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Episodicity of seismicity accompanying melt intrusion into the crust

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[1] Microseismicity from a dike intruding the mid-crust of Iceland occurs episodically on fractures orientated parallel to the dike plane. We interpret it as caused by fragments of chilled magma being broken and pushed along the conduit by melt flow. Laboratory experiments on basalt samples under shear strains of $\sim 10^{-4}$ show that the shear strength of a sample cooled through the glass transition initially increases to a maximum at ~0.7 homologous temperature, but subsequently decreases until failure at ~0.58 homologous temperature. We interpret the failure as due to the connection of microcracks. Episodicity in microseismicity on timescales of hours to days can arise from a cycle in which magma in 0.1–0.5 meter thick dikes first cools and becomes stronger, but then weakens along the dike margins with continued cooling against the country rock. Continued pressure of magma from below may then cause failure along dike-parallel fractures. **Citation:** White, R. S., S. A. T. Redfern, and S.-Y. Chien (2012), Episodicity of seismicity accompanying melt intrusion into the crust, *Geophys. Res. Lett.*, *39*, L08306, doi:10.1029/2012GL051392.

1. Introduction

[2] Extrusion of lava from volcanoes often occurs episodically, on time scales ranging from a few hours through weeks to many months. The most violently episodic are usually silicic volcanoes, but basaltic extrusions can also be time varying. The controls on extrusive processes that cause them to be episodic include pressure loss and replenishment in the magma chamber feeding the flows, interaction of the feeder dike with the elastic country rock [*Costa et al.*, 2007, 2009], stick-slip phase transitions in the melt [*Ozerov et al.*, 2003], degassing processes, changes in crystal content [*Castruccio et al.*, 2010], viscosity changes and solidification [*Wylie et al.*, 1999].

[3] In this paper we suggest mechanisms that may induce episodicity in observed seismicity at deeper levels in the mid-lower crust well below the normal brittle-ductile transition. At these depths, seismicity results from high strain rates caused by magma flow. For sufficiently high strain rates, brittle failure may occur even in high temperature melts, especially if they are silicic [*Webb and Dingwell*, 1990; *Webb and Knoche*, 1996; *Goto*, 1999]. *Tuffen et al.* [2008] show that silicic magmas can fracture seismogenically, even at temperatures as high as 900°C, provided the strain rate is sufficiently high. However, the much lower viscosities of basalt mean that seismogenic fracture is much less likely. In this paper we explore mechanisms that may cause seismicity as a result of changes in the shear strength of the magma as it solidifies and cools through the glass transition.

[4] Precise locations of seismicity accompanying intrusion of a dike in the mid-crust of north Iceland beneath Upptyppingar show that activity occurs in bursts lasting typically a few hours, with longer intervening quiescent periods (Figure 1). This episodic behaviour is typical of the seismicity produced by melt movement through deep crustal dikes elsewhere [*Tarasewicz et al.*, 2012; *Shelly and Hill*, 2011]. The Upptyppingar dike strikes 075°, dips at 50°, and is planar (Figure 1b) [*Martens et al.*, 2010]. A striking feature of the seismicity is that the fault planes of individual microearthquakes constrained by moment tensor solutions lie precisely in the same plane as the macroscopic dip of the dike (Figure 2). Both normal and reverse faults occur in close proximity spatially and temporally [*White et al.*, 2011]. It is likely therefore that the micro-seismicity is caused by high strain rates causing failure within frozen magma in the dike itself. It may do so by breaking plugs of cooled or solidified magma within the dike as pressure from below pushes them up [*White et al.*, 2011].

[5] We show new experimental measurements of the glass transition in basalts which demonstrate that as the melt cools and freezes, its strength first increases rapidly, reaching a peak in the glass phase. Then with further cooling the glass recrystallizes and its strength decreases. This provides a mechanical way of creating episodicity in the microseismicity caused by melt flowing through dikes. If for some reason such as a constriction in the dike, the melt flow slows, then it will first freeze against the sides of the dike. This may slow and eventually stop the flow of magma if the entire conduit freezes. Even if the pressure from below builds up, the high strength of the frozen magma may be sufficient to prevent it from breaking. However, if recrystallization and microcrack formation in the plug occurs as it continues to lose heat to the surrounding country rock, its strength will begin to decrease. Eventually, the melt pressure from below may be sufficient to break the plug free and reopen the dike to allow melt to flow freely. In the process of breaking the plug, it may create microseismicity which we can detect.

[6] In the following sections we discuss first the observations of seismicity in the Upptyppingar dike, and then new experimental results on the strength of basalts as they cool through the glass transition. Finally we combine these two observations to deduce the mechanical processes occurring

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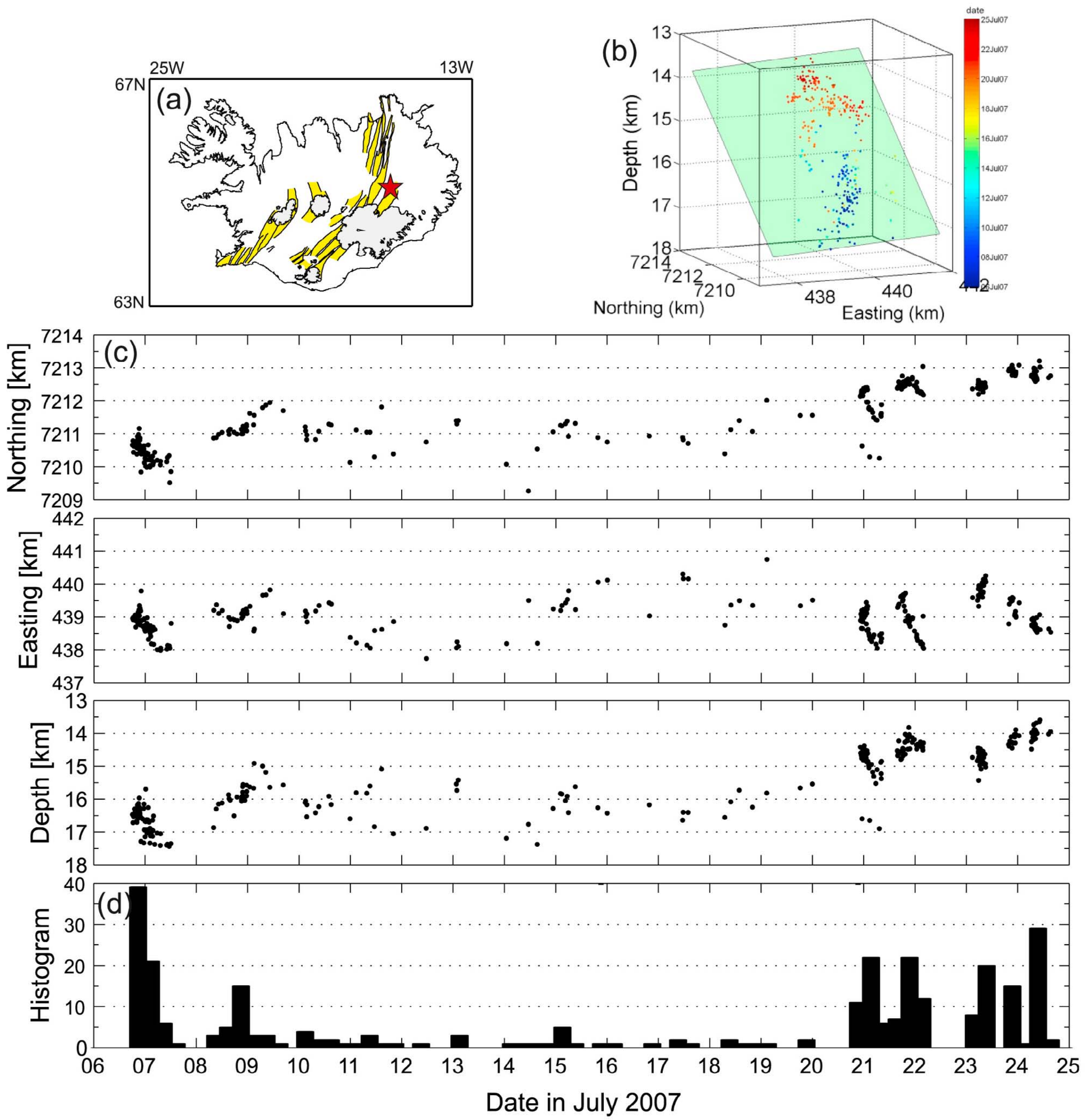

Figure 1. (a) Map showing location of Upptyppingar (red star) within one of the volcanic rift systems (yellow shading) in Iceland. (b) Three-dimensional view from SSW and 10° elevation of hypocenters lying on a plane (green) striking 075° and dipping 50° south. (c) Three panels showing hypocentral locations of microearthquakes during the 6–25 July 2007 period of melt injection at Upptyppingar. Note the migration of hypocenters over distances of ~1 km during periods of a few hours within clusters. Eastings and Northings are in UTM area 28 format and depth is below sea level. (d) Histogram in 4 hour bins showing temporal clustering of seismicity.

to produce episodic microseismicity during melt intrusion through a dike.

2. Microseismicity Produced by Upptyppingar Dike Intrusion

[7] The intrusion lasted one year (March 2007–March 2008), with over 10,000 microearthquakes recorded below Mount Upptyppingar in the Kverkfjöll volcanic system of Iceland. The seismicity was produced by melt injection along a 50° dipping dike which reached a maximum thickness of 1 m [*Jakobsdóttir et al.*, 2008; *Hooper et al.*, 2011]. Starting from a depth of about 18 km below sea level, the deep seismicity migrated both up- and down-dip and also laterally before stopping at about 13 km depth, where the melt apparently froze in situ without eruption. The brittle-ductile boundary in this region is marked by the termination of upper crustal seismicity at 6–7 km depth [*Key et al.*,

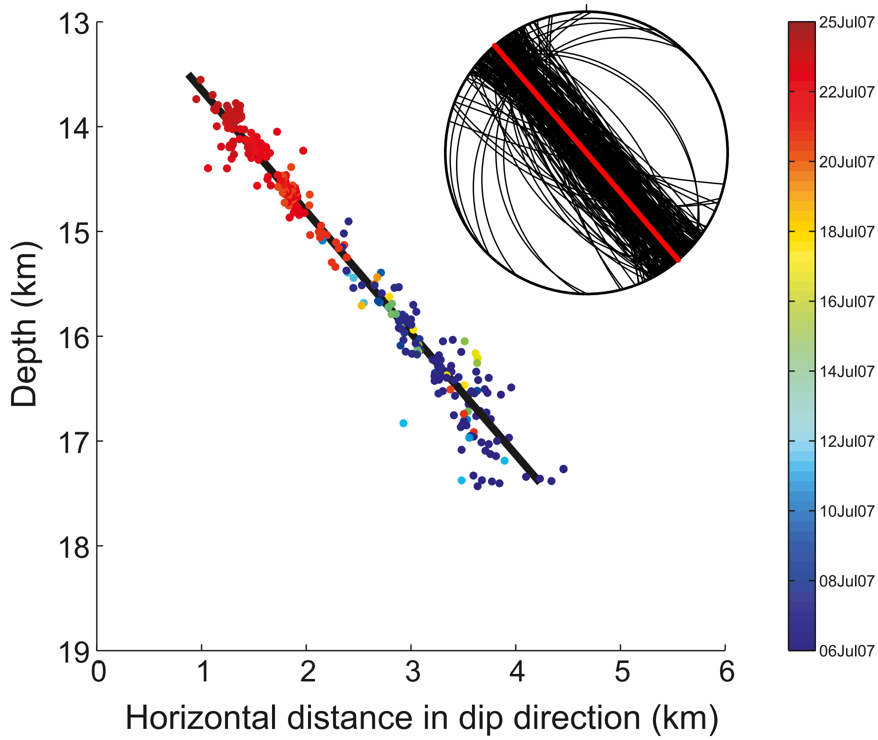

Figure 2. Vertical cross-section of hypocenter locations during the 6–25 July 2007 period of melt injection into a dike beneath Upptyppingar viewed along strike. Solid black line shows best fit plane to hypocentres. Inset shows fault planes from 229 double couple reverse solutions viewed along strike and projected onto an equal area projection orientated vertically and orthogonal to the dike, with red line showing the mean fault plane. Note that the dike plane derived from the hypocentral locations (black line on main diagram) is identical to the mean plane derived from fault plane solutions of individual microearthquakes (red line on inset).

2011]. So the micro-earthquakes at 13–18 km depth caused by the Upptyppingar intrusion are well below the brittle zone, in a normally aseismic region. Nevertheless their appearance is of earthquakes generated by brittle failure, with dominant frequencies around 7 Hz. We postulate that high strain rates produced locally by the magma movement caused the seismicity. The earthquakes typically have local magnitudes of 1.0–1.5. These represent motion of a few tens of millimeters on fault breaks with dimensions of a few meters [*White et al.*, 2011].

[8] An unusual feature of the Upptyppingar seismicity is the sometimes rapid alternation between normal and reverse fault mechanisms, often in the same location within our resolution (c. 60 m) and within minutes of each other [*White et al.*, 2011]. The inferred fault planes of the micro-earthquakes precisely match the macroscopic dip of the dike plane (Figure 2). This suggests that the main faulting mechanism is failure of cooled magma within the dike channel itself, or of country rock immediately adjacent to the dike, since the dip of the dike is constraining the dip of the fractures that generate seismicity. Although some of the seismicity may be caused by faulting at, or adjacent to the propagating dike tip, a significant number of events occur in locations where there has been previous seismicity caused by melt injection. In these locations it is likely that the seismicity accompanies re-opening of an earlier created melt channel. This persistence of seismicity in particular ‘hot spots’ accompanying the flow of basaltic melts has also been observed elsewhere, both at shallower depths on Kilauea [*Klein et al.*, 1987], and at greater depths of 15–30 km within the lower crust and upper mantle in feeder dikes beneath Eyjafjallajökull volcano in Iceland [*Tarasewicz et al.*, 2012] and beneath the Kluchevskoy volcano group in Russia [*Koulakov et al.*, 2011].

3. Observations of Glass Transition in Basalts

[9] Previous studies of the mechanical properties of basaltic lavas have focused principally on the viscoelastic behavior of such material at low strain rates, and at temperatures around and above the glass transition temperature [e.g., *Webb and Dingwell*, 1990; *Lore et al.*, 2000; *James et al.*, 2004]. The mechanical properties of basaltic rocks below the glass transition can be presumed to be a function of cooling history, and of chemical and mineralogical composition. The fracture properties of a series of basalts were investigated by *Balme et al.* [2004] at temperatures below the glass transition and at pressures up to 50 MPa. They proposed a schematic profile of fracture strength that varies according to temperature, dependent on processes including crack blunting and healing on heating to higher temperatures, with a change from brittle to plastic behavior dominating the response. Here we explore further their hypothesis with new measurements of the mechanical response of basalt as a function of temperature on cooling from above the glass transition under dynamic stress.

[10] A sample of hypocrystalline basalt was prepared for mechanical spectroscopy by cutting into a rectangular bar,

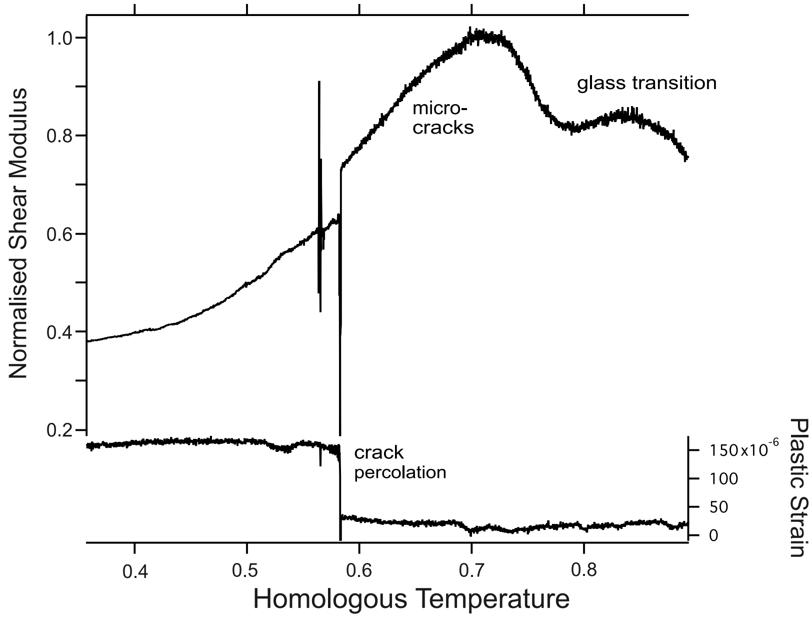

Figure 3. The effective dynamic shear modulus (top curve) and plastic strain (bottom curve) shown by basalt on cooling from a homologous temperature of 0.9. The shear modulus initially increases on cooling from high temperature, as the sample passes through the glass transition. Further cooling below a homologous temperature of 0.7 results in shear weakening, which we attribute to the growth of microcracks and/or stress-induced crystal growth. Percolation of a microcrack network results in a large jump in plastic strain at homologous temperatures around 0.58, with concomitant shear weakening. In this sample the homologous temperatures are based on a melting temperature of 1400 K at 1 atmosphere pressure.

$11 \times 16 \times 2$ mm in size. The specimen is predominantly olivine and pyroxene with around 36% plagioclase and 9% oxides. A few (less than 2%) euhedral olivine and plagioclase phenocrysts lie in small clusters within the homogenous fine grained matrix. Small plagioclase laths are randomly oriented within the groundmass. Clamped within an inverted forced torsion pendulum, the sample was subjected to sinusoidally oscillating shear stress at 1 Hz, so as to generate shear strain of the order of 10^{-4}. The instrument allows the sensitive measurement of variations in the real and imaginary parts of the dynamic shear modulus, expressed as modulus and mechanical loss, tanδ, equivalent to Q^{-1} where Q is the quality factor. The sample was heated to 1250 K and then cooled to 500 K, with the dynamic modulus determined as a function of temperature while cooling at a rate of 1 K/min. All experiments were conducted under high vacuum, around 10^{-3} Pa, and the furnace within the apparatus is manufactured from vanadium which acts as a strong oxygen getter. As a result, the sample shows no oxidative behaviour during the experiments. A typical result is shown in Figure 3.

[11] On cooling from above the glass transition temperature the shear strength of basalt initially increases, shown by the increase in modulus and a transition from a relaxed viscoplastic response to the relaxation-dominated behavior shown by the glassy matrix of the mineral-glass composite. Further cooling, however, results in a weakening of the bulk material due to the onset of microcracking and potentially of stress-induced crystallisation within that matrix.

[12] We have characterised the distribution of energies of such microcracking noise of this basaltic material. It is shown by individual jumps in the energy dissipation, as measured in our experiment by the loss tangent. We have demonstrated that it follows a power law expected for jerky elasticity of material whose mechanical properties are dominated by stochastic behavior [*Chien et al.*, 2010]. This is the temperature region that *Balme et al.* [2004] proposed as weakened due to lower effective crack blunting as the temperature drops. It is also the region in which one anticipates the recrystallization of the glass matrix over a sufficient time scale. Indeed in our experiments we observe the growth of clinopyroxene within the glassy matrix in this temperature interval. It is possible that crystal growth is promoted by the imposed shear stress on the sample, and that jerks in the resultant strain result from rearrangements of the crystalline structure within the glassy matrix.

[13] Eventually, at a homologous temperature of around 0.58 (820 K in this sample), we observe a large jump in the plastic strain of the sample, likely due to the percolation of fractures within the sample across its entire thickness. At the same moment as we see a large increase in plastic strain of the sample we also record a sudden decrease in the measured effective modulus. We have repeatedly cycled the sample, heating up to 1250 K and cooling back down to 500 K. In each temperature cycle the sample recovers its strength on heating and then anneals at the highest temperature by partial melting. Then on cooling the mechanical response is reproduced, with the characteristic weakening seen below a maximum strength which occurs at a homologous temperature of 0.71 (1000 K in this sample) and a sudden large-amplitude jerky behaviour at 0.58 homologous temperature. We interpret this as a triggered local brittle failure of the sample.

4. Application

[14] The flow of basaltic crystal-bearing lava has been treated previously in terms of Bingham fluid models, based

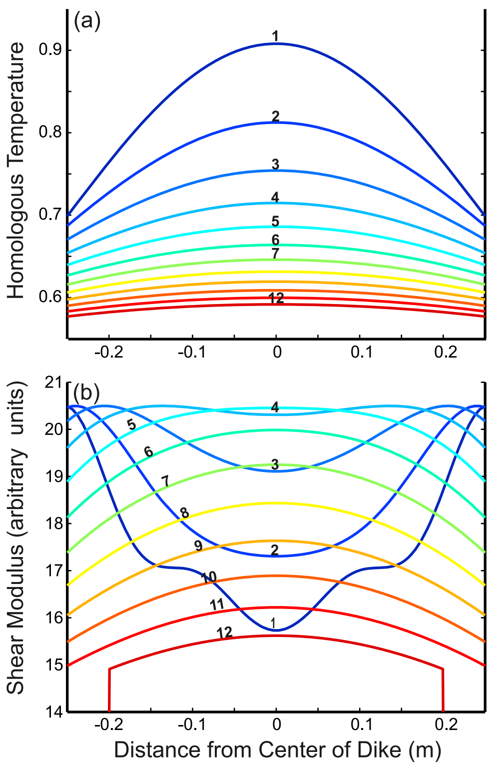

Figure 4. (a) Temperature evolution within a 0.5 m wide dike emplaced in an infinite medium as stationary magma cools from an initial homologous temperature of 1.0 (i.e., at the melting point). Curves are one per day, starting at 1 day. (b) Effective dynamic shear modulus for the temperature distributions using the curve for basalt shown in Figure 3. The magma and country rock have identical thermal properties, with an initial country rock homologous temperature of 0.4.

on the inference that both the absolute viscosity and the strain-rate dependence of the viscosity depend upon the fraction and aspect ratio of crystals within the melt [e.g., *Ishibashi and Sato*, 2010]. Such descriptions provide good explanations of the strain-rate dependence of the properties of these materials above the glass transition. But the mechanical properties below the glass transition, in the region where we find weakening on cooling, are not so clearly amenable to such analysis. We propose that the behavior of the glass-crystal composite state of basalt at lower temperatures is the controlling factor for the cyclic seismicity observed in dikes. In applying our 1 atmosphere experimental observations to the scenario of faults at 15 km depth, the effect of pressure on the glass transition and melt viscosity must be considered, as must the effects of composition of the sample.

[15] The pressure-dependence of these properties in silicate melts is complicated by the fact that melt structure and composition are themselves influenced by pressure [*Lee*, 2011]. For polymerised silica-rich melts the viscosity increases with pressure, and the glass transition temperature decreases, while less-polymerised melts show the opposite behavior. The data of *Del Gaudio and Behrens* [2009] suggest that the glass transition temperature for a pyroxene rich basaltic melt would increase by approximately 30 K on increasing pressure by 500 MPa into the deep crust.

[16] *Ougier-Simonin et al.* [2011] have investigated the influence of pressure on thermal cracking of basalt at relatively low temperatures (up to 570 K) and moderate pressures (up to 50 MPa). They confirm that the role of thermal treatment and temperature is far stronger than any influence of pressure under these conditions. This conclusion is supported by the recent study of *Fortin et al.* [2011] where effective pressures of up to 190 GPa were applied to basaltic samples. Thermal crack networks have a huge impact on the mechanical properties of basalts: *Vinciguerra et al.* [2005] noted a 40% decrease in the P-wave velocity of basalts after thermal stressing to 1170 K, which they attribute to the development of crack networks. *Heap et al.* [2009] demonstrated that repeated stress cycling (for example, induced by pulsing magmatic systems with cyclic cooling and heating combined with cyclic pressure conditions) greatly increased the weakening of basalt through crack propagation.

[17] Brittle failure of the glassy phase on further cooling would be driven by the differential pressure of magma beneath a plug in the conduit. On failure the cooled plug would be reheated and return to its previous state, with further cooling resulting in first a strengthening of the material, and re-formation of the solid glassy plug, and then weakening as it becomes locally brittle and microcrack formation commenced below the glass transition.

[18] In Figure 4 we show a typical example of a 0.5 m thick dike cooling from a subsolidus homologous temperature of 1.0 (i.e., at the melting temperature), after emplacement into cooler country rock initially at a homologous temperature of 0.4. Successive curves are at one day intervals, starting at 1 day. For the first 4 days the interior of the dike is weaker than the margins, which are quenched against the country rock and remain strong because they are close to the peak in strength shown by the curve of shear modulus against temperature (Figure 3). As cooling continues, the temperature of the dike continues to drop because heat conducts into the surrounding rock, but at a decreasing rate (Figure 4a). The core of the dike during this phase remains stronger than the margins because it is warmer, while the shear strength of the margins drops rapidly as they cool down. It is striking that relatively small changes in the temperature of the dike result in large changes in the shear modulus across the dike. Eventually, after 12 days, the rock at the edge of the dike fails as it reaches a homologous temperature of 0.58 if there is an applied stress, such as pressure of melt from below (Figure 4b).

[19] This gives a mechanism for cyclicity in the seismicity in dikes, as well as explaining why the failure planes are parallel to the dike, as we observe in the fault plane solutions (Figure 2 inset). If a drop in the magma flow rate, or perhaps a constriction in the dike allowed the magma to freeze, then its strength would rapidly increase and create a plug of solid rock which could stop the flow and resist further pressure from below. However, continued cooling leads to a weakening in the shear strength, particularly near the walls of the dike where it is coolest, so that failure of the plug may occur if there is persistent pressure from magma deeper in the dike. The plug of frozen melt would likely be cleared from the dike conduit only after frequent small fragments are fractured away from the sides, thus explaining the bursts of seismicity shown in Figure 1.

[20] The duration for the cooling is proportional to the square of the dike thickness. So halving the thickness to

0.25 m would give a timescale of 3 days to cool to failure. Doubling the thickness to 1 m would result in a time to fracture of 1–2 months. In the case of the dike in Iceland from which our seismic data came, modeling of surface deformation suggests that the dike thickness varies from a few centimeters at the edges to about 1 meter in the thickest central region [*Hooper et al.*, 2011], in accord with the range of timescales in episodicity that we observe. It is striking also that we see the clearest 1–2 day cyclicity in the microseismicity near the top and bottom margins of the dike: these are areas where the dike is thinnest and so it is likely that the melt will freeze quickly, whereas in the 1 m thick center of the dike the melt can keep flowing without freezing, and so produces little seismicity.

[21] This modeling is illustrated here only for the simple case of conductive cooling, so that the change in shear strength can be easily visualised. We could also include the effect of release of latent heat during the initial cooling from melt to solid. For a 0.5 m thick dike filled initially with stationary melt, it takes less than a day for the freezing front to extend to the center of the dike. If the melt is flowing, then an equilibrium is reached between the melt temperature and the temperature at the margins of the dike which prevents freezing.

[22] For melt intruded into the mid-crust the temperatures at the margin of the dike remain high during this freezing phase, at slightly above the mean temperature of the melt and country rock. So in general the newly frozen magma remains too strong to fail until it has also cooled conductively as shown in Figure 4. But it is possible that for melt intruded at a very shallow level of the crust, the much lower temperature of the country rock would allow the initial quenched magma to be close to the critical failure criterion of 0.58 homologous temperature. In such a case it is possible that the shear strain exerted by the flowing magma in the still liquid center of the dike could cause shear failure of the solidified rim. This could cause cyclicity in the failure and resultant seismicity, especially if there were small variations in the flow rate that could be magnified by this process. Such a mechanism might, for example, explain the observed repeated seismicity at the same location over a period of several days in a dike flowing at a shallow (3–4 km) depth on Kilauea volcano, Hawaii [*Rubin et al.*, 1998]. Interestingly, the waveforms recorded at a seismometer from this seismicity show complete 180° phase reversals between some events, exactly as observed in the Upptyppingar intrusion [*White et al.*, 2011]. So the same mechanism of failure of fragments of frozen magma along similar planes, but on opposite sides of the dike could also explain the observations from Kilauea.

[23] In some cases the heating of the country rock by intrusion of a dike may cause failure in the country rock rather than within the cooled magma itself, particularly if there is a difference in composition that causes the country rock to be weaker than the cooled magma. Significant heating from an intrusion does not extend far into the country rock, but this might allow country rock failure along planes parallel, and close to the dike if there is a high strain rate caused by melt flowing in the dike. Dike-parallel failures have been reported from kimberlites intruded into dolerites [*Kavanagh and Sparks*, 2011], and could be explained by the mechanism we report here. The dike-parallel failures may also be caused where there are jogs in the alignment of segments of a dike, and again such failure would be facilitated by local heating and weakening of the country rock. In igneous rift zones such as those found in Iceland, the newly intruded dikes are likely to follow the same orientation as earlier dikes, so this pervasive fabric of the country rock may also facilitate dike-parallel faulting in the country rock adjacent to the new dike.

[24] There are of course many variables that could be changed in the simple modeling shown in Figure 4, including the temperature and properties of the country rock. Details of the shear modulus curve shown in Figure 3 would likely change with the composition of the basalt, its pressure and temperature. But we suggest that the essential features of the basalt properties which give rise to the cyclic behaviour, namely an initial increase in shear strength as the magma first cools, followed by weakening and eventual failure as it cools further, are characteristic of intruded basalts.

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