

## Dynamical constraints on kimberlite eruptions

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Eruptions of kimberlite magma involve the rapid ascent of melt from approximately 200 km depth (the base of the lithosphere) (Bailey, 1980; Nixon et al., 1981). Our understanding of the dynamics involved during the eruption is limited by the complex relations observed at the surface in pipe-fill deposits (for example Brown et al., 2008), and the pervasive alteration (Stripp et al., 2006; Mitchell, 1986).

A thermodynamic model of the ascent of kimberlite magma from 200 km depth is described. Temperature-depth profiles are presented for two end-member cases (the ascent of basaltic melt and then carbonatite melt) and also for a kimberlite magma with properties intermediate between these end members. It is assumed that the behaviour of kimberlite magma is encapsulated between these two end-members, and these results are compared with those from the modelling of kimberlite ascent. The effects that the initial temperature of the melt, initial volatile content, and gas solubility law of both pure and mixed gases have on the results are evaluated. For ease of reference, all parameters and notation used in the model are found in Table 1.

### Description of the thermodynamic model

The numerical method that is used to track temperature changes of ascending magmas, as magma exsolves gas and reaches the surface, can be divided into three parts. First the temperature of the melt is calculated, then the temperature of the gas phase (which comprises a temperature balance between any newly exsolved and any preexisting gas), and finally an equilibrium temperature of the magma is calculated. These calculations are repeated for every one hundred metres of magma ascent. Table 2 shows the parameter values used in the model for the three magma compositions. The kimberlite model uses an average thermal expansivity value that is between the basalt and carbonatite values, and a heat capacity value that is similar to that of carbonatite magma.

#### Calculation of the melt temperature

As melt ascends from depth the lithostatic pressure decreases. Up to shallow depths (maybe 5 km depth) the lithostatic pressure is approximately equal to the magma pressure. During its ascent, the melt experiences a pressure decrease and adiabatically expands, resulting in its temperature decreasing:

$$\left(\frac{dT}{dz}\right)_S = \frac{gT\alpha}{C_p}$$

where  $z$  is depth, and  $g$  is the acceleration due to gravity (Jeanloz and Morris, 1986). When applied to melts of different compositions (with different  $\alpha$  and  $C_p$  values), this adiabatic equation calculates the change in melt temperature per kilometre for the specific magma.

	Definition	Units
$C_p$	Heat Capacity	$\text{JK}^{-1}\text{K}^{-1}$
$g$	Gravity	$\text{ms}^{-2}$
$k$	Solubility Coef.	$\text{Pa}^{-0.5}$
$P$	Pressure	Pa
$R$	Universal Gas const.	$\text{JK}^{-1}\text{K}^{-1}$
$S$	Entropy	$\text{JK}^{-1}$
$T$	Temperature	$^{\circ}\text{C}$
$T_m$	Magma Temperature	$^{\circ}\text{C}$
$T_{\text{mix}}$	Equilibrium Temperature	$^{\circ}\text{C}$
$V$	Volume	$\text{m}^3$
$x_g$	Gas weight fraction	wt%
$z$	Depth	m
$\alpha$	Coef. Thermal Expansion	$\text{K}^{-1}$

Table 1. Parameters and notation used in the model.

#### Calculation of the gas temperature

To calculate the temperature of the gas phase of the magmas after each kilometre of ascent, two calculations are made. Firstly, the temperature of any newly exsolved gas is found, and secondly, the new temperature of any pre-existing gas is calculated. An equilibrium calculation is then made to determine the total gas phase. Gases are assumed to obey the ideal gas law.

For every one kilometre the magma ascends, gas of mass  $m_1$  is exsolved according to the solubility law. The gas is released continuously as the pressure decreases, and its temperature is the same as the magma from which it comes. The temperature of this newly exsolved gas  $T_{\text{new}}$  at depth  $z-l$  is calculated from the following equation:

$$T_{\text{new}} = T_{\text{mix}} - \frac{1}{2} \frac{dT}{dz}$$

where  $T_{\text{mix}}$  is the magma temperature at depth  $z$ , and  $dT/dz$  is the adiabatic temperature change calculated.

Any pre-existing gas at depth  $z$  (of mass  $m_2$ ) will ascend to depth  $z-l$ , and cools. As the pressure decreases during ascent, the volume of ideal gases increases adiabatically. The temperature of the pre-existing adiabatically expanded gas  $T_2$  at its new depth  $z-l$  is then calculated:

$$\frac{T_2}{T_1} = \left(\frac{P_2}{P_1}\right)^{\frac{R}{C_p}}$$

where  $T_1$  is the gas temperature at depth  $z$  at volume  $V_1$ , and  $T_2$  is its temperature at depth  $z-1$  where the gas volume has increased to  $V_2$  (Bauman, 1966).

A mass balance equation is used to calculate the temperature of all the exsolved gas at depth  $z-1$ , taking into account the mass and temperature of both the newly exsolved gas ( $T_{new}$ ) and the pre-existing adiabatically expanded gas ( $T_2$ ).

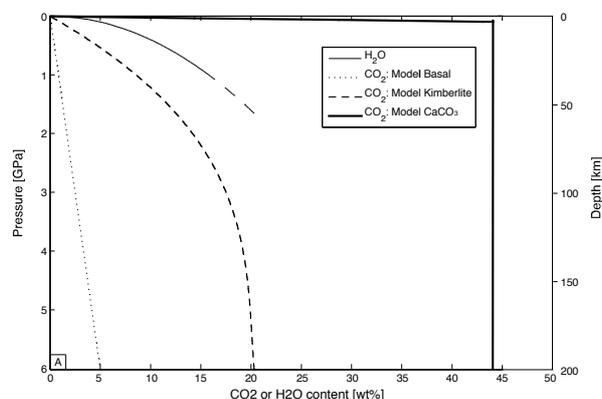


Fig.1 Volatile solubility models of H<sub>2</sub>O and CO<sub>2</sub> in magmas analysed in the thermodynamic model.

#### Calculation of equilibrium temperature

To calculate the temperature of the magma  $T_{mix}$  at the new depth ( $z-1$ ), an equilibrium calculation is made:

$$T_{mix} = \frac{T_g x_g C_{Pg} + T_m (1 - x_g) C_{Pm}}{C_{Pg} x_g + C_{Pm} (1 - x_g)}$$

where  $T_g$  is the gas temperature,  $x_g$  is the weight fraction of exsolved gas,  $C_{Pg}$  is the heat capacity of the gas phase,  $T_m$  is the magma temperature and  $C_{Pm}$  is its heat capacity at fixed. This  $T_{mix}$  calculation is repeated for each one kilometre of magma ascent, from 200 km depth to the surface.

	$\alpha$	$C_p$	$k$	$R$
Basalt	$4 \times 10^{-5}$ *	1522.3 <sup>ψ</sup>	$3 \times 10^{-6}$	
Kimberlite	$11 \times 10^{-5}$	1650	$1 \times 10^{-6}$	
Carbonatite	$14.57 \times 10^{-5}$ <sup>+</sup>	1653.5 <sup>+</sup>	$1 \times 10^{-6}$	
1 GPa				
2 GPa	$18.22 \times 10^{-5}$ <sup>+</sup>	1810.6 <sup>+</sup>	$1 \times 10^{-6}$	
4 GPa	$17.75 \times 10^{-5}$ <sup>+</sup>	1967.9 <sup>+</sup>	$1 \times 10^{-6}$	
6 GPa	$5.89 \times 10^{-5}$ <sup>+</sup>	1967.9 <sup>+</sup>	$1 \times 10^{-6}$	
H <sub>2</sub> O		2333 <sup>ε</sup>		462 <sup>φ</sup>
CO <sub>2</sub>		1250 <sup>ε</sup>		188.97 <sup>φ</sup>

Table 2. Parameter values used in the model for the three tested magma compositions. \*Murase and McBirney (1973), <sup>ψ</sup> Bouhfid et al. (2007), <sup>+</sup> Genge et al. (1995), <sup>ε</sup> Chase (1998), <sup>φ</sup> [1].

## Results

The results from the numerical calculations are presented as temperature-pressure-depth plots for the three magma compositions that have been studied. For clarity, a zoomed section of the final few kilometres of ascent are shown for each composition. The results where the model source temperature is 1350 °C are shown and discussed.

### Basalt Model

The results for the basalt model where a range of volatiles are exsolved are shown in Fig. 2. The adiabatic expansion of the melt phase (0 wt% gas exsolved) accounts for nearly 100 °C cooling of the magma. Any gas that is exsolved during ascent will add to this cooling effect, however the contribution of the gas to the temperature change is negligible until very near the Earth's surface, justifying a posteriori the use of the ideal gas law. The higher the initial volatile content of the basalt is, the lower its temperature at the surface. The exsolution of H<sub>2</sub>O from the basalt has a greater cooling effect than the exsolution of CO<sub>2</sub>. For example, the exsolution of 5 wt% H<sub>2</sub>O results in ~90 °C additional cooling, whereas the exsolution of 5 wt% CO<sub>2</sub> cools the magma by 70 °C. Results from other models where higher source temperatures are used show that increasing the source temperature has no effect on the amount of cooling the basalt undergoes due to gas exsolution, though higher volatile concentrations are then necessary to achieve the same surface temperatures.

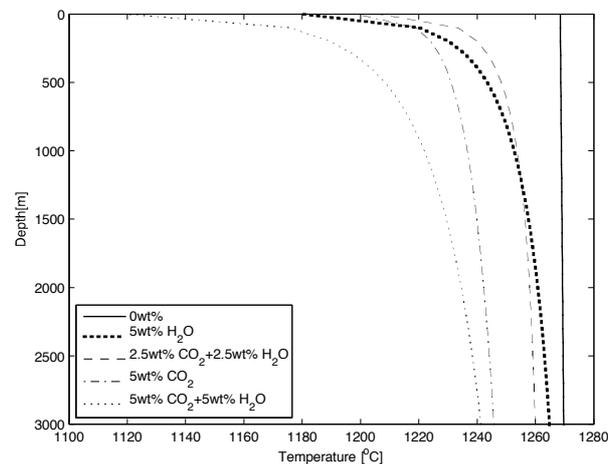


Fig.2 Basalt Model Results. Temperature-depth profile of adiabatically ascending basaltic magma exsolving volatiles.  $T_0 = 1350$  °C.

### Carbonatite Model

The results from the carbonatite model show similar trends to those of the basalt model with the difference that the temperature decrease associated with adiabatic expansion of the melt (0 wt% gas exsolved) is much greater for carbonatite magma, totalling nearly 200 °C. The higher the amount of volatiles that are exsolved from the melt, the lower the surface temperature (see Fig. 3). CO<sub>2</sub> exsolution had a stronger effect on cooling the magma than H<sub>2</sub>O exsolution (at least up to 5 wt%). Increasing the source temperature of the carbonatite model had little effect on the amount of cooling associated with gas exsolution, but a larger amount of gas was needed to produce similar surface temperatures.

### Kimberlite Model

The results of the kimberlite model fall between those of the basalt and carbonatite models. The higher the volatile content of the magma, the larger the amount of

cooling it undergoes during ascent. The adiabatic expansion of the kimberlite melt during its ascent will account for nearly 200 °C of cooling.

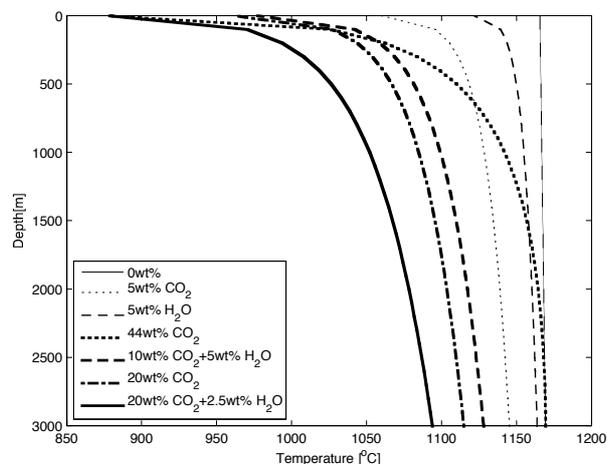


Fig.3 Carbonatite Model Results. Zoomed section: temperature-depth profile of adiabatically ascending carbonatite magma exsolving volatiles.  $T_0 = 1350\text{ }^\circ\text{C}$ .

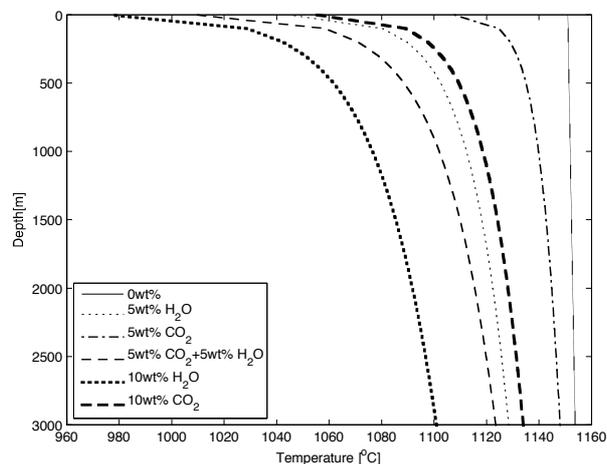


Fig.4 Kimberlite Model Results. Zoomed section of temperature-depth profile of adiabatically ascending kimberlitic magma exsolving volatiles.  $T_0 = 1350\text{ }^\circ\text{C}$ .

When the exsolution of pure  $\text{H}_2\text{O}$  or  $\text{CO}_2$  is considered, for  $<5\text{ wt}\%$  the largest cooling is due to the exsolution of  $\text{H}_2\text{O}$  (Fig. 3). This is also true for when larger volatile contents, such as  $10\text{ wt}\%$ , are exsolved. Additionally, increasing the source temperature of the kimberlite has only a small effect on the amount of cooling associated with adiabatic expansion of the melt, though larger volatile contents are required to reach similar surface temperatures.

## Conclusions

The results so far from the thermodynamic model developed here have some interesting implications for the ascent dynamics of kimberlite magma. One conclusion is that the cooling associated with the adiabatic expansion of the melt phase during ascent is considerable. The model predicts approximately  $200\text{ }^\circ\text{C}$  cooling from this process alone.

The exsolution of gases from kimberlite magma during its ascent can cause substantial cooling as the more volatiles that are released the greater the cooling that occurs. This cooling is largely manifest at shallow depths because of the increasing proportion of exsolved gas.

Current evidence suggests that kimberlite melts have a temperature of  $1050$  and  $1150\text{ }^\circ\text{C}$  in the shallow crust (Canil & Fedortchouk, 1999). Our models can thus place some constraints on the volatile contents and source temperatures of kimberlites. For a source temperature of  $1350\text{ }^\circ\text{C}$  much of the cooling can be explained by decompression of the melt without requiring large proportions of volatiles. Increasing the source temperature allows an increased proportion of volatiles to be exsolved from the melt.

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[1] <http://www.engineeringtoolbox.com>