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Technical Note

Non-adiabatic boundaries and thermal stratification in a confined volume

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Abstract

Heating experiments have been performed during 1.5–18 days in a confined room of volume about 300 m^3 in an underground quarry near Paris (France). During heating, a thermal stratification of the atmosphere is observed. After a time depending on the presence or not of ventilation, this stratification reaches a stationary state during which the temperature difference between the rock and the atmosphere is constant. During this stationary phase, both temperatures increase linearly with time. We propose here a new model of filling box with radiative heat exchanges and cooling by direct contact at the boundaries, which accounts for the observed vertical profiles of temperature in the atmosphere, and for the temperature variation with time. Such contributions of heat transfer at the boundaries are important in situations such as fire in confined cavities, presence of visitors in a painted cave, or in the study of building ventilation.

1. Introduction

The stratification associated with turbulent plumes in confined environments is interesting to study for numerous industrial or geophysical applications [1,2]. In the case of painted caves in the presence of visitors, or of other kinds of artificial heat sources, it is important to estimate the thickness and temperature elevation of the top warm layer. Indeed, this stationary warm layer can affect dramatically the state of the paintings [3]. Similarly, in an underground vault containing warm radioactive containers, it is important to assess properly the thickness of the hot layer to design properly remediation strategies. In the presence of fire in a confined volume [4], it is also important to estimate the size of the hot layer to properly install emergency exhausts.

In 1969, Torrance et al. provided qualitative observations of atmospheric plumes in enclosure containing a small hot spot [4]. The filling box model was then developed by Baines and Turner [5] to describe the evolution of density in a box containing a turbulent plume. This model describes the system by two sets of conservation equations: in the plume, as written by Morton et al. [6], and in a horizontal layer in the environment surrounding the plume. This approach was validated using experiments with water [5,6]. With similar equations and experiments, Linden et al. extended the filling box model to the case of a ventilated volume [7].

However, such models of plume assume a heat transfer by convection only, but, in air, they may need to be modified. Indeed, in air, the contribution of radiative transfer may be important [8,9]. In addition, thermal exchanges with the boundaries have to be considered in the case of non-adiabatic walls [10].

To study these effects, underground sites appear particularly appropriate. First, they are characterized by stable temperature conditions in the absence of perturbations, which is a big advantage while studying the effect of heating with sources of low power. Second, they offer the opportunity to study heat exchanges with boundaries in natural conditions, with volumes significantly larger than laboratory models. In a first series of experiments, we have studied the thermal stratification of the atmosphere in the

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presence of a heat source [10]. The interpretation was limited by a poorly constrained contribution of ventilation. In addition, the experiments were of short duration, which did not allow us to investigate the long term effect of the heating on the rock temperature.

In this paper, we present three long lasting experiments performed in a room of this underground cavity, after an insulating wall has been built to eliminate the effect of ventilation. The thermal stratification of the atmosphere is analysed in details. After 15–30 h of heating in confined conditions, a stationary state is reached, during which the temperature difference between the atmosphere and the rock is constant. We use this observation to propose an improved model based on the filling box mechanism, including two mechanisms of heat exchange with the boundaries: radiative transfer and cooling by contact.

2. The experimental set-up

The underground limestone quarry of Vincennes is located 5 km South-East of the center of Paris (France), at a mean depth of about 18 m. Water drips from the roof and the walls in some places, and the atmosphere is almost saturated with water vapour, with a relative humidity larger than 99.2%. The concentration of carbon dioxide of the air has a yearly cycle related to the natural ventilation [11], and varies from 0.1% in winter to 1% in summer.

The experiments take place in an isolated part of a room with a mean height of 2.2 m (Fig. 1). The mean temperature in this room was 12.7 °C in 2004 with fluctuations smaller than 0.05 °C peak to peak in a day, at least in

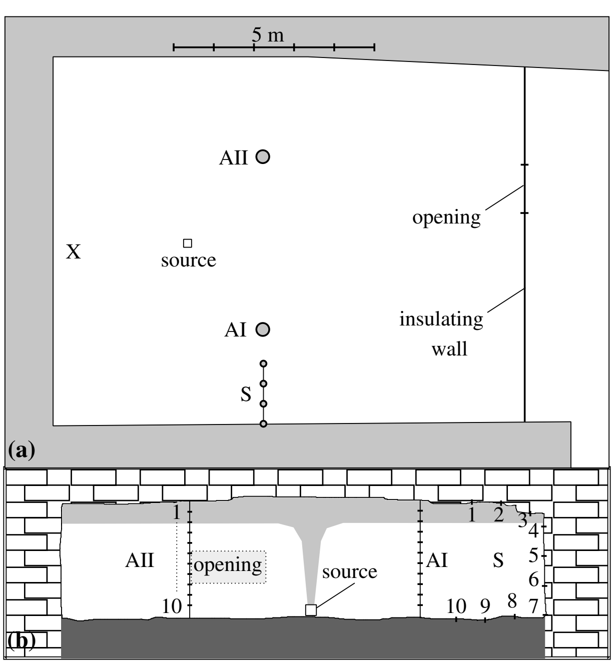

Fig. 1. Sketch of the experimental room showing the location of the insulating wall, the source and the measurements set-ups AI, AII and S (a) on a map, and (b) on a view from point X indicated in (a). A supposed shape for the plume is also depicted in (b).

the absence of perturbations such as visits or heating experiments. These fluctuations are essentially due to atmospheric pressure variations [12]. In addition, the cavity is characterized by a yearly temperature cycle of amplitude 0.07 °C peak to peak. The experimental room has an area S of $12 \times 9\ \mathrm{m}^2$ for a height H varying from 2 to 2.5 m (Fig. 1). The space is bounded by rock at the ceiling and on the three sides of the walls, and on the floor by fillings of 2 m thickness. The fourth side of the room is closed by a Styrofoam™ partition of 5 cm thickness, indicated as the "insulating wall" in Fig. 1, and equipped with a door of area $124 \times 60\ \mathrm{cm}^2$. Because of the confinement, carbon dioxide level is always higher than in the rest of the quarry, and close to 1%.

Three set-ups of 10 thermistors each are installed in the room. Sensors of set-ups labeled AI and AII in Fig. 1 measure the vertical profile of temperature in the atmosphere. Sensors of set-up S are a few centimeters deep in the rock or in the filling covering the floor. The thermistors are inter calibrated with a precision of about 0.005–0.01 °C [13].

Two heat sources have been used. The first one had a power of 100 W, and was surrounded by a polymer screen to limit the direct radiation. The second heat source, with a power of 800 W, was made of four elements of 200 W each. The first heating experiment was performed with the first source, and lasted about 18 days, from 10:00 May 8 to 08:00 May 26, 2004. From the beginning of the experiment to 06:00 May 21, the door in the insulating wall was open; it was then closed until the end of the heating. For the second experiment, the same heat source was turned on from 13:00 December 2 to 08:00 December 7, 2004; the door was kept closed during the whole experiment. The third experiment was performed with the 800 W heat source, from 17:30 April 27 to 06:00 April 29, 2005. During these 1.5 days, the door was also closed.

3. Results

Fig. 2 shows temperature variations measured in the air by sensors of set-up AI during the three heating experiments. A previous study [13] has shown that, in the absence of heating, the vertical variations of temperature in the atmosphere are smaller than a few 10^{-3} °C m^{-1}, hence negligible. Thus, we consider the initial temperature as uniform, and $\Delta T_{\mathrm{a}}(t)$ will be the difference between the temperature at time t and its mean value over the 45 min time span before the beginning of the heating. Note, in the graph relative to the second experiment (Fig. 2b), the perturbations due to the presence of the operators, occurring 2–3 h before the beginning of the heating.

Three phases can be distinguished in the three experiments. The first phase lasts a few hours and shows a fast temperature elevation, with an amplitude increasing with height from 0.03 to 0.5 °C for the two experiments with the 100 W source, and from 0.07 to 1.2 °C for the third experiment with the 800 W source. During the second phase, the atmosphere temperature increase starts leveling

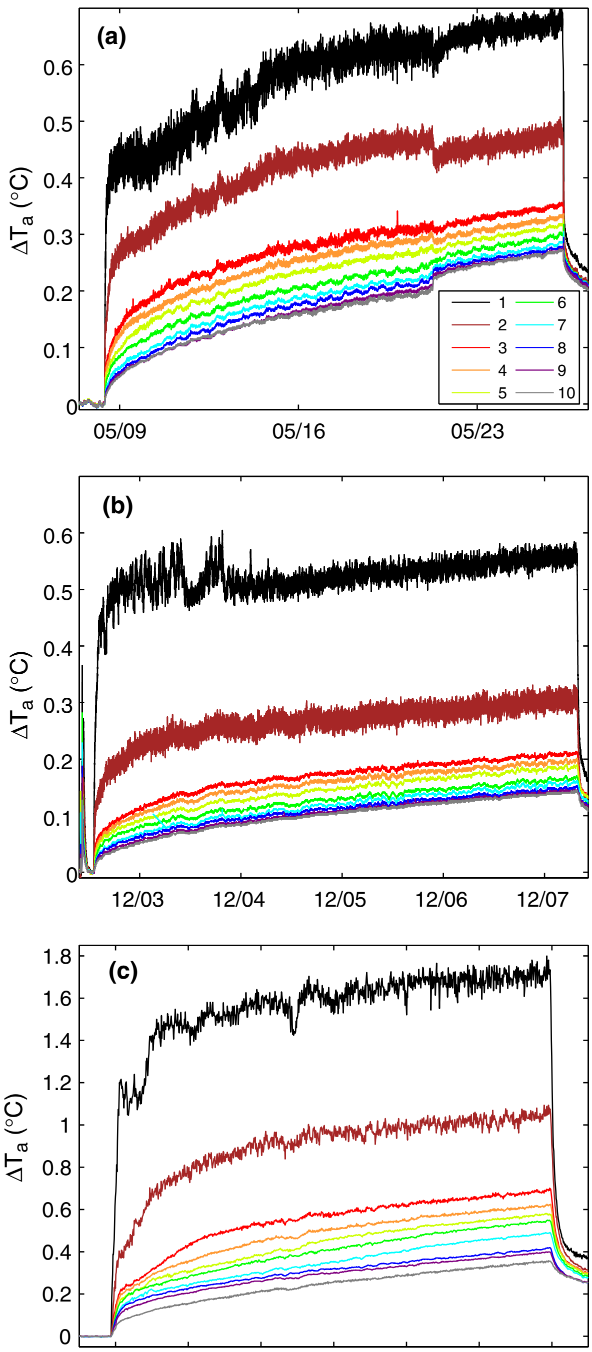

Fig. 2. Temperature variation measured in the atmosphere by sensors of set-up AI during the three heating experiments. In this figure, ΔT_a is the temperature change compared with a temperature reference estimated before heating. Note that time scales are different for the three experiments.

off. This phase lasts about 1 week for experiment 1 (Fig. 2a), 30 h for experiment 2 (Fig. 2b) and 15 h for experiment 3 (Fig. 2c). The large difference between the durations of the second phase in the two first experiments could be due to the fact that, in experiment 1, the door was open during 2 weeks whereas it was closed all along the second experiment. During the third phase, temperature increases linearly with the same rate for all the sensors in the atmosphere and in the rock. The data of experiment 1 (Fig. 2a) clearly show a rupture in the temperatures evolution at 06:00 on 21 May, which corresponds to the closing of the door. The stable linear increase is then recovered after about 1 day. The rate of temperature elevation is almost the same for the two linear increasing phases of experiment 1, with a value of about $7.5 \pm 0.5 \times 10^{-3}$ °C per day. During the third phase of experiment 2, this rate is about $15 \pm 2 \times 10^{-3}$ °C per day, and about 0.12 ± 0.03 °C per day for experiment 3. Note that these values are about 10–100 times larger than the slope associated with the yearly cycle of temperature in the quarry.

Variations of temperature with time scales from 1 h to 1 day are mostly due to atmospheric pressure variations. Such pressure induced temperature variations are regularly observed in the unperturbed state of the room [12], for example visible before starting the heating in Fig. 2a.

Fig. 3 shows temperature variations measured in the rock by sensors of set-up S during the first heating experiment. As in Fig. 2, $\Delta T_r(t)$ is the difference between the temperature at time t and its mean value, calculated over the 45 min time span before the beginning of the heating. As expected, temperature elevation in the rock is slower than in the atmosphere. The first and second phases of heating observed in the air in Fig. 2 correspond to one phase in the rock in Fig. 3. During this phase, $\Delta T_r(t)$ is approximately proportional to $\sqrt{t}$ [10]. The rock temperature then increases almost linearly with time during the last phase.

Fig. 4 represents the evolution of the difference between the atmosphere temperature at four heights and the mean rock temperature defined by

$$T_r(t) = \frac{1}{S_r}\left(\frac{\sum_{i=1}^{3} T_{ri}(t)}{3} S_{\text{ceiling}} + \frac{\sum_{i=4}^{7} T_{ri}(t)}{4} S_{\text{walls}} + \frac{\sum_{i=8}^{10} T_{ri}(t)}{3} S_{\text{floor}}\right), \quad (1)$$

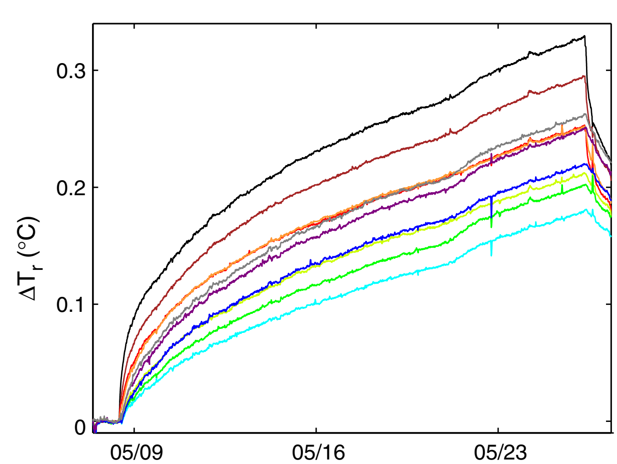

Fig. 3. Temperature variation measured in the rock by sensors of set-up S during the first heating experiment. Similarly to Fig. 2, ΔT_r corresponds to the temperature difference with the value of temperature before a heating. Data are filtered with a moving average of 10 min.

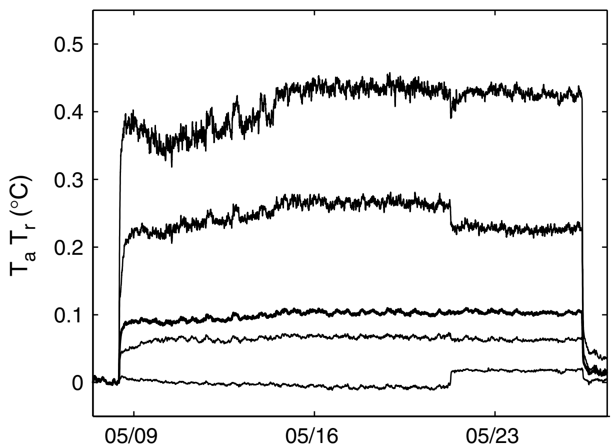

Fig. 4. Temperature difference between the atmosphere and the rock as a function of time for the first heating experiment. Data used for the temperature in the atmosphere are those measured by sensors 1, 2, 5, and 10 of set-up AI. For the rock, the mean temperature is calculated using Eq. (1). The bold line represents the mean difference of temperature between the air and the rock in the whole room. Data are filtered with a moving average of 30 min.

where $S_{\mathrm{walls}} = 79.2\ \mathrm{m}^2$ is the surface of the walls made of rock, $S_{\mathrm{ceiling}} = S_{\mathrm{floor}} = 108\ \mathrm{m}^2$ are respectively the surfaces of the ceiling and of the floor, $S_{\mathrm{r}} = S_{\mathrm{walls}} + S_{\mathrm{ceiling}} + S_{\mathrm{floor}}$ is the total rock surface, and $T_{\mathrm{r}i}$ the temperature measured by sensor i of set-up S (Fig. 1). During the third phase of each experiment, we notice the remarkable fact that the temperature difference between the air and the rock remains constant. The mean value of $T_{\mathrm{a}} - T_{\mathrm{r}}$ is about 0.10 °C for experiment 1, 0.08 °C for experiment 2, and 0.3 °C for experiment 3. Fig. 4 also shows that during the third phase of experiment 1, the mean value of $T_{\mathrm{a}} - T_{\mathrm{r}}$ is constant (bold line), before and after the closing of the door, although differences are observed for individual sensors.

4. Interpretation

To model the stratification and its evolution during the third stationary phase, as in the previous study [10], we have extended the filling box model [5] to the case of a non-adiabatic environment. As a development of this study, we will now take into account the change of rock temperature, and use a more refined description of the heat transfers between the atmosphere and the rock.

As in Baines and Turner model [5], the equations developed by Morton et al. [6] are used to describe the plume. Temperature variations in the surrounding atmosphere, assumed to be stratified, are then calculated using mass and energy conservations in a horizontal layer. This last equation is modified to take into account heat exchanges between the atmosphere and the rock, supposed to take place in two manners. First, when the turbulent plume reaches the ceiling, a layer of warm air spreads onto it and is cooled by direct contact with the rock, before being pushed down as warmer air arrives from the plume. Second, heat exchanges with the rock then proceed by radiative transfer in the whole volume, and by contact with the rock at the boundaries.

In the following, we will limit ourselves to the stationary phase. Density differences are supposed to be small and due to temperature variations; their effect is considered only in the buoyancy term that is defined as follows in the plume:

$$\Delta(z) = g\frac{T(z) - T_{\mathrm{a}}(z)}{T_0}, \tag{2}$$

where g is the gravity, $T(z)$ and $T_{\mathrm{a}}(z)$ respectively the temperatures inside and outside the plume, and T_0 a reference absolute temperature, taken here as the initial temperature.

In the plume, horizontal profiles of buoyancy Δ and vertical velocity w are assumed to be Gaussian:

$$w(z,r) = w(z)\exp\left(-\frac{r^2}{b(z)^2}\right), \tag{3}$$

$$\Delta(z,r) = \Delta(z)\exp\left(-\frac{r^2}{b(z)^2}\right), \tag{4}$$

where $b(z)$ is defined as the radius of the plume at altitude z. Morton et al. [6] have shown that, in a uniform environment, the buoyancy flux $F = \frac{\pi b^2 w\Delta}{2}$ is constant and equal to the source buoyancy flux

$$F_0 = \frac{gP}{\rho_{\mathrm{a}} C_{\mathrm{a}} T_0}, \tag{5}$$

where P is the power of the source, and ρ_{a} and C_{a} are respectively the density and capacity at constant pressure of air. From mass, momentum and energy conservations, they deduce the following expressions for the plume parameters in a uniform environment:

$$b = \frac{6}{5}\alpha z, \tag{6}$$

$$w = \frac{5}{6\alpha}\left(\frac{18}{5\pi}\alpha F_0\right)^{\frac{1}{3}} z^{-\frac{1}{3}}, \tag{7}$$

$$\Delta = \frac{5}{3\pi}\left(\frac{5\pi}{18}\right)^{\frac{1}{3}} \alpha^{-\frac{4}{3}} F_0^{\frac{2}{3}} z^{-\frac{5}{3}}, \tag{8}$$

where α is the entrainment constant. In our experiments, the temperature increase in the environment is small compared with the temperature increase in the plume. Therefore, to calculate the plume parameters, as checked elsewhere [10], it is sufficient to consider a uniform environment.

The vertical velocity u_z of a horizontal layer in the surrounding air can be deduced from mass conservation in a layer, which gives

$$u_z(z) = -\frac{\pi b^2(z) w(z)}{S} = -\mu z^{\frac{5}{3}}, \tag{9}$$

with

$$\mu = \frac{6\pi}{5}\left(\frac{18}{5\pi}\right)^{\frac{1}{3}} \alpha^{\frac{4}{3}} F_0^{\frac{1}{3}} S^{-1}. \tag{10}$$

As temperature differences between the air and the rock are small, the ΔT^4 terms can be linearized in the expression of the radiative heat flux $\phi_r(z)$ from the air to the rock in a horizontal layer, thus

$$\phi_r(z,t) = \lambda_r(T_a(z,t) - T_r(t)), \tag{11}$$

where λ_r is the radiative conductance. Heat exchanges with the floor, made of fillings, probably do not obey exactly the same equation. However, taking them or not into account does not change significantly the results. The mean emissivity ϵ of the air of the room can be deduced from its contents in CO_2 and water vapour, and from the dimension of the room, assuming that all the boundary surfaces behave as black bodies. Using graphs compiled by McAdams [14], we find a value of about 0.23 ± 0.02, which can be used in the following estimation of λ_r:

$$\lambda_r = 4\epsilon\sigma T_0^3, \tag{12}$$

where $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan's constant. The expected value of λ_r is thus about 1.2 ± 0.1 W m^{-2} K^{-1}.

The heat flux $\phi_c(z)$, corresponding to the cooling of the layer by direct contact at the walls is also proportional to the difference of temperature between the air and the rock:

$$\phi_c(z,t) = \lambda_c(T_a(z,t) - T_r(z,t)). \tag{13}$$

This cooling includes the diffusion of the air heat to the rock, and phase changes of water. The heat transfer by condensation of water can indeed be the major contribution [15].

In a first approximation, we will take $T_r(z,t) = T_r(t)$ in this expression. Supposing that both conductances λ_r and λ_c are constant in the whole volume, energy conservation in a layer then gives

$$\begin{aligned}\rho_a C_a S \frac{\partial T_a}{\partial t} &+ \frac{\partial}{\partial z}(\rho_a C_a S u_z (T_a - T_0)) \\ &= \kappa_a \frac{\partial^2 T_a}{\partial z^2} \rho_a C_a S - \rho_a C_a \alpha w 2\pi b (T_a - T_0) \\ &\quad - \left(\lambda_r \frac{S_r}{H} + \lambda_c p\right)(T_a - T_r),\end{aligned} \tag{14}$$

where κ_a is the thermal diffusivity of air, and p the room perimeter. Using the expression of u_z from Eq. (9) and the fact that T_r is independent of z, we can then write

$$\frac{\partial T_a}{\partial t} = \kappa_a \frac{\partial^2 (T_a - T_r)}{\partial z^2} + \mu z^{\frac{5}{3}} \frac{\partial (T_a - T_r)}{\partial z} - \frac{T_a - T_r}{\tau}, \tag{15}$$

where

$$\tau = \frac{\rho_a C_a S}{\left(\lambda_r \frac{S_r}{H} + \lambda_c p\right)} \tag{16}$$

is a characteristic time for heat transfer in a layer.

The data (Fig. 4) strongly suggest that, during the stationary phase, the difference of temperature between the air and the rock is constant everywhere with time, and thus a function of z only: $T_a - T_r = y(z)$. This fact has an immediate consequence in Eq. (15), which is that the time derivative $\frac{\partial T_a}{\partial t}$ is constant. This is actually what we observe during the stationary phase: a linear increase of the temperature in the atmosphere (Fig. 2) as well as in the rock.

The vertical profile of the difference of temperature between the air and the rock, $y(z)$, can then be calculated from

$$\kappa_a \frac{d^2 y}{dz^2} + \mu z^{\frac{5}{3}} \frac{dy}{dz} - \frac{y}{\tau} = \Gamma, \tag{17}$$

with Γ constant. To solve this equation, two boundary conditions are required. Taking $\kappa_a = 2 \times 10^{-5}$ m^2 s^{-1} and reasonable values for $y(0)$ and $y(H)$ from the data, the numerical resolution of Eq. (17) shows that the diffusion term $\kappa_a \frac{d^2 y}{dz^2}$ is negligible, thus the profile is constrained by only one boundary condition. The simplified analytical solution is then

$$y(z) = (y(H) - \Gamma\tau) \exp\left[-\frac{3}{2\mu\tau}\left(z^{-\frac{2}{3}} - H^{-\frac{2}{3}}\right)\right] + \Gamma\tau. \tag{18}$$

The temperature profile thus depends on four parameters: $y(H)$, Γ, τ, and μ. The value of μ, given by Eq. (10), is fixed by the geometry, the source power and the value of the entrainment constant α. An estimation of Γ is made using the time evolution of the temperature in the air and in the rock (Figs. 2 and 3). For the experiments performed with the 100 W source, $\Gamma = 1.4 \pm 0.6 \times 10^{-7}$ °C s^{-1} and for the experiment with the 800 W source, $\Gamma = 1.4 \pm 0.3 \times 10^{-6}$ °C s^{-1}. The rate of temperature increase Γ thus appears to be almost proportional to the power of the source in the considered range of powers. The value of $y(H)$ is also determined from the data (Fig. 5), which gives $y(H) = 0.6 \pm 0.1$ °C for experiments 1 and 2, and $y(H) = 1.8 \pm 0.2$ °C for experiment 3. Thus only one parameter, τ (Eq. (16)), remains unknown.

Fig. 5 compares the data with the results of our computation of $y(z)$ with various assumptions for τ. We have noticed that the contribution of the terms $\Gamma\tau$ in Eq. (18) is always negligible. The estimation of τ is thus constrained by the shape of the exponential term of Eq. (18). Our model is in good agreement with the measured temperature profiles for $\tau = 6 \pm 2$ min, a value which is reasonably compatible with a previous estimate obtained from pressure induced temperature variations [12]. These values correspond to $\lambda_c = 6.5 \pm 2.5$ W m^{-2} K^{-1}.

To estimate the contribution of the cooling of the upper layer as it spreads on the ceiling, we make use of energy conservation in the whole room:

$$P = \frac{\rho_a C_a S}{\tau} \int_0^H y \, dz + \Gamma \rho_a C_a S H + P_{\text{ceiling}}. \tag{19}$$

The first term of this equation represents the power lost by heat transfer from the air layers to the rock. The second term is the heating of the atmosphere with time, and appears to be negligible compared to the other two terms. The third term corresponds to the cooling of the upper layer. This last term is about 47 ± 8 W for experiments 1

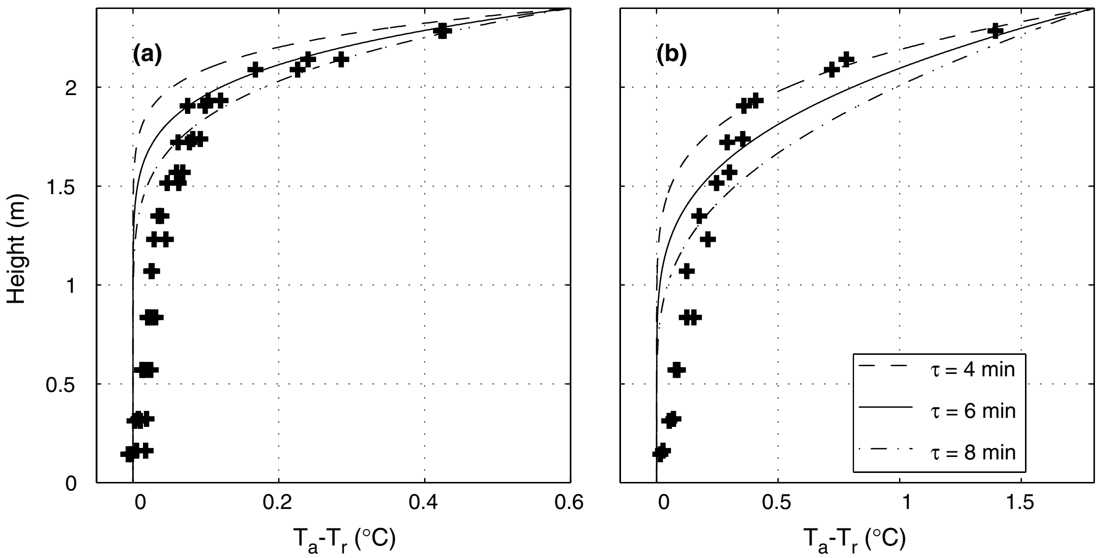

Fig. 5. Vertical profiles of temperature difference between the air and the rock measured during the stationary state in confined conditions for experiments 1 and 2 (a), and for experiment 3 (b). The lines represents the calculated profiles obtained from the model, using three different values of τ and $y(H) = 0.6$ °C for the two first experiments (a), and $y(H) = 1.8$ °C for experiment 3. The following values of the other parameters were used: $\rho_a = 1.2$ kg m^{-3}, $C_a = 1013$ J kg^{-1} K^{-1}, $\alpha = 0.083$, $g = 9.8$ m s^2, $T_0 = 286$ K, $S = 108$ m^2, $H = 2.4$ m, and $P = 100$ W (a) or 800 W (b).

and 2, and about 520 ± 35 W for experiment 3, thus about 65 ± 4% of the power of the source.

5. Discussion

In this paper, we have presented heating experiments performed in an underground quarry with heat sources of power 100 and 800 W. We make two essential observations: first, the temperature difference between the rock and the atmosphere reaches rapidly a stationary profile, which remains constant with time throughout the heating duration; secondly, during this equilibrium regime, the temperatures of both the atmosphere and the rock increase linearly with time. Using a simple model based on conservative physical assumptions, this latter observation can be shown to be a consequence of the former. This model describes the thermal stratification induced by a thermal plume in a confined non-adiabatic volume (see Eq. (15)). The resulting stratification is radically different from the stratification in adiabatic conditions: whereas the hot layer invades the whole space in the classical filling box [5], in the present model, the hot layer remains confined to the upper part of the room.

In this model, the temperature profile, for both source powers 100 and 800 W, results from the plume entrainment and the cooling provided by the boundaries. Its shape can be accounted for by a single parameter, τ (Eq. (16)), whose value provides an estimate of the global heat conductance between the air volume and the rock surface. This effective heat conductance includes radiative transfer and other contributions originating from the direct contact of the convective layer with the wall. Among these latter contributions, phase changes of water seem to play a dominant role. However, the value of the slope of the temperature increase Γ, which we have observed to be the same everywhere in the rock and in the atmosphere, is actually poorly constrained by the energy budget. This may explain why different values of this parameter are observed in subsequent experiments. This slope, in a first approximation, is almost proportional to the power of the source.

We note that the temperature increase in the upper layer is four times larger for the 800 W source than for the 100 W source. This is compatible with the idea that the temperature increase in the upper layer, ΔT_a, is proportional to the temperature excess in the plume, ΔT, at height H, which scales with the 2/3 power of the buoyancy flux F_0 (according to Eq. (8)).

We thus conclude that our model gives a physical description which is satisfactory enough to scale our results from 100 to 800 W. This can be useful in practice, for example to estimate the effect of the presence of six visitors, which corresponds to about 800 W. This situation represents for instance visitors in the Altamira painted cave, whose volume happens to be comparable to our room volume [16]. We have actually confirmed this conclusion by performing another experiment with twelve persons. In this experiment, we have observed a temperature increase of 0.9 °C in the uppermost layer after 10 min, again compatible with our model although the conditions are quite different in this case compared with a single point-like source. This gives some confidence that, in a first approximation, our model remains valid beyond 800 W, which might be extended up to a few kilowatts, relevant in many practical applications in cave preservations or underground waste storage.

To build a comprehensive numerical model, it would be necessary to incorporate the details of the physical processes such as radiative transfers between the rock and the gas volume, and within the gas. Such a model would need to be carefully checked and adjusted against new and dedicated experimental data. Our experiments and model, despite their limitations, provide, to our knowledge, a new insight in the poorly known problem of the effect of heating in a confined and non-adiabatic volume.

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