CONSEQUENCES OF ADIABATIC COOLING WITHIN VOLCANIC CONDUITS ON EARTH AND MARS. K. L. Mitchell¹, L. Wilson¹ and C. J. N. Wilson² ¹Planetary Science Research Group, Environmental Science Department, Lancaster University, Lancaster LA1 4YQ, United Kingdom. (k.l.mitchell@lancaster.ac.uk), ²Institute of Geological and Nuclear Sciences, Wairakei Research Centre, Taupo, New Zealand.

Introduction: We have developed a new model dealing with the physical changes that magma and its associated gases experience when travelling through an eruptive conduit to the surface in explosive volcanic eruptions of silicate materials.

In the past, most such models have assumed isothermal expansion of the magma-gas mixture within the conduit. This paper outlines the new model, and deals with the implications of this new treatment.

Adiabatic expansion in explosive volcanic eruptions: Most numerical models of explosive volcanic eruptions [e.g. 2, 10, 12] considered the expansion of the magmatic gases to be an isothermal process. This assumption has been justified by field evidence which, in general, has shown that erupted magma temperatures are similar to those predicted in the source, typically a magma chamber, by geochemical means. However, such treatments have ignored conservation of energy, which is fundamental to any real system. The increase in kinetic (K.E.) and potential (P.E.) energy has to be balanced by an equal loss from elsewhere, and the only feasible source for this is, directly if not indirectly, the thermal energy contained within the eruptants. Buresti and Casarosa [1] did consider adiabatic expansion, but failed to take into account gas exsolution during magma ascent. Mastin’s model [6] also uses an approximation to an adiabatic treatment and concludes that a typical Kilauean magma of 1200°C and 0.4 wt% water would cool by less than 15°C while decompressing from a few tens of MPa to atmospheric pressure. However, when a wide spectrum of eruptions are considered, the temperature losses cannot be considered insignificant in all cases.

We have used an approximation to the internal energy loss suggested in Wilson, Sparks and Walker [12] to support the isothermal treatment, on the basis that observed erupting magma temperatures are similar to those predicted by geochemical means in the magma chamber. However, we call this into question on the grounds that larger clasts will be protected from the main cooling effects due to the time it takes for heat to propagate through magmatic materials: as a rule of thumb it takes 1 second for the effect of a temperature change to be felt through 1 mm of magma. Therefore observations of magmatic temperatures at the surface will be misleading, as they will not be representative of the average of the bulk products. From this, one can infer that the thermal isolation of the larger clasts means that the finer (typically sub-millimetre) clasts and the gas may be significantly cooler than the averages generated.

This may help explain some anomalies in intraplinian ignimbrite deposits such as those associated with the 1800 B.P. Taupo eruption [9]. Despite being poorly- to non-welded, implying cooling of 200-250 K, atmospheric mixing with this flow cannot have occurred, as depletion of the crystal fraction is not evident. If atmospheric mixing had occurred most of the fine components would have been elutriated out resulting in a greater crystal fraction [8].

Computational model for the eruption of silicate magmas: Our numerical model does not differ greatly from previous treatments. Several assumptions are made in order to solve the series of equations, using a fourth order Runge-Kutta approximation, for the conservation of mass and momentum at successive height increments in the vent:

(1) The mixture of magma and volatiles is homogeneous.

(2) Conduit pressure approximately equals lithostatic pressure, which is given by the following equation, adapted from a self-compacting treatment by Head and Wilson [4], and adjusted by a source driving pressure, if necessary, thus:

\[ P = P_{\text{surf}} + (1/\lambda)\log((V+(1-V))\exp(-\lambda P_{\text{cr}}x)/z_0) + (P_x/z_0) \]

where \( P_{\text{surf}} \) is the atmospheric pressure at the vent, \( \lambda \) is a constant, \( V \) is the surface void space, \( P_{\text{cr}} \) is the density of the fully compacted crust and \( P_x \) is the drive pressure. The driving pressure is added because in many cases on Earth it has been shown to be necessary to introduce a driving pressure in order to overcome the negative buoyancy of some magmas.
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(3) Solubilities of volatiles within the magma are controlled by simple functions of pressure [10], and are independent of the presence of other volatiles.

(4) Bulk viscosity can be treated as equal to magma viscosity below, and gas viscosity above, fragmentation depth (defined as the point where the volume fraction of gas equals 0.75). At present, magma viscosity is considered to be independent of temperature. Temperature dependence will be added at a later date, but we do not anticipate significant departures from our initial results except in cases of extreme cooling, particularly those involving interactions with aquifers.

(5) The erupting mixture decompresses adiabatically to atmospheric pressure. This is the main departure from previous treatments at present.

Mars-Earth comparison and implications for the Martian fines budget: In an attempt to compare eruptions on the Earth and Mars, we have set up simulations of similar basaltic eruptions on both planets. Starting depths ($z_0$) and driving pressures ($P_x$) have been scaled gravitationally such that both are approximately three times greater on Mars than on Earth. Mass flux does not appear to affect eruption temperatures to first order, so we used $1 \times 10^6$ kg/s. We have chosen to include 0.5 wt% CO$_2$ in the source region to ensure the initiation of the eruption. The results are shown below:

Table 1: Mean eruption temperatures of magma in K

<table>
<thead>
<tr>
<th>Planet</th>
<th>H$_2$O wt% contained in magma</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.1</td>
</tr>
<tr>
<td>EARTH</td>
<td>1443</td>
</tr>
<tr>
<td>MARS</td>
<td>1437</td>
</tr>
</tbody>
</table>

Losses approaching 50 K can be seen for the terrestrial case, with greater losses experienced by magmas with higher volatile content. Note that, as mentioned before, the temperature given assumes that none of the magma is thermally isolated. This will not be so if >~ 1 mm clasts are present, and we may see significantly lower gas and fine clast temperatures in such cases. If a significant mass fraction of the eruptive mixture is above these temperatures, expect much cooler gas temperatures at eruption. These temperature figures may also be significantly lower in rhyolites and other more viscous magmas, where the solubilities of gases are higher, thus permitting greater volatile contents.

The case on Mars is more extreme. We see much greater temperature losses due to greater fragmentation depths, which have to be considered significant. The lower gravity and higher exit velocities on Mars means significantly more time between emergence and deposition, suggesting emplacement of pyroclasts at lower temperatures than previously considered. From this it can be inferred that welding and ponding is less likely for an eruption, resulting in finer volcanic products, further supporting the hypothesis that explosive volcanism responsible for the bulk of the Martian fine soil [4]. Also, the significantly lower erupting gas temperatures produced imply much lower plume heights than predicted using isothermal eruption models. This will be investigated further in the future.

Conclusions: Intra-plinian ignimbrite deposits that are poorly- to non-welded may have these properties due to enhanced adiabatic cooling. The grain size and component characteristics of such deposits on the Earth should be re-examined to infer whether adiabatic decompression or mixing with the atmosphere could be responsible for their final emplacement temperature.

Eruption temperatures should be systematically lower on Mars than Earth, making it possible to produce significant quantities of fine-grained, non-welded deposits. Adiabatic cooling may have profound consequences for physical volcanism on Mars.


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