
Volume 301, Issues 1–2
3 January 2011
ISSN 0012-821X
ELSEVIER
EARTH
& PLANETARY
SCIENCE LETTERS
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Earth and Planetary Science Letters 301 (2011) 171–178

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

journal homepage: www.elsevier.com/locate/epsl

# Rise of volcanic plumes to the stratosphere aided by penetrative convection above large lava flows

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ARTICLE INFO

*Article history:*
Received 26 May 2010
Received in revised form 29 October 2010
Accepted 29 October 2010
Available online 24 November 2010

Editor: Y. Ricard

*Keywords:*
basaltic eruptions
volcanic plumes
penetrative convection
continental flood basalts
global environment

ABSTRACT

Turbulent volcanic plumes disperse fine ash particles and toxic gases in the atmosphere and can lead to significant temperature drops in the atmosphere. In the geological past, the emplacement of large continental flood basalts (CFB) has been associated with large changes in the global environment and extinctions of biological species. The variable intensity of environmental changes induced by otherwise similar CFB events, however, begs for a reevaluation of physical controls on the environmental impact of volcanic eruptions. The climatic impact of an eruption depends on its ability to inject gases in the stratosphere and on the eruption rate. Using integral models of turbulent plumes above line and point sources, we find that mass rate estimates for CFBs are in general not large enough for volcanic plumes to reach the stratosphere on their own. Basaltic eruptions, however, are also associated with widespread lava flows which lose large amounts of heat and generate convection in the atmosphere. This form of convection, known as penetrative convection, acts to erode the stably stratified lower atmosphere and generates a thick well-mixed heated atmospheric layer in a few hours. The added buoyancy provided by such a layer almost always ensures that volcanic gases get transported to the stratosphere. The environmental consequences of CFBs are therefore controlled not by the inputs to the atmosphere from individual volcanic plumes, but by the dynamic response of the climate system to a succession of short eruptive pulses within a longer-lasting eruption sequence.



## 1. Introduction

In the Earth's atmosphere, buoyant turbulent plumes are often man-made by-products of industrial processes or catastrophes such as oil fires and nuclear explosions. They are also due to natural phenomena such as forest fires and volcanic eruptions. These plumes consist mainly of vertical turbulent upwellings carrying a number of gaseous or solid components, such as sulfuric gas and ash particles. Material injected in the atmosphere is likely to affect the environment on a global scale if the amount involved is large enough and if it is released in the stratosphere.

Explosive volcanic eruptions are associated with well-documented impacts on the local environment, such as acid rains and thick ash deposits, and the most powerful ones may induce global (or at least zonal and hemispheric) climate changes. For example, the gigantic eruption of Mount Pinatubo (Philippines) in 1991 released more than 25 megatons of aerosols in the northern hemisphere. This led to an increase of the average summer temperature of about 1 °C and to a decrease of the average winter temperature of about 0.5 °C (Kirchner et al., 1999).

Basaltic eruptions are mainly effusive and, as a consequence, have long been considered to be ineffective in perturbing the climate. Yet, these eruptions typically release larger amounts of aerosols than explosive eruptions of more silicic magmas, with potentially more dramatic effects on the global environment. The 1783 Laki eruption (Iceland), which emitted about ten times more $SO_2$ than the 1991 Pinatubo eruption (Thordarson and Self, 2003), led to the reevaluation of basaltic fissure eruptions as agents of climate change. This eruption was followed by anomalously high summer temperatures in Europe and a very cold winter, as well as dramatic air pollution over much of the northern hemisphere (Chenet et al., 2005; Grattan and Charman, 1994; Grattan and Pyatt, 1999; Thordarson and Self, 2003).

Large continental flood basalts (CFB) are paroxysmal basaltic eruptions that emit much larger quantities of volatiles than the Laki one, in the range of thousands of gigatons (e.g. Self et al., 2006), and hence are likely to induce major changes of the global environment. Indeed, the ages of major life extinction episodes and large continental flood basalts (CFB) are strongly correlated with one another over the last 300 Ma, which supports a causal relation between the two types of events (e.g. Courtillot, 1994; Courtillot and Renne, 2003; Keller, 2005; Rampino and Stothers, 1988; Wignall, 2001). Three significant instances are the Cretaceous–Tertiary crisis (≈65.5 Ma) which appears to be in large part due to the Deccan traps

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doi:10.1016/j.epsl.2010.10.037

(Chenet et al., 2007, 2009; Courtillot et al., 1986; Keller, 2003; Keller, 2005; Self et al., 2008a), the Permo-Triassic mass extinction (≈250 Ma), which is synchronous with the Siberian traps (Heydari et al., 2008; Reichow et al., 2009; Xi et al., 2007), and the Triassic–Jurassic biotic event (· 200 Ma), which is associated with the emplacement of the Central Atlantic Magmatic Province (Deenen et al., 2010; Hesselbo et al., 2002; Nomade et al., 2007; Verati et al., 2007). Over the past 500 Ma, one other mass extinction event, at the Guadalupian–Tatarian boundary, is unambiguously related to a CFB (e.g. Courtillot and Renne, 2003; Wignall, 2001; Xi et al., 2007). Most of the other CFBs are associated with global environmental changes of smaller biological impact such as minor extinctions or oceanic anoxia events (see Table 1). For example, the ≈30 Ma Yemen and Ethiopian traps and the ≈183 Ma Karoo traps are associated with climate cooling at the time of the $Oi_2$ (oxygen isotope) event (e.g. Hofmann et al., 1997; Rochette et al., 1998), and with the extinction of 5% of the worldwide biota at the end of the Pliensbachian (Jourdan et al., 2007; Moulin et al., 2009), respectively.

The environmental impact of volcanic activity depends on many different factors: whether eruptions occur subaerially or not, at which latitude they occur, the amounts of $SO_2$ and $CO_2$ they release, as well as the cumulative effects of eruptions that follow one another. Because of the chemical composition of their lavas, their large magnitudes and long durations, CFB eruptions have many of the characteristics that are required for a global influence on the environment. One major requirement, however, is the altitude at which volcanic gases must be injected for maximum impact, which depends on the dynamics of the eruption process and in particular on the mass rate of individual pulses (Robock, 2000; Stothers et al., 1986; Textor et al., 2004).

The aim of this article is to explore the conditions required for a volcanic plume to reach the stratosphere. We show that an isolated volcanic plume rising above a long eruptive fissure is likely to stall below the tropopause, unless unrealistically large values of the mass rate are achieved. We evaluate the effect of *penetrative convection* that is driven by the cooling of large basaltic lava flows. We show that this mechanism generates a thick heated atmospheric boundary layer which provides additional buoyancy that allows a volcanic plume to rise to the stratosphere.

## 2. Rise of turbulent volcanic plumes in a standard atmosphere

Volcanic eruptions eject a mixture of hot materials of different kinds (lava, pyroclasts, ash, and gas). The mixture is initially denser than its surroundings and later becomes buoyant by injection and heating of ambient air. Because the plume rises through a density stratified atmosphere, it eventually encounters ambient air with the same density at some altitude $H$ above the source. At that altitude, the mixture of ash and gas that is carried by the plume ceases to be buoyant and spreads laterally as an umbrella cloud.

The strength of a turbulent volcanic plume depends on the thermal output of the eruption (e.g. Carazzo et al., 2008; Settle, 1978; Suzuki and Koyaguchi, 2009; Woods, 1988, 1993a), and is characterized by the rate of injection of buoyancy $F_0$ defined as follows:

$$F_0 = \Phi g', \tag{1}$$

where $\Phi$ is the volumetric flow rate and $g' = g(\rho_a - \rho)/\rho_r$ is the reduced gravity, with $g$ the acceleration of gravity, $\rho$ and $\rho_a$ the plume and atmospheric densities, respectively, and $\rho_r$ the reference density of the atmosphere. According to the simple and robust theory of turbulent plumes due to Morton et al. (1956), the behavior of a plume is determined by $F_0$, the rate of entrainment of air into the plume, $U_e$, and the density gradient in the atmosphere, or equivalently the atmospheric stratification parameter $S = -g/\rho_r d\rho_a/dz$. The entrainment rate $U_e$, can be considered as the average horizontal velocity at the edge of the plume. It is such that $U_e = \alpha_e W$, where $W$ is the average plume vertical velocity and $\alpha_e$ is a coefficient called the entrainment constant. Within this framework, scaling arguments and laboratory experiments show that $H$ can be written as,

$$H = 2.5(2\alpha_e)^{-\frac{1}{2}} F_0^{\frac{1}{4}} S^{-\frac{3}{8}}. \tag{2}$$

In the following, we review briefly the influence and variability of the three variables involved in this equation.

### 2.1. Source conditions

The most important parameter of Eq. (2) is the emission rate of buoyancy of the plume (i.e. buoyancy flux) $F_0$, which may not be equal to the total value for the eruption $F_{tot}$. Explosive volcanic eruptions emit a mixture of finely fragmented pyroclasts and volcanic gases, that efficiently transfer their thermal energy to ambient air. This efficient thermal coupling ensures that the emission rate of buoyancy of the turbulent plume is equal to the total emission rate, i.e. $F_0 = F_{tot}$. In many basaltic eruptions, however, only part of the magma is fragmented and feeds a plume. Thus, the total emission rate of buoyancy is partitioned between an effusive component associated with lava flows and a plume component with proportions that depend on the degree of magma fragmentation. In basaltic eruptions, the

**Table 1**
Eruptive conditions for Laki and continental flood basalts. P.Lat. = paleo-latitude, $H_{trop}$ = altitude of tropopause. The eruptive mass rates are average values given by the ratio of the volume of erupted magma over the duration of the eruptive episode. The altitude of the tropopause as a function of latitude is from Holton et al. (1995). References: Laki: Thordarson and Self (2003); Guilbaud et al. (2005), Columbia River basalt, Grande Ronde basalt and Roza flow field: Thordarson and Self (1998); Reidel and Tolan (1992); Reidel (1998), Ethiopia, Yemen: Rochette et al. (1998); Courtillot and Renne (2003), Deccan: Keller (2005); Chenet et al. (2008), Karoo: Kosterov and Perrin (1996); Jourdan et al. (2007); Moulin et al. (2009), Central Atlantic: Knight et al. (2004); Whiteside et al. (2007), and Siberia: Courtillot and Renne (2003); Pavlov et al. (2007); Reichow et al. (2009).

| Eruption | Age/date | P.Lat. | $H_{trop}$ (km) | Volume ($km^3$) | Duration (yr) | Eruptive mass rates ($kg\,s^{-1}$) | Environmental effects |
|---|---|---|---|---|---|---|---|
| Laki (Iceland) | AD 1783 | 60N | 9 | 15 | 0.66 | $2.09 \cdot 10^6$ | Perturbations in N. Hemisphere |
| Columbia | ≈15 Ma | 45N | 11 | $1.74 \cdot 10^5$ | $2.5 \cdot 10^6$ | $6.4 \cdot 10^3$ | Large cooling, end of Early Miocene |
| Grande Ronde | id. | id. | id. | $1.48 \cdot 10^5$ | $10^5$ to $10^6$ | $1.35 \cdot 10^6$ to $1.35 \cdot 10^5$ | id. |
| Roza flow | id. | id. | id. | >1300 | 14 | $8.5 \cdot 10^6$ | id. |
| Ethiopia | ≈30 Ma | 10N | 16 | $7.2 \cdot 10^5$ | $8 \cdot 10^5$ | $8.2 \cdot 10^4$ | $Oi_2$ event? |
| Yemen | id. | id. | id. | $4.8 \cdot 10^5$ | $1 \cdot 10^5$ | $4.4 \cdot 10^5$ | id. |
| Deccan (India) | ≈65 Ma | 27S | 15 | $10^6$ to $2.7 \cdot 10^6$ | $5 \cdot 10^5$ | $1.7 \cdot 10^5$ | Cretaceous/Triassic mass extinction |
| Individual pulses | id. | id. | id. | 800 to $1.6 \cdot 10^4$ | 10 to 100 | $7 \cdot 10^4$ to $1.5\ 10^8$ | id. |
| Karoo | ≈183 Ma | 45S | 11 | $2.5 \cdot 10^6$ | $8 \cdot 10^5$ | $2.9 \cdot 10^5$ | Biotic crisis, end of Pliensbachian |
| Central Atlantic | ≈200 Ma | 8S | 14 | $2.5 \cdot 10^6$ | $6.1 \cdot 10^5$ | $3 \cdot 10^5$ | Triassic/Jurassic mass extinction |
| Siberia | ≈250 Ma | 61N | 9 | $3 \cdot 10^6$ | $9 \cdot 10^5$ | $3 \cdot 10^5$ | Permo-Triassic mass extinction |

fraction of fine ash particles typically ranges between 1 and 17 wt.% of the output, and is generally smaller than 10 wt.% (Stothers et al., 1986; Thordarson and Self, 1993), so that $F_0 \leq 0.17 F_{tot}$. According to Eq. (2), a plume generated by a basaltic eruption, with $F_0 \approx 0.1 F_{tot}$, will require twice the mass rate of a Plinian plume ($F_0 \approx F_{tot}$) to reach the same altitude in the atmosphere.

The plume height given in Eq. (2) is only relevant for a plume that is fed from a point source, such that the plume is approximately circular in horizontal cross-section. In a volcanological context, this is valid if the volcanic vent is sufficiently small. In practice, this requires that the vent dimensions are smaller than a tenth of the height of rise, i.e. typically less than a kilometer. This is not appropriate for a fissure-fed eruption, a common feature of basaltic eruptions. In this case, the plume is elongated in one direction. The geometrical configuration of the source affects the efficiency of entrainment through a change of the perimeter to surface ratio of the plume, and hence yields a different scaling for the plume height. Stothers et al. (1986) have shown that the height reached by a 2-D plume scales as the third power of the eruptive mass rate per unit length. From their results and the model of Woods (1993b), one can compare the heights of plumes fed from point sources, $H_{ps}$, and line sources, $H_{line}$:

$$\frac{H_{line}}{H_{ps}} = \beta \left( \frac{Q^{1/4}}{L} \right)^{1/3}, \tag{3}$$

where $Q$ is the eruptive mass rate in kg s$^{-1}$, $L$ is the length of the eruptive fissure in m, and $\beta \approx 2.61$ kg$^{-1/12}$ s$^{1/12}$ m$^{1/3}$. This relationship is illustrated in Figure 1 for fissure lengths that vary within a 1–100 km range. The figure shows that the height reached by line plumes is smaller than the height predicted for a point source, because of the enhanced entrainment.

### 2.2. Efficiency of turbulent entrainment

One control on plume behaviour that has received less attention is the entrainment coefficient $\alpha_e$. Volcanic plumes involve material that is initially denser than the ambient atmosphere and are characterized by large changes of buoyancy during ascent. In such conditions, the simple entrainment model of (Morton et al., 1956) must be reevaluated. Recent analysis, laboratory experiments and fully 3-D numerical calculations demonstrate that the entrainment process is not captured accurately by a constant coefficient $\alpha_e$ (Carazzo et al., 2006; Kaminski et al., 2005; Suzuki and Koyaguchi, 2010). According to these recent studies, entrainment is enhanced in positively buoyant

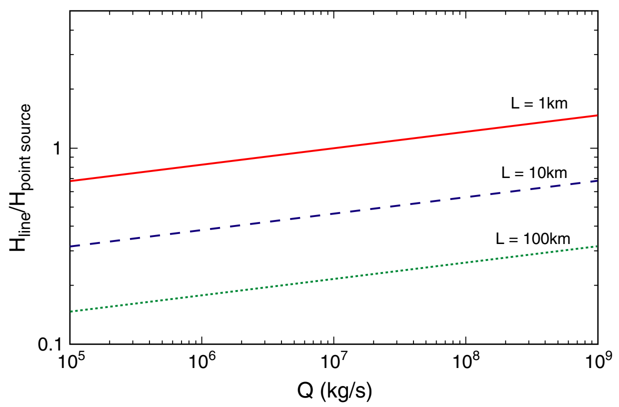


**Fig. 1.** Height reached by a plume issuing from a line source of length $L$ scaled to that for a plume issuing from a point source as a function of eruptive mass rate. For fissure lengths larger than 10 km, line plumes extend to much smaller altitudes than point source ones for the same total mass rate. Thus, for climatic impacts, long eruptive fissures require larger eruption rates than focussed vents.

plumes, which acts to impede plume ascent and hence to decrease the plume height with respect to the prediction of Eq. (2). With more efficient entrainment, plume buoyancy decreases more rapidly during ascent. According to the calculations of Carazzo et al. (2008), this acts to reduce the plume height by about 25%.

### 2.3. Atmospheric stratification

The last parameter in Eq. (2) is $S$, the stratification parameter of the atmosphere,

$$S = N^2 = g\alpha\gamma \tag{4}$$

where $N \approx 0.035$ s$^{-1}$ is the Brunt–Väisälä frequency of the stably stratified atmosphere, $\alpha$ the coefficient of thermal expansion of air and $\gamma$ the stable vertical temperature gradient. On Earth, $S$ varies with latitude with straightforward consequences for the maximum altitude of plumes. All else being equal, plume heights at polar latitudes are smaller by about 20% than those at intermediate and tropical latitudes (Carazzo et al., 2008; Sparks, 1986). The altitude of the tropopause also decreases with latitude, however (from a maximum of 18 km at the equator in the summer to a minimum of 8 km at the poles in winter), so that the ability of a plume to reach the stratosphere is in fact not sensitive to latitude.

An additional parameter related to the structure of the atmosphere is its moisture level. In a moist atmosphere, the condensation of water vapor into liquid releases latent heat and increases the plume buoyancy, thereby enhancing the plume height. Woods (1993a) and Koyaguchi and Woods (1996) have shown that in a saturated atmosphere this effect enhances the plume height by about 2 km if the eruption mass rate is smaller than $10^5$ kg s$^{-1}$, and that it is negligible for eruptive mass rates larger than $10^6$ kg s$^{-1}$.

### 2.4. Summary: dominant controls on plume heights

We conclude that the rise of a volcanic plume is mostly sensitive to source parameters, i.e. the shape and dimensions of the vent and the intensity of fragmentation. Our analysis shows that the rise of plumes above basaltic eruptions is impeded relative to silicic ones, (1) first because fragmentation is less efficient in basaltic eruptions, and (2) second because these eruptions can be fed by rather long fissures. One additional parameter that has not been taken into account in plume models is the change of thermal structure in the atmosphere above a lava flow. A large basaltic eruption will progressively heat the atmosphere and thus decrease the density gradient, i.e. the value of $S$ in Eq. (2), which acts in turn to increase the height of the plume. We thus investigate the impact of heating by large lava flows on the thermal structure of the atmosphere and its consequences on the ascent of volcanic plumes.

## 3. Rise of volcanic plumes through an atmosphere heated by penetrative convection

### 3.1. Penetrative convection above large lava flows

During a basaltic eruption, the lava that is ejected is partitioned into a finely fragmented fraction that gets incorporated into a plume and lava clots that feed a flow. Lava flows that are associated with large basaltic eruptions spread over wide areas and represent large sources of heat. Rates of heat loss in a $10^3$–$10^4$ W m$^{-2}$ range have been recorded above lava lakes and Hawaiian flows (Hardee, 1979; Harris, 2008). The rate of heat loss decreases as a stable crust develops at the top of the lava, but remains significant for a long time. For example, a heat loss of more than 220 W m$^{-2}$ was estimated for a Stromboli lava flow one year after the end of eruption (Gaonac'h et al., 1994). Large amounts of heat are therefore supplied to the atmosphere and are likely to induce strong convective upwellings (Fig. 2). Such convection gradually erodes the stable atmospheric density gradient and generates a well-mixed heated layer that grows with time, and has been called "penetrative convection".

Penetrative convection is a canonical example of the interaction between convective elements and their environment. The details of the flow are complex, but, for our present purposes, it is sufficient to describe the average structure of the atmospheric layer, as shown for example in Figure 3. For a basal heat flux $Q$, the thickness $L$ of the heated layer obeys the following equation (e.g. Lenschow, 1974; Weill et al., 1980):

$$L\frac{dL}{dt} = \frac{Q}{\rho C_p \gamma}, \tag{5}$$

where $\gamma$ is the stable potential temperature gradient in the atmosphere, $\rho$ is the density of air and $C_p$ its specific heat. Heat input from normal ground in the morning generates a well-mixed atmospheric boundary layer that extends to altitudes of ≈1–2 km in a few hours due to penetrative convection (Heidt, 1977; Kaimal, 1976; Lenschow, 1974; Weill et al., 1980), as shown in Figure 4 for a typical ground heat flux of 200 W m$^{-2}$.

For a lava flow, the heat flux into the atmosphere decreases as a function of time due to cooling and to the growth of a crust at the top of the flow. The heat flux through the crust is given by:

$$Q(t) = -k\frac{T_s(t) - T_c}{h_c(t)}, \tag{6}$$

where $T_s(t)$ is the surface temperature, $T_c$ is the temperature of the lava upon emplacement (1150 °C for typical basaltic flows; Harris et al., 2007), $h_c(t)$ is the crust thickness and $k = 2$ W m$^{-1}$ K$^{-1}$ its thermal conductivity (Harris et al., 2007). The detailed physics of the cooling process are complex but can be described by empirical formulae specifying the surface temperature and crust thickness. We consider here the evolution of temperature at the surface of a pahoehoe flow measured by Hon et al. (1994):

$$T_s(t) = -140 ln(t) + 303, \tag{7}$$

where $T_s$ is in °C and time $t$ is in hours. The thickness of the cooling crust is given by the solution of Stefan's problem:

$$h_c(t) = h_0 + 2\lambda\sqrt{\kappa t}, \tag{8}$$

where $h_0$ is the initial thickness of the crust, $\kappa = 7 \cdot 10^{-7}$ m$^2$ s$^{-1}$ is the thermal diffusivity (Harris et al., 2007) and $\lambda = 0.421$ is a dimensionless constant that depends on latent heat (Turcotte and Schubert, 2002). We used $h_0 = 20$ cm as a starting condition, which corresponds to an initial

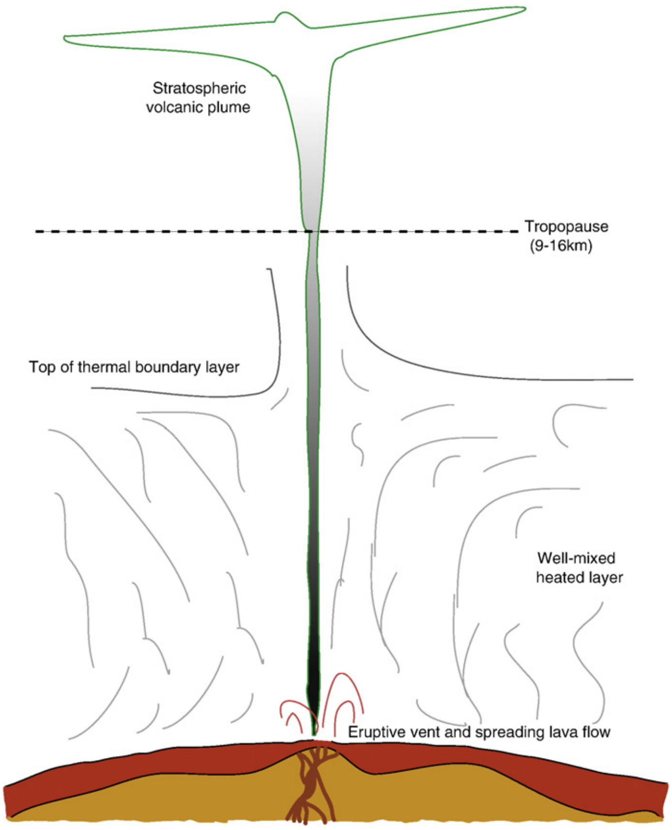


**Fig. 2.** Schematic representation of a plume rising through an atmospheric layer that has been heated by penetrative convection above an active lava flow. The thickening of the heated boundary layer as a function of time enhances the plume buoyancy and increases its final height.

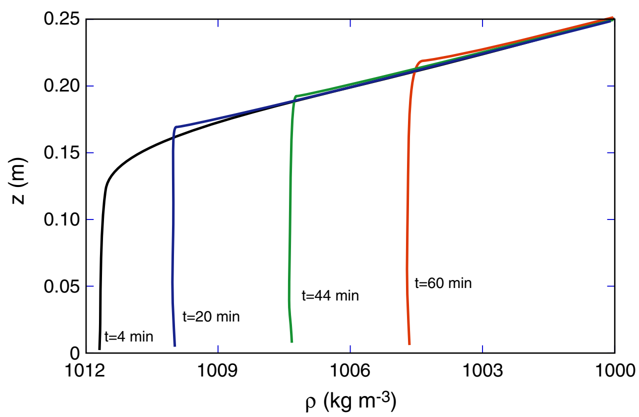


**Fig. 3.** Evolution of the density profile resulting from penetrative convection in a saline convection tank used as an analog for atmospheric convection (adapted from Van Dop et al., 2005). The surface buoyancy flux released in the tank is $4 \cdot 10^{-6}$ m$^2$ s$^{-3}$ during 1 h, which corresponds to the scaled heating of the convective boundary layer during the day (Van Dop et al., 2005). The density profile is initially linear and stable. Penetrative convection generates a layer of uniform density of increasing thickness below an unperturbed stable stratification.

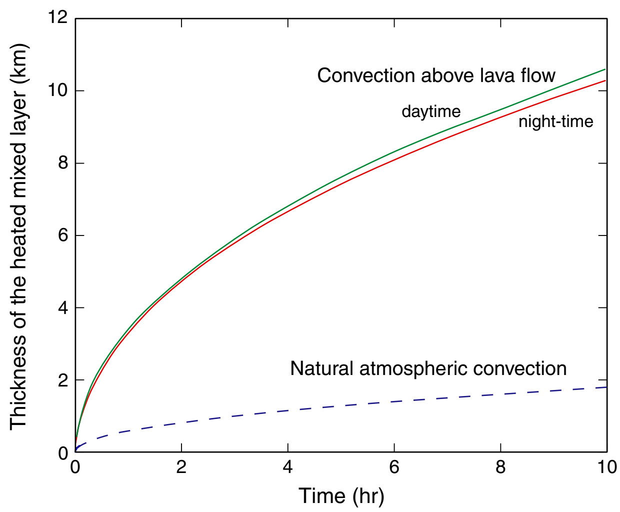


**Fig. 4.** Thickness of the heated boundary layer generated by penetrative convection induced by the solar heat flux (dashed line) and by a lava flow in the daytime and in the night-time (green and red solid lines).

heat flux of $5.5 \cdot 10^3$ W m$^{-2}$. The predicted thickness of the heated atmospheric layer is shown in Figure 4 for these conditions. We find that penetrative convective generates a well-mixed layer of more than 5 km in less than 2 hours and may even reach the stratosphere. Its effect on the rise of volcanic plumes is thus likely to be quite large.

### 3.2. Influence of penetrative convection on the rise of volcanic plumes

At mid latitudes, the average temperature gradient $\gamma_0$ in dry atmosphere is about 6.5 K km$^{-1}$. In a well-mixed layer heated by penetrative convection, the temperature gradient is isentropic,

$$\gamma_{ad} = \frac{g}{C_p} \approx 10\,\mathrm{K\ km}^{-1}, \tag{9}$$

and the temperature of the heated atmosphere is given by

$$T(z) = T_0 + L(\gamma_0 - \gamma_{ad}) + \gamma_{ad} z, \quad \text{for } 0 < z < L \tag{10}$$

with $T_0$ the reference ground temperature. We use this temperature profile and the plume model of Carazzo et al. (2008) to calculate the rise of a turbulent plume through a well-mixed heated layer as a function of the plume mass rate. As the volcanic plume rises through the heated atmospheric layer, entrainment of essentially isothermal air does not modify the buoyancy flux significantly. Above the heated layer, the buoyancy flux decreases as increasingly cold surrounding air gets entrained and the plume eventually reaches its peak height. Results are presented in Figure 5 and clearly demonstrate that the impact of a heated boundary layer on plume ascent is very important for mass rates smaller than $10^6$ kg s$^{-1}$. The effect of the boundary layer increases with increasing thickness, because it reduces the extent of stably stratified air that must be penetrated to reach stratospheric levels. Under normal atmospheric conditions with no heating by lava flows, the boundary layer rarely exceeds 1–2 km in thickness and hence does not affect the rise of volcanic plumes significantly. This explains why there is good agreement between observations and theoretical predictions of Eq. (2) for Plinian eruption columns. If the boundary layer has the same thickness as the troposphere – 10 km in our case – the plume always reaches the stratosphere, even for very low values of the mass rate. This conclusion is illustrated further by Figure 6, where we show the minimum mass rate required to reach the tropopause as a function of the thickness of heated air. For example, for a $L = 8$ km, this minimum mass rate is smaller than that for $L = 0$, i.e. for standard atmosphere that has not been heated by lava flows. As expected, the influence of the plume mass rate becomes negligible when the boundary layer reaches the tropopause. Hence penetrative convection significantly enhances the rise potential of volcanic plumes and can easily allow them to reach the stratosphere.

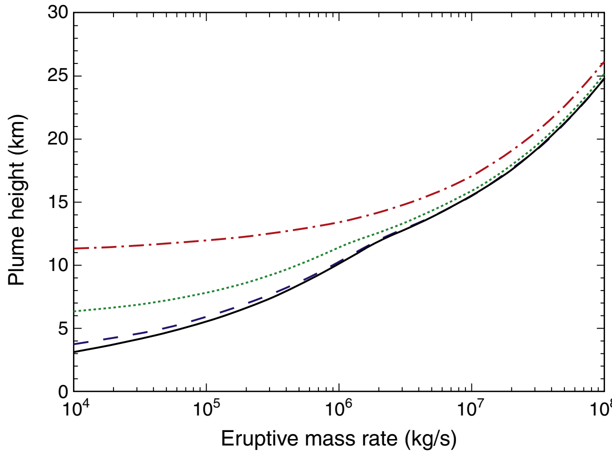


**Fig. 5.** Maximum height of a plume rising through a well-mixed heated atmospheric boundary layer for different thicknesses of the boundary layer: 0 km (solid line), 2 km (dashed line), 5 km (dotted line) and 10 km (dotted and dashed line). The effects are significant for boundary layer thicknesses of 5 km or more.

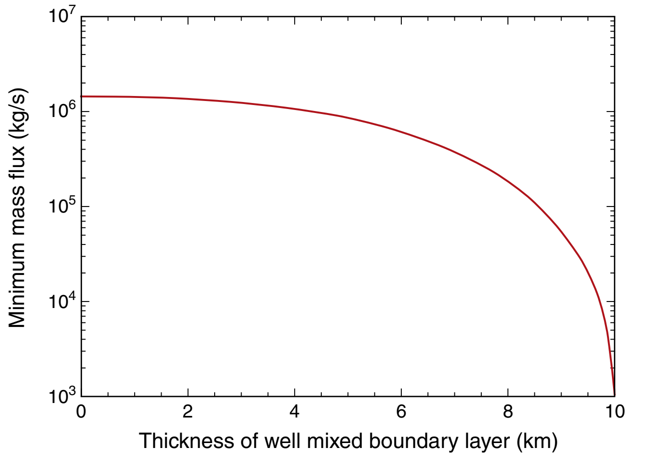


**Fig. 6.** Minimum mass flux required for a volcanic plume to reach the stratosphere as a function of the thickness of the well-mixed boundary layer generated by penetrative convection. For a boundary layer thickness of 8 km, compared to a plume rising through a standard atmosphere, the required mass rate is smaller by one order of magnitude. The influence of the plume mass rate is negligible when the boundary layer extends to the tropopause.

## 4. Discussion: the rise of turbulent plumes above continental flood basalts

For maximum impact on the atmosphere, a volcanic plume must reach the stratosphere, i.e. an altitude that depends on latitude and that is typically about 10 km (Table 1). For a Plinian plume, in which all the erupted material is fragmented, the mass rate must be larger than about $1.5 \cdot 10^6$ kg s$^{-1}$ (e.g. Carazzo et al., 2008; Woods, 1988, and Eq. (2)). For a basaltic plume and a fragmentation efficiency of 10%, the minimum mass rate is much larger, about $10^7$ kg s$^{-1}$ in moist atmosphere (Woods, 1993a). Mass rates of $10^7$ kg s$^{-1}$ or more are not uncommon for Plinian eruptions, but are rare in basaltic eruptions. A first point to consider is thus the value of the mass rate of continental flood basalts.

The 1783 Laki eruption has been used as a (micro-)template for CFB eruptions (e.g. Chenet et al., 2005; Thordarson and Self, 2003). The average mass rate of this eruption was $2 \cdot 10^6$ kg s$^{-1}$, with a peak mass rate of $1.2 \cdot 10^7$ kg s$^{-1}$ at the beginning of activity. These rates yield an average plume height of 7 km, and a peak height of about 10 km, i.e. just above the tropopause. The Laki plume was thus able to inject volcanic aerosols into the stratosphere and hence to affect the climate at the scale of the Northern Hemisphere, but only in the first stages of the eruption. That may explain why a sulfur $\Delta^{33}$S anomaly is recorded only in the first levels of Greenland ice cores that are associated with this eruption (Lanciki et al., 2009). The Laki impact, however, pales in comparison to that of a CFB event. The major difference stems from the fact that individual flow volumes, and amounts of gases emitted in a single eruptive event must be scaled up by factors up to 1000. Moreover, Laki may not be a good reference for a typical CFB because (1) the fragmentation of volcanic material was probably enhanced by the presence of ice and water at Laki (Hamilton et al., 2010) and (2) the length of the Laki eruptive fissure was small enough for the plume source to be treated as a point source (Woods, 1993a).

Fissures feeding CFB eruptions typically extend over tens to hundreds of kilometers. For a long eruptive fissure, the convective plume is best described in 2D, which decreases the plume height as shown in Figure 1. For example the mass rate must be $3 \cdot 10^8$ kg s$^{-1}$ to generate a 10 km high plume from a 10 km eruptive fissure, which may be a lower bound for the very long CFB feeder dykes. Given the important influence of the source characteristics, one must allow for the potential segmentation of an eruptive fissure in a series of smaller eruptive centers. In such conditions, each segment can be treated as an isolated point source if it is sufficiently far from the adjacent ones. For example, a series of one kilometer-long vents separated by more than 10 km along the same eruptive fissure could be treated as independent point sources. Hence if we can conclude that a plume associated with a mass rate smaller than $10^7$ kg s$^{-1}$ will not reach the stratosphere, a larger rate may still not be sufficient depending on the exact geometry of the source.

Magnetostratigraphic and geochronological studies show that most of the volume of CFB events (typically a few million km$^3$) is generally emplaced in 1 Ma or less, (e.g. Courtillot and Renne, 2003). For the eruptions listed in Table 1, we calculate average eruptive mass rates ranging from $2.6\times10^4$ to $2.8\times10^6$ kg s$^{-1}$. Assuming that these eruptions did form explosive turbulent plumes, these average rates would yield plume altitudes below the tropopause. This calculation, however, relies on an average mass rate estimate. In reality, the eruption sequence is likely to have been irregular with large peak values that were reached in a few short pulses. When available, detailed chronology of the eruptive sequence allows refined estimates. Using a combination of physical, volcanological, geological, and paleomagnetic arguments, durations that are as short as 10 to 100 years have been inferred for individual units within the ≈15 Ma Roza flow, Columbia River, (Thordarson and Self, 1998) and the ≈65 Ma Deccan traps (Chenet et al., 2008; Chenet et al., 2009). In the Deccan case, the eruption mass rate value may have been $10^8$ kg s$^{-1}$ or more, which would be sufficient for a plume rising from a point source to reach the stratosphere. If the eruptive fissure was longer than 10 km, however, even such an extreme mass rate would not be sufficient for a stratospheric impact.

According to the previous arguments, the potential ability of CFB eruptions to inject volcanic gases above the troposphere and hence to affect the global environment remains an open question, the answer to which requires additional constraints on the temporal and spatial segmentations of the eruptions: (1) the occurrence of short-lived eruptive peaks with larger flow rates, and (2) the spacing between adjacent eruptive centers. Unfortunately, such constraints are usually out of reach for prehistorical eruptions. One way to overcome these uncertainties is to consider the effect of penetrative convection.

Penetrative convection is naturally present in the environment due to the solar heat flux, and its presence in an enhanced form above lava flows – even rather cold ones – is unavoidable. Strong convection currents associated with lava flows have been witnessed for example at the peak of the Nyiragongo 2002 eruption in Congo (Komorowski et al., 2002). The effusive eruptive mass rate, close to $2 \cdot 10^5$ kg s$^{-1}$ (Houlié et al., 2006), was large enough to induce violent winds that blew out tents and straw houses. The effect was strong but localized, because the areal extent of the lava was only about 3.5 km$^2$. For the large surfaces covered by CFBs, one expects much more dramatic effects and vigourous coinvection leading to a thick well-mixed atmospheric boundary layer. As shown in Figure 5, the thickness of the heated boundary layer is typically 2 km only in a non-volcanic environment, but is expected to increase to 8 km in about 6 h above a lava flow. From Figure 6, one can conclude that thermal plumes above CFBs were always able to reach the stratosphere after a few hours, even if the eruption rate was at the value of the long-term average. Furthermore, it is likely that penetrative convection developed over the whole troposphere during CFB events. In such cases, there is no need for a focussed eruptive plume to carry volcanic gases to the stratosphere. A weak plume, small plumes at the edges of the lava flows, or even no plume at all can account for climatic changes. This conclusion implies in turn that all CFBs did affect the global environment, as observed, regardless of the details of the eruption process. The magnitude of the climatic impact, however, was variable and this must be explained.

We note that all CFBs, including the “smaller” Columbia River one, consist of several large volcanic pulses with similar characteristics, i.e. unit volumes in excess of 1000 km$^3$ and emplacement durations that are less, and possibly much less, than 100 years (Barry et al., 2010; Chenet et al., 2008, 2009; Self et al., 2008a; Thordarson et al., 1996). The impact of a CFB eruption on the global environment is therefore not sensitive to the eruption dynamics (mass rate and magma volatile content — even if those have some influence), but rather results from the dynamical response of the climate system to the intermittent injection of volcanic gases, as proposed for explosive eruptions by (Zielinski, 1995).

The climatic response depends not only on the total amount and composition of the volcanic gases, but also on the duration and time sequence of individual injection events (Chenet et al., 2008, 2009; Courtillot and Renne, 2003; Self et al., 2008b). The 1991 Pinatubo eruption led to global atmospheric cooling for 10 years (Jones et al., 2005). One may expect that several basaltic eruptions, much larger than Laki and following one another in a sequence, could lead to large cumulative (or even “runaway”) atmospheric effects if they occur at intervals of less than 10 years. Such a scenario can be proposed for the exceptionally intense climatic cooling that followed the 1815 Tambora eruption. This eruption, in fact, occurred at the end of an exceptional sequence of large eruptions separated by a year or so: Mayon 1814, Vesuvius 1813, St Vincent and Awu (Indonesia) 1812 and the “unknown” 1809 eruption (Cole-Dai et al., 2009). In a sense, therefore, CFBs may be thought of as producing a similar atmospheric build-up out of a single volcanic system. Vagaries of the eruption sequence would account for the vastly different climate and biotic consequences of otherwise apparently rather similar CFBs (e.g. Siberian traps and Permo-Triassic extinction, the largest of all, vs. Karoo traps and end-Pliensbachian extinction, one of the smaller ones).

## 5. Conclusion

Explosive eruptions affect the environment if the associated volcanic plumes are able to rise to the stratosphere. Such an outcome is favoured for plumes that are fed from a small vent or crater and for eruption rates in excess of $10^7$ kg s$^{-1}$. Huge effusive basaltic eruptions, such as continental flood basalts, are not likely to maintain powerful explosive plumes, but generate buoyant plumes at the top of fire fountains. However, they also cover extremely wide areas with hot lava, which induces penetrative convection in the lower atmosphere. Such convection generates a well-mixed heated atmospheric layer that may extend to the tropopause in about eight hours, which allows the injection of volcanic gases into the stratosphere. The variable effect of CFBs on the global environment is likely to reflect the exact duration and time sequence of eruptive pulses (Graf et al., 2007).

## Acknowledgements

The manuscript benefitted from fruitful comments by Takehiro Koyaguchi and Steve Self, that help to clarify and strengthen the paper. This work is IPGP contribution 3097.

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