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One-dimensional simulation of fire injection heights in contrasted meteorological scenarios with PRM and Meso-NH models

S. Strada[1], S. R. Freitas[2], C. Mari[1], K. M. Longo[2], and R. Paugam[3]

[1]Laboratoire d'Aérologie, University of Toulouse and CNRS, UMR5560, Toulouse, France
[2]Center for Weather Forecasting and Climate Studies, INPE, Cachoeira Paulista, Brazil
[3]Department of Geography, King's College London, London, UK

Correspondence to: S. Strada (susanna.strada@aero.obs-mip.fr), S. R. Freitas
(saulo.freitas@cptec.inpe.br)

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Abstract

Wild-fires release huge amounts of aerosol and hazardous trace gases in the atmosphere. The residence time and the dispersion of fire pollutants in the atmosphere can range from hours to days and from local to continental scales. These various scenarios
5 highly depend on the injection height of smoke plumes. The altitude at which fire products are injected in the atmosphere is controlled by fire characteristics and meteorological conditions. Injection height however is still poorly accounted in chemistry transport models for which fires are sub-grid scale processes which need to be parametrised. Only recently, physically-based approaches for estimating the fire injection heights have
10 been developed which consider both the convective updrafts induced by the release of fire sensible heat and the impact of background meteorological environment on the fire convection dynamics. In this work, two different models are used to simulate fire injection heights in contrasted meteorological scenarios: a Mediterranean arson fire and two Amazonian deforestation fires. A Eddy-Diffusivity/Mass-Flux approach, for-
15 merly developed to reproduce convective boundary layer in the non-hydrostatic meteorological model Meso-NH, is compared to the 1-D Plume Rise Model. For both models, radiosonde data and re-analyses from the European Center for Medium-Range Weather Forecasts (ECMWF) have been used as initial conditions to explore the sensitivity of the models responses to different meteorological forcings. The two models
20 predict injection heights for the Mediterranean fire between 1.7 and 3.3 km with the Meso-NH/EDMF model systematically higher than the 1-D PRM model. Both models show a limited sensitivity to the meteorological forcings with a 20–30 % difference in the injection height between radiosondes and ECMWF data for this case. Injection heights calculated for the two Amazonian fires ranges from 5 to 6.5 km for the 1-D PRM model
25 and from 2 to 4 km for the Meso-NH/EDMF model. The difference of smoke plume heights between the two models can reach 3–4 km. A large difference is obtained for the windy-wet Amazonian fire by the 1-D PRM model with a injection height 1.5 km higher when ECMWF re-analyses are used compared to the run with the radiosonde

forcing. For the Mediterranean case, both models forecast a plume injection height above the boundary layer, although there are evidences that this particular fire propagated near the surface, highlighting the current limitations of the two approaches.

1 Introduction

5 The spatial scale at which fire emissions may impact the chemical composition of the atmosphere depends on their dispersion, a process that is highly influenced by the height of injection of fire products. The smoke plume injection height is defined as the altitude at which the smoke particles are injected into the atmosphere before transport (Kahn et al., 2008). If fire pollutants stay trapped in the the Planetary Boundary

10 Layer (PBL), their residence time can be shortened by removal processes that act more efficiently in the first layer of the atmosphere (Chatfield and Delany, 1990; Stein et al., 2009), and fire products may interact with urban and rural emissions with diverse consequences (e.g. Phuleria et al., 2005; Bytnerowicz et al., 2010). On the contrary, if fire emissions reach the free troposphere, their chemical lifetime increases and fire

15 products are transported by faster winds that allow the spread of their effect on air quality from a local to a regional and occasionally intercontinental scale (Gidel, 1983; Saarikoski et al., 2007; Sofiev et al., 2008; Turquety et al., 2009; Dirksen et al., 2009).

The final height of injection of a smoke plume is a complex parameter to determine (i.e. to measure and to simulate). It is a result of dynamical interactions between the

20 fire induced buoyant flows and the background environment (Freitas et al., 2006; Sofiev et al., 2012). It can change with time according to the fire propagation and evolution of the fire heat fluxes. Wildfires are intense sources of heat that is released in the atmosphere in the form of hot gases and water vapour. The contribution to sensible heat flux from wildland fires can not be neglected: values of sensible heat flux measured

25 during on-field campaigns are nearly 3 orders of magnitude higher than natural fluxes (e.g. Clements et al., 2007; Silvani and Morandini, 2009). Under favourable meteorological conditions, fire-induced sensible heat flux has even the potential to enhance

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deep convection (so-called pyro-convection), leading to direct injection of smoke into the upper troposphere and lower stratosphere, as observed by Fromm et al. (2005) and Damoah et al. (2006), and as modelled by Trentmann et al. (2006). Smoke plumes are also masses of humid air with an increase of water vapour mixing ratio of nearly 30 % over the ambient air within the plume (Clements et al., 2006) due to the release of fuel and "combustion" moisture (Parmar et al., 2008). Also the plume-environment interaction plays an important role in the convection process. The hot fire plume interacts with the cooler surrounding air: this phenomenon can trigger turbulent eddies. Hence, fire-induced turbulence can efficiently mix environmental colder air into the fire plume, cooling the hot plume and reducing its upward movement (Freitas et al., 2006). Furthermore, turbulent motions can bring in the uprising plume some humid air from the environment leading to a gain of extra buoyancy from latent heat of condensation of the water vapour (Freitas et al., 2007). Additional buoyancy may also be gained if the rising haze plume reaches the Lifting Condensation Level (LCL), as already investigated for volcanic plume rise (Graf et al., 1999), beyond which water vapour begins to condense and latent heat is released. Studies on volcanic activities also showed the effect of strong horizontal winds on the final height of plumes (Bursik, 2001). The interaction between the plume and strong winds favours lateral entrainment of air, increasing the horizontal momentum. Particularly for small fires, this phenomenon results in plume bending and might reduce the updraft development because of loosing the additional buoyancy from condensate water vapour (Freitas et al., 2010).

When run at very high resolution, atmospheric models can resolve explicitly convective transport and turbulent motions. Instead, at larger resolution (e.g. meso-scale), several types of atmospheric movements are sub-grid processes, and they are incorporated into atmospheric models through appropriate parametrisation schemes. As explained before, wild-fires can induce direct and rapid transport into the atmosphere, this process may have considerable impacts on the atmospheric dynamics and on pollutants distribution (Luderer et al., 2006; Tressol et al., 2008). Strong updrafts associated with fires are frequently ignored, or their impact is diluted, at the typical resolution of

large-scale models (Rio et al., 2010). Using three chemistry transport models (CTMs) driven by the same meteorology, Elguindi et al. (2010) performed sensitivity tests that underlined the role of low injection heights in the model's poor representation of the CO plumes. Singh et al. (2012) minimally succeeded in the attempts to simulate the com-

⁵ plex interactions of fire emissions with urban and rural air. These peculiar interactions had been documented by airborne observations over California during large wildfires. Among shortcomings in air quality models that may contribute to the disagreement between observations and simulations, the authors cited uncertainties in plume-rise estimation.

¹⁰ Several studies were carried out using remote sensing data to investigate the height to which smoke plumes rise and the variability of this altitude due to fire characteristics. Labonne and Chevallier (2007) assessed the injection height of biomass burning plumes by analyzing the vertical distribution of aerosols, a good marker of fire emissions. They compared released data from the Cloud and Aerosol Lidar for Pathfinder

¹⁵ Spaceborne Observations (CALIPSO) and the mixing layer top diagnosed by the European Center for Meteorological and Weather Forecasting (ECMWF); they concluded that biomass burning plumes were injected within the mixing layer. The same method was used by Amiridis et al. (2009); their results outlined that, under strong fire activity, the ECMWF diagnostic underestimates the BL height. Kahn et al. (2008) suggested

²⁰ to combine lidar observations with stereo imaging to support the modelling of smoke environmental impacts. Mazzoni et al. (2007) utilized stereo imaging from the Terra Multi-angle Imaging Spectro-Radiometer (MISR) and the Moderate Resolution Imaging Spectro-radiometer (MODIS) date to locate fires and their smoke plumes, and they retrieved the injection height generated by fire buoyancy over a 4-month period. This

²⁵ work was extended by ValMartin et al. (2010) that analyzed a 5-yr record of MISR smoke plume injection heights over North America. Their analysis of plume heights indicated that 4–12 % of plumes from fires are injected above the BL; moreover, the MISR plume climatology exhibited larger summertime heights that, once correlated with MODIS Fire Radiative Power (FRP) measurements, seemed to be the result of

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higher fire intensity, likely due to most severe fire intensity during summer. The study of ValMartin et al. (2010) confirmed the conclusions of Sofiev et al. (2009) who showed, over a 2-yr period, the prevalence of fire injection heights within the BL over the US. Gonzi and Palmer (2010) used satellite observations of CO, a tracer of incomplete

5 combustion, in order to estimate the vertical transport of surface fire emissions. Considering boreal and tropical wild-fires, they found that only 10–25 % of emissions are injected above the PBL. Guan et al. (2010) proposed a simple empirical method to identify biomass burning plume heights using the Aerosol Index (AI) measurements as determined by satellite instruments. The authors derived a best-fit relationship between

10 AI and maximum plume height for young plumes that could help to validate the vertical placement of smoke plumes in CTMs.

In CTMs, vertical plume distributions have often been represented by means of empirical or arbitrary procedures such as uniformly distributing fire emissions within a few layers close to the surface up to a prescribed height which varies with the studied region

15 (Liousse et al., 1996; Lamarque et al., 2003; Tressol et al., 2008). Other methodologies are to release fire products at different heights (Elguindi et al., 2010) or directly in the upper atmosphere when it comes to pyro-convection (Hyer et al., 2007). Some studies propose a link between plume height and fire intensity (Lavoué et al., 2000; Miranda, 2004; Hodzic et al., 2007). To improve these techniques, more physical and dynam-

20 ical methods have been developed and implemented in meso-scale models with the scope to parametrise the plume lifting by taking into account fire characteristics and environmental conditions. A 1-D entrainment Plume Rise Model (1-D PRM) was presented by Freitas et al. (2006, 2007). It could be embedded in a host 3-D meso-scale or global model to simulate explicitly the convective transport mechanism associated with

25 wild-fires and determine the final height where fire products, emitted during the flaming phase, would be released. This plume rise algorithm was implemented and successfully tested with the NCAR Community Atmosphere Model for modelling the transport of carbon monoxide (CO) released by Southern Africa biomass burning, showing significant improvements in the vertical distribution of CO (Guan et al., 2008). PRM was

coupled with the Coupled Aerosol and Tracer Transport model to the Brazilian developments on the Regional Atmospheric Modeling System (CATT-BRAMS, Freitas et al., 2009; Longo et al., 2010). The model of Freitas et al. (2007) was also integrated in the WRF model with Chemistry (WRF-Chem) showing significant improvements in weather forecasting (Grell et al., 2011) and in the emissions transport and dispersion (Sessions et al., 2011). The study of pyro-convection was also addressed by the work of Rio et al. (2010) who, starting from the "thermal plume model" of Hourdin et al. (2002), proposed a "pyro-thermal plume model" based on a mixed Eddy Diffusivity/Mass Flux (EDMF) scheme for convective boundary layer plumes. In the EDMF parametrisation, the updraft and the surrounding environment directly interact through local and non-local mixing, respectively associated with the turbulent transport and the mass-flux term; the methodology of the 1-D PRM model relies on the assumption that at rough resolution (grid-scale $\sim$ 30 to 100 km) fires do not have significant effects on the dynamics and the thermodynamics of the host model. The works of Freitas et al. (2010) and Rio et al. (2010) mark the current state of the art in the domain of atmosphere-wildfire interaction and they underline the challenge that remains when it comes to dealing with fire injection height. So far, these approaches have been applied and partially validated for some wild-fires occurred in different scenarios (boreal, Amazonian and African fires), some of which presented elevated injection heights. Recently, Sofiev et al. (2012) proposed a new methodology to evaluate the smoke-injection height from wildfires that is similar to existing Convective Available Potential Energy (CAPE) computations used for describing deep convection. The authors applied their methodology to remote-sensing data for about 2000 fire plumes in North America and Siberia. They also compared their diagnostic of the smoke-injection height with other methods, such as the MISR plume-top, Briggs' plume-rise formulas, the 1-D PRM BUOYANT (Martin et al., 1997), and the prescribed injection height widely used in CTMs. Just selecting a part of the MISR data-set, Sofiev et al. (2012) showed a significant improvement in their diagnostic compared with cited approaches. Moreover, in contrast to other approaches

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(e.g. Freitas et al., 2010), they stated that the wind speed is unimportant for the wild fire plume height.

The main goal of the present study is to delve into the dynamics of strong updrafts associated with wild-fires, focusing on the main actors participating in the smoke plume rise process: heat fluxes (sensible and latent), turbulence and entrainment of ambient air. Moreover, this work aims to investigate the impact of weather conditions on fire evolution, taking into account different meteorological forcings. For this purpose, sensitivity analyses are designed to compare two approaches for predicting the fire injection height, once both numerical models operate with similar environmental and fire conditions. The Meso-NH model is used at a kilo-metric scale in a 1-D configuration to study a typical Mediterranean fire (Lançon-de-Provence 2005) and two deforestation fires burnt in 2002 in the Amazon basin under different meteorological conditions. The capacity of the EDMF scheme in Meso-NH (Pergaud et al., 2009) is here investigated related to strong convective processes associated with wild-fires. Results from the 1-D Meso-NH/EDMF model are compared to corresponding simulations generated by the 1-D PRM model of Freitas et al. (2010). For both models, radiosonde data and re-analyses from the ECMWF are used as initial conditions to explore the sensitivity of both models to different meteorological forcing.

The paper is organized as follows. Section 2 introduces the selected case studies, discussing differences in the atmospheric profiles between radiosondes and ECMWF re-analyses. Section 3 gives a brief description of the two 1-D models (Meso-NH/EDMF and PRM). In Sects. 4 and 5, simulation results are presented and discussed, after having defined an unifying metrics adopted for the comparison. Finally, conclusions are summarized in Sect. 6.

2 Data-sets selected for the inter-comparison

Three wild-fires have been selected as case studies for the comparison exercise and are described in this section: a Mediterranean arson fire and two Amazonian

deforestation fires. These cases differ from one another in vegetation characteristics and in meteorological conditions, suggesting a different evolution of the smoke plume rise.

For each of the three scenarios, initial and boundary meteorological conditions are from radiosondes and operational re-analyses from the ECMWF. The difference in the initial atmospheric profile between a radiosonde and re-analysis field has consequences on the atmosphere that is simulated by the numerical model, leading to different behaviours for the fire plume.

Figure 1 shows the atmospheric conditions in the first kilometres for the Mediterranean fire. The vertical profiles of temperature, wind speed, potential temperature and water vapour mixing ratio are traced up to an height of 8 km for the radiosonde data (dashed line) and the ECMWF analysis (solid line). Data for the Amazonian fires are presented respectively in Figs. 2 and 3.

2.1 Lançon-de-Provence 2005

On 1 July 2005, an arson wild-fire broke out at about 07:40 UTC, 09:40 CEST (Center European Summer Time) to the east of Lançon-de-Provence (south-eastern France, 43.60° N, 5.20° E), threatening downwind inhabited areas and cultivated lands. At 12:00 UTC, on the burning area, firefighters measured a temperature of 26 °C, a wind speed of 46 km h^{-1}, a wind direction of 330° and a relative humidity of 20 %. Documented favourable weather conditions led to the fire spreading easily. After 8 h of burning, the Lançon fire was put out and the burnt area estimated: nearly 626 ha, mainly covered by shrub-land and forest.

Twice a day (at 00:00 and at 12:00 UTC) a radiosonde is launched over Nîmes, a city located 63 km northwest away from Lançon-de-Provence (43° N, 4° E). Looking at radiosonde data measured on 1 July 2005 at 12:00 UTC (14:00 local time), temperatures decrease from 25 to ∼ 0 °C in the first 5 km of the atmosphere (Fig. 1a). A well marked temperature inversion is observed at nearly 760 hPa, around 2.3 km; above this altitude the atmosphere becomes more stable as depicted by the positive slope of the potential

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temperature trend (Fig. 1b). This last graphic highlights unstable conditions at the surface ($\partial_z \theta < 0$), followed by a well developed mixed layer where θ is constant: this is the typical convective boundary layer of a summer early afternoon. Figure 1c shows a dry CBL with water vapour mixing ratio that decreases from $6.5\,\mathrm{g\,kg^{-1}}$ at the surface to nearly $1.0\,\mathrm{g\,kg^{-1}}$ at 700 hPa. The layer above is moister, probably as a result of the radiosonde crossing a cloud; normally this "wet" layer would not interact with the fire plume since strong winds, together with a dry BL, may efficiently prevent fire plume rise, as pointed out by Trelles et al. (1999) and Freitas et al. (2007). Strong northwesterly winds blow over the region with speeds ranging from 8 to $20\,\mathrm{m\,s^{-1}}$ in the first 2 km of the atmosphere (Fig. 1d).

The Lançon-de-Provence fire constituted a benchmark for fire spread models (Fore-Fire and FARSITE, Balbi et al., 2009); moreover, the dynamics and the chemistry downwind of the Lançon fire were investigated by Strada et al. (2012).

2.2 Rondônia 2002

Rondônia is a state in Brazil located in the north-western part of the country and bordered Bolivia. Rondônia landscape underwent a rapid conversion between 1984 and 2002 after the opening of the BR-364 highway and the introduction of pasture in the 70s. With one of the fastest rate of tropical deforestation (Lovejoy, 1991), rain-forests have been soon replaced by agricultural and pasture lands by means of fires (de Barros Ferraz et al., 2005).

In 2002, during the burning season, two radiosondes were launched at 18:00 UTC (on 20 and 27 September) near a deforestation area in Rondônia ($11.0°$ S, $60.0°$ W). Radiosonde time, 18:00 UTC, is 14:00 local time when the diurnal cycle of Amazonian fires reaches its peak (Freitas et al., 2010) and convective structure are well developed (Chou et al., 2007). The selected days differ in wind intensity and atmospheric humidity, for this reason the considered case studies have been renamed as follows: the calm-dry case corresponds to 20 September 2002; the windy-wet case refers to 27 September 2002.

Considering the radiosoundings:

- *Calm-dry case.* On 20 September 2002 temperatures pass from 35 to $\sim 0\,°C$ in the first 5 km of the atmosphere; a strong thermal inversion is observed at around 800 hPa, $\sim 2\,km$ (Fig. 2a). Below 800 hPa the potential temperature and the water vapour mixing ratio are constant in the daytime mixed layer; above the BL is capped by the stably stratified and drier free atmosphere, with r_v decreasing abruptly from 12 to $3\,g\,kg^{-1}$ (Fig. 2b, c). The wind speed accelerates in the first kilometres from 2 to $4\,m\,s^{-1}$, then the speed decreases to $1\,m\,s^{-1}$ at 700 hPa, around 3 km (Fig. 2d). Analysing the zonal and the meridional wind, also a directional wind shear is identified in the first 3 km (not shown).

- *Windy-wet case.* On 27 September 2002 the radiosonde registers a weaker temperature inversion at lower levels (around 870 hPa, $\sim 1.5\,km$, Fig. 3a). The height of the daytime mixed layer is nearly 1 km ($\partial_z \theta \approx 0$, Fig. 3b). Above, in the stable atmosphere, r_v decreases suddenly from 12 to $9\,g\,kg^{-1}$ (Fig. 3c). The wind speed increases with height up to an altitude of 2 km with a strong wind shear from 2 to $6\,m\,s^{-1}$ (Fig. 3d).

These two meteorological situations have already been chosen as case studies for other model comparisons (Freitas et al., 2007, 2010). Significant differences in ambient wind and humidity between the calm-dry and the windy-wet case will permit to better understand the role and the importance of environmental conditions on the smoke rise process, having selected the same fire characteristics for both cases.

2.3 Comparison between radiosondes and ECMWF analyses

2.3.1 Lançon-de-Provence 2005

The ECMWF analysis shows a weaker temperature inversion at a slightly lower altitude compared to the radiosounding: 800 hPa, around 1.8 km (Fig. 1a). The potential

temperature presents a less marked instability at the ground, followed by a mixed layer that stretches up to about 800 hPa (Fig. 1b). The trend of the water vapour mixing ratio describes a drier atmosphere at the ground-level and a moister one in the mixed layer, but in general the ECMWF atmosphere is drier compared to the one described by the radiosonde (Fig. 1c). In Fig. 1d the ECMWF wind speeds are quite similar in the first 1.5 km to those of the radiosonde, except for a relative maximum around 2 km of altitude.

2.3.2   Rondônia 2002

– *Calm-dry case.* A rapid inspection of the ECMWF vertical profiles showed a striking difference between the ECMWF and the radiosonde profiles. In the first 2 km the ECMWF atmosphere looks cooler, more stable, with a water vapour mixing ratio monotonically decreasing (Fig. 2a–c). Moreover, the ECMWF vertical profile is moister in the first kilometer, then it becomes drier. Between the ground surface and an altitude of 2 km, winds are $2\,\mathrm{m\,s^{-1}}$ weaker on the average if compared to the radiosonde atmosphere, in the layer above the magnitude relation inverts. Furthermore, the ECMWF profile shows a weaker wind shear.

– *Windy-wet case.* As for the calm-dry case, the ECMWF vertical profiles look very different compared to those traced using the radiosonde. In the first 2 km the ECMWF atmosphere is cooler, even of $10\,°C$ at the surface (Fig. 2a); moreover, it is more stable with potential temperature monotonically increasing with height, Fig. 2b. The water vapour mixing ratio decreases and the ECMWF atmosphere is highly moister than the radiosonde: at the inversion height the difference in water vapour mixing ratio equals $6\,\mathrm{g\,kg^{-1}}$ (Fig. 2c). Concerning wind speeds, the comparison is quite good for the windy-wet case (Fig. 2d).

Comparing the Lançon-de-Provence case study to the two Amazonian cases, it is worthy to note that the Amazon Basin offers a warmer, moister and less windy atmosphere for the fire starting.

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3 Description of the one-dimensional models

This section is mainly devoted to describe the numerical models that have been used and compared in the present work: the 1-D PRM and the Meso-NH/EDMF. Because of the computational efficiency of a 1-D model and the ability to isolate a column of
5 atmosphere for study, a single column model (SCM) is an ideal environment in which to develop and test parametrisations (Randall et al., 1996).

3.1 The Meso-NH 1-D/EDMF model

Meso-NH is a meteorological research model jointly developed by the Centre National de Recherche Météorologiques (Météo France) and the Centre National de Recherche
10 Scientifique (Laboratoire d'Aérologie) (Lafore et al., 1998). This numerical model was designed to simulate atmospheric motion at different scales (from the large meso down to the micro scale) using the non-hydrostatic assumption and the an-elastic approximation. Meso-NH has a standard three dimensional (3-D) configuration, however it is here run as a SCM, using version 4.8-4. In the present study, cloud micro-physical processes
15 follow a two-moment scheme, using three water phases with five species of precipitating and non-precipitating liquid and solid water (Pinty and Jabouille, 1998). Turbulent motions are represented by the quasi-1-D scheme of Bougeault and Lacarrère (1989). The Eddy-Diffusivity/Kain-Fritschl parametrisation is utilized for representing shallow convection (see Sect. 3.1.2 for further information).
20 The used square grid-mesh has an horizontal resolution of 1 km and the vertical grid has 70 levels, with a level spacing stretching from 40 m near the ground to 600 m at higher altitude. The integration time is one hour with a time-step of one second. Due to the short duration (1 h) of the simulation, radiative processes are neglected (i.e. the downward radiative flux is put to zero) and the Coriolis parameter f is set to zero. The
25 orography is not taken into account, depicting a flat domain.

Dynamical variables are initialized and constrained prescribing a stationary vertical profile (i.e. the initial and the final state of the atmosphere are the same). Two

different types of vertical profile are used: observational soundings recorded in the vicinity of the burnt area, on the day of the fire; and vertical profiles generated from operational re-analyses of the ECMWF, selecting the same UTC hour of the observational radiosonde and the nearest location to the radiosonde launch station. For the three scenarios, radiosondes and ECMWF re-analyses are available at same hours (12:00 UTC for Lançon and 18:00 UTC for the Rondonia fires) when the diurnal cycle of wild-fires is more pronounced (Hodzic et al., 2007; Freitas et al., 2009).

The Interactions Soil-Biosphere-Atmosphere scheme (ISBA, Noilhan and Planton, 1989) is used for parametrising exchanges between the atmosphere and natural or agricultural lands and provides surface energy fluxes to the atmosphere. In order to have the same contribution from the ground, in terms of fluxes, the same kind of vegetation cover has been chosen in Meso-NH for all three scenarios, imposing equal conditions for soil humidity and temperature. The selected cover type is cerrado; temperatures of the surface soil layer, the root zone soil layer and the deep soil layer have been set to 303.53 K; and soil water index (SWI) is zero for the surface soil layer and 0.2 for the root zone and the deep soil layer.

The fire forcing is activated through heat and scalar fluxes that are prescribed at the ground level in the Meso-NH model: Sect. 3.1.3 gives further details on this technique.

3.1.1 1-D Meso-NH general equations

A SCM is a stand-alone model that can be pictured as a single vertical array of grid-point cells placed at a specific geographical location. The column model prognostically calculates the evolution of the vertical structure of some variables based on physical parametrisations. In particular, in the 1-D Meso-NH prognostic variables are: latitudinal and longitudinal wind components (u, v), potential temperature (θ), water vapour (r_v), cloud (r_c) and rain water (r_r) mixing ratios and turbulent kinetic energy (TKE, $\overline{e}$).

The basic equations implemented in the 1-D Meso-NH are:

$$\frac{\partial \overline{u}}{\partial t} = \frac{\partial}{\partial z}\left(K_m \frac{\partial \overline{w'u'}}{\partial z}\right) - w_{ls}\frac{\partial \overline{u}}{\partial z} \tag{1}$$

$$\frac{\partial \overline{v}}{\partial t} = \frac{\partial}{\partial z}\left(K_m \frac{\partial \overline{w'v'}}{\partial z}\right) - w_{ls}\frac{\partial \overline{v}}{\partial z} \tag{2}$$

$$\frac{\partial \overline{\theta}}{\partial t} = \frac{\partial}{\partial z}\left(K_h \frac{\partial \overline{w'\theta'}}{\partial z}\right) - w_{ls}\frac{\partial \overline{\theta}}{\partial z} + Q_\theta^{\text{diab}} \tag{3}$$

$$\frac{\partial \overline{r_j}}{\partial t} = \frac{\partial}{\partial z}\left(K_h \frac{\partial \overline{w'r_j'}}{\partial z}\right) - w_{ls}\frac{\partial \overline{r_j}}{\partial z} + Q_{r_j}^{\text{diab}}; \quad j \in \{v, c, r\} \tag{4}$$

$$\frac{\partial \overline{e}}{\partial t} = \frac{\partial}{\partial z}\left(K_e \frac{\partial \overline{e}}{\partial z}\right) - w_{ls}\frac{\partial \overline{e}}{\partial z} + K_m\left[\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right] + K_h \frac{g}{\theta_V}\overline{w'\theta_V'} - D. \tag{5}$$

Here the primed variables denote perturbations, bar refers to averaged variables. The parameter w_{ls} is the synoptic-scale vertical velocity, K_m, K_h and K_e are the turbulent mixing coefficients for momentum, heat and TKE, respectively. Q_θ^{diab} and $Q_{q_j}^{\text{diab}}$ are the diabatic terms in the heat and humidity equations. The parameter g is the gravitational constant. The temporal evolution of $\overline{e}$ depends on different terms that are, on the right side of Eq. (5), the turbulent transport by eddies, the vertical advection by large scale vertical flow, the shear production or loss term, the buoyancy production, and the dissipation rate of TKE, respectively.

To sum up, in the atmosphere of the SCM the active processes are: vertical advection, turbulent mixing and diabatic exchanges. The vertical velocity is handled as a diagnostic variable.

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3.1.2 The EDMF parametrisation

Recently, a specific parametrisation for shallow convection in the boundary layer has been included in the Meso-NH research model. The new scheme takes into account the impact on the boundary layer dynamics of sub-grid strong convective updrafts gen-
5 erated by surface heating (i.e. dry thermals, Pergaud et al., 2009). The whole concept is based on a scale decomposition where the organized, non-local, strong updrafts are described by an advective Mass-Flux approach (second term on the right side of Eq. 6), whereas the remaining local turbulent field is represented by the Eddy-Diffusivity approach (Siebesma and Teixeira, 2000; Hourdin et al., 2002; Witek et al., 2011; first term
10 on the right side of Eq. 6). As a consequence, the vertical turbulent flux of a conservative variable ϕ is defined as

$$\overline{w'\phi'} = -K\frac{\partial\overline{\phi}}{\partial z} + \frac{M_u}{\rho}(\phi_u - \overline{\phi}), \tag{6}$$

where ρ is the air density, K is the turbulent eddy diffusivity coefficient for the variable ϕ, M_u is the convective mass flux in the updraft $M_u = \rho a_u w_u$ (a_u is the updraft fractional
15 area and w_u is the vertical velocity in the updraft), $\overline{\phi}$ is the mean value and ϕ_u is the updraft value of the variable ϕ. In the following, the subscript u is always used for variables associated with the updraft whereas the subscript e refers to variables associated with the environment.

The basic idea of the EDMF approach is to depict dry thermals as towers of buoyant
20 air rising from the surface and developing in a Convective Boundary Layer (CBL); these strong updraughts are not isolated but they interact with the surrounding environment through turbulent mixing that favours entrainment and detrainment of air masses between the convective parcel and its environment. Therefore, once the EDMF parametrisation is implemented in an atmospheric model, it allows a physical coupling between
25 the updraft and the environmental air: the dynamics and the thermodynamics of both evolve due to a reciprocal influence.

The Mass Flux approach describes the evolution of updraft structures ensuring the mass balance through a diagnostic mass continuity equation:

$$\frac{1}{M_u}\frac{\partial M_u}{\partial z} = \varepsilon - \delta. \tag{7}$$

The mass-flux evolves along the vertical at a rate given by the difference between the entrainment ε and the detrainment δ rate. The definition of entrainment/detrainment rates is the crucial point in EDMF parametrisation: it is at this level that the physical coupling between turbulent mixing and mass flux is done.

The mass-flux profile depends on the vertical velocity of the updraught, whose vertical evolution is affected in turn by a buoyancy term (B_u) and a drag term where the entrainment of environmental air, namely lateral mixing, is accounted for:

$$w_u\frac{\partial w_u}{\partial z} = aB_u - b\varepsilon w_u. \tag{8}$$

The updraft buoyancy acceleration is evaluated related to the difference of virtual potential temperature θ_V between the updraft and its environment, in the absence of phase change in water: $B_u = g(\theta_{u,V} - \overline{\theta_V})/\overline{\theta_V}$; parameters a and b are set to one (Simpson and Wiggert, 1969). The vertical velocity Eq. (8) can be solved to find the top of the updraft imposing $w_u \rightarrow 0$ as boundary condition. Moreover, independent solutions of Eqs. (7) and (8) permit to calculate the vertical variation of the updraft fractional area:

$$a_u = \frac{M_u}{\rho w_u}, \tag{9}$$

that is used to diagnose the cloud fraction, hence to define the sub-grid condensation scheme in the EDMF framework.

Hourdin et al. (2002) noticed a tilt effect on the thermal while introducing the drag effect of apparent wind in the equations for horizontal momentums. However, in the "pyrothermal plume model" of Rio et al. (2010), the momentum transport is not parametrised.

In the EDMF scheme a vertical non-local mixing of momentum is performed by taking into account the contribution of the Mass Flux approach on the horizontal momentum, in addition to the mixing already activated by the turbulent scheme. Hence, the updraft horizontal wind components evolve as

$$\frac{\partial u_u}{\partial z} = -\varepsilon(u_u - \overline{u}) + C_v \frac{\partial \overline{u}}{\partial z}, \tag{10}$$

$$\frac{\partial v_u}{\partial z} = -\varepsilon(v_u - \overline{v}) + C_u \frac{\partial \overline{v}}{\partial z}, \tag{11}$$

where $C_u = C_v = 0.5$; u_u (v_u) is the zonal (meridional) component of wind in the updraft; $\overline{u}$ and $\overline{v}$ are the zonal and meridional mean wind components, respectively.

As pointed out before, the definition of entrainment and detrainment rates characterizes the EDMF parametrisation. Pergaud et al. (2009) chose to draw the definition of lateral mass exchanges from the updraft buoyancy and vertical velocity. Both these parameters are pertinent in shallow convection as they control the mixing rate between the updraft (dry or moist) and its environment. For the dry case, the entrainment/detrainment rate is locally defined as an equilibrium between w_u and B_u:

$$\varepsilon_{\text{dry}}, \delta_{\text{dry}} \propto \frac{B_u}{w_u^2}, \tag{12}$$

For the moist portion of the updraft a different definition of lateral mass exchange is used. In Meso-NH, if the LCL is reached, lateral exchanges are computed using the entraining/detraining plume model of Kain and Fritsch (1990) that considers the cloud radius.

Finally, the scheme initialization is given at the surface (z_{grd}) computing the mass-flux as follows:

$$M_u(z_{\text{grd}}) \propto \rho \left(\frac{g}{\theta_{V,\text{ref}}} \overline{w'\theta'_{V,s}} L_{\text{up}} \right)^{1/3}, \tag{13}$$

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where g is the acceleration of gravity, θ_V is the virtual potential temperature $\overline{w'\theta'_{V,s}}$ is the surface buoyancy flux, L_{up} is the upward mixing length corresponding to the distance that a parcel leaving the ground travels due to buoyancy, as defined by Bougeault and Lacarrère (1989). From the Turbulent Kinetic Energy e at the ground, the vertical velocity of the updraft at the surface results:

$$w_u^2(z_{grd}) = \frac{2}{3}e(z_{grd}).$$ (14)

3.1.3 Fire forcing in Meso-NH

At the horizontal resolution selected for the Meso-NH simulation (1 km), wild-fires are considered as sub-grid processes in the atmospheric model, hence their effect on the atmosphere needs to be parametrise. An off-line coupling is activated between the fire and the atmospheric model (i.e with no feedback of the atmosphere on the fire) by adapting to a 1-D configuration the methodology of Strada et al. (2012) developed for a 3-D simulation. Working in a simple 1-D scenario permits to assess the strength of the fire forcing on the atmospheric dynamics and to discriminate fire effects from other phenomena.

The coupling method consists in providing the burning area (S_b) contained in each Meso-NH grid cell (S_{mnh}) every 2 min, assuming the wild-fire as stationary. At each atmospheric time-step, the surface scheme ISBA accomplishes the fire-atmosphere coupling by computing total wildfire contribution to latent and sensible heat fluxes, taking into account a nominal flux and the surface ratio between the Meso-NH and the total burnt area. Finally, calculated fluxes are taken as inputs at the surface level in the atmospheric model. A step function tunes the starting of the fire during the first five minutes, as done in the 1-D PRM model (Freitas et al., 2007).

The sensible heat flux Φ_S, $kW\,m^{-2}$, is computed as

$$\Phi_S = \phi_s \cdot C \cdot \frac{S_b}{S_{mnh}}.$$ (15)

The nominal value ϕ_S does not separate radiative from convective energy; hence, the multiplication by a reducing factor, C, is necessary to select the percent of total energy effectively available to plume convection. The C parameter ranges from 0.4 to 0.8 depending on fire characteristics and ambient conditions; in order to compare results from the Meso-NH and the 1-D PRM model, the same value selected by Freitas et al. (2010) has been chosen: $C = 0.55$ (from McCarter and Broido, 1965). The surface S_{mnh} measures 100 ha. A constant value is imposed for the burnt area, $S_b = 3.35$ ha, by taking the mean burnt area simulated by a fire propagation model as described in (Strada et al., 2012). This condition insures that during the simulation time step from 12:00 to 13:00, the effective burnt area is 100 ha. The same value is used for the Amazonian fires for which no information on the time evolution of fire burnt area is available.

The latent heat flux Φ_L, $kg\,m^{-2}\,s^{-1}$, is calculated as in the 1-D PRM model:

$$\Phi_L = \frac{\Phi_S}{E_m} \cdot \left[\frac{m}{100} + EF_{H_2O} \right]. \tag{16}$$

The computed sensible heat flux multiplies the heat content E_m ($MJ\,kg^{-1}$) by the sum of fuel moisture m (%) and H2O emission factor (combustion moisture, in $kg\,kg^{-1}$).

A fire tracer is emitted and its flux ($g\,m^{-2}\,s^{-1}$) is defined as follows:

$$\Phi_{fire} = \frac{E_{fire}}{S_{mnh} \cdot \tau}, \tag{17}$$

where the fire emission E_{fire} (in g) are integrated on the Meso-NH grid-mesh, on a period of 2 min, τ. The fire emission is obtained through the equation of Seiler and Crutzen (1980):

$$E_{fire} = S_b \cdot FL \cdot \beta \cdot EF_{fire}, \tag{18}$$

where FL ($kg\,m^{-2}$) is the fuel loading, β (%) is the burning efficiency of the above-ground biomass, and EF_{fire} ($g\,kg^{-1}$) is the emission factor for the fire tracer. From Miranda et al. (2008), FL and β for shrub-lands are $1.00\,kg\,m^{-2}$ and 80 %, respectively.

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The fire tracer is handled as a PM_{10} aerosol ($EF_{fire} = 10\,g\,kg^{-1}$) with no mass (i.e. deposition velocity equals $0\,m\,s^{-1}$).

Table 1 summarizes the fire characteristics used for simulating different fire episodes (for references: Freitas et al., 2007; Silvani and Morandini, 2009; Miranda et al., 2008).
5 The same fuel moisture is selected for Mediterranean and Amazonian vegetation; this choice is justified by the necessity to limit the number of varying parameters between different scenarios. Moreover, recent discussions on the role of fuel moisture state a minor importance of this parameter on fire evolution compared to contributions from the environment in terms of humid air (Luderer et al., 2006; Strada et al., 2012).

10 **3.2 The 1-D PRM model**

Starting from the simple 1-D time-dependent cloud resolving model of Latham (1994), Freitas et al. (2006) proposed a plume rise model for simulating explicitly strong updrafts associated with vegetation fires and for, finally, predicting the fire injection height. The model governing equation for vertical motion includes the entrainment of environ-
15 mental air in the plume, the difference of temperature between the environment and the plume, the upward drag of condensate water vapour, the pyro-convection (i.e. release of latent heat when forming ice), and the effect of horizontal ambient wind (Freitas et al., 2010). The scope of the 1-D PRM model is to make a parametrisation available to 3-D meso-scale or global models in order to describe the sub-grid convective trans-
20 port associated with wild-fires, taking into account fire features, and to better forecast dispersion of fire products (aerosols and trace gases).

The 1-D plume rise model can be embedded in each column of a large-scale atmospheric-chemistry transport model. The coupling between the 1-D PRM and the host model relies on the assumption that at rough resolution (grid-scale ∼ 30 to 100 km)
25 fires do not have significant effects on the dynamics and the thermodynamics of the host model. In this way, the 3-D model passes the environmental large-scale conditions to the 1-D PRM model for initializing and constraining it at the boundaries, under

the hypothesis of a stationary atmosphere. Once the convective energy flux of the fire and the plume radius have been selected, the 1-D PRM model resolves explicitly the vertical extent of the fire plume. For each biome type two values for the fire heat flux are given: a lower and a upper; therefore, the 1-D PRM model computes a lower and a upper injection height. These two results are returned to the host model that homogeneously releases fire tracers emitted during the flaming phase in the vertical range delimited by the lower and the upper height. The 1-D PRM model can be run independently with initial values from a radiosonde.

The fire heat flux is converted into the available convective energy flux E (in $kW\,m^{-2}$) multiplying it by the reducing factor $C = 0.55$, already defined for the Meso-NH model (Sect. 3.1.3). Hence, the buoyancy flux ($m^4\,s^{-3}$) generated at the surface by the fire source is calculated using the following expression:

$$F = \frac{g\mathcal{R}}{c_p P_e} E R^2, \tag{19}$$

where $\mathcal{R}$ is the ideal gas constant ($kg^{-1}\,K^{-1}$), c_p is the specific heat capacity at constant pressure ($J\,kg^{-1}\,K^{-1}$), P_e is the ambient surface pressure (hPa) and R is the plume radius (m), computed at the surface by assuming the burning area as a circle.

Buoyancy triggers the vertical velocity ($w_{f,0}$) and the temperature excess ($T_f - T_e$) of the in-cloud air parcels at a virtual altitude (Morton et al., 1956):

$$w_{f,0} = \frac{5}{6\alpha} \left(\frac{0.9\alpha F}{z_v} \right)^{1/3}, \tag{20}$$

$$\frac{\Delta\rho_0}{\rho_{e,0}} = \frac{5}{6\alpha} \frac{F}{g} \frac{z_v^{-5/3}}{(0.9\alpha F)^{1/3}}, \tag{21}$$

$$T_f = \frac{T_e}{1 - \frac{\Delta\rho}{\rho}}, \tag{22}$$

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where $\alpha = 0.05$, $z_v = (5/6)\alpha^{-1}R$ is the virtual boundary height, and $\Delta\rho_0$ is the density difference between the in-cloud air parcels and environmental air at the surface. The surface water vapour excess is calculated in the same way that is reported for the Meso-NH model (Eq. 16). The heating rate increases following a step function during the first five minutes of the simulation. The time integration is fixed to one hour, as for Meso-NH, even if the steady state is typically reached within 50 min (Freitas et al., 2007). Hereafter, the subscript f is used to identify variables associated with the center of mass of the rising plume.

The 1-D PRM model depicts the evolution of the plume utilizing advection equations for the vertical velocity w_f, the temperature T_f, the water phases parameters $r_{f,v}$, $r_{f,c}$ and $r_{f,\text{ice-rain}}$, the horizontal velocity of the center of mass of the plume at level z (U_f) and the plume radius R. The governing prognostic equations are:

$$\frac{\partial w_f}{\partial t} + w_f \frac{\partial w_f}{\partial z} = \frac{1}{1+\gamma} g B_f - (\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}}) w_f \tag{23}$$

$$\frac{\partial T_f}{\partial t} + w_f \frac{\partial T_f}{\partial z} = -w_f \frac{g}{c_f} - (\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}})(T_f - T_e) + \left(\frac{\partial T_f}{\partial t}\right)_{\mu p} \tag{24}$$

$$\frac{\partial r_{v,f}}{\partial t} + w_f \frac{\partial r_{v,f}}{\partial z} = -(\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}})(r_{v,f} - r_{v,e}) + \left(\frac{\partial r_{v,f}}{\partial t}\right)_{\mu p} \tag{25}$$

$$\frac{\partial r_{c,f}}{\partial t} + w_f \frac{\partial r_{c,f}}{\partial z} = -(\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}}) r_{c,f} + \left(\frac{\partial r_{c,f}}{\partial t}\right)_{\mu p} \tag{26}$$

$$\frac{\partial r_{j,f}}{\partial t} + w_f \frac{\partial r_{j,f}}{\partial z} = -(\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}}) r_{j,f} + \left(\frac{\partial r_{j,f}}{\partial t}\right)_{\mu p} + \text{sedim}_j; \quad j \in \{\text{ice, rain}\} \tag{27}$$

$$\frac{\partial U_f}{\partial t} + w_f \frac{\partial U_f}{\partial z} = -(\varepsilon_{f,\text{lat}} + \varepsilon_{f,\text{dyn}})(U_f - U_e) \tag{28}$$

$$\frac{\partial R}{\partial t} + w_f \frac{\partial R}{\partial z} = \left(\frac{3}{5}\varepsilon_{f,\text{lat}} + \frac{1}{2}\varepsilon_{f,\text{dyn}}\right) R, \tag{29}$$

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where in Eq. (23) $\gamma = 0.5$ compensates for the neglect of non-hydrostatic pressure perturbations (Simpson and Wiggert, 1969) and B_f is the buoyancy term. Compared to the EDMF parametrisation, in the PRM the ice phase, that drives pyro-convection, is taken into account. Furthermore, the buoyancy term includes directly the downward drag of non-precipitating hydrometeors (ice, cloud droplet and hail) and its definition is related to the virtual temperature T_V. Index μp denotes the tendencies from cloud microphysics (Freitas et al., 2007). The plume top is identified by means of the in-cloud vertical velocity: when w_f is less than $1\,\mathrm{m\,s^{-1}}$ over a certain time, the steady state solution is attained and the top of the plume is given by the model.

In the new version of the 1-D PRM model two terms of entrainment have been defined, the classical lateral entrainment

$$\varepsilon_{f,lat} = \frac{2\alpha}{R}|w_f|, \tag{30}$$

and the "dynamic entrainment"

$$\varepsilon_{f,dyn} = \frac{2}{\pi R}(U_e - U_f). \tag{31}$$

This additional entrainment term expresses some physical effects on the plume that are enhanced by strong horizontal winds: the reduction of the in-plume vertical velocity $(-\varepsilon_{f,dyn}w_p)$, the decrease of the buoyancy term due to the loss of temperature excess $(-\varepsilon_{f,dyn}[T_f - T_e])$ and the gain of horizontal velocity of the plume $(\varepsilon_{f,dyn}[U_f - U_e]^2)$ (Freitas et al., 2010).

In the last update of the 1-D PRM model, Freitas et al. (2010) introduced the Vertical Mass Distribution (VMD) to mathematically define an injection layer. The VMD provides a probability vertical mass distribution as a function of the simulated vertical velocity profile (w_f). Premising that the main detrainment mass layer of cumulus convection is situated close to the cloud top, two levels are defined: z_i where w_f starts to decrease, and z_f where w_f is less than $1\,\mathrm{m\,s^{-1}}$. The area included among is the upper half part of

the cumulus (plume). Afterwards, a parabolic function of the height z with roots z_i and z_f is defined. Finally, the function is normalized to 1 in the interval $[z_i, z_f]$.

It is important to highlight that the 1-D PRM model is not a SCM as the 1-D Meso-NH model. In PRM, the evolution of the vertical structure of the prognostic meteorological variables within the plume does not impact the dynamics and the thermodynamics of the surrounding (grid-scale) environment in the host model. The 1-D PRM model is the adaptation of a cloud resolving model, a numerical model conceived to resolve cloud-scale circulation; therefore, it integrates over the whole cloud-area. Once adapted to the study of fire plume, the cloud-area is identified with the burnt area. In the present study, the prescribed burning area corresponds to the whole area that burned during the time integration: 100 ha (Sect. 3.1.3).

4   Results and analysis

In this section, results obtained from the 1-D Meso-NH/EDMF and the 1-D PRM model are presented and analysed, firstly by defining a unifying metrics for the discussion of results (Sect. 4.1), afterward by examining separately the three wild-fire episodes (Sect. 4.2).

4.1   Definition of the metrics used for the comparison

The EDMF and PRM schemes present a strong conceptual difference. The EDMF parametrisation is tightly coupled to the evolution of the grid-scale dynamics and thermodynamics variables through the entrainment and detrainment processes. The EDMF approach recalculates a new vertical distribution of state variables including tracers at each time step of the model in response to the perturbed grid-scale environment. The 1-D PRM model follows a diagnostic approach, in response to a stationary environment, where the fire plume rising stops evolving when all energy from fire characteristics and stationary atmospheric conditions has been used. This conceptual difference

entails a distinct method for simulating fire episodes: Meso-NH/EDMF burnt 100 ha in 1 h, while PRM burnt 100 ha for 1 h. Due to these conceptual differences, the time evolution of the injection height during the simulation time can not be compared between the two models. Rather, the discussion focuses on the injection heights at the end of the simulated time.

Table 2 gathers all variables that have been selected for the comparison between the two numerical models. Regarding these variables, it is important to define a common metrics. First of all, the Meso-NH model distinguishes between grid and updraft variables when the EDMF scheme is activated, otherwise only grid variables are available in the model; on the contrary, the PRM variables only refer to the updraft system. Figure 4 shows several parameters from the 1-D PRM model for the Lançon-de-Provence case. Figure 6 presents the time evolution of the vertical profiles of parameters for the 1-D Meso-NH simulation on the same case. These two figures will be used to illustrate the variables chosen for the comparison study.

- *Water vapour mixing ratio.* In both models, r_v is expressed in $g\,kg^{-1}$. In the PRM model, the water vapour mixing ratio is representative of the updraft ($r_{v,f}$). For the Meso-NH model, $r_{v,env}$ is the value in the environment. It is worth noting that the net contribution of wild-fires is to increase the humidity of the air near the detrainment levels (Figs. 4a and 6a). This moist enrichment of the atmosphere determines the elevation of the vertical level where r_v reduces and keeps constant (i.e. higher cloud base compared with ambient conditions Luderer et al., 2009) along the total integration time (see Figs. 4a and 6a).

- *Updraft vertical velocity.* In both models, vertical velocity is expressed in $m\,s^{-1}$, but different trends are identified due to different theoretical definitions. In the PRM model w_f is initialised at the surface (Eq. 20) by the buoyancy flux and decreases along the vertical (Fig. 4b); whereas in the Meso-NH model, w_u is a prognostic variable whose trend strongly depends on the updraft fraction a_u and the updraft

mass flux M_u behaviour:

$$w_u = \frac{M_u}{\rho a_u},$$

where a_u increases near the surface then diminishes along the vertical, while M_u maximises in the mixing layer (Fig. 6b).

– *Buoyancy.* The buoyancy acceleration is given in $m\,s^{-2}$, and it shows similar trends and values for both models (Figs. 4c and 6c). In general, the buoyancy acceleration slightly increases near the surface, where the heat source is active and feeds the rising of the in-cloud parcels, then it reduces until the sign inversion, when downdraft movements start. Near the cloud top, equilibrium is attained (buoyancy equals zero).

– *Turbulent parameters.* Updraft turbulent fluxes ($\langle w_u'\theta_u'\rangle$ in $K\,m\,s^{-1}$, and $\langle w_u'r_{v,u}'\rangle$ in $kg\,kg^{-1}\,m\,s^{-1}$) and turbulent kinetic energy (in $m^2\,s^{-2}$) are only available for the Meso-NH model. The updraft turbulent kinetic sensible heat flux $\langle w_u'\theta_u'\rangle$ represents the buoyancy source in the equation of the TKE (Eq. 5), therefore its trend nearly resembles that of buoyancy with a deeper increment near the surface heating (in the first kilometres of the atmosphere), followed by a faster decrease until the equilibrium is reached ($\langle w_u'\theta_u'\rangle = 0$, Fig. 6d). The temporal evolution of the updraft turbulent kinetic latent heat flux $\langle w_u'r_{v,u}'\rangle$ shows a clear rising of the altitude at which the maximum is placed that finally matches with the location of maximal fluctuations of w_u and $r_{v,env}$ in the upper part of the updraft (Fig. 6e).

– *Entrainment.* In Meso-NH, the entrainment rate ε_u is measured in m^{-1} (Eq. 12), whereas in the PRM model the two entrainment rates are both expressed in s^{-1} (Eqs. 30 and 31). For consistency, the entrainment fluxes of Meso-NH/EDMF ($E_u = \varepsilon M_u$, in $kg\,m^{-3}\,s^{-1}$) are multiplied by the density of dry air (ρ, in $kg\,m^{-3}$) in order to have a common metrics for the entrainment coefficients: s^{-1}. Although

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the definitions are slightly different, the lateral entrainment $\varepsilon_{f,lat}$ of the PRM model (Eq. 30) can be compared to the entrainment rate in the EDMF scheme of Meso-NH, Eq. (12): their values and exponential trends are similar (Figs. 4f and 6f). In the PRM model, $\varepsilon_{f,lat}$ has the same value at the surface through the whole set of simulations (between 0.3×10^{-2} and $0.4 \times 10^{-2}\,\mathrm{s}^{-1}$), likely due to the huge burnt area (100 ha) that homogenizes the existing differences in terms of fire forcings between Mediterranean and tropical fires (role of the plume radius in Eqs. 19 and 30). In the Meso-NH model, the entrainment coefficient at the surface has a constant value of $0.2 \times 10^{-2}\,\mathrm{s}^{-1}$ that is comparable with the lateral entrainment in PRM.

Concerning the dynamic entrainment $\varepsilon_{f,dyn}$ of the PRM model (Eq. 31) a similar parameter is not available in the Meso-NH model, therefore we decided to show in the following the sum of the two entrainment terms in order to illustrate the total entrainment of ambient air that feed, or slow down, the rising of the fire plume in the PRM model. Changes in the vertical profile of $\varepsilon_{f,dyn}$ are driven by the fluctuations of environmental wind. In Fig. 4d, $\varepsilon_{f,dyn}$ has a quasi monotonic decrease compatible with the wind profile in Fig. 1d.

– *Detrainment.* As done for the entrainment rate, the Meso-NH detrainment coefficient δ_u is converted from m^{-1} to s^{-1}. The trend of δ_u points out the coexistence of entrainment/detrainment in the CBL that both feed the vertical evolution of the mass flux; when ε_u goes to zero, δ_u maximises (Fig. 6f). In Meso-NH, we define the detrainement zone (or injection layer) as the vertical range where δ_u maximises. The injection height is the altitude at which the detrainment is maximal. Since the PRM model does not have a detrainment rate or zone among its output variables, the Meso-NH detrainment layer is compared to the vertical range enclosed by the VMD in the 1-D PRM simulations. The injection height is identified as the altitude where VMD maximises in the PRM model (Fig. 4f). In the graphics, the injection layers are compared by overlaying the VMD for the steady

state solution of the PRM (dot filled area) on the plot of the Meso-NH detrainment rate (Fig. 6f). The mathematical definition of the VMD implies that the PRM model vents the fire products (from the flaming phase) away from the surface inducing a depletion of the lower levels of the atmosphere in terms of fire pollutants (Fig. 4f) which are no more available for turbulent mixing. In the Meso-NH/EDMF model, a part of the released fire tracer is mixed in the first kilometres of the atmosphere by the turbulence, the rest is vertically transported by the thermal plume (Fig. 6f).

– *Scalar.* As explained before (Sect. 3.1.3), a fire tracer is released in the Meso-NH simulations. Its mixing ratio is normalized by its maximal value at each temporal session. In general, once released at the surface, the fire tracer is partly transported high in the atmosphere and released near the top of the updraft (Fig. 6d) showing a characteristic "C-shape" profile.

4.2 Comparison of fire forced simulations

Table 3 recapitulates simulations that are here shown: RSOU stands for simulations forced by radiosonde data, ECMWF means that the meteorological forcing is taken from the ECMWF re-analyses. For the 1-D PRM model, only results obtained considering the environmental wind drag are presented since the difference between a simulation with the wind effect on/off was already discussed in the work of Freitas et al. (2010). The top of the fire plume as predicted by the PRM is illustrated on all graphics by an horizontal solid line, the horizontal dashed line refers to the plume top when the environmental wind effect is off in the PRM model.

For the 1-D PRM model, the vertical profiles obtained after 10 min of simulation are drawn with a dashed line, while the solid line depicts the attained steady state solution. Results from the 1-D Meso-NH model are presented at the temporal session of the PRM model steady state solution (dashed line) and at the end of the simulation (after 60 min, solid line).

4.3 Lançon 2005 wild-fire

4.3.1 1-D PRM results

Using the radiosounding of Nîmes as meteorological forcing, the 1-D PRM model predicts a plume top near 2.5 km including the environmental wind effect (Fig. 4b); looking at the VMD, the main injection layer is localized between 1 and 2.5 km and it maximises near 1.7 km (Fig. 4f). This steady state solution is obtained after around 20 min. Above 1.5 km of altitude, $r_{v,f}$ records a gain of 1 g kg^{-1} between the first output (after 10 min) and the steady state solution (Fig. 4a). This increase in $r_{v,f}$ documents a growth of humidity inside the fire plume that come from combustion itself (fuel and combustion moisture) and from the entrainment of ambient air.

Forcing the 1-D PRM model by the ECMWF re-analyses, the forecast plume top is located at 3 km (wind on) after nearly 20 min of simulation. The main injection layer reaches out nearly 1.5 km, from 1.3 to 3 km, with its maximum attained around 2.2 km (Fig. 5c). Compared with values obtained using the radiosonde forcing, the plume top is 500 m higher, the injection layer has the same width with an injection height 700 m higher than the RSOU case. Diverse factors may lead to these differences. As observed in Sect. 2, the radiosounding has a stronger temperature inversion than the ECMWF atmospheric profile, an atmospheric parameter that may efficiently control the injection height of the fire plume, preventing it from reaching higher altitudes (Trentmann et al., 2003). Moreover, the radiosonde measured a stronger wind velocity at the surface: this results in a quasi-doubled dynamic entrainment ($e_{f,dyn} = 0.011\,\text{s}^{-1}$ in the RSOU case versus $e_{f,dyn} = 0.006\,\text{s}^{-1}$ using the ECMWF forcing, not shown), and in a faster decrease of B_f and w_f (not shown for the ECMWF case). In the first kilometres of the atmosphere, in the ECMWF forced simulation, B_f even increases before starting to diminish (Fig. 4c versus Fig. 5a). The weaker wind drag in the ECMWF simulation has a consequence also on $r_{v,f}$; since the fire plume can rise higher and faster in the ECMWF case, the growth of water vapour mixing ratio within the plume begins higher

than the radiosonde case and it is even more significant: 3 km above the surface $r_{v,f}$ increases of $3\,g\,kg^{-1}$ during 10 min (not shown).

4.3.2 1-D Meso-NH/EDMF results

When forced by the radiosounding of Nîmes, after 60 min of simulation, the 1-D Meso-
NH/EDMF model simulates an updraft that has its top near 3.8 km (Fig. 6b). This al-
titude is comparable to the plume top predicted by the 1-D PRM model without the
wind drag (3.5 km, horizontal dashed line on Fig. 6b). The detrainment zone is local-
ized between 2.7 and 3.7 km (Fig. 6f), above the turbulent mixing stops to be active
(Fig. 6d–e), and the detrainment maximises at 3.3 km. The final injection height as pre-
dicted by Meso-NH is 1.6 km higher than the value simulated by the 1-D PRM model
using the RSOU forcing. This difference reduces ($< 1\,km$) when we consider Meso-
NH outputs after 20 min (PRM steady state solution). The illustrated dissimilarity can
be ascribed to the weaker entrainment rate at the surface in the 1-D Meso-NH/EDMF
model: $\varepsilon_u = 0.002\,s^{-1}$ (Fig. 6f, positive values) against $0.004\,s^{-1}$ for $\varepsilon_{f,lat}$ in the respec-
tive 1-D PRM simulation (Fig. 4e). The Meso-NH updraft entrains less environmental
air, therefore the cooling of the fire plume due to the mixing with ambient air is weaker
resulting in a more convective updraft than the PRM plume. The same case study has
been simulated by the 1-D Meso-NH without the EDMF parametrisation. The vertical
spread is weaker because only the local mixing due to turbulence is considered: after
60 min, the TKE falls to zero at 3.1 km (Fig. 7b), as a consequence the fire tracer is not
transported higher than this level (Fig. 7c). Even if a detrainment zone is not defined for
this case, the relative VMD overlays the normalized vertical profile of the tracer mixing
ratio just below the level where the diminution of the scalar mixing ratio becomes faster
(2.6 km on Fig. 7c). The comparison between Fig. 6 (with EDMF) and Fig. 7 (without
EDMF) shows the contribution of the mass flux approach which transport efficiently the
boundary layer products to the higher altitudes.

Once forced by the ECMWF profile, at the end of the simulation, the updraft in the 1-D Meso-NH/EDMF model stops rising at around 3.2 km, and it detrains between 2.2 and 3.2 km with its maximum at 2.7 km (Fig. 8b). In this case, the comparison of the injection layer between the two models is better than for the RSOU forcing with a difference of only 500 m. For both models, the buoyancy acceleration starts reducing around 2 km of altitude (Fig. 8c versus Fig. 5a). The Meso-NH entrainment rate (Fig. 5b) is still the half than the PRM lateral entrainment (not shown); considering the total entrainment for the PRM model (Fig. 5b), Meso-NH nearly entrains one fourth of the ambient air compared to PRM (Fig. 5b). As before, the Meso-NH simulation without the EDMF scheme predicts a lower injection height (2.5 km after 60 min of integration time, not shown).

4.3.3 General comments

The 1-D PRM model predicts a higher injection height (by nearly 500 m) when it is forced by the ECMWF re-analysis because this meteorological forcing has weaker winds at the surface than the radiosonde. Using the 1-D Meso-NH/EDMF model, the result is opposite with nearly the same variability (600 m): the final injection height is higher for the RSOU case rather than the ECMWF case. The 1-D PRM would pre-scribe to a host CTM a fire injection height that represents a fire plume escaping the PBL. The measurements recorded downwind of the Lançon fire by the air quality mon-itoring network gave evidences of a fire plume kept in contact with the surface (Strada et al., 2012). When the EDMF parametrisation is not activated, the 1-D Meso-NH model predicts a lower plume top that is more comparable with the air quality observations reported for the Lançon fire in the work of Strada et al. (2012).

4.4 Rondonia 2002 wild-fire: calm-dry case

4.4.1 1-D PRM results

Once initialised by the radiosonde, the 1-D PRM predicts a plume top at 6.9 km (wind on) with a maximum of 4 % of mass around 5 km and the main injection layer com-
prised between 3.3 and 6.9 km (Fig. 9c). This maximum is substantially lower than the 9 % maximum obtained for the Lançon-de-Provence fire but the mass is distributed in a deeper layer than the Mediterranean case. The 1-D PRM steady state solution is obtained after around 30 min. At the surface, the buoyancy acceleration is $0.03\,\mathrm{m\,s^{-2}}$ stronger than for the Lançon fire (Fig. 9a). In Eq. (19) the convective energy flux E
has diminished and the plume radius R keeps the same, therefore this difference can only be attributed to the lower ambient surface pressure P_e recorded in Rondônia. The dynamic entrainment is critically reduced compared to the Lançon case, leading to a reduction in the total entrainment (Fig. 9b). This is due to the non negligible variation in the ambient wind speed between the Mediterranean and the Amazonian atmospheric
background (Sect. 2; Fig. 1 versus Fig. 2).

The 1-D PRM simulation forced with the ECMWF profile has a 700 m lower plume top (6.2 km, when including the wind drag) after around 40 min, and the injection layer is 600 m less deeper (from 3 to 6 km) with a maximum of 5 % of mass around 4.8 km (Fig. 10c). At 1 km of altitude, the dynamic entrainment maximises using both forc-
ings but with different values: $0.3 \times 10^{-2}\,\mathrm{s^{-1}}$ for the RSOU case, $0.1 \times 10^{-2}\,\mathrm{s^{-1}}$ for the ECMWF case. The influence of this dissimilar contribution is evident on the vertical profile of the total entrainment that decreases quasi monotonically along the vertical (Fig. 10b) compared to the more contrasted profile of the RSOU case (Fig. 9b). These trends reflect contrasts in terms of wind speed between the two forcings as highlighted
in Sect. 2.

4.4.2 1-D Meso-NH/EDMF results

The 1-D Meso-NH/EDMF simulation forced by the radiosounding reproduces an updraft top at 4.5 km. The detrainment zone is localized between 3 and 4.5 km, just overlaying the lower one third of the PRM injection layer (Fig. 11b). The injection height is 1.5 km

5 lower than the value predicted by the 1-D PRM model. Although the well developed daytime mixed layer (Fig. 2b–c) and the surface heating associated with the fire, the buoyancy acceleration at the surface is nearly the half of the same parameter in the PRM model (Fig. 11a versus Fig. 9a). In the Meso-NH model, the influence of the ambient surface pressure seems not to be accounted for as observed for the 1-D PRM

10 model.

Using the ECMWF forcing, the convective updraft in the 1-D Meso-NH/EDMF model rises up to around 4.5 km and it detrains between 2.2 and 3.5 km, maximising at 2.7 km (Fig. 12b). Also in this case, the comparison of the injection layer is unsatisfactory with a gap of around 2 km. The less turbulent atmospheric background reproduced by the

15 ECMWF re-analyses (Fig. 2b–c) determines weaker turbulent kinetic heat fluxes than the Meso-NH simulation forced by the radiosounding (not shown).

4.4.3 General comments

In contrast with general conclusions drawn for the Lançon fire, the 1-D PRM model predicts a higher injection height when forced by the radiosounding: even if the radiosonde

20 recorded a stronger wind speed at the surface than the ECMWF re-analyses (hence a stronger dynamic entrainment), a stronger buoyancy acceleration is produced at the surface. Using the 1-D Meso-NH/EDMF model, the updraft stops rising at nearly the same altitude (4.5 km) for both forcings, while the detrainment zone and its maximun are located at different levels: 800 m higher when the environment is more turbulent

25 (RSOU forcing). Concerning the comparison of the fire injection height as predicted by the two numerical models, the result is unsatisfactory: the Meso-NH/EDMF model is

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between 1.1 and 1.5 km lower than the PRM model. The main injection layers predicted by the two models partly overlay only for the case study initialised by the radiosounding.

4.5 Rondonia 2002 wild-fire: windy-wet case

4.5.1 1-D PRM results

Forced by the radiosonde, the 1-D PRM model simulates a plume top at 6.7 km (wind on) after around 50 min (steady state solution). The main injection layer is localized between 3.2 and 6.7 km and the injection height is around 5 km (Fig. 13c). The dynamic entrainment is $0.1 \times 10^2 \, s^{-1}$ stronger than the calm-dry case implying a similar discrepancy in the total entrainment (Fig. 13b versus Fig. 9b). At the surface, the buoyancy acceleration is about $0.06 \, m \, s^{-2}$ (Fig. 13a), slightly smaller than the Rondônia calm-dry case forced by the radiosounding (Fig. 9a).

In Fig. 14, the 1-D PRM simulation initialised by the ECMWF profile has a fire plume that rises up to 8.9 km (wind on) after the whole integration time (60 min). The meteorological scenario prescribed by the ECMWF profile offers favourable conditions for the plume rise: weak ambient winds (lower values for the total entrainment in Fig. 14b versus Fig. 13b) and high humidity in the air (Fig. 3c). The main injection layer reaches out 4.7 km, from 4.2 to 8.9 km, and 3 % of mass are injected at 6.5 km (Fig. 14c).

4.5.2 1-D Meso-NH/EDMF results

When forced by the radiosounding, the 1-D Meso-NH/EDMF model simulates an updraft top at 4.5 km (15). Once again, the 1-D Meso-NH/EDMF plume top is notably lower than the one predicted by the 1-D PRM model using the radiosonde. The detrainment zone is localized between 2.5 and 4.5 km and it shows two maxima: at 3 and 4.1 km (Fig. 15b). The higher maximum and the associated detrainment overlays the lower half injection layer simulated by the respective 1-D PRM simulation.

Figure 14 shows the results of the 1-D Meso-NH/EDMF forced by the ECMWF profile. The updrfat top is located at 5 km. The detrainment process presents two main zones of activity: a stronger one at 2 km, a weaker one at 4 km (Fig. 14b). This feature influences the vertical profile of the scalar mixing ratio: above the first detrainment zone, the fire tracer is rapidly transported along the vertical up to 5 km (Fig. 14c). The presence of two detrainment zones and of a deeper vertical transport above the turbulent mixing layer document the role of the mass flux approach in the EDMF scheme. The mass flux term is responsible for non-local mixing that feeds the convective tower along the whole column. Although the detrainment zone stretches from 1.6 to 4.8 km, it does not overlay the vertical range of the respective VMD (Fig. 14b).

4.5.3 General comments

Regarding the PRM model, there is a difference of 1.5 km between the RSOU and the ECMWF case. This gap confirms the sensitivity of the model of Freitas et al. (2010) to the humidity and the wind pattern of the meteorological background: the ECMWF profiles show an atmosphere moister and less windy than the radiosounding (Fig. 3c, d). As observed for the calm-dry case, the Meso-NH/EDMF results do not record sensible variations, probably due to intrinsic limitations of the EDMF scheme, as it will be discussed in the following. The 1-D Meso-NH/EDMF model is not able to rise higher than 5 km, therefore the comparison with the 1-D PRM model is less satisfactory than for the calm-dry case. As for the calm-dry case, the comparison between the two numerical models is unsatisfactory, slightly better when using the radiosonde as meteorological forcing rather than the ECMWF re-analyses.

5 General discussion

In this section differences between the two approaches and the sensitivity to the meteorological forcing are discussed.

Figure 17 summarizes the injection heights and layers as predicted by the PRM and the Meso-NH/EDMF models for the three documented fire cases. Looking at this graphic, at first glance an outstanding difference is observed between the Mediterranean (Lançon 2005) and the Amazonian wild-fires (Rondônia 2002): although the
5　fire-induced heat fluxes for the Mediterranean case have higher values than the Amazonian ones (Table 1), the windy and dry meteorological conditions of the Mediterranean Basin efficiently constrain the vertical development of the fire plume. This sensitivity is evident for the PRM model where the Mediterranean fire has an injection height that is 3–4 km lower than the Amazonian values, and the Mediterranean injection layer
10　is nearly the half of those obtained for the Amazonian cases. The difference Mediterranean/Amazon is less definite for the Meso-NH/EDMF model that predicts fire injection heights in a range between 2 and 4 km, with a quite similar width of the injection layer (except for the Rondônia windy-wet case where two well distinct detrainment zones are observed).
15　The kind of meteorological forcing also influences the evolution of the convective updrafts. For the 1-D PRM, the injection height records a variation in a range between 500 m and 1.5 km for the same wild-fire, whereas the extent of the injection layer is quite similar for both meteorological forcings, except for the Rondônia windy-wet case where the main injection layer is 1.2 km wider using the ECMWF re-analyses rather than the
20　radiosounding. The identified variabilities for the PRM model are highly influenced by the intensity of the ambient wind speed that determines, in turn, the intensity of the dynamic entrainment that governs the effectiveness of the wind drag; the humidity of the ambient air also plays an important role since, once moist air mixes with the updraft, the net result is to lighten the rising plume ($\rho_{dry} < \rho_{moist}$, e.g. Fig. 13 versus Fig. 14). For
25　the 1-D Meso-NH/EDMF model, the level of maximum detrainment can vary between 500 m and 1 km between the two forcings. The 1-D Meso-NH/EDMF model is partly influenced by the atmospheric conditions in terms of turbulence that locally feed the turbulent flux of conservative variables (Eq. 6): the less turbulent environment for the Amazonian cases depicted by the ECMWF re-analysed (Figs. 2b, c and 3b, c) leads to

lower injection heights. However, significant values observed for the updraft turbulent kinetic latent heat flux in the Rondônia calm-dry case do not lead to higher injection heights compared to other results. This model response is coherent with the thesis of Luderer et al. (2009) who state that the fire-released latent heat is of much lesser importance than the fire-released sensible heat.

Comparing results from the two numerical models, the Meso-NH/EDMF model simulates 0.5–1 km higher injection heights for the Mediterranean wild-fire, and 1.5–3.5 km lower values for the Amazonian cases. These gaps can be ascribed to the different intensity of the entrainment of ambient air in the two approaches. The Meso-NH model takes into account only the lateral entrainment, while the PRM model includes the effect of ambient wind among the environmental factors that may feed the lateral mixing of the rising plume. In the PRM model, this approach results in a total entrainment coefficient at the surface that is always the double of the Meso-NH value. As a consequences, the in-cloud parcels mix more efficiently with the ambient air in the PRM frame; hence, if the surrounding atmosphere is dry and windy (as for the Mediterranean fire), the net result is a drag force, while a humid and less windy atmosphere can feed the rising of the plume by triggering the pyro-convection (as for the Amazonian fires). These conclusions state the important role of ambient wind on the fire plume rise in contrast with the recent work of Sofiev et al. (2012). Both models used a total burnt area of 100 ha that burned in 1 h for Meso-NH, while for 1 h for PRM. This distinct design of simulations implied different fire fluxes at the surface for the two models with Meso-NH receiving 1/30 of the fire fluxes prescribed at the surface in PRM. In spite of this, the Meso-NH response is not systematically lower than PRM diagnostic.

For the Lançon fire, the two models simulate a fire injection height above the PBL, in contrast with the existing observations for the Lançon fire (Strada et al., 2012). For the Amazonian fires, there exist considerable differences between the two models. In particular, the 1-D Meso-NH/EDMF model simulates the injection heights in the lower half of the troposphere, below the zero isothermal (at about 5 km). This is an intrinsic limitation of the current version of EDMF in Meso-NH. The EDMF parametrisation was

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implemented in the Meso-NH model to reproduce strato-cumulus clouds. Its design implies some important features that strongly limit the vertical evolution of the updraft: (1) the altitude of the zero-isothermal is a vertical limit in the rising of the updraft, (2) the ice phase is not yet activated, and (3) the cloud layer can not exceed a fixed 3 km extent. Similarly, Rio et al. (2010) discussed the use of the EDMF scheme in configurations, such as wild-fire episodes, for which this parametrisation has not been initially developed for, possibly leading to deep convection.

In general, it is important to underline the intrinsic limitations of the design that has been chosen for the present study. The choice of a 1 km horizontal grid-mesh was justified by the aim to study the vertical evolution of a fire plume in the same config-uration of the Lançon-de-Provence 2005 case study (Strada et al., 2012). However, kilometric horizontal resolutions are intermediate scales for turbulence movements where these processes are not mainly resolved neither entirely parametrised. Honnert et al. (2011) investigated the behaviour of atmospheric models at intermediate scales (the so called "Terra Incognita" of the turbulence, Wyngaard, 2004) and they identified some misleading results of atmospheric models due to the presence of too many re-solved movements, when the turbulence scheme parametrises the subgrid thermal, or an overestimation of the subgrid part, when a mass-flux scheme is introduced. Con-cerning the PRM model, a burnt area that measures 100 ha induces a strong decrease of the entrainment of ambient air (see Eqs. 30 and 31), hence it leads to a fire plume that rises fast in the atmosphere and does not mix properly with the surrounding air. Utilizing radiosondes to force the PRM model a difference of only 200 m is observed between the calm-dry and the windy-wet case. While forcing the PRM model by the same radiosoundings and using a burnt area of 10 ha, Freitas et al. (2010) observed a difference of nearly 1 km in terms of injection height between the calm-dry and the windy-wet case; such a significant difference was not observed when the authors con-sidered a burnt area of 50 ha. In addition, at 1 km scale, the hypothesis of the grid size environmental air not impacted by the fire becomes questionable in the PRM model

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since the horizontal scale is comparable with the horizontal scale of global and meso-scale models.

6 Conclusions and perspectives

When modelling wild-fires, the height of injection of fire products has a crucial importance for determining the distance and the direction the smoke will travel (Guan et al., 2010), and also for understanding interactions between fire emissions and urban pollution (Singh et al., 2012). The parameter of the fire injection height highlights the tight link that exists between the dynamics and the chemistry of a wild-fire: it depends on fire characteristics and meteorological conditions, and it determines the chemistry that will act on the fire plume and on the environmental air. Nowadays, several studies have been carried out to define a database of seasonally and regionally diverse plume heights in order to prescribe, or just validate, the fire injection height in CTMs (Labonne and Chevallier, 2007; ValMartin et al., 2010; Gonzi and Palmer, 2010; Guan et al., 2010). Furthermore, some physical approaches have been developed to predict the fire plume rise, these go from simple (Miranda, 2004; Hodzic et al., 2007) to more complex schemes such as the plume rise models (Freitas et al., 2010; Sofiev et al., 2012) or EDMF approaches (Pergaud et al., 2009; Rio et al., 2010).

In this work, sensitivity tests have been realised to compare the fire plume top predicted by the 1-D PRM and the 1-D Meso-NH/EDMF models. Three wild-fires have been chosen: a Mediterranean arson fire and two deforestation Amazonian fires. They distinguish from one another in terms of fire features and meteorological scenarios. Moreover, for each case, two meteorological forcings have been used to initialise each model: a radiosounding and a vertical profile taken from ECMWF re-analyses.

The predicted injection heights showed considerable differences from one model to the other regarding the kind of fire and, for a given model, between ECMWF and radiosounding forcings. A definite sensitivity to the type of fire is observed with predicted injection heights strongly driven by distinct meteorological conditions in the Mediter-

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ranean and Amazonian Basin. For both models, sensitivity to meteorological forcing is less pronounced for the Mediterranean fire with a difference of 20–30 % for both models (using as reference an injection height of 1 km; Strada et al., 2012); a similar bias is estimated for Amazonian fires (using results from Freitas et al., 2010, as reference) except for the Rondônia windy-wet case for which a difference of ∼ 40 % is observed for PRM results. Compared to the 1-D Meso-NH/EDMF model, the 1-D PRM model simulates lower plumes for the Mediterranean case, higher for the Amazonian cases. The difference among models can attain 3–4 km for the Amazonian fires. The comparison for the Mediterranean fire gives injection heights between 2 and 3 km for both models. For the Lançon case, both models forecast a plume top above the BL, although there are evidences that the Lançon fire plume propagated near the surface (Strada et al., 2012).

The evolution of the fire plume in the 1-D PRM model seems to be mainly influenced by the ambient wind and humidity. There exists an important difference in the theoretical definitions of the two numerical models: the 1-D PRM model is designed to reach a steady state solution because it does not influence the dynamics and the thermodynamics of the environment; while the 1-D Meso-NH model acts as a single column model where the evolution of the convective updraft perturbs the atmosphere during the whole integration time. Moreover, the PRM model always predicts a fire plume that takes off from the ground. The PRM was designed to feed a 3-D host (CTM) model with the information of the fire injection height; hence, the prescribed elevation of the fire plume may lead to an artificial depletion of the mixing layer in the host (CTM) model in terms of fire products coming from the flaming phase.

Some limits have been identified for both models concerning the chosen configuration: they are forced in the "Terra Incognita" (Wyngaard, 2004; Honnert et al., 2011) of the turbulence, and at a horizontal resolution where the fire is supposed to impact the environmental air in global and meso-scale models. Actually, some developments are in progress concerning the activation of the ice phase in the EDMF frame (Riette, personal communication, 2013) to properly induce the transition from shallow to

deep convection. These developments are expected to significantly improve the accuracy of Meso-NH. Moreover, sensitivity tests have to be run on more documented fire episodes (e.g. the Quinault fire; Trentmann et al., 2003; Freitas et al., 2007), and at resolutions higher than kilometric horizontal scales in order to study the behaviour of Meso-NH when the atmospheric dynamics is fully resolved. Regarding the 1-D Plume Rise Model, perspectives rely on the numerical coupling between the model and the FRP information provided by MODIS, thought first studies (ValMartin et al., 2012) show that the model input are difficult to constrain with the current accuracy of the available data (i.e. MISR, BL estimation). However such a development is attractive, as this evolution of PRM could take advantage of the collaborative project about Monitoring Atmospheric Composition and Climate (MACC II) in the frame of which FRP observations from MODIS are assimilated in the Global Fire Assimilation System (GFAS, Kaiser et al., 2012). The coupling between the PRM model and satellite observations (e.g. FRP) would permit to provide to PRM more accurate information concerning fire characteristics. This technique could improve PRM predictions of fire injection heights that are passed to chemical transport models.

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Table 1. Fire and fuel characteristics for the considered case studies.

Fire features	Lançon 2005	Rondônia 2002 Calm-Dry	Windy-wet
Burnt Area (ha)	100	100	100
Sensible Heat Flux, ϕ_{sh} (kW m^{-2})	100.0	80.0	80.0
Heat Content, E_m (MJ kg^{-1})	19.6	15.5	15.5
Fuel Moisture, FM (%)	10	10	10
Water Emission Factor, EF_{H_2O} (kg kg^{-1})	0.5	0.5	0.5
Fire Tracer Flux, Φ_s (10^{-3} g m^{-2} s^{-1})	2.23	2.23	2.23

Table 2. Environmental and updraft variables selected in the Meso-NH and Plume Rise models for discussing the simulation results.

Variables	Meso-NH		PRM
	Grid	Updraft	
Water vapour mixing ratio ($g\,kg^{-1}$)	r_v		$r_{v,f}$
Vertical velocity ($m\,s^{-1}$)		w_u	w_f
Buoyancy ($m\,s^{-2}$)		B_u	B_f
Turbulent heat flux ($K\,m\,s^{-1}$)		$\langle w'_u \theta'_u \rangle$	
Turbulent moist flux ($kg\,kg^{-1}\,m\,s^{-1}$)		$\langle w'_u r'_{v,u} \rangle$	
Turbulent kinetic energy ($m^2\,s^{-2}$)	TKE		
Lateral entrainment rate (s^{-1})		ε_u	$\varepsilon_{f,lat}$
Dynamic entrainment rate (s^{-1})			$\varepsilon_{f,dyn}$
Detrainment rate (s^{-1})		δ_u	
Injection layer		$[\varepsilon_u]_{max}$	VMD (%)
Normalized scalar mixing ratio	r_s		

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Table 3. Summary of numerical experiments that were performed with the 1-D models (Meso-NH and Plume Rise Model) and discussed in the present study.

Fire episode		Atmospheric forcing	Meso-NH		PRM Environmental wind effect
			TURB	EDMF	
Lançon 2005		RSOU	On On	On Off	On
		ECMWF	On On	On Off	On
Rondônia 2002	Calm-dry	RSOU	On	On	On
		ECMWF	On	On	
	Windy-wet	RSOU	On	On	On
		ECMWF	On	On	
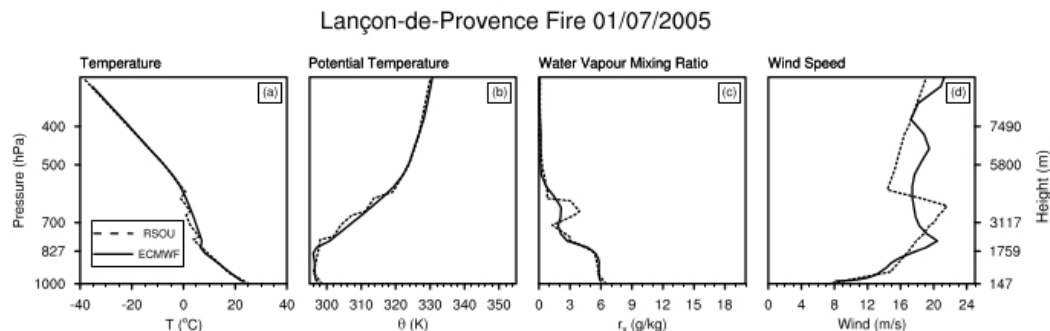
Fig. 1. Comparison between the radiosonde data (dashed line) and the analyses from the European Center for Meteorological and Weather Forecasting (solid line) for the Lançon-de-Provence 2005. The four graphics represent the vertical profile of: **(a)** temperature T (°C), **(b)** potential temperature θ (K), **(c)** water vapour mixing ratio r_v (g kg^{-1}) and **(d)** wind speed (m s^{-1}).
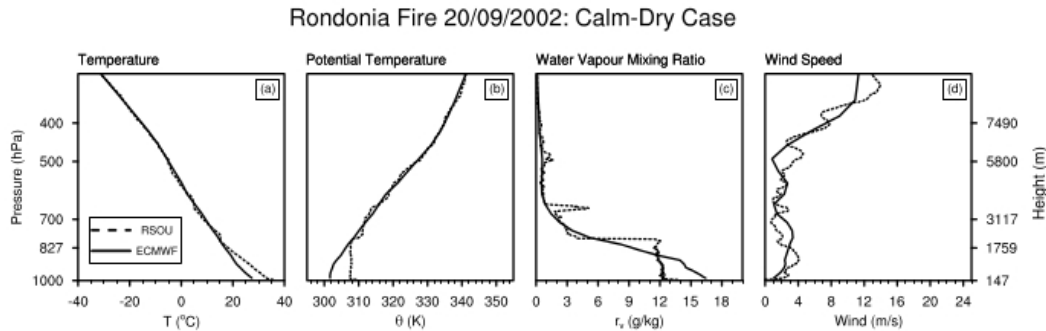
Fig. 2. The same as Fig. 1 for the calm-dry case of Rondônia 2002.
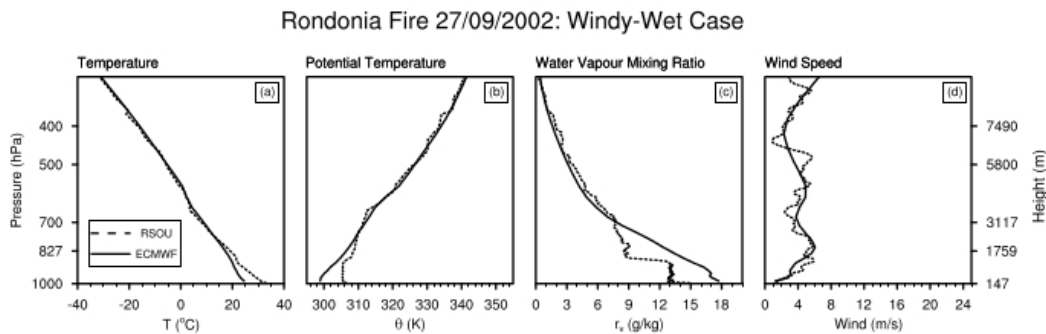
Fig. 3. The same as Fig. 1 for the windy-wet case of Rondônia 2002.

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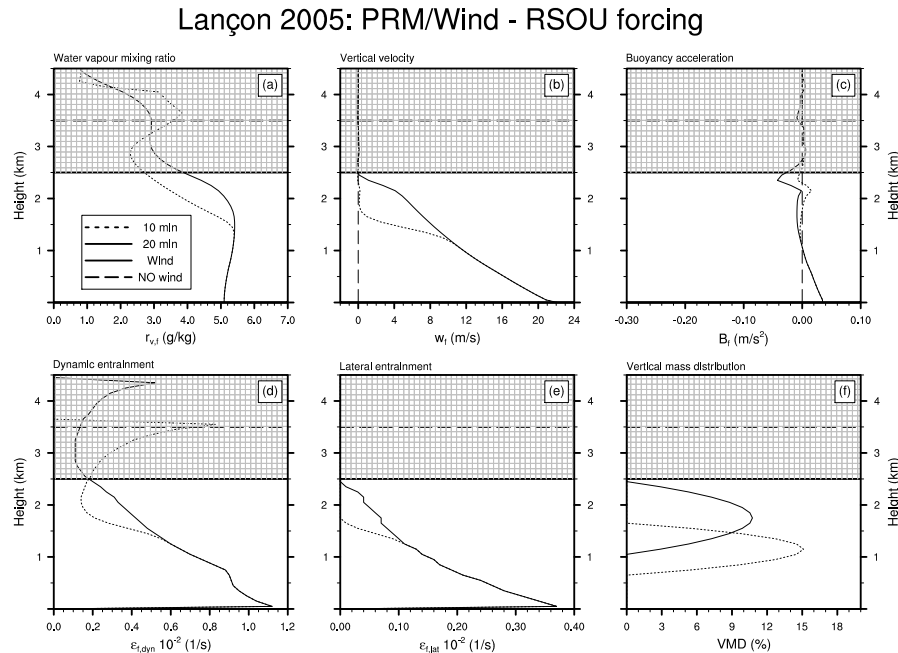
Fig. 4. 1-D Plume Rise Model results for the Lançon fire using the radiosounding of Nîmes as meteorological forcing. The quantities are: **(a)** water vapour mixing ratio ($r_{v,f}$, $g\,kg^{-1}$); **(b)** vertical velocity (w_f, $m\,s^{-1}$); **(c)** buoyancy acceleration (B_f, $m\,s^{-2}$); **(d)** dynamic entrainment ($\varepsilon_{f,dyn}$, s^{-1}); **(e)** lateral entrainment ($\varepsilon_{f,lat}$, s^{-1}); **(f)** vertical mass distribution (VMD, %). The solid line indicates the plume top obtained including the environmental wind effect; the horizontal dashed line is the plume top when the wind effect is off.
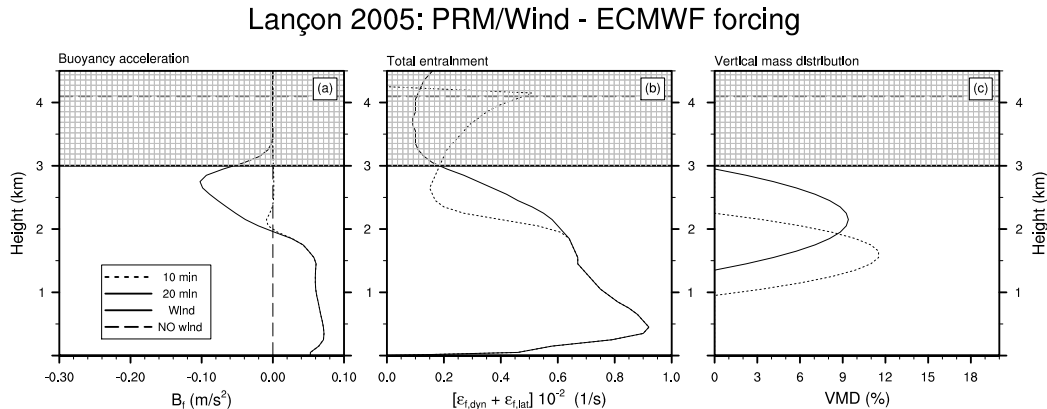
Fig. 5. 1-D Plume Rise Model results for the Lançon fire using the ECMWF re-analyses as meteorological forcing. The quantities are: **(a)** buoyancy acceleration (B_f, $m\,s^{-2}$); **(b)** total entrainment ($\varepsilon_{f,dyn} + \varepsilon_{f,lat}$, s^{-1}); **(c)** vertical mass distribution (VMD, %). The solid line indicates the plume top obtained including the environmental wind effect; the horizontal dashed line is the plume top when the wind effect is off.
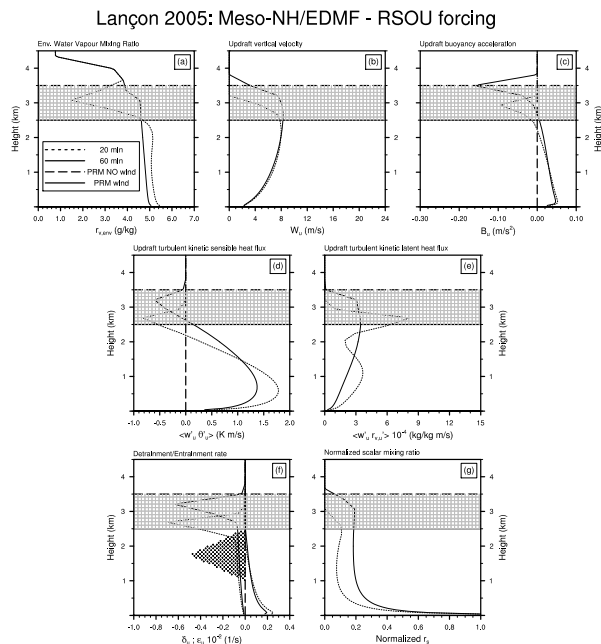
Fig. 6. 1-D Meso-NH/Eddy Diffusivity–Mass Flux results for the Lançon fire using the radiosounding of Nîmes as meteorological forcing. The quantities are: **(a)** environmental water vapour mixing ratio ($r_{v,env}$, $g\,kg^{-1}$); **(b)** updraft vertical velocity (w_u, $m\,s^{-1}$); **(c)** updraft buoyancy acceleration (B_u, $m\,s^{-2}$); **(d)** updraft turbulent kinetic sensible heat flux ($\langle w_u' \theta_u' \rangle$, $K\,m\,s^{-1}$); **(e)** updraft turbulent kinetic latent heat flux ($\langle w_u' r_{v,u}' \rangle$, $kg\,kg^{-1}\,m\,s^{-1}$); **(f)** detrainment rate when values are negative (δ_u, s^{-1}), entrainment rate for positive values (ε_u, s^{-1}); **(g)** normalized scalar flux. The horizontal solid (dashed) line is the plume top obtained with the environmental wind effect on (off) in the respective 1-D Plume Rise Model simulation. The dot filled area indicates where the VMD has positive values.
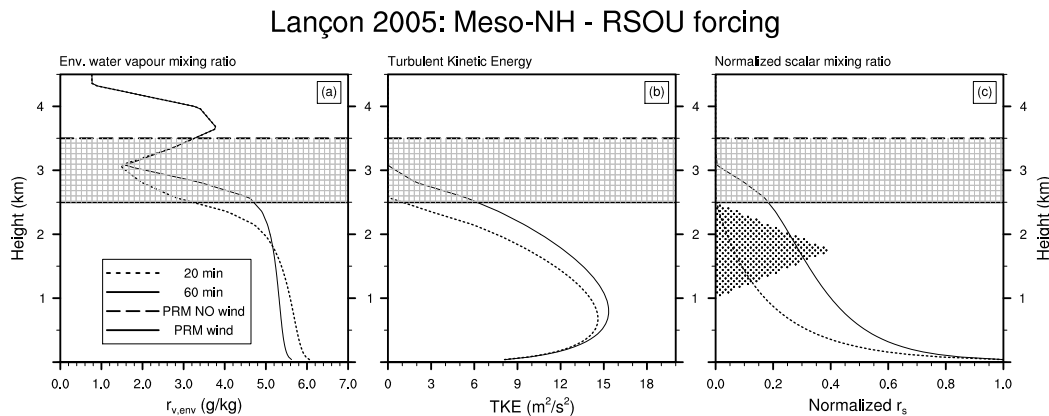
Fig. 7. 1-D Meso-NH results for the Lançon fire using the radiosounding of Nîmes as meteorological forcing, without activating the Eddy Diffusivity–Mass Flux scheme. The quantities are only referred to the environment: **(a)** environmental water vapour mixing ratio ($r_{v,env}$, $g\,kg^{-1}$); **(b)** turbulent kinetic energy (TKE, $m^2\,s^{-2}$); **(c)** normalized scalar mixing ratio. The horizontal solid (dashed) line is the plume top obtained with the environmental wind effect on (off) in the respective 1-D Plume Rise Model simulation. The dot filled area indicates where the VMD has positive values.
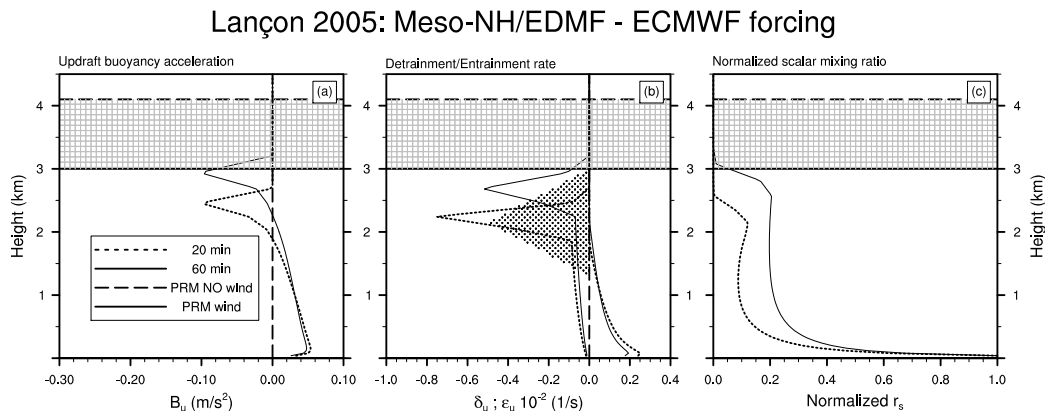
Fig. 8. 1-D Meso-NH/Eddy Diffusivity–Mass Flux results for the Lançon fire using the ECMWF re-analyses as meteorological forcing. The quantities are: **(a)** updraft buoyancy acceleration (B_u, ms^{-2}); **(b)** detrainment rate when values are negative (δ_u, s^{-1}), entrainment rate for positive values (ε_u, s^{-1}); **(c)** normalized scalar mixing ratio. The horizontal solid (dashed) line is the plume top obtained with the environmental wind effect on (off) in the respective 1-D Plume Rise Model simulation. The dot filled area indicates where the VMD has positive values.

Fig. 9. The same as Fig. 5 for the calm-dry case of Rondônia (20 September 2002) using the radiosounding as meteorological forcing.

Fig. 10. The same as Fig. 5 for the calm-dry case of Rondônia (20 September 2002) using the ECMWF re-analyses as meteorological forcing.

Fig. 11. The same as Fig. 8 for the calm-dry case of Rondônia (20 September 2002) using the radiosounding as meteorological forcing.

Fig. 12. The same as Fig. 8 for the calm-dry case of Rondônia fires (20 September 2002) using the ECMWF re-analyses as meteorological forcing.

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Fig. 13. The same as Fig. 5 for the windy-wet case of Rondônia fires (27 September 2002) using the radiosounding as meteorological forcing.

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Rondônia Windy-Wet: PRM/Wind -ECMWF forcing

Fig. 14. The same as Fig. 5 for the windy-wet case of Rondônia fires (27 September 2002) using the ECMWF re-analyses as meteorological forcing.

Fig. 15. The same as Fig. 8 for the windy-wet case of Rondônia fires (27 September 2002) using the radiosounding as meteorological forcing.

Rondônia Windy-Wet: Meso-NH/EDMF - ECMWF forcing

Fig. 16. The same as Fig. 8 for the windy-wet case of Rondônia fires (27 September 2002) using the ECMWF re-analyses as meteorological forcing.

Fig. 17. Fire injection height and plume base and top as predicted by the two numerical models. For each of the three wild-fires, the two numerical forcings are distinguished: RSOU for the radiosonde, ECMWF for the re-analyses. The blue dots show the steady state solution attained by the PRM model, and they correspond to the level of maximum VMD with the associated width of the injection layer (blue error bar). The red triangles represent the final result given by the Meso-NH/EDMF model (after 60 min of simulation): they correspond to the level of maximum detrainment with the associated extent of the injection layer (red error bar). Red crosses are the intermediate results of the Meso-NH/EDMF model at the time of the steady state solution of the PRM model.

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