japan', *Geology* **20**(July), 657–660.
- Harayama, S. (1994), 'Cooling History of the youngest exposed pluton in the World - The Plio-Pleistocene Takidani Granodiorite (Japan Alps, central Japan)', *Memoirs of the Geological Society of Japan* **43**, 87–97.
- Harayama, S. (1999), 'Multistage upheaval and tectonics in the Hida Mountain Range', *Monthly Earth* **21**(9), 603–607.
- Harayama, S., Takeuchi, M., Nakano, S., Sato, T. & Takizawa, F. (1991), 'Geology of the Yurigatake district with geological sheet map at 1:50000'.
- Harayama, S., Wada, H. & Yamaguchi, Y. (2003), 'Trip A1 Quaternary and Pliocene granites in the Northern Japan Alps'.
- Harrison, T. M., Watson, E. B. & Aikman, A. B. (2007), 'Temperature spectra of zircon crystallization in plutonic rocks', *Geology* **35**(7), 635–638.
- Hartung, E., Caricchi, L., Floess, D., Wallis, S., Harayama, S., Kouzmanov, K. & Chiaradia, M. (2017), 'Evidence for residual melt extraction in the Takidani Pluton, Central Japan', *Journal of Petrology* **58**(4), 763–788.
- Hess, K. & Dingwell, D. (1996), 'Viscosities of hydrous leucogranitic melts: A non-arrhenian model', *American Mineralogist* **81**(9-10), 1297–1300.
- Hildreth, W. (1981), 'Gradients in silicic magma chambers: Implications for lithospheric magmatism', *Journal of Geophysical Research: Solid Earth* **86**(B11), 10153–10192.
- Hildreth, W. (2004), 'Volcanological perspectives on Long Valley , Mammoth Mountain , and Mono Craters : several contiguous but discrete systems', *Journal of Volcanology and Geothermal Research* **136**, 169–198.
- Hildreth, W. & Wilson, C. J. N. (2007), 'Compositional Zoning of the Bishop Tuff', *Journal of Petrology* **48**(5), 951–999.
- Holland, T. & Blundy, J. D. (1994), 'Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry', *Contributions to Mineralogy and Petrology* **116**, 433–447.

- Holness, M. B. (2018), 'Melt segregation from silicic crystal mushes: a critical appraisal of possible mechanisms and their microstructural record', *Contributions to Mineralogy and Petrology* **173**(6), 1–17.
- Holness, M. B., Vukmanovic, Z. & Mariani, E. (2017), 'Assessing the role of compaction in the formation of adcumulates: A microstructural perspective', *Journal of Petrology* **58**(4), 643–674.
- Holtz, F., Sato, H., Lewis, J., Behrens, H. & Nakada, S. (2005), 'Experimental petrology of the 1991–1995 Unzen dacite, Japan. Part I: Phase relations, phase composition and pre-eruptive conditions', *Journal of Petrology* **46**(2), 319–337.
- Housh, T. B. & Luhr, J. F. (1991), 'Plagioclase-melt equilibria in hydrous systems', *American Mineralogist* **76**(3–4), 477–492.
- Huang, H.-H., Lin, F.-C., Schmandt, B., Farrell, J., Smith, R. B. & Tsai, V. C. (2015), 'The Yellowstone magmatic system from the mantle plume to the upper crust', *Science* **348**(6236), 773 LP – 776.
- Huber, C., Bachmann, O. & Dufek, J. (2010), 'The limitations of melting on the reactivation of silicic mushes', *Journal of Volcanology and Geothermal Research* **195**(2–4), 97–105.
- Huber, C., Bachmann, O. & Dufek, J. (2011), 'Thermo-mechanical reactivation of locked crystal mushes: Melting-induced internal fracturing and assimilation processes in magmas', *Earth and Planetary Science Letters* **304**(3–4), 443–454.
- Huber, C., Bachmann, O. & Dufek, J. (2012), 'Crystal-poor versus crystal-rich ignimbrites: A competition between stirring and reactivation', *Geology* **40**(2), 115.
- Huber, C., Bachmann, O. & Manga, M. (2009), 'Homogenization processes in silicic magma chambers by stirring and mushification (latent heat buffering)', *Earth and Planetary Science Letters* **283**, 38–47.
- Huber, C. & Parmigiani, A. (2018), 'A Physical Model for Three-Phase Compaction in Silicic Magma Reservoirs', *Journal of Geophysical Research: Solid Earth* pp. 2685–2705.
- Hutton, D. H. W. (1988), 'Granite emplacement mechanisms and tectonic controls: inferences from deformation studies', *Transactions of the Royal Society of Edinburgh: Earth Sciences* **79**, 245–255.
- Ito, H., Spencer, C. J., Danišík, M. & Hoiland, C. W. (2017), 'Magmatic tempo of Earth's youngest exposed plutons as revealed by detrital zircon U-Pb geochronology', *Scientific reports* **7**(1), 12457.
- Jackson, S. E. (2008a), Calibration Strategies for Elemental Analysis by LA-ICP-MS, in P. J. Sylvester, ed., 'Laser Ablation-ICP-Mass Spectrometry in the Earth Sciences: Current Practices and Outstanding Issues', 40 edn, Mineralogical Association of Canada Short Course Series, pp. 169–188.

- Jackson, S. E. (2008b), LAMTRACE data reduction software for LA-ICP-MS, in P. J. Sylvester, ed., 'Laser Ablation-ICP-Mass Spectrometry in the Earth Sciences: Current Practices and Outstanding Issues', 40 edn, Mineralogical Association of Canada Short Course Series, pp. 305–307.
- Jackson, S. E., Longerich, H. P., Dunning, G. R. & Fryer, B. J. (1992), 'The application of laser-ablation microprobe-inductively coupled plasma-mass spectrometry (LAM-ICP-MS) to in situ trace element determination in minerals', *Canadian Mineralogist* **30**, 1049–1064.
- Jahn, B.-M. (2010), 'Accretionary orogen and evolution of the Japanese islands: Implications from a Sr-Nd isotopic study of the Phanerozoic granitoids from SW Japan', *American Journal of Science* **310**(10), 1210–1249.
- Jenkins, R. (1995), *Quantitative X-ray spectrometry*, CRC Press.
- Jenkins, R. (2000), 'X-ray techniques: overview', *Encyclopedia of analytical chemistry* pp. 13269–13288.
- Johannes, W. & Holtz, F. (1996), *Petrogenesis and Experimental Petrology of Granitic Rocks*, Springer-Verlag, Berlin, Heidelberg, New York.
- Johnson, M. C. & Rutherford, M. J. (1989), 'Experimentally Determined Conditions in the Fish Canyon Tuff, Colorado, Magma Chamber', *Journal of Petrology* **30**(3), 711–737.
- Kano, S., Kishi, K., Saito, K., Shiro, C., Tsuchiya, N. & Nakatsuka, K. (2000), 'Field observation of a cooling joint system in the Quarternary Takidani Pluton, Japan', *Proceedings World Geothermal Congress 2000* pp. 3761–3766.
- Kataoka, K., Nagahashi, Y. & Yoshikawa, S. (2001), 'An extremely large magnitude eruption close to the Plio-Pleistocene boundary: reconstruction of eruptive style and history of the Ebisutoge-Fukuda tephra, central Japan', *Journal of Volcanology and Geothermal Research* **107**(1-3), 47–69.
- Kavanagh, J. L., Menand, T. & Sparks, R. S. J. (2006), 'An experimental investigation of sill formation and propagation in layered elastic media', *Earth and Planetary Science Letters* **245**(3-4), 799–813.
- Kawakatsu, K. & Yamaguchi, Y. (1987), 'Successive zoning of amphiboles during progressive oxidation in the Daito-Yokota granitic complex, San-in belt, southwest Japan', *Geochimica et Cosmochimica Acta* **51**(3), 535–540.
- Kawano, Y. & Ueda, Y. (1965), 'K-A dating on the igneous rocks in Japan (iii)', *The Journal of the Japanese Association of Mineralogists, Petrologists and Economic Geologists* **54**(5), 162–172.

- Kim, H., Cho, H., Ok Jeong, J., Son, M. & Sohn, Y. (2015), 'Magnetic fabric changes through thermal treatment: a case study on the cretaceous gusandong tuff in the gyeongsang basin, korea', *51*, 171–190.
- Kimura, J. I., Manton, W. I., Sun, C. H., Iizumi, S., Yoshida, T. & Stern, R. J. (2002), 'Chemical diversity of the Ueno Basalts, central Japan: Identification of mantle and crustal contributions to arc basalts', *Journal of Petrology* **43**(10), 1923–1946.
- Kimura, J.-I. & Nagahashi, Y. (2007), 'Origin of a voluminous iron-enriched high-K rhyolite magma erupted in the North Japan Alps at 1.75 Ma: Evidence for upper crustal melting', *Journal of Volcanology and Geothermal Research* **167**(1-4), 81–99.
- Kimura, J.-I., Nagahashi, Y., Satoguchi, Y. & Chang, Q. (2015), 'Origins of felsic magmas in Japanese subduction zone: Geochemical characterization of tephra from caldera-forming eruptions <5 Ma', *Geochemistry, Geophysics, Geosystems* **18**(1-2), 2147–2174.
- Kimura, J.-I. & Yoshida, T. (1999), 'Mantle diapir-induced arc volcanism: The Ueno Basalts, Nomugi-Toge and Hida volcanic suites, central Japan', *The Island Arc* **8**, 304–322.
- Lange, R. A. (1994), 'The Effects of H₂O, CO₂ and F on the Density and Viscosity of Silicate Melts', *Reviews in Mineralogy and Geochemistry* **30**(Chapter 9), 331–370.
- Launeau, P. & Cruden, A. R. (1998), 'Magmatic fabric acquisition mechanisms in a syenite: Results of a combined anisotropy of magnetic susceptibility and image analysis study', *Journal of Geophysical Research: Solid Earth* **103**(B3), 5067–5089.
- Le Maitre, R. W. (1989), *A Classification of Igneous Rocks and Glossary of Terms: Recommendations of the International Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks*, Language in society, international edn, Blackwell.
- Leake, B. E., Woolley, A. R., Arps, C. E. S., Birch, W. D., Gilbert, M. C., Grice, J. D., Hawthorne, F. C., Kato, A., Kisch, H. J., Krivovichev, V. G., Linthout, K., Laird, J., Mandarino, J. A., Maresch, W. V., Nickel, E. H., Rock, N. M. S., Schumacher, J. C., Smith, D. C., Stephenson, N. C. N., Ungaretti, L., Whittaker, E. J. W. & Youzhi, G. (1997), 'Nomenclature of Amphiboles: Report of the Subcommittee on Amphiboles of the International Mineralogical Association Commission on New Minerals and Mineral Names', *The Canadian Mineralogist* **35**, 219–246.
- Lee, C.-T. A. & Morton, D. M. (2015), 'High silica granites: Terminal porosity and crystal settling in shallow magma chambers', *Earth and Planetary Science Letters* **409**, 23–31.
- Lee, C.-T. A., Morton, D. M., Farner, M. J. & Moitra, P. (2015), 'Field and model constraints on silicic melt segregation by compaction/hindered settling: The role of water and its effect on latent heat release', *American Mineralogist* **100**(8-9), 1762–1777.

- Leuthold, J., Müntener, O., Baumgartner, L. P., Putlitz, B., Ovtcharova, M. & Schaltegger, U. (2012), 'Time resolved construction of a bimodal laccolith (Torres del Paine, Patagonia)', *Earth and Planetary Science Letters* **325-326**, 85–92.
- Lipman, P., Dungan, M. & Bachmann, O. (1997), 'Comagmatic granophyric granite in the Fish Canyon Tuff, Colorado: Implications for magma-chamber processes during a large ash-flow eruption', *Geology* **25**(10), 915–918.
- Lipman, P. W. (2007), 'Incremental assembly and prolonged consolidation of Cordilleran magma chambers: Evidence from the Southern Rocky Mountain volcanic field', *Geosphere* **3**(1), 42–70.
- Longerich, H. (2008), 'Laser ablation-Inductively coupled plasma-mass spectrometry (LA-ICP-MS); an introduction', *Mineralogical Association of Canada Short Course Series* **40**, 1–18.
- Longerich, H. P., Gunther, D. & Jackson, S. E. (1996), 'Elemental fractionation in laser ablation inductively coupled plasma mass spectrometry', *Fresenius Journal of Analytical Chemistry* **355**, 538–542.
- Lundstrom, C. C. & Glazner, A. F. (2016), 'Silicic Magmatism and the Volcanic–Plutonic Connection', *Elements* **12**(2), 91–96.
- Lyell, C. (1838), *Elements of Geology*, John Murray, London.
- Mamtani, M. A. & Greiling, R. (2005), 'Granite emplacement and its relation with regional deformation in the aravalli mountain belt (india)?inferences from magnetic fabric', *Journal of structural Geology* **27**(11), 2008–2029.
- Mamtani, M. A., Piazzolo, S., Greiling, R. O., Kontny, A. & Hrouda, F. (2011), 'Process of magnetite fabric development during granite deformation', *Earth and Planetary Science Letters* **308**(1-2), 77–89.
- Marsh, B. D. (1981), 'On the crystallinity, probability of occurrence, and rheology of lava and magma', *Contributions to Mineralogy and Petrology* **78**(1), 85–98.
- Martin, R. F. (2007), 'Amphiboles in the Igneous Environment', *Reviews in Mineralogy and Geochemistry* **67**(Chapter 9), 323–358.
- Matzel, J. E. P., Bowring, S. A. & Miller, R. B. (2006), 'Time scales of pluton construction at differing crustal levels: Examples from the Mount Stuart and Tenpeak intrusions, North Cascades, Washington', *Bulletin of the Geological Society of America* **118**(11-12), 1412–1430.
- Maughan, L. L., Christiansen, E. H., Best, M. G., Gromme, C. S., Deino, A. L. & Tingey, D. G. (2002), 'The Oligocene Lund Tuff, Great Basin, USA : a very large volume monotonous intermediate', *Journal of Volcanology and Geothermal Research* **113**, 129–157.

- McDonough, W. F. & Sun, S.-S. (1995), 'The composition of the earth', *Chemical geology* **120**(3-4), 223–253.
- McKenzie, D. (1984), 'The Generation and Compaction of Partially Molten Rock', *Journal of Petrology* **25**(3), 713–765.
- Melekhova, E., Annen, C. & Blundy, J. (2013), 'Compositional gaps in igneous rock suites controlled by magma system heat and water content', *Nature Geoscience* **6**(5), 385–390.
- Menand, T. (2008), 'The mechanics and dynamics of sills in layered elastic rocks and their implications for the growth of laccoliths and other igneous complexes', *Earth and Planetary Science Letters* **267**(1-2), 93–99.
- Menand, T. (2011), 'Physical controls and depth of emplacement of igneous bodies: A review', *Tectonophysics* **500**(1-4), 11–19.
- Menand, T., Annen, C. & de Saint Blanquat, M. (2015), 'Rates of magma transfer in the crust: Insights into magma reservoir recharge and pluton growth', *Geology* **43**(3), 199–203.
- Michel, J., Baumgartner, L., Putlitz, B., Schaltegger, U. & Ovtcharova, M. (2008), 'Incremental growth of the Patagonian Torres del Paine laccolith over 90 k.y', *Geology* **36**(6), 459–462.
- Miller, C. & Wark, D. (2008), 'SUPERVOLCANOES AND THEIR EXPLOSIVE SUPERERUPTIONS', *Elements* **4**(1), 11.
- Morioka, K., Tainosho, Y. & Kagami, H. (2000), 'Rb–sr isochron ages of the cretaceous granitoids in the ryoke belt, kinki district, southwest japan', *Island Arc* **9**(1), 46–54.
- Müntener, O. & Ulmer, P. (2018), 'Arc crust formation and differentiation constrained by experimental petrology', *American Journal of Science* **318**(1), 64–89.
- Murphy, M. D., Sparks, R. S. J., Barclay, J., Carroll, M. R. & Brewer, T. S. (2000), 'Remobilization of Andesite Magma by Intrusion of Mafic Magma at the Soufriere Hills Volcano, Montserrat, West Indies', *Journal of Petrology* **41**(1), 21–42.
- Nagahashi, Y. (1995), 'Plio-Pleistocene volcanoclastic formation in the Takayama Basin, Gifu prefecture - Stratigraphy and Petrography -', *Earth Science (Chikyu Kagaku)* **2**(49), 109–124.
- Nagahashi, Y., Kosaka, T. & Hibi, N. (1996), 'Correlation of the late Pliocene large-scale pyroclastic flow deposits, Gifu and Nagano prefectures, central Japan - case study of the Nyukawa Pyroclastic Flow Deposit, Ebisutoge Pyroclastic Deposits and their correlative deposits', *Earth Science (Chikyu Kagaku)* **50**, 29–42.

- Nagahashi, Y., Satoguchi, Y. & Yoshikawa, S. (2000), 'Correlation and stratigraphic eruption age of the pyroclastic flow deposits and wide spread volcanic ashes intercalated in the Pliocene-Pleistocene strata, central Japan', *The Journal of the Geological Society of Japan* **106**(1), 51–69.
- Nakajima, T. (1994), 'The ryoke plutonometamorphic belt: crustal section of the cretaceous eurasian continental margin', *Lithos* **33**(1), 51–66. Tectonics, metamorphism and magmatism in island arcs.
- Nandedkar, R. H., Ulmer, P. & Müntener, O. (2014), 'Fractional crystallization of primitive, hydrous arc magmas: an experimental study at 0.7 GPa', *Contributions to Mineralogy and Petrology* **167**(6), 1015.
- Newman, S. & Lowenstern, J. B. (2002), 'VolatileCalc: a silicate melt–H₂O–CO₂ solution model written in Visual Basic for excel', *Computers & Geosciences* **28**(5), 597–604.
- Oikawa, T. (2003), 'The Spatial and Temporal Relationship between Uplifting and Magmatism in the Hida Mountain Range, Central Japan', *The Quaternary Research* **42**(3), 141–156.
- Otten, M. T. (1984), 'The origin of brown hornblende in the Artfjället gabbro and dolerites', *Contributions to Mineralogy and Petrology* **86**(2), 189–199.
- Pallister, J. S., Hoblitt, R. P. & Reyes, A. G. (1992), 'A basalt trigger for the 1991 eruptions of Pinatubo volcano?', *Nature* **356**(2 April), 426–428.
- Pamukcu, A. S., Gualda, G. A. R., Ghiorso, M. S., Miller, C. F. & McCracken, R. G. (2015), 'Phase-equilibrium geobarometers for silicic rocks based on rhyolite?MELTS?Part 3: Application to the Peach Spring Tuff (Arizona?California?Nevada, USA)', *Contributions to Mineralogy and Petrology* **169**(33), 1–17.
- Papale, P. & Barbato, D. (2006), 'The compositional dependence of the saturation surface of H₂O+CO₂ fluids in silicate melts', *Chemical Geology* **229**(1), 78–95.
- Parmigiani, A., Faroughi, S., Huber, C., Bachmann, O. & Su, Y. (2016), 'Bubble accumulation and its role in the evolution of magma reservoirs in the upper crust', *Nature* **532**(7600), 492–495.
- Parmigiani, A., Huber, C. & Bachmann, O. (2014), 'Mush microphysics and the reactivation of crystal-rich magma reservoirs', *Journal of Geophysical Research: Solid Earth* pp. n/a–n/a.
- Pasquarè, F. & Tibaldi, A. (2007), 'Structure of a sheet-laccolith system revealing the interplay between tectonic and magma stresses at Stardalur Volcano, Iceland', *Journal of Volcanology and Geothermal Research* **161**(1-2), 131–150.

- Paterson, S. R., Fowler, T., Schmidt, K. L., Yoshinobu, A. S., Yuan, E. & Miller, R. B. (1998), 'Interpreting magmatic fabric patterns in plutons', *Lithos* **44**(1-2), 53–82.
- Payacán, I., Gutiérrez, F., Gelman, S. E., Bachmann, O. & Parada, M. Á. (2014), 'Comparing magnetic and magmatic fabrics to constrain the magma flow record in La Gloria pluton, central Chile', *Journal of Structural Geology* **69**(PA), 32–46.
- Pirrie, D., Butcher, A. R., Power, M. R., Gottlieb, P. & Miller, G. L. (2004), 'Rapid quantitative mineral and phase analysis using automated scanning electron microscopy (QemSCAN); potential applications in forensic geoscience', *Geological Society, London, Special Publications* **232**(1).
- Pistone, M., Arzilli, F., Dobson, K. J., Cordonnier, B., Reusser, E., Ulmer, P., Marone, F., Whittington, A. G., Mancini, L., Fife, J. L. & Blundy, J. D. (2015), 'Gas-driven filter pressing in magmas: Insights into in-situ melt segregation from crystal mushes', *Geology* **43**(8), G36766.1.
- Riker, J. M., Blundy, J. D., Rust, A. C., Botcharnikov, R. E. & Humphreys, M. C. S. (2015), 'Experimental phase equilibria of a Mount St. Helens rhyodacite: a framework for interpreting crystallization paths in degassing silicic magmas', *Contributions to Mineralogy and Petrology* **170**(1), 6.
- Robinson, P., Schumacher, J. C. & Spear, F. S. (1982), Formulation of electron probe analyses, in R. P. Veblen DR, ed., 'Amphiboles: petrology and experimental phase relations', Mineralogical Society of America, pp. 6–9.
- Röntgen, W. C. (1895), *Ueber eine neue Art von Strahlen*, Phys.-med. Gesellschaft.
- Rose, W. I. & Chesner, C. A. (1987), 'Dispersal of ash in the great Toba eruption, 75 ka', *Geology* **15**(October), 913–917.
- Rudnick, R. & Gao, S. (2003), Composition of the Continental Crust, in 'Treatise on Geochemistry', Vol. 3, Elsevier, pp. 1–64.
- Rueden, C. T., Schindelin, J., Hiner, M. C., DeZonia, B. E., Walter, A. E., Arena, E. T. & Eliceiri, K. W. (2017), 'ImageJ2: ImageJ for the next generation of scientific image data', *BMC bioinformatics* **18**(1), 529.
- Rust, A. C. & Cashman, K. V. (2011), 'Permeability controls on expansion and size distributions of pyroclasts', *Journal of Geophysical Research: Solid Earth* **116**(11), 1–17.
- Rutherford, M. & Devine, J. D. (1996), Preeruption Pressure-Temperature conditions and volatiles in the 1991 dacitic magma of Mount Pinatubo, in 'Fire and mud: eruptions and lahars of Mount Pinatubo, Philippines', PHIVOLCS and University of Washington, Seattle, pp. 751–766.
- Sano, Y., Tsutsumi, Y., Terada, K. & Kaneoka, I. (2002), 'Ion microprobe U-Pb dating of Quaternary zircon: implication for magma cooling and residence time', *Journal of Volcanology and Geothermal Research* **117**, 285–296.

- Scailliet, B. & Evans, B. W. (1999), 'The 15 June 1991 Eruption of Mount Pinatubo. I. Phase Equilibria and Pre-eruption P-T-fO₂-fH₂O Conditions of the Dacite Magma', *Journal of Petrology* **40**(3), 381–411.
- Schaen, A. J., Cottle, J. M., Singer, B. S., Brenhin Keller, C., Garibaldi, N. & Schoene, B. (2017), 'Complementary crystal accumulation and rhyolite melt segregation in a late Miocene Andean pluton', *Geology* **45**(9), 835–838.
- Schindelin, J., Arganda-Carreras, I., Frise, E., Kaynig, V., Longair, M., Pietzsch, T., Preibisch, S., Rueden, C., Saalfeld, S., Schmid, B. et al. (2012), 'Fiji: an open-source platform for biological-image analysis', *Nature methods* **9**(7), 676.
- Schmidt, M. W. (1992), 'Amphibole composition in tonalite as a function of pressure: an experimental calibration of the Al-in-hornblende barometer', *Contributions to Mineralogy and Petrology* **110**(2-3), 304–310.
- Schöpa, A. & Annen, C. (2013), 'The effects of magma flux variations on the formation and lifetime of large silicic magma chambers', *Journal of Geophysical Research: Solid Earth* **118**(3), 926–942.
- Schöpa, A., Floess, D., de Saint Blanquat, M., Annen, C. & Launeau, P. (2015), 'The relation between magnetite and silicate fabric in granitoids of the Adamello Batholith', *Tectonophysics* **642**, 1–15.
- Sekine, K., Bignall, G., Tsuchiya, N. & Nakatsuka, K. (2001), 'Evidence for Fluid Flow in Non-Brittle Takidani Granite Implications for Utilization of DSGRs', *Geothermal Resources Council Transactions* **25**(August 26-29), 243–248.
- Shane, P. & Smith, V. C. (2013), 'Using amphibole crystals to reconstruct magma storage temperatures and pressures for the post-caldera collapse volcanism at Okataina volcano', *Lithos* **156-159**, 159–170.
- Sharpe, M. R. (1985), 'Strontium isotope evidence for preserved density stratification in the main zone of the Bushveld Complex, South Africa', *Nature* **316**(6024), 119–126.
- Shibata, K. & Ishihara, S. (1979), 'Rb-sr whole-rock and k-ar mineral ages of granitic rocks in Japan', *GEOCHEMICAL JOURNAL* **13**(3), 113–119.
- Sisson, T. W. & Bacon, C. R. (1999), 'Gas-driven filter pressing in magmas', *Geology* **27**(7), 613–616.
- Sisson, T. W. & Grove, T. L. (1993), 'Experimental investigations of the role of H₂O in calc-alkaline differentiation and subduction zone magmatism', *Contributions to Mineralogy and Petrology* **113**, 143–166.
- Sisson, T. W., Ratajeski, A. K., Hankins, A. W. B. & Glazner, A. F. (2005), 'Voluminous granitic magmas from common basaltic sources', *Contributions to Mineralogy and Petrology* **148**, 635–661.

- Sliwinski, J. T., Bachmann, O., Dungan, M. A., Huber, C., Deering, C. D., Lipman, P. W., Martin, L. H. J. & Liebske, C. (2017), 'Rapid pre-eruptive thermal rejuvenation in a large silicic magma body: the case of the Masonic Park Tuff, Southern Rocky Mountain volcanic field, CO, USA', *Contributions to Mineralogy and Petrology* **172**(5), 30.
- Solano, J. M. S., Jackson, M. D., Sparks, R. S. J., Blundy, J. D. & Annen, C. (2012), 'Melt segregation in deep crustal hot zones: A mechanism for chemical differentiation, crustal assimilation and the formation of evolved magmas', *Journal of Petrology* **53**(10), 1999–2026.
- Sparks, R. & Marshall, L. (1986), 'Thermal and mechanical constraints on mixing between mafic and silicic magmas', *Journal of Volcanology and Geothermal Research* **29**(1-4), 99–124.
- Spear, F. S. (1981), 'An experimental study of hornblende stability and compositional variability in amphibolite.', *American Journal of Science* **281**(6), 697–734.
- Spear, F. S. & Kimball, K. L. (1984), 'RECOMP-A FORTRAN IV Program for estimating Fe³⁺ contents in amphiboles', *Computers & Geosciences* **10**(2-3), 317–325.
- Srogi, L. & Lutz, T. M. (1996), 'The role of residual melt migration in producing compositional diversity in a suite of granitic rocks', *Earth and Planetary Science Letters* **144**, 563–576.
- Sylvester, P. J. (2008), *Laser ablation-ICP-MS in the earth sciences: Current practices and outstanding issues*, Vol. 40, Mineralogical Association of Canada.
- Taira, A., Ohara, Y., Wallis, S., Ishiwatari, A. & Iryu, Y. (2016), 'Geological evolution of Japan: an overview', *The geology of Japan. Geological Society of London, London* pp. 1–24.
- Tanaka, T., Togashi, S., Kamioka, H., Amakawa, H., Kagami, H., Hamamoto, T., Yuhara, M., Orihashi, Y., Yoneda, S., Shimizu, H., Kunimaru, T., Takahashi, K., Yanagi, T., Nakano, T., Fujimaki, H., Shinjo, R., Asahara, Y., Tanimizu, M. & Dragusanu, C. (2000), 'JNdi-1: a neodymium isotopic reference in consistency with LaJolla neodymium', *Chemical Geology* **168**(3-4), 279–281.
- Tarling, D. H. & Hrouda, F. (1993), *The Magnetic Anisotropy of Rocks*, first edn, Chapman & Hall, London.
- Thomas, W. M. & Ernst, W. G. (1990), The aluminium content of hornblende in calc-alkaline granitic rocks: a mineralogic barometer calibrated experimentally to 12 kbars*, in R. J. Spencer & I.-M. Chou, eds, 'Fluid-Mineral Interactions: A Tribute to H.P. Eugster', The Geochemical Society, Special Publication No. 2, pp. 59–63.

- Trubač, J., Žák, J., Chlupáčová, M. & Janoušek, V. (2009), 'Magnetic fabric of the Říčany granite, Bohemian Massif: A record of helical magma flow?', *Journal of Volcanology and Geothermal Research* **181**(1), 25–34.
- Tsuchiya, N. & Fujino, K. (2000), 'Evaluation of cooling history of the Quarternary Takidani Pluton using thermoluminescence technique', *Proceedings World Geothermal Congress 2000* pp. 3939–3944.
- Ulmer, P., Kaegi, R. & Müntener, O. (2018), 'Experimentally derived intermediate to silica-rich arc magmas by fractional and equilibrium crystallization at 1.0 GPa: An evaluation of phase relationships, compositions, liquid lines of descent and oxygen fugacity', *Journal of Petrology* **59**(1), 11–58.
- Uo, M., Wada, T. & Sugiyama, T. (2015), 'Applications of x-ray fluorescence analysis (xrf) to dental and medical specimens', *Japanese Dental Science Review* **51**(1), 2 – 9.
- Vernon, R. (2000), 'Review of Microstructural Evidence of Magmatic and Solid-State Flow', *Visual Geosciences* **5**(2), 1–23.
- Vignerresse, J.-L. (2014), 'Textures and melt-crystal-gas interactions in granites', *Geoscience Frontiers* .
- Vignerresse, J. L., Barbey, P. & Cuney, M. (1996), 'Rheological Transitions During Partial Melting and Crystallization with Application to Felsic Magma Segregation and Transfer', *Journal of Petrology* **37**(6), 1579–1600.
- Wakita, K. (2013), 'Geology and tectonics of Japanese islands: A review - The key to understanding the geology of Asia', *Journal of Air Transport Management* **31**, 75–87.
- Wakita, K. (2018), Geology of the japanese islands: an outline, in 'Natural Heritage of Japan', Springer, pp. 9–17.
- Walker, B. A., Bergantz, G. W., Otamendi, J. E., Ducea, M. N. & Cristofolini, E. A. (2015), 'A MASH zone revealed: The mafic complex of the Sierra Valle Fértil', *Journal of Petrology* **56**(9), 1863–1896.
- Watson, E. B. & Harrison, T. M. (1983), 'Zircon saturation revisited: temperature and composition effects in a variety of crustal magma types', *Earth and Planetary Science Letters* **64**(2), 295–304.
- Whitney, J. A. (1975), 'The Effects of Pressure, Temperature, and xH₂O on Phase Assemblage in Four Synthetic Rock Compositions', *The Journal of Geology* **83**(1), 1–31.
- Whitney, J. A. (1988), 'The origin of granite: The role and source of water in the evolution of granitic magmas', *Geological Society of America Bulletin* **100**, 1886–1897.

- Whittington, A. G., Hofmeister, A. M. & Nabelek, P. I. (2009), 'Temperature-dependent thermal diffusivity of the Earth's crust and implications for magmatism.', *Nature* **458**(7236), 319–21.
- Wilke, M. & Behrens, H. (1999), 'The dependence of the partitioning of iron and europium between plagioclase and hydrous tonalitic melt on oxygen fugacity', *Contributions to Mineralogy and Petrology* **137**(1-2), 102–114.
- Wilson, C. J. & Charlier, B. L. (2016), 'The life and times of silicic volcanic systems', *Elements* **12**(2), 103–108.
- Wotzlaw, J.-F., Bindeman, I. N., Watts, K. E., Schmitt, A. K., Caricchi, L. & Schaltegger, U. (2014), 'Linking rapid magma reservoir assembly and eruption trigger mechanisms at evolved Yellowstone-type supervolcanoes', *Geology* **42**(9), 807–810.
- Yamada, N., Kato, H., Ono, K. & Iwata, O. (1985), 'K-Ar ages of some Pliocene to Pleistocene felsic volcanic rocks around the Japanese Northern Alps, central Honshu.', *Bulletin Geological Survey Japan* **36**(10), 539–549.
- Yoshikawa, S., Satoguchi, Y. & Nagahashi, Y. (1996), 'A widespread volcanic ash bed in the horizon close to the Pliocene-Pleistocene boundary : Fukuda-Tsujimatagawa-Kd38 volcanic ash bed occurring in central Japan', *Journal of Geological Society of Japan* **102**(3), 258–270.
- Zak, J., Paterson, S. R. & Memeti, V. (2007), 'Four magmatic fabrics in the Tuolumne batholith, central Sierra Nevada, California (USA): Implications for interpreting fabric patterns in plutons and evolution of magma chambers in the upper crust', *Geological Society of America Bulletin* **119**(1-2), 184–201.

Appendix A

Analytical Conditions

TABLE A.1: Analytical conditions of EPMA analyses of plagioclase and alkali feldspar

Element	Standard	Crystal	Time on peak	Time on background
Si	Wollastonite & Albite	TAP	20 s	10 s
K	Orthoclase	PETH	20 s	10 s
Na	Jadeite & Albite	TAP	20 s	10 s
Ca	Wollastonite	PETJ	30 s	15 s
Fe	Fayalite	LIFH	40 s	20 s
Al	Andalusite & Albite	TAP	20 s	10 s

TABLE A.2: Analytical conditions of EPMA analyses of pyroxene, amphibole and biotite

Element	Standard	Crystal	Time on peak	Time on background
Si	Wollastonite	TAP	30 s	15 s
K	Orthoclase	PETH	20 s	10 s
Na	Jadeite	TAP	20 s	10 s
Ca	Wollastonite	PETJ	30 s	15 s
Fe	Fayalite & Olivine	LIFH	30 s	15 s
Al	Andalusite	TAP	30 s	15 s
Cl	Scapolite	PETH	30 s	15 s
Mg	Olivine	TAP	30 s	15 s
Ti	MnTi-oxide	PETJ	30 s	15 s
Mn	MnTi-oxide	LIFH	30 s	15 s

TABLE A.3: Analytical conditions of EPMA analyses of Fe-oxides

Element	Standard	Crystal	Time on peak	Time on background
Si	Wollastonite	TAP	30 s	15 s
Ca	Wollastonite	PETH	30 s	15 s
Mg	MgO	TAP	30 s	15 s
Ti	MnTi-oxide	PETJ	30 s	15 s
Fe	Hematite	LIFH	30 s	15 s
Al	Andalusite	TAP	30 s	15 s
Ni	NiO	LIFH	30 s	15 s
Cr	Cr ₂ O ₃	PETJ	30 s	15 s
Mn	MnTi-oxide	LIFH	30 s	15 s

TABLE A.4: Analytical conditions of EPMA analyses of pumice and matrix glasses

Element	Standard	Crystal	Time on peak	Time on background
Na	Albite	TAP	20 s	10 s
K	Orthoclase	PETH	20 s	10 s
Si	Wollastonite	TEP	30 s	15 s
Ca	Wollastonite	PETJ	30 s	15 s
Fe	Fayalite	LIFH	30 s	15 s
Mg	Olivine	TAP	40 s	20 s
P	Apatite	PETH	40 s	20 s
Al	Andalusite	TAP	30 s	15 s
Ti	MnTi-oxide	PETJ	30 s	15 s
Mn	MnTi-oxide	LIFH	30 s	15 s

Appendix B

Appendix Chapter 3

TABLE B.1: Major and trace elements and radiogenic isotopes of Takidani pluton along Shiradashi-zawa

[illegible]

TABLE B.2: (Continued)

Unit:	pGT	pGT	mGT	mGT	pGT	pGT	pGT
ID:	EH69	EH70	EH71	EH72	EH74	EH75	EH76
Latitude:	36.2916	36.2918	36.2907	36.2901	36.2918	36.2919	36.2921
Longitude:	137.6294	137.6293	137.6299	137.6301	137.6291	137.6290	137.6287
Elevation:	2040	2035	2110	2160	2020	2010	1995
<i>wt%</i>							
SiO ₂	68.88	69.25	67.68	67.41	67.38	66.43	67.03
TiO ₂	0.48	0.41	0.55	0.59	0.54	0.56	0.54
Al ₂ O ₃	14.67	14.72	14.64	14.88	14.95	15.14	14.84
Fe ₂ O ₃	3.45	3.27	4.11	4.22	4.01	4.23	4.02
MnO	0.05	0.06	0.08	0.08	0.06	0.05	0.06
MgO	1.37	1.27	1.51	1.58	1.63	1.70	1.60
CaO	3.28	3.25	3.47	3.60	3.79	3.66	3.67
Na ₂ O	3.12	3.28	3.40	3.64	3.08	2.83	3.31
K ₂ O	3.28	3.50	3.26	3.04	3.29	2.99	3.29
P ₂ O ₅	0.13	0.12	0.16	0.17	0.15	0.16	0.15
LOI	0.46	0.35	0.37	0.27	0.41	1.39	0.50
Total	99.18	99.47	99.22	99.47	99.29	99.14	99.02
<i>ppm</i>							
TiO ₂	0.56	0.50	0.64	0.67	0.63	0.65	0.63
MnO	0.05	0.05	0.08	0.08	0.05	0.04	0.06
Sc	10	10	11	11	12	12	12
Co	7	6	8	8	8	9	7
Ni	4	4	4	4	5	4	5
Cu	19	5	5	3	29	28	8
Zn	25	29	35	41	29	28	28
Rb	119	108	95	94	107	92	104
Sr	267	266	302	314	288	306	292
Y	17	16	18	19	18	17	18
Zr	130	131	149	153	137	135	139
Nb	8	8	8	9	8	7	8
Mo	12	2	1	0	6	0	0
Cs	6	5	3	3	5	4	4
Ba	470	471	546	533	445	480	491
La	31	29	37	29	29	23	27
Ce	57	53	69	56	55	45	51
Pr	6	5	7	6	6	5	5
Nd	20	18	25	22	20	18	19
Sm	3	3	4	4	4	3	4
Eu	1	1	1	1	1	1	1
Gd	3	3	3	4	3	3	3
Tb	0	0	0	1	0	0	0
Dy	3	3	3	4	3	3	3
Ho	1	1	1	1	1	1	1
Er	2	2	2	2	2	2	2
Tm	0	0	0	0	0	0	0
Yb	2	2	2	2	2	2	2
Lu	0	0	0	0	0	0	0
Hf	4	4	4	4	4	4	4
Ta	1	1	1	1	1	1	1
W	1	1	0	0	0	1	1
Pb	9	13	9	9	11	12	10
Th	18	17	14	11	16	14	15
U	4	4	2	2	4	3	3
¹⁴³ Nd/ ¹⁴⁴ Nd		0.512486		0.512526			0.512540
⁸⁷ Sr/ ⁸⁶ Sr		0.707347		0.707511			0.707324
²⁰⁶ Pb/ ²⁰⁴ Pb		18.40263		18.40788			18.40258
²⁰⁷ Pb/ ²⁰⁴ Pb		15.61703		15.62514			15.61689
²⁰⁸ Pb/ ²⁰⁴ Pb		38.78160		38.80612			38.77832

TABLE B.3: Average composition of plagioclase from TK

	Pop I	Pop II	Pop III	Rim I	Rim II	Rim III	Matrix Plg	Kfs	Matrix Kfs
n:	488	672	442	29	145	99	29	25	33
<i>wt%</i>									
SiO ₂	59.77 (0.81)	57.53 (1.47)	56.82 (1.48)	62.35 (1.38)	63.23 (1.58)	63.00 (1.61)	62.31 (1.06)	65.47 (0.23)	65.27 (0.40)
Al ₂ O ₃	24.66 (0.57)	26.17 (1.02)	26.89 (0.97)	23.00 (0.88)	22.57 (0.96)	22.98 (1.05)	23.07 (0.85)	18.14 (0.19)	18.13 (0.22)
FeO	0.19 (0.02)	0.26 (0.07)	0.32 (0.05)	0.17 (0.03)	0.15 (0.03)	0.18 (0.04)	0.15 (0.03)	0.09 (0.02)	0.10 (0.03)
CaO	7.09 (0.62)	8.89 (1.16)	9.57 (1.17)	5.13 (1.04)	4.50 (1.16)	4.70 (1.26)	4.96 (0.89)	0.07 (0.04)	0.04 (0.03)
Na ₂ O	7.26 (0.34)	6.16 (0.68)	5.99 (0.64)	8.49 (0.60)	8.67 (0.61)	8.78 (0.77)	8.70 (0.49)	1.47 (0.29)	0.12 (0.37)
K ₂ O	0.51 (0.25)	0.44 (0.68)	0.32 (0.13)	0.45 (0.09)	0.50 (0.15)	0.52 (0.14)	0.31 (0.09)	14.23 (0.45)	14.80 (0.51)
Total	99.49 (0.35)	99.45 (0.29)	99.91 (0.51)	99.59 (0.44)	99.61 (0.30)	100.15 (0.49)	99.49 (0.57)	99.47 (0.39)	99.55 (0.48)
<i>mol</i>									
An	0.34 (0.03)	0.43 (0.06)	0.46 (0.06)	0.24 (0.05)	0.22 (0.06)	0.22 (0.06)	0.24 (0.04)	0.00 (0.00)	0.00 (0.00)
Ab	0.63 (0.03)	0.54 (0.06)	0.52 (0.05)	0.73 (0.05)	0.76 (0.05)	0.75 (0.06)	0.75 (0.04)	0.14 (0.03)	0.11 (0.03)
Or	0.03 (0.01)	0.03 (0.04)	0.02 (0.01)	0.03 (0.01)	0.03 (0.01)	0.03 (0.01)	0.02 (0.00)	0.86 (0.03)	0.89 (0.03)
Fe x 10	0.29 (0.03)	0.41 (0.04)	0.48 (0.08)	0.25 (0.04)	0.23 (0.05)	0.26 (0.07)	0.22 (0.05)	0.14 (0.03)	0.15 (0.05)

TABLE B.4: Average composition of high-Al amphibole (5.8–8.0 wt.% Al_2O_3) of TK

n:	mGT 7		pGT 27		GDT 24		GRT 4	
<i>wt%</i>								
SiO ₂	46.64	(0.72)	47.15	(0.52)	47.32	(0.55)	47.47	(0.51)
TiO ₂	1.49	(0.12)	1.65	(0.10)	1.61	(0.15)	1.57	(0.23)
Al ₂ O ₃	7.12	(0.57)	6.80	(0.43)	6.58	(0.36)	6.25	(0.47)
FeO	17.17	(0.33)	15.07	(0.47)	15.36	(0.47)	14.70	(1.29)
MgO	11.71	(0.31)	13.34	(0.44)	13.32	(0.41)	13.91	(0.73)
MnO	0.70	(0.05)	0.31	(0.05)	0.36	(0.09)	0.33	(0.08)
CaO	11.11	(0.08)	11.17	(0.09)	11.14	(0.09)	11.24	(0.12)
Na ₂ O	1.52	(0.11)	1.52	(0.16)	1.46	(0.08)	1.50	(0.12)
K ₂ O	0.64	(0.09)	0.67	(0.06)	0.65	(0.06)	0.65	(0.04)
Cl	0.15	(0.02)	0.14	(0.02)	0.14	(0.04)	0.13	(0.04)
Total	98.24	(0.29)	97.83	(0.38)	97.95	(0.37)	97.74	(0.10)
<i>T-sites</i>								
Si	6.87	(0.09)	6.91	(0.06)	6.94	(0.06)	6.97	(0.07)
Al ^{IV}	1.13	(0.09)	1.09	(0.06)	1.06	(0.06)	1.03	(0.07)
	8.00		8.00		8.00		8.00	
<i>M1-3 sites</i>								
Al ^{VI}	0.11	(0.01)	0.09	(0.03)	0.07	(0.01)	0.05	(0.01)
Ti	0.16	(0.01)	0.18	(0.01)	0.18	(0.02)	0.17	(0.03)
Fe ³⁺	0.51	(0.02)	0.44	(0.06)	0.41	(0.05)	0.30	(0.08)
Mg	2.57	(0.06)	2.91	(0.09)	2.91	(0.08)	3.04	(0.14)
Mn	0.09	(0.01)	0.04	(0.01)	0.04	(0.01)	0.04	(0.01)
Fe ²⁺	1.55	(0.05)	1.34	(0.10)	1.38	(0.07)	1.39	(0.21)
Ca	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
	5.00		5.00		5.00		5.00	
<i>M4 site</i>								
Fe	0.05	(0.01)	0.07	(0.03)	0.09	(0.02)	0.12	(0.04)
Ca	1.75	(0.01)	1.75	(0.02)	1.75	(0.02)	1.77	(0.01)
Na	0.19	(0.01)	0.18	(0.03)	0.16	(0.03)	0.11	(0.05)
	2.00		2.00		2.00		2.00	
<i>A site</i>								
Ca	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
Na	0.24	(0.03)	0.25	(0.04)	0.26	(0.03)	0.32	(0.05)
K	0.12	(0.02)	0.13	(0.01)	0.12	(0.01)	0.12	(0.01)
Sum cations	15.36	(0.05)	15.38	(0.05)	15.38	(0.04)	15.44	(0.05)
<i>OH site</i>								
O	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
OH	1.96	(0.01)	1.96	(0.01)	1.96	(0.01)	1.97	(0.01)
F	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
Cl	0.04	(0.01)	0.04	(0.01)	0.04	(0.01)	0.03	(0.01)
Mg#	0.51	(0.01)	0.58	(0.02)	0.58	(0.02)	0.61	(0.03)

TABLE B.5: Average composition of low-Al amphibole (3.6–5.8 wt.% Al_2O_3) of TK

	mGT 11		pGT 16		GDT 31		GRT 12	
<i>wt%</i>								
SiO ₂	49.89	(1.22)	49.14	(0.65)	49.46	(0.67)	49.19	(0.65)
TiO ₂	0.81	(0.22)	1.00	(0.21)	0.93	(0.17)	0.93	(0.12)
Al ₂ O ₃	4.56	(0.59)	4.57	(0.48)	4.52	(0.53)	4.46	(0.54)
FeO	15.54	(0.83)	16.90	(0.46)	16.47	(0.49)	16.95	(0.42)
MgO	13.50	(0.64)	12.84	(0.45)	13.36	(0.37)	13.09	(0.42)
MnO	0.74	(0.07)	0.82	(0.07)	0.70	(0.08)	0.63	(0.07)
CaO	11.48	(0.38)	10.69	(0.21)	10.82	(0.26)	10.76	(0.25)
Na ₂ O	0.96	(0.15)	1.23	(0.17)	1.14	(0.15)	1.11	(0.16)
K ₂ O	0.40	(0.10)	0.43	(0.08)	0.42	(0.06)	0.44	(0.07)
Cl	0.16	(0.07)	0.20	(0.03)	0.17	(0.04)	0.17	(0.02)
Total	98.06	(0.36)	97.81	(0.22)	98.01	(0.50)	97.72	(0.31)
<i>T-sites</i>								
Si	7.28	(0.12)	7.26	(0.08)	7.27	(0.08)	7.27	(0.08)
Al ^{IV}	0.72	(0.12)	0.74	(0.08)	0.73	(0.08)	0.73	(0.08)
	8.00		8.00		8.00		8.00	
<i>M1-3sites</i>								
Al ^{VI}	0.07	(0.02)	0.05	(0.02)	0.05	(0.02)	0.04	(0.02)
Ti	0.09	(0.02)	0.11	(0.02)	0.10	(0.02)	0.10	(0.01)
Fe ³⁺	0.35	(0.06)	0.29	(0.10)	0.29	(0.09)	0.27	(0.08)
Mg	2.94	(0.12)	2.83	(0.10)	2.93	(0.08)	2.88	(0.09)
Mn	0.09	(0.01)	0.10	(0.01)	0.09	(0.01)	0.08	(0.01)
Fe ²⁺	1.46	(0.07)	1.62	(0.07)	1.55	(0.07)	1.62	(0.06)
Ca	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
	5.00		5.00		5.00		5.00	
<i>M4 site</i>								
Fe	0.09	(0.05)	0.18	(0.05)	0.19	(0.05)	0.20	(0.05)
Ca	1.80	(0.05)	1.69	(0.03)	1.70	(0.04)	1.70	(0.04)
Na	0.12	(0.02)	0.13	(0.05)	0.11	(0.05)	0.09	(0.05)
	2.00		2.00		2.00		2.00	
<i>A site</i>								
Ca	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
Na	0.16	(0.05)	0.22	(0.06)	0.22	(0.05)	0.22	(0.04)
K	0.08	(0.02)	0.08	(0.01)	0.08	(0.01)	0.08	(0.01)
Sum cations	15.23	(0.06)	15.30	(0.07)	15.30	(0.05)	15.31	(0.04)
<i>OH site</i>								
O	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
OH	1.96	(0.02)	1.95	(0.01)	1.96	(0.01)	1.96	(0.01)
F	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)	0.00	(0.00)
Cl	0.04	(0.02)	0.05	(0.01)	0.04	(0.01)	0.04	(0.01)
Mg#	0.59	(0.02)	0.57	(0.02)	0.59	(0.02)	0.58	(0.02)

TABLE B.6: Modal abundances (area%) of mineral phases in GDT, pGT and mGT units

Unit: Sample:	GDT EH16	GDT EH23	pGT EH12	pGT EH76	pGT EH70	mGT EH5
Plagioclase feldspar	43.25	49.27	43.47	41.10	36.57	46.64
Quartz	26.07	21.44	26.52	28.52	31.13	24.12
K-feldspar	14.64	14.29	15.58	19.04	21.93	16.07
Amphibole	5.24	5.89	3.90	3.17	2.59	5.69
Biotite/phlogopite	7.71	6.00	7.59	4.95	5.67	5.28
Chlorite	0.70	1.40	0.75	1.62	0.75	0.65
Fe-oxides	0.56	0.37	0.50	0.29	0.18	0.60
Ilmenite	0.18	0.14	0.13	0.08	0.12	0.22
Apatite	0.28	0.24	0.29	0.23	0.17	0.27
Titanite	0.13	0.25	0.13	0.36	0.12	0.17
Epidote	0.27	0.12	0.26	0.16	0.12	0.02
Zircon	0.02	0.02	0.02	0.01	0.02	0.02
Kaolinite	0.00	0.00	0.00	0.02	0.00	0.12
Others	0.96	0.56	0.86	0.45	0.63	0.12

TABLE B.7: Distribution of anorthite content (area%) in GDT, pGT and mGT created from Ca maps and EPMA analyses

Unit: Sample:	GDT EH16	GDT EH23	pGT EH12	pGT EH76	pGT EH70	mGT EH5
An ₁₀₋₃₀	11.7	14.1	15.3	15.9	15.2	18.0
An ₃₀₋₅₀	21.6	23.2	18.4	16.8	14.1	26.8
An ₅₀₋₇₀	8.4	10.6	6.9	4.9	4.0	1.4
An ₇₀₋₉₀	0.7	0.2	0.4	0.2	0.3	0.0
Total (anorthite maps)	42.5	48.1	41.0	37.8	33.6	46.2
Total (QEMSCAN maps)	43.3	49.3	43.5	41.1	36.6	46.6
<i>Normalised (anorthite maps)</i>						
An ₁₀₋₃₀	27.6	29.3	37.3	42.1	45.2	39.0
An ₃₀₋₅₀	50.8	48.3	44.8	44.6	42.1	58.0
An ₅₀₋₇₀	19.8	22.0	16.8	12.9	11.8	3.0
An ₇₀₋₉₀	1.7	0.4	1.1	0.4	0.9	0.0
<i>Normalised (plagioclase content of mineral maps)</i>						
An ₁₀₋₃₀	27.1	28.6	35.2	38.7	41.5	38.6
An ₃₀₋₅₀	49.9	47.2	42.3	40.9	38.6	57.4
An ₅₀₋₇₀	19.5	21.5	15.9	11.8	10.8	3.0
An ₇₀₋₉₀	1.7	0.4	1.0	0.4	0.8	0.0

TABLE B.8: Abundance of plagioclase population (%) across TK

Population:		P1	P2	P3
mGT	<i>n</i> = 6	100	–	–
pGT	<i>n</i> = 17	24	70	6
GDT	<i>n</i> = 11	–	91	9
GRT	<i>n</i> = 8	–	–	100
fGT	<i>n.a.</i>	<i>n.a.</i>	<i>n.a.</i>	<i>n.a.</i>

TABLE B.9: Temperature evaluation (°C) at 200 MPa for low-Al and high-Al amphibole population

Unit:		mGT		pGT		GDT		GRT	
<i>Holland and Blundy (1994) - equation A</i>									
low-Al	An ₃₀	686	(37)	709	(24)	709	(20)	712	(16)
low-Al	An ₃₅	693	(38)	717	(25)	717	(21)	719	(16)
low-Al	An ₄₀	703	(39)	728	(26)	727	(22)	730	(17)
low-Al	An ₄₅	714	(40)	740	(27)	739	(22)	742	(17)
high-Al	An ₃₀	766	(18)	767	(15)	767	(15)	773	(12)
high-Al	An ₃₅	775	(18)	777	(16)	777	(15)	782	(13)
high-Al	An ₄₀	788	(19)	789	(16)	789	(16)	795	(13)
high-Al	An ₄₅	802	(20)	803	(17)	803	(16)	810	(14)
<i>Holland and Blundy (1994) - equation B</i>									
low-Al	An ₃₀	694	(18)	712	(24)	699	(26)	685	(55)
low-Al	An ₃₅	712	(18)	731	(25)	717	(27)	703	(57)
low-Al	An ₄₀	731	(18)	751	(26)	736	(28)	722	(59)
low-Al	An ₄₅	750	(18)	771	(27)	755	(30)	740	(61)
high-Al	An ₃₀	754	(9)	754	(11)	749	(8)	733	(21)
high-Al	An ₃₅	774	(9)	773	(11)	769	(9)	751	(22)
high-Al	An ₄₀	794	(9)	794	(11)	789	(9)	771	(23)
high-Al	An ₄₅	814	(9)	814	(12)	808	(9)	790	(24)
<i>Blundy and Holland (1990)</i>									
low-Al	An ₃₀	638	(29)	645	(19)	642	(20)	642	(20)
low-Al	An ₃₅	647	(30)	654	(19)	651	(21)	651	(20)
low-Al	An ₄₀	659	(30)	667	(20)	664	(21)	664	(21)
low-Al	An ₄₅	673	(31)	681	(21)	678	(22)	678	(21)
high-Al	An ₃₀	732	(20)	724	(13)	719	(14)	711	(15)
high-Al	An ₃₅	743	(21)	735	(14)	729	(14)	721	(15)
high-Al	An ₄₀	758	(21)	750	(14)	744	(14)	736	(16)
high-Al	An ₄₅	776	(22)	767	(15)	760	(15)	752	(16)
<i>Olten et al. (1984)</i>									
low-Al		652	(30)	679	(28)	669	(22)	669	(17)
high-Al		743	(16)	763	(13)	758	(20)	753	(30)

TABLE B.10: Chemical composition of alkali feldspar from TK

<i>n</i> :	phenocrysts		matrix	
	25		33	
wt%				
SiO ₂	65.47	(0.23)	65.27	(0.40)
Al ₂ O ₃	18.14	(0.19)	18.13	(0.22)
FeO	0.09	(0.02)	0.10	(0.03)
CaO	0.07	(0.04)	0.04	(0.03)
Na ₂ O	1.47	(0.29)	0.12	(0.37)
K ₂ O	14.23	(0.45)	14.80	(0.51)
Total	99.47	(0.39)	99.55	(0.48)
An	0.00	(0.00)	0.00	(0.00)
Ab	0.14	(0.03)	0.11	(0.03)
Or	0.86	(0.03)	0.89	(0.03)
Fe x 10	0.14	(0.03)	0.15	(0.05)

TABLE B.11: Average chemical composition of biotite of TK

Unit:	mGT	mGT	pGT	pGT	pGT	pGT	GDT	GDT	GRT
Sample:	EH5	EH72	EH6	EH11	EH70	EH76	EH23	EH16	EH14
n:	21	18	12	12	6	6	17	19	16
<i>wtf</i> %									
SiO ₂	37.47 (0.28)	37.2 (0.28)	37.77 (0.70)	37.26 (0.26)	37.15 (0.32)	36.67 (0.32)	37.43 (0.35)	37.54 (0.26)	37.83 (0.32)
TiO ₂	4.61 (0.31)	4.58 (0.34)	4.32 (0.44)	4.78 (0.37)	4.73 (0.18)	4.74 (0.11)	4.58 (0.23)	4.62 (0.32)	4.71 (0.45)
Al ₂ O ₃	13.3 (0.16)	13 (0.14)	12.56 (0.41)	12.93 (0.26)	13.2 (0.22)	12.83 (0.21)	12.98 (0.22)	13.33 (0.19)	13.1 (0.23)
FeO	18.69 (0.58)	20.26 (0.42)	21.91 (0.50)	20.93 (0.30)	21.5 (0.53)	21.6 (0.57)	20.18 (0.38)	19.75 (0.40)	19.39 (0.70)
MgO	12.12 (0.34)	11.1 (0.34)	10.18 (0.63)	10.62 (0.28)	10.18 (0.53)	10.21 (0.35)	11.14 (0.30)	11.39 (0.27)	11.57 (0.46)
MnO	0.32 (0.04)	0.36 (0.03)	0.46 (0.03)	0.39 (0.01)	0.34 (0.02)	0.29 (0.05)	0.32 (0.03)	0.3 (0.02)	0.27 (0.02)
CaO	0.02 (0.03)	0.02 (0.03)	0.02 (0.02)	0.01 (0.01)	0.01 (0.01)	0.01 (0.02)	0.01 (0.01)	0.04 (0.05)	0.15 (0.50)
Na ₂ O	0.16 (0.03)	0.14 (0.03)	0.1 (0.04)	0.16 (0.07)	0.14 (0.04)	0.15 (0.02)	0.15 (0.03)	0.15 (0.03)	0.14 (0.02)
K ₂ O	9.14 (0.13)	9.11 (0.12)	9.05 (0.11)	9.01 (0.15)	9.11 (0.12)	9 (0.14)	9.01 (0.12)	9.02 (0.18)	9.14 (0.27)
F	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Cl	0.35 (0.02)	0.35 (0.02)	0.36 (0.03)	0.33 (0.07)	0.34 (0.04)	0.35 (0.06)	0.38 (0.07)	0.37 (0.07)	0.49 (0.08)
Total	96.18 (0.40)	96.12 (0.31)	96.73 (0.37)	96.42 (0.36)	96.7 (0.34)	95.85 (0.32)	96.17 (0.45)	96.51 (0.47)	96.78 (0.45)
Mg #	0.54 (0.01)	0.49 (0.01)	0.45 (0.02)	0.47 (0.01)	0.46 (0.02)	0.46 (0.01)	0.5 (0.01)	0.51 (0.01)	0.52 (0.02)

TABLE B.12: Average chemical composition of magnetite of TK

Unit: Sample: n:	mGT EH5 3	mGT EH72 4	pGT EH6 14	pGT EH8 2	pGT EH11 4	GDT EH23 3	GDT EH16 5	GRT EH14 5
<i>wtf%</i>								
SiO ₂	0.04 (0.01)	0.04 (0.02)	0.09 (0.08)	0.11 (0.10)	0.04 (0.01)	0.05 (0.01)	0.06 (0.02)	0.07 (0.03)
TiO ₂	0.86 (0.39)	0.78 (0.20)	1.22 (1.03)	0.62 (0.33)	1.27 (0.32)	0.59 (0.16)	2.23 (0.95)	0.66 (0.26)
Al ₂ O ₃	0.11 (0.05)	0.21 (0.10)	0.32 (0.41)	0.11 (0.03)	0.09 (0.05)	0.19 (0.09)	0.36 (0.18)	0.14 (0.09)
FeO	92.27 (0.15)	92.65 (0.21)	91.49 (1.31)	92.54 (0.05)	92.09 (0.34)	92.07 (0.33)	90.44 (1.29)	92.31 (0.29)
MgO	0.02 (0.02)	0.02 (0.01)	0.04 (0.07)	0.00 (0.00)	0.01 (0.01)	0.01 (0.02)	0.03 (0.02)	0.01 (0.02)
MnO	0.14 (0.03)	0.12 (0.03)	0.12 (0.03)	0.10 (0.01)	0.15 (0.03)	0.10 (0.03)	0.28 (0.22)	0.09 (0.01)
CaO	0.04 (0.01)	0.06 (0.08)	0.07 (0.05)	0.03 (0.00)	0.07 (0.04)	0.03 (0.01)	0.02 (0.02)	0.05 (0.02)
Cr ₂ O ₃	0.07 (0.03)	0.11 (0.02)	0.11 (0.05)	0.05 (0.00)	0.14 (0.07)	0.13 (0.04)	0.23 (0.09)	0.17 (0.03)
NiO	0.02 (0.03)	0.02 (0.02)	0.01 (0.01)	0.00 (0.00)	0.01 (0.01)	0.01 (0.02)	0.02 (0.03)	0.02 (0.03)
Total	93.57 (0.34)	94.02 (0.23)	93.45 (0.45)	93.55 (0.15)	93.86 (0.46)	93.18 (0.30)	93.67 (0.26)	93.53 (0.39)

TABLE B.13: Average chemical composition of ilmenite of TK

Unit: Sample: n:	mGT EH5 3	mGT EH72 3	pGT EH6 10	pGT EH8 3	pGT EH11 4	GDT EH23 5	GDT EH16 3	GRT EH14 3
<i>wt%</i>								
SiO ₂	0.02 (0.02)	0.03 (0.02)	0.03 (0.01)	0.02 (0.02)	0.02 (0.02)	0.00 (0.01)	0.00 (0.01)	0.05 (0.03)
TiO ₂	50.15 (0.54)	49.35 (0.20)	50.63 (0.51)	49.51 (1.38)	50.42 (0.41)	50.50 (0.37)	49.72 (0.19)	50.18 (0.97)
Al ₂ O ₃	0.00 (0.00)	0.00 (0.00)	0.01 (0.01)	0.00 (0.00)	0.00 (0.00)	0.01 (0.01)	0.00 (0.01)	0.02 (0.03)
FeO	43.17 (0.34)	45.35 (1.21)	44.52 (1.69)	45.51 (0.39)	43.88 (0.45)	42.73 (0.70)	44.74 (0.84)	44.95 (0.12)
MgO	0.03 (0.01)	0.15 (0.02)	0.20 (0.24)	0.04 (0.01)	0.06 (0.02)	0.08 (0.02)	0.11 (0.06)	0.04 (0.01)
MnO	6.62 (0.42)	5.44 (0.83)	4.59 (1.72)	4.52 (0.61)	5.84 (0.42)	6.48 (0.47)	5.07 (1.09)	4.83 (0.28)
CaO	0.06 (0.03)	0.05 (0.05)	0.06 (0.05)	0.14 (0.22)	0.08 (0.06)	0.02 (0.02)	0.04 (0.06)	0.10 (0.07)
Cr ₂ O ₃	0.02 (0.02)	0.05 (0.02)	0.05 (0.05)	0.04 (0.03)	0.04 (0.03)	0.06 (0.02)	0.06 (0.03)	0.05 (0.02)
NiO	0.01 (0.02)	0.00 (0.00)	0.01 (0.01)	0.00 (0.00)	0.01 (0.01)	0.01 (0.02)	0.02 (0.02)	0.01 (0.02)
Total	100.08 (0.20)	100.42 (0.23)	100.07 (0.46)	99.78 (0.62)	100.34 (0.31)	99.90 (0.38)	99.76 (0.08)	100.23 (0.61)

TABLE B.14: Mean temperature and pressure for amphibole of TK

Unit:	mgT		pGT		GDT		GRT						
n:	18		43		55		16						
Temperature (°C; <i>BH90</i>)		min		min		min		min					
		max		max		max		max					
	An30	683 (43)	612	749	700 (33)	632	742	685 (33)	631	741	673 (28)	635	727
	An35	694 (45)	620	764	712 (35)	640	757	698 (36)	639	774	683 (30)	644	741
Pressure (MPa; <i>AS95</i>)		min		min		min		min					
		max		max		max		max					
	An30	134 (85)	5	245	161 (67)	17	271	128 (68)	5	226	99 (59)	19	204
	An35	124 (74)	6	216	147 (59)	18	251	120 (61)	6	224	92 (53)	19	184

TABLE B.15: Mean temperature and pressure for low-Al and high-Al amphibole population

Unit:		mGT		pGT		GDT		GRT	
Temperature (°C; <i>BH90</i>)									
low-Al	An ₃₀	654	(24)	660	(15)	658	(16)	659	(15)
low-Al	An ₃₅	663	(25)	670	(16)	668	(16)	668	(15)
high-Al	An ₃₀	729	(18)	723	(11)	719	(12)	714	(12)
high-Al	An ₃₅	743	(19)	737	(12)	731	(13)	727	(13)
Pressure (MPa; <i>AS95</i>)									
low-Al	An ₃₀	76	(48)	82	(36)	76	(40)	73	(42)
low-Al	An ₃₅	73	(44)	78	(34)	72	(37)	69	(39)
high-Al	An ₃₀	226	(21)	208	(22)	196	(15)	177	(20)
high-Al	An ₃₅	204	(15)	188	(20)	178	(12)	161	(17)

Appendix C

Appendix Chapter 4

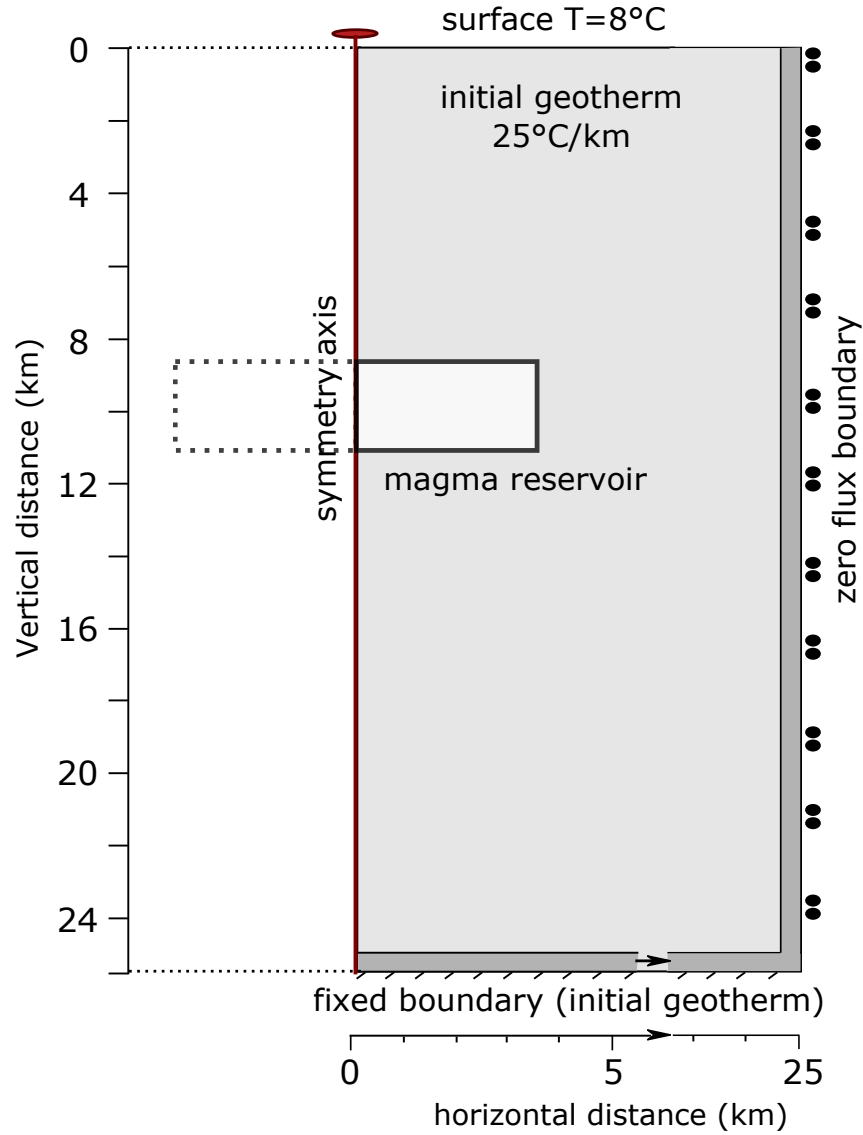


FIGURE C.1: Geometry of the axisymmetric thermal model. Instantaneous intrusion of a cylindrical magma reservoir at 10 km depth is modeled for magma volumes of 100, 1000 and 10.000 km³. The surface temperature is fixed at 8°C and the initial geothermal gradient is set to 25°C km⁻¹. Upper and lower boundaries are fixed temperature, while a zero flux condition is imposed on the right boundary. The left boundary of the model is the plane of symmetry. This figure was provided by Gregor Weber at the University of Geneva.

TABLE C.1: Normalised mineral abundance from QEMSCAN

Sample	Qtz	Kfs	Plg<An30
EH16	26.1	14.6	11.7
EH23	21.4	14.3	14.1
EH12	26.5	15.6	15.3
EH76	28.5	19.0	15.9
EH70	31.1	21.9	15.2

TABLE C.2: Calculated interstitial melt composition (EPMA calibrated)

Sample	Qtz		Or		Ab	
EH16	0.50	(0.01)	0.25	(0.02)	0.25	(0.04)
EH23	0.43	(0.01)	0.27	(0.03)	0.30	(0.05)
EH12	0.46	(0.01)	0.25	(0.03)	0.29	(0.06)
EH76	0.45	(0.01)	0.28	(0.04)	0.27	(0.07)
EH70	0.46	(0.01)	0.29	(0.03)	0.25	(0.07)

TABLE C.3: Parameters used in thermal modelling

Specific heat	1000	$\text{J kg}^{-1} \text{K}^{-1}$
Thermal conductivity	as in Whittington et al. (2009)	
Latent heat	3.13×10^5	J kg^{-1}
Density (melt and solid)	2700	kg m^{-3}
Initial geothermal gradient	25	$^{\circ}\text{C km}^{-1}$
Surface temperature	8	$^{\circ}\text{C}$
Intrusion depth	10	km
Intrusion volumes	100, 1000, 10.000	km^3

TABLE C.4: Polynomial fits

<i>Aspect ratio 10</i>	
H ₂ O-saturated	$y = -3\text{E-}13x^3 + 2\text{E-}07x^2 + 0.0053x + 0.9947$
3 wt.% H ₂ O	$y = -1\text{E-}13x^3 + 1\text{E-}07x^2 + 0.0033x + 0.9967$
2 wt.% H ₂ O	$y = -1\text{E-}13x^3 + 9\text{E-}08x^2 + 0.0031x + 0.9969$
<i>Aspect ratio 2</i>	
H ₂ O-saturated	$y = 2\text{E-}08x^2 + 0.0020x + 0.9980$
3 wt.% H ₂ O	$y = 1\text{E-}08x^2 + 0.0014x + 0.9986$
2 wt.% H ₂ O	$y = 8\text{E-}09x^2 + 0.0012x + 0.9988$

Appendix D

Appendix Chapter 5

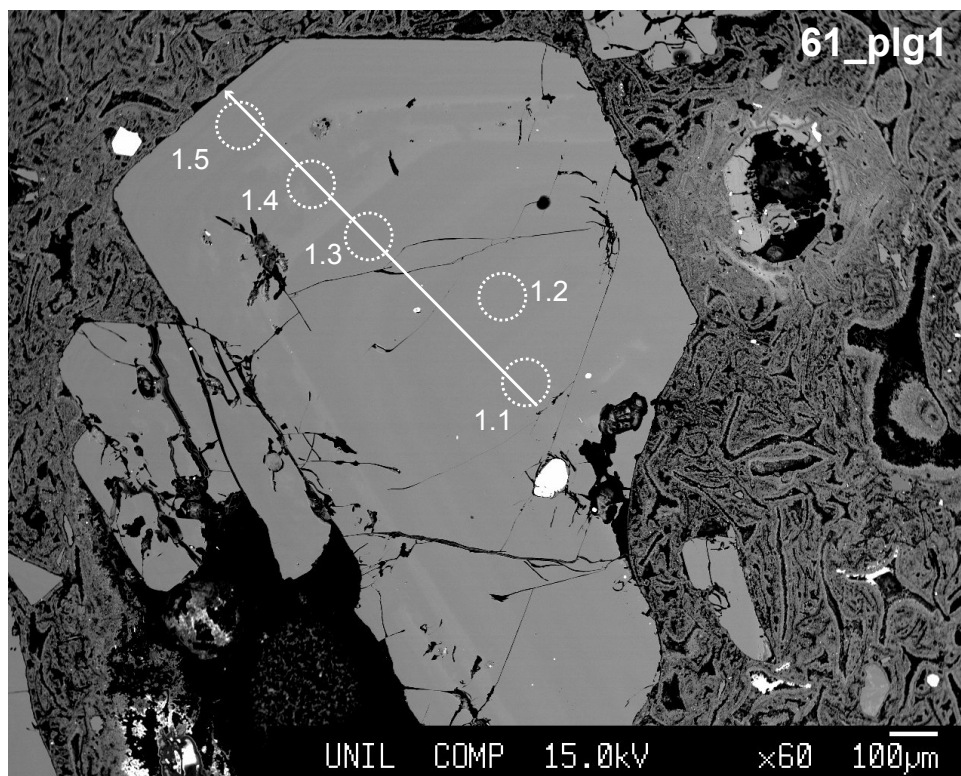


FIGURE D.1: Nyukawa PFD - EH61 - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

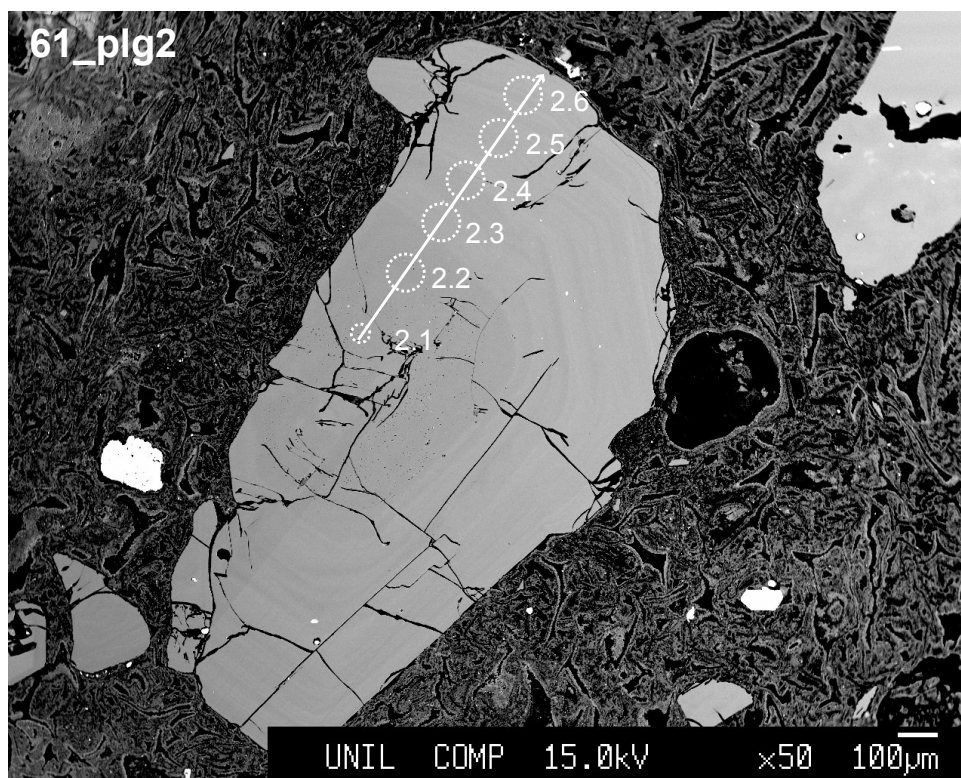


FIGURE D.2: Nyukawa PFD - EH61 - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

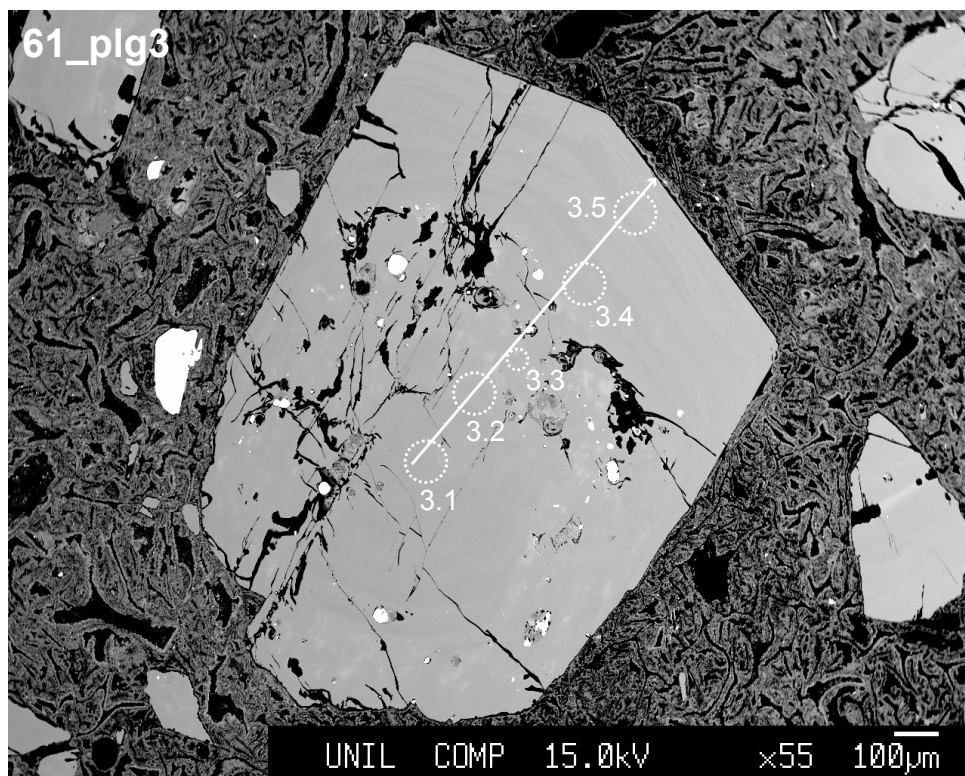


FIGURE D.3: Nyukawa PFD - EH61 - plg3; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

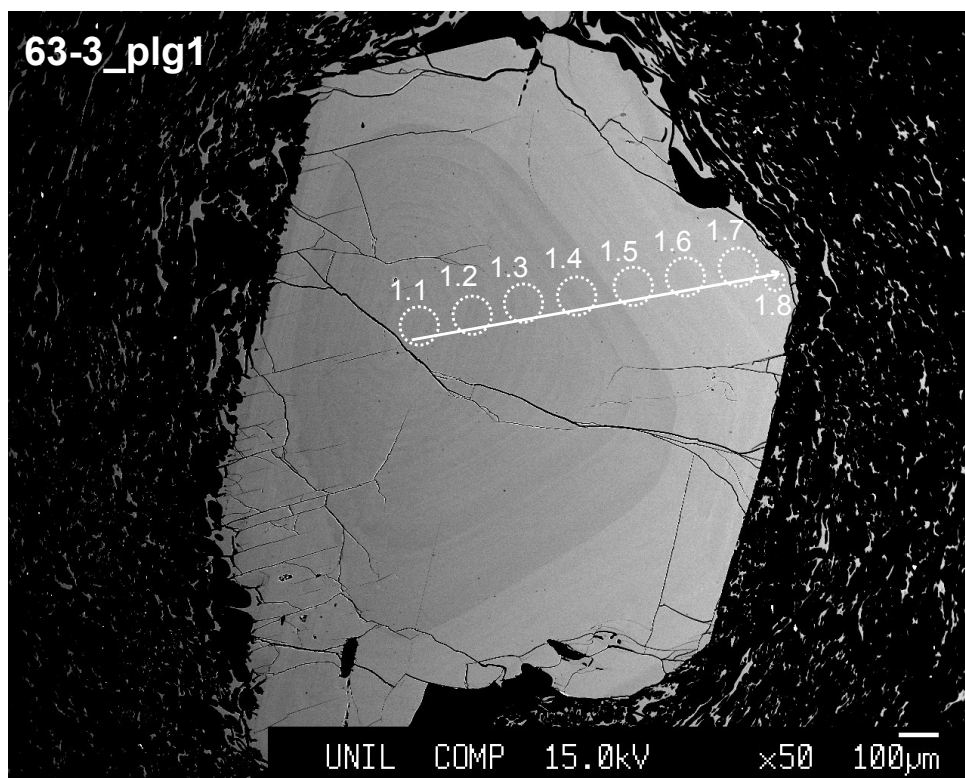


FIGURE D.4: Ebisutoge PD - EH63-3 - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

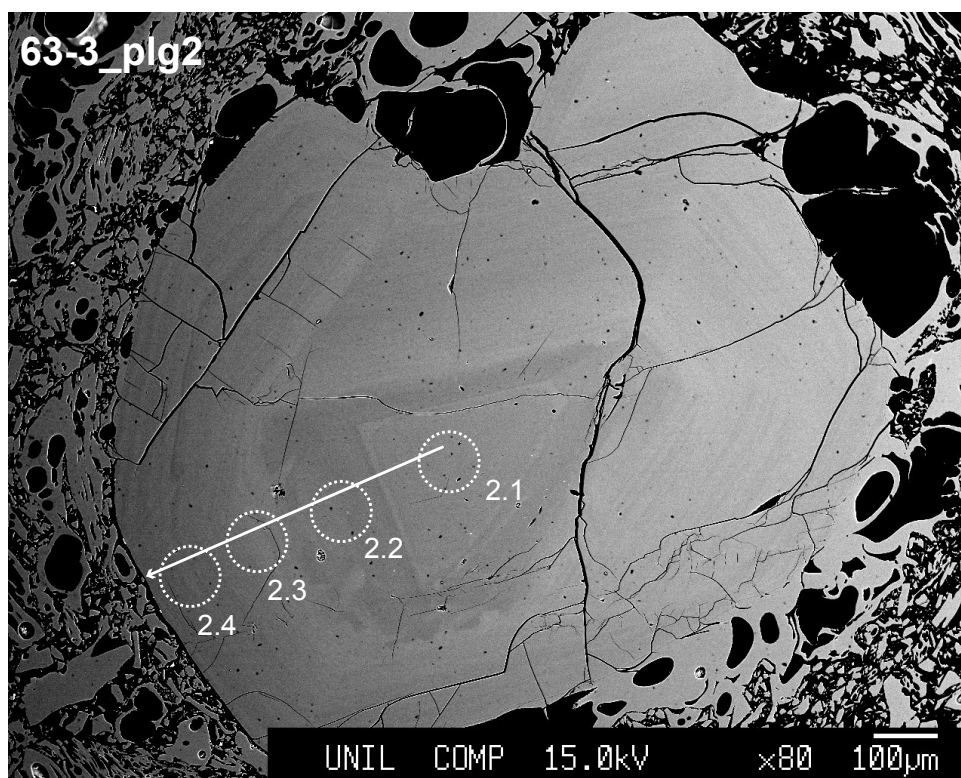


FIGURE D.5: Ebisutoge PD - EH63-3 - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

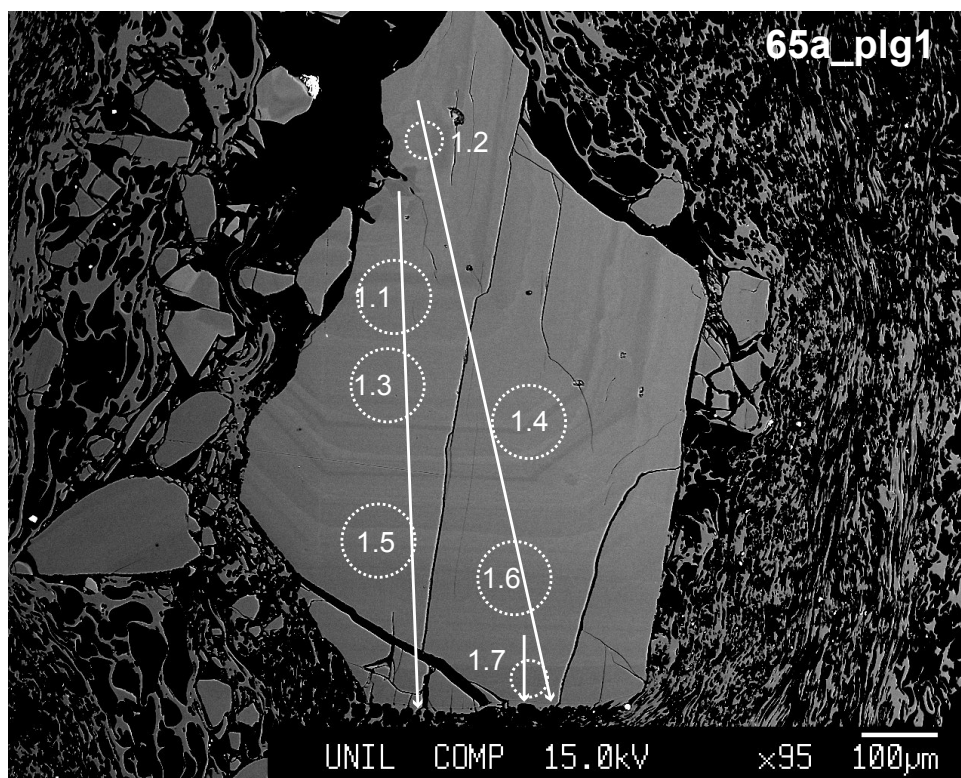


FIGURE D.6: Ebisutoge PD - EH65a - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

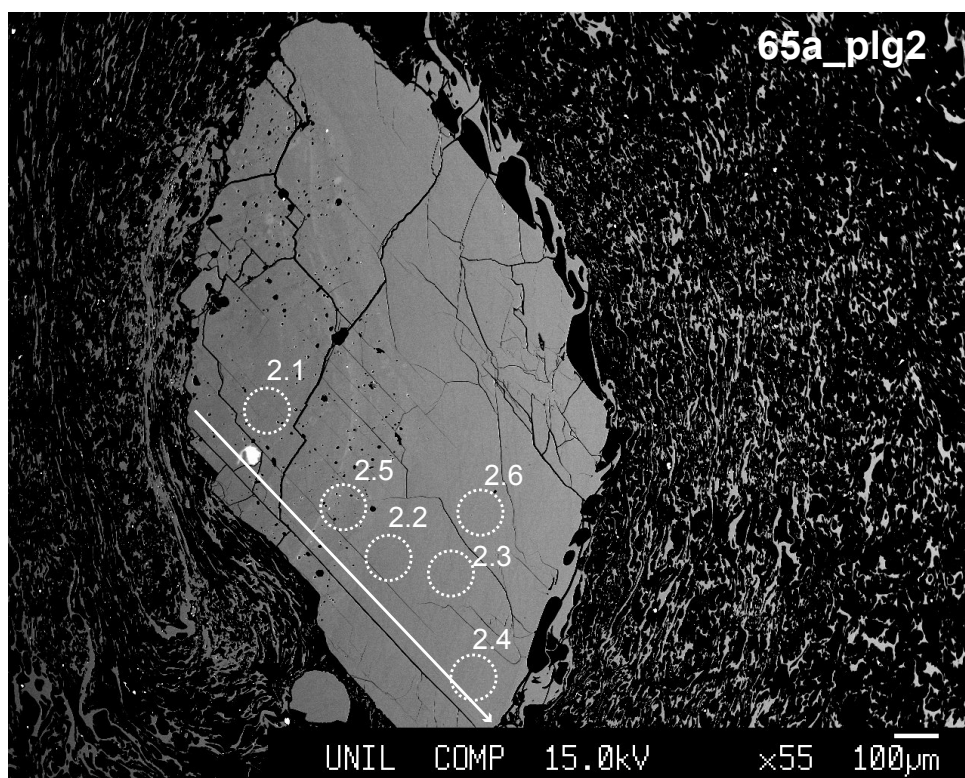


FIGURE D.7: Ebisutoge PD - EH65a - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

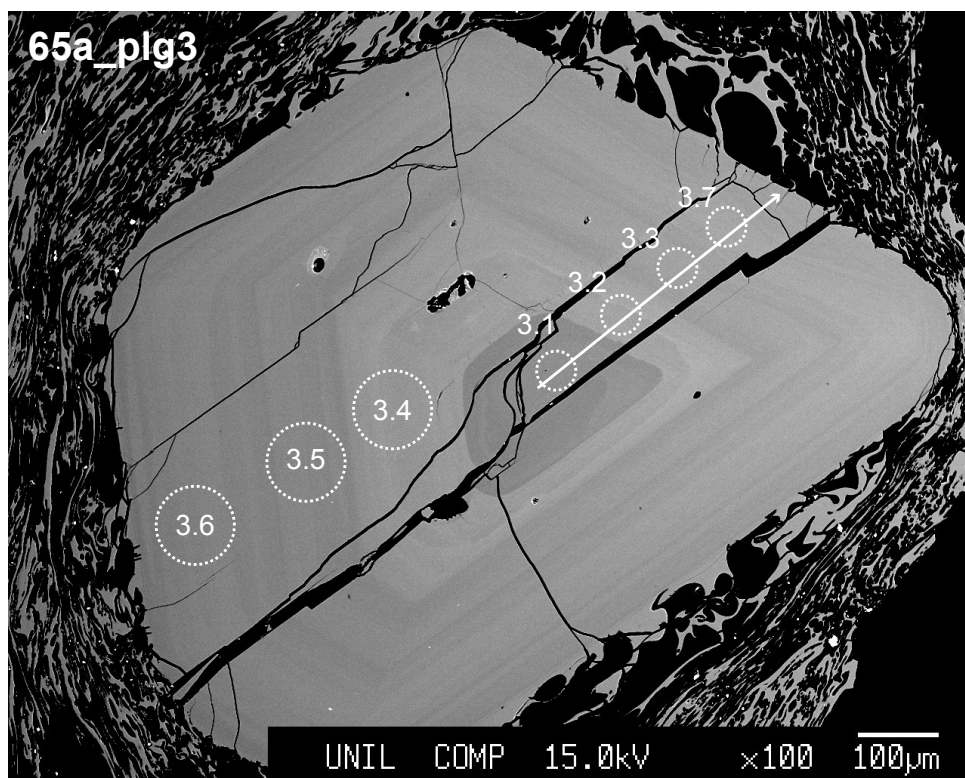


FIGURE D.8: Ebisutoge PD - EH65a - plg3; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

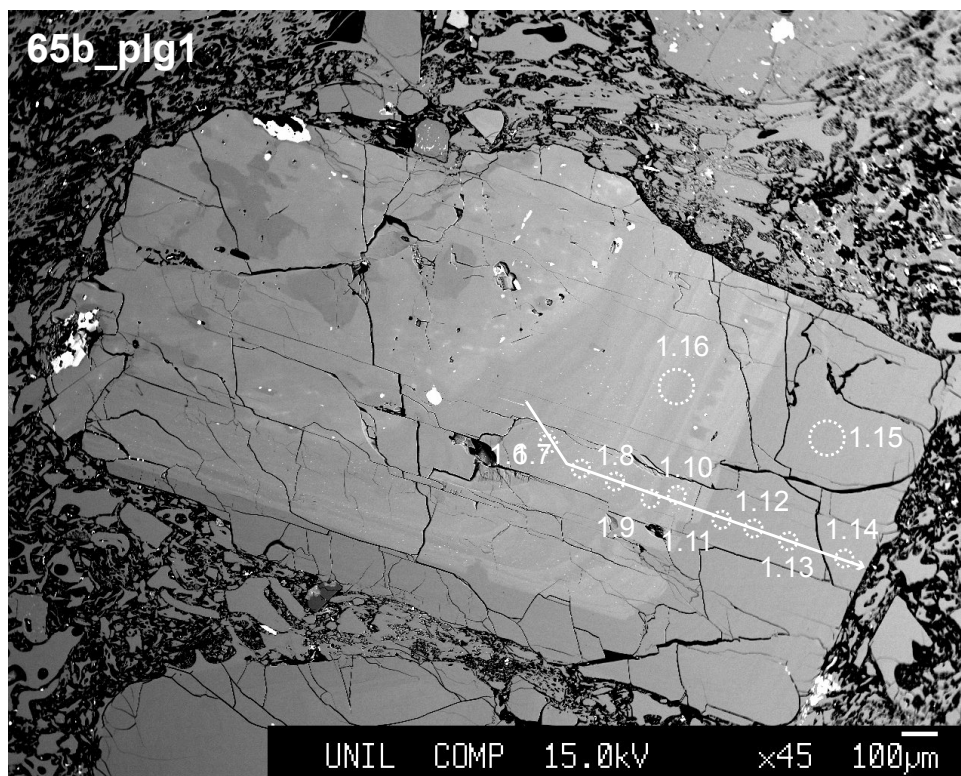


FIGURE D.9: Ebisutoge PD - EH65b - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

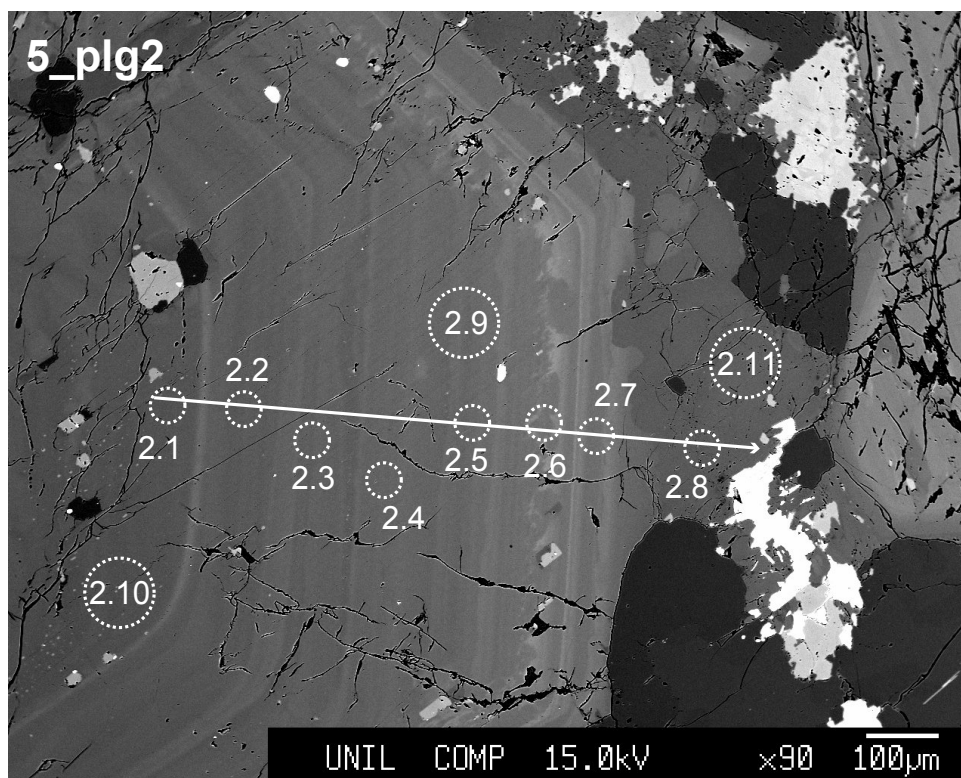


FIGURE D.10: Takidani Pluton - EH5 - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

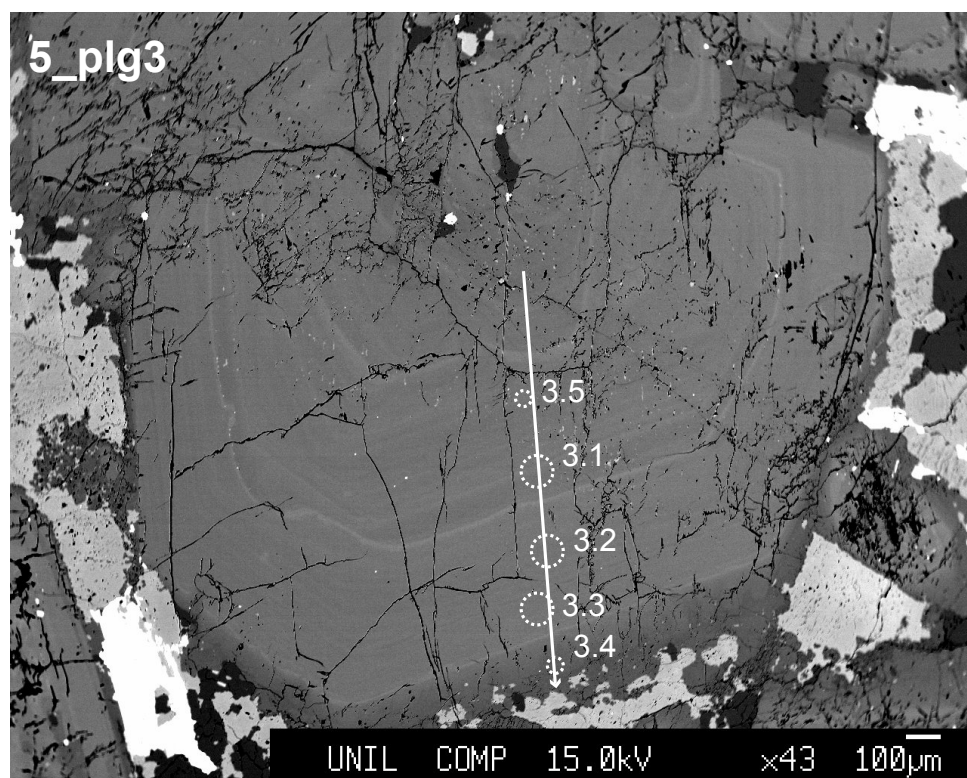


FIGURE D.11: Takidani Pluton - EH5 - plg3; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

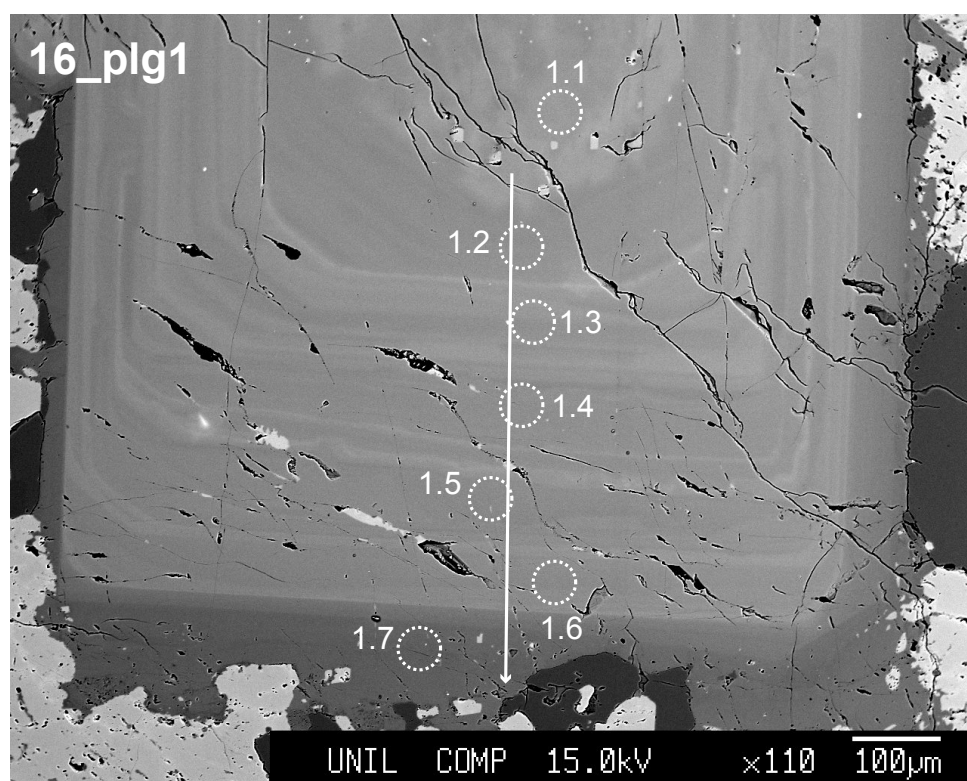


FIGURE D.12: Takidani Pluton - EH16 - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

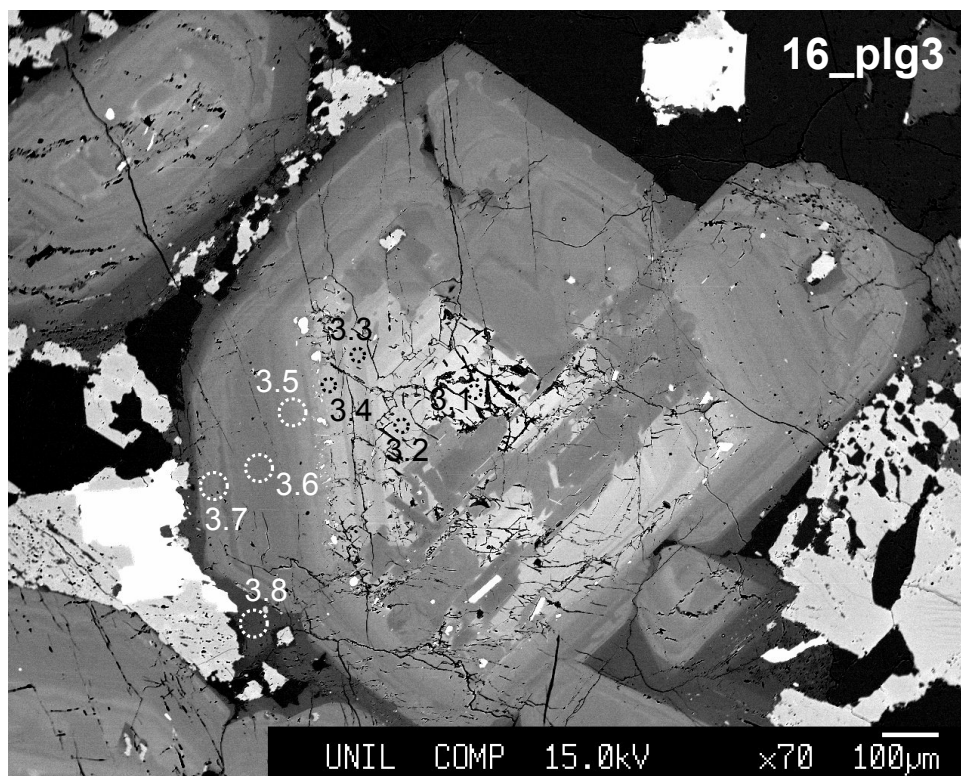


FIGURE D.13: Takidani Pluton - EH16 - plg3; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

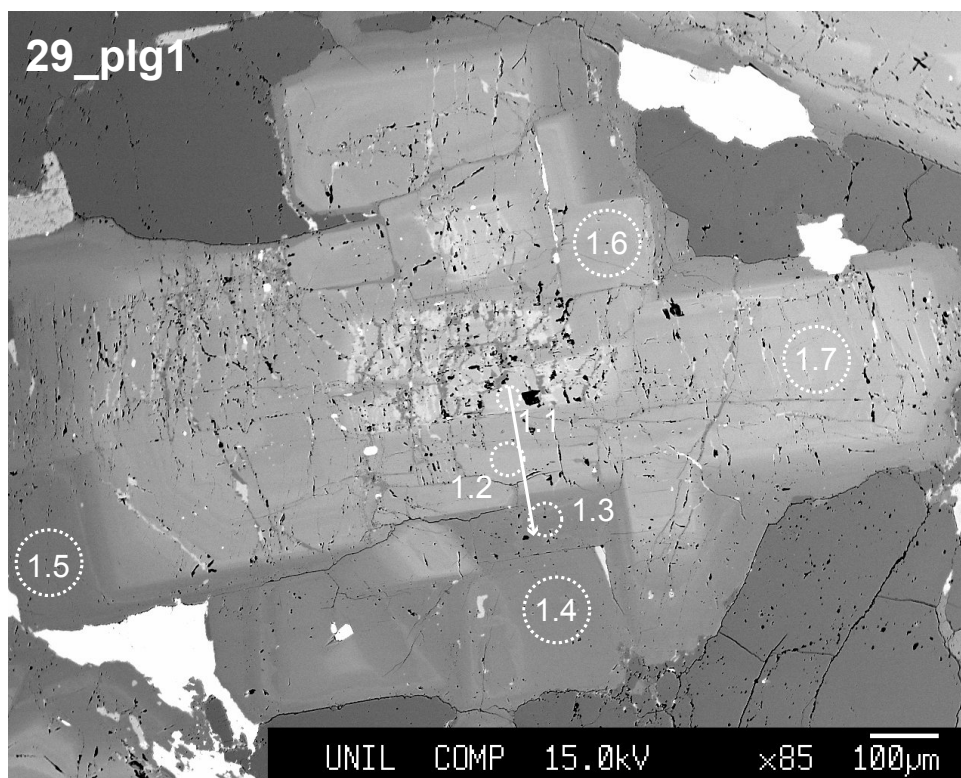


FIGURE D.14: Takidani Pluton - EH29 - plg1; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

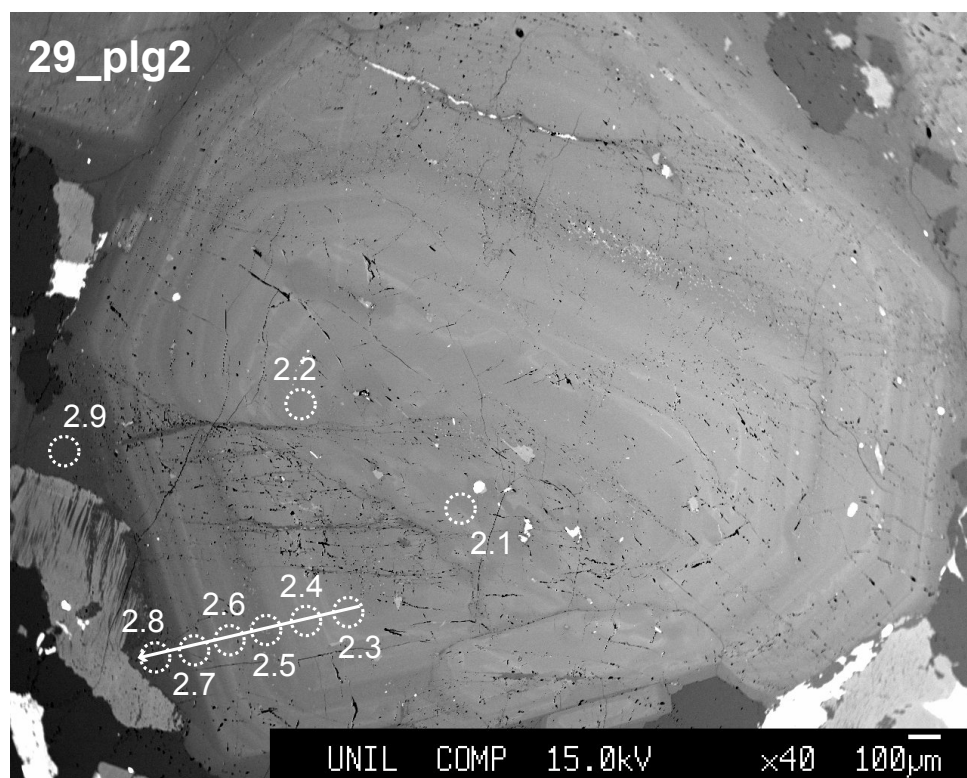


FIGURE D.15: Takidani Pluton - EH29 - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

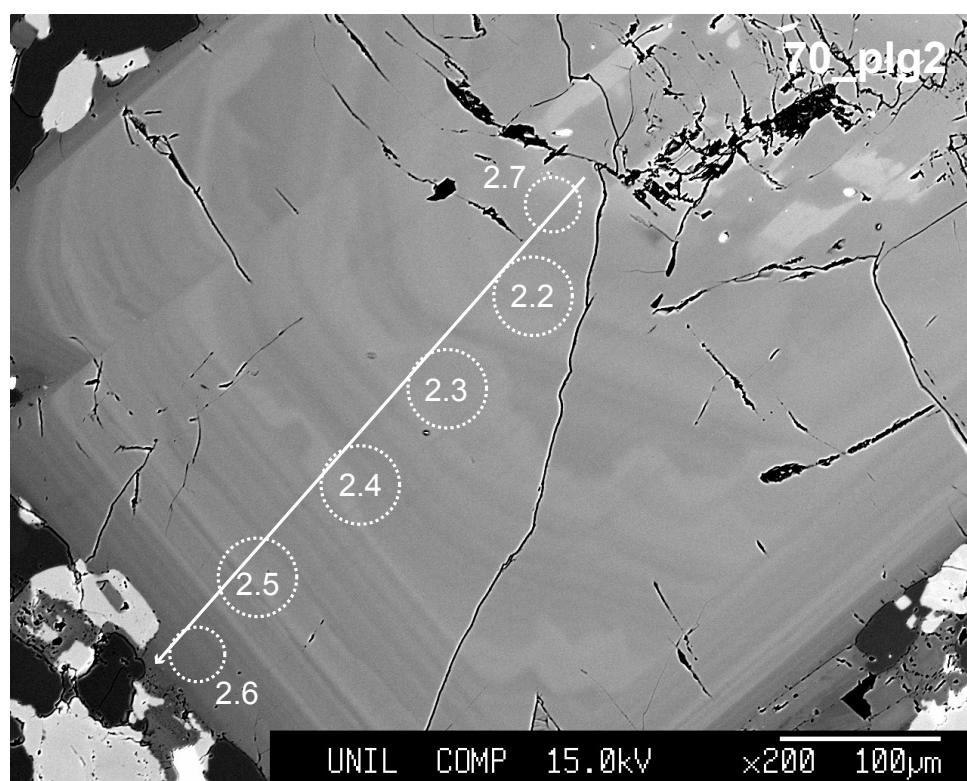


FIGURE D.16: Takidani Pluton - EH70 - plg2; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

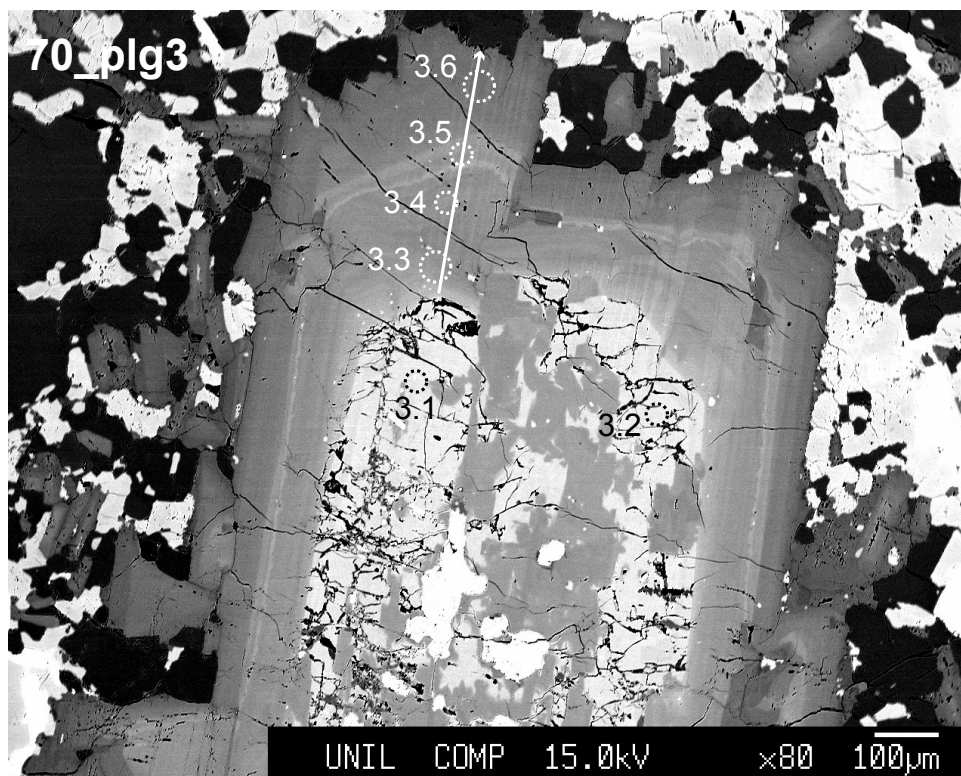


FIGURE D.17: Takidani Pluton - EH70 - plg3; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

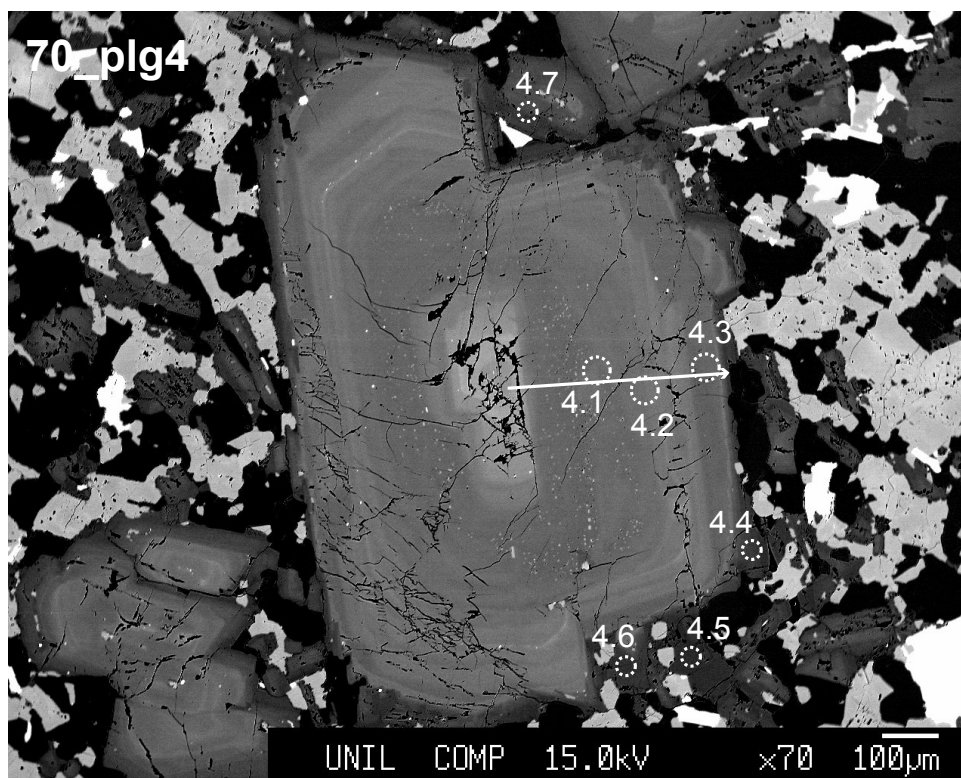


FIGURE D.18: Takidani Pluton - EH70 - plg4; Location of EPMA (solid line) and ICPMS (dashed ring) analyses

TABLE D.1: Average major and trace element composition of the granodiorite, granites, enclaves and aplites of TK

Unit: n:	Granodiorites 40		Granites 6		Enclaves 23		Aplites 8	
<i>wt%</i>								
SiO ₂	67.23	(1.63)	73.69	(1.18)	57.65	(1.94)	76.64	(0.78)
TiO ₂	0.55	(0.08)	0.28	(0.07)	1.00	(0.13)	0.13	(0.06)
Al ₂ O ₃	14.96	(0.39)	13.35	(0.53)	16.84	(0.46)	12.35	(0.25)
Fe ₂ O ₃	4.12	(0.59)	2.07	(0.32)	7.71	(0.79)	0.99	(0.38)
MnO	0.08	(0.02)	0.05	(0.01)	0.18	(0.05)	0.02	(0.01)
MgO	1.64	(0.31)	0.57	(0.21)	3.27	(0.71)	0.22	(0.16)
CaO	3.75	(0.55)	1.60	(0.47)	6.01	(0.98)	0.94	(0.34)
Na ₂ O	3.37	(0.18)	3.24	(0.11)	3.98	(0.47)	3.03	(0.20)
K ₂ O	3.23	(0.31)	4.63	(0.31)	2.15	(0.38)	5.09	(0.25)
P ₂ O ₅	0.15	(0.02)	0.07	(0.02)	0.33	(0.05)	0.03	(0.02)
LOI	0.47	(0.21)	0.39	(0.09)	0.80	(0.35)	0.24	(0.06)
Total	99.57	(0.31)	99.93	(0.09)	99.92	(0.09)	99.69	(0.32)
<i>ppm</i>								
Sc	12.3	(1.6)	8.9	(0.6)	20.7	(3.8)	5.3	(1.7)
Co	8.2	(1.5)	3.0	(0.8)	14.1	(3.9)	1.4	(0.7)
Ni	5.4	(1.3)	4.9	(2.0)	10.3	(9.1)	4.7	(2.2)
Cu	10.2	(8.5)	13.4	(15.3)	22.5	(16.8)	4.6	(1.3)
Zn	55	(24)	71	(7)	116	(13)	36	(18)
Rb	100	(9)	168	(35)	104	(34)	133	(23)
Sr	300	(31)	156	(43)	362	(88)	76	(46)
Y	18	(2)	22	(3)	36	(11)	10	(5)
Zr	143	(13)	144	(17)	168	(24)	83	(12)
Nb	7.8	(0.8)	9.9	(1.5)	12.6	(3.3)	6.8	(3.2)
Mo	1.2	(2.3)	0.7	(0.3)	1.0	(0.8)	0.8	(0.8)
Cs	3.6	(1.0)	5.1	(2.2)	5.5	(4.7)	2.1	(0.8)
Ba	497	(39)	584	(50)	361	(118)	233	(137)
La	27	(4)	31	(3)	28	(8)	26	(8)
Ce	51	(8)	64	(6)	65	(18)	46	(16)
Pr	5.4	(0.7)	6.8	(0.7)	7.9	(2.1)	4.1	(1.5)
Nd	20	(3)	25	(3)	33	(8)	13	(4)
Sm	3.6	(0.5)	4.5	(0.3)	7.0	(1.8)	1.9	(0.7)
Eu	0.9	(0.1)	0.6	(0.1)	1.3	(0.2)	0.3	(0.1)
Gd	3.2	(0.4)	3.8	(0.4)	6.4	(1.7)	1.5	(0.8)
Tb	0.5	(0.1)	0.6	(0.1)	1.0	(0.3)	0.2	(0.1)
Dy	3.1	(0.5)	3.7	(0.5)	6.4	(1.9)	1.4	(0.7)
Ho	0.6	(0.1)	0.7	(0.1)	1.3	(0.4)	0.3	(0.2)
Er	1.9	(0.2)	2.2	(0.3)	3.7	(1.2)	1.0	(0.5)
Tm	0.3	(0.0)	0.3	(0.0)	0.6	(0.2)	0.2	(0.1)
Yb	2.0	(0.2)	2.3	(0.3)	4.0	(1.3)	1.4	(0.8)
Lu	0.3	(0.0)	0.3	(0.0)	0.6	(0.2)	0.2	(0.1)
Hf	4.0	(0.3)	4.3	(0.5)	4.3	(0.6)	3.5	(0.8)
Ta	0.7	(0.1)	1.1	(0.3)	0.9	(0.3)	1.3	(0.5)
W	0.8	(0.3)	0.7	(0.3)	0.9	(0.5)	0.9	(0.5)
Pb	11.7	(2.2)	16.6	(5.0)	8.9	(2.0)	15.0	(1.9)
Th	13.7	(2.7)	14.7	(1.1)	9.3	(3.4)	32.0	(10.3)
U	3.0	(0.6)	4.3	(1.3)	2.3	(0.8)	6.7	(2.4)

TABLE D.2: Average major and trace element composition of Hotaka Andesite, Nyukawa PFD and Ebisutoge PD

Unit: Analysis: n:	Hotaka Andesite bulk 4		Nyukawa PFD bulk 2		Chayano-Ebisutoge PD bulk glass 7 9			
<i>wt%</i>								
SiO ₂	63.62	(0.97)	66.69	(1.17)	70.57	(1.77)	76.71	(0.14)
TiO ₂	0.71	(0.04)	0.70	(0.06)	0.16	(0.02)	0.09	(0.01)
Al ₂ O ₃	16.29	(0.47)	16.11	(0.68)	14.56	(1.14)	12.70	(0.06)
Fe ₂ O ₃	5.28	(0.19)	4.81	(0.08)	2.25	(0.24)		
FeO							1.31	(0.17)
MnO	0.10	(0.02)	0.09	(0.00)	0.07	(0.01)	0.05	(0.01)
MgO	2.10	(0.10)	0.85	(0.06)	0.12	(0.04)	0.05	(0.01)
CaO	5.05	(0.33)	2.86	(0.46)	1.27	(0.16)	0.90	(0.05)
Na ₂ O	3.57	(0.06)	2.83	(0.10)	2.95	(0.56)	3.61	(0.25)
K ₂ O	2.27	(0.40)	3.26	(0.32)	4.46	(0.52)	4.55	(0.34)
P ₂ O ₅	0.20	(0.01)	0.17	(0.01)	0.01	(0.00)	0.03	(0.00)
Total	99.88	(0.13)	99.80	(0.19)	99.46	(0.22)	95.57	(0.49)
<i>ppm</i>								
Sc	14.9	(0.8)	13.33	(0.8)	8.6	(0.4)		
Co	8.9	(1.4)	8.72	(0.3)	0.9	(0.2)		
Ni	3.9	(0.5)	3.61	(0.6)	2.0	(0.2)		
Cu	19.4	(10.8)	8.02	(1.1)	5.9	(0.9)		
Zn	96.8	(14.5)	102.87	(4.4)	95.4	(7.4)		
Rb	72.1	(20.2)	86.24	(37.8)	121.5	(7.7)		
Sr	382.1	(26.5)	259.33	(37.9)	149.7	(22.0)		
Y	17.5	(0.9)	22.26	(1.1)	30.9	(6.5)		
Zr	166.9	(5.3)	188.06	(4.6)	186.0	(17.1)		
Nb	7.8	(0.2)	9.48	(0.4)	10.6	(1.1)		
Mo	0.6	(0.2)	1.46	(0.5)	2.2	(0.2)		
Cs	2.5	(1.0)	1.84	(1.7)	5.5	(0.6)		
Ba	494.1	(54.4)	709.42	(27.3)	727.6	(24.7)		
La	24.7	(1.2)	32.70	(2.0)	42.9	(14.1)		
Ce	48.5	(2.1)	58.24	(6.2)	72.4	(5.2)		
Pr	5.3	(0.2)	6.92	(0.3)	9.2	(3.5)		
Nd	19.8	(0.8)	25.66	(1.2)	33.5	(11.9)		
Sm	3.7	(0.2)	4.65	(0.2)	6.1	(2.0)		
Eu	1.2	(0.1)	1.11	(0.2)	1.2	(0.3)		
Gd	3.2	(0.2)	4.15	(0.2)	5.3	(1.4)		
Tb	0.5	(0.0)	0.59	(0.0)	0.8	(0.2)		
Dy	3.1	(0.2)	3.80	(0.1)	5.3	(1.2)		
Ho	0.6	(0.0)	0.76	(0.0)	1.1	(0.2)		
Er	1.8	(0.1)	2.19	(0.0)	3.2	(0.6)		
Tm	0.3	(0.0)	0.33	(0.0)	0.5	(0.1)		
Yb	2.0	(0.1)	2.41	(0.0)	3.5	(0.6)		
Lu	0.3	(0.0)	0.35	(0.0)	0.5	(0.1)		
Hf	4.2	(0.2)	4.86	(0.1)	5.3	(0.4)		
Ta	0.6	(0.0)	0.74	(0.0)	0.8	(0.1)		
W	0.8	(0.2)	0.88	(0.3)	1.4	(0.1)		
Pb	21.3	(23.3)	16.33	(2.5)	17.9	(1.7)		
Th	7.9	(0.8)	11.66	(0.2)	13.8	(1.8)		
U	1.7	(0.2)	2.38	(0.2)	2.9	(0.2)		

TABLE D.3: Major and trace element concentration of the Hotaka Andesite, Nyukawa PFD and Ebisutoge PD

Unit:	Hotaka Andesite				Nyukawa PFD				Ebisutoge PD			
Subunit:	Wm	Wm	Wm	Wm	Wm	Wm	Wm	Wm	B	B	C	D
Sample:	EH4H	EH52	EH53	EH54	EH61	EH62	EH63-1	EH63-2	EH63-3	EH65A	EH65B	EH65C
Latitude:	36.2937	36.2790	36.2785	36.2743	36.1556	36.1614	36.1971	36.1971	36.1971	36.1971	36.1971	36.1971
Longitude:	137.6319	137.6290	137.6286	137.6264	137.2942	137.3231	137.3359	137.3359	137.3359	137.3359	137.3359	137.3359
wt%												
SiO ₂	62.70	64.86	63.91	63.03	65.86	67.51	71.48	70.51	70.68	66.69	71.46	71.74
TiO ₂	0.72	0.69	0.68	0.76	0.74	0.66	0.13	0.14	0.14	0.17	0.17	0.17
Al ₂ O ₃	16.68	15.62	16.31	16.57	16.59	15.63	14.01	14.45	14.36	17.10	13.95	13.82
Fe ₂ O ₃	5.36	5.11	5.15	5.51	4.87	4.75	2.02	2.18	2.09	2.77	2.24	2.19
MnO	0.11	0.10	0.07	0.11	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	2.21	1.97	2.07	2.15	0.09	0.09	0.07	0.06	0.06	0.06	0.09	0.05
CaO	5.27	4.57	5.11	5.25	0.81	0.89	0.08	0.09	0.09	0.10	0.17	0.14
Na ₂ O	3.58	3.52	3.66	3.54	3.18	2.53	1.16	1.16	1.12	1.13	1.45	1.45
K ₂ O	2.33	2.76	1.79	2.19	2.76	2.40	2.65	2.40	2.43	2.53	3.44	3.58
P ₂ O ₅	0.21	0.19	0.20	0.21	3.03	3.48	4.99	5.00	5.02	3.93	4.21	4.11
LOI	0.55	0.57	0.89	0.68	0.18	0.16	0.01	0.01	0.01	0.01	0.02	0.02
Total	99.71	99.96	99.85	100.00	99.66	99.93	99.74	99.45	99.40	99.06	99.59	99.34
ppm												
Sc	15.9	14.4	14.1	15.3	13.9	12.8	8.6	8.6	8.5	9.5	8.5	8.4
Co	10.3	8.8	7.0	9.3	8.9	8.5	0.9	0.7	0.7	0.8	1.3	1.0
Ni	4.5	3.5	4.2	3.6	3.2	4.1	2.3	2.1	1.9	1.9	2.1	1.9
Cu	19.1	9.5	14.6	34.4	8.8	7.3	7.0	6.4	5.5	4.9	4.7	6.5
Zn	106	94	78	110	106	100	91	106	100	103	92	86
Rb	95	78	47	68	60	113	127	130	128	107	121	121
Y	17	17	17	17	21	23	26	26	26	26	29	41
Zr	165	174	161	167	185	191	163	189	181	220	186	180
Nb	7.6	8.0	7.7	7.9	9.7	9.2	10.5	10.8	10.5	12.9	9.7	9.8
Mo	0.6	0.6	0.4	0.9	1.1	1.8	2.6	2.4	2.3	2.2	2.2	2.0
Cs	3.9	2.6	1.5	2.0	0.6	3.1	5.0	5.5	5.6	6.8	4.9	5.3
Ba	532	541	422	482	729	690	755	736	736	716	754	708
La	24	26	25	24	36	31	36	37	35	34	42	74
Ce	46	51	50	47	63	54	74	75	69	65	68	79
Pr	5.1	5.5	5.4	5.1	7.1	6.7	7.6	8.0	7.4	7.5	8.9	17.1
Nd	19	21	20	19	27	25	27	29	28	28	32	60
Sm	3.5	4.0	3.6	3.6	4.8	4.5	5.3	5.5	5.2	5.2	5.8	10.7
Eu	1.2	1.1	1.2	1.2	1.2	1.0	1.0	1.1	0.9	1.0	1.2	1.2
Gd	3.2	3.5	3.1	3.2	4.3	4.0	4.5	4.8	4.5	4.4	4.9	8.5
Tb	0.5	0.5	0.5	0.5	0.6	0.6	0.7	0.7	0.7	0.7	0.8	1.3
Dy	3.0	3.4	3.0	3.1	3.9	3.7	4.5	4.9	4.5	4.5	5.1	7.9
Ho	0.6	0.7	0.6	0.6	0.8	0.8	1.0	1.0	0.9	1.0	1.0	1.5
Er	1.8	1.8	1.8	1.8	2.2	2.2	2.8	3.0	2.7	2.7	2.9	4.3
Tm	0.3	0.3	0.3	0.3	0.3	0.3	0.4	0.5	0.4	0.4	0.5	0.7
Yb	1.9	2.2	1.9	2.0	2.4	2.4	3.2	3.3	3.1	3.1	3.4	4.7
Lu	0.3	0.3	0.3	0.3	0.4	0.4	0.5	0.5	0.5	0.5	0.5	0.7
Hf	4.1	4.5	4.2	4.2	4.8	5.0	5.0	5.4	5.2	6.2	5.3	5.2
Ta	0.5	0.6	0.6	0.6	0.8	0.7	0.8	0.9	0.8	1.1	0.8	0.8
W	0.5	0.9	0.6	1.1	0.7	1.1	1.6	1.5	1.5	1.3	1.4	1.5
Pb	9.0	8.8	56.3	11.2	18.1	14.6	20.7	na	na	18.3	17.7	16.6
Th	7.2	8.9	8.0	7.4	11.5	11.8	14.0	14.4	13.9	17.3	12.1	12.3
U	1.5	1.9	1.6	1.6	2.2	2.5	3.0	3.0	3.0	3.1	2.7	3.0

TABLE D.5: (Continued)

[illegible]

TABLE D.6: (Continued)

Unit:	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave	Enclave
Sample:	EH66	EH80A	EH80B	EH80C	EH80E	EH80F	EH80G	EH80H	EH80I	EH80J	EH80K	EH80L	EH80M	EH80N	EH80P	EH80R			
Latitude:	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
Longitude:	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na	na
<i>wt%</i>																			
SiO ₂	61.61	57.21	56.83	55.95	57.82	56.63	57.86	56.38	57.02	54.30	56.37	57.90	56.30	53.49	59.34	58.66			
SiO ₂	0.72	0.90	1.07	0.98	1.05	1.17	1.04	0.98	0.94	1.11	1.21	1.00	0.92	1.36	0.89	0.92			
Al ₂ O ₃	16.48	16.94	17.09	16.86	16.74	17.29	16.55	17.11	15.60	17.01	16.76	16.99	16.11	18.17	17.09	16.79			
Fe ₂ O ₃	6.21	0.11	8.21	8.34	8.04	8.50	7.43	7.76	9.28	8.92	8.18	7.59	8.14	8.79	6.89	6.84			
MnO	0.11	0.19	0.21	0.22	0.16	0.19	0.23	0.16	0.22	0.24	0.17	0.17	0.25	0.18	0.15	0.21			
MgO	2.28	3.38	3.21	4.00	3.00	3.09	3.62	3.99	4.53	4.25	3.65	3.16	4.48	3.53	2.43	3.23			
CaO	5.58	6.36	5.18	6.32	5.74	6.16	6.48	8.19	4.98	6.80	7.76	5.62	7.11	7.67	5.21	6.20			
Na ₂ O	3.59	4.14	4.28	4.23	4.30	4.20	4.16	2.86	3.91	4.12	4.32	3.91	3.71	3.24	4.50	3.94			
K ₂ O	2.32	1.94	2.98	2.04	1.66	1.84	2.18	1.60	2.36	2.09	1.63	2.02	1.87	2.15	2.42	2.42			
P ₂ O ₅	0.42	0.27	0.33	0.28	0.32	0.36	0.29	0.27	0.26	0.31	0.31	0.41	0.30	0.38	0.39	0.31			
LOI	0.65	0.93	0.54	0.62	1.02	0.53	0.05	0.58	0.92	0.83	0.70	0.92	0.57	0.81	0.57	0.51			
Total	99.96	99.78	99.94	99.85	99.84	99.96	99.89	99.90	100.06	99.98	99.83	100.10	99.79	99.77	99.88	100.04			
<i>ppm</i>																			
Sc	10.8	20.8	21.7	24.7	22.8	22.0	23.9	25.1	23.1	25.1	22.8	17.1	24.3	23.2	16.3	19.3			
Co	10.5	14.6	14.5	17.8	15.3	14.3	14.8	18.9	12.8	19.2	19.1	12.3	18.0	20.9	11.1	14.1			
Ni	3.2	8.4	7.3	22.6	8.1	4.1	5.9	14.0	32.4	20.3	6.5	8.1	35.7	3.9	4.4	14.5			
Cu	23.4	14.2	9.2	9.9	60.2	52.5	7.7	22.4	36.3	8.3	45.1	18.4	9.5	39.2	5.4	16.4			
Zn	85	112	128	120	116	122	124	110	144	114	126	109	124	128	105	116			
Rb	74	85	127	107	100	94	79	85	230	96	69	108	85	126	111	103			
Sr	22	34	303	301	336	382	296	428	208	306	472	320	334	472	320	397			
Y	46	34	35	53	28	47	32	22	46	46	21	36	45	23	29	31			
Zr	162	152	182	187	185	155	160	166	135	152	165	202	183	120	192	160			
Nb	7.5	10.7	12.8	18.0	11.6	16.4	11.0	8.0	15.9	16.2	8.2	14.5	13.7	9.0	11.8	10.7			
Mo	0.9	0.7	3.5	0.4	0.7	0.7	0.7	0.4	2.0	1.5	0.5	0.4	0.4	1.6	0.5	0.4			
Cs	3.8	4.5	6.3	6.3	6.8	4.6	2.3	4.6	25.2	4.8	2.7	5.3	4.1	5.8	5.9	3.3			
Ba	366	376	621	414	274	260	367	189	115	441	316	438	301	271	410	380			
La	26	27	42	43	27	16	30	23	32	26	21	22	37	23	22	26			
Ce	54	62	90	102	57	43	71	48	66	70	46	55	89	49	53	60			
Pr	6.2	7.5	10.2	12.4	6.5	6.1	8.3	5.6	7.5	9.2	5.4	7.0	11.1	5.9	6.7	7.1			
Nd	25	31	40	50	26	29	33	23	28	40	23	31	45	25	27	29			
Sm	4.8	6.2	7.5	10.2	5.3	7.6	6.7	4.7	6.4	9.1	4.8	6.7	9.2	5.2	5.9	6.1			
Eu	1.3	1.3	1.3	1.4	1.2	1.4	1.5	1.3	0.7	1.5	1.3	1.2	1.5	1.4	1.2	1.2			
Gd	4.1	5.8	6.7	9.3	5.1	7.5	5.8	4.3	6.5	8.2	4.3	6.4	8.2	4.8	5.4	5.4			
Tb	0.6	0.9	1.0	1.4	0.7	1.2	0.9	0.6	1.1	1.2	0.6	1.0	1.2	0.7	0.8	0.8			
Dy	0.8	5.8	6.3	9.1	5.0	8.2	5.6	4.1	7.7	8.2	3.9	6.3	8.0	4.4	5.2	5.3			
Ho	3.8	1.2	1.2	1.8	1.0	1.7	1.1	0.8	1.6	1.6	0.7	1.3	1.6	0.8	1.0	1.1			
Er	2.2	3.5	3.6	5.4	2.9	5.0	3.3	2.1	4.9	4.8	2.1	3.7	4.6	2.4	2.9	3.2			
Tm	0.3	0.5	0.5	0.8	0.4	0.8	0.5	0.3	0.8	0.7	0.3	0.6	0.7	0.3	0.4	0.5			
Yb	2.4	3.8	3.9	6.0	3.0	5.5	3.7	2.3	6.1	5.3	2.2	4.0	4.8	2.4	3.3	3.4			
Lu	0.4	0.6	0.6	0.9	0.5	0.8	0.5	0.4	0.9	0.8	0.3	0.6	0.7	0.4	0.5	0.5			
Hf	3.9	4.2	4.8	5.2	4.8	4.1	4.2	4.0	3.4	4.2	4.1	4.8	4.7	3.2	5.0	4.2			
Ta	0.5	0.9	0.9	1.4	0.9	1.2	0.9	0.5	0.8	1.1	0.5	0.9	1.0	0.5	0.9	0.8			
W	0.6	0.9	1.1	0.5	1.0	0.7	0.9	0.8	2.6	1.3	0.7	0.8	0.8	1.5	0.7	0.5			
Pb	12.5	11.7	8.4	6.6	10.8	6.1	9.2	6.1	10.1	6.7	7.5	8.3	7.1	6.4	8.2	9.2			
Th	7.5	12.0	13.4	17.4	11.2	7.7	10.0	4.8	5.8	7.5	5.6	9.4	10.6	4.6	13.9	10.1			
U	1.7	2.9	3.0	4.2	2.8	2.3	2.0	1.1	2.4	2.0	1.2	2.6	2.4	1.0	3.1	2.5			

TABLE D.7: Average glass composition of the Ebisutoge PD

Unit: Sample:	Unit B EH63-3 pumice 26	Unit B EH63-4 pumice 23	Unit C EH65a pumice 50	Unit D EH65b fiamme 38	Unit D EH65b matrix 39	Unit D EH65c fiamme 40	Unit D EH65c matrix 39	Unit D EH65d fiamme 39	Unit D EH65d matrix 38
n:									
wtf%									
SiO ₂	76.65 (0.30)	76.55 (0.29)	76.55 (0.38)	76.79 (0.31)	76.87 (0.34)	76.93 (0.40)	76.79 (0.35)	76.66 (0.36)	76.60 (0.44)
TiO ₂	0.07 (0.04)	0.10 (0.05)	0.09 (0.05)	0.09 (0.04)	0.11 (0.05)	0.09 (0.04)	0.10 (0.04)	0.09 (0.04)	0.10 (0.04)
Al ₂ O ₃	12.67 (0.20)	12.67 (0.20)	12.70 (0.20)	12.72 (0.18)	12.66 (0.20)	12.72 (0.20)	12.65 (0.19)	12.85 (0.26)	12.70 (0.24)
FeO	1.47 (0.07)	1.47 (0.10)	1.45 (0.13)	1.36 (0.16)	1.33 (0.25)	1.01 (0.30)	1.31 (0.25)	1.04 (0.33)	1.30 (0.27)
MnO	0.05 (0.04)	0.06 (0.03)	0.05 (0.03)	0.04 (0.03)	0.06 (0.04)	0.05 (0.03)	0.07 (0.03)	0.05 (0.04)	0.05 (0.04)
MgO	0.06 (0.02)	0.05 (0.02)	0.05 (0.02)	0.04 (0.02)	0.05 (0.03)	0.03 (0.03)	0.05 (0.03)	0.04 (0.02)	0.06 (0.02)
CaO	0.91 (0.05)	0.92 (0.07)	0.95 (0.06)	0.94 (0.06)	0.91 (0.06)	0.91 (0.06)	0.83 (0.06)	0.91 (0.07)	0.82 (0.06)
Na ₂ O	3.90 (0.17)	3.92 (0.15)	3.86 (0.19)	3.76 (0.16)	3.56 (0.16)	3.36 (0.19)	3.38 (0.14)	3.36 (0.27)	3.37 (0.17)
K ₂ O	4.20 (0.12)	4.23 (0.10)	4.27 (0.16)	4.22 (0.13)	4.44 (0.15)	4.87 (0.19)	4.80 (0.16)	4.96 (0.22)	4.96 (0.20)
P ₂ O ₅	0.02 (0.04)	0.02 (0.04)	0.03 (0.03)	0.03 (0.04)	0.02 (0.03)	0.03 (0.04)	0.03 (0.04)	0.02 (0.03)	0.02 (0.04)
Total	95.06 (0.52)	94.96 (0.52)	94.81 (0.43)	95.88 (0.54)	96.04 (0.46)	95.93 (0.58)	95.90 (0.70)	95.58 (0.79)	95.99 (0.53)

TABLE D.8: Glass composition (*wt.%*) of Ebisutoge PD Unit B, EH63-3 (pumice)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_8	gl_9	gl_10	gl_11
SiO ₂	76.46	77.14	76.55	76.50	76.86	76.67	76.16	76.58	77.10	76.68
TiO ₂	0.08	0.08	0.04	0.03	0.00	0.13	0.12	0.01	0.07	0.11
Al ₂ O ₃	12.98	12.27	12.75	12.71	12.76	12.53	12.91	12.68	12.19	12.68
FeO	1.48	1.43	1.34	1.51	1.34	1.52	1.47	1.51	1.56	1.49
MnO	0.05	0.08	0.00	0.02	0.00	0.03	0.08	0.01	0.07	0.04
MgO	0.09	0.07	0.06	0.07	0.05	0.04	0.07	0.08	0.05	0.06
CaO	0.94	0.84	0.93	0.91	0.88	0.93	0.93	0.86	0.93	0.93
Na ₂ O	3.71	3.98	4.05	4.03	3.72	3.90	4.05	4.01	4.04	3.97
K ₂ O	4.22	4.08	4.27	4.21	4.27	4.18	4.20	4.11	3.99	4.05
P ₂ O ₅	0.00	0.03	0.00	0.00	0.11	0.07	0.00	0.14	0.00	0.00
Total	94.12	95.57	95.70	95.44	94.61	95.03	94.32	95.05	94.29	94.90
ID:	gl_12	gl_13	gl_15	gl_16	gl_18	gl_19	gl_20	gl_21	gl_22	gl_23
SiO ₂	76.63	76.50	77.24	76.17	76.62	76.94	76.65	76.13	76.54	77.06
TiO ₂	0.06	0.06	0.00	0.04	0.04	0.14	0.09	0.06	0.05	0.12
Al ₂ O ₃	12.43	12.89	12.59	12.95	12.62	12.47	12.75	12.93	12.56	12.47
FeO	1.52	1.54	1.53	1.56	1.44	1.47	1.45	1.53	1.45	1.52
MnO	0.09	0.04	0.04	0.00	0.03	0.09	0.04	0.03	0.03	0.08
MgO	0.07	0.06	0.05	0.09	0.06	0.05	0.06	0.05	0.05	0.02
CaO	0.95	0.93	0.88	0.98	0.96	0.88	0.95	0.91	0.92	0.87
Na ₂ O	4.12	4.00	3.66	3.95	3.84	3.66	3.58	4.18	3.95	3.62
K ₂ O	4.12	3.99	4.00	4.27	4.39	4.24	4.40	4.14	4.36	4.22
P ₂ O ₅	0.00	0.00	0.01	0.00	0.00	0.05	0.05	0.05	0.09	0.02
Total	96.03	95.29	94.95	94.20	95.29	94.68	95.07	95.24	95.17	95.13
ID:	gl_24	gl_25	gl_26	gl_28	gl_29	gl_30				
SiO ₂	76.26	76.70	76.53	76.91	76.87	76.37				
TiO ₂	0.10	0.11	0.03	0.04	0.06	0.16				
Al ₂ O ₃	12.86	12.61	12.79	12.78	12.67	12.68				
FeO	1.36	1.40	1.40	1.35	1.46	1.53				
MnO	0.08	0.01	0.14	0.09	0.08	0.07				
MgO	0.02	0.03	0.02	0.06	0.06	0.08				
CaO	0.99	0.96	0.94	0.87	0.79	0.87				
Na ₂ O	3.97	3.97	3.86	3.71	3.86	4.07				
K ₂ O	4.37	4.21	4.30	4.20	4.14	4.17				
P ₂ O ₅	0.00	0.00	0.00	0.00	0.01	0.00				
Total	95.14	94.79	95.22	95.48	94.77	96.20				

TABLE D.10: Glass composition (*wt. %*) of Ebisutoge PD Unit C, EH65a (pumice)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10	gl_11	gl_12	gl_14	gl_15	gl_16	gl_17	gl_19
SiO ₂	76.26	75.71	76.39	76.32	76.76	75.92	76.15	76.44	76.96	76.14	76.26	76.80	77.09	76.79	76.95	76.43	76.67
TiO ₂	0.09	0.18	0.12	0.08	0.17	0.12	0.07	0.16	0.13	0.10	0.07	0.07	0.17	0.10	0.09	0.06	0.12
Al ₂ O ₃	12.68	13.00	12.71	12.78	12.86	12.77	12.99	12.61	12.57	12.94	12.79	12.53	12.53	12.88	12.44	12.86	12.55
FeO	1.51	1.50	1.42	1.55	1.29	1.40	1.43	1.47	1.31	1.63	1.61	1.49	1.41	1.43	1.42	1.38	1.43
MnO	0.07	0.07	0.00	0.04	0.07	0.00	0.06	0.04	0.01	0.00	0.04	0.07	0.03	0.05	0.02	0.05	0.06
MgO	0.06	0.09	0.03	0.09	0.04	0.04	0.04	0.05	0.07	0.02	0.06	0.05	0.10	0.04	0.04	0.02	0.06
CaO	0.95	1.01	0.94	0.90	0.86	1.01	0.84	1.00	1.04	0.96	1.02	1.01	0.93	0.91	0.92	1.03	0.88
Na ₂ O	3.93	4.10	4.00	4.00	3.62	4.27	4.00	4.01	3.52	4.02	3.53	3.58	3.78	3.76	3.88	4.06	3.94
K ₂ O	4.41	4.34	4.30	4.21	4.33	4.40	4.42	4.15	4.41	4.19	4.62	4.39	3.90	4.04	4.23	4.12	4.26
P ₂ O ₅	0.04	0.00	0.09	0.03	0.00	0.07	0.00	0.07	0.00	0.00	0.00	0.00	0.07	0.00	0.03	0.01	0.04
Total	94.95	94.43	94.74	95.58	94.22	94.45	94.43	94.70	94.41	94.89	94.90	94.63	95.56	94.94	94.64	94.89	94.99

ID:	gl_20	gl_21	gl_22	gl_23	gl_25	gl_27	gl_28	gl_29	gl_31	gl_33	gl_34	gl_35	gl_36	gl_37	gl_38	gl_40	gl_43
SiO ₂	76.17	76.38	76.28	76.30	76.22	77.31	76.17	76.16	76.64	76.12	76.39	76.65	76.55	76.86	76.47	76.40	77.12
TiO ₂	0.09	0.10	0.03	0.12	0.14	0.05	0.05	0.06	0.17	0.04	0.07	0.08	0.12	0.00	0.00	0.00	0.03
Al ₂ O ₃	12.66	12.90	12.84	12.80	12.85	12.33	12.82	13.10	12.80	13.00	12.89	12.64	12.68	12.84	12.69	12.78	12.32
FeO	1.71	1.53	1.48	1.36	1.41	1.40	1.41	1.47	1.41	1.38	1.49	1.44	1.53	0.89	1.60	1.51	1.31
MnO	0.07	0.08	0.06	0.04	0.07	0.00	0.00	0.00	0.08	0.09	0.00	0.13	0.04	0.04	0.00	0.11	0.10
MgO	0.03	0.07	0.02	0.00	0.04	0.07	0.05	0.04	0.05	0.01	0.03	0.12	0.04	0.05	0.07	0.08	0.02
CaO	1.08	0.96	1.03	1.00	1.00	0.98	1.02	0.81	0.96	0.86	1.01	1.00	0.90	0.89	0.95	0.91	0.89
Na ₂ O	3.90	3.75	3.65	3.95	3.99	3.67	3.95	4.09	3.70	4.27	3.79	3.76	3.94	3.54	3.84	3.99	3.95
K ₂ O	4.29	4.23	4.62	4.38	4.17	4.15	4.47	4.22	4.17	4.20	4.26	4.18	4.20	4.87	4.31	4.19	4.16
P ₂ O ₅	0.01	0.00	0.00	0.04	0.10	0.04	0.05	0.05	0.03	0.02	0.07	0.00	0.00	0.01	0.07	0.03	0.11
Total	95.67	94.60	94.55	94.52	94.92	95.51	94.59	94.80	94.27	94.75	94.09	94.40	94.78	94.16	94.47	95.20	95.05

ID:	gl_44	gl_45	gl_46	gl_47	gl_48	gl_49	gl_50	gl_52	gl_53	gl_54	gl_55	gl_56	gl_57	gl_58	gl_59	gl_60
SiO ₂	76.89	76.09	76.42	76.81	76.90	77.39	76.56	77.07	76.98	76.10	76.80	76.25	77.30	76.87	76.40	76.48
TiO ₂	0.06	0.08	0.07	0.08	0.11	0.13	0.08	0.09	0.00	0.19	0.10	0.10	0.05	0.12	0.12	0.08
Al ₂ O ₃	12.42	12.84	12.65	12.76	12.47	12.45	12.47	12.38	12.52	12.84	12.57	12.66	12.33	12.43	12.87	12.71
FeO	1.56	1.51	1.49	1.49	1.31	1.28	1.59	1.43	1.47	1.72	1.48	1.49	1.47	1.38	1.44	1.47
MnO	0.02	0.05	0.07	0.10	0.08	0.04	0.05	0.09	0.02	0.01	0.08	0.07	0.00	0.07	0.06	0.12
MgO	0.08	0.04	0.06	0.05	0.04	0.04	0.08	0.07	0.03	0.09	0.03	0.04	0.06	0.04	0.07	0.05
CaO	0.86	1.02	0.93	0.88	0.93	0.92	0.93	0.96	1.05	0.91	0.89	0.95	0.98	0.98	0.87	0.85
Na ₂ O	3.95	4.06	4.10	3.60	3.89	3.64	3.92	3.73	3.53	3.90	3.84	3.93	3.64	3.92	3.87	3.81
K ₂ O	4.14	4.33	4.22	4.21	4.18	4.07	4.29	4.17	4.36	4.23	4.22	4.42	4.18	4.20	4.29	4.44
P ₂ O ₅	0.01	0.00	0.00	0.01	0.09	0.04	0.02	0.00	0.05	0.01	0.00	0.09	0.00	0.00	0.01	0.00
Total	94.88	94.30	94.14	94.21	95.16	94.67	95.15	94.81	95.31	95.24	94.64	94.76	94.83	95.72	95.33	95.60

TABLE D.11: Glass composition (wt.%) of Ebisutoge PD Unit D, EH65b (fiamme)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10
SiO ₂	76.81	76.97	76.55	76.61	76.90	76.79	76.34	76.79	77.49	76.91
TiO ₂	0.14	0.11	0.09	0.12	0.16	0.05	0.07	0.07	0.11	0.05
Al ₂ O ₃	12.83	12.57	12.82	12.82	12.82	13.00	12.79	12.64	12.57	12.62
FeO	1.16	1.23	1.42	1.38	1.18	1.17	1.60	1.45	0.69	1.43
MnO	0.05	0.00	0.07	0.02	0.06	0.06	0.00	0.01	0.01	0.00
MgO	0.09	0.02	0.05	0.03	0.02	0.04	0.06	0.03	0.01	0.03
CaO	0.85	0.93	1.00	0.94	0.90	0.96	1.04	0.93	1.02	0.96
Na ₂ O	3.77	3.92	3.94	4.00	3.72	3.50	3.85	3.92	3.54	3.89
K ₂ O	4.28	4.16	4.07	4.09	4.20	4.33	4.25	4.16	4.46	4.10
P ₂ O ₅	0.01	0.08	0.00	0.00	0.04	0.12	0.00	0.00	0.11	0.00
Total	96.24	95.09	95.84	96.09	95.80	95.42	95.14	95.16	95.71	96.74
ID:	gl_11	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_21
SiO ₂	76.84	76.47	76.61	77.18	77.00	77.12	76.64	76.41	76.65	77.02
TiO ₂	0.08	0.14	0.11	0.06	0.12	0.05	0.06	0.03	0.13	0.04
Al ₂ O ₃	12.48	12.88	12.81	12.44	12.34	12.72	12.70	13.21	12.63	12.63
FeO	1.51	1.40	1.54	1.43	1.49	1.44	1.45	1.45	1.48	1.29
MnO	0.00	0.03	0.02	0.05	0.03	0.00	0.09	0.04	0.03	0.07
MgO	0.07	0.03	0.03	0.03	0.03	0.04	0.06	0.02	0.01	0.03
CaO	0.90	1.01	0.85	0.84	0.94	0.88	0.86	0.98	0.94	0.96
Na ₂ O	3.98	3.76	3.88	3.85	3.76	3.52	3.85	3.58	3.92	3.89
K ₂ O	4.13	4.26	4.14	4.08	4.28	4.23	4.28	4.25	4.10	4.06
P ₂ O ₅	0.00	0.01	0.00	0.04	0.00	0.00	0.01	0.02	0.13	0.00
Total	96.62	95.04	95.81	96.22	96.02	95.25	96.46	95.13	96.55	96.70
ID:	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_30	gl_31
SiO ₂	77.28	77.49	76.53	76.66	76.88	76.67	76.73	76.60	76.79	76.91
TiO ₂	0.12	0.01	0.12	0.11	0.13	0.11	0.12	0.01	0.14	0.07
Al ₂ O ₃	12.46	12.67	13.02	12.84	12.78	12.71	12.66	12.86	12.70	12.76
FeO	1.34	1.05	1.46	1.36	1.46	1.43	1.40	1.36	1.32	1.40
MnO	0.07	0.03	0.05	0.09	0.00	0.05	0.01	0.05	0.00	0.00
MgO	0.00	0.00	0.06	0.05	0.05	0.03	0.03	0.07	0.04	0.09
CaO	0.93	0.88	1.01	0.85	0.84	0.96	0.90	0.93	1.04	0.96
Na ₂ O	3.67	3.65	3.59	3.80	3.62	3.94	3.92	3.93	3.66	3.60
K ₂ O	4.13	4.20	4.16	4.13	4.19	4.04	4.17	4.20	4.31	4.20
P ₂ O ₅	0.00	0.03	0.00	0.11	0.05	0.06	0.06	0.00	0.00	0.00
Total	95.74	96.19	95.09	95.54	96.45	96.69	96.13	96.05	95.34	96.37
ID:	gl_32	gl_33	gl_34	gl_35	gl_36	gl_37	gl_39	gl_40		
SiO ₂	76.30	76.67	76.53	77.41	76.82	77.00	76.38	76.21		
TiO ₂	0.16	0.09	0.10	0.04	0.10	0.11	0.09	0.08		
Al ₂ O ₃	12.77	12.93	12.78	12.54	12.45	12.48	12.90	12.91		
FeO	1.50	1.19	1.41	1.26	1.38	1.33	1.46	1.51		
MnO	0.05	0.06	0.06	0.05	0.12	0.03	0.12	0.06		
MgO	0.07	0.06	0.06	0.03	0.03	0.06	0.07	0.06		
CaO	0.92	0.96	1.00	0.98	0.91	0.93	1.01	1.01		
Na ₂ O	3.90	3.48	3.64	3.55	3.64	3.76	3.53	3.79		
K ₂ O	4.33	4.54	4.32	4.06	4.49	4.31	4.46	4.37		
P ₂ O ₅	0.00	0.02	0.11	0.08	0.06	0.00	0.00	0.00		
Total	95.87	95.68	95.00	95.96	96.40	96.51	95.59	95.75		

TABLE D.12: Glass composition (*wt. %*) of Ebisutoge PD Unit D, EH65b (matrix)

ID:	gl_1	gl_2	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10	gl_11
SiO ₂	76.78	76.87	77.13	76.95	76.96	76.69	76.53	77.30	76.75	76.65
TiO ₂	0.14	0.15	0.06	0.10	0.05	0.14	0.05	0.11	0.08	0.14
Al ₂ O ₃	12.53	12.63	12.60	12.65	12.77	12.67	12.70	12.51	12.80	12.89
FeO	1.57	1.38	1.27	1.50	1.39	1.50	1.55	1.33	1.08	1.44
MnO	0.08	0.11	0.03	0.04	0.09	0.05	0.06	0.09	0.11	0.02
MgO	0.00	0.09	0.09	0.04	0.08	0.06	0.08	0.01	0.03	0.05
CaO	1.00	0.96	0.85	0.94	0.86	0.85	0.86	0.85	1.06	0.83
Na ₂ O	3.48	3.56	3.58	3.55	3.44	3.80	3.85	3.52	3.49	3.47
K ₂ O	4.43	4.24	4.39	4.22	4.28	4.24	4.31	4.24	4.57	4.50
P ₂ O ₅	0.00	0.00	0.00	0.00	0.08	0.00	0.01	0.03	0.02	0.00
Total	95.64	96.31	96.05	96.35	96.09	96.42	96.56	95.81	95.16	95.09
ID:	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_20	gl_21
SiO ₂	76.45	76.82	76.92	76.85	76.31	76.88	76.59	76.96	77.08	76.85
TiO ₂	0.19	0.10	0.08	0.09	0.04	0.07	0.15	0.04	0.07	0.13
Al ₂ O ₃	12.98	12.77	12.64	12.66	13.06	12.70	12.57	12.71	12.38	12.43
FeO	1.43	1.14	1.40	1.51	1.43	1.42	1.74	1.41	1.39	1.50
MnO	0.05	0.13	0.09	0.04	0.13	0.05	0.03	0.06	0.14	0.11
MgO	0.05	0.05	0.07	0.03	0.04	0.08	0.05	0.04	0.02	0.05
CaO	0.94	0.97	0.91	0.89	0.99	0.95	0.90	0.87	1.01	0.94
Na ₂ O	3.48	3.51	3.44	3.54	3.71	3.58	3.56	3.60	3.48	3.55
K ₂ O	4.41	4.47	4.35	4.34	4.30	4.26	4.36	4.33	4.43	4.42
P ₂ O ₅	0.02	0.03	0.09	0.04	0.00	0.00	0.04	0.00	0.00	0.04
Total	95.53	96.45	95.55	96.33	95.40	95.74	96.24	95.92	95.93	96.23
ID:	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_30	gl_31
SiO ₂	76.56	77.15	77.43	77.51	76.56	76.78	77.34	77.08	77.53	77.09
TiO ₂	0.15	0.09	0.08	0.11	0.09	0.07	0.18	0.07	0.09	0.14
Al ₂ O ₃	12.76	12.37	12.34	12.53	12.60	12.55	12.70	12.76	12.15	12.23
FeO	1.33	1.46	1.39	0.79	1.35	1.31	0.81	1.23	1.39	1.51
MnO	0.05	0.09	0.05	0.04	0.01	0.02	0.03	0.08	0.00	0.02
MgO	0.02	0.04	0.03	0.02	0.03	0.07	0.04	0.04	0.06	0.09
CaO	0.85	0.83	0.87	0.87	0.90	0.96	0.86	0.81	0.90	0.85
Na ₂ O	3.72	3.54	3.33	3.66	3.93	3.69	3.40	3.51	3.36	3.65
K ₂ O	4.52	4.41	4.48	4.44	4.50	4.46	4.65	4.42	4.48	4.42
P ₂ O ₅	0.03	0.03	0.00	0.02	0.03	0.10	0.00	0.00	0.05	0.00
Total	96.65	96.88	95.37	95.67	96.64	96.69	96.10	95.99	96.25	96.69
ID:	gl_32	gl_33	gl_34	gl_35	gl_36	gl_37	gl_38	gl_39	gl_40	
SiO ₂	76.67	76.68	76.64	76.55	76.23	76.55	76.53	77.70	76.97	
TiO ₂	0.15	0.08	0.08	0.04	0.25	0.06	0.12	0.20	0.17	
Al ₂ O ₃	12.73	12.54	12.90	12.86	12.93	12.69	12.69	12.65	12.93	
FeO	1.32	1.45	0.98	1.51	1.51	1.30	1.40	0.55	0.79	
MnO	0.08	0.05	0.00	0.02	0.03	0.00	0.04	0.07	0.08	
MgO	0.05	0.08	0.00	0.08	0.03	0.08	0.05	0.04	0.01	
CaO	0.98	0.97	0.92	0.98	0.90	0.88	0.89	0.76	0.89	
Na ₂ O	3.53	3.63	3.61	3.34	3.69	3.86	3.77	3.20	3.31	
K ₂ O	4.49	4.53	4.86	4.55	4.42	4.58	4.45	4.80	4.78	
P ₂ O ₅	0.00	0.00	0.00	0.07	0.00	0.00	0.05	0.02	0.08	
Total	95.93	95.95	95.25	95.51	96.48	96.40	96.21	96.18	96.07	

TABLE D.13: Glass composition (wt.%) of Ebisutoge PD Unit D, EH65c (fiamme)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10
SiO ₂	76.18	76.55	76.59	76.43	77.12	76.95	77.40	76.72	76.88	76.94
TiO ₂	0.10	0.09	0.12	0.12	0.06	0.08	0.06	0.14	0.10	0.15
Al ₂ O ₃	12.92	12.72	12.82	12.94	12.71	12.86	12.90	12.97	12.82	12.54
FeO	1.29	1.18	1.11	0.86	0.92	0.75	0.60	0.65	0.82	0.98
MnO	0.10	0.07	0.07	0.07	0.00	0.01	0.04	0.04	0.06	0.09
MgO	0.06	0.09	0.07	0.05	0.00	0.00	0.01	0.00	0.06	0.00
CaO	0.91	0.91	0.86	0.99	0.94	0.96	0.90	0.82	0.85	0.87
Na ₂ O	3.56	3.44	2.81	3.54	3.32	3.49	3.30	3.07	3.45	3.44
K ₂ O	4.86	4.93	5.52	5.00	4.94	4.90	4.79	5.46	4.92	4.99
P ₂ O ₅	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.13	0.04	0.00
Total	95.83	96.13	95.57	95.72	96.49	94.88	96.22	95.83	96.40	95.67
ID:	gl_11	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_20
SiO ₂	76.71	76.85	77.05	76.98	76.31	76.96	76.65	76.11	76.65	77.08
TiO ₂	0.13	0.10	0.14	0.03	0.04	0.10	0.00	0.11	0.08	0.07
Al ₂ O ₃	12.66	12.76	12.53	12.69	13.19	12.71	12.76	13.18	12.82	12.83
FeO	1.30	1.09	1.09	0.64	0.70	0.74	1.06	0.98	1.07	0.78
MnO	0.07	0.09	0.00	0.08	0.05	0.06	0.06	0.05	0.03	0.04
MgO	0.03	0.02	0.06	0.00	0.04	0.05	0.04	0.02	0.00	0.01
CaO	0.90	0.81	0.93	0.97	0.94	0.95	0.97	0.90	0.98	0.90
Na ₂ O	3.37	3.32	3.36	3.60	3.74	3.47	3.53	3.72	3.45	3.47
K ₂ O	4.80	4.97	4.85	4.94	4.92	4.92	4.83	4.90	4.93	4.82
P ₂ O ₅	0.03	0.00	0.00	0.07	0.08	0.05	0.08	0.04	0.00	0.01
Total	96.47	95.24	95.64	95.91	94.24	95.80	95.42	95.76	95.18	96.38
ID:	gl_21	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_30
SiO ₂	76.72	77.47	77.18	76.75	77.23	76.70	77.83	77.02	76.85	77.19
TiO ₂	0.00	0.09	0.05	0.11	0.10	0.10	0.06	0.10	0.06	0.01
Al ₂ O ₃	12.63	12.46	12.51	12.42	12.56	12.52	12.48	12.61	12.71	12.71
FeO	1.32	0.71	1.13	1.57	1.55	1.52	0.39	1.33	1.23	1.11
MnO	0.11	0.06	0.04	0.07	0.04	0.04	0.08	0.05	0.06	0.00
MgO	0.05	0.00	0.06	0.02	0.04	0.05	0.03	0.05	0.00	0.01
CaO	0.89	0.91	0.97	0.97	0.95	0.91	0.95	0.94	0.99	0.92
Na ₂ O	3.49	3.29	3.24	3.30	2.88	3.20	3.40	3.21	3.36	3.29
K ₂ O	4.74	4.99	4.73	4.80	4.62	4.86	4.77	4.65	4.69	4.75
P ₂ O ₅	0.04	0.02	0.10	0.00	0.04	0.10	0.00	0.04	0.04	0.01
Total	96.27	96.49	96.58	96.75	96.33	96.56	96.48	96.08	96.82	95.95
ID:	gl_31	gl_32	gl_33	gl_34	gl_35	gl_36	gl_37	gl_38	gl_39	gl_40
SiO ₂	77.48	77.61	77.48	76.95	77.04	77.03	76.68	77.73	76.98	76.35
TiO ₂	0.14	0.06	0.12	0.12	0.10	0.04	0.03	0.12	0.11	0.10
Al ₂ O ₃	12.35	12.54	12.57	12.82	12.71	12.62	12.93	12.46	12.75	13.02
FeO	1.03	0.91	0.86	0.57	1.17	1.16	1.15	0.43	1.32	1.29
MnO	0.01	0.02	0.15	0.10	0.00	0.07	0.02	0.03	0.02	0.04
MgO	0.05	0.00	0.04	0.07	0.06	0.00	0.02	0.00	0.04	0.05
CaO	0.85	0.87	0.80	0.75	0.87	0.97	0.94	0.92	0.82	0.84
Na ₂ O	3.35	3.39	3.30	3.36	3.20	3.30	3.50	3.12	3.30	3.51
K ₂ O	4.74	4.61	4.68	5.18	4.85	4.77	4.73	5.09	4.67	4.70
P ₂ O ₅	0.00	0.00	0.00	0.07	0.00	0.03	0.02	0.10	0.00	0.11
Total	95.29	96.28	96.34	95.30	95.66	96.03	95.22	96.12	96.77	95.20

TABLE D.14: Glass composition (*wt. %*) of Ebisutoge PD Unit D, EH65c (matrix)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10
SiO ₂	76.24	76.72	76.52	76.76	76.13	77.41	76.98	76.61	76.59	76.44
TiO ₂	0.10	0.11	0.05	0.10	0.08	0.14	0.08	0.13	0.04	0.06
Al ₂ O ₃	12.80	12.55	12.62	12.74	12.94	12.35	12.39	12.59	12.62	12.92
FeO	1.39	1.43	1.47	1.04	1.49	0.88	1.44	1.41	1.47	1.52
MnO	0.11	0.11	0.05	0.01	0.08	0.07	0.04	0.07	0.10	0.12
MgO	0.08	0.06	0.04	0.17	0.03	0.01	0.03	0.06	0.10	0.06
CaO	0.77	0.79	0.69	0.83	0.78	0.81	0.77	0.87	0.85	0.82
Na ₂ O	3.46	3.32	3.76	3.04	3.54	3.36	3.53	3.49	3.46	3.47
K ₂ O	5.06	4.91	4.80	5.31	4.94	4.85	4.74	4.71	4.69	4.56
P ₂ O ₅	0.00	0.00	0.00	0.01	0.00	0.12	0.00	0.06	0.09	0.04
Total	95.66	95.95	96.21	94.85	95.21	96.05	95.80	96.27	95.74	96.29
ID:	gl_11	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_20
SiO ₂	76.81	77.16	77.07	77.59	77.00	76.30	76.55	76.51	76.68	76.37
TiO ₂	0.12	0.09	0.06	0.04	0.04	0.04	0.14	0.12	0.11	0.12
Al ₂ O ₃	12.75	12.63	12.39	12.78	12.78	12.81	12.84	12.66	12.67	13.02
FeO	1.34	0.98	1.34	0.57	0.90	1.47	1.03	1.47	1.28	1.56
MnO	0.07	0.06	0.12	0.03	0.10	0.09	0.07	0.14	0.06	0.05
MgO	0.02	0.03	0.08	0.07	0.04	0.03	0.04	0.05	0.03	0.04
CaO	0.86	0.78	0.88	0.81	0.83	0.97	0.90	0.81	0.90	0.87
Na ₂ O	3.39	3.39	3.46	3.43	3.40	3.45	3.44	3.59	3.50	3.20
K ₂ O	4.61	4.78	4.61	4.68	4.91	4.80	4.97	4.64	4.77	4.77
P ₂ O ₅	0.04	0.11	0.00	0.00	0.00	0.03	0.00	0.00	0.00	0.00
Total	95.56	95.88	95.98	95.56	94.84	94.40	94.05	96.33	94.78	94.69
ID:	gl_21	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_30
SiO ₂	77.47	77.14	76.92	76.84	77.00	77.07	76.76	76.65	76.53	76.53
TiO ₂	0.11	0.11	0.07	0.05	0.15	0.13	0.09	0.09	0.11	0.07
Al ₂ O ₃	12.37	12.65	12.88	12.51	12.52	12.49	12.62	12.87	12.80	12.79
FeO	0.94	1.14	1.31	1.31	1.43	1.44	1.48	1.52	1.28	1.32
MnO	0.08	0.03	0.11	0.06	0.06	0.00	0.10	0.00	0.08	0.13
MgO	0.08	0.08	0.03	0.06	0.07	0.06	0.06	0.07	0.06	0.08
CaO	0.81	0.87	0.89	0.81	0.81	0.80	0.75	0.96	0.82	0.86
Na ₂ O	3.55	3.32	3.19	3.46	3.41	3.32	3.28	3.08	3.35	3.49
K ₂ O	4.60	4.59	4.62	4.84	4.53	4.65	4.85	4.76	4.92	4.73
P ₂ O ₅	0.00	0.08	0.00	0.06	0.01	0.06	0.00	0.00	0.04	0.01
Total	96.67	96.15	96.35	96.99	96.62	96.44	96.34	95.89	95.74	96.04
ID:	gl_31	gl_32	gl_33	gl_34	gl_35	gl_37	gl_38	gl_39	gl_40	
SiO ₂	76.84	76.47	76.64	76.54	77.12	76.61	77.06	77.39	76.90	
TiO ₂	0.06	0.19	0.16	0.21	0.08	0.14	0.07	0.05	0.04	
Al ₂ O ₃	12.40	13.00	12.61	12.51	12.37	12.52	12.39	12.72	12.42	
FeO	1.58	1.18	1.53	1.56	1.40	1.63	1.49	0.74	1.51	
MnO	0.03	0.02	0.07	0.06	0.05	0.06	0.05	0.04	0.07	
MgO	0.03	0.06	0.05	0.06	0.06	0.01	0.04	0.01	0.07	
CaO	0.88	0.82	0.80	0.80	0.92	0.79	0.82	0.77	0.70	
Na ₂ O	3.35	3.27	3.24	3.40	3.22	3.32	3.15	3.23	3.35	
K ₂ O	4.83	4.97	4.82	4.76	4.80	4.88	4.93	5.05	4.94	
P ₂ O ₅	0.00	0.02	0.10	0.10	0.00	0.04	0.01	0.00	0.01	
Total	96.44	94.98	96.38	96.43	96.69	96.31	96.48	96.39	96.73	

TABLE D.15: Glass composition (wt.%) of Ebisutoge PD Unit D, EH65d (fiamme)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10
SiO ₂	76.84	76.50	76.80	76.71	76.44	76.27	76.66	76.79	76.44	76.10
TiO ₂	0.05	0.04	0.10	0.11	0.07	0.06	0.07	0.21	0.09	0.11
Al ₂ O ₃	12.95	13.02	13.14	12.79	13.12	13.08	13.07	12.93	13.19	13.21
FeO	0.42	0.67	0.74	0.90	0.86	0.89	0.86	0.41	0.77	0.76
MnO	0.10	0.01	0.05	0.10	0.04	0.05	0.04	0.00	0.11	0.00
MgO	0.02	0.00	0.03	0.00	0.09	0.04	0.03	0.00	0.02	0.06
CaO	0.94	1.11	0.96	0.91	0.86	0.88	1.00	1.00	0.97	0.97
Na ₂ O	3.65	3.58	3.26	3.49	3.47	3.54	3.49	3.35	3.42	3.66
K ₂ O	4.99	5.04	4.92	5.00	4.96	5.08	4.75	5.30	4.94	5.14
P ₂ O ₅	0.04	0.03	0.00	0.00	0.08	0.11	0.02	0.00	0.05	0.00
Total	95.38	94.42	94.74	94.68	94.33	94.43	95.47	95.25	94.14	94.71
ID:	gl_11	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_20
SiO ₂	76.14	76.00	76.52	76.09	76.52	76.50	76.89	76.76	77.02	76.58
TiO ₂	0.18	0.14	0.09	0.09	0.09	0.08	0.10	0.04	0.05	0.06
Al ₂ O ₃	13.01	13.10	12.98	13.10	13.20	12.99	13.15	12.89	12.77	12.98
FeO	1.06	1.30	0.90	0.96	1.07	0.67	0.57	0.90	0.66	0.41
MnO	0.04	0.09	0.09	0.12	0.00	0.07	0.00	0.00	0.04	0.11
MgO	0.01	0.01	0.03	0.06	0.05	0.03	0.04	0.00	0.07	0.01
CaO	0.94	0.99	0.84	1.00	0.80	0.89	0.83	0.94	0.84	0.97
Na ₂ O	3.51	3.41	3.46	3.48	3.32	3.68	3.40	3.38	3.50	3.88
K ₂ O	5.09	4.93	5.04	5.03	4.93	5.06	5.02	5.03	5.04	4.97
P ₂ O ₅	0.03	0.04	0.04	0.08	0.00	0.03	0.00	0.06	0.00	0.03
Total	94.48	94.20	95.16	95.98	95.07	95.17	95.61	94.96	95.60	95.26
ID:	gl_21	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_30
SiO ₂	76.32	76.30	76.11	76.64	76.63	76.77	76.55	76.84	76.51	76.71
TiO ₂	0.05	0.01	0.07	0.12	0.16	0.09	0.15	0.09	0.13	0.11
Al ₂ O ₃	13.03	12.99	12.85	12.88	12.27	12.88	12.53	12.45	12.83	12.58
FeO	1.36	1.51	1.50	0.91	1.49	1.29	1.48	1.28	1.42	1.38
MnO	0.02	0.09	0.05	0.11	0.05	0.10	0.00	0.06	0.07	0.06
MgO	0.01	0.05	0.07	0.04	0.02	0.04	0.08	0.07	0.05	0.02
CaO	0.86	0.91	0.87	0.95	0.95	0.89	1.03	0.85	0.87	0.93
Na ₂ O	3.32	3.32	3.42	2.92	3.49	3.15	3.33	3.55	2.87	3.44
K ₂ O	4.96	4.82	5.01	5.43	4.87	4.79	4.86	4.80	5.24	4.75
P ₂ O ₅	0.06	0.00	0.04	0.01	0.07	0.01	0.00	0.02	0.00	0.02
Total	96.14	95.65	95.94	95.94	96.07	96.31	96.16	96.53	96.39	95.90
ID:	gl_32	gl_33	gl_34	gl_35	gl_36	gl_37	gl_38	gl_39	gl_40	
SiO ₂	76.88	76.97	77.49	77.28	77.39	77.29	76.83	76.84	76.95	
TiO ₂	0.07	0.13	0.06	0.07	0.02	0.11	0.09	0.10	0.05	
Al ₂ O ₃	12.83	12.38	12.57	12.47	12.69	12.55	12.66	12.57	12.54	
FeO	1.30	1.50	1.03	1.00	1.18	1.15	1.39	1.35	1.34	
MnO	0.14	0.03	0.04	0.06	0.04	0.00	0.05	0.06	0.06	
MgO	0.00	0.04	0.03	0.03	0.08	0.04	0.05	0.02	0.04	
CaO	0.93	0.88	0.81	0.88	0.88	0.82	0.91	0.85	0.92	
Na ₂ O	3.15	2.74	3.30	3.44	2.98	2.41	3.39	3.42	3.46	
K ₂ O	4.70	5.34	4.68	4.78	4.71	5.62	4.57	4.79	4.63	
P ₂ O ₅	0.00	0.00	0.00	0.00	0.03	0.00	0.04	0.00	0.00	
Total	95.67	96.79	95.98	96.49	96.17	96.74	96.93	96.40	96.32	

TABLE D.16: Glass composition (*wt. %*) of Ebisutoge PD Unit D, EH65d (matrix)

ID:	gl_1	gl_2	gl_3	gl_4	gl_5	gl_6	gl_7	gl_8	gl_9	gl_10
SiO ₂	76.04	76.34	76.64	76.38	76.96	76.47	77.12	76.97	75.92	76.62
TiO ₂	0.10	0.07	0.07	0.16	0.12	0.08	0.11	0.10	0.03	0.10
Al ₂ O ₃	12.93	13.01	13.01	12.54	12.63	12.50	12.92	12.67	12.59	12.52
FeO	1.26	1.49	1.25	1.51	1.16	1.60	0.78	1.07	1.76	1.36
MnO	0.05	0.00	0.03	0.11	0.03	0.07	0.04	0.06	0.09	0.02
MgO	0.00	0.00	0.07	0.04	0.06	0.06	0.06	0.04	0.06	0.05
CaO	0.85	0.84	0.89	0.88	0.75	0.86	0.82	0.79	0.91	0.88
Na ₂ O	3.81	3.33	3.24	3.48	3.24	3.44	3.35	3.37	3.46	3.49
K ₂ O	4.89	4.91	4.73	4.86	5.03	4.91	4.80	4.91	5.04	4.95
P ₂ O ₅	0.06	0.00	0.06	0.05	0.02	0.01	0.00	0.02	0.15	0.01
Total	96.28	94.66	95.60	96.92	96.28	96.01	95.77	95.96	96.23	96.15
ID:	gl_11	gl_12	gl_13	gl_14	gl_15	gl_16	gl_17	gl_18	gl_19	gl_20
SiO ₂	76.37	76.35	76.79	76.55	76.10	76.04	76.73	76.28	76.36	75.90
TiO ₂	0.07	0.15	0.13	0.11	0.05	0.07	0.06	0.17	0.16	0.11
Al ₂ O ₃	13.09	12.86	12.98	12.69	12.98	12.91	12.66	12.84	12.67	12.68
FeO	0.90	1.34	0.63	1.40	1.47	1.62	0.96	1.30	1.26	1.68
MnO	0.07	0.07	0.14	0.06	0.03	0.07	0.00	0.02	0.05	0.13
MgO	0.11	0.07	0.06	0.03	0.08	0.08	0.02	0.05	0.05	0.05
CaO	0.89	0.86	0.75	0.87	0.83	0.73	0.74	0.85	0.81	0.84
Na ₂ O	2.80	3.41	3.17	3.20	3.38	3.43	3.50	3.39	3.35	3.58
K ₂ O	5.70	4.88	5.34	5.08	5.08	5.05	5.31	5.12	5.21	5.00
P ₂ O ₅	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.08	0.03
Total	95.62	96.40	96.05	95.11	95.68	95.84	95.63	94.77	96.68	95.91
ID:	gl_21	gl_22	gl_23	gl_24	gl_25	gl_26	gl_27	gl_28	gl_29	gl_32
SiO ₂	76.67	76.81	76.32	76.93	76.68	78.03	76.94	76.44	76.54	76.53
TiO ₂	0.11	0.10	0.20	0.12	0.10	0.08	0.08	0.13	0.16	0.10
Al ₂ O ₃	12.66	12.42	12.63	12.74	12.42	11.90	12.35	12.41	12.63	12.62
FeO	1.41	1.32	1.26	1.29	1.60	1.23	1.43	1.54	1.48	1.49
MnO	0.08	0.06	0.04	0.00	0.05	0.08	0.00	0.02	0.08	0.04
MgO	0.05	0.07	0.08	0.05	0.03	0.06	0.09	0.09	0.05	0.05
CaO	0.81	0.84	0.84	0.81	0.74	0.69	0.94	0.89	0.80	0.79
Na ₂ O	3.44	3.45	3.56	3.33	3.46	3.28	3.40	3.50	3.37	3.49
K ₂ O	4.76	4.94	4.94	4.72	4.86	4.62	4.76	4.93	4.89	4.83
P ₂ O ₅	0.00	0.00	0.14	0.01	0.06	0.03	0.00	0.06	0.00	0.06
Total	96.02	95.67	96.90	96.44	96.83	96.25	96.19	96.38	96.07	95.16
ID:	gl_33	gl_34	gl_35	gl_36	gl_37	gl_38	gl_39	gl_40		
SiO ₂	77.37	76.87	77.35	76.85	76.37	76.77	76.54	75.91		
TiO ₂	0.07	0.11	0.18	0.04	0.07	0.15	0.07	0.07		
Al ₂ O ₃	12.54	12.96	12.84	12.71	12.75	12.67	12.65	13.14		
FeO	0.81	1.12	0.77	1.30	1.55	1.25	1.49	1.45		
MnO	0.00	0.02	0.07	0.09	0.05	0.02	0.11	0.13		
MgO	0.04	0.09	0.07	0.04	0.05	0.01	0.06	0.07		
CaO	0.81	0.81	0.81	0.91	0.78	0.78	0.71	0.83		
Na ₂ O	3.39	2.95	3.15	3.17	3.40	3.36	3.41	3.42		
K ₂ O	4.96	5.08	4.75	4.89	4.98	4.96	4.96	4.97		
P ₂ O ₅	0.00	0.00	0.01	0.00	0.00	0.04	0.00	0.00		
Total	96.61	95.56	95.99	95.65	96.04	96.45	96.36	95.61		

TABLE D.17: Composition and temperature and oxygen fugacity of magnetite-ilmenite pairs of the Nyukawa PFD

Sample:	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61	EH61
Magnetite wt%	mag1.2	mag1.3	mag1.4	mag1.5	mag2.1	mag2.2	mag2.3	mag2.4	mag2.5	mag3.1	mag3.2	mag3.3	mag3.4	mag3.5	mag4.1	mag4.2	mag4.4	mag4.5	mag5.3
SiO ₂	0.08	0.04	0.04	0.03	0.05	0.10	0.12	0.07	0.07	0.07	0.05	0.08	0.06	0.07	0.06	0.05	0.07	0.06	0.05
TiO ₂	9.56	9.53	8.56	9.57	9.61	9.96	9.82	9.89	9.63	11.07	11.59	9.18	12.74	12.47	11.36	8.71	9.93	9.37	10.08
Al ₂ O ₃	2.07	2.01	1.96	2.11	2.46	2.39	2.41	2.40	2.49	2.45	2.89	2.18	1.76	1.80	1.80	1.97	2.02	2.09	2.09
FeO(T)	80.23	82.13	82.95	80.59	81.51	80.28	80.92	81.35	81.59	78.16	79.89	82.20	79.90	79.06	79.07	82.89	81.34	80.79	80.79
MnO	0.76	0.87	0.79	0.86	0.81	0.81	0.77	0.79	0.76	0.50	0.49	0.44	0.53	0.53	0.83	0.73	0.71	0.68	0.51
MgO	0.29	0.40	0.39	0.40	0.37	0.42	0.40	0.38	0.38	0.41	0.42	0.36	0.43	0.43	0.36	0.33	0.37	0.38	0.41
CaO	0.01	0.01	0.04	0.08	0.02	0.03	0.05	0.07	0.05	0.01	0.05	0.07	0.02	0.00	0.05	0.08	0.05	0.05	0.00
Cr ₂ O ₃	0.08	0.10	0.08	0.08	0.12	0.11	0.10	0.11	0.10	0.10	0.09	0.07	0.10	0.08	0.10	0.10	0.09	0.10	0.10
NiO	0.01	0.01	0.02	0.03	0.03	0.03	0.03	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.03	0.10
Total	93.20	95.09	94.84	93.74	94.95	94.14	94.62	95.06	95.08	92.77	94.61	94.59	95.54	94.45	93.64	94.87	94.59	94.68	94.03
Recalculated after Carmichael (1967)																			
Fe ₂ O ₃ wt%	46.54	48.17	50.07	47.00	47.28	45.93	46.52	46.84	47.28	42.64	43.54	48.24	42.14	41.82	43.50	49.72	46.88	48.05	46.13
FeO wt%	38.35	38.79	37.90	38.30	38.97	38.96	39.06	39.21	39.05	39.79	40.71	38.80	41.98	41.43	39.93	38.15	39.16	38.71	39.28
Total	97.86	99.91	99.85	98.45	99.69	98.74	99.28	99.76	99.82	97.05	98.97	99.42	99.76	98.64	97.99	99.85	99.28	99.50	98.65
Ilmenite wt%																			
SiO ₂	0.00	il2	il4	il5	il2.1	il2.2	il2.3	il2.4	il2.5	il3.1	il3.2	il3.3	il3.4	il3.5	il4.1	il4.2	il4.4	il4.5	il5.3
TiO ₂	44.54	44.77	44.78	47.52	43.32	43.30	43.28	43.28	43.73	43.06	43.37	43.82	43.51	43.47	44.37	44.19	44.64	44.25	44.40
Al ₂ O ₃	0.15	0.13	0.11	0.09	0.16	0.20	0.20	0.20	0.38	0.34	0.31	0.18	0.15	0.21	0.18	0.13	0.16	0.15	0.12
FeO(T)	51.67	51.52	51.20	48.37	52.66	52.69	52.73	52.72	52.16	52.46	52.52	52.95	52.82	52.73	51.65	51.46	51.59	51.74	52.10
MnO	1.74	1.81	1.84	2.27	1.60	1.61	1.57	1.50	1.67	1.00	0.97	1.01	1.02	1.02	1.56	1.53	1.48	1.45	1.13
MgO	0.53	0.55	0.55	0.66	0.47	0.48	0.46	0.50	0.53	0.54	0.56	0.54	0.53	0.52	0.52	0.58	0.53	0.54	0.51
CaO	0.00	0.01	0.07	0.09	0.05	0.08	0.11	0.07	0.05	0.06	0.08	0.09	0.02	0.03	0.02	0.04	0.04	0.02	0.02
Cr ₂ O ₃	0.01	0.00	0.02	0.03	0.03	0.04	0.03	0.04	0.02	0.08	0.03	0.03	0.01	0.04	0.03	0.03	0.03	0.02	0.04
NiO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.00	0.00	0.05	0.00	0.03	0.01	0.00	0.00	0.02
Total	98.65	98.81	98.57	99.05	98.29	98.42	98.37	98.32	98.58	98.32	98.38	98.59	98.10	98.31	98.38	97.98	98.47	98.17	98.34
Recalculated after Carmichael (1967)																			
Fe ₂ O ₃ wt%	15.91	15.64	15.38	10.24	18.02	18.13	18.17	18.09	17.23	16.52	16.61	17.31	17.39	16.98	15.87	15.92	15.48	15.99	15.83
FeO wt%	37.36	37.45	37.36	39.16	36.45	36.37	36.38	36.44	36.66	37.59	37.58	37.38	37.17	37.45	37.37	37.13	37.66	37.36	37.86
Total	100.24	100.38	100.11	100.07	100.09	100.24	100.19	100.13	100.31	99.98	100.04	100.32	99.85	100.01	99.97	99.57	100.02	99.77	99.93
Giorso and Evans (2008)																			
Temperature (°C)	852	848	811	774	871	883	878	879	867	916	914	849	937	934	896	823	853	843	863
Fe ₂ O ₃ (wt%)	0.51	0.49	0.55	-0.17	0.66	0.64	0.65	0.65	0.61	0.55	0.51	0.67	0.45	0.46	0.39	0.60	0.47	0.56	0.52

TABLE D.19: (Continued)

Sample:	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62	EH62
	mag1.3	mag1.4	mag1.5	mag2.1	mag2.2	mag2.3	mag2.4	mag3.1	mag3.2	mag3.3	mag3.4	mag3.5	mag4.1	mag4.2	mag4.4	mag5.1	mag5.2	mag5.3	mag5.4
Magnetite wt%																			
SiO ₂	0.05	0.04	0.06	0.06	0.07	0.08	0.27	0.09	0.06	0.04	0.05	0.08	0.18	0.09	0.06	0.06	0.06	0.04	0.08
TiO ₂	11.99	12.35	12.67	11.88	12.30	11.20	11.92	12.42	12.22	12.48	12.42	11.54	12.15	10.83	11.50	11.95	12.37	12.33	12.03
Al ₂ O ₃	1.85	1.81	1.81	1.78	1.80	1.89	1.88	1.81	1.82	1.83	1.86	1.91	1.74	1.95	1.99	1.84	1.81	1.83	1.84
FeO(T)	80.20	79.87	79.59	79.83	79.59	80.50	77.37	79.67	80.10	79.78	80.00	80.62	78.04	81.15	80.61	80.31	79.97	80.20	80.12
MnO	0.56	0.59	0.55	0.55	0.56	0.50	0.55	0.57	0.51	0.56	0.56	0.69	0.54	0.48	0.52	0.58	0.54	0.56	0.59
MgO	0.63	0.63	0.63	0.61	0.63	0.57	0.63	0.59	0.63	0.65	0.62	0.57	0.60	0.54	0.60	0.62	0.60	0.65	0.61
CaO	0.01	0.01	0.02	0.02	0.03	0.10	0.08	0.02	0.02	0.01	0.02	0.01	0.21	0.25	0.04	0.04	0.01	0.03	0.10
Cr ₂ O ₃	0.07	0.09	0.09	0.03	0.06	0.05	0.06	0.05	0.06	0.05	0.05	0.05	0.07	0.09	0.04	0.07	0.07	0.10	0.08
NiO	0.01	0.00	0.00	0.03	0.01	0.02	0.04	0.00	0.05	0.00	0.02	0.03	0.00	0.00	0.04	0.04	0.00	0.03	0.01
Total	95.37	95.41	95.45	94.79	95.04	94.92	92.73	95.22	95.46	95.39	95.60	95.33	93.54	95.39	95.39	95.50	95.43	95.78	95.45
Recalculated after Carmichael (1967)																			
Fe ₂ O ₃ wt%	43.62	42.96	42.29	43.49	42.76	44.79	41.17	42.58	43.22	42.73	42.89	44.35	41.78	45.81	44.46	43.77	42.89	43.23	43.54
FeO wt%	40.95	41.22	41.53	40.70	41.11	40.20	40.33	41.36	41.21	41.33	41.40	40.71	40.45	39.93	40.61	40.92	41.37	41.30	40.94
Total	99.74	99.71	99.69	99.15	99.32	99.41	96.85	99.49	99.79	99.67	99.90	99.77	97.72	99.98	99.85	99.89	99.73	100.11	99.81
Ilmenite wt%																			
SiO ₂	0.01	0.02	0.02	0.01	0.00	0.01	0.02	0.01	0.02	0.01	0.00	0.03	0.00	0.01	0.01	0.00	0.01	0.00	0.03
TiO ₂	44.65	44.73	44.90	44.44	44.34	45.65	47.08	45.25	46.39	45.27	44.94	45.19	46.43	48.46	47.47	47.90	47.24	46.90	47.87
Al ₂ O ₃	0.13	0.15	0.14	0.15	0.13	0.10	0.09	0.15	0.13	0.14	0.16	0.14	0.11	0.07	0.10	0.10	0.11	0.09	0.09
FeO(T)	51.65	51.73	51.51	51.91	51.78	50.84	49.61	51.05	50.17	51.47	51.37	51.48	50.75	48.48	49.48	49.27	50.17	50.20	49.10
MnO	0.72	0.71	0.72	0.67	0.68	0.79	0.80	0.68	0.77	0.70	0.68	0.69	0.75	0.86	0.75	0.82	0.77	0.82	0.84
MgO	1.10	1.09	1.06	1.04	1.06	1.15	1.21	1.19	1.28	1.14	1.07	1.11	1.19	1.32	1.26	1.25	1.25	1.24	1.26
CaO	0.01	0.00	0.01	0.02	0.06	0.04	0.09	0.01	0.02	0.07	0.04	0.02	0.07	0.07	0.10	0.05	0.01	0.01	0.08
Cr ₂ O ₃	0.02	0.02	0.01	0.02	0.03	0.03	0.00	0.06	0.02	0.00	0.00	0.03	0.00	0.02	0.01	0.02	0.03	0.00	0.05
NiO	0.00	0.04	0.00	0.01	0.01	0.04	0.00	0.04	0.01	0.00	0.01	0.00	0.03	0.00	0.01	0.04	0.03	0.00	0.00
Total	98.29	98.47	98.37	98.27	98.09	98.64	98.90	98.44	98.80	98.78	98.27	98.69	98.74	99.29	99.19	99.45	99.63	99.26	99.32
Recalculated after Carmichael (1967)																			
Fe ₂ O ₃ wt%	15.76	15.72	15.26	16.09	16.17	14.07	11.47	14.62	12.78	15.08	15.10	15.01	12.63	9.11	11.00	10.34	11.88	12.30	10.21
FeO wt%	37.47	37.59	37.78	37.43	37.23	38.18	39.29	37.89	38.67	37.90	37.78	37.98	38.79	40.29	39.58	39.97	39.48	39.13	39.91
Total	99.87	100.05	99.90	99.88	99.71	100.05	100.05	99.90	100.08	100.29	99.78	100.20	100.01	100.20	100.30	100.49	100.82	100.49	100.34
Giorso and Evans (2008)																			
Temperature (°C)	902	911	913	903	915	876	852	902	870	904	905	884	872	775	825	822	859	863	822
wt% FeO (NNO)	0.39	0.37	0.32	0.42	0.40	0.26	-0.07	0.27	0.10	0.31	0.32	0.35	0.07	-0.39	-0.10	-0.22	-0.02	0.04	-0.25

TABLE D.20: Composition and temperature and oxygen fugacity of magnetite-ilmenite pairs of the Ebisutoge PD

Sample:	EH63-4	EH63-4	EH63-4	EH63-3	EH63-3	EH63-3	EH63-3	EH63-3	EH65c	EH65c	EH65c
Magnetite wt%	mag1	mag2	mag3	63ii_mag4.1	mag4.2	mag4.3	mag4.4	mag4.5	mag.6.1cpl	mag.6.2cpl	mag.6.3cpl
SiO ₂	0.09	0.07	0.09	0.08	0.11	0.13	0.07	0.10	0.07	0.08	0.11
TiO ₂	16.82	16.81	16.63	15.88	16.90	17.91	17.00	16.02	16.87	16.83	17.09
Al ₂ O ₃	1.48	1.53	1.49	1.60	1.56	1.53	1.46	1.54	1.45	1.50	1.46
FeO(T)	77.25	76.93	77.18	76.93	76.03	74.51	76.80	77.54	76.83	76.66	76.60
MnO	0.70	0.71	0.74	0.70	0.76	0.73	0.71	0.71	0.71	0.74	0.73
MgO	0.33	0.33	0.33	0.34	0.32	0.33	0.34	0.33	0.36	0.33	0.32
CaO	0.01	0.00	0.01	0.01	0.00	0.00	0.01	0.02	0.00	0.02	0.01
Cr ₂ O ₃	0.00	0.00	0.02	0.04	0.01	0.00	0.00	0.02	0.04	0.04	0.02
NiO	0.02	0.00	0.03	0.04	0.00	0.00	0.00	0.00	0.00	0.01	0.00
Total	96.70	96.38	96.52	95.62	95.68	95.13	96.39	96.27	96.33	96.20	96.33
<i>Recalculated after Carmichael (1967)</i>											
FeO wt%	34.84	34.63	35.08	35.83	33.77	31.29	34.36	36.09	34.55	34.43	33.99
FeO wt%	45.90	45.77	45.62	44.69	45.64	46.36	45.88	45.07	45.74	45.68	46.02
Total	100.19	99.85	100.04	99.21	99.07	98.27	99.83	99.89	99.80	99.65	99.74
<i>Ilmenite</i>											
wt%	il1	il2	il3	il4.1	il4.2-4.3	il4.2-4.3	il4.4	il4.5	il6.1cpl	il6.2cpl	il6.2cpl
SiO ₂	0.02	0.01	0.00	0.02	0.00	0.00	0.00	0.03	0.01	0.01	0.01
TiO ₂	50.83	50.97	50.77	50.52	50.76	50.76	50.55	50.66	50.54	50.32	50.32
Al ₂ O ₃	0.06	0.06	0.04	0.05	0.05	0.05	0.05	0.06	0.07	0.08	0.08
FeO(T)	47.41	47.35	47.45	47.11	47.67	47.67	47.72	47.72	47.63	47.66	47.66
MnO	1.09	1.07	1.07	1.47	1.12	1.12	1.16	1.12	1.12	1.12	1.12
MgO	0.60	0.61	0.56	0.65	0.66	0.66	0.64	0.63	0.57	0.61	0.61
CaO	0.02	0.01	0.00	0.02	0.01	0.01	0.00	0.01	0.01	0.03	0.03
Cr ₂ O ₃	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.02	0.02	0.02
NiO	0.01	0.07	0.05	0.02	0.00	0.00	0.01	0.01	0.02	0.03	0.03
Total	100.05	100.16	99.94	99.87	100.27	100.27	100.13	100.24	99.98	99.88	99.88
<i>Recalculated after Carmichael (1967)</i>											
FeO wt%	4.29	4.10	4.28	4.80	4.81	4.81	5.06	4.86	4.80	5.16	5.16
FeO wt%	43.55	43.66	43.60	42.79	43.34	43.34	43.17	43.34	43.31	43.02	43.02
Total	100.48	100.57	100.37	100.35	100.75	100.75	100.63	100.73	100.46	100.40	100.40
<i>Chiono and Evans (2008)</i>											
Temperature (°C)	728	723	728	740	756	773	766	745	754	769	773
log ₁₀ fO ₂ (NNO)	-2.120	-2.190	-2.090	-1.880	-1.880	-1.900	-1.760	-1.810	-1.860	-1.710	-1.720

TABLE D.21: Composition of orthopyroxene of the Nyukawa PFD

Sample: ID:	EH61 opx_1	EH61 opx_1r	EH61 opx_2c	EH61 opx_2r	EH61 opx_3c	EH61 opx_3r	EH61 opx_4c	EH61 opx_4r	EH61 opx_5c	EH61 opx_5r	EH61 opx_6c	EH61 opx_6r	EH61 opx_7c	EH61 opx_7.2	EH61 opx_8	EH61 opx_9	EH61 opx_9r	EH61 opx_10c	EH61 opx_10r	EH61 opx_11c
wt%																				
SiO ₂	52.46	52.62	52.69	53.05	52.63	53.13	52.56	52.92	52.34	52.29	52.45	52.50	52.59	52.64	52.66	52.12	52.40	52.27	52.36	51.92
K ₂ O	0.01	0.02	0.00	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.01	0.00
Na ₂ O	0.02	0.01	0.02	0.04	0.02	0.02	0.00	0.00	0.01	0.04	0.00	0.04	0.01	0.00	0.03	0.00	0.01	0.00	0.01	0.02
CaO	1.05	1.15	1.00	1.01	1.02	1.04	1.00	1.04	1.05	1.04	1.02	1.06	1.04	1.01	0.87	1.06	1.12	1.07	1.08	1.04
FeO	24.98	24.73	24.98	25.00	25.05	24.47	24.96	24.85	25.65	24.69	25.35	24.51	25.03	25.40	24.93	25.29	24.68	25.12	24.15	25.47
Al ₂ O ₃	0.48	0.61	0.43	0.40	0.57	0.47	0.44	0.50	0.57	0.63	0.56	0.49	0.57	0.47	0.51	0.56	0.58	0.54	0.66	0.50
MgO	20.74	20.56	20.46	20.35	20.91	20.72	20.57	20.76	19.95	20.66	20.32	20.54	20.65	20.43	20.65	20.58	20.85	20.65	21.22	20.33
TiO ₂	0.11	0.19	0.13	0.12	0.13	0.10	0.09	0.13	0.13	0.12	0.12	0.11	0.15	0.10	0.10	0.09	0.15	0.10	0.13	0.14
MnO	0.90	0.88	0.97	0.90	0.93	0.93	0.94	0.93	1.01	0.83	0.97	0.94	0.91	1.00	1.06	0.92	0.88	0.90	0.85	0.94
Total	100.74	100.76	100.68	100.89	101.27	100.90	100.58	101.13	100.72	100.31	100.78	100.18	100.96	101.06	100.81	100.63	100.67	100.66	100.47	100.36
En	0.58	0.58	0.58	0.58	0.59	0.59	0.58	0.59	0.57	0.59	0.58	0.59	0.58	0.58	0.59	0.58	0.59	0.58	0.60	0.57
Fs	0.39	0.39	0.40	0.40	0.39	0.39	0.40	0.39	0.41	0.39	0.40	0.39	0.40	0.40	0.40	0.40	0.39	0.40	0.38	0.40
Wo	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Mg#	0.60	0.60	0.59	0.59	0.60	0.60	0.59	0.60	0.58	0.60	0.59	0.60	0.60	0.59	0.60	0.59	0.60	0.59	0.61	0.59

Sample: ID:	EH61 opx_11r	EH62 opx_1	EH62 opx_2c	EH62 opx_3c	EH62 opx_4c	EH62 opx_4r	EH62 opx_5c	EH62 opx_5r	EH62 opx_6c	EH62 opx_6r
wt%										
SiO ₂	51.89	52.79	52.45	52.47	52.12	52.70	52.27	52.61	52.07	52.42
K ₂ O	0.01	0.00	0.01	0.00	0.01	0.01	0.00	0.02	0.00	0.00
Na ₂ O	0.02	0.02	0.02	0.05	0.02	0.04	0.00	0.01	0.00	0.01
CaO	1.12	0.99	1.04	1.01	0.99	1.05	1.00	1.05	0.98	1.07
FeO	24.82	24.88	24.87	24.95	25.06	24.49	25.19	24.67	25.17	24.79
Al ₂ O ₃	0.48	0.42	0.48	0.42	0.42	0.44	0.42	0.45	0.54	0.47
MgO	20.71	20.40	20.41	20.43	20.65	20.98	20.65	20.60	20.22	20.75
TiO ₂	0.12	0.13	0.09	0.09	0.10	0.11	0.13	0.11	0.13	0.14
MnO	0.88	0.89	0.89	0.97	0.90	0.89	0.92	0.86	0.88	0.95
Total	100.04	100.52	100.26	100.39	100.28	100.71	100.58	100.40	99.99	100.60
En	0.58	0.58	0.58	0.58	0.58	0.59	0.58	0.59	0.58	0.59
Fs	0.39	0.40	0.40	0.40	0.40	0.39	0.40	0.39	0.40	0.39
Wo	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Mg#	0.60	0.59	0.59	0.59	0.59	0.60	0.59	0.60	0.59	0.60

TABLE D.22: Composition of orthopyroxene of the Ebisutoge PD

Sample:	EH65b	EH65b	EH65b	EH65b	EH65b	EH65c	EH65c	EH65c	EH65c	EH65c
ID:	opx_4	opx_5	opx_6	opx_7	opx_11	opx_5	opx_6	opx_9	opx_10	
<i>wtf%</i>										
SiO ₂	49.55	49.32	49.57	49.20	49.22	49.29	49.93	48.72	49.22	
K ₂ O	0.01	0.01	0.02	0.00	0.01	0.02	0.03	0.01	0.00	
Na ₂ O	0.01	0.02	0.02	0.04	0.00	0.04	0.03	0.04	0.02	
CaO	1.07	1.19	1.18	1.14	1.19	1.22	1.03	1.25	1.06	
FeO	38.55	39.75	40.04	39.86	39.83	39.51	38.04	39.25	37.85	
Al ₂ O ₃	0.37	0.28	0.18	0.27	0.18	0.30	0.32	0.33	0.39	
MgO	9.90	8.56	8.76	8.61	8.88	8.75	10.01	8.88	10.91	
TiO ₂	0.10	0.12	0.04	0.10	0.07	0.12	0.10	0.08	0.10	
MnO	1.50	1.72	1.57	1.67	1.55	1.73	1.49	1.57	1.45	
Total	101.07	100.98	101.37	100.89	100.94	100.99	100.98	100.13	101.00	
En	0.31	0.27	0.27	0.27	0.28	0.28	0.31	0.28	0.28	
Fs	0.67	0.70	0.70	0.70	0.70	0.70	0.66	0.69	0.70	
Wo	0.02	0.03	0.03	0.03	0.03	0.03	0.02	0.03	0.03	
Mg#	0.31	0.28	0.28	0.28	0.28	0.28	0.32	0.29	0.29	
Sample:	EH65c	EH65c	EH65c	EH65c	EH65b	EH65b	EH65b	EH65b	EH65b	
ID:	opx_11	opx_16	opx_17	opx_8	opx_1	opx_2	opx_8	opx_9		
<i>wtf%</i>										
SiO ₂	48.83	50.04	49.46	52.64	52.87	52.51	52.62	52.69		
K ₂ O	0.02	0.00	0.00	0.00	0.01	0.04	0.01	0.02		
Na ₂ O	0.03	0.00	0.01	0.00	0.01	0.00	0.01	0.00		
CaO	1.14	1.00	1.21	1.02	1.36	1.15	1.04	1.14		
FeO	39.87	37.44	39.78	24.96	16.17	21.21	25.29	24.35		
Al ₂ O ₃	0.22	0.26	0.30	0.55	2.38	1.53	0.49	0.53		
MgO	8.95	11.22	8.69	20.57	26.29	22.36	20.07	20.50		
TiO ₂	0.10	0.11	0.09	0.11	0.31	0.30	0.14	0.14		
MnO	1.65	1.41	1.75	0.95	0.41	0.66	0.95	0.89		
Total	100.81	101.48	101.29	100.81	99.81	99.77	100.62	100.27		
En	0.28	0.34	0.27	0.58	0.72	0.64	0.57	0.59		
Fs	0.70	0.64	0.70	0.40	0.25	0.34	0.41	0.39		
Wo	0.03	0.02	0.03	0.02	0.03	0.02	0.02	0.02		
Mg#	0.29	0.35	0.28	0.59	0.74	0.65	0.59	0.60		

TABLE D.23: Composition of amphibole inclusion in orthopyroxene in Ebisutoge PD

Sample: ID:	EH65c am_1	EH65c am_2	EH65c am_3	EH65c am_4	EH65c am_5
<i>wt%</i>					
SiO ₂	43.52	43.41	44.15	44.39	44.04
TiO ₂	1.86	1.84	1.80	1.78	1.81
Al ₂ O ₃	8.54	8.95	8.16	8.01	8.09
FeO	23.74	23.84	23.73	22.86	23.25
MnO	0.46	0.50	0.51	0.46	0.47
MgO	7.65	7.52	8.02	8.53	8.28
CaO	9.88	9.94	9.76	10.02	9.96
Na ₂ O	1.94	2.06	1.96	2.04	1.99
K ₂ O	0.72	0.75	0.65	0.74	0.75
Total	98.31	98.81	98.74	98.83	98.64
Mg#	0.36	0.36	0.38	0.40	0.39

TABLE D.24: Trace element concentration of plagioclase phenocrysts in the Nyukawa PFD

Sample: ID:	EH61 pl1.1	EH61 pl1.2	EH61 pl1.3	EH61 pl1.4	EH61 pl1.5	EH61 pl2.1	EH61 pl2.2	EH61 pl2.3	EH61 pl2.4	EH61 pl2.5	EH61 pl2.6	EH61 pl3.1	EH61 pl3.2	EH61 pl3.3	EH61 pl3.4	EH61 pl3.5
<i>ppm</i>																
Sc	<0.101	0.109	0.265	0.210	0.240	<0.288	0.272	0.284	0.226	0.244	0.292	0.324	0.256	0.492	0.301	0.291
Rb	0.718	0.797	0.816	0.721	0.713	4.122	2.727	1.804	0.709	0.661	0.735	1.105	0.826	0.653	0.862	0.775
Sr	824	840	899	842	871	803	815	923	870	843	850	891	815	819	848	869
Y	0.230	0.198	0.211	0.342	0.371	0.343	0.307	0.242	1.617	0.623	0.350	0.260	0.281	0.623	0.424	0.386
Zr	0.027	0.056	0.087	0.356	0.026	<0.062	0.139	0.099	0.020	0.057	0.072	0.084	0.024	0.566	<0.015	<0.011
Nb	0.005	<0.002	<0.002	0.003	0.014	0.052	0.017	0.007	<0.002	0.005	0.003	0.007	0.011	0.177	0.007	0.002
Cs	<0.014	<0.014	<0.026	<0.018	<0.022	<0.067	0.019	<0.017	<0.012	<0.011	<0.013	<0.017	<0.014	<0.033	<0.014	<0.016
Ba	314	352	361	333	362	360	365	406	343	351	375	383	341	298	377	377
La	10.4	10.8	9.5	11.1	11.0	11.9	13.1	13.4	12.9	10.7	11.5	10.5	10.2	11.6	11.8	11.6
Ce	13.6	14.4	12.7	15.0	16.0	14.5	16.3	16.8	19.5	15.5	16.0	13.4	13.6	17.5	16.3	16.3
Pt	1.01	1.05	0.93	1.20	1.26	1.05	1.19	1.20	1.71	1.25	1.23	0.99	1.02	1.41	1.32	1.29
Nd	2.84	2.78	2.51	3.61	3.88	2.99	3.38	3.19	5.69	3.80	3.72	2.86	2.93	4.16	3.81	3.87
Sm	0.240	0.236	0.233	0.343	0.409	0.257	0.291	0.281	0.696	0.427	0.384	0.239	0.291	0.472	0.402	0.438
Eu	1.45	1.47	1.63	1.70	1.86	1.59	1.68	1.87	1.90	1.81	1.84	1.82	1.64	1.66	1.86	1.92
Gd	0.116	0.102	0.161	0.198	0.222	0.135	0.138	0.118	0.537	0.269	0.202	0.127	0.141	0.263	0.198	0.224
Tb	0.012	0.012	0.010	<0.068	0.022	0.022	0.016	0.014	0.059	0.029	0.019	0.014	0.015	0.026	0.021	0.019
Dy	0.050	0.085	0.049	0.086	0.103	0.076	0.069	0.060	0.328	0.160	0.113	0.070	0.063	0.137	0.110	0.089
Ho	0.011	0.008	0.009	0.013	0.017	0.013	0.019	0.008	0.055	0.022	0.015	0.010	0.011	0.021	0.017	0.020
Er	0.019	0.017	0.019	0.028	0.033	0.034	0.029	0.023	0.145	0.058	0.028	0.020	0.030	0.062	0.035	0.028
Tm	0.003	0.002	0.002	0.003	0.003	0.003	0.003	0.002	0.019	0.006	0.005	0.003	0.006	0.008	0.004	0.003
Yb	0.013	0.010	0.009	0.011	0.018	0.026	0.022	0.029	0.127	0.036	0.021	0.016	0.055	0.026	0.022	0.046
Lu	0.002	0.002	0.002	0.003	0.001	0.004	0.004	0.002	0.017	0.004	0.002	<0.002	0.002	0.005	0.002	0.003
Pb	8.14	8.20	8.54	7.63	7.80	6.87	7.54	8.37	7.80	7.35	7.05	8.29	7.51	7.60	8.09	7.80
Th	0.003	0.005	0.001	0.001	0.007	0.033	0.032	0.013	0.078	0.015	0.001	0.008	0.013	0.023	0.007	0.004
U	0.002	0	0	0	0.001	0.006	0.008	0.003	0.017	0.004	0.001	0.002	0.004	0.005	0.004	0.001
An	0.44	0.43	0.46	0.46	0.45	0.40	0.39	0.44	0.46	0.45	0.44	0.45	0.44	0.51	0.45	0.46

TABLE D.25: Trace element concentration of plagioclase phenocrysts in the Ebisutoge PD

Unit: Sample: ID:	Unit B EH63 pl1.1	Unit B EH63 pl1.2	Unit B EH63 pl1.3	Unit B EH63 pl1.4	Unit B EH63 pl1.5	Unit B EH63 pl1.6	Unit B EH63 pl1.7	Unit B EH63 pl1.8	Unit B EH63 pl2.1	Unit B EH63 pl2.2	Unit B EH63 pl2.3	Unit B EH63 pl2.4
<i>ppm</i>												
Sc	<0.075	0.247	0.236	0.269	0.248	0.286	0.258	0.258	0.314	0.276	0.316	0.311
Rb	0.786	0.849	0.839	0.862	0.65	0.549	0.449	0.665	0.782	0.843	0.764	0.78
Sr	907	887	884	889	1109	1127	1030	822	918	913	978	848
Y	0.161	0.159	0.159	0.15	0.157	0.163	0.154	0.166	0.167	0.151	0.189	0.162
Zr	0.014	<0.015	<0.009	<0.085	<0.011	0.034	<0.008	0.06	<0.008	<0.013	0.049	0.054
Nb	<0.001	0.003	0.008	<0.002	<0.002	0.003	0.002	<0.073	<0.001	<0.001	0.002	<0.002
Cs	<0.011	<0.011	<0.012	<0.010	<0.012	<0.015	<0.016	<0.043	<0.015	<0.014	<0.011	<0.012
Ba	542	553	526	532	361	300	255	389	480	512	490	479
La	9.3	9.3	8.8	8.3	6.5	5.8	5.2	7.3	9.5	9.6	9.9	8.4
Ce	11.8	11.9	11.4	11.1	9.0	8.0	7.4	9.5	11.8	12.2	13.0	11.0
Pr	0.922	0.917	0.879	0.855	0.727	0.653	0.608	0.76	0.908	0.914	0.987	0.839
Nd	2.68	2.66	2.55	2.50	2.21	2.09	1.93	2.38	2.63	2.72	2.91	2.52
Sm	0.267	0.259	0.246	0.228	0.22	0.198	0.189	0.215	0.234	0.224	0.258	0.227
Eu	4.00	4.14	4.09	4.12	4.43	4.29	3.97	3.59	3.34	3.56	3.77	3.82
Gd	0.113	0.139	0.141	0.119	0.112	0.105	0.103	0.144	0.102	0.109	0.113	0.122
Tb	0.011	0.009	0.010	0.011	0.012	0.010	0.009	0.014	0.013	0.009	0.012	0.012
Dy	0.057	0.069	0.050	0.046	0.042	0.055	0.046	0.058	0.046	0.050	0.053	0.049
Ho	0.009	0.008	0.007	0.009	0.005	0.006	0.007	0.008	0.006	0.007	0.008	0.005
Er	0.014	0.017	0.014	0.018	0.018	0.014	0.013	0.021	0.012	0.014	0.017	0.015
Tm	0.001	0.002	0.001	<0.001	0.001	0.002	0.002	<0.001	0.001	0.001	0.008	0.001
Yb	0.007	0.007	0.043	0.005	0.004	0.007	0.007	<0.012	<0.006	0.004	0.006	0.004
Lu	<0.001	<0.001	0.001	0.001	0.001	0.001	0.001	<0.002	0.001	0.001	0.001	0.001
Pb	8.682	8.961	8.611	8.414	6.192	5.279	4.650	5.800	7.905	7.792	7.865	7.049
Th	0.001	0	0.001	0	0	0	0	0.004	0	0.001	0	0
U	0	0	0	0	0	0	0	0.001	0	0	0.001	0.001
An	0.30	0.29	0.30	0.29	0.35	0.36	0.34	0.30	0.31	0.30	0.34	0.30
Unit: Sample: ID:	Unit C EH65a pl1.1	Unit C EH65a pl1.2	Unit C EH65a pl1.3	Unit C EH65a pl1.4	Unit C EH65a pl1.5	Unit C EH65a pl1.6	Unit C EH65a pl1.7	Unit C EH65a pl2.2	Unit C EH65a pl2.3	Unit C EH65a pl2.4	Unit C EH65a pl2.6	Unit C EH65a pl3.1
<i>ppm</i>												
Sc	<0.069	0.313	0.229	0.293	0.386	0.369	0.449	0.386	0.322	0.335	<0.104	<0.581
Rb	0.393	0.985	0.412	0.45	0.431	0.468	0.722	0.904	0.837	0.799	0.836	0.981
Sr	1209	1284	1284	1304	1221	1163	970	922	858	852	861	842
Y	0.185	0.279	0.188	0.179	0.174	0.17	0.185	0.167	0.15	0.161	0.16	0.236
Zr	0.021	2.07	0.076	0.014	0.056	0.016	<0.035	0.031	0.038	<0.010	0.024	0.49
Nb	<0.003	0.069	0.018	0.009	<0.002	<0.002	<0.007	<0.002	<0.002	<0.002	<0.003	0.044
Cs	<0.012	<0.064	<0.012	<0.013	<0.016	<0.016	<0.057	<0.018	<0.020	<0.013	<0.013	<0.114
Ba	226	265	244	253	278	301	413	543	515	492	518	532
La	5.0	5.1	5.2	5.1	5.9	6.2	8.1	9.3	8.5	8.3	8.5	8.2
Ce	7.1	7.4	7.6	7.3	8.4	8.6	10.4	11.7	11.1	10.9	11.2	10.1
Pr	0.606	0.614	0.64	0.617	0.686	0.699	0.834	0.89	0.843	0.839	0.843	0.798
Nd	1.99	1.94	2.04	1.88	2.22	2.17	2.50	2.67	2.45	2.49	2.48	2.28
Sm	0.214	0.199	0.201	0.218	0.215	0.206	0.277	0.258	0.223	0.238	0.226	0.224
Eu	3.87	3.78	4.13	4.15	4.32	4.31	3.91	4.09	3.95	3.99	3.86	3.52
Gd	0.123	0.152	0.114	0.113	0.115	0.117	0.116	0.123	0.114	0.113	0.114	0.574
Tb	0.011	0.014	0.011	0.011	0.011	0.014	0.014	0.010	0.010	0.011	0.010	0.012
Dy	0.050	0.063	0.048	0.055	0.049	0.041	0.053	0.049	0.049	0.041	0.040	0.020
Ho	0.007	0.011	0.008	0.006	0.008	0.007	0.007	0.007	0.006	0.013	0.006	0.012
Er	0.017	0.021	0.017	0.015	0.016	0.014	0.015	0.014	0.012	0.013	0.014	0.018
Tm	0.002	0.004	0.002	0.002	0.002	0.002	<0.002	0.002	0.001	0.002	0.001	0.006
Yb	0.010	0.017	0.010	0.007	0.006	0.011	0.012	0.006	0.004	0.009	0.004	0.004
Lu	0.001	0.004	0.001	0.002	0.001	<0.001	<0.003	0.005	0.007	0.001	<0.001	0.006
Pb	3.917	3.665	3.830	3.661	4.305	4.604	5.937	7.950	7.101	6.942	7.447	6.124
Th	0.002	0.043	0.008	0.002	0.001	0	<0.002	0.003	0.001	<0.001	0.006	0.030
U	0.001	0.019	0.008	0.003	0.001	0	0	0	0	0	0.007	0.012
An	0.40	0.40	0.40	0.40	0.39	0.36	0.33	0.30	0.29	0.29	0.29	0.28

TABLE D.26: (Continued)

Unit: Sample: ID:	Unit C EH65a pl3.2	Unit C EH65a pl3.3	Unit C EH65a pl3.4	Unit C EH65a pl3.5	Unit C EH65a pl3.6	Unit B EH65a pl3.7	Unit D EH65b pl1.11	Unit D EH65b pl1.12	Unit D EH65b pl1.13	Unit D EH65b pl1.14	Unit D EH65b pl1.16
<i>ppm</i>											
Sc	<0.535	<0.473	0.193	0.319	0.257	<0.534	<0.250	<0.289	<0.255	<0.253	0.292
Rb	0.479	0.476	0.385	0.569	0.503	0.583	0.828	0.766	0.809	0.773	0.867
Sr	1334	1137	1067	1217	1058	1117	856	870	877	860	857
Y	0.188	0.17	0.175	0.164	0.171	0.147	0.184	0.159	0.175	0.164	0.55
Zr	<0.077	0.352	0.045	0.017	<0.013	<0.094	<0.032	<0.038	<0.063	0.052	0.323
Nb	<0.019	<0.014	0.01	<0.003	<0.003	<0.014	<0.006	<0.006	<0.007	<0.006	0.049
Cs	<0.114	<0.115	<0.018	<0.015	<0.012	<0.130	<0.049	<0.028	<0.048	<0.038	<0.012
Ba	319	312	217	325	274	319	494	510	499	495	544
La	5.6	5.8	4.8	6.2	5.6	5.8	8.8	8.7	8.5	8.5	9.0
Ce	7.5	7.7	6.6	8.3	7.8	7.6	11.3	11.1	10.7	11.0	13.1
Pr	0.625	0.616	0.559	0.676	0.633	0.638	0.897	0.839	0.849	0.855	0.927
Nd	1.96	1.96	1.77	2.14	1.97	1.92	2.55	2.61	2.51	2.44	2.79
Sm	0.247	0.208	0.167	0.192	0.205	0.191	0.244	0.235	0.234	0.198	0.317
Eu	3.84	3.86	3.23	4.19	3.74	3.58	3.65	3.88	3.75	3.68	3.97
Gd	0.156	0.459	0.102	0.113	0.107	0.072	0.127	0.123	0.225	0.154	0.182
Tb	0.009	0.012	0.012	0.010	0.011	0.010	0.010	0.013	0.012	0.012	0.022
Dy	0.053	0.080	0.079	0.049	0.050	0.045	0.043	0.048	0.055	0.047	0.132
Ho	0.032	0.008	0.006	0.008	0.007	<0.005	0.012	0.005	0.005	0.006	0.026
Er	0.086	0.017	0.016	0.015	0.018	0.011	0.016	0.018	0.018	0.019	0.067
Tm	<0.005	<0.005	0.001	0.002	0.003	<0.003	0.004	<0.003	0.004	0.003	0.011
Yb	0.539	<0.017	0.009	0.007	0.008	0.462	0.010	0.009	0.005	<0.009	0.080
Lu	0.006	<0.005	0.007	0.002	0.001	<0.007	<0.003	<0.002	<0.003	<0.002	0.013
Pb	3.978	4.248	3.675	4.957	4.751	4.257	6.998	6.666	6.351	6.502	7.622
Th	0.003	<0.003	0	0	0.008	<0.005	<0.002	<0.001	0.002	<0.003	0.032
U	0.011	0.001	0	0	0	0	0	0	0	0.001	0.027
An	0.39	0.35	0.35	0.38	0.35	0.35	0.30	0.30	0.30	0.30	0.29

TABLE D.27: Trace element concentration of plagioclase phenocrysts in TK

Unit: Sample: ID:	pGT EH70 pl2.1	pGT EH70 pl2.2	pGT EH70 pl2.3	pGT EH70 pl2.4	pGT EH70 pl2.5	pGT EH70 pl2.6	pGT EH70 pl2.7	pGT EH70 pl3.1	pGT EH70 pl3.2	pGT EH70 pl3.3	pGT EH70 pl3.4	pGT EH70 pl3.5
<i>ppm</i>												
Sc	<0.518	<0.365	<0.388	<0.381	<0.460	<0.486	<0.676	<1.081	<0.658	<0.282	<0.457	<0.680
Rb	0.598	0.602	0.584	0.809	0.908	1.311	0.539	0.78	7.88	0.532	0.839	0.884
Sr	915	826	843	877	846	679	794	827	854	816	866	920
Y	0.262	0.341	0.289	0.25	0.192	0.144	0.344	0.752	1.924	0.313	0.219	0.31
Zr	<3.543	<0.061	<0.038	0.206	<0.054	<0.069	<0.053	0.667	3.346	0.198	<0.100	<0.107
Nb	<0.012	<0.010	<0.010	<0.216	<0.016	<0.019	<0.018	0.122	0.273	0.061	<0.014	<0.017
Cs	<0.063	<0.048	<0.044	<0.055	<0.049	<0.072	<0.065	<0.211	9.394	0.038	<0.090	<0.079
Ba	316	283	304	361	336	269	248	103	122	257	326	281
La	11.0	10.4	11.7	11.8	11.4	11.2	8.9	3.8	2.8	9.1	11.7	11.8
Ce	14.5	13.4	15.3	14.7	13.6	12.3	11.9	6.6	5.1	11.6	14.1	14.5
Pr	1.064	1.04	1.176	1.068	0.897	0.804	0.944	0.647	0.544	0.883	0.994	1.016
Nd	3.20	2.95	3.49	3.06	2.62	1.95	2.76	2.39	2.05	2.57	2.57	2.80
Sm	0.335	0.353	0.335	0.259	0.23	0.188	0.393	0.421	0.467	0.266	0.224	0.351
Eu	1.64	1.52	1.53	1.57	1.39	1.48	1.42	0.73	0.65	1.33	1.38	1.37
Gd	0.188	0.144	0.19	0.209	0.134	0.09	0.167	0.372	0.38	0.151	0.132	0.214
Tb	0.059	0.021	0.02	0.016	0.009	0.009	0.014	0.024	0.063	0.011	0.016	0.014
Dy	0.074	0.075	0.084	0.049	0.052	<0.057	0.109	0.11	0.324	0.067	0.027	0.072
Ho	0.013	0.069	<0.115	0.014	0.012	0.005	0.014	0.017	0.076	0.014	<0.008	0.042
Er	0.047	0.02	0.039	0.014	0.009	0.019	0.027	0.067	0.175	0.106	0.238	0.017
Tm	0.005	0.061	0.005	<0.005	<0.002	0.006	0.04	0.005	0.027	0.018	<0.003	0.057
Yb	<0.014	0.025	<0.009	0.14	<0.013	<0.011	<0.033	0.118	0.176	0.015	0.04	0.025
Lu	0.004	<0.008	<0.002	<0.004	<0.002	<0.004	<0.008	<0.008	0.028	<0.002	<0.007	<0.006
Pb	8.51	7.46	7.55	8.32	8.48	10.75	6.81	5.93	7.02	8.31	9.00	11.03
Th	<0.002	0.004	0.056	<0.002	<0.004	0.006	0.019	0.052	0.077	0.006	<0.003	0.005
U	0	0.001	0	0	0.001	0.005	<0.002	0.004	0.009	0.002	0.001	<0.170
An	0.51	0.47	0.45	0.44	0.40	0.30	0.46	0.82	0.82	0.44	0.41	0.46
Unit: Sample: ID:	pGT EH70 pl3.6	pGT EH70 pl4.1	pGT EH70 pl4.2	pGT EH70 pl4.3	pGT EH70 pl4.4	pGT EH70 pl4.6	pGT EH70 pl4.7	GDT EH16 pl1.1	GDT EH16 pl1.2	GDT EH16 pl1.3	GDT EH16 pl1.4	GDT EH16 pl1.5
<i>ppm</i>												
Sc	<0.288	<0.277	<0.334	<0.233	<1.050	<0.909	<0.531	<0.542	<0.358	<0.326	<0.305	<0.246
Rb	1.23	0.867	0.523	1.006	1.307	1.824	0.933	0.307	0.587	0.711	1.054	3.243
Sr	726	839	857	854	242	204	387	801	859	969	961	916
Y	0.191	0.166	0.311	0.148	<0.180	0.21	0.2	0.219	0.289	0.329	0.248	0.19
Zr	0.528	<0.077	0.271	<0.027	<0.165	<0.138	0.082	<0.082	0.042	<0.033	<0.045	<0.045
Nb	<0.149	<0.007	<0.007	<0.007	<0.033	<0.029	<0.019	<0.015	<0.011	<0.009	<0.007	<0.009
Cs	<0.041	<0.048	<0.042	<0.035	<0.159	<0.135	0.129	<0.058	<0.039	<0.038	0.047	0.087
Ba	331	269	281	232	38	32	182	228	302	357	414	413
La	14.0	12.5	11.0	11.0	7.9	19.8	13.8	9.4	10.4	12.2	12.5	11.5
Ce	15.3	13.9	13.7	12.7	9.2	21.8	13.8	11.9	13.1	16.4	15.4	14.4
Pr	0.981	0.992	1.035	0.848	0.611	1.289	0.896	0.884	0.937	1.238	1.079	1.022
Nd	2.46	2.75	2.82	2.17	1.65	2.83	2.32	2.25	2.54	3.39	3.03	2.64
Sm	0.178	0.194	0.302	0.179	0.297	0.333	0.189	0.18	0.271	0.389	0.33	0.212
Eu	1.75	1.47	1.62	1.31	0.64	0.75	1.08	1.21	1.52	1.68	1.67	1.47
Gd	0.102	0.293	0.167	0.349	0.142	<0.109	0.1	0.073	0.171	0.152	0.151	0.12
Tb	0.045	0.026	0.017	<0.009	<0.014	0.017	<0.005	0.01	0.012	0.016	0.071	0.012
Dy	0.023	0.124	0.081	0.083	0.017	0.064	0.075	0.047	0.06	0.062	0.05	0.052
Ho	0.007	0.006	0.035	0.01	0.01	<0.003	0.018	0.014	0.011	0.014	0.011	0.006
Er	0.17	0.014	0.028	<0.047	<0.039	<0.010	<0.018	<0.008	0.023	0.028	0.027	0.022
Tm	<0.002	<0.003	0.003	0.031	<0.004	0.002	<0.003	0.005	0.004	0.004	<0.002	0.002
Yb	0.218	0.011	0.013	0.262	0.021	0	0	0	<0.018	<0.016	0.218	0.014
Lu	<0.004	<0.004	<0.002	<0.003	<0.011	0.009	0.083	<0.006	<0.002	<0.005	0.012	<0.004
Pb	13.06	7.59	8.02	9.58	10.89	13.49	11.47	5.72	6.81	7.89	9.04	10.06
Th	0.022	0.011	0.005	0	<0.007	<0.002	0.013	0.011	0.006	0.001	<0.001	<0.002
U	0.001	0.004	0.001	0	<0.003	0	<0.002	0.001	0.001	<0.001	0	0
An	0.30	0.42	0.46	0.39	0.23	0.23	0.23	0.44	0.43	0.50	0.46	0.41

TABLE D.28: (Continued)

Unit: Sample: ID:	GDT EH16 pl1.6	GDT EH16 pl1.7	GDT EH16 pl3.1	GDT EH16 pl3.2	GDT EH16 pl3.3	GDT EH16 pl3.4	GDT EH16 pl3.5	GDT EH16 pl3.6	GDT EH16 pl3.7	GDT EH16 pl3.8	mGT EH5 pl2.1	mGT EH5 pl2.2
<i>ppm</i>												
Sc	<0.248	<0.523	<1.230	<1.235	<1.857	2.932	<0.399	<0.321	<0.337	<0.374	<0.519	<0.345
Rb	0.955	4.101	0.41	19.018	<0.201	0.881	0.856	0.73	1.661	3.736	1.23	0.829
Sr	898	293	734	763	835	890	1015	805	768	149	814	813
Y	0.174	0.203	0.261	0.428	0.406	0.832	0.248	0.157	0.183	0.074	0.44	0.17
Zr	<0.035	<0.096	<0.231	<0.194	<0.140	0.309	<0.048	<0.035	0.232	<0.073	<0.079	0.086
Nb	<0.007	0.071	0.025	<0.035	<0.027	<0.044	<0.009	0.016	<0.006	<0.012	<0.013	<0.172
Cs	<0.042	0.081	<0.170	4.527	<0.116	<0.213	<0.043	<0.027	0.075	0.135	0.076	0.058
Ba	392	57	43	78	131	158	365	315	277	44	341	308
La	12.0	24.6	0.9	2.2	5.3	5.7	12.3	10.7	11.7	10.2	8.6	9.6
Ce	14.3	25.8	1.6	3.8	8.1	9.2	15.8	13.0	14.2	11.2	10.4	11.3
Pr	0.975	1.568	0.172	0.35	0.744	0.737	1.143	0.913	0.976	0.643	0.855	0.836
Nd	2.37	3.47	0.75	1.30	3.23	3.58	3.23	2.30	2.50	1.29	2.27	2.24
Sm	0.208	0.233	0.163	0.27	0.304	0.284	0.571	0.201	0.321	0.301	0.264	0.168
Eu	1.46	1.04	0.39	0.64	1.09	1.50	1.56	1.29	1.38	0.63	1.82	1.72
Gd	0.121	0.174	0.13	0.675	0.102	0.221	0.143	0.097	0.099	0.061	0.128	0.105
Tb	0.011	0.01	<0.008	<0.019	0.019	0.035	0.015	0.009	0.006	0.001	0.01	0.008
Dy	0.157	0.02	0.229	0.089	0.073	0.121	0.059	0.27	0.05	0.023	0.077	0.046
Ho	0.004	0.005	<0.009	0.011	0.015	0.037	0.009	0.007	0.008	0.006	0.015	0.01
Er	0.011	0.021	0.539	<0.055	0.037	0.139	0.017	0.01	0.016	0.014	0.028	<0.011
Tm	<0.002	0.081	<0.015	0.008	<0.010	<0.012	<0.002	0.003	<0.003	0.003	<0.003	0.004
Yb	0.01	0.021	<0.080	0.027	<0.059	0.223	<0.013	0.015	0.009	0	0.031	<0.012
Lu	<0.003	<0.004	<0.010	<0.019	<0.011	<0.014	<0.004	<0.003	<0.003	<0.003	0.005	<0.005
Pb	8.63	19.96	4.20	4.67	5.53	5.80	8.30	7.23	10.78	11.47	11.09	10.60
Th	<0.001	0.009	0.168	0.031	0.006	0.079	0.005	0.024	0.029	0.001	0.011	0.003
U	0	0.001	0	0.012	0.001	0.011	0	0.02	0.017	0.002	0.004	0
An	0.42	0.38	0.86	0.74	0.67	0.68	0.51	0.38	0.42	0.18	0.32	0.32
Unit: Sample: ID:	mGT EH5 pl2.3	mGT EH5 pl2.4	mGT EH5 pl2.5	mGT EH5 pl2.6	mGT EH5 pl2.7	mGT EH5 pl2.8	mGT EH5 pl2.9	mGT EH5 pl2.10	mGT EH5 pl2.11	mGT EH5 pl3.1	mGT EH5 pl3.2	mGT EH5 pl3.3
<i>ppm</i>												
Sc	<0.617	<0.370	<0.379	<0.554	<0.380	<0.529	0.106	0.175	0.231	0.132	0.19	0.178
Rb	1.114	1.301	0.875	0.737	0.681	1.034	0.825	1.141	1.795	0.839	0.899	0.921
Sr	829	873	896	907	808	236	872	796	214	869	884	845
Y	0.176	1.736	0.147	0.229	0.187	0.532	0.152	0.209	0.416	0.152	0.158	0.168
Zr	<0.071	<0.042	<0.053	0.159	<0.043	<0.118	<0.022	0.062	<0.016	<0.013	<0.009	<0.007
Nb	<0.011	<0.011	<0.008	<0.013	<0.008	0.015	<0.003	0.005	0.004	<0.002	<0.002	<0.001
Cs	0.063	0.197	<0.039	<0.065	<0.033	0.173	0.028	0.015	0.105	0.016	<0.009	<0.009
Ba	285	296	298	237	281	53	285	341	45	289	264	250
La	11.6	14.9	13.2	13.0	13.3	9.2	13.1	8.1	6.4	13.2	13.0	12.8
Ce	13.7	18.8	15.3	15.7	15.1	9.9	15.7	10.4	8.1	15.2	15.2	15.1
Pr	0.952	1.576	1.011	1.095	1.017	0.661	1.009	0.804	0.589	1.079	1.021	1.008
Nd	2.41	4.71	2.52	2.87	2.57	1.94	2.63	2.30	1.57	2.79	2.52	2.51
Sm	0.244	0.549	0.189	0.222	0.196	0.201	0.213	0.197	0.148	0.213	0.194	0.199
Eu	1.84	1.73	1.76	1.63	1.71	0.78	1.74	1.89	0.82	1.69	1.60	1.58
Gd	0.65	0.457	0.087	0.125	0.136	0.156	0.102	0.096	0.087	0.095	0.092	0.086
Tb	0.109	0.044	0.013	0.011	0.011	0.012	0.011	0.012	0.009	0.008	0.012	0.007
Dy	0.029	0.239	0.044	0.107	0.066	0.065	0.041	0.055	0.055	0.035	0.057	0.036
Ho	<0.003	0.051	0.004	0.007	0.034	0.017	0.008	0.014	0.012	0.007	0.007	0.005
Er	0.025	0.144	0.007	0.04	0.022	0.035	0.012	0.02	0.026	0.014	0.015	0.015
Tm	<0.003	0.018	0.003	0.129	0.006	0.022	0.002	0.002	0.004	0.001	0.001	0.002
Yb	<0.013	0.095	<0.012	<0.017	0.011	<0.028	0.014	0.018	0.017	0.01	0.009	0.011
Lu	<0.005	0.01	<0.002	<0.003	0.003	0.005	<0.002	<0.001	0.003	0.001	0.007	0.001
Pb	11.29	11.74	10.85	11.83	11.70	8.92	11.17	10.99	10.68	11.79	11.33	11.43
Th	<0.002	0.05	<0.001	0.015	0.002	0.02	0.001	0.012	0.006	0.001	0.001	0.001
U	<0.001	0.028	0	0.001	0.043	0.016	0.001	0.003	0.006	0.001	<0.000	0
An	0.34	0.35	0.34	0.41	0.36	0.21	0.34	0.32	0.21	0.35	0.35	0.36

TABLE D.29: (Continued)

Unit: Sample: ID:	mGT EH5 pl3.4	mGT EH5 pl3.5	GRT EH29 pl1.1	GRT EH29 pl1.2	GRT EH29 pl1.3	GRT EH29 pl1.4	GRT EH29 pl1.5	GRT EH29 pl1.6	GRT EH29 pl1.7	GRT EH29 pl2.1	GRT EH29 pl2.2	GRT EH29 pl2.3
ppm												
Sc	0.266	<0.351	<1.199	<0.654	<0.548	<0.249	<0.107	0.157	0.215	0.143	0.178	0.206
Rb	1.329	0.989	25.731	0.947	3.413	10.17	2.53	1.795	6.005	2.685	0.64	1.364
Sr	275	782	1088	953	98	195	128	1057	974	1115	1006	1013
Y	0.364	0.144	3.713	0.494	0.177	0.117	0.139	0.435	0.477	0.53	0.838	0.683
Zr	<0.045	<0.041	0.087	0.071	<0.091	<0.034	<0.014	0.058	0.051	0.043	0.228	0.579
Nb	0.012	<0.005	<0.028	<0.014	<0.013	<0.006	0.002	<0.004	0.003	<0.002	0.015	0.012
Cs	0.046	<0.028	0.123	0.053	0.158	0.102	0.064	0.115	0.158	0.14	0.018	0.011
Ba	45	263	269	302	23	53	40	379	334	302	242	261
La	16.9	10.7	9.7	10.9	12.0	11.1	11.2	12.9	11.7	10.6	8.5	8.9
Ce	17.6	12.4	17.8	15.6	14.5	13.1	13.2	18.0	17.0	15.6	13.3	13.5
Pr	1.097	0.846	1.87	1.355	0.926	0.856	0.835	1.505	1.399	1.372	1.239	1.244
Nd	2.57	2.27	8.30	4.15	2.12	1.84	1.91	4.73	4.52	4.55	4.33	4.16
Sm	0.189	0.217	1.495	0.447	0.176	0.125	0.159	0.439	0.484	0.494	0.577	0.493
Eu	1.07	1.60	2.08	1.76	0.53	0.66	0.78	2.00	1.97	1.90	1.73	1.62
Gd	0.096	0.1	1.183	0.304	0.108	0.072	0.07	0.234	0.277	0.286	0.348	0.308
Tb	0.013	0.011	0.123	0.033	0.006	0.006	0.008	0.023	0.03	0.035	0.042	0.033
Dy	0.054	0.033	0.689	0.127	0.051	0.045	0.036	0.121	0.122	0.124	0.218	0.167
Ho	0.013	0.037	0.157	0.02	0.003	0.003	0.005	0.017	0.02	0.024	0.034	0.028
Er	0.037	0.021	0.371	0.027	0.081	0.005	0.01	0.045	0.035	0.078	0.082	0.065
Tm	0.003	<0.001	0.047	0.009	<0.002	<0.002	<0.002	0.005	0.006	0.004	0.011	0.008
Yb	<0.017	<0.014	0.304	0.013	<0.026	0.008	0.01	0.036	0.023	0.026	0.067	0.039
Lu	0.003	<0.003	0.024	<0.004	<0.004	0.003	0.001	0.007	0.003	0.004	0.007	0.004
Pb	10.86	9.81	8.07	6.99	17.84	10.91	16.22	9.88	9.09	9.10	7.07	7.14
Th	0.017	0.002	0.164	<0.001	0.001	0.001	0.004	0.002	0.012	0.006	0.047	0.037
U	0.003	0.002	0.025	0	0.004	0.001	0.002	0.002	0.001	0.002	0.01	0.008
An	0.25	0.32	0.67	0.47	0.17	0.17	0.17	0.47	0.47	0.51	0.51	0.51
Unit: Sample: ID:	GRT EH29 pl2.4	GRT EH29 pl2.5	GRT EH29 pl2.6	GRT EH29 pl2.7	GRT EH29 pl2.8	GRT EH29 pl2.9						
ppm												
Sc	0.183	0.207	0.259	0.245	0.249	0.233						
Rb	0.397	1.128	0.968	3.142	2.771	1.913						
Sr	960	842	971	773	136	146						
Y	0.517	0.316	0.292	0.29	0.237	0.065						
Zr	0.055	0.153	0.043	0.522	0.356	<0.013						
Nb	0.003	0.011	0.006	<0.002	0.013	<0.001						
Cs	<0.011	0.022	0.011	0.074	0.275	0.231						
Ba	220	270	349	351	57	63						
La	8.5	8.8	11.6	11.1	8.5	5.6						
Ce	14.2	13.1	15.0	14.4	10.5	6.7						
Pr	1.267	1.07	1.12	1.167	0.673	0.446						
Nd	4.21	3.30	3.12	3.41	1.64	1.08						
Sm	0.534	0.337	0.309	0.342	0.16	0.092						
Eu	1.57	1.37	1.76	1.54	0.64	0.85						
Gd	0.281	0.173	0.153	0.174	0.075	0.045						
Tb	0.029	0.018	0.015	0.019	0.008	0.003						
Dy	0.131	0.072	0.073	0.074	0.046	0.016						
Ho	0.022	0.013	0.011	0.015	0.007	0.003						
Er	0.047	0.029	0.026	0.022	0.014	0.009						
Tm	0.004	0.003	0.004	0.004	0.002	<0.001						
Yb	0.039	0.021	0.015	0.016	0.011	<0.006						
Lu	0.004	0.003	0.002	0.003	0.003	0.001						
Pb	7.17	7.13	8.23	8.69	11.95	11.66						
Th	0.004	0.046	0.004	0.004	0.021	0.001						
U	0.001	0.013	0.001	0.002	0.022	0.002						
An	0.57	0.44	0.47	0.33	0.15	0.15						

TABLE D.30: Trace element concentration of plagioclase xenocrysts in the Ebisutoge PD

Unit: Sample: ID:	Unit B EH65a pl2.1	Unit B EH65a pl2.5	Unit D EH65b pl1.1	Unit D EH65b pl1.2	Unit D EH65b pl1.3	Unit D EH65b pl1.4	Unit D EH65b pl1.5	Unit D EH65b pl1.6	Unit D EH65b pl1.7	Unit D EH65b pl1.8	Unit D EH65b pl1.9	Unit D EH65b pl1.10	Unit D EH65b pl1.15
<i>ppm</i>													
Sc	0.381	0.342	<0.432	<0.310	<0.209	<0.224	<0.204	<0.274	<0.291	0.37	<0.241	0.314	0.214
Rb	2.269	1.278	1.446	0.682	0.388	0.55	0.459	0.9	0.739	0.583	0.568	0.642	0.542
Sr	654	711	738	790	661	798	576	847	821	843	819	804	873
Y	0.453	0.531	0.141	0.519	0.415	0.337	0.26	0.335	0.492	0.544	0.523	0.571	0.346
Zr	0.292	0.355	<0.068	0.212	0.276	0.136	0.095	<0.058	0.053	<0.054	0.099	<0.056	0.043
Nb	0.051	0.052	<0.010	0.156	0.055	0.017	0.01	0.016	<0.007	<0.006	0.012	<0.014	0.005
Cs	0.038	0.02	<0.069	<0.040	<0.047	<0.049	<0.032	<0.050	<0.034	<0.042	<0.052	<0.055	<0.014
Ba	411	372	346	268	198	292	229	343	302	380	380	338	391
La	12.3	11.7	4.447	8.463	7.111	8.696	7.242	11.061	10.204	11.929	10.943	11.772	11.9
Ce	16.0	16.6	5.481	12.865	11.075	12.186	9.576	14.214	14.796	15.818	14.723	16.6	15.9
Pr	1.24	1.29	0.41	1.08	0.95	0.95	0.74	1.13	1.21	1.25	1.18	1.39	1.27
Nd	3.68	3.84	1.068	3.475	2.885	2.814	1.98	3.181	3.775	3.871	3.524	4.064	3.75
Sm	0.38	0.405	0.145	0.417	0.43	0.321	0.221	0.33	0.413	0.39	0.36	0.458	0.351
Eu	2.08	2.05	1.40	1.40	1.19	1.47	1.07	1.65	1.67	1.78	1.78	1.91	1.89
Gd	0.177	0.226	0.073	0.232	0.181	0.171	0.117	0.194	0.178	0.183	0.222	0.276	0.177
Tb	0.021	0.024	0.016	0.023	0.022	0.02	0.013	0.018	0.025	0.026	0.024	0.03	0.020
Dy	0.105	0.140	0.065	0.123	0.161	0.091	0.076	0.082	0.119	0.112	0.137	0.154	0.097
Ho	0.016	0.019	<0.006	0.023	0.018	0.016	0.007	0.016	0.018	0.019	0.02	0.016	0.013
Er	0.047	0.043	0.007	0.036	0.038	0.037	0.022	0.037	0.033	0.041	0.06	0.055	0.032
Tm	0.006	0.005	<0.006	0.004	0.005	0.003	0.003	0.002	0.006	0.007	0.005	0.004	0.004
Yb	0.037	0.041	<0.017	0.023	0.029	<0.018	0.025	0.02	<0.029	0.034	0.022	0.093	0.052
Lu	0.006	0.006	<0.004	0.004	0.005	<0.003	0.003	<0.003	0.005	0.004	0.009	0.005	0.002
Pb	10.88	10.28	6.28	6.26	5.16	6.27	4.72	7.07	6.82	7.41	8.35	9.40	8.20
Th	0.117	0.063	<0.010	0.024	0.046	0.018	0.006	0.013	0.015	0.015	0.027	0.01	0.007
U	0.014	0.014	0	0.003	0.019	0.008	0.004	0.013	0.003	0.003	0.008	0.002	0.003
An	0.28	0.31	0.31	0.46	0.41	0.44	0.30	0.45	0.49	0.43	0.40	0.45	0.44

Appendix E

AMS measurements

Magnetic and magmatic fabrics are determined by measurements of anisotropy of magnetic susceptibility (AMS) and measurements of mineral orientations in the field, respectively, to identify the structures associated with magma emplacement of the Takidani Pluton. Mineral alignments in igneous rocks occur as result of deformation and magma flow (Bhattacharyya 1966, Bouchez 1997, Callahan & Markley 2003, Mamtani & Greiling 2005, Mamtani et al. 2011). The direction of the mineral alignment largely depends on the aspect ratio of the mineral, the rheology of the matrix material, the concentration and interaction of particles and strain. Quantification of rock fabrics is used to determine the orientations of the principle axes of strain that can be related to magma flow and emplacement (Archanjo et al. 2012, Bouchez 1997, Hutton 1988, Launeau & Cruden 1998, Paterson et al. 1998, Payacán et al. 2014, de Saint-Blanquat et al. 2001, Schöpa et al. 2015, Trubač et al. 2009, Vernon 2000).

Magmatic rock fabrics are determined by measurement of mineral lineation within a foliation plane. This can be achieved by directly measuring mineral alignment in the field and by indirect measurements of shape-preferred orientations (SPO) of minerals in oriented thin sections. Granitic rocks are often deprived of visible mineral alignment. Determination of magnetic fabrics, thus, is used to select samples to further constrain magmatic fabrics.

Magnetic susceptibility of rocks was measured in the field using a hand-held magnetic susceptibility meter (KT-10 v2) from Terraplus, and in the laboratory using a MFK1 Kappabridges from AGICO. Measurements in the field were carried out on flat rock surfaces mainly along 30 cm to 100 cm transects using the scanning function of the Kappameter. Where surface exposure is limited, shorter or single point measurements were taken. In total, over two hundred sites were measured with more than ten thousand bulk susceptibility readings combined. Measurements in the laboratory were carried out on cylindrical specimens (25.4 x 22 mm) using peak field intensity of 300 Am^{-1} .

The orientation and magnitude of the three principal axes of the magnetic susceptibility ellipsoid (K1, K2, K3) were determined for 240 oriented cores from 49 rock

samples. The magnitude of the principal directions is used to calculate the mean susceptibility (K_m), degree of magnetic anisotropy (P), magnitude of magnetic lineation (L), magnetic foliation (F) and shape parameter (T). The results are summarised in Figure E.1 and E.2. The pole to the magnetic foliation is attributed to K3 and the magnetic lineation is attributed to K1. The mean bulk susceptibility overall varies between 7.9×10^{-5} and 4.4×10^{-2} SI across the different lithological units of the Takidani Pluton and mimics the major and trace elements composition variability measured in the whole rock samples (Tables ??, ??, D.3- D.5). The magnetic fabrics of the Takidani Pluton are generally characterised by triaxial ellipsoids of magnetic susceptibility (Figure E.3). Strong magnetic foliations are observed at the western contact of the pluton in the fine-grained granite (fGT) and base of the granodiorite (GRT) potentially indicating shearing and deformation along the tectonic contact. Magnetic lineations and foliations are oriented in NE-SW direction (Figures E.4, E.5).

Amphibole lineation were measured at several locations of the Takidani pluton. The magmatic foliations were determined using the Stereonet 10.0 software (Allmendinger et al. 2011, Cardozo & Allmendinger 2013). Magmatic foliations are oriented in NE-SW direction and sub-parallel to the magnetic foliation of the Takidani Pluton (Figures E.5, E.6). Shape preferred orientations of amphibole was additionally measured by image analysis of scanned oriented thin sections using the ImageJ software package (Rueden et al. 2017, Schindelin et al. 2012). The results of SPO investigations are currently being processed.

Field and petrological investigations suggest that the chemical zoning of minerals in various portion of the intrusion and the lack of clear internal contacts within the Takidani Pluton are the result of successive magma emplacement indicating that the magnetic and magmatic fabrics likely formed during magma emplacement of the felsic magmas of the Takidani Pluton. This project complements our petrological investigations and aims to improve our understanding of the construction and evolution of zoned magma reservoirs in the upper crust. Ongoing investigations of SPO will help to better constrain the fabric strength of magmatic rocks and to identify strain gradients in the Takidani Pluton.

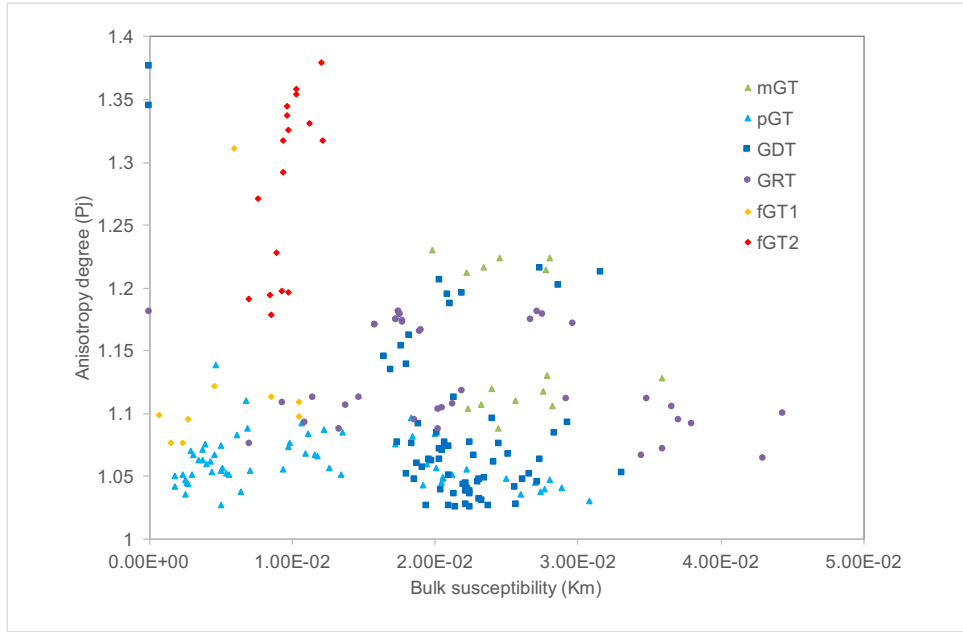


FIGURE E.1: Anisotropy degree (P_j) and bulk susceptibility of the Takidani Pluton

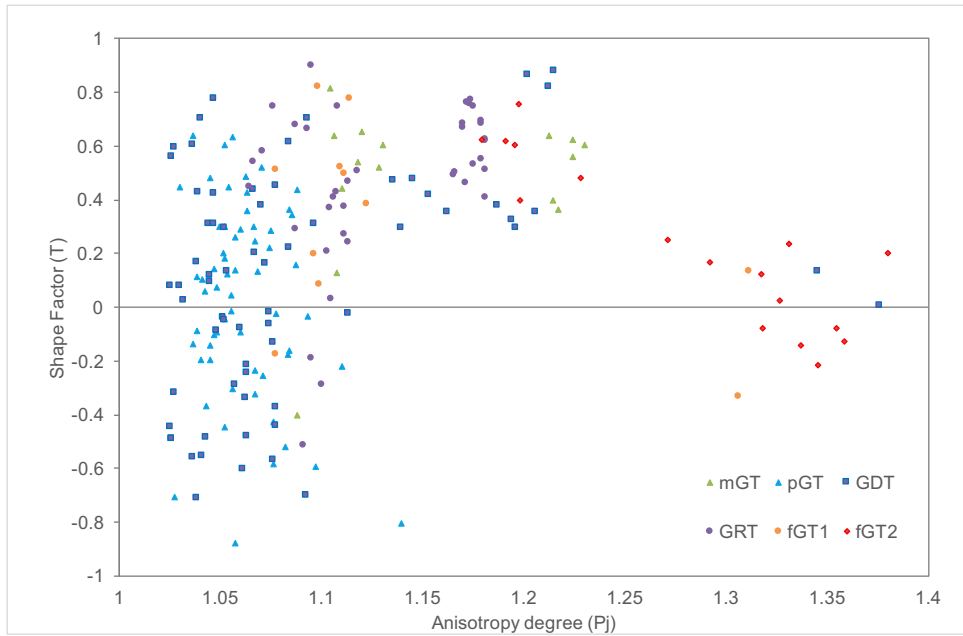


FIGURE E.2: Shape factor (T) versus degree of anisotropy of the Takidani Pluton

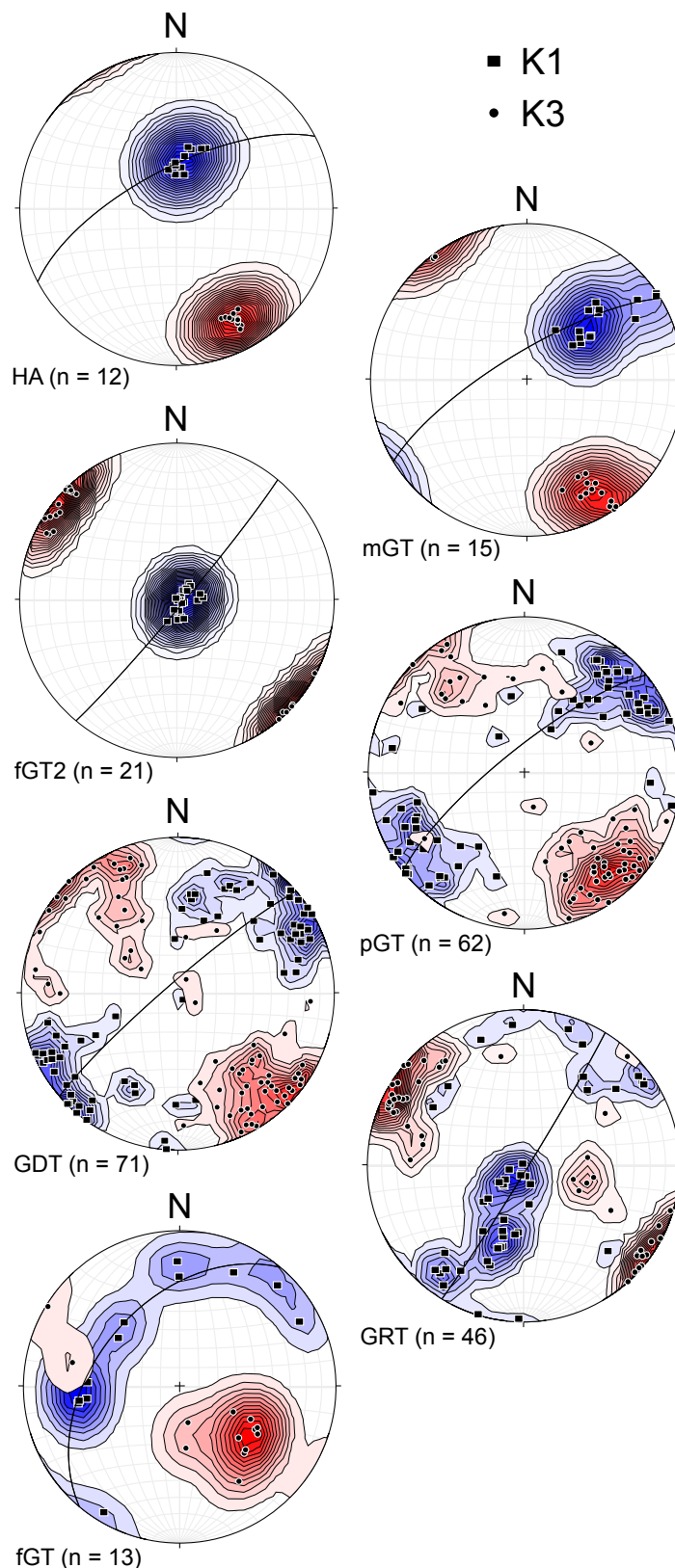


FIGURE E.3: K1 and K3 principle axis of the magnetic susceptibility ellipsoid plotted in stereonet projections (Allmendinger et al. 2011, Cardozo & Allmendinger 2013). Stereonet projections show the magnetic fabric of the Hotaka Andesite and the six lithological units of the Takidani Pluton. Kamb contours are created for K1 (blue) and K3 (red) using an interval and significance level of 1. Contour lines are smoothed. Solid curved line indicates the magnetic foliations determined for each lithological unit using Bingham analysis.

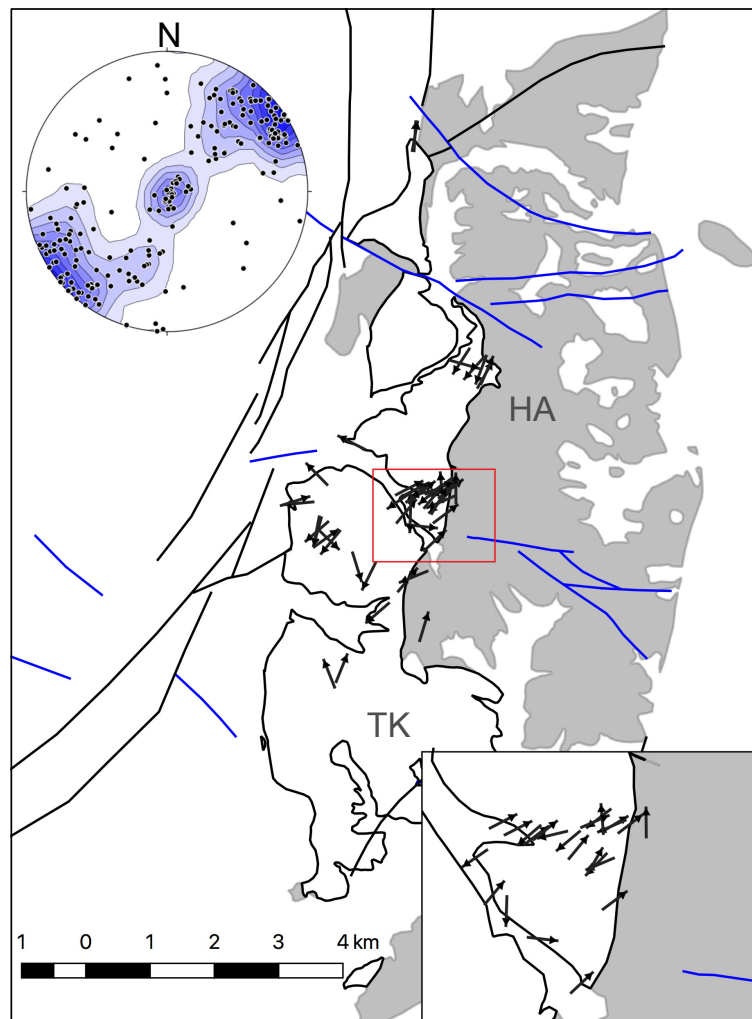


FIGURE E.4: Orientation of magnetic lineation of the Takidani Pluton in Central Japan. Magnetic lineations are determined by measurements of anisotropy of magnetic susceptibility (AMS) of oriented cylindrical samples. Black lines show NE-SW running faults. Blue lines indicate NW-SE running faults.

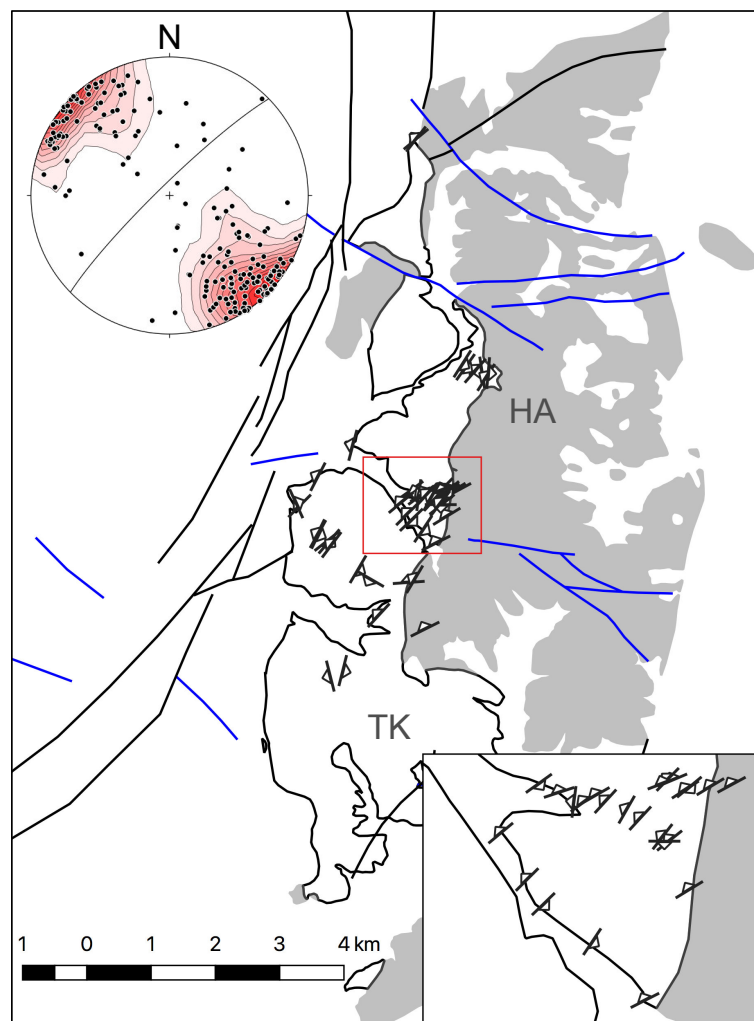


FIGURE E.5: Orientation of magnetic foliation of the Takidani Pluton in Central Japan. Magnetic foliations are determined by measurements of anisotropy of magnetic susceptibility (AMS) of oriented cylindrical samples. Black lines show NE-SW running faults. Blue lines indicate NW-SE running faults.

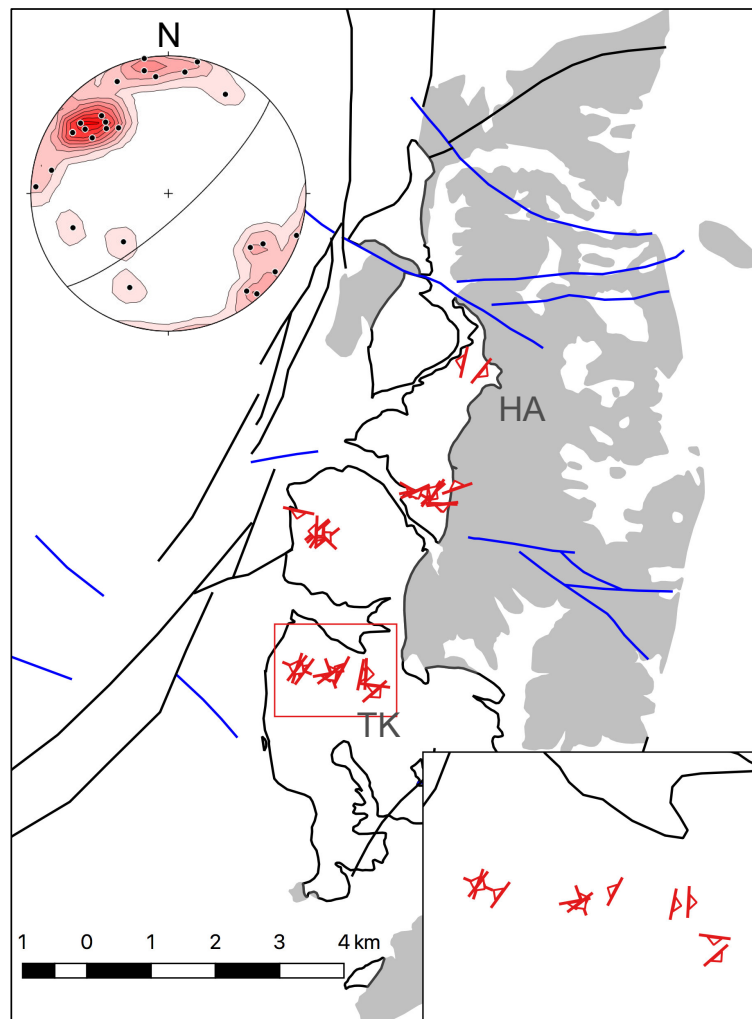


FIGURE E.6: Orientation of magmatic foliation of the Takidani Pluton in Central Japan. Magmatic foliations are established from direct measurements of amphibole lineation in the field using Stereonet. Black lines show NE-SW running faults. Blue lines indicate NW-SE running faults.

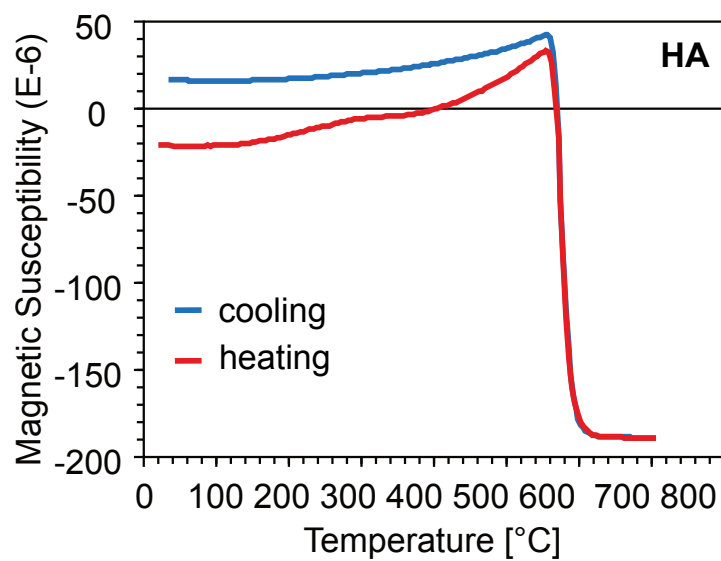


FIGURE E.7: Bulk susceptibility measurements of Hotaka Andesite during cooling and heating

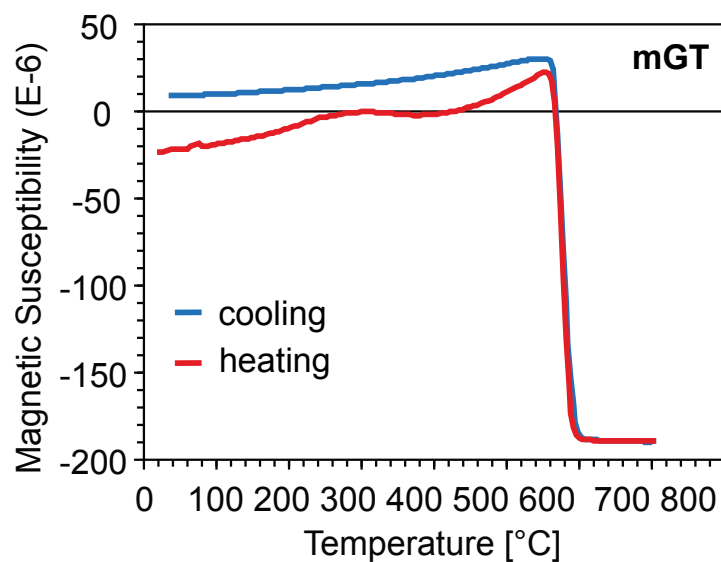


FIGURE E.8: Bulk susceptibility measurements of the mGT unit during cooling and heating

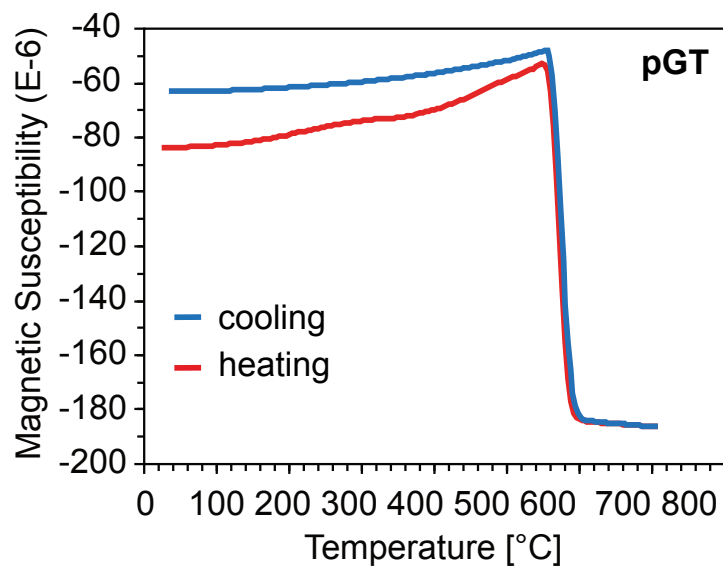


FIGURE E.9: Bulk susceptibility measurements of the pGT unit during cooling and heating

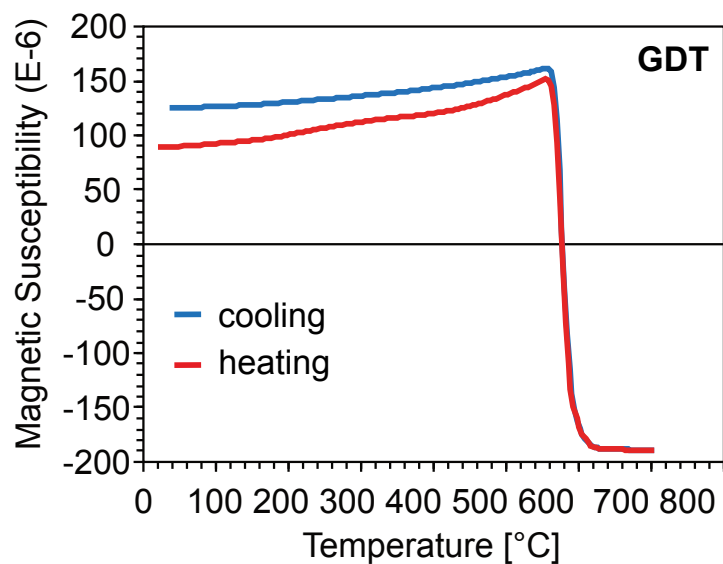


FIGURE E.10: Bulk susceptibility measurements of the pGT unit during cooling and heating

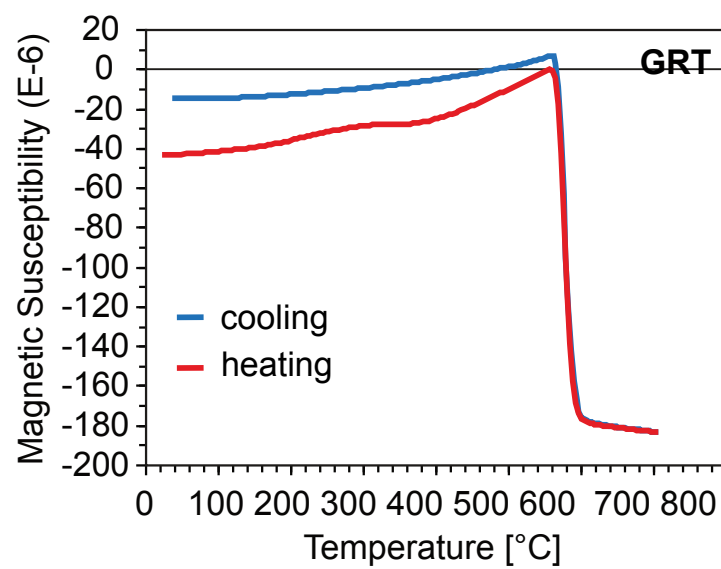


FIGURE E.11: Bulk susceptibility measurements of the pGT unit during cooling and heating

TABLE E.1: AMS measurements of the Takidani Pluton

Unit: Sample: ID	HA EH4 A1	HA EH4 A2	HA EH4 B0	HA EH4 B1	HA EH4 B2	HA EH4 C1	HA EH4 C2	HA EH54 A1	HA EH54 A2	HA EH54 A3	HA EH54 B1	HA EH54 B2	mGT EH5 A1	mGT EH5 A2	mGT EH5 B2	mGT EH5 B3	mGT EH39 A	mGT EH39 B
Km	2.60E-02	2.24E-02	3.02E-02	1.85E-02	2.99E-02	3.01E-02	2.09E-02	3.48E-02	2.51E-02	3.81E-02	3.94E-02	2.27E-02	2.78E-02	2.56E-02	3.59E-02	2.75E-02	2.32E-02	2.44E-02
L	1.146	1.145	1.126	1.126	1.139	1.116	1.129	1.02	1.086	1.023	1.025	1.023	1.023	1.029	1.028	1.025	1.045	1.06
F	1.245	1.249	1.247	1.283	1.252	1.229	1.239	1.078	1.086	1.078	1.083	1.077	1.097	1.076	1.091	1.085	1.059	1.025
P	1.426	1.431	1.404	1.445	1.426	1.426	1.398	1.1	1.108	1.102	1.111	1.101	1.122	1.107	1.122	1.112	1.107	1.086
Pj	1.431	1.436	1.411	1.456	1.431	1.426	1.404	1.106	1.115	1.107	1.115	1.106	1.113	1.11	1.128	1.118	1.107	1.088
T	0.234	0.242	0.301	0.354	0.267	0.165	0.277	0.579	0.613	0.539	0.531	0.536	0.604	0.444	0.52	0.541	0.128	-0.402
U	0.149	0.157	0.222	0.272	0.183	0.078	0.198	0.563	0.597	0.522	0.511	0.518	0.586	0.423	0.499	0.522	0.102	-0.419
Kld	358.4	12.8	358.8	356.1	6.5	348.7	359	26	22.5	8.8	13.9	10.7	56.2	57	41.7	54	41.6	60.7
Kl1	72.1	71.7	65.7	67.3	68.2	69.2	67.3	54.4	55.7	62	57.8	56.9	0.4	1.5	34	13.7	40.2	21.5
K2d	243.2	245	247.1	244.1	245.6	243.2	244.6	259.3	256.4	250.3	250.7	248.2	325.2	324.3	267.7	303	259.1	297.4
K3d	7.9	11.5	9.5	8.9	11.6	5.8	9.8	23.2	21.9	14.2	19	19.4	68.2	60	45.8	55.9	43.2	54.4
K3d	150.9	152	153.2	150.7	151.8	151.1	151	157.6	155.6	154	151.4	148.4	146.4	147.9	149.7	152.2	149.3	162.2
K3i	16	14.1	22.1	20.7	18.2	19.9	20.2	25.4	25	23.6	24.9	25.8	21.8	30	24.6	30.6	19.8	26.9
Mfd	330.9	332	333.2	330.7	331.8	331.1	331	337.6	335.6	334	331.4	328.4	326.4	327.9	329.7	332.2	329.3	342.2
MFi	74	75.9	67.9	69.3	71.8	70.1	69.8	64.6	65	66.4	65.1	64.2	68.2	60	65.4	59.4	70.2	63.1
AK1	81.8	78.2	79.8	80.5	77.8	83.8	79.6	64.1	65.7	74.5	68.9	68.4	0.4	1.8	37.9	15.9	43.3	24.2
Unit: Sample: ID	mGT EH39 C	mGT EH39 D	mGT EH39 E	mGT EH72 A1	mGT EH72 A2	mGT EH72 B1	mGT EH72 B2	mGT EH72 C1	mGT EH72 C2	mGT EH6 A1	mGT EH6 A2	mGT EH6 A3	mGT EH6 B1	mGT EH6 B2	mGT EH6 B3	mGT EH7 A1	mGT EH7 A2	mGT EH7 B1
Km	2.40E-02	2.82E-02	2.23E-02	2.80E-02	1.98E-02	2.45E-02	2.22E-02	2.78E-02	2.34E-02	2.88E-02	2.73E-02	2.60E-02	2.70E-02	2.80E-02	2.77E-02	2.05E-02	2.06E-02	3.08E-02
L	1.018	1.017	1.008	1.043	1.039	1.036	1.033	1.058	1.063	1.018	1.021	1.02	1.027	1.026	1.024	1.025	1.022	1.008
F	1.092	1.081	1.085	1.162	1.17	1.167	1.159	1.141	1.14	1.022	1.017	1.015	1.018	1.021	1.016	1.019	1.025	1.021
P	1.112	1.099	1.094	1.212	1.216	1.209	1.198	1.208	1.212	1.041	1.038	1.036	1.045	1.047	1.04	1.045	1.048	1.029
Pj	1.12	1.106	1.104	1.224	1.23	1.224	1.212	1.214	1.217	1.041	1.038	1.036	1.045	1.047	1.04	1.045	1.048	1.03
T	0.656	0.639	0.816	0.562	0.604	0.625	0.639	0.4	0.363	0.102	-0.089	-0.139	-0.196	-0.102	-0.198	-0.141	0.073	0.447
U	0.64	0.625	0.808	0.529	0.572	0.596	0.611	0.36	0.321	0.092	-0.098	-0.147	-0.207	-0.113	-0.207	-0.151	0.062	0.441
Kld	47	29.8	45.8	55.6	51.5	53	47.1	55.2	55.7	59	62.1	61.7	57.8	59.9	53.1	247.9	245.6	325
Kl1	37.5	59.9	35.8	55.6	54.4	59.9	52.2	48.8	49.5	16.4	11.8	17.4	10.6	25.3	17.5	26.6	17.2	66.7
K2d	263.9	253.5	254.8	236.2	239.4	235.5	239.9	230.2	231.7	178.3	173.2	202.9	160.2	181.4	171.6	19.9	27.5	55.4
K2i	46.2	22.8	50.5	34.4	35.3	30.1	37.1	41.1	40.4	59	60	68.2	48.9	47.9	56.5	53.2	68.6	0.1
K3d	152.5	155.4	146.5	146	146.8	144.9	145.2	322.4	323.4	320.9	325.9	327.6	319	313.3	313.6	145.4	151.7	145.4
K3i	19.3	18.7	14.5	0.3	3.7	1.1	6.2	2.4	2	25.5	27.2	12.9	39.2	31.2	27.6	23.5	12.4	23.3
Mfd	332.5	335.4	326.5	326	326.8	324.9	325.2	142.4	143.4	140.9	145.9	147.6	139	133.3	133.6	325.4	331.7	325.4
MFi	70.7	71.3	75.5	89.7	86.3	88.9	83.8	87.6	88	64.5	62.8	77.1	50.8	58.8	62.4	66.5	77.6	66.7
AK1	40.1	65.9	37.2	55.6	54.6	59.9	52.7	48.9	49.6	18.2	13.2	17.8	13.7	30	19.9	29.2	17.6	89.8
Unit: Sample: ID	mGT EH72 B2	mGT EH72 C1	mGT EH72 C2	mGT EH6 A1	mGT EH6 A2	mGT EH6 A3	mGT EH6 B1	mGT EH6 B2	mGT EH6 B3	mGT EH7 A1	mGT EH7 A2	mGT EH7 B1	mGT EH7 B2	mGT EH8 A0	mGT EH8 A1	mGT EH8 B1	mGT EH8 C1	mGT EH8 D1
Km	2.22E-02	2.78E-02	2.34E-02	2.88E-02	2.73E-02	2.60E-02	2.70E-02	2.80E-02	2.77E-02	2.05E-02	2.06E-02	3.08E-02	2.11E-02	2.88E-03	3.72E-03	3.88E-03	2.59E-03	2.51E-03
L	1.033	1.058	1.063	1.018	1.021	1.02	1.027	1.026	1.024	1.025	1.022	1.008	1.009	1.016	1.043	1.056	1.011	1.006
F	1.159	1.141	1.14	1.022	1.017	1.015	1.018	1.021	1.016	1.019	1.025	1.021	1.039	1.05	1.026	1.015	1.032	1.027
P	1.198	1.208	1.212	1.041	1.038	1.036	1.045	1.047	1.04	1.045	1.048	1.029	1.049	1.067	1.071	1.072	1.043	1.033
Pj	1.212	1.214	1.217	1.041	1.038	1.036	1.045	1.047	1.04	1.045	1.048	1.03	1.052	1.07	1.071	1.076	1.045	1.036
T	0.639	0.4	0.363	0.102	-0.089	-0.139	-0.196	-0.102	-0.198	-0.141	0.073	0.447	0.603	0.521	-0.256	-0.583	0.484	0.637
U	0.611	0.36	0.321	0.092	-0.098	-0.147	-0.207	-0.113	-0.207	-0.151	0.062	0.441	0.595	0.509	-0.271	-0.595	0.476	0.632
Kld	47.1	55.2	55.7	59	61.7	61.7	57.8	59.9	53.1	247.9	245.6	325	246.5	330.2	58.5	302.1	262.7	94.4
Kl1	52.2	48.8	49.5	16.4	11.8	17.4	10.6	25.3	17.2	26.6	17.2	66.7	24.9	48.7	15.6	23.8	2.5	20.3
K2d	239.9	230.2	231.7	178.3	173.2	202.9	160.2	181.4	171.6	19.9	27.5	55.4	17.2	258.1	291.8	198.3	353.5	302.8
K2i	37.1	41.1	40.4	59	60	68.2	48.9	47.9	56.5	53.2	68.6	18.1	65	28.4	65	28.4	18.1	67.2
K3d	145.2	322.4	323.4	320.9	325.9	327.6	319	313.3	313.6	145.4	151.7	144.9	166.6	154	154	65.7	165.3	188.1
K3i	6.2	2.4	2	25.5	27.2	12.9	39.2	31.2	27.6	23.5	12.4	23.3	23.5	41.3	19.1	51.4	71.7	10
Mfd	325.2	142.4	143.4	140.9	145.9	147.6	139	133.3	133.6	325.4	331.7	325.4	324.9	346.6	334	345.7	345.3	8.1
MFi	83.8	87.6	88	64.5	62.8	77.1	50.8	58.8	62.4	66.5	77.6	66.5	66.7	70.9	38.6	38.6	48.7	80
AK1	52.7	48.9	49.6	18.2	13.2	17.8	13.7	30	19.9	29.2	17.6	89.8	27.3	87.6	16.5	40.4	7.8	20.6

TABLE E.2: (Continued)

Unit Sample:	PGT EH9 A1	PGT EH9 A2	PGT EH9 B1	PGT EH9 C1	PGT EH9 C2	PGT EH9 D0	PGT EH9 D1	PGT EH9 D2	PGT EH10 A1	PGT EH10 B1	PGT EH10 C1	PGT EH11 A1	PGT EH11 A2	PGT EH11 A3	PGT EH11 A4	PGT EH11 B1	PGT EH11 B2	PGT EH11 B3
Km	2.30E-03	2.48E-03	3.45E-03	1.80E-03	1.79E-03	2.94E-03	3.05E-03	2.74E-03	1.11E-02	1.22E-02	1.06E-02	1.17E-02	1.34E-02	1.09E-02	1.26E-02	1.16E-02	9.79E-03	1.35E-02
L	1.018	1.02	1.017	1.02	1.02	1.021	1.041	1.019	1.026	1.036	1.047	1.022	1.027	1.029	1.024	1.043	1.038	1.027
F	1.033	1.026	1.043	1.022	1.03	1.03	1.025	1.024	1.056	1.049	1.044	1.042	1.025	1.038	1.032	1.022	1.037	1.055
P	1.052	1.047	1.061	1.042	1.05	1.052	1.067	1.044	1.083	1.087	1.093	1.065	1.052	1.068	1.057	1.065	1.077	1.084
P ₁	1.052	1.047	1.063	1.042	1.051	1.052	1.067	1.044	1.084	1.087	1.093	1.066	1.052	1.068	1.057	1.067	1.077	1.085
T	0.299	0.141	0.426	0.058	0.2	0.181	-0.236	0.124	0.363	0.158	-0.035	0.301	-0.045	0.132	0.139	-0.323	-0.023	0.343
U	0.288	0.13	0.414	0.048	0.188	0.168	-0.251	0.113	0.346	0.138	-0.057	0.286	-0.057	0.116	0.125	-0.338	-0.042	0.325
Kid	220.3	237.3	243	238.7	231.1	251.2	218.7	217.4	64.4	70.2	64.7	41.6	42.5	44.4	37.6	39.1	36.5	34.1
KH	6.3	0.4	23.3	15.7	4.3	7.2	15	16.9	13	1.9	8.2	12.3	21.1	24.5	17.1	17.8	22.7	14.8
K2d	322.9	327.7	5.7	348.6	325.1	348.2	335.7	335.7	296.3	332.2	319.7	267.1	278.8	223.2	260.2	232.3	240.3	246.9
K2i	63.4	43.1	51.5	50.5	43.1	43.7	62.2	57.4	69.5	76.4	60.9	72.8	55.2	65.5	67.3	71.7	65.4	72.6
K3d	127.3	146.9	139.3	137.3	136.5	153.8	122.9	118.5	158.1	160.6	159	134.2	143.5	134.2	132.2	130.3	130.3	126.5
K3i	25.8	46.5	28.8	35.1	46.5	45.4	20.5	27	15.5	13.5	22.7	11.9	26.3	0.4	14.4	3.9	8.9	9.1
M2d	307.3	326.9	319.3	317.3	316.5	333.8	302.9	298.5	338.1	340.6	339	314.2	323.5	134.2	312.2	310.3	310.3	306.5
M2i	64.2	43.5	61.2	54.9	43.5	44.6	69.5	63	76.5	76.5	62.3	78.1	63.7	89.6	75.6	86.1	81.1	80.9
AK1	6.9	0.6	26.8	19.3	6.3	10.3	16.1	19	13.5	2	9.2	12.5	23.7	24.5	17.7	17.9	23	15
Unit Sample:	PGT EH11 B4	PGT EH12 A1	PGT EH12 A2	PGT EH12 A3	PGT EH12 B1	PGT EH12 B2	PGT EH12 B3	PGT EH12 A1	PGT EH12 A2	PGT EH12 A3	PGT EH12 B1	PGT EH12 B2	PGT EH12 B3	PGT EH12 A1	PGT EH12 A2	PGT EH12 A3	PGT EH12 B1	PGT EH12 B2
Km	9.69E-03	2.00E-02	2.06E-02	2.50E-02	1.94E-02	1.91E-02	2.01E-02	5.36E-03	6.11E-03	4.01E-03	7.09E-03	6.73E-03	5.08E-03	4.41E-03	4.25E-03	6.36E-03	3.71E-03	4.51E-03
L	1.028	1.048	1.017	1.026	1.032	1.029	1.047	1.023	1.048	1.021	1.026	1.065	1.021	1.014	1.015	1.017	1.019	1.025
F	1.045	1.034	1.031	1.021	1.027	1.013	1.003	1.029	1.033	1.038	1.028	1.041	1.035	1.037	1.044	1.021	1.041	1.041
P	1.074	1.083	1.048	1.048	1.06	1.042	1.05	1.053	1.083	1.059	1.055	1.109	1.057	1.052	1.059	1.058	1.061	1.066
P ₁	1.074	1.084	1.048	1.048	1.06	1.043	1.057	1.053	1.083	1.06	1.055	1.11	1.057	1.054	1.062	1.038	1.063	1.067
T	0.223	-0.162	0.3	-0.093	-0.091	-0.37	-0.879	0.123	-0.174	0.291	0.044	-0.219	0.262	0.445	0.486	0.112	0.357	0.246
U	0.206	-0.181	0.289	-0.104	-0.106	-0.379	-0.881	0.11	-0.193	0.278	0.031	-0.244	0.269	0.435	0.475	0.103	0.344	0.23
Kid	40.4	219.4	238.1	219.4	237.5	236.7	232.4	45.5	42.7	229.8	231.1	253.9	229	254.7	211.9	280.7	195.1	233
KH	37.8	27.7	5.2	15.7	147	27.5	30.2	45.5	13.2	0.5	2.8	19	1.5	29.7	43.8	17.5	32	20.1
K2d	328.5	23.9	348.8	328.5	54	67.3	66	266.6	32.1	67.6	331.3	33.3	322.2	67.8	60.1	100.8	89.4	118
K2i	51.8	62.2	71.3	49.3	66	62.1	59.1	79.6	70.8	74.7	140.4	65.5	65.8	63.4	20.5	43.8	23.4	49.1
K3d	134	124	146.4	117.4	147	329	325.9	136.4	136	139.6	151	14.8	15.1	163	353.1	24.9	330.1	337.2
K3i	4.7	2.3	17.9	36.3	1.3	4.4	6	6.7	13.6	22.4	15.1	14.8	24.2	3	39.1	42.3	48.6	33.8
M2d	314	304	326.4	327	322	149	145.9	316.4	316	319.6	320.4	338.7	318.3	343	173.1	204.9	150.1	157.2
M2i	85.3	87.7	72.1	85.7	83.3	85.6	84	83.3	76.4	67.6	74.9	75.2	65.8	87	50.9	47.7	41.4	56.2
AK1	38	27.7	5.5	19.6	23.9	27.6	30.4	7.9	13.6	0.6	2.9	19.7	1.6	29.8	63.1	20.6	53.2	24.5
Unit Sample:	PGT EH74 B3	PGT EH75 A1	PGT EH75 A2	PGT EH75 B1	PGT EH75 B2	PGT EH75 C1	PGT EH75 C2	GDT EH16 A1	GDT EH16 A2	GDT EH16 B1	GDT EH16 B2	GDT EH16 C1	GDT EH16 C2	GDT EH18 A1	GDT EH18 A2	GDT EH18 B1	GDT EH18 B2	GDT EH18 C1
Km	6.86E-03	4.99E-03	5.56E-03	5.02E-03	9.34E-03	4.67E-03	4.99E-03	2.57E-02	2.22E-02	2.37E-02	2.23E-02	2.22E-02	2.22E-02	1.95E-02	2.25E-02	2.10E-02	2.14E-02	2.31E-02
L	1.023	1.028	1.036	1.026	1.036	1.113	1.022	1.017	1.005	1.005	1.005	1.015	1.016	1.019	1.007	1.019	1.027	1.016
F	1.061	1.027	1.014	1.047	1.019	1.011	1.004	1.009	1.021	1.019	1.032	1.028	1.022	1.006	1.027	1.006	1.027	1.016
P	1.085	1.055	1.052	1.075	1.055	1.125	1.025	1.026	1.026	1.025	1.037	1.043	1.038	1.025	1.034	1.025	1.034	1.032
P ₁	1.088	1.055	1.052	1.075	1.055	1.139	1.027	1.027	1.027	1.025	1.04	1.044	1.038	1.026	1.036	1.026	1.036	1.032
T	0.438	-0.014	-0.445	0.285	-0.306	-0.807	-0.705	-0.317	0.593	0.558	0.203	0.311	0.166	-0.491	0.606	-0.493	-0.557	0.027
U	0.422	-0.027	-0.455	0.269	-0.318	-0.817	-0.709	-0.323	0.589	0.554	0.698	0.301	0.157	-0.495	0.601	-0.498	-0.563	0.019
Kid	102.5	31.8	32.6	28.8	17.1	221.2	54.5	64	78.9	49.2	242.8	63.8	59.6	62.6	246	74.2	59.3	247.4
K2i	4.1	16.2	15.1	20.3	13.6	37.8	38.1	13.6	6.4	6.4	3.7	8.1	24.7	16.1	2.7	21.2	0.5	1.8
K2d	198.9	266.4	262.5	210.8	243.2	248.8	248.8	162	219	141.8	142.7	164.3	181.5	168.9	145.5	229.2	316	343.7
K2i	57	63.4	67.2	69.1	32.2	40.6	51	29.9	51.5	22.1	33.4	52	69.7	44.2	75.7	66.8	88	73.6
K3d	32.7	20.5	127.2	108.8	142.3	108.4	150	312.5	336.1	304	334.1	327.7	340.9	317.9	336.7	340.7	149.3	156.9
K3i	189.9	32.3	288.8	322.3	163	28.6	7	56.5	19.9	66.9	20	36.9	30.5	41.4	14	8.9	1.9	16.3
M2d	57.3	69.5	73.4	85.5	32.3	63.4	83	132.5	156.1	124	154.1	147.7	133.9	137.9	156.7	160.7	329.3	73.7
M2i	4.9	17.3	15.8	20.4	56.2	43.3	38.4	25.3	33.7	16.4	4	10.1	29	21.7	2.7	21.5	0.5	1.9

TABLE E.3: (Continued)

Unit: Sample: ID	GDT EH18 C2	GDT EH20 A1	GDT EH20 A2	GDT EH20 B1	GDT EH20 B2	GDT EH21 A1	GDT EH21 A2	GDT EH21 A3	GDT EH21 B1	GDT EH21 B2	GDT EH21 B3	GDT EH23 A1	GDT EH23 A2	GDT EH23 A3	GDT EH23 B1	GDT EH23 B2	GDT EH24 A1	GDT EH24 A2
Km	2.15E-02	2.20E-02	2.33E-02	2.25E-02	2.42E-02	2.05E-02	2.28E-02	1.80E-02	1.88E-02	1.91E-02	1.86E-02	2.45E-02	2.74E-02	2.41E-02	2.03E-02	1.98E-02	2.56E-02	2.10E-02
L	1.017	1.031	1.014	1.004	1.016	1.021	1.045	1.018	1.032	1.036	1.016	1.056	1.037	1.031	1.044	1.04	1.03	1.026
F	1.007	1.011	1.016	1.005	1.011	1.047	1.045	1.033	1.027	1.02	1.03	1.015	1.024	1.061	1.015	1.02	1.008	1.024
P	1.024	1.042	1.03	1.035	1.057	1.069	1.064	1.052	1.06	1.056	1.046	1.076	1.062	1.094	1.063	1.061	1.039	1.051
Pj	1.025	1.043	1.03	1.038	1.061	1.07	1.066	1.052	1.06	1.057	1.047	1.076	1.063	1.096	1.063	1.062	1.041	1.051
T	-0.446	-0.486	0.079	-0.71	-0.605	0.377	0.436	0.297	-0.077	-0.29	0.312	-0.568	-0.214	0.31	-0.337	-0.337	-0.555	-0.04
U	-0.451	-0.494	0.072	-0.714	-0.613	0.363	0.424	0.286	-0.091	-0.303	0.302	-0.579	-0.228	0.289	-0.493	-0.35	-0.561	-0.053
K1d	64.2	227.7	181.1	246	243	46.2	47.1	45.9	224.1	59.3	56.6	63.2	61.3	24.1	47.5	57.7	242.5	251
K1i	21.1	7.3	29.7	6.4	7.2	33.3	2.9	18.6	4.2	14.6	37.4	16.7	17.1	27.5	2.4	5.1	10.3	18.9
K2d	253.4	337.6	272.8	152.4	248.3	248.3	294.1	225.8	348.2	257.6	222.7	301.9	270.1	278.2	309.8	302.8	116.1	359.9
K2i	68.6	53.3	3	7.9	54.7	54.7	82.7	71.4	82.6	74.7	51.8	59.9	77	27.7	72.3	78	72.9	43.4
K3d	155.4	142.4	8	117.1	30.1	143.2	137.4	315.9	133.7	150.5	321.4	160.9	152.5	150.9	138.2	148.7	335	144
K3i	3.1	35.7	60.1	79.8	81.4	10.5	6.7	0	6.1	4.6	6.8	24.2	61	49	17.5	10.8	13.5	40.5
MFd	335.4	322.4	188	297.1	210.1	323.2	317.4	135.9	313.7	330.5	141.4	340.9	332.5	330.9	318.2	328.7	155	324
MFi	86.9	54.3	29.9	10.2	8.6	79.5	83.3	90	83.9	85.4	83.2	65.8	83.9	41	72.5	79.2	76.5	49.5
AK1	21.1	9	84	39.3	57.3	33.9	2.9	18.6	4.2	14.7	37.7	18.4	11.5	44.8	2.6	5.2	10.6	25.2
Unit: Sample: ID	GDT EH24 B1	GDT EH24 B2	GDT EH24 C1	GDT EH24 D1	GDT EH24 D2	GDT EH37 A1	GDT EH37 B	GDT EH37 C	GDT EH44 A	GDT EH44 B	GDT EH44 C	GDT EH44 D	GDT EH44 E	GDT EH44 F	GDT EH46 A	GDT EH46 B	GDT EH46 C	GDT EH46 D
Km	2.24E-02	2.05E-02	2.51E-02	2.30E-02	2.35E-02	2.66E-02	2.72E-02	3.31E-02	8.80E-07	8.48E-07	9.09E-07	8.29E-07	8.01E-07	9.28E-07	2.02E-02	2.04E-02	2.14E-02	2.07E-02
L	1.012	1.011	1.026	1.02	1.026	1.027	1.019	1.022	1.277	1.172	1.044	1.226	1.249	1.228	1.032	1.029	1.056	1.02
F	1.014	1.027	1.04	1.022	1.022	1.027	1.025	1.029	1.114	1.174	1.174	1.254	1.269	1.241	1.05	1.041	1.053	1.053
P	1.025	1.038	1.067	1.045	1.048	1.052	1.045	1.053	1.423	1.376	1.537	1.585	1.585	1.524	1.084	1.072	1.113	1.074
Pj	1.025	1.039	1.067	1.045	1.048	1.052	1.045	1.053	1.435	1.376	1.537	1.585	1.585	1.524	1.084	1.072	1.113	1.077
T	0.081	0.426	0.201	0.095	-0.088	-0.049	0.12	0.133	-0.385	0.005	0.895	0.054	0.035	0.025	0.221	0.165	-0.026	0.45
U	0.075	0.418	0.186	0.084	-0.1	-0.062	0.11	0.121	-0.459	-0.074	0.849	-0.054	-0.08	-0.08	0.202	0.148	-0.053	0.435
K1d	241.6	52.1	56.4	245.1	257.4	244.9	231	230.6	203.2	184.2	243.2	250	170.5	185.1	215.8	226.6	220.3	217
K1i	15.4	2.7	0.5	12.3	15.9	21.3	16.6	8.9	31.4	0.4	40.8	55.2	6.3	6.3	0.6	6.7	0.2	7.4
K2d	91.3	316.9	154.7	10.4	8.8	26	357.3	340.1	95.7	93.1	20.9	10	274	295.6	309.9	337.5	312.9	15.8
K2i	72.4	62.7	86.7	69.3	52	63.4	63.3	64.9	26.3	72	40.6	19.1	20.7	72.5	81.6	71.7	85.3	82
K3d	333.9	143.4	326.4	151.5	156.6	148.8	134.6	136.8	333.9	274.3	131.9	110.6	31.6	93.3	125.8	134.6	130.3	126.7
K3i	8.3	27.2	3.3	16.4	33.4	15.2	20.3	23.3	46.9	18	22.8	27.8	50.8	16.2	8.4	17	4.7	2.8
MFd	153.9	323.4	146.4	331.5	336.6	328.8	314.6	316.8	153.9	94.3	311.9	290.6	211.6	273.3	305.8	314.6	310.3	306.7
MFi	81.7	62.8	86.7	73.6	56.6	74.8	69.7	66.7	43.1	72	67.2	62.2	39.2	73.8	81.6	73	85.3	87.2
AK1	15.6	3	0.5	12.9	19.2	22.1	17.7	9.7	49.7	0.4	45.1	68.2	56	6.6	0.6	7	0.2	7.4
Unit: Sample: ID	GDT EH46 E	GDT EH46 F	GDT EH46 G	GDT EH46 H	GDT EH47 A	GDT EH47 B	GDT EH47 C	GDT EH47 D	GDT EH47 E	GDT EH47 F	GDT EH14 A1	GDT EH14 A2	GDT EH14 B1	GRT EH14 B2	GRT EH14 C1	GRT EH14 C2	GRT EH27 A	GRT EH27 B
Km	1.96E-02	2.07E-02	2.10E-02	1.84E-02	9.36E-07	8.59E-07	9.58E-07	8.33E-07	1.20E-06	7.17E-07	2.02E-02	1.86E-02	1.10E-02	7.02E-03	9.40E-03	1.34E-02	2.76E-02	2.72E-02
L	1.038	1.039	1.037	1.042	1.137	1.134	1.302	1.198	1.27	1.36	1.03	1.004	1.014	1.009	1.012	1.013	1.036	1.04
F	1.023	1.034	1.036	1.032	1.183	1.3	1.082	1.303	1.146	2.664	1.054	1.079	1.071	1.061	1.086	1.068	1.129	1.128
P	1.063	1.074	1.074	1.076	1.344	1.475	1.409	1.561	1.455	3.624	1.086	1.084	1.086	1.07	1.099	1.081	1.17	1.173
Pj	1.063	1.074	1.074	1.076	1.345	1.486	1.432	1.565	1.462	3.837	1.087	1.095	1.093	1.076	1.108	1.087	1.179	1.181
T	-0.244	-0.062	-0.018	-0.133	0.135	0.351	-0.54	0.19	-0.275	0.522	0.29	0.899	0.665	0.746	0.745	0.679	0.552	0.509
U	-0.258	-0.08	-0.036	-0.151	0.062	0.264	-0.599	0.082	-0.36	0.268	0.271	0.895	0.654	0.739	0.734	0.669	0.524	0.479
K1d	224.8	221.3	225	38.7	67.5	67.3	244.2	123	79.7	205.3	213.8	355.5	341.3	210.9	302.9	316.5	194.4	184.5
K1i	1.2	5	0.4	2.8	9.6	15.8	31.8	33.5	24.2	35.6	8.7	10.4	15.9	22.7	31.6	30.4	45.8	54.8
K2d	320.1	343.9	343.8	326.9	193.6	183	96.7	213.3	298.8	87.1	314.8	258	245	311.5	210.5	217.6	42.9	31.1
K2i	77.3	80.8	89.2	86.9	47.6	56.9	53.7	0.4	60	33.5	51.4	35.5	21.3	23.9	3.9	14.7	40.5	32.3
K3d	134.5	130.6	135	308.7	165.9	328.5	344.3	303.9	177.5	327.6	117.1	99.4	105.3	82.6	114.3	105.3	300	293
K3i	12.6	7.7	0.7	1.3	40.8	28.2	15.7	56.5	16.7	36.7	37.3	52.5	63	56	58.1	55.5	14.7	12.6
MFd	314.5	310.6	315	128.7	345.9	148.5	164.3	123.9	357.5	147.6	297.1	279.4	285.3	262.6	294.3	285.3	120	113
MFi	77.4	82.3	89.3	88.7	49.2	61.8	74.3	33.5	73.3	53.3	52.7	37.5	27	34	31.9	34.5	75.3	77.4
AK1	1.2	5	0.4	2.8	12.7	18	33.2	89.2	25.3	46.5	11	17.3	37.1	43.6	82.7	63.4	47.8	56.8

TABLE E.4: (Continued)

Unit: Sample: ID	GRT EH27 C	GRT EH27 D	GRT EH29 A	GRT EH29 B	GRT EH29 C	GRT EH29 D	GRT EH29 E	GRT EH29 F	GRT EH31 A1	GRT EH31 A2	GRT EH31 A3	GRT EH31 B1	GRT EH31 B2	GRT EH31 B3	GRT EH32 A	GRT EH32 B	GRT EH32 C	GRT EH32 D
Km	2.67E-02	2.97E-02	1.90E-02	1.75E-02	1.73E-02	1.78E-02	1.76E-02	1.58E-02	3.60E-02	3.45E-02	4.30E-02	-7.91E-05	1.01E-06	1.17E-06	9.45E-07	9.16E-07	1.00E-06	9.78E-07
L	1.037	1.042	1.038	1.03	1.019	1.018	1.024	1.025	1.014	1.014	1.017	1.843	1.377	1.049	1.005	1.266	1.167	1.212
F	1.125	1.118	1.116	1.135	1.138	1.138	1.138	1.13	1.053	1.048	1.044	-0.286	1.161	1.121	1.637	1.351	1.257	1.323
P	1.167	1.165	1.159	1.159	1.159	1.158	1.166	1.158	1.068	1.063	1.062	-0.528	1.599	1.176	1.644	1.711	1.467	1.604
Pj	1.175	1.171	1.165	1.181	1.175	1.173	1.179	1.17	1.071	1.066	1.064	0	1.616	1.181	1.644	1.713	1.471	1.609
T	0.532	0.461	0.492	0.617	0.745	0.759	0.692	0.668	0.582	0.543	0.447	0	-0.364	0.41	0.982	0.121	0.192	0.185
U	0.504	0.43	0.464	0.592	0.728	0.744	0.672	0.648	0.571	0.532	0.435	0.684	-0.462	0.376	0.977	-0.012	0.099	0.07
Kid	195.9	187.1	196.1	197.3	198.7	200.8	200.1	197.8	54.4	55.9	52	50.2	135.3	211.8	33.9	40.1	181.8	115.4
Kih	45.1	53.5	51.1	43.9	33.1	35	59.8	53.9	7.8	7.8	7.4	24.1	24.2	49.2	31.9	36.8	2.3	47.1
K2d	40.6	29.9	36	32.3	33	32.5	31.2	29.6	16.5	16.9	163.2	209.4	305	317.1	170	219.9	87.4	240.7
K2i	42.2	34.3	37.1	45.1	56	54.5	29.8	35.5	71.8	70.8	319.5	64.4	65.5	12.9	49.2	53.2	61.7	28.3
K3d	298.9	292.3	298.5	294.6	293.1	294.6	298.4	295.5	322.2	323.4	316.6	316.6	43.5	57.3	288.8	310	273	348.3
K3i	12.7	11	9.8	7.5	6.6	5.5	4.8	5.6	16.5	17.4	18.1	8	3.9	37.9	22.7	0.1	28.2	29.3
Mfd	118.9	112.3	118.5	114.6	113.1	114.6	118.4	115.5	142.2	143.4	139.5	136.6	223.5	237.3	108.8	130	93	168.3
Mfi	77.3	79	80.2	82.5	83.4	84.5	85.2	84.4	73.5	72.6	71.9	82	86.1	52.1	67.3	89.9	61.8	60.7
AK1	46.5	55	52.2	44.4	33.4	35.2	60.1	54.3	7.7	8.2	7.8	24.3	24.2	73.6	34.9	36.8	2.7	57.1
Unit: Sample: ID	GRT EH32 E	GRT EH32 F	GRT EH26 26A	GRT EH26 26B	GRT EH26 26D	GRT EH26 26E	GRT EH38 38A	GRT EH38 38B	GRT EH38 38C	GRT EH38 38D	GRT EH38 38E	GRT EH38 38F	GRT EH40 40A	GRT EH40 40B	GRT EH40 40C	GRT EH40 40D	GRT EH40 40E	GRT EH41 41A
Km	8.93E-07	1.02E-06	4.73E-03	2.71E-03	5.94E-03	4.59E-03	9.61E-03	9.63E-03	1.08E-02	9.72E-03	9.32E-03	1.03E-02	1.20E-02	1.12E-02	1.21E-02	9.32E-03	7.60E-03	9.28E-03
L	1.11	1.532	1.037	1.045	1.035	1.044	1.196	1.18	1.188	1.148	1.16	1.177	1.136	1.115	1.128	1.112	1.093	1.054
F	1.487	1.414	1.056	1.031	1.081	1.053	1.122	1.132	1.143	1.156	1.136	1.15	1.212	1.191	1.167	1.161	1.16	1.131
P	1.65	2.166	1.095	1.077	1.119	1.099	1.342	1.336	1.357	1.326	1.318	1.354	1.377	1.328	1.316	1.291	1.268	1.192
Pj	1.696	2.169	1.096	1.077	1.122	1.099	1.345	1.337	1.358	1.326	1.318	1.354	1.377	1.328	1.316	1.292	1.271	1.198
T	0.584	-0.104	0.199	-0.178	0.385	0.085	-0.218	-0.141	-0.128	0.024	-0.076	-0.076	0.203	0.234	0.125	0.167	0.251	0.398
U	-0.291	-0.291	0.177	-0.196	0.361	0.062	-0.287	-0.211	-0.202	-0.046	-0.144	-0.151	0.125	0.167	0.057	0.104	0.195	0.361
Kid	171.8	197	272.9	260.8	262.8	262.2	34.9	32	43.4	49.7	38	47.4	166.2	89.7	83.6	106.1	74.7	81
Kih	69.7	0.7	40.6	39.8	35.3	35.3	79.3	83.9	79.8	87.2	82.2	82.8	81.1	79.3	76.4	86.3	77.6	88.3
K2d	42.3	306.9	24.4	28.1	9.8	15	226.9	222.2	225.7	226.1	222.8	229.4	35.9	206.6	207.5	219	215.6	226.5
K2i	13.2	87.9	23.2	41.4	19.3	28.7	10.4	6	10.2	2.8	7.8	7.2	5.8	4.9	7.7	1.5	9.7	1.4
K3d	308.7	107	135.8	147.9	119.5	133.7	136.5	132.1	135.6	136.1	132.7	139.4	305.2	297.4	299.1	309.1	306.9	316.5
K3i	15.1	2	40.5	29.4	43.9	41.3	2.2	1.1	0.4	0.2	0.6	0.3	6.8	9.5	11.2	3.4	7.7	1
Mfd	128.7	287	315.8	327.9	299.5	313.7	316.5	312.1	315.6	136.1	312.7	319.4	125.2	117.4	119.1	129.1	126.9	136.5
Mfi	74.9	88	49.5	60.6	46.1	48.7	87.8	88.9	89.6	89.8	89.7	89.7	83.2	80.5	78.8	86.6	82.3	89
AK1	76.3	0.7	58.9	40.6	62.7	50.3	79.6	84	79.8	87.2	82.2	82.8	84.2	85.1	82.1	88.5	80.2	88.6
Unit: Sample: ID	FCI2 EH41 B	FCI2 EH41 C	FCI2 EH41 D	FCI2 EH41 E	FCI2 EH41 F													
Km	8.91E-03	8.46E-03	6.97E-03	9.73E-03	8.57E-03													
L	1.053	1.034	1.032	1.02	1.03													
F	1.158	1.145	1.143	1.155	1.134													
P	1.219	1.183	1.179	1.179	1.167													
Pj	1.228	1.195	1.191	1.197	1.179													
T	0.481	0.607	0.62	0.757	0.622													
U	0.443	0.58	0.594	0.739	0.598													
Kid	176.6	107.1	120.7	204.3	192.9													
Kih	80	89.5	87	78	84													
K2d	47.4	224.8	225.4	46.7	46.4													
K2i	6.3	0.2	0.8	11.1	5													
K3d	316.6	314.8	315.5	315.9	316.1													
K3i	7.7	0.4	2.9	4.5	3.3													
Mfd	136.6	134.8	135.5	135.9	136.1													
Mfi	82.3	89.6	87.1	85.5	86.7													
AK1	83.6	89.8	89.2	78.9	85													

TABLE E.5: AMS measurements of the Takidani Pluton (samples from Satoru Harayama

Unit: Sample: ID	IGT Np0 A	IGT Np0 B	IGT Np9 A	IGT Np1 A	IGT Np1 B	IGT Np1 C	IGT Np2 A	IGT Np2 B	IGT Np2 C	IGT Np4 D	IGT Np4 2	IGT Np5 2	IGT Np5 2.2	IGT Np6 3	IGT Np6 3A	IGT Np6 3B	IGT Np6 3C
Km	6.66E-04	2.36E-03	1.80E-03	3.49E-02	3.66E-02	2.92E-02	2.19E-02	2.13E-02	2.03E-02	2.05E-02	2.11E-02	2.04E-02	2.09E-02	2.22E-02	1.83E-02	1.84E-02	1.72E-02
L	1.017	1.167	1.044	1.039	1.05	1.033	1.027	1.029	1.039	1.031	1.053	1.061	1.064	1.009	1.072	1.059	1.052
F	1.055	1.518	1.385	1.068	1.053	1.074	1.084	1.073	1.106	1.069	1.121	1.132	1.122	1.042	1.018	1.018	1.021
P	1.074	1.771	1.446	1.11	1.105	1.109	1.114	1.104	1.102	1.102	1.182	1.201	1.193	1.052	1.091	1.079	1.074
Pj	1.077	1.807	1.496	1.111	1.105	1.111	1.118	1.107	1.103	1.104	1.187	1.206	1.196	1.056	1.097	1.082	1.076
T	0.512	0.46	0.766	0.272	0.028	0.375	0.507	0.429	0.206	0.369	0.378	0.353	0.325	0.632	-0.594	-0.522	-0.428
U	0.499	0.343	0.726	0.248	0.003	0.353	0.487	0.409	0.183	0.348	0.342	0.256	0.285	0.624	-0.608	-0.536	-0.443
K1d	211.2	61.4	43.9	194.3	224.7	255.4	222.1	217.9	216.9	233.6	159.4	169	338.1	312	39.2	44.9	33.6
K1i	6.9	13.8	10.7	85.9	75.5	82.3	25.8	13.9	18.3	13.6	79.8	84.5	88.8	63.2	43.6	34.1	43
K2d	43.2	323.6	310.8	41.5	40	30.2	21	62	26.5	71.5	28.1	33.6	31.9	208	142	278.7	42.1
K2i	83	28.7	15.6	3.6	14.4	5.4	62.7	73.8	7.15	74.2	6.8	3.9	0.8	10.2	44.6	41	42.1
K3d	301.4	174.2	167	311.4	130.3	120.7	127.9	125.8	125.9	128.4	297.2	303.4	117.9	236.7	299.5	158.2	140.2
K3i	1.4	57.6	70.9	1.9	1.1	5.4	8.5	8.2	3.1	8	7.6	3.9	0.9	24.5	10	30.3	17
MFd	121.4	354.2	347	131.4	310.3	300.7	307.9	305.8	305.9	308.4	117.2	123.4	297.9	56.7	119.5	338.2	320.2
MFi	88.6	32.4	19.1	88.1	88.9	84.6	81.5	81.8	86.9	82	82.4	86.1	89.1	65.5	80	59.7	73
AK1	6.9	26.5	34.7	86.4	75.6	84.6	26.1	14	18.3	13.7	83.2	86	89.2	78.8	44.5	40.5	45.5

Unit: Sample: ID	HA Np7 2C	HA Np7 3B	IGT Np9 A	IGT Np9 B	IGT Np12 A	IGT Np12 B	IGT Np12 C	IGT Np0 D	IGT KH816-1 A	IGT KH816-1 B	IGT KH816-2 A	IGT KH816-2 B	IGT KH816-2 C	IGT 911009-1 A	IGT 911009-1 B	IGT 911009-1 C	GDT 911009-2
Km	1.50E-02	1.86E-02	2.84E-02	2.93E-02	6.38E-03	8.53E-03	1.05E-02	1.05E-02	4.44E-02	3.80E-02	1.15E-02	1.47E-02	1.38E-02	1.74E-02	1.89E-02	2.25E-02	2.62E-02
L	1.116	1.137	1.015	1.012	1.011	1.008	1.024	1.026	1.062	1.065	1.055	1.041	1.03	1.053	1.072	1.051	1.013
F	1.252	1.29	1.063	1.073	1.091	1.08	1.078	1.079	1.034	1.037	1.068	1.079	1.072	1.02	1.012	1.023	1.032
P	1.398	1.467	1.079	1.086	1.103	1.088	1.104	1.107	1.098	1.087	1.112	1.109	1.103	1.075	1.085	1.075	1.045
Pj	1.407	1.477	1.084	1.093	1.114	1.098	1.111	1.111	1.1	1.091	1.095	1.113	1.106	1.077	1.092	1.077	1.047
T	0.342	0.331	0.615	0.705	0.777	0.819	0.519	0.499	-0.287	-0.517	-0.19	0.24	0.465	-0.443	-0.7	-0.375	0.424
U	0.267	0.244	0.603	0.694	0.767	0.812	0.501	0.48	-0.309	-0.532	-0.212	0.215	0.386	-0.457	-0.71	-0.39	0.415
K1d	237.9	246.5	25.9	18.1	272.2	358.9	319.3	308.6	356.7	2.8	8.2	13.7	9.8	25.7	33.3	29.6	19.5
K1i	87.5	87.4	14.7	8.8	29	20.7	44.7	48.4	61.3	44.8	39	21.7	36	24.4	16.5	24.8	48.9
K2d	33.7	29.7	136.6	120.5	7.9	254.3	218.2	214.5	243.2	247.7	248.8	263.5	251.3	228.8	229.6	230.9	182.4
K2i	2.3	2.1	53.5	54	10.1	33.8	11	3.7	12.3	23.1	31.2	41	33.7	63.8	72.9	63.6	39.8
K3d	123.7	119.8	286.3	281.9	115.1	114.5	117.7	121.2	147.2	139.5	133.3	124.1	127.2	119.8	124.7	123.5	279.6
K3i	1	1.6	32.6	34.5	59	48.7	43.2	45.5	25.5	36.3	35.4	41.1	34.6	9.1	4.5	8.4	8.6
MFd	303.7	299.8	106.3	101.9	295.1	294.5	301.2	301.2	327.2	319.5	313.3	304.1	311.7	299.8	304.7	303.5	99.6
MFi	89	88.4	57.4	55.5	31	41.3	46.8	48.7	64.5	53.7	54.6	48.9	53.5	80.9	85.5	81.6	81.4
AK1	87.7	87.9	17.5	10.8	70.1	32.4	74.8	85.1	76.3	60.9	50.5	29.4	46.4	24.7	16.5	25.1	49.6

Unit: Sample: ID	GDT 911009-2 2.2	GDT 911009-4 1B	GDT 911009-4 2D	GDT 911009-4 A	GDT 911010-2 A	GDT 911010-2 B	GDT 911010-2 C	GDT 911010-2 D
Km	2.31E-02	2.74E-02	2.87E-02	3.16E-02	1.82E-02	1.77E-02	1.65E-02	1.69E-02
L	1.005	1.01	1.011	1.016	1.049	1.041	1.035	1.033
F	1.038	1.177	1.177	1.172	1.105	1.103	1.101	1.094
P	1.043	1.19	1.179	1.119	1.159	1.148	1.14	1.13
Pj	1.047	1.215	1.202	1.212	1.162	1.153	1.145	1.135
T	0.775	0.881	0.867	0.819	0.353	0.417	0.477	0.473
U	0.771	0.871	0.857	0.804	0.321	0.388	0.451	0.449
K1d	27.3	211.3	139.5	218.7	227.5	236	228.2	227.2
K1i	43.6	34.8	85.9	13.7	75.7	74.7	61.7	64.7
K2d	160.1	36	24.3	65.3	41.1	36	47.9	43
K2i	35.5	55.1	1.8	74.8	14.2	14.4	28.3	25.2
K3d	270.1	302.9	294.1	310.3	131.5	127.3	138	133.7
K3i	25.7	2.2	6.6	1.5	5	5	0.1	1.6
MFd	90.1	122.9	114.1	130.3	311.5	307.3	318	313.7
MFi	64.3	87.8	86.3	88.5	88.5	89.9	88.4	87.8
AK1	49.9	34.8	88.2	13.8	75.6	75.6	61.7	64.8

TABLE E.6: Field measurements of magnetic susceptibility

ID:	233	234	235	236	237	238	239	240	241	242	243	244	245	246	247	248	249	250
Latitude:	36.2938	36.2936	36.2936	36.2937	36.2937	36.2937	36.2936	36.2936	36.2936	36.2936	36.2935	36.2935	36.2938	36.2938	36.2938	36.2938	36.2940	36.2941
Longitude:	137.6300	137.6306	137.6306	137.6319	137.6319	137.6319	137.6319	137.6310	137.6301	137.6301	137.6300	137.6300	137.6294	137.6294	137.6294	137.6294	137.6293	137.6295
Elevation:	2099	2182	2182	2203	2203	2203	2186	2186	2113	2113	2092	2092	2051	2051	2051	2051	2043	2056
Km (avg)	2.24E-02	3.83E-02	3.36E-02	1.91E-02	1.59E-02	1.89E-02	1.85E-02	2.23E-02	1.62E-02	2.68E-02	1.98E-02	2.03E-02	7.46E-04	7.00E-04	3.59E-03	7.77E-04	1.32E-02	1.35E-02
Km (std)	2.18E-03	5.54E-03	6.42E-03	4.70E-03	6.82E-03	5.99E-03	1.60E-03	1.60E-03	1.38E-03	2.08E-03	1.19E-03	2.51E-03	5.69E-05	8.38E-05	4.41E-04	1.05E-04	1.62E-03	1.96E-03
Km (max)	2.69E-02	4.78E-02	4.72E-02	2.66E-02	2.67E-02	2.27E-02	2.54E-02	2.25E-02	1.89E-02	2.93E-02	2.31E-02	2.50E-02	8.61E-04	9.60E-04	4.84E-03	9.94E-04	1.72E-02	1.73E-02
Km (min)	1.71E-02	3.18E-02	2.73E-02	9.18E-03	3.79E-03	1.05E-02	1.54E-02	1.86E-02	1.37E-02	2.41E-02	1.80E-02	1.68E-02	6.17E-04	5.76E-04	2.70E-03	5.73E-04	8.13E-03	9.35E-03
ID:	251	252	253	254	255	256	257	258	259	260	261	262	263	264	265	266	267	268
Latitude:	36.2940	36.2938	36.2938	36.2930	36.2930	36.2930	36.2936	36.2936	36.2936	36.2936	36.3000	36.2997	36.2976	36.2976	36.2937	36.2937	36.2935	36.2935
Longitude:	137.6294	137.6293	137.6293	137.6276	137.6276	137.6276	137.6138	137.6138	137.6138	137.6138	137.6149	137.6151	137.6181	137.6181	137.6239	137.6239	137.6244	137.6244
Elevation:	2059	2040	2040	1896	1896	1896	1549	1549	1549	1549	1568	1573	1573	1805	1805	1805	1816	1816
Km (avg)	1.50E-02	2.57E-03	2.59E-03	1.51E-02	1.95E-02	1.84E-02	1.80E-02	2.58E-02	5.45E-02	3.06E-02	1.57E-02	1.65E-02	1.56E-02	2.95E-02	2.17E-02	2.35E-02	2.72E-02	2.46E-02
Km (std)	3.02E-03	7.31E-04	6.67E-04	1.40E-03	1.34E-03	2.49E-03	2.61E-03	2.27E-03	2.28E-02	3.39E-03	1.55E-03	1.64E-03	2.88E-03	1.12E-02	1.66E-03	2.71E-03	2.28E-03	2.87E-03
Km (max)	2.20E-02	3.93E-03	3.86E-03	1.74E-02	2.27E-02	2.46E-02	2.29E-02	3.13E-02	8.55E-02	3.68E-02	1.86E-02	2.01E-02	4.88E-02	4.88E-02	2.62E-02	2.90E-02	3.17E-02	2.94E-02
Km (min)	9.85E-03	1.59E-03	1.60E-03	1.79E-02	1.75E-02	1.47E-02	2.16E-02	2.16E-02	2.91E-02	2.35E-02	1.13E-02	1.17E-02	1.03E-02	1.39E-02	1.83E-02	1.94E-02	2.00E-02	1.93E-02
ID:	269	270	271	272	273	274	275	276	277	278	279	280	281	282	283	284	285	286
Latitude:	36.2934	36.2934	36.2934	36.2934	36.2934	36.2933	36.2933	36.2933	36.2933	36.2933	36.2931	36.2931	36.2931	36.2931	36.2931	36.2931	36.2931	36.2931
Longitude:	137.6246	137.6246	137.6246	137.6246	137.6246	137.6247	137.6247	137.6247	137.6248	137.6248	137.6254	137.6254	137.6254	137.6255	137.6255	137.6255	137.6257	137.6257
Elevation:	1820	1820	1820	1820	1820	1827	1827	1827	1828	1828	1849	1849	1849	1857	1857	1857	1861	1861
Km (avg)	2.22E-02	2.14E-02	2.08E-02	2.88E-02	3.06E-02	2.45E-02	2.48E-02	2.60E-02	2.20E-02	2.16E-02	2.25E-02	2.32E-02	2.21E-02	1.81E-02	1.68E-02	1.93E-02	2.28E-02	2.27E-02
Km (std)	1.15E-03	1.05E-03	1.16E-03	1.68E-03	1.38E-03	1.35E-03	1.35E-03	1.77E-03	1.35E-03	1.85E-03	1.43E-03	1.37E-03	1.65E-03	9.05E-04	1.22E-03	2.14E-03	1.16E-03	1.17E-03
Km (max)	2.51E-02	2.37E-02	2.30E-02	2.21E-02	2.76E-02	2.76E-02	2.92E-02	2.59E-02	2.72E-02	2.66E-02	2.64E-02	2.66E-02	2.66E-02	2.03E-02	1.89E-02	2.34E-02	2.62E-02	2.59E-02
Km (min)	1.99E-02	1.87E-02	1.83E-02	2.49E-02	2.27E-02	2.18E-02	2.14E-02	2.29E-02	1.92E-02	1.79E-02	1.95E-02	2.04E-02	1.88E-02	1.61E-02	1.34E-02	1.54E-02	2.03E-02	2.05E-02
ID:	287	288	289	290	291	292	293	294	295	296	297	298	299	300	301	302	303	304
Latitude:	36.2932	36.2932	36.2932	36.2932	36.2932	36.2932	36.2931	36.2931	36.2931	36.2931	36.2929	36.2931	36.2931	36.2918	36.2918	36.2918	36.2918	36.2918
Longitude:	137.6257	137.6258	137.6258	137.6261	137.6261	137.6261	137.6261	137.6263	137.6263	137.6263	137.6266	137.6268	137.6268	137.6268	137.6268	137.6268	137.6268	137.6268
Elevation:	1867	1868	1868	1873	1873	1870	1870	1878	1878	1878	1879	1891	1891	1472	1472	1472	1472	1472
Km (avg)	2.25E-02	2.06E-02	2.01E-02	2.24E-02	2.10E-02	1.97E-02	1.63E-02	2.02E-02	1.94E-02	2.13E-02	2.18E-02	2.42E-02	2.44E-02	2.70E-04	1.71E-03	6.14E-04	1.58E-04	3.04E-05
Km (std)	1.01E-03	2.05E-03	1.63E-03	2.11E-03	2.13E-03	1.24E-03	4.50E-03	1.53E-03	6.72E-04	1.52E-03	1.48E-03	1.13E-03	1.72E-03	5.06E-05	3.73E-04	4.33E-05	1.14E-03	2.40E-05
Km (max)	2.46E-02	2.60E-02	2.39E-02	2.21E-02	2.26E-02	2.24E-02	2.21E-02	2.34E-02	1.98E-02	2.61E-02	2.49E-02	2.75E-02	2.80E-02	3.74E-04	2.22E-03	6.89E-04	1.78E-04	3.06E-03
Km (min)	2.07E-02	1.63E-02	1.65E-02	1.91E-02	1.71E-02	1.57E-02	7.35E-03	1.72E-02	1.89E-02	1.80E-02	1.97E-02	2.19E-02	2.03E-02	1.66E-04	9.73E-04	5.32E-04	1.32E-04	3.00E-03
ID:	305	306	307	308	309	310	311	312	313	314	315	316	317	318	319	320	321	322
Latitude:	36.2918	36.2918	36.2918	36.2918	36.2916	36.2877	36.2877	36.2877	36.2877	36.2877	36.2877	36.2877	36.2877	36.2877	36.2877	36.2876	36.2875	36.2875
Longitude:	137.6042	137.6042	137.6042	137.6042	137.6048	137.6076	137.6076	137.6076	137.6076	137.6076	137.6076	137.6076	137.6076	137.6076	137.6076	137.6078	137.6079	137.6079
Elevation:	1472	1472	1472	1472	1490	1600	1600	1598	1598	1598	1598	1598	1598	1598	1598	1599	1602	1602
Km (avg)	3.04E-03	2.08E-03	7.66E-03	5.67E-03	5.36E-03	3.57E-02	3.29E-02	3.06E-02	2.33E-02	2.39E-02	2.44E-02	2.85E-02	2.97E-02	1.94E-02	1.66E-02	2.52E-02	1.56E-02	1.97E-02
Km (std)	3.53E-04	3.53E-04	2.19E-05	1.41E-04	1.14E-03	3.69E-03	2.19E-03	3.27E-03	2.64E-02	2.34E-03	6.01E-03	1.45E-03	4.07E-03	2.53E-03	1.71E-03	2.34E-03	1.24E-03	2.63E-03
Km (max)	3.24E-03	2.52E-03	7.69E-03	5.78E-03	7.29E-03	4.82E-02	3.39E-02	3.58E-02	3.00E-02	3.14E-02	3.48E-02	3.05E-02	3.91E-02	2.41E-02	2.26E-02	2.99E-02	1.79E-02	2.99E-02
Km (min)	2.32E-03	1.54E-03	7.66E-03	5.42E-03	3.89E-03	3.12E-02	2.65E-02	2.36E-02	1.94E-02	2.09E-02	1.66E-02	2.60E-02	1.99E-02	1.46E-02	1.37E-02	2.12E-02	1.33E-02	1.58E-02
ID:	323	324	325	326	327	328	329	330	331	332	333	334	335	336	337	338	339	340
Latitude:	36.2875	36.2875	36.2868	36.2866	36.2866	36.2865	36.2864	36.2865	36.2865	36.2936	36.2936	36.2936	36.2936	36.2936	36.2937	36.2921	36.2921	36.2905
Longitude:	137.6080	137.6080	137.6091	137.6094	137.6095	137.6098	137.6098	137.6100	137.6100	137.6079	137.6079	137.6079	137.6079	137.6079	137.6079	137.6221	137.6221	137.6233
Elevation:	1605	1605	1646	1658	1661	1665	1669	1669	1669	1447	1447	1447	1447	1447	1446	1851	1851	1941
Km (avg)	1.47E-02	1.42E-02	1.09E-02	2.70E-02	3.28E-02	2.93E-02	2.47E-02	2.52E-02	9.37E-05	7.06E-03	9.54E-03	5.46E-03	2.36E-03	2.14E-03	1.04E-02	2.70E-03	2.57E-02	8.03E-03
Km (std)	1.01E-03	6.19E-04	1.41E-02	3.24E-02	4.42E-02	3.37E-02	2.88E-02	3.10E-02	8.74E-06	1.13E-04	1.11E-02	6.82E-03	4.30E-03	2.72E-04	1.87E-03	3.37E-02	3.01E-02	1.42E-03
Km (max)	1.66E-02	1.55E-02	1.33E-02	3.24E-02	4.58E-02	3.37E-02	2.88E-02	3.10E-02	1.13E-04	8.85E-03	1.11E-02	6.82E-03	4.30E-03	2.50E-03	1.41E-02	3.37E-02	3.01E-02	1.13E-02
Km (min)	1.28E-02	1.31E-02	6.66E-03	2.38E-02	2.55E-02	2.51E-02	2.14E-02	1.74E-02	7.90E-05	3.94E-03	6.93E-03	3.67E-03	9.23E-04	1.75E-03	7.19E-03	1.75E-02	2.25E-02	6.49E-03

ID:	341	342	343	344	345	346	347	348	349	350	351	352	353	354	355	356	357	358
Latitude:	36.2905	36.2884	36.2864	36.2864	36.2864	36.2918	36.2918	36.2918	36.3101	36.3101	36.3102	36.3102	36.3101	36.3104	36.3105	36.3109	36.3109	36.3109
Longitude:	137.6233	137.6282	137.6282	137.6282	137.6282	137.6225	137.6225	137.6225	137.6370	137.6370	137.6361	137.6361	137.6362	137.6356	137.6350	137.6346	137.6346	137.6346
Elevation:	1941	2115	2258	2258	2051	2049	2049	2049	1881	1881	1854	1854	1854	1843	1839	1819	1819	1819
Km (avg)	9.77E-03	1.15E-02	2.14E-02	1.64E-02	5.89E-03	1.13E-02	9.42E-03	6.96E-03	2.33E-02	2.84E-02	1.88E-02	1.97E-02	1.93E-02	1.07E-02	1.71E-02	1.95E-02	2.14E-02	1.58E-02
Km (std)	7.66E-04	1.20E-03	2.23E-03	4.51E-03	7.26E-04	2.01E-03	2.19E-03	7.56E-04	2.96E-03	3.63E-03	1.25E-03	1.87E-03	1.92E-03	8.16E-04	1.25E-03	2.31E-03	1.97E-03	4.30E-03
Km (max)	1.13E-02	2.71E-02	2.34E-02	2.34E-02	8.22E-03	1.50E-02	1.93E-02	8.87E-03	2.93E-02	3.63E-02	2.20E-02	2.35E-02	2.42E-02	1.21E-02	1.97E-02	2.60E-02	2.56E-02	2.37E-02
Km (min)	8.20E-03	8.23E-03	1.83E-02	5.37E-03	4.48E-03	7.30E-03	6.12E-03	5.99E-03	1.82E-02	2.13E-02	1.57E-02	1.64E-02	1.44E-02	8.77E-03	1.48E-02	1.52E-02	1.65E-02	9.15E-03
ID:	359	360	361	362	363	364	365	366	367	368	369	370	371	372	373	374	375	376
Latitude:	36.3117	36.2688	36.2790	36.2786	36.2786	36.2786	36.2785	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743
Longitude:	137.6327	137.6220	137.6291	137.6286	137.6286	137.6286	137.6286	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263
Elevation:	1819	2464	2909	2872	2872	2872	2873	2740	2740	2740	2740	2740	2740	2740	2740	2740	2740	2740
Km (avg)	3.34E-02	2.83E-02	3.08E-02	2.39E-02	2.40E-02	2.64E-02	1.65E-02	3.18E-02	2.70E-02	2.59E-02	3.36E-02	2.35E-02	3.00E-02	2.88E-02	2.13E-02	3.14E-02	3.25E-02	2.69E-02
Km (std)	1.75E-03	4.60E-03	2.29E-03	1.10E-03	2.17E-03	1.70E-03	1.97E-03	2.49E-03	2.58E-03	1.56E-03	8.82E-04	2.73E-03	1.89E-03	1.80E-03	2.48E-03	1.59E-03	1.90E-03	2.99E-03
Km (max)	3.87E-02	4.15E-02	3.49E-02	2.59E-02	2.73E-02	2.99E-02	2.94E-02	3.89E-02	3.30E-02	2.94E-02	3.53E-02	2.93E-02	2.38E-02	3.39E-02	2.47E-02	3.37E-02	3.72E-02	3.12E-02
Km (min)	2.97E-02	1.92E-02	2.40E-02	2.16E-02	2.07E-02	2.38E-02	1.14E-02	2.88E-02	2.26E-02	2.33E-02	3.20E-02	1.90E-02	2.45E-02	2.60E-02	1.63E-02	2.72E-02	2.88E-02	1.89E-02
ID:	377	378	379	380	381	382	383	385	386	387	388	389	390	392	393	394	395	396
Latitude:	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2743	36.2677	36.2677	36.2677	36.2677	36.2680	36.2680	36.2916	36.2916	36.2916	36.2916	36.2916
Longitude:	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6263	137.6100	137.6100	137.6100	137.6090	137.						