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Estimating the total mass emitted by the eruption of Eyjafjallajökull in 2010 using plume-rise models

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ABSTRACT

The volcanic plume-rise model of Devenish (2013) is applied to the duration of the 39-day eruption of Eyjafjallajökull in 2010 to produce a time series of the estimated source mass flux. This in turn is integrated to give the total emitted mass, which is found to lie within the error bounds of an observational estimate made by Gudmundsson et al. (2012). The calculation uses realistic profiles of key atmospheric variables such as wind speed, temperature, and humidity taken from a numerical weather prediction model and appropriate to the time of the eruption. The sensitivity of the model results to changes in the values of the entrainment coefficients is discussed. It is shown that including the radius of the plume (when it is strongly affected by the wind) in the comparison of the modelled and observed rise heights not only improves the accuracy of the estimated total emitted mass (compared with neglecting the radius) but also reduces the sensitivity of this estimate to the value of the entrainment coefficient associated with velocity differences normal to the plume axis.

1. Introduction

The duration and impact of the eruption of Eyjafjallajökull in 2010 made it one of the best observed eruptions. As a result, estimates of the total mass emitted have been possible using a mixture of ground surveys and remote sensing. In particular, Gudmundsson et al. (2012) estimated the total mass emitted to be $3.84 \pm 0.96 \cdot 10^{11}$ kg (excluding the mass emitted in the form of lava) using a combination of remote sensing and ground surveys, and Stohl et al. (2011) used satellite column loads and an inversion model to estimate the mass contained in the fine particle fraction (2.8–28 μm diameter) that survives into the far field to be $8.3 \pm 4.2 \cdot 10^9$ kg. Taking these two values together gives an estimate of the distal fine ash fraction (i.e., the fraction of mass that survives into the far field) of 2%. There is, however, a lot of uncertainty regarding the distal fine ash fraction with estimates ranging from 0.1% to 10% (e.g., Rose et al., 2000; Dacre et al., 2011; Devenish et al., 2012a, 2012b; Webster et al., 2012); this dominates any error in the estimate of the total mass from the fine particle fraction.

The volcanic plume-rise model of Devenish (2013), which includes the effects of ambient wind and moisture, was applied iteratively to a short period of the Eyjafjallajökull eruption in mid-May 2010 using realistic atmospheric profiles appropriate to the time of the eruption in order to estimate the source mass flux for a given rise height. It was shown that accounting for the prevailing meteorology can lead to significant differences in estimates of the source mass flux compared with empirical relationships between the rise height of the eruption column and the source mass flux (Mastin et al., 2009; Sparks et al., 1997, §5.2). For example, if the volcanic plume is strongly bent over by the ambient wind, then using one of the empirical relationships quoted above is likely to lead to an underestimate of the source mass flux. Conversely, moisture can add significantly to the energy of a volcanic plume via latent heating and so can potentially lead to an overestimate of the source mass flux. Furthermore, since the stability of the troposphere is less than that of the stratosphere, an empirical relationship of the form proposed by Mastin et al. (2009) or Sparks et al. (1997) is also likely to overestimate the source mass flux (all else being equal). Similar results were also obtained for the eruption of Eyjafjallajökull in 2010 by, e.g., Woodhouse et al. (2013); Mastin (2014).

This paper is organized as follows: in the next section, the mass flux is calculated using the plume-rise model of Devenish (2013) for the duration of the 39-day eruption of Eyjafjallajökull in 2010. This allows the total mass emitted to be calculated which is then compared with the observational estimate. The calculation uses realistic profiles of key atmospheric variables such as wind speed, temperature, and humidity taken from a numerical weather prediction model and appropriate to the time of the eruption. The sensitivity of the model results to changes in the entrainment coefficients is considered in §3.

2. Calculation of mass emitted by Eyjafjallajökull in 2010

An initial estimate of the source mass flux is calculated from the empirical relationship between the observed rise height (above the

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volcano summit), z_{obs}, and the source mass flux, Q_m, proposed by Mastin et al. (2009):

$$Q_m = 141 z_{obs}^{4.15} \quad (1)$$

where z_{obs} is measured in km and Q_m in kg s $^{-1}$. The model of Devenish (2013) is applied iteratively to determine a revised source mass flux for a given rise height that accounts for the prevailing atmospheric conditions. The observed rise height appropriate to the time of interest is kept fixed, and a bisection method is used to refine the value of Q_m. It should be noted that the source mass flux estimated from Eq. (2) or from the model of Devenish (2013) is for the total emitted mass; the model of Devenish (2013) makes no reference to the size of the ash particles nor does it allow for fall out of ash from the eruption column.

The model of Devenish (2013) has been modified so that here the observed height of the eruption column is matched with

$$Z_{top} = \min(Z_{max}|_{U=0}, Z_{max} + b_{max}) \quad (2)$$

where z_{max} is the height of the eruption column, determined by the model, at which plume rise terminates (i.e., the vertical velocity, w, first becomes zero), $z_{max}|_{U=0}$ is the value of z_{max} in the absence of a crossflow, and b_{max} is the plume radius (as defined by, e.g., Devenish et al., 2010a, 2010b) at the maximum rise height. Previously, z_{obs} was matched directly with z_{max} (Devenish, 2013), but this can lead to an overestimation of the source mass flux for a strongly bent-over plume since; as explained by Mastin (2014), the observed height of a strongly bent-over plume corresponds to $z_{max} + b_{max}$ (if measured correctly). However, since $b_{max} \to \infty$ as $U \to 0$, $z_{max} + b_{max}$ can exceed $z_{max}|_{U=0}$ in a weak wind, and hence it is necessary to provide a limiting value for z_{top} in the form of $z_{max}|_{U=0}$. The transition between $z_{max}|_{U=0}$ and $z_{max} + b_{max}$ depends on the relative magnitudes of the ambient wind speed, U, and the vertical-velocity scale of the plume which form a convenient dimensionless wind speed

$$\tilde{U} = \frac{U}{(F_0 N)^{1/4}}$$

where F_0 is the effective buoyancy flux at the source and N is a constant buoyancy frequency. (We note that since F_0 is negative for a volcanic plume, it is more usefully taken to be an effective buoyancy flux assuming all the heat is transferred to the gas phase to ensure a positive buoyancy flux. In a realistic atmosphere, neither N nor U is constant, but reasonable estimates can be obtained from respectively a least-squares fit to the potential temperature profile over the depth of the plume above the volcano summit and the average wind speed over the same depth. Both F_0 and N assume a constant reference temperature of 273 K.) Eq. (2) is illustrated for a simple plume model, of the form used by Devenish et al. (2010a), in Fig. 1 for a uniform crosswind with speed U. It can be seen that the transition from $z_{max}|_{U=0}$ to $z_{max} + b_{max}$ occurs for $\tilde{U} = O(1)$. For $\tilde{U} \gg 1$, the maximum rise height for a bent-over plume in a uniform wind is given by (e.g., Briggs, 1984; Devenish et al., 2010a)

$$Z_{max} = \left(\frac{6}{\pi \beta^2} \frac{F_0}{N^2 U} \right)^{1/3} \quad (3)$$

where β is the entrainment coefficient associated with strongly bent-over plumes. When Eq. (3) is scaled by $\alpha^{-1/2} F_0^{1/4} N^{-3/4}$, the scale height for a vertically rising plume in which α is an entrainment coefficient associated with vertically rising plumes, $z_{max} \propto \tilde{U}^{-1/3}$. The model used to produce Fig. 1 employs $\alpha = 0.1$ and $\beta = 0.5$. It is found that in the calculations with a volcanic plume model in a realistic atmosphere, using Eq. (2) instead of z_{max} alone (as was used by Devenish, 2013) can lead to reductions in the revised source mass flux of approximately a factor of three.

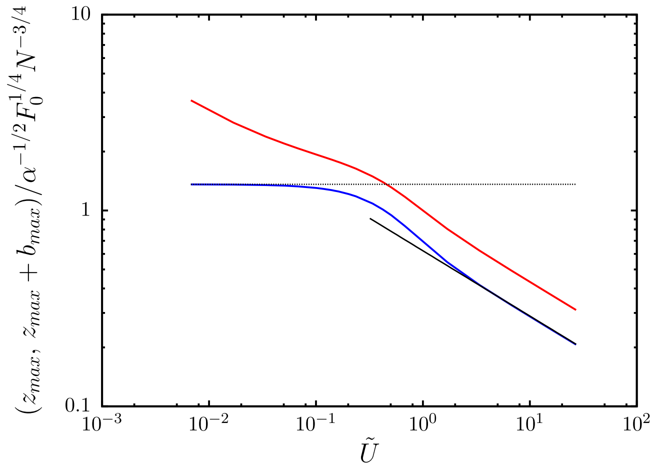

Fig. 1. The variation of z_{max} (blue) and $z_{max} + b_{max}$ (red) with $\tilde{U}$ computed from a simple plume model with a uniform wind speed (i.e., constant with height) and constant F_0 and N. The dotted horizontal black line is $z_{max}/\alpha^{-1/2} F_0^{1/4} N^{-3/4} \approx 1.36$ as given by Morton et al. (1956) for a vertically rising plume ($\tilde{U} = 0$). The solid black line is $\alpha^{-1/2} (6/\pi\beta^2)^{1/3} \tilde{U}^{-1/3}$.

The model of Devenish (2013) has been coupled to the Met Office's operational dispersion model NAME (Numerical Atmospheric-Dispersion Modelling Environment; version 6.5, see, e.g., Jones et al., 2007) to facilitate frequent updates of the prevailing meteorology, which are provided by the Met Office's numerical weather prediction model, the unified model (UM). The global configuration of the UM is used which at the time of the eruption had a horizontal grid spacing of about 25 km in the mid-latitudes and 70 unequally spaced vertical levels extending into the mesosphere with a typical resolution of 300–400 m in the mid-troposphere. The ambient meteorology is updated at three hourly intervals and interpolation in time and space gives appropriate profiles for Eyjafjallajökull at the time of interest. In this study, the eruption was considered to have started at 0900 h on 14th April 2010 and to have finished at 1800 h on 23rd May. The rise height of the eruption column was determined by the Icelandic Meteorological Office using a combination of methods of which radar was the most important (Arason et al., 2011); a piece-wise constant fit of their data is used here (see Fig. 3 of Webster et al., 2012).

Using the same parameter values as used by Devenish (2013) (i.e., an initial temperature of 1273 K, exit velocity of 100 ms $^{-1}$, initial gas fraction of 3%, $\alpha = 0.1$, $\beta = 0.5$) and neglecting any source moisture, the source mass flux was calculated at three hourly intervals. It is shown in Fig. 2 along with the empirical formula of Mastin et al. (2009). Distinct periods of relatively vigorous volcanic activity can be identified of which the most important are the initial period, during which the largest values of Q_m occured, and further periods in early and mid-May. As may be expected, this is consistent with the analysis of Arason et al. (2011) and Gudmundsson et al. (2012), who divide the duration of the eruption into two distinct phases, 14–18 April and 5–17 May. Fig. 3 shows that Q_m estimated using a plume-rise model is usually larger – sometimes more than ten times as large – than that calculated from Mastin's empirical formula using the rise height alone but can also be smaller. Of course, greater variability in Q_m(original) could also have been achieved using Eq. (1) and a more frequently varying time series of the rise height. While it is difficult to make an *a priori* assessment of the magnitude of this additional variability, it is worth noting that the time series proposed by Webster et al. (2012) and used here averages out (in some sense) many of the small fluctuations in the rise height over longer time periods than the six hourly averaged time series in Fig. 7 of Arason et al. (2011). Thus, in some cases, from 18 to 25 April, for example, where the time series of Webster et al. (2012) shows a generally higher rise height than the time series of Arason et al. (2011), one might have expected a lower mass emission rate on average