Constraining the duration of eruptions of the Rangitoto volcano, New Zealand, using paleomagnetism

Linda Aulin

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Abstract: The volcano Rangitoto belongs to the Auckland volcanic field and is situated north east of Auckland city, New Zealand. The volcanos of the Auckland volcanic field have previously thought to be of monogenetic origin, but recent studies indicate that Rangitoto has a long history of eruptions, which would suggest that the volcanic system may have entered a new phase. The last eruption is dated to 500-550 years BP (1950 AD), however there is little evidence to support the possibility that Rangitoto will not erupt again or that a new volcano on a similar scale will erupt elsewhere in the Auckland area. Reconstruction of past eruptions of the volcano is of great importance to prepare for future volcanic hazards in the area. Two hypotheses of the eruption rate are presented; I) the majority of the eruptions took place within 150 years, ca 650-500 BP, and II) the eruptions were scattered over a period of a few thousand years. To test these hypotheses, paleomagnetic analyses of samples of lava flows from Rangitoto were used to compare variations in the recorded geomagnetic field with models based on regional independently dated paleomagnetic data. Paleointensity experiments were conducted on eight samples of volcanic rocks from Rangitoto according to the Thellier technique. The experiments resulted in two acceptable results and six rejected samples due to multidomain behaviour and thermochemical alteration. The accepted results, combined with additional paleointensity measurements based on an alternating microwave technique as well as independently measured paleomagnetic directions, have been compared to a regional paleomagnetic dataset. I conclude that the data are not sufficient to reject either hypothesis. Further paleomagnetic experiments are suggested.

Keywords: Paleomagnetism, paleointensity experiment, Rangitoto, Auckland volcanic field, thermal demagnetization

Supervisors: Andreas Nilsson and Megan Allington

Subject: Quaternary Geology

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Bestämning av eruptionshistorien av vulkanen Rangitoto, Nya Zeeland, med hjälp av paleomagnetism

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Sammanfattning: Vulkanen Rangitoto tillhör Aucklands vulkaniska fält, och är belägen nordöst om Auckland, Nya Zeeland. Vulkanerna inom Aucklands vulkanfält har tidigare antagits vara monogenetiska, men nya studier visar på att Rangitoto har varit aktiv under en lång tid, vilket tyder på att vulkanen kan ha gått in i en ny fas. Det senaste utbrottet är daterat till 500-550 år BP (1950 AD), och det finns inget belägg för ifall det kommer ske igen, eller om en annan vulkan i samma storlek kommer att få utbrott någon annanstans inom regionen. Att undersöka Rangitotos eruptionshistoria och utveckling är av stor betydelse för att kunna förbereda och göra riskplaneringar inför kommande utbrott.

Två hypoteser fanns i åtanke under projektet: I) majoriteten av utbrotten ägde rum inom ett tidsspann på ca 150 år, ca 650–500 år BP, och II) utbrotten var spridda över en längre period på några tusen år. För att testa dessa hypoteser har paleomagnetiska studier gjorts på vulkaniska prover från Rangitotos lavaflöden och jämförts med modeller över regionala paleomagnetiska data. I denna studie har åtta prover från lavaflöden från Rangitoto undersökts enligt Thellier-teknik. Experimentet resulterade i två lyckade resultat, och sex avvisade prover pga. att korn uppvisade beteende av multidomänstruktur och kemisk förändring av mineralen under upphettning. De godkända resultaten, tillsammans med paleointensitetsmätningar framtagna genom microvågsteknik, går att passa in i båda hypoteserna, och är därför inte tillräckliga för att göra några trovärdiga slutsatser. Jag föreslår fortsatta studier av paleomagnetiska experiment för att säkerställa en hypotes.

Nyckelord: Paleomagnetism, paleointensitetsexperiment, Rangitoto, Auckland vulkanfält, termal avmagnetisering

Handledare: Andreas Nilsson och Megan Allington

Ämnesinriktning: Kvartärgeologi

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1 Introduction

Rangitoto belongs to the Auckland volcanic field (AVF) (also called Auckland monogenetic field). The origins of these monogenetic volcanoes are highly debated. Some claim that the volcanoes of the AVF evolved from a mantle plume (Heming 1980), with no magma reservoir in the crust, whereas others suggest that high U-Th isotope ratios and low upwelling rates of melt implies a source region for the melt in the upper mantle (Huang et al. 1997). A shared property of volcanoes in the AVF is that basaltic magma quickly rises through the crust. The AVF is typically considered to be “monogenetic”, which means that every edifice has its own conduit and that each volcano was active from days to no more than a year (Haraldur Sigurðsson et al. 1999; Shane et al. 2013). Rangitoto is the largest and youngest of the AVF volcanoes, and comprise about 50 percent of the volcanic material in the area (Smith et al. 1993).

Unlike other volcanos in the AVF, new studies indicate that Rangitoto could have erupted over a period of about 1000 years, which means that it could not be considered a “monogenetic” volcano (Shane et al. 2013). This means that the volcanic activity in the field must have undergone some change, or that the concept of the monogenetic volcanism in the AVF is misunderstood. The AVF has always posed a threat to the society of Auckland, but now this new type of volcanism must be in consideration to be able to forecast future volcanic hazards in the area and it is important to gain an understanding of how different volcanic structures behave to be able to interpret and plan for future hazards.

Thermal remanent magnetizations (TRM) carried by volcanic rocks can provide information about the direction and intensity of the ancient geomagnetic field (Korte & Constable 2006). Our knowledge about the evolution of the geomagnetic field on millennial time scales comes primarily from indirect measurements of natural remanent magnetizations (NRMs), e.g. TRM, in geological and archaeological materials, so-called paleomagnetic data. Historic records of direct observations of the geomagnetic field directions exist for the past four centuries but it was not until 1832 that Carl Freidrich Gauss developed the method for obtaining direct measurements of absolute field intensity (Valet 2003).

It is important to understand the geomagnetic field and how it varies in time for a better understanding of geodynamic processes in the core and the physical processes that control variations on a larger scale. Models of the geomagnetic field, developed for these types of deep Earth studies, can be constructed using paleomagnetic data from archaeological artefacts, volcanic rocks and marine and lacustrine sediments (Korte & Constable 2005; Korte & Constable 2006). Such models could also be used to indirectly date volcanic rocks by correlating paleomagnetic measurements with model predictions for the coordinates of the volcano.In February 2014, a drilling core was sam-

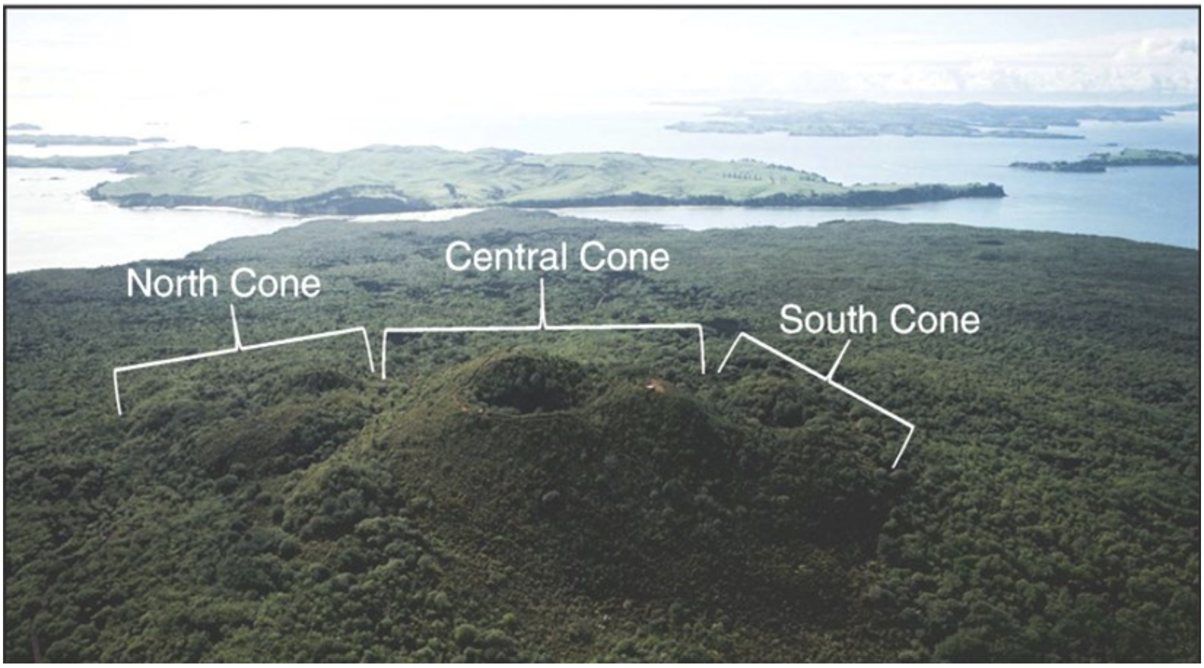

Fig. 1. overview of the scoria cones on top of Rangitoto, situated 8 km northeast of Auckland, New Zealand. Rangitoto belongs to the Auckland volcanic field which is comprised of about 50 volcanoes. Figure: A.J Needham et al. /Journal of Volcanology and Geothermal Research 201 (2001) 126.142

pled from Rangitoto by the University of Auckland. The core was drilled on the upper flank of the volcano, from 150 m down to below sea level. 128 m basaltic rock was collected of at least 53 separate lava flows. The lava flows have a thickness of 0.5-7 m and overlay 8 m of marine sediments mixed with pyroclastic deposits consisting of phreatomagmatic ash and lapilli (Linnell et al. 2016). The pyroclastic deposits probably represent the subaquatic birth of the volcano (pers. com. Augustinus).

Eruptions began about 6000 B.P, and ^{14}C-dating suggest a major shield building phase around 650-500 BP. Lack of sediments and paleosols between the lava flows suggest a continuity of the eruptions during the 100- year time span. However, the conclusions from the radiocarbon dating is based on the youngest radiocarbon date from a complicated and inverted dataset comprising age reversals and generally scattered ages, which is presumably due to sediment mixing/ redeposition during the subaqueous birth of the volcano. (Linnell et al. 2016).

In contrast, ^{14}C-dating of tephra layers suggests a period of eruptions of at least 1000 years (Shane et al. 2013).

Based on the previous studies of Rangitoto, two hypotheses have been the background for this project:

(I) there was a series of eruptions during a shorter period of time, when the shield of Rangitoto was formed (dated to 650-500 BP).

(II) Rangitoto had a prolonged period of eruptions, scattered over more than 6000 years, and numerous small volcanoes formed and are enclosed within the Rangitoto shield.

The aim of this study is to provide more information to make a better and stronger age model of Rangitotos' eruption history. The material will complement existing radiocarbon dates and other paleomagnetic analyses. The information is to be used in further hazard planning in the Auckland region.

2 Study area

Rangitoto is situated 8 km north east of Auckland, New Zealand. It is the youngest of about 50 volcanoes that is connected to the active Auckland volcanic field and rises 260 m above sea level, and occupies an area of 2300 ha (Linnell et al. 2016)

The AVF is situated on an intraplate tectonic setting, which sits across a back-arc extension, in transition to a "regular" back-arc setting on the continental landmass of New Zealand (Horspool et al. 2006). It is located just behind the margin of the Hikurangi subduction zone, where the Pacific plate is subducting under the Australian plate (Walcott 1978). AVF display extensional faulting as a result from rifting from Gondwana. The basement consists of Mesozoic greywacke overlain by Cretaceous and Oligocene sediments (Spoerli & Eastwood 1997). Rangitoto consists mainly of pyroclastic material covered by a thin layer of lava (Linnell et al. 2016).

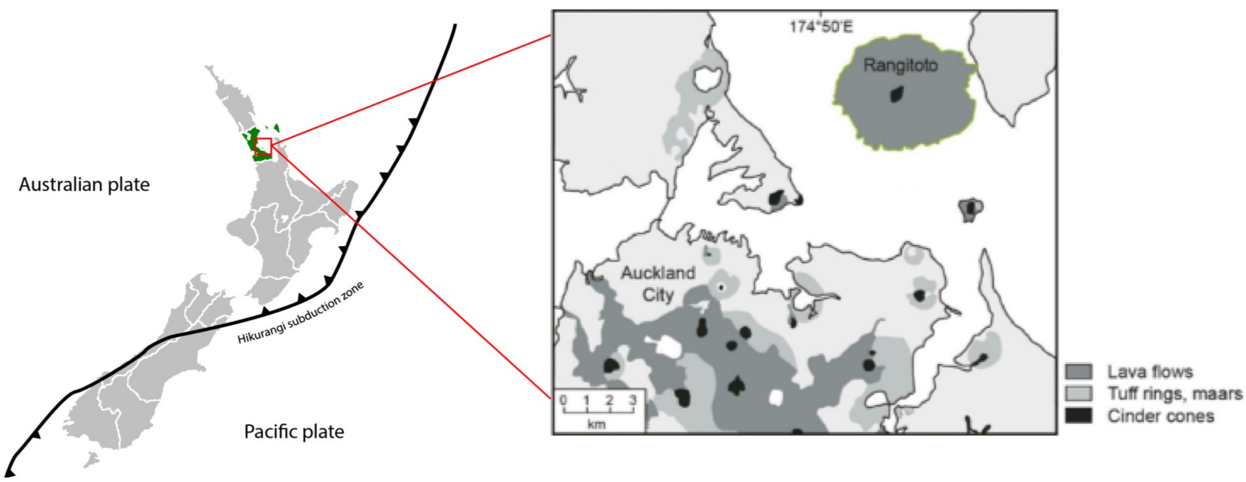

Fig. 2. Overview of New Zealand and map over the Auckland region and Rangitoto. Rangitoto is located 8 km northeast of Auckland, just behind the margin of the Hikurangi subduction zone. Figure: Andreas Nilsson, modified by Steven Kulicke

3 Theoretical background

A basic understanding of some magnetic terms is required in order to understand some of the principals of using paleomagnetism as a method for dating. Some important concepts will be explained in the following sections.

3.1 The geomagnetic field

The Earth magnetic field is generated through convection in the liquid iron-rich outer core. The geocentric axial dipole (GAD) concept is a model that is commonly used for describing the geomagnetic field, in which the field can be illustrated as a single magnetic dipole in the middle of the Earth aligned to the rotation axis (Butler 1992). The field lines of the magnetic field go from the magnetic South Pole to the magnetic north pole. The geomagnetic field fluctuates over time, which means that the magnetic north pole is moving. This is noticeable for example on the compass, which in Lund, Sweden, presently deviates by roughly +4° from geographic North. This angle of the horizontal component of the geomagnetic field direction is referred to as the magnetic declination. The angular deviation from the horizontal plane is called inclination (Fig. 3).

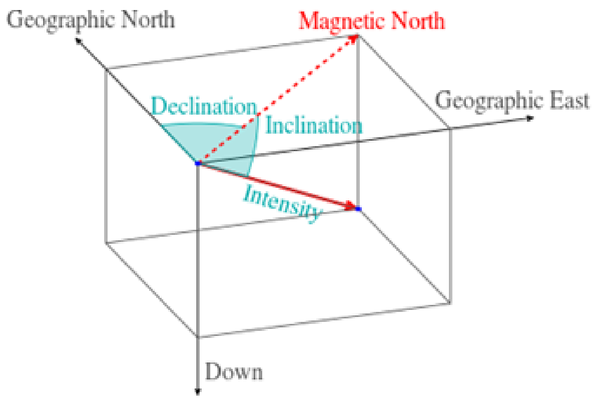

Fig.3. Schematic diagram of declination and inclination. Declination is the angle between magnetic north and geographic north, and inclination is the dip of the intensity from the magnetic north. Figure: Cymaera, cc by-sa 3.0, modified by Linda Aulin.

3.2 Paleomagnetic secular variations

The orientation of the geomagnetic field changes over time. Changes that occur between 1 to 10,000 years are called geomagnetic secular variations, and is similar over subcontinental regions (Butler 1992). The main reason for the magnetic field to change over time is flow-variation in the outer core (Constable & Constable 2013).

3.3 Natural remanent magnetiszation (NRM)

The NRM is the magnetization that is saved in a rock as it cools down below a specific temperature, e.g. 580°C for magnetite, which is its Curie temperature. It is dependent on the direction and strength of the local geomagnetic field at the time the rock formed, the mineralogic composition of the rock and secondary geological processes such as weathering. It is normally composed of a primary and one or more secondary components. A secondary NRM can be acquired in a number of ways; chemical alterations in the minerals, lightning strikes or long exposure to the geomagnetic field after formation. There are different kinds of NRM, for example thermoremanent magnetization, chemical remanent magnetization (CRM) and depositional remanent magnetization (DRM) (Butler 1992).

The magnetic remanence is preserved in ferromagnetic mineral grains in rocks and sediment. Magnetite, hematite and maghemite are the most important minerals that often only comprise a few percent of the rock, but provide the information needed for the paleointensity experiments (Gupta 2011). In this study, TRM will be examined.

3.4 Blocking temperature

The unblocking temperature is the temperature where the magnetization of a certain mineral gets randomized upon heating and vice versa blocked upon cooling (Valet 2003). The different components in the rock do not have the same blocking temperature, thus in each zero-field temperature step, only some components are demagnetized. These components will in the next temperature step within a laboratory magnetic field obtain a new NRM.

3.5 Absolute paleointensity measurements

Absolute paleointensity measurements can only be made on materials that have been exposed to the geomagnetic field and acquired a thermoremanent magnetisation at the time of formation, which can be reproduced in the laboratory. Materials that may be used for these measurements include fired archaeological materials, and igneous and metamorphic rocks that have been reheated (Gupta 2011).

3.6 Paleointensity experiments

Paleointensity experiments depends on a comparison between the NRM, and the TRM acquired in a known field in the lab (Valet 2003). This is explained by the equation;

$$\frac{H_{anc}}{H_{lab}} = \frac{M_{anc}}{M_{lab}} => H_{anc} = \frac{M_{anc}}{M_{lab}} \times H_{lab}$$

where M_{anc}= the intensity of the NRM, H_{anc}= the ancient magnetic field, M_{lab}= the new TRM from the lab and H_{lab}= the magnetic field in the lab. The components of NRM and TRM are successively removed and gained in certain temperature steps. The mineral grains with blocking temperature less than the present heating step will acquire a randomized magnetization in the zero field, and then gain a new magnetization in the presence of a known field. The results are presented in an Arai plot (see figure 4) and the final field value of the ancient magnetic field determined using the above equation (Butler 1992).

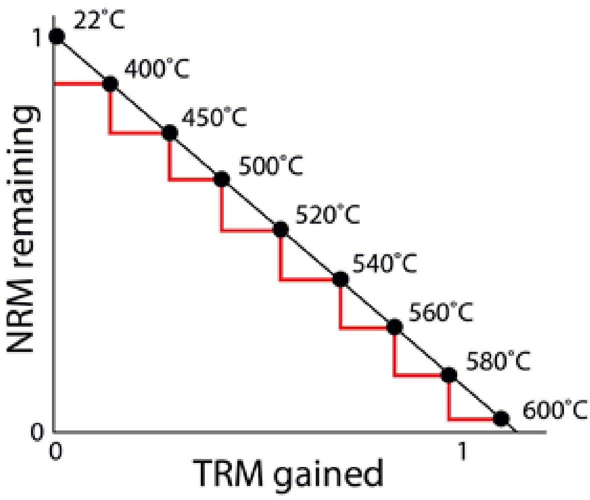

Fig. 4. The figure pictures an Arai plot, with pTRM checks. The heating started at 400 °C and continued till 600 °C. a pTRM check is made in a previous temperature step to make sure nothing has happened to the sample. in the figure the pTRM checks shown in red are overlapping the temperature steps, which means that there has been no alteration (nothing has changed since the last temperature step) and the test is therefore accepted. If the pTRM checks are not overlapping the next temperature step however, it is evidence for alternation of the grains. Figure: Steven Kulicke.

3.7 Alteration

Thermal demagnetization does not always get the best outcome, and have a generally low success rate. This depends mainly on thermochemical alteration during heating of samples. Not all rocks have the optimal grain-size spectra or are stable enough to produce reliable paleointensity measurements (Fabian & Leonhardt 2010). Alteration of the grains can up bring a new stable chemical remanent magnetization (CRM), which can be detected using pTRM checks (fig 4). pTRM-checks are carried out by repeating the experiment at a lower temperature and comparing with the initial results. If there is a significant difference, there grains have been altered, and further values obtained at higher temperature should be rejected (Coe et al. 1978).

3.8 Single and multidomain grains

The NRM in volcanic rocks suitable for paleointensity experiments should ideally be carried by very small (no more than tens of microns) so called single domain (SD), or pseudo-single domain (PSD) grains. As grains get coarser it is likely they get more than one magnetic domain due to a single domain becoming too unstable with size and splitting into two or more different magnetic domains, which will align themselves in opposite direction, and thereby cancel each other out and give smaller or zero net remanence. These are called multidomain (MD) grains, and they are less stable against alteration of the magnetization than single domains, and are therefore not suitable for paleointensity studies (Gupta 2011). In addition, MD grains can have different blocking and unblocking temperatures which cause a so-called "tail effect". This means that when the magnetic particles get activated during demagnetization, a different amount gets activated during remagnetization, which causes problems for the paleointensity experiments.

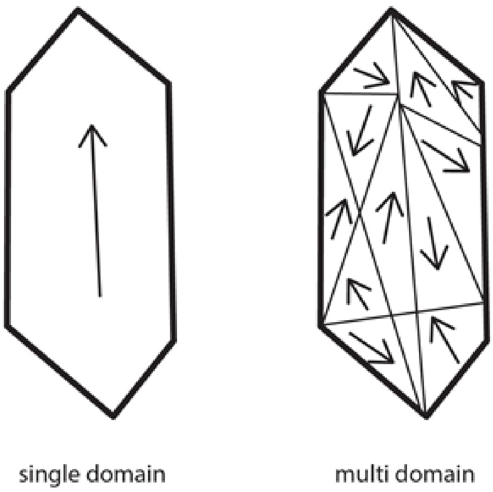

Fig. 5. A schematic figure of a single and a multidomain grain. the arrows represent the direction of the magnetic domains. Figure: Steven Kulicke.

5 Methods

Previous studies of Rangitoto include ^{14}C- dating, which have been problematic due to the fact that there was not enough organic material found in between the lava flows. Shell fragments were found in the deepest segments of the flows, derived from the volcanos subaquatic phase, although presumably redeposition of the sediments gave inverted ages and prevent reliable results, but indicate volcanic activity during a period of 150 years (Linnell et al. 2016). Also, ^{14}C-dating have been made on tephra particles from lake sediments, which indicates volcanic activity from ca 1500 BP to 500 BP (Shane et al. 2013).

The drill core was collected in February 2014 by Auckland University, by drilling through the upper flank of the volcano. The drilling resulted in a drill core of 128 m lava flows, separated by thin layers of basaltic breccia (Linnell et al. 2016). Of these 128 m, 8 samples representing ca 70 m has been examined in this study (Fig 6).

Paleointensity experiments have been conducted according to the Thellier technique (Paterson et al. 2016) using the IZZI protocol (Yu et al. 2004). The samples were subsampled and taken to Lund University for measuring. In order to get reference values, the susceptibility was first measured with a Kappabridge KLY-2, before any heating step was started. The demagnetization and measuring was carried out in the paleomagnetic laboratory at Lund University, with a 2G 760 Squid Superconducting Rock Magnetometer and a Magnetic Measurements Thermal Demagnetizer MMTD. The heating steps began at 400 °C in a zero field and continued in cycles with an infield of 50 μT, chosen to fit the expected paleointensity value. The measurements cycles were designed as: one cycle = zero field in T_1, infield in T_1, infield in T_2, zero field in T_2, pTRM test in T_1. The temperatures progressed from 400, 450, 500, 520, 540, 560, 580 and 600 °C. When the maximum temperature was reached during each step, the samples were kept at that temperature for 15 minutes to ensure that the entire sample reached desired temperature. During the experiments, pTRM checks were conducted in order to detect alteration. The susceptibility and magnetization was also measured after each step in a cycle in order to detect alteration.

The measurements have been made according to the SELCRIT2(MOD) criterions:

Fraction factor *(f)* is the "NRM fraction used for the best-fit on an Arai diagram" (Coe et al. 1978, cit. in Paterson et al. 2014). This shows how much of the initial NRM was used in calculation the paleointensity.

"The gap factor (*g*) is a measure of the average NRM lost between successive temperature steps of the segment chosen for the best-fit line on the Arai plot. The gap reflects the average spacing of the selected Arai plot points along the best-fit line." (Paterson et al. 2014a)

"The quality factor *(q)* is a measure of the overall quality of the paleointensity estimate and combines the relative scatter of the best-fit line, the NRM fraction and the gap factor." (Coe et al. 1978, cit. in Paterson et al. 2014)

"*(β)* is a measure of the relative data scatter around the best-fit line and is the ratio of the standard error of the slope to the absolute value of the slope". (Coe et al. 1978, cit. in Paterson et al. 2014)

(DRAT) is the "maximum absolute difference produced by a pTRM check, normalized by the length of the best-fit line" (Selkin et al. 2000, cit. in Paterson et al. 2014)

(MAD) is the "Maximum Angular Deviation of the anchored and free- floating, respectively, directional fits to the paleomagnetic vector on a vector component diagram". It is determined from the demagnetization steps. (Kirschvink 1980, cit. in Paterson et al. 2014)

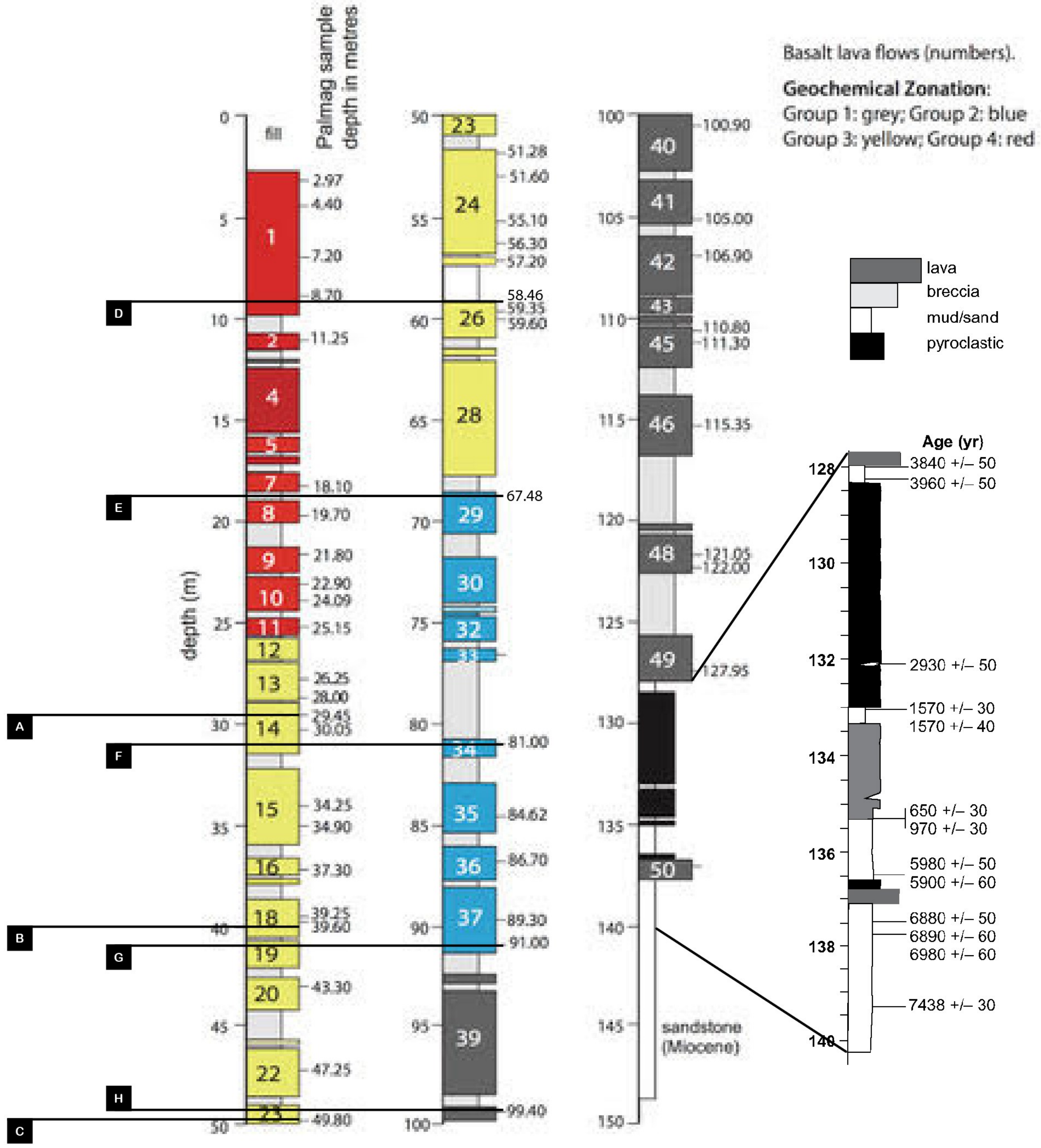

Fig. 6. The drill core log from Rangitoto, divided into geochemical zones and number of lava flows. A-H marks where the eight samples were collected. The sample depths range from 29.45 to 99.40 m and hence covers all of the zonations. To the right the ^{14}C-dating by Linell et al. is showed. Figure: based on Linnell 2016, modified by Steven Kulicke.

6 Results

The susceptibility measurements indicate some change after the first heating to 400 °C, but stay rather stable over the next heatings (see fig. 7). The change is indicative of alteration of the grains, but does not have to change the reliability of the results, if the ferromagnetic assemblage is unaltered.

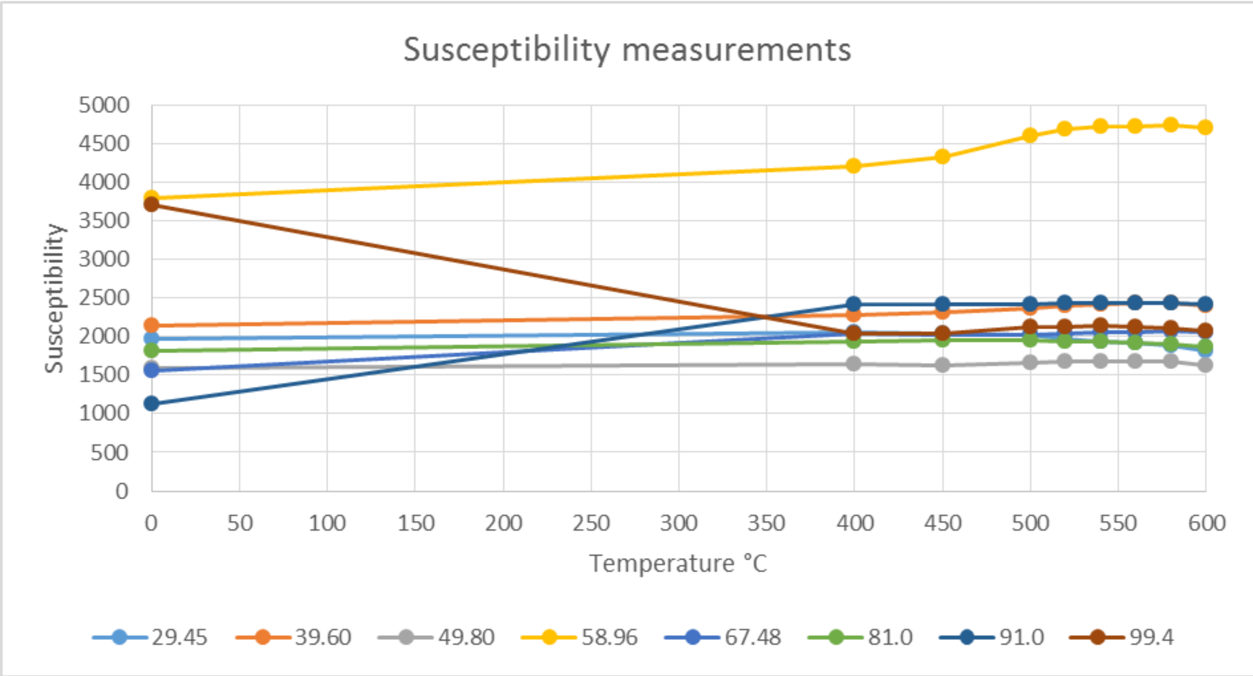

Fig. 7. Graph over the susceptibility measurements for each sample, named according to their depth in the drill core.

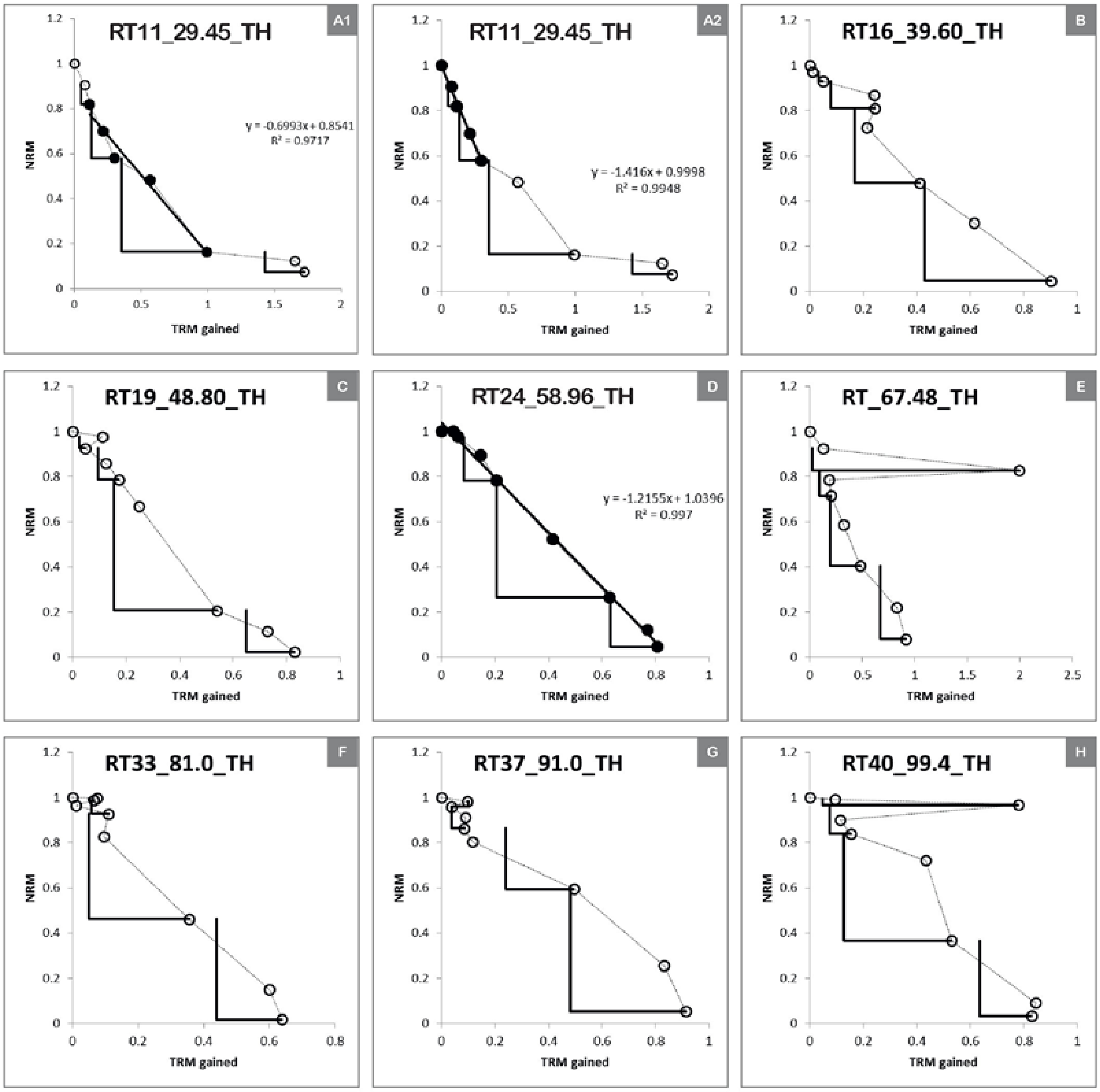

Fig. 8. Shows arai plots for all eight paleointensity experiments. plot A1 and A2 show two possible interpretations of sample RT11_29.45_TH. The arai plots show the decrease of NRM and gaining of TRM at each temperature step during the experiment. The arai plots show how successful the experiment has been and through calculations obtains the intensity values.

Sample RT19_48.80_TH RT33_81.0_TH, and RT37_91.0_TH (Fig 8 C, F, G) shows tendencies of a zigzag pattern starting at 400 °C which indicates presence of MD grains and are therefore rejected. Other samples (Fig 8 B, E, H) fails the pTRM checks which imply alteration due to overlapping unblocking temperatures and are hence rejected. Alterations occur at 450-500 °C as can be seen in figure 7 B, E, H. The TRMs then proceed back to “normal” values, but this is for the new altered mineral, and does not show the NRM and TRM of the original mineral, and can thereby not be accepted.

Sample RT24_58.96_TH (Fig 8D) shows an exemplary result of the paleointensity experiment, where all pTRM checks are successful. This yielded an intensity value (F) of 60.9 µT.

Sample RT11_29.45_TH have two possible acceptable results, one high temperature component and one low temperature component (Fig 8), which gave an intensity value spectra of 35.5 µT up to 71 µT. Archaeomagnetic experiments also gave two possible results with this sample, one low temperature component which gave an intensity value of around 70 µT and a high temperature component which gave a value of around 30 µT. The high temperature component results were

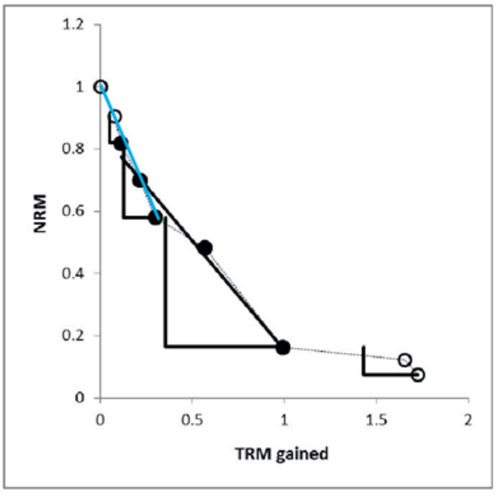

Fig. 9. Combined results of the arai plot of sample rt11_29.45_th. The blue line represent the rejected low temperature component whereas the black line represent the accepted high temperature component.

the one which fit the best according to the SELCRIT2 (MOD) criterions, and in a previously made intensity data plot of regional data (Pers. comm. Andreas Nilsson).

Criterions of SELCRIT2(MOD) are as follows (Paterson et al. 2014b):

$f = \geq 0.35$

$\beta = \leq 0.1$

$q = \geq 1$

$DRAT = \leq 10$ %

$MAD = \leq 15$

Where the successful samples gave following values:

RT11_29.45_TH (A1)	RT11_29.45_TH (A2)	RT24_58.96_TH
F = 35.5 µT	F = 71.0 µT	F = 60.9 µT
f = 0.744	f = 0.452	f = 0.930
β = 0.097	β = 0.041	β = 0.021
q = 5.1	q = 7.5	q = 36.5
DRAT = 5.2%	DRAT = 4.7%	DRAT = 1.7%
MAD = 3.3	MAD = 3.2	MAD = 1.1

6 Discussion

The most successful sample also shows the highest susceptibility values, possible related to higher SD components.

In order to test the initial hypotheses, the results from this study are combined with previous unpublished measurements from the same Rangitoto drill core (pers. comm. Andreas Nilsson 2017). The Rangitoto paleomagnetic data are compared to regional archeomagnetic and sedimentary paleomagnetic data available from the online GEOMAGIA.v50 database (Brown et al. 2015a; Brown et al. 2015b) as well as unpublished paleointensity data from ceramic fragments from the SW Pacific (pers. comm. Mimi Hill, 2017). The results achieved from this experiment, merged with the other available data are shown in fig. 10 and fig. 11 and are compared to the regional dataset. The inclination data shows no apparent structure and is therefore easier to fit into a small time slot in the reference dataset. It fits well together with the regional inclination data over a time span of ~150 years. The intensity data shows a stronger structure and has therefore a narrower time span where it can fit in the reference dataset. After comparisons between the two datasets it is fitted into a time span of ~1000 years. Based on these comparisons I conclude that neither hypothesis can be rejected.

A possible explanation for the two components in paleointensity experiment RT11_29.45 could potentially be the phenomena of an overprint, where a secondary component of NRM acquired after the formation of the rock overprint the primary components (Gupta 2011). The sample was positioned at the end of a lava flow, which means it could have been reheated by the next erupted lava flow, or it could have been oxidized by being in contact with air for too long, and consequently acquired a secondary chemical remanence. However, in general, there is no correlation between the successful samples and the geochemical zones in the core log, or the position of the samples.

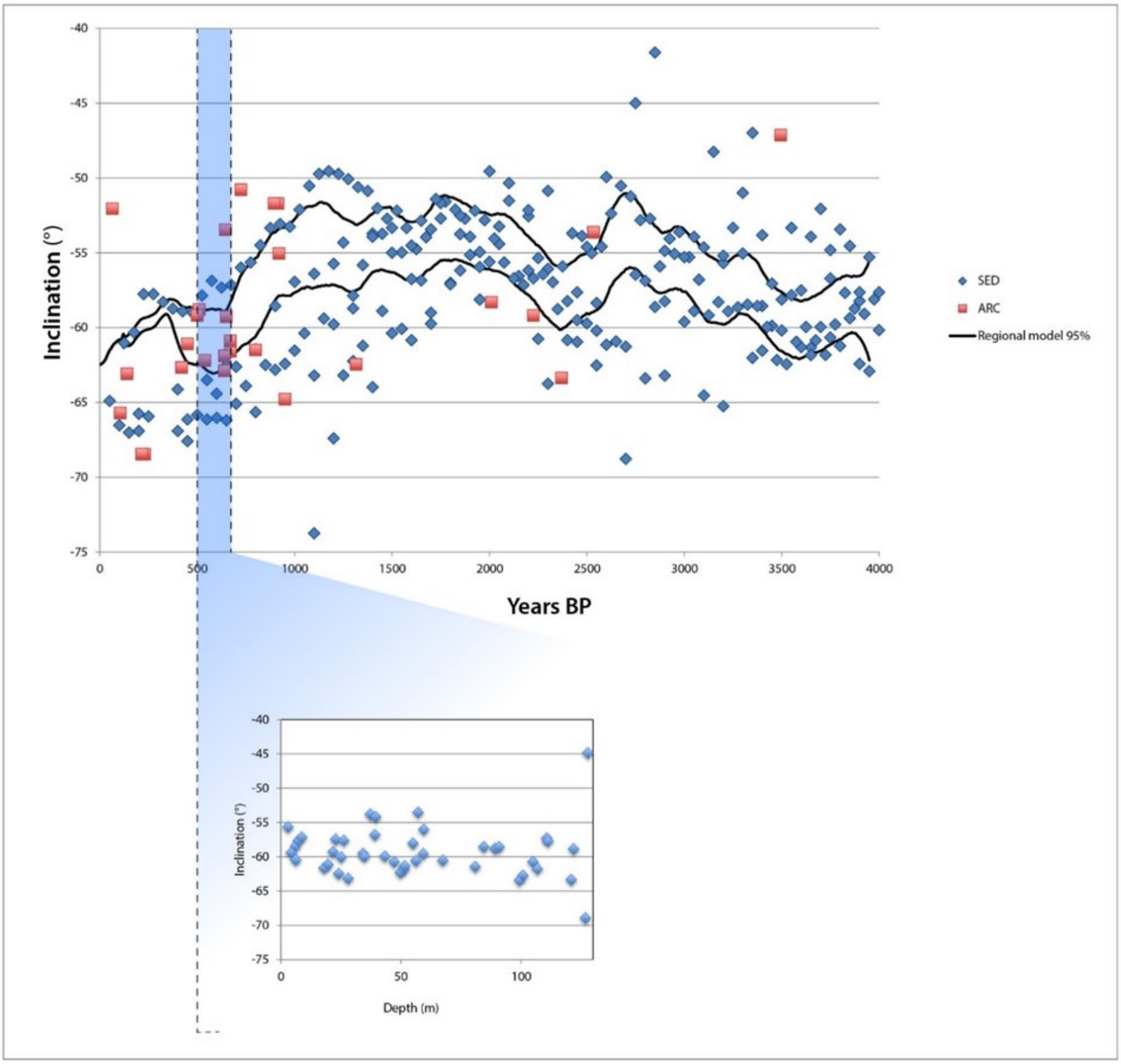

Fig. 10. inclination data from unpublished paleointensity data of ceramic fragments from the SW Pacific are correlated to a regional inclination data set from the online database GEOMAGIA.v50. The inclination data shows no apparent structure and can fit into a timespan of 150 years.

The inclination/years dataset from archaeomagnetic and sedimentary data is received from the online database GEOMAGIA.v50 and correlated with unpublished inclination/depth paleointensity data of ceramic fragments from the SW Pacific The additional data can be extrapolated to fit within a time span of 150 years in the model of the intensity/years data (fig 10), and thus support the first hypothesis that the majority of the eruptions occurred during a shorter period of time (~500-650 BP). This hypothesis is also supported by ^{14}C dating, but due to mixing of the sediments during the subaqueous phase of the volcano, the ages has been reversed and therefor the age determinations are unreliable.

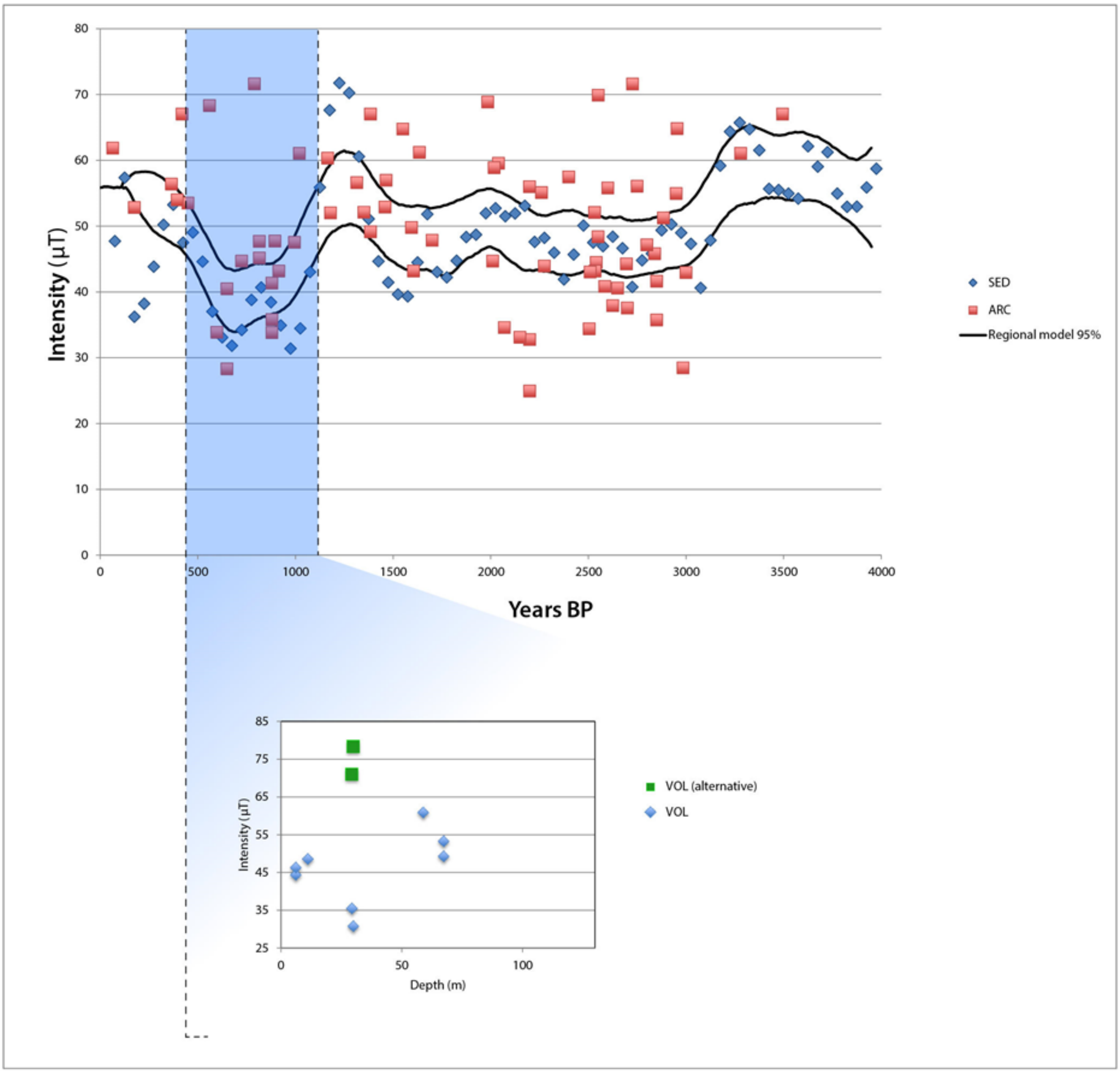

Fig. 11. The intensity data from this study and results from microwave experiments from the same drill core are plotted to fit the dataset of regional intensity measurements of archaeological and sedimentary samples. The alternative values show the nonreliable data from this study and the microwave experiment.

The intensity dataset is received from microwave experiments on samples from all identified lava flows from Rangitoto. Microwave experiments are another method of obtaining paleomagnetic data, which does not show the same high failure rate due to alteration as to thermal demagnetization.

The intensity data implies that the eruptions would have happened during a longer period of time, and rather demonstrate a timespan of about 1000 years (see fig 11), and thus support the second hypothesis. The green squares in the intensity diagram represent the low temperature component value of RT11_29.45_TH and the alternative value from the microwave experiment. The reason for the two different intensity values could, as previously mentioned, be the result of an overprint.

8 Conclusions

(I) The paleointensity experiments had a success rate of 25% and presented a paleointensity sequence of 30.5-71 μT.

(II) The inclination data supports a history of a great eruptive phase in the latest active stage of the volcano of about 150 years. This would represent the main shield building period that stood for the majority of the erupted material.
The intensity data tells a different story however, and supports the second hypothesis that the eruptions occurred sporadically over a longer period of time, namely ~1000 years.

(III) Since the inclination data does not show an apparent structure it is easier to fit it into the regional model. The intensity data on the other hand shows a strong structure which makes it harder and more specific to fit into the model.

(IV) The results of this experiment have not given any clear answers to the questions as to Rangitotos eruption history, and are not sufficient to reject either hypothesis. There are not enough data to make any reliable conclusions, and in order to establish a more consistent age model of the lava flows it is recommended to do more measurements of the paleointensity, to be able to do more correlations. This should be done using alternating microwave experiments as to the lower rate of alteration of the samples. To determine the history of the volcano and its conduit system it is also recommended to do geophysical surveying in order to get more information about the underlying geology and the crustal structure of the volcano.

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