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Cooling history and emplacement dynamics within rubbly lava flows, southern Deccan Traps: insights from textural variations and crystal size distributions

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Abstract

We analyse two representative rubbly pāhoehoe lavas (F3 and F5) from drill cores at Tural-Rajwadi, southwest of Koyna, in the southern Deccan Traps. Low vesicle deformation (0.1 to 0.4) indicates that both lavas ultimately cooled under a low-stress regime. The crystal size distributions (CSDs) of most samples from F5 (especially those from within the core) are not linear but instead show kinks. These kinks are attributed to a rise in plagioclase nucleation due to degassing following the brecciation of the crust. Since it is difficult to constrain cooling time for ancient lava flows, we used the products of nucleation rates (Jt, 1.64×10^{-8} to 1.45×10^{-5} μm^{-3}) and growth rates (Gt, 2.1 to 156 μm) with time. When compared with natural analogues as well as experimental results for basalt crystallisation, these values suggest a much faster lava cooling rate (~1 to 7℃/hr) than a conductive cooling model (≤0.1 ℃/hr). The CSDs for F3 fan with depth suggesting that the lava flow might represent local accumulation (ponding?) in a transitional lava flow field. CSDs for F5 show little variation with depth, with the exception of kinks for samples from the lower crust and core. The relatively higher number density of plagioclase microcrysts in our rubbly pāhoehoe (F5) and their CSD patterns are similar to those measured for transitional lavas by Katz and Cashman (2003). The vesicle data and CSDs indicate that brittle deformation was the primary mode of transition within these lavas. Identifying occurrence of thick ponded lavas within vertical stacks of rubbly pāhoehoe flows in the upper stratigraphic levels of the Deccan Traps are critically important as they demonstrate complex cooling styles, crystallisation histories, and

emplacement dynamics. Transitional lavas such as rubbly pāhoehoe are important components of large CFB provinces such as the Deccan Traps and constitute nearly 46 to 85% of all lava types. Modelling of continental flood basalt provinces should therefore account for these diversities within lavas, and any oversimplified version using end-member morphotypes is unrealistic and untenable.

Keywords: Lava flows, Rubbly pāhoehoe, Crystal size distribution, Nucleation rates, Growth rates, Deccan Traps

Introduction

Continental Flood Basalts (CFBs) represent the largest, most voluminous eruptions on Earth. Individual lava flows have been known to contain over 1000 km^3 and are believed to extend up to 1000 km in length (Vye-Brown et al. 2013; Self et al. 2008; Fendley et al. 2020). Such extensive lava flows can potentially have significant impacts on global atmospheric compositions due to volatile escape during their cooling and, thus, paleoclimate (e.g. Courtillot and Renne 2003; Self et al. 2014; Burgess et al. 2017; Glaze et al. 2017; Jones et al. 2019). Despite this critical importance and years of study, many aspects (e.g. eruption rate and duration, time of emplacement of individual lava flows, etc.) regarding the emplacement of CFB lavas remain poorly understood, and cooling histories of only a few individual flows are known (e.g. Thoradarson and Self 1998). In the absence of modern analogues, the morphology of these lavas provides vital insights into their emplacement, which can be used to infer the evolution of the entire province (Bondre et al. 2004). Thus, it is critical to study the lava morphotypes present within CFB provinces and map their distribution (the frequency with which they occur and their thicknesses).

Lava types have often been distinguished based on their surface morphology into pāhoehoe and 'a'ā (Dutton 1884; Dana 1891; MacDonald 1953; Harris et al. 2017). Pāhoehoe lavas have a smooth undisrupted crust and are ‘crust dominated’, with crusts comprising between 44-58% of the total thickness of the flow, while 'a'ā flows have disrupted crusts and bases, and are 'core dominated', with intact crustal thicknesses of ≤14% (Duraiswami et al. 2014). Apart from these two end-members, transitional lava types (e.g. rubbly pāhoehoe, slabby pāhoehoe, etc.) are commonly recognised across shields, stratovolcanoes and flood basalt terrains (Peterson and Tilling 1980; Kilburn 1981; Wilmoth and Walker 1993; Duraiswami et al. 2003; Bondre et al. 2004; Guilbaud et al. 2005; Sheth et al. 2011; Harris et al. 2017; Óskarsson and Riishuus 2014; Murcia et al. 2014). Katz and Cashman (2003) use the term 'transitional' to refer to lavas identified in drill cores that cannot be classified as either pāhoehoe or 'a'ā. According to them, transitional lavas often have rubbly or brecciated tops and intact bases and are finely crystalline throughout like 'a'ā, but contain near-spherical vesicles, like pāhoehoe. Rubbly pāhoehoe morphotypes from CFBs, similar to these 'transitional' lavas, are 'core dominated' and recognised by the presence of rubble at the top of at least a significant portion of the flow, with an intact base (Keszthelyi 2002; Duraiswami et al. 2003, 2008, 2014; Óskarsson and Riishuus 2014). The spaces between rubble can be occupied by unoxidised glass and/or bole (Duraiswami et al. 2020 and reference therein).

Rubbly pāhoehoe lavas are of particular interest because they contain characteristics of both end-members. The exact conditions that lead lavas to be emplaced as a rubbly pāhoehoe are of great interest considering their observed widespread occurrences (as will be demonstrated in the present study). Such widespread occurrences imply that the conditions that lead CFB lavas to transition are common, while they are much less abundant in other types of eruptions like shield volcanoes (e.g. Katz and Cashman 2003) and stratovolcanoes (e.g. Branca et al. 2011; Lanzafame et al. 2013). Rubbly pāhoehoe lavas also commonly form in the more voluminous Icelandic eruptions (as compared to central and shield lavas, Guilbaud et al. 2005; Óskarsson and Riishuus 2014), and have even been identified on other terrestrial bodies like Moon and Mars (Keszthelyi et al. 2004, 2006). Duraiswami et al. (2008, 2014) and Murcia et al. (2014) have hypothesised that the rubbly pāhoehoe lavas are initially emplaced as inflated pāhoehoe, but as their yield strength increases, they eventually autobrecciate only in the upper surface to form rubble or flow top breccias due to increased cooling and degassing during flow. Detailed flow scale datasets (e.g. vesicle distribution and textural variations) need to be collected to analyse whether this hypothesis applies to all rubbly pāhoehoe lavas. These datasets can also be used to study the effect of strain rate on the crust, which together with temperature, is believed to be one of the primary drivers of transition from pāhoehoe to 'a'ā lavas (Peterson and Tilling 1980; Di Fiore et al. 2021).

Gas bubbles or vesicles form patterns reminiscent of inflation, degassing, and cooling, thereby providing information on emplacement mechanisms and lava rheology (Walker 1993; Cashman and Kauahikaua 1997). Vesicles also preserve the strain rate experienced within the crust at the time of solidification through their shape (Taylor 1934). Similarly, nucleation and growth rates of crystals in a lava flow are sensitive to various factors like the morphology of lava, cooling and strain rate (e.g. Thordarson and Self 1998). By studying these rates or their variations using tools like crystal size distributions (CSD), we can infer the impact of crystallisation on lava morphology and vice-versa (Marsh 1988; Cashman and Marsh 1988). CSDs can also be used to estimate the cooling rate by comparing the patterns of growth and nucleation with experimentally derived CSDs whose cooling rates are known (e.g. Giuliani et al. 2020).

The Deccan Traps CFB province (Fig. 1a) occupies over 500,000 km^2 in Western India (Krishnamurthy 2020), with good exposures of highly dissected lava flows. It is believed to have erupted over ~1 Ma between 66.3 and 65.3 Ma (Sprain et al. 2019; Schoene et al. 2019) and formed a ~1 to 2 km thick volcanic edifice in the Western Ghats. This province provides an excellent opportunity to study CFB lavas due to their highly dissected nature. Thus, in the present study, we present the morphology of lava flows exposed on the surface and within select boreholes from the southern Deccan Traps to show the widespread occurrence of rubbly pāhoehoe lavas. We then use drill cores of two representative rubbly lavas to study their textural diversity, vesicle, and crystal size distributions. We use the vesicle data to estimate the strain rate within these lavas qualitatively. We use the textures and CSDs to gain insights into the cooling rate and emplacement history of the lava flow. We also generate a conductive cooling model using the equations of Wright and Marsh (2016) to compare the cooling rate of an insulated flow with a rubbly pāhoehoe flow of a similar thickness. Finally, we discuss the implications of the data in the present study on the nature of transition of flows within CFBs.

Geological setting

We studied the lavas in thick ghat (mountain pass) sections south of Pune (Fig. 1b, Table 1) that included important locations around Mahad, Mahabaleshhwar, Satara and Koyna in southern Deccan Traps. In the Sinhagad–Bhuleshwar hill range (Location 1, 2 and 3 in Fig. 1b), south of Pune, rubbly pāhoehoe flows dominate in the Sinhagad Fort (Wilkins et al. 1994), Katraj ghat and Diveghat sections (Duraiswami et al. 2014). Rubbly pāhoehoe flows occur as thick (20-50 m) sheet flows (Fig. 2a) or as thick (30–70 m) compound flow fields with multiple flow lobes that may be topped by fine (Fig. 2b) or coarse rubble (e.g. Fig. 2c). In the Khambatki–Mandhar Devi section (Location 4 and 5 in Fig. 1b), south of Bhor, the rubbly pāhoehoe flow fields, belonging to the Ambenali and Mahabaleshwar Formations are separated by a compound megacrystic (M4) lava flow (Shandilya et al. 2020). This stack of rubbly pāhoehoe flows, belonging to the Wai Subgroup, overlie the strongly compound pāhoehoe flow fields of the lower stratigraphic formations in the Varandha ghat near Mahad (Subbarao et al. 2000). Thick sheet rubbly pāhoehoe flows with jointed cores and separated by red bole horizons (e.g. see Fig. 2d, e) are recorded in the Ambenali-Mahabaleshwar and Parsarni Ghats (Locations 6 and 7 in Fig. 1b; Jay and Widdowson 2008). The plateaux of Kas and Thoseghar, west of Satara and around Koynanagar (Patan–Dapharwadi ghat, Arle–Gudhe ghat), and the Kumbharli ghat in the Western Ghat escarpment (Location 8 to 12 in Fig. 1b) expose thick and morphologically diverse pāhoehoe, transitional (slabby and rubbly pāhoehoe) and 'a'ā' flow fields (Duraiswami et al. 2017). The surface lava flows and those in the boreholes at Tural–Rajwadi (Location 13 in Fig. 1b) were delineated as rubbly pāhoehoe (Monteiro et al. 2019). The Koyna deep boreholes (Location 14 to 15 in Fig. 1b; Gupta et al. 2017; Roy 2017) located at Phansavle, Khadi, southeast of Tural–Rajwadi also contain rubbly pāhoehoe lavas.

The rubbly pāhoehoe flows from southern Deccan Traps constitute about 46 to 85% of all lavas morphotypes (Table 1). They, thus appear to be more abundant than the Hawaiian 'transitional' lavas studied by Katz and Cashman (2003), which are similar in morphology to CFB rubbly pāhoehoe lavas. In Hawaii, Katz and Cashman (2003) recorded an abundance by volume of 8 to 11% for 'transitional' flows from a proximal (SOH-4) and a distal (KP-1) borehole. By contrast, in the southern Deccan Traps the proportion of rubbly pāhoehoe flows varies from a maximum of 94.9% in the KBH-09 borehole from Khadi in the Koyna region to about 35.8% in the Tural-Rajwadi borehole (Table 1). This is comparable to higher proportion of surface rubbly pāhoehoe flows exposed in the region, thus making it an important morphotype that needs to be represented in the study of flood basalt provinces.

Materials and methods

Drilling and logging

The Tural–Rajwadi borehole was drilled up to 203 m below ground level (bgl) close to the Koyna deep boreholes. The similarity in morphology and thickness of the Tural-Rajwadi and Koyna lavas indicates they were likely emplaced under similar conditions. Logging was undertaken based on sound modern concepts in physical volcanology, keeping in mind the internal geometry of transitional lavas based on our experiences gained from surface outcrops in the area. Two of the lava flows namely F3 and F5 of six identified are rubbly (Fig. 3), and were selected for detailed volcanological descriptions and further studies.

Probability distribution function (PDF)

The thickness of selected lava flows was measured using steel tape. We plotted probability distribution function of the thicknesses of the flows (Fig. 4a) using a Python script to compare the flows from Koyna with those from Tural-Rajwadi and elsewhere. Specifically, the SciPy, NumPy, and Matplotlib libraries were used to define kernel densities (SciPy and NumPy) and plot (Matplotlib) the PDF. The thickness data for the surface lava flows were taken from Duraiswami et al. (2017). For the subsurface drill cores, thickness data from the KBH-1 borehole was taken from Sinha et al. (2017) and for the KBH-7 borehole, from Mishra et al. (2017).

Vesicle distribution

Vesicles and their distribution in lavas are crucial for understanding the emplacement dynamics of lava flows (Aubele et al.1988; Walker 1993; Cashman and Kauahikaua 1997; Self et al. 1998). We attempt quantitative measurements of vesicle distribution, wherein the vesicles from drill cores were physically studied for their size and frequency. We used the length and width of these vesicles to estimate strain qualitatively.

Vesicles of these samples were studied by tracing images (Supplementary Fig. 1) obtained from drum scanner and measuring their long (length) and short (width) axes using a Vernier's callipers. Most vesicles were approximately spherical, and it was assumed that the cross-sections of the measured vesicles were across the diameter of these vesicles, since this is the most probable result for a plane passing through randomly distributed spheres in a three-dimensional body (Saltikov 1967; Higgins 2000). The frequency of the vesicles or number of vesicles per unit area (number/cm^2) and density of vesicles (area of vesicles per unit area of the sample, unitless) were also calculated using these tracings. The length (l) and width (w) of individual vesicles were then used to calculate the deformation index (reported as mean deformation index for individual cores) given by Taylor (1934) as:

$$D = \frac{(l - w)}{(l + w)} \quad (1)$$

Textural variations

We selected samples for textural and petrographic studies based on macroscopic observations of drill cores from the two lava flows. Selected samples were cut into circular discs using a slab saw, and thin sections were prepared. These thin sections were studied using a Nikon Research microscope fitted with a digital camera at the Department of Geology, Savitribai Phule Pune University, Pune. A total of 31 thin sections from F3 and 15 from F5 were prepared to study the textural variations within the flow. Representative photographs documenting the textures observed within the two lava flows are presented in Figs. 6 and 7.

Thermal model

There exist well-developed conductive cooling models developed for lava lakes (e.g. Kirkpatrick 1977; Wright and Marsh 2016), but little work has been done for inflating sheet lobes (e.g. Keszthelyi and Denlinger 1996, describe a thermal model by studying initial cooling in a small lava toe). However, Hon et al. (1994) demonstrated that an inflating hummocky pāhoehoe lobe shows a similar relationship of cooling rate with depth as that of a lava lake. Therefore, we use the more robust model of Wright and Marsh (2016) to model the cooling rate with depth of a lava body with a thickness comparable with that of lava flow F3 (50 m).

Wright and Marsh (2016) developed a conductive cooling model by studying the emplacement and cooling of active Hawaiian lava lakes. This model allows for a relatively accurate estimation of the advance of isotherms within a natural body since it was developed by studying such systems. The depth of a given isotherm is given by the formula:

$$S(t) = 2b(\kappa t)^{\frac{1}{2}} \quad (2)$$

where κ is thermal diffusivity, *t* is time, and *b* is a constant of order unity, which is a function of melting temperature, latent heat (H), specific heat (C_P), contact temperature (T_O) and magma temperature (T_m).

The value of the constant b can be determined using the equation

$$b = \frac{6 \times 10^{-8} T^2}{\kappa^{-1/2}} \quad (3)$$

where *T* is the temperature of the isotherm in degree Celsius.

Lavas cool faster at the upper surface than the base of the flow due to radiative cooling and the effects of wind and rainfall. Thus, the final depth of solidification occurs at ~60% of the total flow depth (based on the Alae Lava Lake data; Wright et al. 1976). To account for this, the isotherm depth, S(t), has been multiplied by a factor of 2/3 for the isotherms moving upwards from the base of the flow, effectively slowing down the advance of isotherms from the bottom in the model. Equations (2) and (3) were used in a Python script to generate a conductive cooling model for an insulated 50 m thick body.

Crystal size distributions (CSD)

Representative areas within thin sections (23 from the lava flow F3, 14 from lava flow F5) free of xenocrysts and zoned plagioclase crystals (if any) were chosen to measure plagioclase CSDs. This approach was taken to maximise the measurements of crystals that formed within the system, though a minor contribution from small sections of larger pre-existing crystals cannot be ruled out. In order to investigate the crystal growth changes as the lava flows cooled, a set of thin-sections were taken at noted depth intervals within the flows. Observations in the present study are constrained by the fact that these thin sections represent frozen melt and they record the nucleation and growth of crystals at the flow depths at which they are sampled before the flow stagnated (~60% crystallisation according to Marsh 2013). Any interstitial crystallisation thereafter may not reflect the emplacement dynamics of the lava flow, since at this stage, melt cannot travel as freely and crystallisation is more a function of interstitial melt composition than cooling rate (this is the closure problem, which is discussed in detail by Higgins 2002). This crystallisation would be recorded by the smallest crystal sizes. According to Higgins (2000), populations for the smaller crystal sizes may not be correctly recorded in thin sections due to their intersection probabilities being low. Use of scanning electron microscopy (SEM) and other imaging techniques to try and mitigate these effects may help, but they do not completely eliminate this problem, since it is a function of how the rock is cut and not the resolution of the image analysis. However, as mentioned above, these smaller size classes may not entirely preserve the emplacement dynamics of the lava flows, which is what we hope to examine in this study, and this is why CSDs were calculated only from microscope images.

Plagioclase from the digitally photographed thin sections were isolated and measured using Fiji (Schindelin et al. 2012), an open-source statistical image analysis software (also see Antonelli et al. 2019 for a description of the image analysis methodology). After semi-automatically thresholding the plagioclase crystals, those that were touching were inspected and separated manually by drawing a pixel-wide line and re-binarising them (for efficacy of this method, see Supplementary Fig. 2). Crystal sizes recorded were the length of the best-fit ellipse. In total, over 200 crystals were measured for each slide (Supplementary Table 2), allowing for statistically reliable measurements of crystal size distributions (Mock and Jerram. 2005, Morgan and Jerram 2006). These recorded sizes were binned by creating geometrical size classes based on the maximum and minimum crystal sizes for each sample using a Python script. Volumetric corrections were made in line with Zieg and Marsh (2012), since our aim, similar to theirs, is to ascertain variations within individual lava flows, i.e.

$$\text{Populationdensity, } n = \frac{Na}{L_i \Delta L_i} \quad (4)$$

where *Na* is the number of crystals per unit area within a particular size range (within a bin), L_i is the midpoint of the bin and ΔL_i is the width of that bin.

Fiji was also used to measure the modal proportions of the glass and major mineral components of the selected lava flows.

We also plotted the best-fit lines through the generated CSD graphs to quantify effective nucleation and growth rates. Since our CSDs are slightly curved, provisions were made to calculate and compare one, two and three best-fit lines for each sample CSD curve to see which one would fit best, using an algorithm in Python. The algorithm determined the best-fit line by plotting all permutations of lines possible and measuring the R^2 coefficient of each line. The line (or pair of lines) with the best mean R^2 coefficient was selected as the best-fit line(s). Most samples required only one best-fit line, while some required two best-fit lines and no samples required three best-fit lines.

The calculation of effective nucleation and growth rates require independent quantification of time. Since the Tural–Rajwadi lavas belong to an ancient volcanic province (~66 Ma) which cooled prehistorically, it is difficult to estimate the time taken for their complete emplacement. Equations such as that of Hon et al. (1994) and Kirkpatrick (1977) have morphological constraints on them, which allow for them to be used only for certain types of lavas, or in specific initial conditions. For instance, the equation by Hon et al. (1994) is only applicable for sheet or hummocky pāhoehoe lavas which are insulated. Hence, we used the formulae of Marsh (1988) and Zieg et al. (2002) to calculate the product of growth and nucleation rates with time. Such quantities have been previously reported from CSD studies, when it was difficult to constrain time (e.g. Crisp et al. 1994). These quantities allow for meaningful comparisons of nucleation and growth rates and their variations from within the same flow. The formulae used are:

$$\text{Productofgrowthrateandtime, } Gt(\mu m/s) = \frac{-1}{m} \quad (5)$$

$$\text{Productofnucleationrateandtime, } Jt(/\mu m^3) = \frac{-e^c}{m} \quad (6)$$

where *m* is slope and *c* is intercept determined from the best-fit lines.

Results

Physical volcanology

Lava flow F3 is 54.4 m thick (75 to 129.5 m bgl) and shows a simple sheet geometry. It has a 3 m thick crust, 50.7 m thick core and 0.7 m thick basal vesicular zone. The low crust to core ratio (0.06) indicates that the flow is 'core dominated', a common feature of transitional or 'a'ā flows (Duraiswami et al. 2014). Since there is no breccia either at the top or at the bottom of the flow, it is clear that the flow is not an 'a'ā. Therefore, this flow is classified by us as a transitional flow. The vesicles in the crust of F3 are spherical and increase in size from the top to bottom of the crust, indicating a low shear strain during emplacement (Duraiswami et al. 2008, 2014). The core of F3 is fine-grained and massive with glomeroporphyritic and microporphyrytic plagioclase aggregates. F3 shows moderately spaced joints up to 100 m bgl, after which it is intensely jointed up to 107 m bgl. Intense jointing indicates possible post emplacement fracturing. The lower basal vesicular zone has few elongated vesicles. The flow is separated from F4 by a red bole. Lava flow F5 is an 18.2 m thick flow (170 to 188.2 m bgl), with a 6.2 m thick crust, 11.4 m thick core and a 0.6 m thick basal vesicular zone. The upper part of the crust (up to 5.3 m) is brecciated, while the lower crust (0.7 m) is intact. The presence of brecciation suggests a high flow rate (Duraiswami et al. 2003). We classify this flow as a rubbly pāhoehoe flow (Duraiswami et al. 2003) based on the presence of flow top breccia (Duraiswami et al. 2008, 2014). A thin (~2 cm) glassy zone marks the contact between F5 and F6.

Probability distribution function (PDF)

Compared to the thickness of the lava flows in the Koyna region exposed either on the surface or in drill cores (Fig. 4a), we find that the two chosen flows F3 and F5, are not outliers but instead represent the rubbly lava flows from the Southern Deccan Traps. Further, inferences about emplacement may be applicable to many lava flows in the southern Deccan Traps and elsewhere. Comparing with the lava core to total thickness ratios for other CFBs (Fig. 4b), the flow F3 is unique for its thickness, although there are thinner aphyric rubbly pāhoehoe lavas in the Icelandic Neogene flood basalt province (Kumlafell, Hólmatindur, and Hjálmadalur group; Óskarsson and Riishuus, 2014) with similar crust to thickness ratios (Fig. 4b). Furthermore, the thicknesses of lava flows measured in the present study are similar to those for Columbia River Basalt Province, North Atlantic Igneous Province, Karoo, Emeishan and other CFB provinces (Self et al. 2021 and references therein).

Vesicle distribution

The vesicle frequency and number density for both flows are presented in Supplementary Table 3. For F3, the vesicles were found in the upper 2.8 m of the crust, and the lower 3.5 m of the flow (Fig. 3) suggesting that the vesicle pattern is reminiscent of a thick sheet inflated pāhoehoe (Aubele et al. 1988). The vesicles in the crust range in size from 0.2 to 5.4 cm with an average length of 0.8 cm. The vesicle frequency ranges from 0.2 to 1.8 cm^{-2} and vesicular density ranges from 0.01 to 0.08 for the lava flow. Two distinct populations of vesicles are observed near the base of the flow. The upper population, at a depth of around 51.4 m within the flow, shows maximum vesicle length of 1.6 cm and average vesicle length of 0.5 cm. In contrast, the lower vesiculation, seen at a depth of 53.8 m, shows maximum vesicle length of 0.8 cm and average vesicle length of 0.3 cm. Lava flow F5 shows an erratic pattern of vesicle sizes and frequencies in the upper parts of the crust. These variations are attributed to the presence of rubbly clasts above the crust due to localised auto-brecciation.

The vesicle deformation index (VDI) for the two flows is presented in Supplementary Table 4. For lava flow F3, samples from the crust (sample no. 50 and 53) showed a mean deformation index of 0.4 and 0.2 while the VDI is 0.3 and 0.2 at the base (for sample no. 336 and 345), respectively. The vesicles at the top of the crust, especially the top-most sample (sample no. 50) with the largest mean vesicle length and VDI, show a higher degree of vesicle deformation. For lava flow F5, the maximum deformation index is 0.3 (for samples 500, 510, 512, 514, 520), with most samples showing a deformation index of 0.1–0.2, including samples from within the rubble. Similar ranges and variations of vesicle deformation index were observed in the case of other slabby and rubbly pāhoehoe flows from the Deccan Traps (Duraiswami et al. 2003, 2014).

Textural diversity

Lava flow F3 shows large textural diversity and variations in modal proportion of glass and crystallising phases (Fig. 5, Supplementary Table 5). The thin crust (~3m) of the flow is fine-grained and contains a greater proportion of glass with respect to the lower core. The uppermost 1.7 m of the crust is extremely fine-grained, glassy with plagioclase microcrysts (Fig. 6a). Lower down, the crust continues to be fine-grained but contains fine needle-like quenched plagioclase microcrysts, suggesting a higher degree of undercooling (ΔT) of ~101 to 200°C; (Lofgren 1974; Corrigan 1982; Shea and Hammer 2013; Fig. 6b). Glomeroporphyritic microphenocrysts of plagioclase (Fig. 6c) within intersertal groundmass occur at the base of the crust, at a depth of 3 m. The upper part of the lava core at 7.56 m shows fine-grained diktytaxitic texture with little intersertal glass (Fig. 6d). Beyond this depth, the groundmass of the lava core shows a typical subophitic texture (Fig. 6d), where clinopyroxenes partially enclose plagioclase crystals. At a depth of 17.61 m, the core exhibits a relatively coarse-grained texture, with a large amount of glass. The coarse grain size is a result of the relatively slow cooling rate and low nucleation rate. Microporphyritic plagioclase crystals (Fig. 6e), some of which exhibit very feeble zoning (Fig. 6f), can be seen at various depths (10.45 m, 14.29 m, 30.71 m, 32.36 m, and 41.28 m) in the lava core. Few plagioclase crystals in the lava core also exhibit skeletal morphologies suggesting reabsorption during cooling.

The relative thickness of the crust of lava flow F5 (i.e. 6.2 m) is much larger than that of F3 (See Fig. 3), yielding a smaller lava core to total flow ratio (Fig. 4b). The crust exhibits a predominantly fine-grained glassy texture. The microvesicles within the rubbly clast overlying the intact crust range in size from 250 to 500 µm and are more or less spherical (Fig. 7a) to slightly oval (Fig. 7b). In the vesicular crust, sporadic microphenocrysts of plagioclase with melt inclusions are observed (Fig. 7c). Patches of devitrified glass in the fine-grained groundmass are also observed. At the base of the flow crust (6.26 m), the microvesicles are oval (Fig. 7d) and significantly larger (500 µm–2 mm) than those in the upper part of the crust. Plagioclase microphenocrysts are also seen within the lava crust and core. The core of the flow beyond 6.26 m exhibits a relatively coarser groundmass (Fig. 7e) compared to that of the crust. Below the depth of 14.86 m, the basalt from the core generally shows larger plagioclase and opaque crystals in the groundmass (Fig. 7f).

Crystal size distributions

The lengths and widths of the measured plagioclase crystals and their summary statistics are presented in Supplementary Table 2. The depth-wise measured CSDs of various samples from lava flows F3 and F5 show different patterns (Fig. 8). In the thicker lava flow F3 (Fig. 8a), the CSD lines from the crust (TW050, TW053) are clearly distinguished from those of the core. They also contain the largest plagioclase crystals, higher population density of the smallest crystal sizes (Fig. 8a) and generally lower average crystal size (Fig. 5, Supplementary Table 5). In the thick lava core of F3, the upper parts and lower core have different trends. Those from the former tend to mimic the patterns of the crust, suggesting higher rates of cooling. Overall, the samples from F3 show a distinct fanning pattern (Cashman and Marsh, 1988; Higgins 2002), indicating a general decrease in growth and nucleation rate with depth and longer cooling time scales.

In contrast, samples from F5 show little variation with depth. Samples from the breccia and crust show similar CSD patterns, which contrast with the lower core that shows distinct kinks related to a sudden rise in growth and nucleation rates. The calculated slopes and intercepts for these CSDs for both flows are presented in Supplementary Table 6. The best-fit lines based on the computed slopes and intercepts are plotted in Supplementary Fig. 3. Where there are two slopes for a single sample, the algorithm plotted two best-fit lines, with Slope 1 and Intercept 1 representing the best-fit line for the smaller crystal populations and Slope 2 and Intercept 2 for the larger crystal populations. It follows that Slope 2 could be influenced, at least partially, by inherited crystals and phenocrysts that did not crystallise within the lava flow (Jerram et al. 2003; Marsh 2013). In general, most samples from F3 needed only one best-fit line, while most samples from F5 required two.

The slope and intercept values were used to calculate the product of nucleation (Jt) and growth rate (Gt) with time (Table 2). For lava flow F3, the product of effective nucleation rate with time (Jt) increase from 1.64×10^{-8} µm^{-3} to 2.49×10^{-6} µm^{-3} up to a depth of 8.35 m (Fig. 9). The product of growth rate with time (Gt), however, reduces from 156 to 36.85 µm over the same range. Little variation for both Jt and Gt with depth can be observed for the core beyond ~8 m (Fig. 9). For lava flow F5, a more erratic pattern is seen within the rubble for Jt, while Gt remains more or less constant. In the intact crust beneath the rubble however, the product of nucleation and growth rates with time show a similar pattern as that of F3 (Fig. 9). Jt varies from 8.31×10^{-8} µm^{-3} at a depth of 4.87 m to 7.25×10^{-6} µm^{-3} at a depth of 6.26 m, while Gt varies from 70.96 to 21.07 µm at the same depths. Jt shows low values (Fig. 9) in the core for larger crystal populations (initial crystallisation). Note that in Fig. 9, where two data points are present for a particular sample, products of nucleation and growth rates with time were plotted for both slopes and intercepts. Slope 1 and Intercept 1 from the samples were plotted against each other in Fig. 10. These plots reveal a very interesting trend, with higher slope and intercept values recorded at shallower depths for F3, and a reverse trend for F5, where samples from the core, i.e. deeper within the lava flow show higher slope and intercept values in general.

The product of growth (Gt) and nucleation rates (Jt) with time (Fig. 10) show an approximately linear relationship (though there are certain outliers). For F3, Jt shows a general decreasing relation with depth, while the opposite is true for F5. Similarly, as depth increases, F3 shows a generally increasing Gt while F5 shows a general decreasing Gt with depth. These trends indicate the possibility of insulation for F3, where the core has relatively lower nucleation and growth than the crust. Similarly, F5 shows a higher nucleation rate in the core than in the crust for the smaller crystal sizes (left of the kink), which indicates that the core experienced degassing induced nucleation. It should be noted that the plots use Gt and Jt for the smaller crystal sizes where kinks are present. Complementarily, the initial (larger crystal sizes on the right of the kink) Jt values for F5 are lower in the core, resulting from the insulation of the flow during initial emplacement as a pāhoehoe.

Thermal model

Two plots were generated: a colour gradient plot depicting the top solidification front cooling history before it substantially interacts with the bottom solidification front (Fig. 11a) and a plot showing variation in average cooling rate with depth (Fig. 11b). The conductive cooling model in Fig. 11a shows how the isotherms advance to different depths over time. In general, as depth increases, the rate of isotherm advance decreases exponentially. The shaded portion in the figure shows the range over which crystallisation would occur within basaltic lava (1000□ to 800□). The model reveals that the cooling rate reduces exponentially with depth, from the order of about 100 □/hr at the extreme top of the flow to about 0.1 to 1 □/hr within what would be the crust of the flow, to the order of 0.01–0.005 □/hr well within the interior of the flow (Fig. 11b).

Discussion

Our study provides new datasets on the internal structure, textures, modal variations, vesicle distribution, and plagioclase CSDs from two rubbly pāhoehoe lavas from southern Deccan Traps. We use these to provide important insights on cooling histories and emplacement dynamics that can be used to improve the existing models for rubbly pāhoehoe flows in CFB setups.

Inferences on emplacement style

Lava flow F3 shows a pattern of increasing vesicle size and decrease in frequency due to growth and coalescence of gas bubbles (see Supplementary Fig. 1), similar to that of a typical pāhoehoe crust (Aubele et al. 1988; Wilmoth and Walker 1993; Duraiswami et al. 2002). The vesicularity for the flow (8 %) is too low compared to the Alae lava lake (~40% from near the surface to about ~10% at 3 m depth; Peck 1978; Wright and Peck 1978). The relative thickness of the crust is too low for a pāhoehoe crust (Katz and Cashman 2003; Duraiswami et al. 2014). Quenched plagioclase crystals (e.g. Slide TW053, Fig. 7b) formed during escaping volatiles (degassing) and undercooling, similar to what was reported by Lipman et al. (1985), are present deep within the crust. These observations make us believe that despite the low deformation of the vesicles in the undisrupted crust, the flow is in its incipient stage of transition and most likely grades laterally into a rubbly pāhoehoe. Flow F3 also shows a general fanning for CSDs with depth, similar to what was observed for the Makaopuhi lava lake (Cashman and Marsh 1988). The CSDs thus indicate that the flow F3 from the Tural–Rajwadi drill core may represent a local accumulation or ponding. This could easily account for the unusually low thickness of the crust as well as the low strain regime interpreted from the vesicle distribution. Apart from this, there is immense textural variation within the flow which provides evidence for the diverse cooling regimes prevalent during crystallisation. These regimes (textural variations) recorded in the flow could indicate episodes of crust stabilisation and overturning as observed in Hawaiian lava lakes (Cashman pers. comm. 2020).

Lava flow F5 shows a more irregular distribution of vesicle size and frequency that is expected when the crust disrupts (Macdonald 1953). The vesicles, however, show a low deformation index, indicating that these fragments formed under relatively low strain conditions, similar to pāhoehoe (Peterson and Tilling 1980). This could imply that flow F5 was initially emplaced as a pāhoehoe flow, inflated and was brecciated well after the crust had cooled. The CSD patterns show little variation with depth, a character of transitional lava flows (Katz and Cashman 2003). The CSDs from the rubble do not show any kinks, indicating that they had crystallised to a solid crust well before strain related fracturing and induced degassing occurred. When subjected to increased strain, the solid crust underwent brittle deformation, leading to autobrecciation. Increased autobrecciation promoted greater volatile escape (increased degassing) that led to an increase in nucleation rates. All these factors contributed to changes in lava morphology and internal structure leading finally to preservation as rubbly pāhoehoe. We envision that the crustal brecciation process cannot occur too rapidly or at the same time within the entire lava flow. If this happens, the lava will cool too quickly and produce a juvenile crust that would prevent volatile escape and preservation of the vesicles, since the lava flow viscosity increases exponentially during crystal nucleation and growth at subliquidus conditions (Marsh 2013).

Inferences on cooling rates

The CSDs for both lavas show a complex variation in slope and intercept with depth. Such variations are most likely a result of several dynamic factors acting prior to eruption (pre-eruption) in the plumbing system, during fountaining (syn-eruption e.g. Crisp et al. 1994; Lipman et al. 1985) or during emplacement (post-eruption) along the transport path due to processes like changes in volatile content, occurrences of breakouts and blockages (e.g. Swanson 1973; Wilmoth and Walker 1993). The resulting spatiotemporal variations affect crystallisation of lava emplaced at different depths differently, giving rise to variations in slopes and intercepts for CSDs with depth.

A comparison of the slope and intercept, along with Gt and Jt calculated for other basaltic lavas of different morphologies viz. pāhoehoe lava lake (Cashman and Marsh 1988) and an 'a'ā flow field (Crisp et al. 1994), as well as experimental set ups measuring cooling rates and their effect on textures (Giuliani et al. 2020), is shown in Fig. 12. Only long axis measurements of plagioclase crystals from the dataset of Crisp et al. (1994) were chosen (since this is similar to what we measured in the present study) and split into three categories which are 1) partially insulated samples consisting of channel and pāhoehoe pond samples; 2) the non-insulated samples consisting of samples from 'a'ā lobes and spatters collected syn-emplacement; and 3) the samples collected post-emplacement. The trend of samples from the present study (flows F3 and F5) match well with the trend of the flow field from Crisp et al. (1994), implying that the samples from the present study may have experienced a cooling history similar to that of their pāhoehoe-'a'ā flow field. The data from the Makaopuhi lava lake (Cashman and Marsh 1988) do not share the same trend as the other data, showing lower Gt and Jt values, which would be expected, since the lake was insulated. An interesting phenomenon observed here is that the slopes and intercepts of our flows are comparable with those of Giuliani et al. (2020) when they subjected the melt to a cooling rate of 1 and 7 □/hr. Since cooling rate has an important influence on the size and texture of the crystals within the lava flow, we propose that both F3 and F5 had cooling rates comparable with these experimental set ups, i.e. the order of 1–7 □/hr. The modelled rate of cooling (0.005 to 0.01 °C/hr; Fig. 11) is much lower than what is observed through the textures and crystal size distributions (Fig. 11), indicating the role of a lack of insulation of the crust in increasing the cooling rate of lava. It should be noted that we should not expect the growth and nucleation rates to match precisely with these experimental set-ups due to differences in emplacement and cooling duration, volatile content, initial temperature of eruption, etc.

Inferences on the nature of transition in CFB rubbly pāhoehoe flows

The most commonly proposed mechanism for transition between pāhoehoe and 'a'ā lava flows is a viscous rupture of the crust under high shear stress exceeding the yield strength occurring within lavas (Peterson and Tilling 1980; Di Fiore et al. 2021). The vesicle data results in the present study indicate that the crust for F5 cooled (and solidified) at a relatively low-stress regime. The solid crust then experienced brecciation by undergoing brittle deformation, by experiencing shear stress beyond the yield strength of the crust. We imagine two possible scenarios under which such a transition could occur:

i) The crust's yield strength reduces as it cools from plastic lava to an increasingly brittle crust. Under such a condition, continued endogenous transfer of lava could lead to continued inflation and increasing shear stress experienced by the crust. The cooling crust could have its yield strength drop below the shear stress it undergoes and, being brittle, would autobrecciate to rubble. In this scenario, it is reasonable to expect that the sheet lobe would brecciate episodically as the yield strength and shear stress vary spatially (e.g. hotter, more plastic crust is expected towards the flow front). There might also be local regions of low shear stress, e.g. due to ponding such as that envisaged in F3 where no brecciation would occur.

ii) Variable lava supply due to variable eruption rates at the vent or blockages, surges, etc., within the flow field could lead to increased shear stress on the crust, causing autobrecciation to a rubbly top.

The first scenario can be expected of high volume CFB lavas, where lava flow emplacement is likely to progress for extended periods, spanning several years (e.g. Thoradorson and Self 1998). Such a scenario can account for the rare transition over a complete spectrum of pāhoehoe—slabby pāhoehoe—rubbly pāhoehoe—cauliflower 'a'ā—'a'ā transitions recorded in a CFB set-up (e.g. Murcia et al. 2014). It can also explain the significantly higher proportion of rubbly pāhoehoe lavas documented in the present study and in other CFBs (e.g. Keszthelyi et al. 2001; Rosetti et al. 2018). The second scenario can account for sporadic brecciated lobes within dominantly pāhoehoe flow fields (pāhoehoe—rubbly pāhoehoe and pāhoehoe—'a'ā; e.g. Brown et al. 2011; Duraiswami et al. 2014). Thus, within CFBs, it is possible to have large brecciated sheet lobes emplaced over an extended period of time as well as sporadically brecciated flow lobes emplaced due to variable lava supply.

Viscous rupture as a primary mechanism for the transition of lavas within CFBs seems likely, but is a rarely documented phenomenon, probably due to a high lava flux sustained over time. Under these conditions, viscous ruptures resulting from shear would tend to anneal rather than propagate, as recorded by Duraiswami et al. (2003). Viscous ruptures tend to occur when the crust is still in the plastic stage, and thus, the vesicles would tend to preserve this deformation (Taylor 1934). The subsequent increase in crystallisation and an associated increase in viscosity also have serious implications for the distance that lavas can flow post-rupture. Thus, while viscous rupture can be a suitable mechanism for pāhoehoe-'a'ā transitions on shield volcanoes and high slope stratovolcanoes, it may seldom take place in CFB lavas which are orders of magnitude more voluminous and flow over greater distances on extremely low gradients. In any case, post-brecciation increase in nucleation rates increase the viscosity of flows, making it vital to reconsider some of the long distances theorised for the upper lava flows in the Deccan, including long distances across sub-provinces (e.g. Mandla lobe) and Rajahmundry Traps (e.g. Fendley et al. 2020).

Conclusions

- In the present study, we explore the distribution of lava flows from within the southern Deccan Traps. The results reveal that rubbly pāhoehoe lavas dominate this region, implying that conditions that led these lava flows to transition were the norm during their emplacement.
- We also studied the effects of crystallisation on the morphology of the flow and vice versa within rubbly pāhoehoe lavas. Evidence from vesicle and crystal size distributions and textures shows that F5 was initially emplaced as a pāhoehoe flow and later transitioned to a rubbly pāhoehoe flow, increasing nucleation rates around the order of 10 to 100 times. F3 is a rubbly pāhoehoe flow that shows local ponding. The cooling rates of these flows are much higher than those calculated through the conductive cooling model, signifying that they cooled much faster. Post brecciation increase in nucleation rates for rubbly pāhoehoe lavas accounts for the higher cooling rates observed.
- Higher nucleation rates correspond to a faster increase in viscosity, which tends to retard flowing and accelerate stagnation of lava flows. The dominance of rubbly pāhoehoe lavas observed in the southern Deccan Traps has important implications for how far these flows can travel from their eruption source.
- Modelling of flood basalt provinces should also account for the diverse cooling and

emplacement histories of transitional lavas to correctly estimate their impact (palaeoclimatic, tectonic, ecological, etc.). Understanding the nature of transition within these lavas can have crucial implications for characterising the rate of extrusion of these lavas. Identifying thick ponded lavas within vertical stacks of rubbly pāhoehoe flows in the upper stratigraphic levels of the Deccan Traps is critically important as they demonstrate complex cooling styles, crystallisation histories and emplacement dynamics that are prevalent within CFB transitional lavas.

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Fig. 1 **a** The map of the Deccan Traps (modified after Deshmukh 1988). **b** Shaded relief map depicting locations of surface sections (depicted in circles) and boreholes (depicted in triangles) whose proportion and thickness of morphotypes is discussed in the present study. The location numbers correspond to locations as mentioned below, while the percentage values in parentheses indicate proportion of rubbly pāhoehoe flows (see Table 1). 1—Sinhagad Fort; 2—Katraj Ghat; 3—Diveghat; 4—Bhor-Mandhardevi Ghat; 5—Khambatki Ghat; 6—Ambenali-Mahabaleshwar Ghat; 7—Pasarni Ghat; 8—Kas; 9—Thosegarh; 10—Kumbharli Ghat; 11—Patan Dapharwadi Ghat; 12—Arle- Gudhe Ghat; 13—Tural; 14—Phansavle; 15—Khadi

Fig. 2 Characteristic features of rubbly pāhoehoe flows in the Deccan Traps. **a** Thick core of a rubbly pāhoehoe sheet overlying a red bole horizon. **b** Close up of brecciated crust. **c** Angular clasts with red bole between individual clasts. **d** Red bole forming as a result of devitrification of the glassy margin of the flow. **e** Colonnade and entablature structure of a rubbly pāhoehoe flow core. Such jointing pattern is common in ponded flows

Fig. 3 Internal structure and morphology for flows F3 and F5, along with the samples selected for present study. The graph to the right of each log shows the variation of maximum vesicle length (mm) with depth. Abbreviations: Cr—crust, Co—core, BVZ—basal vesicular zone

Fig. 4 **a** Univariate kernel density estimator to calculate probability density function (PDF) of lava lobe thicknesses in three data sets: Koyna drill core KBH-7 (Mishra et al. 2017), Koyna drill core KBH-1 (Sinha et al. 2017), and surface exposures of lava flows in the Koyna area (Duraiswami et al. 2017). Vertical scale is a percentage of the probability distribution function (PDF), and area under plot is one. Plot gives useful view of whole thickness distribution at a glance with the values for the Flow F3 and F5 illustrated by vertical lines. **b** Range of lava flow core to lava flow total thickness ratios for Icelandic Neogene flood basalt lava flows: porphyritic group (Grænavatn group, Óskarsson et al. 2017), olivine basalt groups (Hólmar and Grjótá group; Óskarsson and Riishuus 2013), aphyric groups (Kumlafell, Hólmatindur and Hjálmadalur; Óskarsson and Riishuus 2014) and Columbia River basalt lava flows (Thordarson and Self 1998; Vye-Brown et al. 2013)

Fig. 5 Variation in modal percentages (after Monteiro et al. 2019), plagioclase number densities (per μm^2) and average plagioclase sizes (μm) for flows F3 and F5. Abbreviations: Gl—glass, opq—opaque minerals, Cpx—clinopyroxene, Pl—plagioclase

Fig. 6 Photomicrographs depicting textural diversity within lava flow F3. **a** TW050 showing a fine grained and glassy texture, along with the presence of vesicles. **b** TW053 exhibiting distinctive quench morphologies for plagioclase crystals in a fine grained and glassy groundmass. The morphologies of plagioclase probably indicate crystallisation under degassing along with undercooling, similar to what was reported by Lipman et al. (1985). **c** TW063 (BXN) showing a glomeroporphyritic phenocryst surrounded by a coarser groundmass of opaque minerals, plagioclase and clinopyroxene. **d** TW084 showing typical diktytaxitic textures with a larger number of crystals as compared to the groundmass TW063, indicating a higher nucleation rate at this depth. The glass content is also higher here, which might be indicative of a faster cooling rate. **e** A large plagioclase phenocryst in TW107. **f** TW288 (BXN) showing zoned plagioclase microphenocryst, indicating crystallisation prior to extrusion of flow. Abbreviations as in Fig. 5

Fig. 7 Photomicrographs depicting textural diversity within lava flow F5. **a** TW501 showing fine grained glassy textures, along with a slightly deformed vesicle in the basaltic rubble clast. **b** TW507 showing a number of more or less aligned vesicles. **c** TW510 showing a stray plagioclase phenocryst having a corroded appearance. **d** TW524 showing slightly deformed vesicles that are larger in size. Also note the fine, glassy groundmass around subhedral plagioclase phenocrysts, in contrast to the coarser groundmass between the large vesicles. **e** TW557 showing coarser groundmass suggesting slower cooling than the previous samples. **f** TW578 showing plagioclase microphenocrysts within coarse grained groundmass. Abbreviations: Ves—vesicle

Fig. 8 Measured crystal size distributions (CSDs) of various samples with depth for **a** lava flow F3, and **b** lava flow F5

Fig. 9 Variation in product of nucleation and growth rate with time for the two flows. Where two values are available for the same depth, the hollow circles represent larger crystal populations (on the right of the kink formed by the regression lines), while the filled circles represent smaller crystal populations (on the left of the kink)

Fig. 10 Plots of slope vs intercept and product of growth rate and time against product of nucleation rate and time for lava flow F3 and F5. The samples have been shaded with depth, with lighter shades indicating deeper samples

Fig. 11 A theoretical model for a conductive cooling of a 40 m thick ponded lava after Wright and Marsh (2016). These equations use cooling data from real lava systems for their calibration, and therefore, account for the release of latent heat of crystallisation during cooling. **a** Temperature contours on a plot for depth versus time indicating the rate of advance of solidus (lower end of the shaded region) and liquidus (upper end of the shaded region). The shaded region indicates the range over which crystallisation occurs, clearly depicting that as depth increases crystallisation occurs over longer duration of time. **b** Variation in maximum and mean temperature change with depth

Fig. 12 Slope versus intercept, and Gt versus Jt plots for samples from the present study (F3, F5) compared with lavas of different morphologies ('a'ā flows from Crisp et al. 1994 split into i) partly insulated comprising of pāhoehoe ponds and channel portions, ii) non-insulated portions consisting of spatter and 'a'ā flow lobes and iii) post-emplacement samples; and pāhoehoe lava lake from Cashman and Marsh 1988) with experimental measurements of Giuliani et al. (2020)

Table 1 Representative measured sections and statistics related to rubbly pāhoehoe lavas from the southern Deccan Traps

	Surface Lavas exposed in Southern Deccan Traps												**Tural-Rajwadi**		**Koyna Drillhole**	
Measured Ghat Sections	**Sinhagad Fort**	**Katraj Ghat**	**Diveghat**	**Bhor-Mandhardevi Ghat**	**Khambatki Ghat**	**Ambenali-Mahabaleshwar Ghat**	**Pasarni Ghat (Wai-Pachghani)**	**Kas**	**Thoseghar (Mandve-Narewadi)**	**Kumbharli Ghat**	**Patan-Dapharwadi Ghat**	**Arle-Gudhe Ghat**	**Surface flows**	**Borehole (TW-01) Total depth: 216 m**	**Phansavle (KBH-05) Total depth: 980 m**	**Khadi (KBH-09) Total depth: 1135 m**
Lat/Long	18° 21' 57.62" N 73° 45' 35.33"	18° 24' 3.14" N, 73° 51' 24.65"	18° 24' 55.04" N, 73° 59' 22.54"	18° 2'43.91" N, 73° 52' 33.84"	17° 0'3.76" N, 73° 46' 37.21" E	17° 55'20.44"N, 73° 37'22.23"E	17° 56' 35.36" N, 73° 50' 48.38"	17° 42' 42.28" N, 73° 48' 54.41"	17° 35' 30.24" N, 73° 50' 53.86"	17° 25' 34.81" N, 73° 38' 58.17"	17° 22' 36.81" N, 73° 54' 3.13"	17° 14' 1.17" N, 73° 59' 30.38"	17° 14' 19.97" N, 73° 33' 33.49"	17° 14' 19.97" N, 73° 33' 33.	17° 8'44.47" N, 73°40'7	17° 1'39.59" N, 73° 46' 30.32"

	E	E	E	E			E	E	E	E	E	E	E	49" E	.37 "E	E
Elevation (m amsl)	618 – 1317	550-1124	550 - 959	599 - 1234	720 – 954	693-1200	762 - 1189	743 - 1264	757 - 1119	340 - 828	646 - 1060	609-1060	13-76	13-(-203)	161-501	144-(-405)
Total basalt Thickness (m)	699	574	409	635	234	507	427	521	362	488	414	451	63	203	662	549
Details of lava flows	25	25	18	22	12	18	17	9	10	12	9	7	4	6	12	13
Thickness Range(m)	4 - 80	6 - 44	2 - 54	2 - 86	6 - 46	8 - 65	9 - 46	21 - 122	8 - 108	12 - 168	11 - 90	39 - 89	2 - 30	6.5 -54	10 - 55	10 - 60
Lava morphotypes	Rb Ph, Aa	Ph, Rb Ph, Aa	Ph, Rb Ph, Aa	Ph, Rb Ph, Aa	Rb Ph, Aa	Ph, RbPh	Ph, Rb Ph	Ph, Rb Ph, Aa	Ph, Rb Ph	Ph, Rb Ph, Aa	Ph, Rb Ph	Ph, Rb Ph, Aa	Ph, Rb Ph	Ph, Rb Ph	Rb Ph, Ph	Ph, Rb Ph
Pāhoehoe flows																
Number	--	5	4	1	--	9	4	1	2	5	1	2	1	4	4	--
Cumu	--	91	119	58	--	271	74	67	58	134	86	128	2	130.4	107.8	--

lative thickness (m)																
Percentage	--	15.8	24.2	9.1	--	53.5	17.3	12.9	16	27.5	20.8	28.4	3.2	64.2	16.3	--
A'a' flows																
Number	5	4	3	1	1	--	--	1	--	1	--	1	--	--	--	3
Cumulative thickness (m)	259	97	107	39	46	--	--	21	--	24	--	65	--	--	--	28
Percentage	37.1	16.9	26.2	6.1	19.7	--	--	4.0	--	4.9	--	14.4	--	--	--	5.1
Rubbly pāhoehoe flows																
Number	20	16	11	20	11	9	13	7	8	6	8	4	3	2	8	10
Cumulative thickness (m	440	386	203	538	188	236	352	433	308	350	328	251.8	61	72.6	263	521

)																
Percentage	62.9	67.3	49.6	84.7	80.3	46.5	82.4	83.1	85.0	71.7	79.2	55.8	96.8	35.8	39.7	94.9

Table 2 Product of nucleation and growth rates with time for lava flows F3 and F5. For samples with two values, the first set of quantities represent smaller crystal sizes (on the left of the kink), while the second set represents larger crystal sizes (on the right of the kink)

Flow No.	**Sample**	**Depth (m)**	**Growth (μm)**		**Nucleation (μm^{-3})**	
			G_1t	**G_2t**	**J_1t**	**J_2t**
F3	TW050	1.6	156.01	26.03	1.64×10^{-8}	4.80×10^{-6}
	TW053	1.92	110.74	31.76	5.08×10^{-8}	2.44×10^{-6}
	TW063	3.09	49.43	--	8.76×10^{-7}	--
	TW084	7.56	49.16	--	1.45×10^{-6}	--
	TW092	8.35	36.85	--	2.49×10^{-6}	--
	TW107	10.45	48.92	--	9.72×10^{-7}	--
	TW139	14.29	63.82	--	5.32×10^{-7}	--
	TW142	14.5	64.23	--	5.25×10^{-7}	--
	TW145	14.84	55.77	--	8.42×10^{-7}	--
	TW159	17.61	63.73	--	6.61×10^{-7}	--
	TW176	20.38	64.68	14.04	6.61×10^{-7}	1.18×10^{-6}
	TW187	23.66	66.58	24.68	3.21×10^{-7}	1.84×10^{-6}
	TW213	26.65	75.64	--	4.50×10^{-7}	--
	TW236	30.71	60.35	--	6.00×10^{-7}	--
	TW247	32.36	60.64	--	5.55×10^{-7}	--
	TW256	36.19	70.02	--	4.08×10^{-7}	--
	TW259	36.69	59.49	--	6.72×10^{-7}	--
	TW263	37.5	72.57	--	4.01×10^{-7}	--
	TW269	38.43	77.64	2.10	4.50×10^{-7}	5.62×10^{-6}
	TW281	40.46	76.86	--	4.21×10^{-7}	--
	TW288	41.28	73.96	--	3.76×10^{-7}	--
	TW304	43.5	71.38	--	4.51×10^{-7}	--
	TW328	48.58	69.11	--	4.82×10^{-7}	--
F5	TW501	1.53	26.19	--	5.55×10^{-6}	--
	TW505	2.59	26.53	--	5.09×10^{-6}	--
	TW507	3.76	25.77	--	1.08×10^{-5}	--
	TW512	4.48	20.90	--	6.91×10^{-6}	--
	TW514	4.87	70.96	27.16	8.31×10^{-8}	3.36×10^{-6}
	TW519	5.49	20.01	--	4.80×10^{-6}	--
	TW524	6.26	21.07	--	7.25×10^{-6}	--

TW532	7.12	29.94	15.80	1.07×10^{-6}	1.45×10^{-5}
TW547	9.28	26.04	--	3.77×10^{-6}	--
TW552	10.15	122.27	20.86	2.08×10^{-8}	8.63×10^{-6}
TW557	11.24	40.51	17.88	6.14×10^{-7}	1.12×10^{-5}
TW562	11.96	62.39	20.67	6.68×10^{-8}	8.54×10^{-6}
TW571	12.9	32.17	18.16	2.22×10^{-6}	1.10×10^{-5}
TW578	14.86	26.21	--	3.60×10^{-8}	3.66×10^{-6}