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Magma plumbing beneath Anak Krakatau volcano, Indonesia: evidence for multiple magma storage regions

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Abstract

Understanding magma plumbing is essential for predicting the behaviour of explosive volcanoes. We investigate magma plumbing at the highly active Anak Krakatau volcano (Indonesia), situated on the rim of the 1883 Krakatau caldera by employing a suite of thermobarometric models. These include, clinopyroxene-melt thermobarometry, plagioclase-melt thermobarometry, clinopyroxene composition barometry and olivine-melt thermometry. Petrological studies have previously identified shallow magma storage in the region of 2–8 km beneath Krakatau, while existing seismic evidence points towards mid- to deep-crustal storage zone(s), at 9 and 22 km respectively. Our results show that clinopyroxene in Anak Krakatau lavas crystallized at a depth of 7–12 km, while plagioclase record both shallow crustal (3–7 km) and sub-Moho (23–28 km) levels of crystallisation. These magma storage regions coincide with well constrained major lithological boundaries in the crust, implying that magma ascent and storage at Anak Krakatau is strongly controlled by crustal properties. A tandem seismic tomography survey independently identified a separate upper crustal (< 7 km) and a lower to mid-crustal magma storage region (> 7 km). Both petrological and seismic methods are sensitive in detecting magma bodies in the crust, but suffer from various limitations. Combined geophysical and petrological surveys, in turn, offer increased potential for a comprehensive characterization of magma plumbing at active volcanic complexes.

Keywords: *Anak Krakatau; thermobarometry; clinopyroxene; plagioclase; magma plumbing; seismic tomography.*

1. Introduction

The Krakatau volcanic complex (Fig. 1), western Java (Indonesia), is one of the most infamous volcanoes worldwide due to the caldera forming eruption of 1883 during which Krakatau island collapsed into the sea. Inside the remnants of the old structure, a new volcanic cone breached the ocean surface in 1927, earning the name Anak Krakatau, “child of Krakatau” (Stehn 1929).

The bimodal nature of the Krakatau complex, i.e. extended periods of basalt and/or basaltic-andesite eruptions developing towards, and culminating in, colossal high-silica caldera-forming ignimbrite eruptions, was originally noted by van Bemmelen (1949). The topic has been further discussed by other authors (e.g. Camus et al. 1987; Mandeville et al. 1996a), but still lacks a comprehensive treatment. Improved knowledge of the magma plumbing system beneath Anak Krakatau will improve our understanding and better allow for prediction of future activity at this highly dynamic volcanic complex.

The aim of this investigation is therefore to constrain the current depth of magma storage region(s) beneath Anak Krakatau. We approach this problem by employing pressure and temperature modelling calculations that use measured mineral and rock composition data and calibrated thermodynamic formulations. The focus will be on clinopyroxene-melt and plagioclase-melt thermobarometry (Putirka et al. 2003; Putirka 2005; 2008), but data derived from the application of these thermobarometers will be compared with those obtained from clinopyroxene composition barometry and olivine-melt thermometry (Nimis 1999; Putirka et al. 2007; Putirka 2008). The mineral dataset consists of electron microprobe (EMP) and X-ray fluorescence (XRF) analyses of minerals and rocks erupted between 1883 and 2002. The results will serve as an independent test of previous estimates and constraints on magma storage derived by geophysical means, plagioclase-melt geobarometry

1.

Figure 1 Simplified geological and structural map of the Sunda Straits, based on Mandeville et al. (1996a), Schlüter et al. (2002), Lunt et al. (2009) and Susilohadi et al. (2009). The locations of volcanoes that together with Anak Krakatau define a north-south trending volcanic lineament are marked with triangles. The hatched area in the centre represents the volcanic lineament of quaternary volcanoes discussed in Nishimura and Harjono (1992). Inset (a) shows location relative to regional geography and the boundary between the subducting Indo-Australian plate and the Eurasian Plate. Inset (b) shows details of Krakatau volcano.

and in-situ isotope stratigraphy (Camus et al. 1987; Harjono et al. 1989; Mandeville et al. 1996a; Gardner et al. 2006; in press). Our thermobarometric approach will be integrated with the results of a geophysical survey, using local earthquake tomographic inversion, aimed at investigating the seismic structure beneath the Krakatau volcanic complex and surrounding areas (Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011).

Our study applies, for the first time, barometry based on clinopyroxene, to provide an independent and complementary test to isotopic, geophysical and geobarometric depth constraints for Anak Krakatau.

2. Geotectonic setting

The Krakatau archipelago lies in the Sunda arc where the Indo-Australian plate is subducted beneath the Eurasian one (Fig. 1, inset a). In west Java, subduction is occurring at a rate of 67 ± 7 mm per year (Tregoning et al. 1994). The Sunda arc is an active volcanic region, with the Krakatau volcanic complex being one of the most active parts. Since 1927, when it emerged above sea level, Anak Krakatau has had numerous eruptions, and has grown to a height of ~ 315 m (Hoffmann-Rothe et al. 2006), which translates to an average growth rate of ~ 7.5 cm per week. The contemporary Krakatau archipelago and volcanic complex consists of four islands; Rakata, Sertung, Panjang and Anak Krakatau. The only exposed remnant of the pre-1883 edifice of Krakatau is the dissected cone of Rakata, previously the southernmost peak of the pre-1883 volcanic complex. The submarine caldera formed in the 1883 eruption is visible as a ~ 100 m deep depression in the sea floor (Deplus et al. 1995). Although presently largely covered with 1883 eruptive products, the islands of

Sertung and Panjang are remnants of an earlier “proto-Krakatau” eruption. The records in local Javanese folk stories, e.g. the Book of Kings, or ‘Pararaton’ (Judd 1889; Nishimura et al. 1986; Camus et al. 1987), describe heavy rains of stone in the year 338 Saka (416 AD). While there is no direct evidence for an eruption of this size at that time, it may be a mistaken date for another eruption in 535 AD (Wohletz 2000). In addition, there is evidence for a yet older large ignimbrite eruption at $\sim 60,000$ BC (Ninkovich 1979), indicating a cyclic nature to behaviour of the Krakatau complex.

The Krakatau volcanic complex as a whole is part of a NNE-SSW trending lineament of quaternary volcanic edifices which lies approximately perpendicular to the Java trench (Nishimura and Harjono, 1992) (Fig. 1). This lineament is related to a north-south trending fracture zone that is manifested in a shallow seismic belt with foci depths predominantly in the range of 0–20 km (Harjono et al. 1989; Nishimura and Harjono 1992; Špičák et al. 2002). Furthermore, the Krakatau volcanic complex is located at the intersection of this volcanic lineament and a fault connecting Krakatau to the Sunda Strait graben (Harjono et al. 1989; Deplus et al. 1995). In fact, the whole of the Sunda Strait is subject to extensive faulting and extensional rifting, attributed to the clockwise rotation of Sumatra relative to Java by $\sim 20^\circ$ since the late Cenozoic (Ninkovich 1976; Nishimura et al. 1986; Harjono et al. 1991; Schlüter et al. 2002). The angle of plate convergence changes from near perpendicular (13°) in front of Java to oblique (55°) in front of Sumatra (Jarrard 1986). The Sumatran rotation has resulted in extensional thinning of the crust to ~ 20 km in the Sunda Strait, as compared to 25–30 km in Sumatra and west Java (Harjono et al. 1991; Nishimura and Harjono, 1992). The micro-seismic study by Harjono et al. (1989)

Figure 2 (a) Inferred stratigraphy of the bedrock below Anak Krakatau. The lithologies are constrained by data obtained from xenoliths (Oba et al. 1983; Camus et al. 1987; Mandeville et al. 1996b; Gardner et al. in press), drill holes, and seismic studies (Harjono et al. 1989; Kopp et al. 2001). (b) Cross section through the crust showing seismic attenuation zones detected below Anak Krakatau (redrawn from Harjono et al. 1989). Circles represent earthquake foci with (open circles) and without (filled circles) S-wave attenuation. The inferred magma storage regions are represented by the grey shapes. Note that a lack of data from the lower crust in this study meant that it was not possible to decide whether or not these two attenuation zones are connected.

estimated the crustal thickness directly below Anak Krakatau to be ~22 km.

The influence of the rifting is manifested in Sukadana (Fig. 1), where 0.8–1.2 Ma MORB-type basalts are found (Nishimura and Harjono 1992). The magmatism in the Sunda Strait is thus not exclusively subduction zone related, but can also be a function of extensional tectonics.

Drill cores obtained during hydrocarbon exploration by Pertamina/Aminoil provide information on the bedrock at depth in the Sunda Strait (Noujiam 1976; Mandeville et al. 1996a). The closest of these drilled wells (C-1SX) is located ~30 km southeast of Anak Krakatau (Fig. 1). The C-1SX well penetrated a sequence of sediments and sedimentary rocks of Quaternary to upper Pliocene age. The upper part (0–600 m) consists of unconsolidated marine clays followed by clay-dominated siliciclastic rocks interbedded with volcanoclastic material to a depth of at least 3000 m (Nishimura and Harjono 1992; Mandeville et al. 1996a). Findings by Lelgemann et al. (2000) suggest that the extension and rapid subsidence of the Sunda Strait have created space for up to 6 km of graben fill. Thus, the total depth of the sediments and sedimentary rocks in this part of the Sunda Straits can be constrained to between 3 and 6 km. The Pertamina-Aminoil wells all failed to reach the basement below the sedimentary sequence, but other wells to the southeast of Sumatra and northwest of Java have drilled Cretaceous granites and quartz-monzonites (Hamilton 1979). The assumption of a sedimentary sequence underlain by a plutonic basement below Krakatau (Harjono et al. 1991) is supported by the presence of sedimentary (Mandeville et al. 1996b; Gardner et al. in press), granitic (Oba et

al. 1983) as well as dioritic, gabbroic and meta-basic xenoliths (Oba et al. 1983; Gardner et al. in press) in Krakatau lavas and pyroclastic flows. The micro-seismic study by Harjono et al. (1989) identified three boundaries in the crust below the Krakatau complex, with unique crustal velocities, which should correspond with lithological boundaries (Fig. 2). The upper boundary (4 km) very likely represents the sedimentary-plutonic transition. The boundary at 9 km represents a density contrast that implies a change in lithology from low density (e.g. granite) to a higher density plutonic rock (e.g. diorite or gabbro) and the lowermost boundary at 22 km represents the Moho.

2.1 Whole rock Geochemistry

Fig. 3 shows whole rock geochemistry of lava flows, bombs and ash erupted from Krakatau and Anak Krakatau in terms of total alkalis versus silica (TAS). Data is taken from Zen and Hadikusumo (1964), Self (1982), Camus et al. (1987), Mandeville et al. (1996a) and Gardner et al. (2006; in press). All oxide values have been normalized to 100%, (volatile free), and iron content is FeO_t . Note that the lava flows and bombs of the 1990 to 2002 described in Gardner et al. (2006; in press), are the same samples that are used for the petrographic and microprobe analyses in this study. Analyses of rocks erupted during the 1883 rhyodacitic eruption and between 1960–1981 at Anak Krakatau are also plotted in Fig. 3.

The rocks plot in two main groups in the TAS diagram. The pumices and obsidians of the 1883 eruption plot in the dacite-rhyolite field, while the lava flows and bombs from Anak Krakatau belong to a rather homogenous suite of basaltic-andesites. The exceptions to this would be the 1960–1963 and 1981 eruptions (basalts and acidic-andesites, respectively), but these seem to represent isolated events with anomalous chemistry compared to the majority of eruptions. Note also that several basaltic dyke rocks from the island of Rakata with a similar composition to the 1963 basaltic flows have been reported (Camus et al. 1987). The early history of Anak Krakatau is, unfortunately, not well documented with few analyses performed on rocks erupted prior to 1960. However, there are indications that the early Anak Krakatau lavas did not differ significantly in composition from more recent eruptives, as silica content in ashes and bombs erupted in the period 1928–1935 have been reported to be in the range of 51.81 to 54.76 wt. % (van Bemmelen 1949), overlapping with the SiO_2 content of the recent basaltic-andesites (Fig. 3). This corresponds well with the observation of Camus et al. (1987) that the composition of the tuff ring and lava flows on Anak Krakatau appeared to belong invariably to the basaltic-andesite suite. Field observations by the two lead authors in 2008 also supports this, as the lava bombs of the 2007–2008 eruptions appear to be virtually identical petrographically to the 2002 bombs described by Gardner et al. (2006; in press). Thus, all whole rock analyses and field observations are consistent with the conclusion that the bulk of Anak

Krakatau is made up of basaltic-andesite, with minor components of basalt plus sparse acidic-andesite.

3. Previous estimates of magma storage depth

Estimates and constraints of magma storage beneath the Krakatau complex have been previously derived using a.) plagioclase-melt thermobarometry, b.) chlorine content in melt inclusions, c.) loci of seismic attenuation zones, and d.) in-situ crystal isotope stratigraphy.

Mandeville et al. (1996a) employed plagioclase-melt thermobarometry (after Housh and Luhr 1991) and chlorine content in melt inclusions (after Metrich and Rutherford, 1992) to estimate the depth of the 1883 magma chamber. Their results indicate shallow depths of magma storage, with pressure estimates in the range of 100 to 200 MPa (~4 to 8 km) for plagioclase crystallisation, and ca. 100 MPa (~4 km) derived from chlorine inclusions. Camus et al. (1987) used plagioclase-melt thermobarometry (after Kudo and Weill, 1970) to estimate pressures of crystallization for plagioclase in rocks erupted in 1883 (Krakatau) and between 1973 and 1981 (Anak Krakatau), arriving at an estimate of 50 to 200 MPa (~2 to 8 km) for both Krakatau and Anak Krakatau lavas. Gardner et al. (2006; in press) carried out in situ $^{87}\text{Sr}/^{86}\text{Sr}$ analyses employing LA-ICPMS on plagioclase from the 2002 eruptions of Anak Krakatau. This work illustrates that late crystallization of many plagioclase grains must have taken place during assimilation of sedimentary country-rock. This observation constrains the depth of late stage plagioclase crystallization, on simple stratigraphic grounds (Fig. 2a), to within the upper 3-6 kilometres (Nishimura and Harjono 1992; Lelgemann et al. 2000), which is in good agreement with the plagioclase-melt thermobarometry results discussed above.

Harjono et al. (1989) analysed the seismic signature from 14 earthquakes near Anak Krakatau in 1984, using data from analogue seismograms. Two seismic attenuation zones beneath the volcanic edifice were

identified in that study, one a small and irregular zone at a depth of approximately 9 km, and another much larger area at 22 km (Fig. 2b). These were thought to represent the granite – diorite/gabbro transition and the Moho, respectively (Fig. 2a). However, a lack of data from the lower crust meant that it was not possible to decide whether or not these two attenuation zones are connected. It also remains unresolved if these two zones represent large-volume chambers or a concentration of smaller melt pockets, which is relevant for estimating long term magma storage volumes and hence the availability of magma for eruption. Harjono et al. (1989) did, however, speculate that the mid-crustal storage region (9 km) is made up of separate magma pockets with a combined volume large enough to differentiate an eruptive volume in the order of 10-15 km³, i.e. comparable to the volume erupted in the caldera forming eruption of 1883. The deep storage region (22 km), on the other hand, was thought to be considerably larger, both in terms of volume and extension (Fig. 2b, Harjono et al. 1989).

In summary, petrological studies have previously identified shallow magma storage in the region of 2-8 km beneath Krakatau (Camus et al. 1987; Mandeville et al. 1996a; Gardner et al. 2006; in press), whereas seismic evidence points towards mid- to deep-crustal storage zone(s), at 9 and 22 km respectively (Harjono et al. 1989). This discrepancy between the results is likely due to inherent limitations in both methods. Seismic studies cannot resolve small volume magma bodies and petrological methods are dependent on mineral phases and melt inclusions retaining a complete record of pressure, temperature and compositional parameters that have changed during the various stages of magma ascent and evolution.

4. Methods

4.1 EMP analytical procedure

Mineral chemistry as well as glass and groundmass

Figure 3 Total alkalis versus silica (TAS) diagram (Le Bas et al. 1986) plotting bulk composition of rocks from Anak Krakatau (circles) and Krakatau (triangles) (data from Zen & Hadikusumo, 1964; Self, 1982; Camus et al. 1987; Mandeville et al. 1996a; Gardner et al. in press). The Krakatau rocks include products from the 1883 rhyodacite eruption (open triangles) as well as older basaltic dyke rocks (filled triangles). The vast majority of the rocks erupted from Anak Krakatau plot in a very narrow range in the basaltic-andesite field, with single eruptive events only in the basalt and andesite fields

compositions from basaltic andesites erupted in the period 1990-2002 were analysed at two different Electron Microprobe (EMP) facilities: a Cameca SX 50 at Uppsala University (Sweden), and a JEOL JXA-8200 Superprobe at Copenhagen University (Denmark). For the Cameca EMP, an accelerating voltage of 20 kV and a current of 15 nA was used, while the JEOL EMP was calibrated using an accelerating voltage of 15 kV and a current of 15 nA. The beam diameter was generally 1-2 μm , though a defocused spot was used for the analysis of groundmass and glass composition. These 'defocused spot' analyses of groundmass included glass and microcrysts, but avoided phenocrysts. Glass compositions were not only analysed from the groundmass, but also in melt inclusions. International reference materials were used for calibration and standardisation (e.g. Andersson 1997). For elements with abundances greater than 1 wt%, the reproducibility is $< \pm 10\%$, while for abundances less than 1 wt% the reproducibility is $> \pm 10\%$.

The analysed dataset consists of 297 clinopyroxene, 375 plagioclase, 39 olivine and 92 orthopyroxene spot analyses, collected from 63 clinopyroxene, 71 plagioclase, 11 olivine and 36 orthopyroxene distinct crystals. Also, 14 analyses of glass in groundmass and melt inclusions were collected as well as another 14 defocused spot ($\sim 10 \times 10 \mu\text{m}$) analyses of groundmass, in order to calculate an average groundmass composition that includes both glass and microcrystalline phases. All analytical data are reported in the electronic appendix.

4.2. Clinopyroxene-melt thermobarometers

Two models based on clinopyroxene-melt equilibria have been applied. The first thermobarometer is based on the jadeite-diopside/hedenbergite exchange equilibria between clinopyroxene and co-existing melt (Putirka et al. 2003). This model has proved to be able to recreate P-T conditions for a wide range of magma compositions, within a reasonable margin of error, and has been frequently used in the last decade (e.g. Shaw and Klügel 2002; Putirka and Condit 2003; Schwarz et al. 2004; Caprarelli and Riedel 2005; Klügel et al. 2005; Galipp et al. 2006; Mordick and Glazner 2006; Longpré et al. 2008; Barker et al. 2009). Crucially, it has proved to be applicable to analogue (high P-T) experiments (Putirka 2008) while also correlating well with independent methods for natural samples (Klügel and Klein 2005). The Putirka et al. (2003) thermobarometer will henceforth be termed PTB03. The standard errors of estimate (SEE) for PTB03 are $\pm 33^\circ\text{C}$ and $\pm 170\text{ MPa}$ (Putirka et al. 2003).

The second clinopyroxene-melt model to be used for comparison is a barometer based on the Al partitioning between melt and clinopyroxene, calibrated especially for hydrous systems, requiring the input of a specific H_2O concentration (Putirka 2008, eqn. 32c). This model will be referred to as PB08. The PB08 formulation also requires the input of a temperature value, which can be provided by the PTB03 model. Note that PB08 is not as firmly tested

as PTB03, but the SEE of $\pm 150\text{ MPa}$ is thought to be superior to that of PTB03 (Putirka 2008).

As PTB03 and PB08 are based on independent clinopyroxene-melt exchange equilibria (Na and Al, respectively), these models are used in conjunction to provide an internal validation of results.

4.3 Clinopyroxene composition barometer

To test the results of the Putirka clinopyroxene-melt thermobarometry (PTB03 and PB08), a clinopyroxene barometer not requiring the input of a coexisting melt can be used. The clinopyroxene composition barometer developed by Nimis (1995; 1999) and Nimis and Ulmer (1998) is widely used, despite a reported tendency of systematically underestimating pressures when applied to hydrous systems (Putirka 2008). To eliminate this systematic error, this barometer was re-calibrated for hydrous systems by Putirka (2008, eqn. 32b), with the added requirement of an H_2O estimate in addition to the temperature estimate already needed. This barometer will be referred to as NimCal08. The SEE for NimCal08 is estimated to $\pm 260\text{ MPa}$ (Putirka 2008).

4.4 Plagioclase-melt thermobarometer

Since the first plagioclase thermometer was formulated (Kudo and Weill 1970), the approach has been developed further by various workers including geobarometers (Housh and Luhr 1991; Sugawara 2001; Ghiorso et al. 2002; Putirka 2005; 2008). Putirka (2005; 2008) calibrated the plagioclase-melt thermobarometer for hydrous systems, requiring the input of a H_2O estimate in the modelling calculations. However, the uncertainty in plagioclase-melt geobarometry remains significant and, occasionally and apparently randomly, produces very poor results from some data sets (Putirka 2008). The SEE for the plagioclase-melt thermobarometer is $\pm 36^\circ\text{C}$ and $\pm 247\text{ MPa}$ (Putirka 2008).

4.5 Olivine-melt thermometer

The Putirka et al. (2007) olivine-melt thermometer, specifically calibrated for hydrous systems, will be used in this study. The SEE of this olivine-melt thermometer is $\pm 29^\circ\text{C}$ (Putirka 2008).

We have employed this thermometer to provide an independent test for the reliability of our clinopyroxene-melt thermobarometry. As olivine is assumed to have formed prior to and/or together with clinopyroxene, calculated olivine crystallisation temperatures are expected to overlap or be slightly higher than calculated clinopyroxene temperatures. Such a match would indicate reliable results (e.g. Longpré et al. 2008).

4.6 Seismic tomography

The seismic structure beneath the Krakatau complex and surrounding areas has been studied using local earthquake tomographic inversion (Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011). The data for this

study were collected within the seismological part of the KRAKMON Project (Hoffmann-Rothe et al. 2006; Ibs-von Seht 2008). Fourteen temporary seismic stations were installed on the islands of the Krakatau archipelago and in surrounding areas of the Sunda Strait, Sumatra and Java. The survey was conducted between June 2005 and January 2006. The dataset used in this study includes more than 700 local events with arrival times of 3128 P- and 2050 S-phases. The data processing is based on the LOTOS code (Koulakov 2009) which is used for the non-linear tomographic inversion of P and S arrival times to the distributions of P and S velocities, as well as Vp/Vs ratio and source locations. The resolution limitations due to unfavourable configuration of the network were thoroughly investigated in Jaxybulatov et al. (2011) using a series of different synthetic tests.

5. Input parameters

5.1 Mineral-melt equilibrium tests

Finding the appropriate equilibrium melt compositions for mineral-melt thermobarometric models is of key importance for accurate results. This is the single largest source of error in mineral-melt equilibria models, which means that rigorous evaluation of any nominal melt of choice for equilibrium is critically important. Equilibrium tests are commonly performed using exchange coefficients of elements between mineral and melt (e.g. Klügel and Klein 2005; Longpré et al. 2008; Barker et al. 2009). For clinopyroxene and olivine, the equilibrium exchange coefficients for iron and magnesium ($Kd_{[FeMg]}$) are expected to be 0.28 ± 0.08 and 0.30 ± 0.03 respectively (Roeder and Emslie 1970; Duke 1976; Putirka 2008). Mineral-melt pairs with $Kd_{[FeMg]}$ values falling outside these boundaries should be discarded. For clinopyroxene, it is also useful to consider the equilibrium partitioning of other components such as Na-Ca-Al (Rhodes et al 1979; Putirka 1999; 2008). The equilibrium test suggested by Putirka (1999) predicts the relative amounts of different clinopyroxene mineral components that would crystallize from a given equilibrium melt at the estimated P-T conditions. The “predicted mineral components” (PMC) can then be compared to the “observed mineral components” (OMC). A close match would support equilibrium conditions. The recommended equilibrium test for plagioclase-melt thermobarometry uses the ratio of the partitioning coefficients of the anorthite and albite components, $Kd_{[An-Ab]}$ (Putirka 2005; 2008). This is expected to be 0.10 ± 0.05 at $T < 1050$ °C and 0.27 ± 0.11 at $T > 1050$ °C (Putirka 2008). Only the plagioclase datapoints closest to equilibrium with the selected nominal melt should be considered reliable. As a further test for equilibrium, the temperature estimate will be compared to a plagioclase saturation surface temperature calculated for the nominal melt, with a SEE of ± 37 °C (Putirka 2008, eqn. 26). The

plagioclase saturation surface temperature is the lowest possible temperature for the nominal melt, at a given pressure, before plagioclase starts crystallising. A close match between the temperature estimates of the plagioclase-melt and plagioclase saturation thermometers is expected for close to perfect equilibrium conditions (Putirka 2008).

5.2 Pre-eruptive H₂O content

The H₂O content is a very influential parameter in a number of the thermobarometers that are used in this study. The pre-eruptive volatile content of a rock can be approximated from the mass deficiency in EMP analyses of groundmass glass and melt inclusions (‘the difference method’), as described in Devine et al. (1995), provided that the volatiles make up $>1\%$. Anak Krakatau melt inclusions represent the melt composition of the magma during different stages of its evolution prior to degassing and eruption. The mass deficiency in the glass inclusions measured here ranges from 1.2 to 4.9 wt% (average = 2.4 wt%; $n = 9$). However, the precision of the ‘difference method’ is not very high (generally around $\pm 0.5\%$; Devine et al. 1995). Mandeville et al. (1996a), also using the ‘difference method’, estimated the pre-eruptive volatile content in 1883 Krakatau rhyolites and dacites to be 4 ± 0.5 wt. %. Basalts and basaltic andesites should be H₂O saturated at a lower H₂O content than is the case for more evolved compositions, leading to the assumption that melt inclusions with low volatile content likely represents melt captured at an earlier stage of magma evolution. Therefore we will use a range of values from 2 to 4 wt% H₂O for the thermobarometric calculations to follow. The higher end of that range (3–4 wt. %) will be considered for basaltic-andesite whole rock compositions, while the lower end (2 to 3 wt. %) will be considered for basaltic compositions.

5.3 Bedrock density

For the conversion of pressure estimates (MPa) to depth (km), the approximate densities of the respective stratigraphic units below Krakatau have been estimated. In the study by Kopp et al. (2001), a seismic line over the Java trench was investigated, ending just ~ 10 km south of Krakatau. The stratigraphy proposed in that study will be used here as the basis for estimating the densities of bedrock units beneath Anak Krakatau (Table 1).

6. Results

6.1 Petrography and mineral chemistry

The basaltic andesite lavas examined are all highly porphyritic (30–35%), dark in appearance, and in part vesicular. Plutonic as well as sedimentary xenoliths occur. The homogeneity of the whole rock chemistry

Table 1. Estimated densities of crustal rocks below Anak Krakatau

Inferred rock types	Depth (km) (Harjono et al. 1989)	Density (g cm ⁻³) (Kopp et al. 2001)
Sedimentary succession	0-4	2.32 (2.23 - 2.4)
Granitoids	4-9	2.8
Diorite/gabbro	9-22	2.95
Mantle	>22	3.37

Table 2. Representative mineral compositions from Anak Krakatau eruptives

Mineral	plag	plag	plag	plag	plag	cpx	cpx	Ol
notes	core, An ₉₀	rim, An ₇₉	rim, An ₆₆	rim, An ₅₁	core, An ₄₅	Mg# = 0.61	Mg# = 0.63	core, Fo ₇₂
Analysis no:	kcs24plag44a	4_AKXCI 11 tr2 11	3_AK1 1 fsp p13	4_AKXCI 11 tr3 63	4_AKXCI 11 tr3 11	cpx 6	cpx 11	3_AKXIC 04 p35
SiO ₂	45.06	48.22	50.94	55.54	57.05	51.62	51.38	37.08
TiO ₂	0.027	0.00	0.057	0.04	0.01	0.69	0.64	0.02
Al ₂ O ₃	34.47	31.95	30.06	27.06	25.77	2.23	2.26	-
FeOt	0.527	0.70	0.74	0.54	0.48	9.77	9.17	25.23
MnO	0.024	-	-	-	-	0.35	0.33	0.42
MgO	0.087	-	-	-	-	15.34	15.63	36.73
CaO	18.16	16.13	13.90	10.85	9.73	19.80	19.86	0.21
Na ₂ O	1.046	2.28	3.84	5.49	6.31	0.20	0.19	-
K ₂ O	0.028	0.06	0.09	0.17	0.20	-	-	-
Cr ₂ O ₃	0.015	-	-	-	-	0.01	0.01	-
Total	99.44	99.34	99.63	99.69	99.55	100.0	99.47	99.69

is reflected in the petrographic features. All thin sections examined share the same characteristics, as well as a relatively homogenous mineral chemistry, especially with respect to clinopyroxene. In order of relative abundance the mineral phases in the rocks are, plagioclase > clinopyroxene > orthopyroxene > titanomagnetite > olivine. The modal composition is on average, 70% groundmass, 25% plagioclase, 4% clinopyroxene and less than 1% olivine phenocrysts or micro-phenocrysts (cf. Camus et al. 1987; Gardner et al. in press). These phenocryst phases will be described below. Selected microphotographs and electron backscatter images are displayed in Fig. 4. Representative compositions of plagioclase, clinopyroxene and olivine are reported in Table 2. All EMP analyses are available as an electronic appendix.

6.1.1 Plagioclase

Plagioclase phenocrysts are mostly subhedral, but a few are anhedral. Sieve-like textures are a common

feature with numerous melt inclusions present. The size of the plagioclase crystals is on the order of 0.5-2 mm. Many plagioclase crystals, when viewed under polarized light, appear to have experienced stages of growth, dissolution, and overgrowth. Some cores are rounded and appear partially resorbed. Such cores are often surrounded by sieve-like (or cellular) zones ending with a thin rim (e.g. Fig. 4a, c). Other crystals lack resorbed cores, but carry numerous melt inclusions (Fig. 4b). Normal, complex and reverse zoning have been documented. This is illustrated by the fact the lowest anorthite concentrations (An₄₅) as well as the highest (An₉₀) were found in plagioclase cores, though the majority of the low-An composition was found in the rims, and the highest An contents (~An₈₀₋₉₀) were only found in the cores. Similarly, the rims also vary widely in their compositional range, between An₇₉ to An₅₀. However, plagioclase compositions in the range of ~An₆₀₋₇₀ are typically found in sieve-like zones, rims and groundmass microcrysts, implying that this composition is closest

Figure 4 Representative microphotographs (crossed polars) and BSE images of (a) Subhedral plagioclase crystal with highly sieve-textured core and rim regions, reflecting a dynamic magma system. (b) Sieve-textured plagioclase crystal with numerous melt inclusions. Note also the very thin overgrowth rim. (c) Plagioclase crystal with resorbed core. The outer regions are sieve-textured with a thin overgrowth on the rims (cf. 4a, 4b). (d) A plagioclase crystal intergrown with several smaller clinopyroxene crystals. (e) Euhedral, twinned clinopyroxene crystal. In the bottom left corner, there is a small, partly resorbed olivine with very high interference colours. Note that the olivine has a discontinuous reaction rim with orthopyroxene and titanomagnetite, as can also be seen in Fig. 4h below. (f) Euhedral clinopyroxene crystal. Dark brown, vesicular groundmass consisting of acicular plagioclase, orthopyroxene, titanomagnetite and glass. (g) BSE image of intergrown plagioclase (bottom) and clinopyroxene (top). (h) BSE image of a partly resorbed olivine (center of image), covered by discontinuous overgrowths of orthopyroxene and titanomagnetite. The olivine crystal is surrounded by acicular plagioclase (dark grey laths), anhedral orthopyroxene (light grey), titanomagnetite (white crystals) and glass (irregular dark grey fields).

to equilibrium with the host rock. The composition of all plagioclase datapoints analysed for this study is illustrated in Fig. 5a-b.

6.1.2 Clinopyroxene

The clinopyroxene crystals are, with few exceptions, euhedral. Most grains have melt inclusions, though considerably fewer than in plagioclase. The clinopyroxene crystals are slightly smaller than plagioclase, in the region of 0.2-1.0 mm. Also, it is common to find plagioclase that has grown around clinopyroxene, implying that plagioclase growth occurred largely after clinopyroxene crystallization (Fig. 4d). The two mineral phases may thus record different stages of magma storage and crystallization. For reference, compositional data of clinopyroxenes from Mandeville et al. (1996a) and Camus et al.

(1987) have been included as a sub-set in the model calculations. The mineral chemistry of clinopyroxene is very homogenous with very little variation within individual grains. Therefore, in the case of clinopyroxene, the average composition of each individual grain was calculated and used in the thermobarometric calculations.

Following Morimoto et al. (1988), the compositions of pyroxene are plotted in Fig. 5c. All the analysed clinopyroxenes belong to the same compositional family. Core-rim variations are minor, with a tendency for normal zoning towards slightly more Fe-rich rims, but the opposite has also been observed in some grains. The recent (1990-2002) and old (1883-1981) clinopyroxenes are compositionally similar, despite occurring in basaltic andesite, andesite and even rhyodacite host rocks, and erupted over >100 years (i.e. 1883 and 2002).

6.1.3 Olivine

Olivine is a relatively rare phase in our analysed samples (<1 %) and has not been identified in all thin sections. The olivines present are small, with diameters in the order of 0.02 to 0.15 mm, and are best identified using electron backscatter images and EMP analyses. All olivines observed are anhedral (resorbed), frequently with rims covered by overgrowths of orthopyroxene and occasional titanomagnetite (Fig. 4h). There is a gradual normal zoning in all the olivine crystals investigated with generally more Fe-rich rims. Forsterite content ranges between Fo₆₃₋₈₀, with most olivine rims falling below Fo₇₀.

6.2 Pressures and temperatures from clinopyroxene-melt thermobarometry

The clinopyroxenes exhibit no clear petrographic indicators of being out of equilibrium with the host melt, as they are commonly euhedral and lack strong compositional zoning. However, euhedral habit and the lack of zoning does not by itself verify equilibrium between clinopyroxene and host melt, especially considering the sluggishness of clinopyroxene re-equilibration, which is on the order of months to years for a 5-10 μm rim (Cashman 1990). The rather short repose time at Anak Krakatau volcano, often with a dozen eruptive pulses during a single year (Siebert and

Simkin 2002), are unlikely to allow sufficient time for full re-equilibration during shallow residence of the magma.

The most commonly used nominal melts are whole rock, groundmass and glass compositions (e.g. Shaw and Klügel 2002; Putirka and Condit 2003; Putirka et al. 2003; Caprarello and Riedel 2005; Klügel et al. 2005; Longpré et al. 2008). In addition to these, a more primitive whole rock from an Anak Krakatau lava flow from the eruptive period 1960-1963 (henceforth labeled 1963 basalt) will be evaluated here (Zen and Hadikusumo 1964). In Fig. 6a, the $K_d_{[\text{FeMg}]}$ of the four nominal melt options are compared. Only the 2002 basaltic andesite and the 1963 basalt plot within the $K_d_{[\text{FeMg}]} = 0.28 \pm 0.08$ interval, and can thus be assumed to represent liquids in potential equilibrium with the analysed clinopyroxenes (cf. Putirka 2008).

In Fig. 6b, the observed and predicted Diopside-Hedenbergite (DiHd) components are plotted, using (a) the 1963 basalt and (b) the 2002 basaltic andesite composition. Both compositions result in virtually identical distributions, with 77% of the datapoints plotting within 5% of the theoretical equilibrium. It is not possible, using only these methods, to determine whether the 1963 basalt or 2002 basaltic andesite is more appropriate as the equilibrium melt. However, using the 2002 basaltic andesite, as much as 78% of the datapoints result in negative pressures (range = -1296 to 187 MPa, average = -203 MPa). Therefore, the 2002 basaltic andesite is a less viable nominal

Figure 5 (a) Composition of all analyzed plagioclase datapoints (n=375). Plagioclase composition varies between An₄₅ and An₉₀. (b) Histogram of all analysed plagioclase. (c) Composition of clinopyroxene (filled circles, n=312) and orthopyroxene (open circles, n=92) crystals. The clinopyroxene dataset includes compositions of minerals analyzed for this study (n=297) and data from the literature (n=15; Camus et al. 1987; Mandeville et al. 1996a). All clinopyroxene data plots in a narrow region in the augite field. The compositions of the orthopyroxene crystals are slightly more heterogeneous, spreading over the enstatite and pigeonite fields

melt, despite successfully passing the equilibrium tests. The 1963 basalt, producing more realistic pressures, is therefore taken as the most reliable nominal melt option to use in our calculations, i.e. the clinopyroxenes crystallised from a basaltic magma. All mineral-melt pairs that fall out of the equilibrium range in the DiHd test are ignored for the P-T calculations.

The three sets of calculated clinopyroxene crystallisation pressures using the 1963 basalt as nominal melt range from 59 to 377 MPa, with a high degree of overlap between the results from PTB03 and PB08 (Fig. 7). The PTB03 and PB08 models were also applied to published clinopyroxene compositions (Camus et al. 1987; Mandeville et al. 1996a), erupted between 1883 and 1981 (also using the 1963 basalt composition as nominal melt). These older clinopyroxenes were also subjected to the array of equilibrium tests described above (Fig. 6). The recorded P-T conditions of these older clinopyroxenes spread over a much larger interval of calculated pressures (116 to 759 MPa), broadly in agreement with the results obtained for clinopyroxenes analysed in this study. Generally, the older clinopyroxenes record slightly higher pressures and temperatures as compared to those obtained for lavas erupted between 1990 and 2002 (Fig. 7c). This potentially indicates that clinopyroxenes in lavas erupted prior to 1981 crystallized at a greater depth than the more recently erupted clinopyroxenes, pointing toward a shallowing of the magma system over the last 30 years. However, this remains somewhat speculative, and further testing is required to substantiate a long term pattern. All results of PTB03 and PB08 are reported in Table 3.

6.3 Pressure estimates from clinopyroxene barometry

The NimCal08 barometer (Putirka 2008, eqn. 32b), making use of temperature estimates calculated through the PTB03 model (see above), yields pressures slightly lower than the PTB03 and PB08 models, but with a significant overlap (Table 3, Fig. 7b,7d). Estimates of $2_{\text{H}_2\text{O}}$ and $3_{\text{H}_2\text{O}}$ result in average pressures of 103 MPa (-57 to 214) and 148 MPa (-11 to 259), respectively. As mentioned above, this model is not deemed particularly precise nor accurate, with a tendency for systematically underestimating crystallisation pressures (Putirka 2008). However, the overall overlap with model results from PTB03 and PB08 lends credence to the calculations. For a comparison of the results of PTB03, PB08 and NimCal08 models, see Fig. 7d.

6.4 Pressures and temperatures from plagioclase-melt thermobarometry

The $K_{\text{d}_{[\text{Ab-An}]}}$ equilibrium test (Fig. 8a) indicates that plagioclase in the compositional range of An_{56-76} (68% of analysed datapoints) are in equilibrium with the 2002 basaltic andesite, whereas the high anorthite plagioclase (An_{77-88} , 11% of analysed datapoints) are closer to equilibrium with the 1963 basalt. The low-

An compositional range (An_{45-56} , 21% of analysed datapoints), are out of equilibrium with all tested nominal melts, and these compositions are not considered for the model calculations. The medium anorthite compositions (An_{56-76}) that are in equilibrium with the whole rock are commonly found in sieve-like zones, rims and groundmass microcrysts, providing a textural indication that this compositional range should be closest to equilibrium with the host rock. Furthermore, using an input of >3.5 wt% H_2O , the temperature estimates calculated are invariably below 1050 °C, requiring a different set of $K_{\text{d}_{[\text{Ab-An}]}}$ equilibrium conditions (Putirka 2008). This means that the plagioclase is not in equilibrium with the whole rock at >3.5 wt% H_2O effectively constraining pre-eruptive H_2O content in the recent basaltic andesites to ≤ 3.5 wt% H_2O . A further indication of the suitability of the selected nominal melts for plagioclase-melt thermobarometry is the fact that the plagioclase saturation surface temperatures calculated are only 3-15 °C higher than the temperature estimated using the plagioclase-melt thermometer, which is well within the SEE for the plagioclase saturation model of ± 37 °C (Putirka 2008). There is a strong correlation between $K_{\text{d}_{[\text{Ab-An}]}}$ and the P-T estimates. The highest and lowest P-T estimates correlate with the $K_{\text{d}_{[\text{Ab-An}]}}$ farthest removed from the ideal value of 0.27. Therefore, the average P-T estimate for each data set $\pm 1\sigma$ will be considered, as this removes the less reliable results. Looking exclusively at these “best fit” plagioclases, two very tight ranges of P-T estimates are derived (Fig. 8b, Table 4). The plagioclase in equilibrium with the 2002 basaltic andesite (An_{56-76}), records pressures between 23 and 186 MPa. The high anorthite plagioclase (An_{77-88}) in equilibrium with the 1963 basalt, in contrast, record pressures between 645 to 807 MPa. In other words, these high anorthite plagioclase compositions formed in a deeper storage region, after which the magma was transported to shallower levels where crystallisation of more albitic plagioclase took place. Calculated plagioclase and clinopyroxene crystallisation depths are compared in Fig. 10 and Table 6.

6.5 Temperature estimates from olivine-melt thermometry

Due to the petrological evidence of resorption, olivine is not considered a stable phase in the basaltic andesite host rock. Thus, neither the 2002 basaltic andesite whole rock, nor its groundmass compositions can be considered feasible equilibrium melts for olivine-melt thermometry. However, the olivine present is likely to have crystallised in equilibrium with a more primitive melt, similar in composition to the 1963 basalt lava. The $K_{\text{d}_{[\text{FeMg}]}}$ values for olivine and the 1963 basalt whole rock composition (Zen and Hadikusumo 1964) were compared with the equilibrium $K_{\text{d}_{[\text{FeMg}]}}$ value (0.30 ± 0.03) of Roeder and Emslie (1970). The 1963 basalt fulfills the equilibrium requirements with olivine in the very narrow compositional range of

Fo₇₀₋₇₂, representing only 28% of all analysed olivine datapoints (Fig. 9a). Using the 1963 basalt and H₂O contents of 2 and 3 wt. %, the olivine melt thermometry results in a range of temperatures from 1124 to 1145 °C (see Table 5). The results, and a comparison with the calculated clinopyroxene and plagioclase crystallisation temperatures, are plotted in Fig. 9b. The temperatures estimated for olivine overlap with the temperatures estimated for clinopyroxene as well as for high anorthite plagioclase – all of which appear to be in equilibrium with the 1963 basalt. Meanwhile, the medium anorthite plagioclase (An₅₆₋₇₆) in equilibrium with the 2002 basaltic andesite form a distinct cluster of lower temperatures. Note also that the calculated olivine temperatures are marginally higher than for clinopyroxene and high anorthite plagioclase, which would fit well with the modelling of liquid line of descent from Putirka et al. (2007), suggesting that olivine would be the first phase to precipitate from this type of liquid. Crucially, the fact that the olivine-melt, clinopyroxene-melt and plagioclase-melt thermometers are consistent with each other provides a strong indication that the results are reliable (cf. Longpré et al. 2008).

6.6 Seismic tomography

In the tandem seismic tomography study (Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011) the structure of P and S velocities, as well as the distribution of Vp/Vs ratios have been obtained for the Krakatau area. The data distribution is not favourable for tomographic inversion, so thorough testing was performed to separate robust features from possible artefacts. In particular, it was shown that details of structures outside the Krakatau complex are strongly smeared and should be interpreted with prudence. At the same time, the checkerboard tests show that the shapes of the main patterns beneath the Krakatau complex, where most of the seismic events and stations are located, define real features. The amplitudes of P and S velocity anomalies were estimated using synthetic tests with realistic configurations of anomalies and noise. The general magnitudes of computed P and S anomalies beneath Krakatau are not exceptionally high, being approximately 8% and 13% respectively. Beneath the Krakatau complex down to a depth of 5 km, higher P-velocity anomalies anticorrelate with lower S-anomalies. Below 7 km depth, both P and S anomalies are negative, but the magnitude of S anomalies is much stronger. These velocity distributions result in extremely high Vp/Vs ratios (up to 2.1-2.2) in an area of ~5 km diameter beneath Anak Krakatau. In a vertical section (Fig. 10), a transition between two high-Vp/Vs zones at a depth of about 7 km becomes visible. The robustness of this feature of the modelling is tested in a special synthetic model which reveals satisfactory vertical resolution (Jaxybulatov et al. 2011). The Vp/Vs distribution in the vertical section is interpreted to reflect two distinct magma storage

Figure 6 (a) Test for equilibrium using the $K_d[\text{FeMg}]$ between clinopyroxene and melt. The 1963 basalt and 2002 basaltic andesite both result in $K_d[\text{FeMg}]$ values close to the ideal of 0.28 (Putirka, 2008). They have therefore been selected as the melt compositions with which the clinopyroxenes are most likely to have equilibrated. (b) The predicted vs. observed mineral components of diopside + hedenbergite, derived using the two nominal equilibrium melts 1963 basalt (top) and 2002 basaltic andesite (bottom). Both nominal melts give very similar results, in terms of predicted mineral compositions, indicating that both need to be considered viable as potential equilibrium melts.

Figure 7 Results of PTB03 (circles), PB08 (triangles) and NimCal08 (squares) calculations. (a) PTB03 and PB08 results for clinopyroxene erupted 1990-2002. Note that the results of the PTB03 and PB08 models overlap, justifying the choice of the 1963 basalt as an appropriate equilibrium melt. SEE for PTB03 are $\pm 33^\circ \text{C}$ and $\pm 170 \text{ MPa}$. SEE for PB08 is $\pm 150 \text{ MPa}$. (b) Results of the NimCal08 barometer partially overlap with those obtained from the PTB03 and PB08 models. SEE for NimCal08 is at 260 MPa . (c) Pressures and temperatures calculated for the older (1883-1981) clinopyroxenes are slightly higher than for the younger (1990-2002) clinopyroxenes. (d) The P-T fields representing the three sets of calculated pressures and temperatures described in Fig. 7a-c

Table 3. Clinopyroxene-melt thermobarometry and clinopyroxene barometry for Anak Krakatau eruptives. Calculated P and T values are given as average and range

Model	wt. % H ₂ O	Recent clinopyroxene (1990-2002)		Older clinopyroxenes (1883-1981)	
		P (MPa)	T (°C)	P (MPa)	T (°C)
PTB03	n/a	301 (58 to 546)	1117 (1097 to 1140)	492 (272 to 759)	1131 (1112 to 1154)
PB08	2	236 (60 to 389)	n/a	280 (116 to 563)	n/a
	3	303 (127 to 456)		347 (183 to 630)	
NimCal08	2	129 (-110 to 387)	n/a	241 (-172 to 566)	n/a
	3	84 (-155 to 342)		286 (-127 to 611)	

regions, an upper crustal one at <7 km and a mid-crustal one at >7 km. For full details on the seismic tomography, see Jaxybulatov et al. (2011).

7. Discussion

Key factors controlling the ascent of silicate magmas include a.) fractures, b.) states of stress, c.) mechanical properties of the lithosphere, and d.) density contrast between magma and country rock, where lower density lithologies act as barriers for denser magma (cf. Shaw 1980; Stolper and Walker 1980; Gudmundsson 1986; Kuntz 1992; Yang et al. 1999; Putirka et al. 2003). Below Anak Krakatau, the crust can be broadly divided into three main lithological packages (Fig. 2a), and is heavily fractured and faulted, making this volcano particularly well suited for investigating magma ascent and evolution.

Our clinopyroxene-melt and plagioclase-melt thermobarometry produces three distinct and separate sets of P-T estimates which are also independently detected by seismic studies (Harjono et al. 1989; Jaxybulatov et al. 2011; this study), and are consistent with in-situ crystal isotope stratigraphy (Gardner et al. in press) as well as chlorine content in melt inclusions (Mandeville et al. 1996a). This strongly indicates three separate levels of crystallisation and thus three different levels of magma storage. In Fig. 10 and Table 6, the results from plagioclase-melt and clinopyroxene-melt barometry are compared. The plagioclase crystallization is focused in the region of 3-7 (An₅₆₋₇₆) and 23-28 (An₇₇₋₈₈) km while the majority of clinopyroxenes appear to have formed in the region of 7-12 km.

The large compositional variation of plagioclase (An₄₅₋₉₀), the sieve-like textures and complex zoning patterns are features commonly observed in basaltic and andesitic, subduction-related volcanic rocks, and there is a widely held view that they reflect crystallisation in a highly dynamic magma system (e.g. Tepley et al. 1999; Troll et al. 2004; Chadwick et al. 2007). Sieve-like textures may develop as a result of crystallographic-controlled dissolution of crystals in a system where the particular plagioclase composition has become unstable. This can occur by external heating to temperatures above solidus, i.e. during partial melting (e.g. Johannes et al. 1994; Petcovic & Grunder 2003; Årebäck et al. 2008) or by mixing of more Ab-rich plagioclase xenocrysts from a felsic magma into a (hotter) more mafic magma (e.g. Tsuchiyama 1985; Andersson & Eklund 1994; Hattori & Sato 1996). Alternatively, sieve-like textures may arise by rapid 'skeletal' growth at conditions of supersaturation due to rapid undercooling and/or mixing of two magmas to an intermediate composition (e.g. Lofgren 1974; 1980; Kuo & Kirkpatrick 1982; Landi et al. 2004; O'Driscoll et al. 2007). The most sodic plagioclase compositions (An₄₅₋₅₅) did not equilibrate with the basaltic andesite magma but they have compositions that overlap with those found in 1883 rhyodacite and 1981 dacite rocks (Camus et al. 1987; Mandeville et al. 1996a), suggesting a possible origin as resorbed xenocrysts from felsic volcanics

Figure 8 (a) Test for plagioclase and four possible equilibrium melt options. (b) Results of plagioclase-melt thermobarometry, recording two separate regions of plagioclase crystallisation. Plagioclase with medium anorthite content (An₅₆₋₇₆) record shallow crustal storage (3-7 km), while the high anorthite plagioclase crystallised below Moho (23-28 km). SEE for the plagioclase-melt thermobarometer are ± 36 °C and ± 247 MPa, and so the results for 3 and 3.5 wt.% H₂O overlap within the uncertainties.

and intrusives, or from assimilation of felsic crust. Similarly, the most calcic plagioclase (An₇₇₋₈₈) was equilibrated in a basaltic melt similar to the 1963 basalt rather than in a basaltic andesite, as suggested would be the case in Camus et al. (1987). Thus, textural evidence (as also discussed by Gardner et al. 2006; in press) together with the equilibrium tests suggest that although plagioclase crystals of various origins are present, only those of intermediate composition (~An₅₆₋₇₆) crystallized from a melt similar in composition to that of the basaltic andesite in which they occur. The depths calculated for crystallisation of these medium anorthite plagioclases (3-7 km) fit well with previous results of plagioclase-melt geobarometry (Camus et al. 1987; Mandeville et al. 1996a), as well as evidence from crystal isotope

Table 4. Results from plagioclase-melt thermobarometry for Anak Krakatau eruptives. Calculated *P* and *T* values are given as average and range, where appropriate

T (°C)	Saturation surface T (°C)	P (MPa)	wt. % H ₂ O	Nominal melt
1053 (1047 to 1058)	1062	70 (23 to 117)	3.5	2002 basaltic andesite
1069 (1063 to 1074)	1077	136 (81 to 186)	3	2002 basaltic andesite
1098 (1097-1098)	1104	651 (645 to 658)	3	1963 basalt
1132 (1131-1133)	1138	800 (793 to 807)	2	1963 basalt

Figure 9 (a) Test for olivine-melt equilibrium. Only the olivines in equilibrium with the 1963 basalt (Fo_{70-72}) are used in the calculations. (b) Results of all mineral-melt thermometers employed. The temperatures calculated using olivine-melt thermometry overlap with but are also slightly higher than the temperatures calculated for clinopyroxene as well for high anorthite (An_{77-88}) plagioclase, all of which are in equilibrium with the 1963 basalt. The temperatures calculated for plagioclase in equilibrium with the more evolved 2002 basaltic andesite (An_{56-76}) are significantly lower. The SEE for olivine-melt thermometry is at $\pm 29^\circ\text{C}$. Also refer to Fig. 7 and Fig. 8b.

stratigraphy (Gardner et al. 2006; in press) and confirms the presence of a shallow magma storage region beneath Anak Krakatau. The high anorthite plagioclase record a depth of crystallisation that is very much in agreement with the sub-Moho level detected by Harjono et al (1989).

Similarly, the zone of clinopyroxene crystallization (7–12 km) coincides with the magma chamber system identified by Harjono et al. (1989) at a depth of ~ 9 km (with an unknown downward extension). These contrasting depth levels calculated for plagioclase and clinopyroxene crystallization are not contradictory. In contrast, An_{56-76} plagioclase is more likely to preserve a later stage of crystallization than clinopyroxene in mafic magmas (e.g. Bowen 1928), as is also confirmed by the observed textures (Fig. 4d). Therefore, the results of our thermobarometry are in agreement with this order of crystallisation, indicating that the late stage, medium anorthite plagioclase form at shallow level in the crust beneath Anak Krakatau. Harjono et al. (1989) was not able to identify a seismic attenuation zone at depths less than 9 km, but this does not exclude shallower magma storage. As discussed by Gardner et al. (2006; in press), a diffuse zone of small pockets of magma would avoid detection by a broad-scale micro-seismic study such as that of Harjono et al. (1989).

The tomography study performed in tandem with this work, in turn, detected two broad but separate crustal magma storage regions. These are constrained to depths of < 7 km, and > 7 km respectively (Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011), and are consistent with the upper two levels of magma storage detected by our thermobarometry investigation. The tomography study unfortunately suffered from limited resolution at depths greater than ~ 15 km (Jaxybulatov et al. 2011), and the very deep, sub-Moho storage region was not resolvable.

Therefore, the combined thermobarometric, geophysical and isotopic evidence point towards three distinct magma storage regions that feed Anak Krakatau. One at a depth of approximately 3–7 km (An_{56-76} plagioclase crystallisation), another at 7–12 km (clinopyroxene crystallisation), plus yet another very deep storage region at ≥ 22 km (crystallisation of An_{77-88} plagioclase and olivine)

Table 5. Results of olivine-melt thermometry.
Calculated T values are given as average and range

Model	wt% H ₂ O	T (°C) Fo ₇₀₋₇₂
PO07	2	1144 (1141 to 1145)
	3	1127 (1124 to 1128)

7.1 A model for Anak Krakatau

Our results and those from previous work (Camus et al. 1987; Harjono et al. 1989; Mandeville et al. 1996a; Gardner et al. 2006; in press; Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011) indicate that magma emplacement and storage at Anak Krakatau volcano coincides with the major lateral lithological

envisaged to have been characterised through the following stages (Fig. 11):

1) Partial melting of the mantle wedge producing a primary basaltic magma, followed by transport of magma to the mantle – crust boundary. There, magma ascent is halted due to the density contrast between the mantle and the lower crust (at ~22 km depth). The initial melt composition may be considerably influenced by decompressional melting, due to the extensional character of the Sunda Strait (Harjono et al. 1991). The extent of the seismic attenuation zone detected by Harjono et al (1989) implies that this deep storage region is large-scale and likely interconnected. This would, in part, account for the semi-continuous supply of magma to the Anak Krakatau system. Analogues to this may be the “MASH” zones proposed by Hildreth and Moorbath (1988), or the

Figure 10 A comparison of the results of the thermobarometry with the tandem seismic tomography study. The V_p/V_s ratio is obtained from real data inversion. Red dots represent seismic events below the Krakatau complex during the period of study. For full details, see Jaxybulatov et al. (2011). The tomography indicates two distinct magma storage regions in the crust, one above and one below approximately 7 km, which is broadly consistent with the results of the thermobarometry as both sets of results indicate the presence of shallow crustal plus mid-crustal storage regions. The sub-Moho storage region (23–28 km) detected by the thermobarometry was not detected by the tomography due to limitations in resolution for depths greater than 15 km.

boundaries in the crust at 4, 9 and 22 km depth (Fig. 2a, 10). There, density contrasts between the different lithologies is likely a controlling parameter, causing ascending dense magma to stall. Below each of these lithological/density boundaries, lateral transport will cause magma pockets to grow and evolve further. The evolution of magma beneath Anak Krakatau is

more recent “deep crustal hot zones” proposed by e.g. Annen et al. (2006). At this level, crystallisation of high anorthite plagioclase and also olivine takes place. 2) Ascent of basalt takes place either when magma density has decreased by fractionation to below that of the lower crust (~2.95 g cm⁻³, Table 1), or when replenishment of fresh basalt, with associated volatile

release, into the magma chambers from below forces ascent to higher levels.

3) The ascending magma then stalls at a mid-crustal level due to the next major density contrast, this time at a depth of about 9 km (Fig. 11, granite – diorite transition), where the dominant phase of crystallization of clinopyroxene takes place. The euhedral habitus and homogenous composition of all observed and analyzed clinopyroxenes would indicate a sizeable and stable storage region with a semi-continuous supply of magma. The bulk composition of the magma at this level is likely close to the evolved basalt documented in Zen and Hadikusumo (1964), as indicated by clinopyroxene being in apparent equilibrium with this composition at the calculated thermodynamic conditions (Fig. 6). Interestingly, Camus et al. (1987) noted that the composition of the clinopyroxenes found in the 1981 acidic-andesites does not differ significantly from those in the basaltic-andesites, implying a common source for the clinopyroxenes, despite differences in host rock chemistry of different eruptions.

4) Further ascent of magma is triggered by continued crystal fractionation (and possibly also assimilation/mixing with felsic crust/magmas), shifting the magma towards yet lower density (e.g. basaltic-andesite composition) until that of the middle crust ($\sim 2.75 \text{ g cm}^{-3}$) is approached. After this, magma will begin to rise again. Alternatively, replenishment with fresh magma and associated release of gases from below may help force ascent.

5) At a depth of $\sim 4 \text{ km}$, the magma ascent is stalled once more, this time at the granite-sedimentary lithological boundary (Fig. 11). A major phase of An_{56-76} plagioclase crystallization takes place at this level. This shallow storage region is likely made up of a plexus of more or less interconnected pockets of magma dispersed in the crust, as it was not detected in the low resolution micro-seismic study of Harjono et al. (1989). In contrast, the seismic tomography reported here appears to pick up signals of this shallow storage level (cf. Hoffmann-Rothe et al. 2006; Jaxybulatov et al. 2011). It is at this level that magma evolves to its final pre-eruptive composition (i.e. evolved basaltic-andesite), as indicated by the late-crystallized plagioclase of intermediate composition being in equilibrium with this bulk composition at the calculated thermodynamic conditions.

Thermal preconditioning of the upper crust by mafic to intermediate magmas has been suggested to be a major factor in the production of rhyolitic magmas (e.g. Troll et al. 2004; Price et al. 2005; Annen et al. 2006). Such a petrogenic model could be applicable to the Krakatau complex, considering its history of recurring major dacitic-rhyolitic eruptions with intermittent periods of mafic to intermediate magmatism. A steep geothermal gradient and continuous heating of the crust eventually leads to large scale assimilation of country rock, as well as partial melting and recycling of comagmatic intrusives. The idea that the uppermost magma storage region detected ($\sim 3\text{--}7 \text{ km}$) is currently made up of a plexus of magma pockets resulting in a very high

surface-to-volume ratio of the stored magma at that depth, would implicate an efficient heat transfer from magma to crust. The high magma throughput at Anak Krakatau, evident in the average extrusive growth of $\sim 7.5 \text{ cm/week}$, adds to the efficient heating of the crust. The geothermal gradient in one of the 3 km deep wells, about 30 km south-southeast of Krakatau, has been estimated to be as high as 67°C/km (Nishimura et al. 1986). With the model proposed by Price et al. (2005) and the cyclicity of van Bemmelen (1949) (Fig. 11b) in mind, the possible shallowing of the plumbing system detected in this study (Fig. 7c) could be an indication that Krakatau is again in the process of evolving towards a larger and higher silica system culminating in a major explosive eruption. Therefore, continuous seismic and petrological monitoring of Anak Krakatau remains of utmost relevance. This is especially true considering that Indonesia today has the world's fourth largest population, with a much more densely populated proximal area compared to that in 1883 when the latest major ignimbrite eruption claimed 36,000 lives.

Figure 11 (a) Schematic illustration of the magma plumbing system at Anak Krakatau based on thermobarometric and geophysical data. (b) Eruption cyclicity of Krakatau, redrawn and modified from van Bemmelen (1949). Since the ignimbrite eruption in 1883, all analysed rocks erupted from Anak Krakatau have been of basaltic or basaltic-andesite composition, with the notable exception of a single event in 1981, when acidic-andesites were erupted.

Table 6. Calculated pressures converted to depth. Calculated depth values are given as average and range

Model	Mineral phase	wt% H ₂ O	Depth of crystallisation (km)
PTB03	Clinopyroxene 1990-2002	n/a	11.46 (2.55 to 19.96)
PB08	Clinopyroxene 1990-2002	2	11.59 (5.32 to 16.85)
		3	9.23 (2.19 to 14.54)
PTB03	Clinopyroxene 1883-1981	n/a	18.06 (10.49 to 26.68)
PB08	Clinopyroxene 1883-1981	2	13.09 (7.36 to 22.78)
		3	10.75 (4.92 to 20.57)
Plagioclase-melt	Plagioclase An ₅₆₋₇₆	3	5.51 (3.60 to 7.40)
		3.5	2.98 (1.01 to 4.88)
Plagioclase-melt	Plagioclase An ₇₇₋₈₈	2	28.05 (27.98 to 28.12)
		3	23.37 (23.22 to 23.60)

8. Conclusions

Recent magmas erupted from Anak Krakatau record clinopyroxene crystallization at mid-crustal level (7–12 km), within a magma storage region previously identified in micro-seismic and tomographic studies (Harjono et al. 1989; Jaxybulatov et al. 2011). In contrast, plagioclase records both sub-Moho (23–28 km) and upper crustal (3–7 km) depths of crystallisation. High anorthite plagioclase (An₇₇₋₈₈, 11% of datapoints), in equilibrium with a basaltic melt, forms at approximately the crust-mantle interface, coinciding with the deep storage region detected in Harjono et al. (1989). The much more abundant medium anorthite plagioclase (An₅₆₋₇₆, 68% of datapoints), in equilibrium with a basaltic andesite melt, forms in the upper crust, in agreement with the upper storage region detected by Jaxybulatov et al. (2011). This is also in agreement with previous estimates based on plagioclase barometry, chlorine in fluid inclusions barometry and supported by crystal isotope stratigraphy (Camus et al. 1987; Mandeville et al. 1996a; Gardner et al. 2006; in press).

The magma storage regions detected beneath Krakatau coincide with major lithological boundaries in the crust, implying that magma ascent at Anak Krakatau is in part controlled by lateral crustal discontinuities. This, in turn, indicates that, at this volcano, density contrast between magma and bedrock is an important

parameter for magma ascent. Additionally, the extensional character and heavily faulted bedrock in the Sunda Straits (Nishimura et al. 1986), is likely fundamental in providing vertical pathways for magma ascent.

Our study demonstrates that petrology is able to detect magma storage zones in the crust where conventional seismic surveys (e.g. Harjono et al. 1989) fail due to limitations in resolution (e.g. the shallow reservoir below Anak Krakatau). In contrast, seismic surveys may pick up signals of deeper storage regions from which none of the appropriate phenocryst phases remain in the eruptive products. This may be due to e.g. cumulate capture, resorption and/or replacement of minerals. Combined geophysical and petrological surveys, thus offer the highest potential for a thorough characterization of magma plumbing at active volcanic complexes.

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References

- Andersson, U.B. (1997) Petrogenesis of some Proterozoic granitoid suites and associated basic rocks in Sweden (Geochemistry and isotope geology). *SGU Rapp & Medd* 91:216
- Andersson, U.B. & Eklund, O. (1994) Cellular plagioclase intergrowths as a result of crystal-magma mixing in the Proterozoic Åland rapakivi batholith, SW Finland. *Contrib Mineral Petrol* 117:124–136.
- Annen, C., Blundy, J.D. & Sparks, R.S.J. (2006) The Genesis of Intermediate and Silicic Magmas in Deep Crustal Hot Zones. *J Petrol* 47:505–539
- Årebäck, H., Andersson, U. B. & Petersson, J. (2008) Petrological evidence for crustal melting, unmixing, and undercooling in an alkali-calcic, high-level intrusion: the late Sveconorwegian Vinga intrusion, SW Sweden. *Mineral Petrol* 93:1–46.
- Barker, A.K., Holm, P.M., Peate, D.W. & Baker, J.A. (2009) Geochemical Stratigraphy of Submarine Lavas (3[^]5 Ma) from the Flamengos Valley, Santiago, Southern Cape Verde Islands. *J Petrol* 50:169–193
- Bowen, N.L. (1928) The evolution of the igneous rocks: Princeton, New Jersey, Princeton University Press, 334 p
- Camus, G., Gourgard, A. & Vincent, P.M. (1987) Petrologic evolution of Krakatau (Indonesia): Implications for a future activity. *J Volcanol Geotherm Res* 33:299–316

- Caprarelli, G. & Riedel, S.P. (2005) A clinopyroxene–basalt geothermobarometry perspective of Columbia Plateau (NW-USA) Miocene magmatism. *Terra Nova* 17:265-277
- Cashman, K.V. (1990) Textural constraints on the kinetics of crystallization of igneous rocks. In: J. Nicholls & J.K. and Russell, eds. *Modern methods of igneous petrology: Understanding magmatic processes*. Mineral Soc Am Rev Mineral 24:259-314
- Chadwick, J.P., Troll, V.R., Ginibre, C., Morgan, D., Gertisser, R., Waight, T.E. & Davidson J.P. (2007) Carbonate Assimilation at Merapi Volcano, Java, Indonesia: Insights from Crystal Isotope Stratigraphy. *J Petrol* 48:1793-1812
- Deplus, C., Bonvalot, S., Dahrin, D., Diamant, M., Harjono, H. & Dubois, J. (1995) Inner structure of the Krakatau volcanic complex (Indonesia) from gravity and bathymetry data. *J Volcanol Geotherm Res* 64:23-52
- Devine, J.D., Gardner, J.E., Brack, H.P., Layne, G.D. & Rutherford, M.J. (1995) Comparison of microanalytical methods for estimating H₂O contents of silicic volcanic glasses. *Am Mineral* 80:319-328
- Duke, J.M. (1976) Distribution of the Period Four Transition Elements among Olivine, Calcic Clinopyroxene and Mafic Silicate Liquid: Experimental Results. *J Petrol* 17:499-521
- Galipp, K., Klügel, A. & Hansteen, T.H. (2006) Changing depths of magma fractionation and stagnation during the evolution of an oceanic island volcano: La Palma (Canary Islands). *J Volcanol Geotherm Res* 155:258-306
- Gardner, M.F., Troll, V.R., Hart, G., Gertisser, R., Wolf, J.A., & Gamble, J.A. (2006) Shallow-level processes at Anak Krakatau: Crystallisation and late stage crustal contamination. *Goldschmidt Conference Abstracts* A194
- Gardner, M.F., Troll, V.R., Gamble, J.A., Gertisser, R., Hart, G.L., Ellam, R.M., Harris, C. & Wolf, J.A. (in press) Shallow level differentiation processes at Krakatau: evidence for late-stage crustal contamination. *J Petrol* doi:10.1029/2001GC000217
- Ghiorso, M.S., Hirschmann, M.M., Reiners P.W. & Kress, V.C.I. (2002) The pMELTS: A revision of MELTS for improved calculation of phase relations and major element partitioning related to partial melting of the mantle to 3 GPa. *Geochem Geophys Geosystems* 3
- Gudmundsson, A. (1986) Formation of crustal magma chambers in Iceland. *Geology* 14:164-166.
- Hamilton, W.B. (1979) *Tectonics of the Indonesian Region*. United States Geological Survey Professional Paper 1078
- Harjono, H., Diamant, M., Dubois, J. & Larue, M. (1991) Seismicity of the Sunda strait: evidence for crustal extension and volcanological implications. *Tectonics* 10:17-30
- Harjono, H., Diamant, M., Nouaili, L. & Dubois, J. (1989) Detection of magma bodies beneath Krakatau volcano (Indonesia) from anomalous shear waves. *J Volcanol Geotherm Res* 39:335-348
- Hattori, K. & Sato, H. (1996). Magma evolution recorded in plagioclase zoning in 1991 Pinatubo eruption products. *Am Mineral* 81:982-994.
- Hildreth, W. & Moorbath, S. (1988) Crustal contributions to arc magmatism in the Andes of Central Chile. *Contrib Mineral Petrol* 98:455-489
- Hoffmann-Rothe, A., Ibs-von Seth, M., Kneiß, R., Faber, E., Klinge, K. & Reichert, R. (2006) Monitoring Anak Krakatau volcano in Indonesia. *EOS* 87:581-586
- Housh, T.B. & Luhr, J.F. (1991) Plagioclase-melt equilibria in hydrous systems. *Am Mineral* 76:477-492
- Ibs-von Seht, M. (2008) Detection and identification of seismic signals recorded at Krakatau volcano (Indonesia) using artificial neural networks. *J Volcanol Geotherm Res* 176:448-456
- Jarrard, R.D. (1986) Relations Among Subduction Parameters. *Rev Geophys* 24:217-284
- Jaxybulatov, K., Koulakov, I., Ibs-von Seht, M., Klinge, K., Reichert, C., Dahren, B. & Troll, V.R. (2011) Evidence for high fluid/melt content beneath Krakatau volcano (Indonesia) from local earthquake tomography. *J Volcanol Geotherm Res* 206:96-105
- Johannes, W., Koepke, J. & Behrens, H. (1994) Partial melting reactions of plagioclase and plagioclase-bearing systems. In: Parson, I., (ed.) *Feldspars and their reactions*. Kluwer Academic Publishers 161-194.
- Judd, J.W. (1889) The earlier eruptions of Krakatau. *Nature* 40:365-366
- Koulakov, I. (2009) LOTOS Code for Local Earthquake Tomographic Inversion: Benchmarks for Testing Tomographic Algorithms. *Bull Seismol Soc Am* 99:194-214
- Klügel, A., Hansteen, T.H. & Galipp, K. (2005) Magma storage and underplating beneath Cumbre Vieja volcano, La Palma (Canary Islands). *EPSL* 236:211-226
- Klügel, A. & Klein, F. (2005) Complex magma storage and ascent at embryonic submarine volcanoes from the Madeira Archipelago. *Geology* 34:337-340
- Kopp, H., Flueh, E.R., Klaeschen, D., Bialas, J. & Reichert, C. (2001) Crustal structure of the central Sunda margin at the onset of oblique subduction. *Geophys J Int* 147:449-474
- Kudo, A.M. & Weill, D.F. (1970) An igneous plagioclase thermometer. *Contrib Mineral Petrol* 25:52-65
- Kuntz, M. (1992) A model-based perspective of basaltic volcanism, eastern Snake River Plain, Idaho. In P.K. Link, M.A. Kuntz, and L.B. Plat, Eds., *Regional geology of eastern Idaho and Western Wyoming*, *Geol Soc Am Memoir* 179:289-304
- Kuo, L. C. & Kirkpatrick, R.J. (1982) Pre-eruption history of phryic basalts from DSDP legs 45 and 46: evidence from morphology and zoning patterns in plagioclase. *Contrib Mineral Petrol* 79:13-27
- Landi, P., Métrich, N., Bertagnini, A. & Rosi, M. (2004) Dynamics of magma mixing and degassing recorded in plagioclase at Stromboli (Aeolian Archipelago, Italy). *Contrib Mineral Petrol* 147:213-227.
- le Bas, M.J., le Maitre, R.W., Streckeisen, A. & Zanettin, B. (1986) A chemical classification of volcanic rocks based on the total alkali-silica diagram. *J Petrol* 27:745-750
- Lelgemann, H., Gutscher, M., Bialas, J., Flueh, E., Weinrebe, W. & Reichert, C. (2000) Transtensional basins in the western Sunda Strait. *Geophys Res Lett* 27:3545-3548
- Lofgren, G.E. (1974). An experimental study of plagioclase crystal morphology: isothermal crystallization. *Am J Sci* 274:243-273
- Lofgren, G.E. (1980) Experimental studies on the dynamic crystallization of silicate melts. In: RB Hargraves (ed) *Physics of Magmatic Processes*, Princeton University Press, Princeton, New Jersey

- Longpré, M.A., Troll, V.R. & Hansteen, T.H. (2008) Upper mantle magma storage and transport under a Canarian shield-volcano, Teno, Tenerife (Spain). *J Geophys Res* 113 doi:10.1029/2007JB005422
- Lunt, P., Burgon, G. & Baky, A. (2009) The Pemali Formation of Central Java and equivalents: Indicators of sedimentation on an active plate margin. *J Asian Earth Sci* 34:100-113
- Mandeville, C.W., Carey, S. & Sigurdsson, H. (1996a) Magma mixing, fractional crystallization and volatile degassing during the 1883 eruption of Krakatau volcano, Indonesia. *J Volcanol Geotherm Res* 74:243-274
- Mandeville, C.W., Carey, S. & Sigurdsson, H., (1996b) Sedimentology of the Krakatau 1883 submarine pyroclastic deposits. *Bull Volcanol* 96:512-529
- Metrich, N. & Rutherford, M.J. (1992) Experimental study of chlorine behaviour in hydrous silicic melts. *Geochimica et Cosmochimica Acta* 56:607-616
- Mordick, B.E. & Glazner, A.F. (2006) Clinopyroxene thermobarometry of basalts from the Coso and Big Pine volcanic fields, California. *Contrib Mineral Petrol* 152:111-24
- Morimoto, N., Fabries, J., Ferguson, A.K., Ginzburg, I.V., Ross, M., Seifert, F.A., Zussman, J., Aoki, K. & Gottardi, G. (1988) Nomenclature of Pyroxenes. *Mineral Petrol* 39:55-76
- Nimis, P. (1995) A clinopyroxene geobarometer for basaltic systems based on crystal-structure modelling. *Contrib Mineral Petrol* 121:115-125
- Nimis, P. (1999) Clinopyroxene geobarometry of magmatic rocks. Part 2. Structural geobarometers for basic to acid, tholeiitic and mildly alkaline magmatic systems. *Contrib Mineral Petrol* 135:62-74
- Nimis, P. & Ulmer, P. (1998) Clinopyroxene geobarometry of magmatic rocks Part 1: An expanded structural geobarometer for anhydrous and hydrous, basic and ultrabasic systems. *Contrib Mineral Petrol* 133:122-135
- Ninkovich, D. (1976) Late Cenozoic clockwise rotation of Sumatra. *EPSL* 29:269-275
- Ninkovich, D. (1979) Distribution, age and chemical composition of tephra layers in deep-sea sediments off western Indonesia. *J Volcanol Geotherm Res* 5:67-86
- Nishimura, S. & Harjono, H. (1992) The Krakatau Islands: The Geotectonic Setting. *Geojournal* 28:87-98
- Nishimura, S., Nishida, J., Yokoyama, T. & Hehuwat, F. (1986) Neo-tectonics of the Straits of Sunda, Indonesia. *J Southeast Asian Earth Sci* 1:81-91
- Noujiam, A. (1976) Drilling in a high temperature and overpressured area Sunda Straits, Indonesia. *Proc Fifth Annu Conv Indonesian Pet Assoc, Jakarta*, 211-214.
- Oba N, Tomita K, Yamamoto M, Istidjab M, Badruddin M, Parlin M, Sadjiman, Djuwandi A, Sudradjat A, Suhanda T. (1983) Geochemical study of lava flows, ejecta and pyroclastic flows from the Krakatau group, Indonesia. *Rept Fac Sci Kagoshima Univ* 16:21-41
- O'Driscoll B, Donaldson CH, Troll VR, Jerram DA, and Emeleus CH (2007) An origin for harrisitic and granular olivine in the Rum Layered Suite, NW Scotland: A Crystal Size Distribution study. *J Petrol* 48:253-270
- Petcovic, H.L. & Grunder, A.L. (2003). Textural and thermal history of partial melting in tonalitic wallrock at the margin of a basalt dike, Wallowa Mountains, Oregon. *J Petrol* 44:2287-2312
- Price, R.C., Gamble, J.A., Smith, I.E.M., Stewart, R.B., Eggins, S. & Wright, I.C. (2005) An integrated model for the temporal evolution of andesites and rhyolites and crustal development in New Zealand's North Island. *J Volcanol Geotherm Res* 140:1-24
- Putirka, K.D. (1999) Clinopyroxene+liquid equilibrium to 100 kbar and 2450 K. *Contrib Mineral Petrol* 135:151-163
- Putirka, K.D. (2005) Igneous thermometers and barometers based on plagioclase + liquid equilibria: test of some existing models and new calibrations. *Am Mineral* 90:336-346
- Putirka, K.D. (2008) Thermometers and barometers for volcanic systems. 69, p61-120. In: Putirka, K.D. & Tepley, F.E. *Rev Mineral Geochem* 69:61-120
- Putirka, K.D. & Condit, C. (2003) A cross section of a magma conduit system at the margins of the Colorado Plateau. *Geology* 31:701-704
- Putirka, K.D., Mikaelian, H., Ryerson, F. & Shaw, H. (2003) New clinopyroxene-liquid thermobarometers for mafic, evolved, and volatile-bearing lava compositions, with applications to lavas from Tibet and the Snake River Plain, Idaho. *Am Mineral* 88:1542-1554
- Putirka, K.D., Perfit, M., Ryerson, F.J., & Jackson, M.G., (2007) Ambient and excess mantle temperatures, olivine thermometry, and active vs. passive upwelling. *Chem Geol* 241:177-206
- Rhodes, J.M., Dungan, M.A., Blanchard, D.P. & Long, P.E. (1979) Magma mixing at mid-ocean ridges: Evidence from basalts drilled near 22° N on the mid-atlantic ridge. *Tectonophysics* 55:35-61
- Roeder, P.L. & Emslie, R.F. (1970) Olivine-Liquid Equilibrium. *Contrib Mineral Petrol* 29:275-289
- Schlüter, H.U., Gaedicke, C., Roeser, H. A., Schreckenberger, B., Meyer, H., Reichert, C., Djajadihardja, Y., Prexl, A. (2002) Tectonic features of the southern Sumatra-western Javan forearc of Indonesia. *Tectonics* 21 doi:10.1029/2001TC901048
- Schwarz, S., Klügel, A. & Wohlgemuth-Ueberwasser, C. (2004) Melt extraction pathways and stagnation depths beneath the Madeira and Desertas rift zones (NE Atlantic) inferred from barometric studies. *Contrib Mineral Petrol* 147:228-240
- Self, S. (1982) Krakatau Revisited: The Course of Events and Interpretation of the 1883 Eruption. *GeoJournal* 28:109-121
- Shaw, H.R. (1980) The fracture mechanisms of magma transport from the mantle to the surface. In R.B. Hargraves, Ed., *Physics of magmatic processes*, Princeton University Press 201-264
- Shaw, C.S.J. & Klügel, A. (2002) The pressure and temperature conditions and timing for glass formation in mantle-derived xenoliths from Baarley, West Eifel, Germany: The case for amphibole breakdown, lava infiltration and mineral-melt reactions. *Mineral Petrol* 74:163-187
- Siebert L., & Simkin, T. (2002-) *Volcanoes of the World: an Illustrated Catalog of Holocene Volcanoes and their Eruptions*. Smithsonian Institution, Global Volcanism Program, Digital Information Series, GVP-3, (<http://www.volcano.si.edu/world/>)
- Špičák, Ā., Václav, H. & Vaněk, J. (2002) Seismic activity around and under Krakatau volcano, Sunda arc: constraints to the source region of island arc volcanics. *Stud Geophys Geod* 46:545-565

- Stehn, C.E. (1929) The geology and volcanism of the Krakatau group. In: Guidebook for 4th Pac Sci Congr 1-55
- Stolper, E. and Walker, D. (1980) Melt density and the average composition of basalt. *Contrib Mineral Petrol* 74:7-12.
- Sugawara, T. (2001) Ferric iron partitioning between plagioclase and silicate liquid: thermodynamics and petrological applications. *Contrib Mineral Petrol* 141:659-686
- Susilohadi, S., Gaedicke, C. & Djajadihardja, D. (2009) Structures and sedimentary deposition in the Sunda Strait, Indonesia. *Tectonophysics* 467:55-71
- Tepley III, F.J., Davidson, J.P. & Clyne, M.A. (1999) Magmatic interactions as recorded in plagioclase phenocrysts of Chaos Crags, Lassen Volcanic Centre, California. *J Petrol* 40:787-806
- Tregoning, P., Brunner, F.K., Bock, Y., Puntodewo, S.S.O., McCaffrey, R., Genrich, J.F., Calais, E., Rais, J. & Subarya, C. (1994) First geodetic measurement of convergence across the Java Trench. *Geophys Res Lett* 21:2135-2138
- Troll, V.R., Donaldson, C.H. & Emeleus, C.H. (2004) Pre-eruptive magma mixing in ash-flow deposits of the tertiary Rum Igneous Centre, Scotland. *Contrib Mineral Petrol* 147:722-739
- Tsuchiyama, A. (1985). Dissolution kinetics of plagioclase in the melt of the system diopside-albite-anorthite, and origin of dusty plagioclase in andesites. *Contrib Mineral Petrol* 89:1-16.
- van Bemmelen, R. (1949) The Geology of Indonesia, 732 pp. Gov Print Off, The Hague, Netherlands
- Wohletz, K.H. (2000) Were the Dark Ages triggered by volcano-related climate changes in the 6th century? *EOS* 48:F1305
- Yang, H-J., Frey, F.A., Clague, D.A., and Garcia, M.O. (1999) Mineral chemistry of submarine lavas from Hilo Ridge, Hawaii; implications for magmatic processes within Hawaiian rift zones. *Contrib Mineral Petrol* 135:355-372.
- Zen, M.T. & Hadikusumo, D. (1964) Recent changes in the Anak Krakatau Volcano. *Bull Volcanol* 27:259-268