Geoscientific
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The implementation of a MiXed Layer model (MXL, v1.0) for the dynamics of the atmospheric boundary layer in the Modular Earth Submodel System (MESSy)

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GMDD

7, 7197–7238, 2014

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Abstract

We present a new submodel for the Modular Earth Submodel System (MESSy): the MiXed Layer (MXL) model for the diurnal dynamics of the convective boundary layer, including explicit representations of entrainment and surface fluxes. Through the MESSy interface, MXL is coupled with modules that represent other processes relevant to chemistry in the atmospheric boundary layer (ABL). In combination, these provide a computationally inexpensive tool that is ideally suited for the analysis of field data, for evaluating new parametrizations for 3-D models, and for performing systematic sensitivity analyses. A case study for the DOMINO campaign in Southern Spain is shown to demonstrate the use and performance of MXL/MESSy in reproducing and analysing field observations.

1 Introduction

In atmospheric chemistry, various types of models are used for studies on different spatio-temporal scales, ranging from box models representing a single point in space (e.g. Sander et al., 2011) to regional and global 3-D models (e.g. Jöckel et al., 2006). Here, we describe the implementation of a MiXed Layer (MXL) submodel for the dynamics of the atmospheric boundary layer (ABL) in the Modular Earth Submodel System (MESSy, Jöckel et al., 2010). MXL represents the dynamics of the convective boundary layer at diurnal time-scales with two vertically changing layers, one for the ABL and one for the free troposphere (FT). In combination with MESSy submodels that represent the other processes relevant to atmospheric chemistry, it provides a missing link between box and 3-D atmospheric chemistry models, at a spatio-temporal scale typical for measurements of atmospheric chemistry. It is especially suited for studying the diurnal evolution of chemical species that have chemical lifetimes similar to the time scales of atmospheric mixing, so for which the effects of chemistry and dynamics should be studied simultaneously (Fig. 1). Since the typical time-scale of mixing of

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a convective boundary layer is about 15 min., the evolution of species with a chemical time scale of minutes to several hours (e.g. O_3, isoprene, NO, and NO_2) will be to a similar extent driven by both dynamics and chemistry.

The MXL model has been developed in the 1960's and 1970's (Lilly, 1968; Tennekes, 1973) to study the dynamics of clear and cloudy boundary layers and was first applied to atmospheric chemistry by Vilà-Guerau de Arellano et al. (2009), as MXLCH (MXL-CHemistry). Afterwards MXLCH has been applied in studies of the diurnal evolution of gas-phase chemistry (Vilà-Guerau de Arellano et al., 2011; Ouwersloot et al., 2012; Van Stratum et al., 2012) and organic aerosol (Janssen et al., 2012, 2013) in mid-latitude, tropical and boreal areas. The model explicitly accounts for entrainment (ABL-FT exchange) and parametrizes turbulence in a simplified way. Although MXL uses simple parametrizations for turbulence and entrainment, the ABL dynamics as simulated by the model compare well with results from a turbulence resolving Large Eddy Simulation model (Pino et al., 2006) and with vertical profiles as observed from radio soundings (Ouwersloot et al., 2012).

With the implementation of MXL in MESSy (henceforth MXL/MESSy) as dynamical core and the subsequent coupling to the submodels that represent the other processes relevant to the evolution of chemical species in the troposphere, a versatile boundary layer chemistry model is created. MXL/MESSy can be used for different types of studies. First, it is well-suited to support the interpretation of field observations in an eulerian framework, e.g. observations from a tower. After obtaining a best fit with the observational data, a budget analysis can be performed, dividing the total tendency of species into the contributions of gas-phase chemistry, ABL dynamics, emission, deposition and/or gas/particle partitioning. The chemistry term can be further subdivided into the total production and loss terms, and the contributions of individual reactions to these terms.

Secondly, MXL/MESSy is well suited for performing systematic sensitivity analyses, in which the parameter space is explored and through which non-linearities can be studied. Because it is computationally inexpensive, detailed and systematic sensitivity

runs are possible. In that way, it can serve as a test bed for new parametrizations, especially in combination with a direct comparison to field data. It thus forms a link between theoretical/lab study results and 3-D model applications. Finally, it can be used in theoretical/conceptual studies, for a qualitative evaluation of feedbacks and forcings (e.g. Vilà-Guerau de Arellano et al., 2012; Janssen et al., 2012).

In Sect. 2 we describe the MXL model and give its governing equations, in Sect. 3 we focus on the implementation of the MXL model in MESSy and in Sect. 4 we show an example application of MXL/MESSy using observations from the DOMINO campaign.

2 MXL submodel description

The version of the MiXed Layer model that we implement here has been developed in several stages: Lilly (1968) developed the mixed-layer equations and Tennekes (1973) the turbulence closure that we apply in this work. The model for the dynamics of the convective boundary layer that resulted from this work was later coupled to an atmospheric chemistry module by Vilà-Guerau de Arellano et al. (2009) and a land surface scheme by Van Heerwaarden et al. (2009, 2010).

The mixed layer theory states that under convective conditions, strong turbulent flow, driven by the surface heat fluxes, causes perfect mixing of quantities over the entire depth of the ABL (Vilà-Guerau de Arellano et al., 2015). Therefore, scalars and reactants in the convective boundary layer are characterized by well-mixed vertical profile over the whole depth of the ABL. The transition between the well-mixed ABL and the free troposphere is marked by an infinitesimally thin inversion layer. Typical profiles of potential temperature (θ), specific humidity (q) and chemical species in the mixed-layer model are shown in Fig. 2. It shows that the ABL is represented by bulk values of scalars and chemical species mixing ratios, and capped by a infinitesimally thin layer over which these strongly change. Above this inversion layer lies the residual layer (the remainder of the convective boundary layer of the previous day) during the early morning and the free troposphere later during the day, which are characterized in the model

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by a lapse rate for heat and moisture, and a concentration profile of chemical species which is constant with height.

In addition to the local surface heat fluxes, the boundary layer dynamics can be influenced by large-scale atmospheric flows which act as external forcings on the ABL development: for instance, Janssen et al. (2013) and Pietersen et al. (2014) studied cases for which the observed ABL dynamics could only be reproduced if the influence of advection, caused by meso-scale flows, and/or subsidence, caused by a high-pressure system, were taken into account. Therefore, advection and subsidence can be prescribed as forcings to MXL.

2.1 Governing equations for the heat budget

The main variable in the dynamics of the convective boundary layer is the potential temperature (θ), since it is used to quantify the convective turbulence. The evolution of the potential temperature of a dry convective boundary layer is driven by the heat input at the surface (surface heat flux), at boundary layer top (entrainment heat flux) and at the lateral boundaries (advection):

$$\frac{\partial \langle \theta \rangle}{\partial t} = \overbrace{\frac{\overline{(w'\theta')}_s}{h}}^{\text{surface heat flux}} - \overbrace{\frac{\overline{\left(w'\theta'\right)}_e}{h}}^{\text{entrainment heat flux}} + \overbrace{\text{adv}_\theta}^{\text{heat advection}} \qquad (1)$$

Equation (1) is the result of a vertical integration of the 1-dimensional equation of the heat budget and we have assumed that the vertical profile of θ is in quasi-steady state (Lilly, 1968), which causes a linear vertical gradient of the heat flux. In this equation, $\overline{w'\theta'}_s$ is the kinematic heat flux at the surface, which is related to the sensible heat flux (SH) as: $\overline{w'\theta'}_s = \text{SH}/(\rho \cdot c_p)$, with ρ the density of air and c_p the specific heat of air. The entrainment process, represented by $\overline{\left(w'\theta'\right)}_e$, is defined as the process whereby air from the FT is mixed in into the mixed layer, and it is therefore related to the θ jump

at the inversion. The evolution of $\langle\theta\rangle$ thus equals the input of heat into the ABL at the surface and due to entrainment over the ABL height h. Additionally, a heat advection term (adv_θ), which is a large-scale forcing on the boundary layer dynamics, can be prescribed to the model.

When calculating the entrainment flux, we assume that the transition from the ABL to the FT, the inversion, is represented by a sharp discontinuity, namely the zero-order jump closure (ZOJ, Tennekes, 1973). ZOJ closure defines this jump as $\Delta\theta = \theta_{FT} - \langle\theta\rangle$ over an infinitely thin inversion layer. In this ZOJ approach, the entrainment flux is the product of the entrainment velocity w_e (defined positive in the upward direction) and the potential temperature jump $\Delta\theta$ at the inversion. First-order closure approaches exist, which include the explicit representation of the depth of the entrainment zone, but the ZOJ approach already gives satisfactory results (Pino et al., 2006).

The equation for the potential temperature entrainment flux reads:

$$\left(\overline{w'\theta'}\right)_e = -\left(\frac{\partial h}{\partial t} - w_s\right)\Delta\theta = -w_e \cdot \Delta\theta \tag{2}$$

In our model, we calculate the subsidence velocity as:

$$w_s = -\varpi h \tag{3}$$

where ϖ represents the large scale vertical velocity that is a function of the horizontal wind divergence in s^{-1}, i.e. $\varpi = -\text{Div}(\boldsymbol{U})$; where $\boldsymbol{U}$ is the horizontal wind. It can be thought of as the fraction with which the ABL is pushed down per second, due to large-scale subsiding air motions. Therefore, w_s is per definition negative.

By rewriting Eq. (2), we obtain an expression for the boundary layer growth ($\frac{\partial h}{\partial t}$) as a function of the entrainment flux $\left(\overline{w'\theta'}\right)_e$, the potential temperature jump ($\Delta\theta$) and the subsidence velocity (w_s). It reads:

$$\frac{\partial h}{\partial t} = -\frac{\left(\overline{w'\theta'}\right)_e}{\Delta\theta} + w_s = w_e + w_s \tag{4}$$

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This equation states that the mixed layer grows by entrainment of warm air from the free atmosphere ($w_e > 0$) and that it is opposed by the vertical subsidence velocity ($w_s < 0$) driven by high pressure situations. This growth is limited by the presence of a stably stratified layer defined by a potential temperature jump on top of the convective ABL. This jump at the inversion, represented by $\Delta\theta$, changes during the growth of the mixed-layer. Consequently, depending on a positive or negative tendency of $\Delta\theta$, the ABL growth increases or decreases.

As Eq. (2) shows, the entrainment flux for heat depends on the entrainment velocity and the potential temperature jump at the inversion. It is therefore necessary to obtain a prognostic equation for the potential temperature jump. This equation reads:

$$\frac{\partial \Delta\theta}{\partial t} = \frac{\partial \theta_{FT}}{\partial t} - \frac{\partial \langle\theta\rangle}{\partial t} = \gamma_\theta \left(\frac{\partial h}{\partial t} - w_s \right) - \frac{\partial \langle\theta\rangle}{\partial t} \tag{5}$$

where γ_θ is the lapse rate (change with height) of θ in the FT. The final set of equations, which describes the heat budget in the diurnal atmospheric boundary layer, is therefore composed by Eqs. (1), (4) and (5). The three prognostic variables are $\langle\theta\rangle$, $\Delta\theta$ and h.

In this set of equations, four variables still remain as unknowns. The surface heat flux $\left(\overline{w'\theta'} \right)_s$ is the result of land–atmosphere interactions, but can be prescribed based on observations. The external variables w_s and γ_θ represent the influence of the free tropospheric conditions on the ABL development and their values are imposed on the mixed-layer model. Finally, the entrainment heat flux $\left(\overline{w'\theta'} \right)_e$ remains. Here, we assume an important closure and we relate the entrainment flux to the surface heat flux as:

$$\left(\overline{w'\theta'} \right)_e = -\beta \left(\overline{w'\theta'} \right)_s \tag{6}$$

where the coefficient β can be imposed as a constant or as a parameter depending on the thermodynamic characteristics at the inversion, for instance the presence of shear (Tennekes and Driedonks, 1981; Pino et al., 2003; Conzemius and Fedorovich, 2006).

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In our model, we assume a value of β equal to 0.2, which physically means that the contribution of entrainment to the heat budget equals 20 % of the contribution of the surface heat flux. The potential temperature is normally underestimated if this contribution is neglected. As we will show later on when introducing the moisture budget,
5 Eq. (6) needs to be modified to make entrainment dependent on the buoyancy flux.

2.2 Governing equations for the moisture budget

By adding the moisture budget to the heat budget, we incorporate the effect of moisture on the buoyancy of air parcels and therewith complete the configuration of the thermodynamic variables in the ABL. The inclusion of the moisture effects on the dynamics of
10 the ABL requires the introduction of two new equations. The first one (similar to Eq. 1) is the evolution of the mixed-layer specific humidity $\langle q \rangle$:

$$\frac{\partial \langle q \rangle}{\partial t} = \overbrace{\frac{\left(\overline{w'q'}\right)_{s}}{h}}^{\text{surface moisture flux}} - \overbrace{\frac{\left(\overline{w'q'}\right)_{e}}{h}}^{\text{entrainment moisture flux}} + \overbrace{\text{adv}_{q}}^{\text{moisture advection}} \tag{7}$$

where $\left(\overline{w'q'}\right)_{s}$ represents the surface specific moisture flux, $\left(\overline{w'q'}\right)_{e}$ the entrainment flux of moisture and adv_{q} an optionally prescribed moisture advection term. The spe-
15 cific moisture flux is related to the latent heat flux (LE) following $\left(\overline{w'q'}\right)_{s} = \text{LE}/(\rho \cdot L_{v})$, with L_{v} the latent heat of vaporization. Similarly to Eq. (2), we represent the entrainment flux as:

$$\left(\overline{w'q'}\right)_{e} = -\left(\frac{\partial h}{\partial t} - w_{s}\right)\Delta q = -w_{e} \cdot \Delta q. \tag{8}$$

This equation relates the dynamics of the boundary layer growth, represented by the
20 entrainment velocity, with the specific humidity jump at the interface between the ABL and the FT.

Equation (8) requires an additional equation for the temporal evolution of the jump of q at the ABL-FT interface (Δq). It reads:

$$\frac{\partial \Delta q}{\partial t} = \frac{\partial q_{FT}}{\partial t} - \frac{\partial \langle q \rangle}{\partial t} = \gamma_q \left(\frac{\partial h}{\partial t} - w_s \right) - \frac{\partial \langle q \rangle}{\partial t} \tag{9}$$

where γ_q is the lapse rate of q in the FT.

At this point, we need to introduce a new variable, the virtual potential temperature, which accounts for the both changes in the heat and moisture budgets and is used to quantify buoyancy. It is defined as:

$$\theta_v = \theta(1 + 0.61q) \tag{10}$$

The virtual potential temperature thus account for the effect of water vapour on the density of air. Since moist air is less dense than dry air at the same conditions of temperature and pressure, θ_v is always greater than the actual temperature, but only by a few degrees. The turbulent transport of this variable, the buoyancy flux, combines in one quantity the information of the potential temperature flux and the specific moisture flux. It reads:

$$\overline{w'\theta_v'} = \overline{w'\theta'} + 0.61 \left(\langle \theta \rangle \overline{w'q'} + \langle q \rangle \overline{w'\theta'} + \overline{w'\theta'q'} \right) \approx \overline{w'\theta'} + 0.61 \left(\langle \theta \rangle \overline{w'q'} \right). \tag{11}$$

This buoyancy flux expresses the production of turbulent kinetic energy in the ABL due to density differences. Turbulence driven by shear (mechanical turbulence) on the mixed-layer thermodynamic equations is not dealt with in our model.

The buoyancy flux directly enters in the ABL growth formulated in Eq. (2). Therefore, we rewrite Eq. (2) in the definitive form as:

$$\left(\overline{w'\theta_v'} \right)_e = - \left(\frac{\partial h}{\partial t} - w_s \right) \Delta \theta_v = -w_e \cdot \Delta \theta_v, \tag{12}$$

where $\Delta \theta_v$ is expressed in terms of the characteristics of the θ- and q-budgets:

$$\Delta \theta_v = \Delta \theta + 0.61 \left(\langle q \rangle \Delta \theta + \langle \theta \rangle \Delta q + \Delta \theta \Delta q \right) \tag{13}$$

By introducing the buoyancy flux as the driver of the turbulent process in the determination of the boundary layer growth, we complete the main framework of our model formulation based on mixed-layer theory.

In summary, the combined heat and moisture system is composed by the following 6 equations:

1. prognostic budget equations for $\langle \theta \rangle$ and $\langle q \rangle$ (Eqs. 1 and 7),

2. boundary layer growth (Eq. 12) rewritten as:

$$\frac{\partial h}{\partial t} = -\frac{\left(\overline{w'\theta_v'}\right)_e}{\Delta\theta_v} + w_s, \tag{14}$$

3. prognostic equations for $\Delta\theta$ and Δq (Eqs. 5 and 9),

4. closure assumption relating the surface buoyancy flux to the entrainment buoyancy flux $\left(\overline{w'\theta_v'}\right)_e = -\beta \left(\overline{w'\theta_v'}\right)_s$.

2.3 Governing equations for the horizontal wind budget

As the last part of the mixed-layer dynamics, we introduce the horizontal wind, or the momentum budget. The wind speed is important for calculating surface–atmosphere exchange, and appears for instance in the calculations of the aerodynamic resistance which governs evapotranspiration (Sect. 2.7.2) and dry deposition.

Similar expressions as for $\langle \theta \rangle$ and $\langle q \rangle$ can be used for the two horizontal components of the wind speed. These equations also contain the Coriolis force that takes into

account the rotation of the Earth. This gives another four equations:

$$\frac{d\langle u \rangle}{dt} = \frac{\left(\overline{u'w'}\right)_s}{h} + \frac{\left(\overline{u'w'}\right)_e}{h} - f_c \Delta v \tag{15}$$

$$\frac{d\langle v \rangle}{dt} = \frac{\left(\overline{v'w'}\right)_s}{h} + \frac{\left(\overline{v'w'}\right)_e}{h} + f_c \Delta u \tag{16}$$

$$\frac{d\Delta u}{dt} = \gamma_u w_e - \frac{d\langle u \rangle}{dt} \tag{17}$$

$$\frac{d\Delta v}{dt} = \gamma_v w_e - \frac{d\langle v \rangle}{dt} \tag{18}$$

where $\langle u \rangle$ and $\langle v \rangle$ are the mixed-layer wind velocities in x and y direction, Δu and Δv are the jumps of these two variables, γ_u and γ_v are the lapse rates in the FT and f_c is the Coriolis parameter. The free tropospheric wind velocities are assumed to be equal to the geostrophic wind at their height. Also for the wind budget, we assume zero-order closure:

$$\overline{(u'w')}_e = -w_e \cdot \Delta u \tag{19}$$

$$\overline{(v'w')}_e = -w_e \cdot \Delta v \tag{20}$$

For applications such as in Eq. (41), u and v are combined to form the total horizontal wind speed:

$$U = \sqrt{\langle u^2 \rangle + \langle v^2 \rangle + w_*^2} \tag{21}$$

where w_* is the free convection scaling velocity:

$$w_* = \left(\frac{g \cdot h}{\theta_v} \overline{w'\theta'_{v,s}} \right)^{1/3} \tag{22}$$

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2.4 Governing equations for chemical species

The mixed-layer model allows us to investigate the influence of the boundary layer dynamics on reactive atmospheric compounds during daytime. It is important to stress that in mixed-layer theory, we assume that species are mixed instantaneously as soon as they are chemically produced, emitted or entrained from the FT. In other words, the MXL model acts as a reactive chamber with the additional advantage of accounting for boundary layer growth and ABL-FT exchange.

As for scalars, the inclusion of reactive species requires the introduction of two additional equations for each species. The expression for the evolution of the generic species C is similar to that for potential temperature (Eq. 1) and moisture (Eq. 7), but includes a term for the chemical transformation:

$$\frac{\partial \langle C \rangle}{\partial t} = \overbrace{\frac{\left(\overline{w'C'}\right)_s}{h}}^{\text{emis./dep. flux}} - \overbrace{\frac{\left(\overline{w'C'}\right)_e}{h}}^{\text{entrainment flux}} + \overbrace{S_C}^{\text{chemistry}} \tag{23}$$

By solving Eq. (23), we determine how C varies over time as a function of emission/deposition processes at the surface (represented by the term $(\overline{w'C'})_s$), the dynamic effects (h and $(\overline{w'C'})_e$) and the chemical transformation (S_C). Note that the surface fluxes of chemical species (emission or deposition) can be either prescribed (Sect. 2.5) to MXL/MESSy or calculated by other MESSy submodels (Sect. 3.1, Table 1). Advection of tracers is not considered, because horizontal gradients in species concentrations, which are necessary to calculate advection, are generally not constrained by observations. We are planning to implement the TNUDGE submodel (Kerkweg et al., 2006b) in the future to allow the relaxation of chemical species concentrations towards observations, which will also compensate for sources or sinks that are the result of advection of air masses with different concentrations of species.

The flux at the top of the boundary layer is represented in the same way as the entrainment flux for buoyancy, moisture and wind. For C it reads:

$$\left(\overline{w'C'}\right)_e = -w_e \cdot \Delta C \tag{24}$$

By representing the exchange of C as the product of the entrainment velocity and the jump of C, we account for both the dynamics and chemistry, which together determine the exchange between the ABL and FT.

Equation (24) requires an additional prognostic equation to solve the evolution of ΔC:

$$\frac{\partial \Delta C}{\partial t} = \frac{\partial C_{FT}}{\partial t} - \frac{\partial \langle C \rangle}{\partial t} \tag{25}$$

where C_{FT} is the concentration of C in the FT with $\partial C_{FT}/\partial t$ driven by chemical production and loss only in MXL/MESSy.

By assuming instantaneous mixing, we assume that the intensity of segregation of chemical species due to possible inefficient turbulent mixing is zero. It has been hypothesized that this segregation might be important for explaining model-measurements discrepancies for certain species, e.g. isoprene and the hydroxyl radical (Butler et al., 2008; Pugh et al., 2010). Later studies, however, showed that the effects of imperfect mixing are only small, with a maximum reduction of the effective rate constant for the isoprene and OH by less than 15 % (Ouwersloot et al., 2011; Pugh et al., 2011), and not by as much as 50 %, which was necessary to explain the measurement-model discrepancy. Moreover, Large Eddy Simulation results show well-mixed profiles throughout the convective boundary layer for species like isoprene and OH (Vilà-Guerau de Arellano et al., 2009, 2011; Ouwersloot et al., 2011) under homogeneous land surface conditions. A parametrization for the effect of the segregation of species on the effective rate constants was developed by Vinuesa and Vilà-Guerau de Arellano (2003) and can in principle be implemented to account for this effect.

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2.5 Simple emission functions

Besides online calculation of emissions, MXL/MESSy provides the possibility to prescribe emissions, following simplified diurnal emission patterns. These include constant, sinusoidal and cosine shaped emission fluxes, respectively:

$$\left(\overline{w'C'}\right)_{\text{s}} = F_C \tag{26}$$

F_C is the (maximum) daily emission flux ($\text{ppb}\,\text{m}\,\text{s}^{-1}$),

$$\left(\overline{w'C'}\right)_{\text{s}} = F_C \cdot \sin\left(\frac{\pi t}{t_{\text{d}}}\right) \tag{27}$$

where t is the time of day and t_{d} the day length and

$$\left(\overline{w'C'}\right)_{\text{s}} = F_C \cdot \left(1 - \cos\left(\frac{2\pi t}{t_{\text{d}}}\right)\right). \tag{28}$$

For these functions, F_C can be set by the user in the MXL namelist.

2.6 Radiation model

In this section, we describe a simple radiation model that is implemented along with MXL/MESSy. The MESSy radiation submodel RAD is intended to be implemented in the future as a second option.

The net radiation at the surface R_{n} is defined as the sum of the four radiation components:

$$R_{\text{n}} = S_{\text{in}} - S_{\text{out}} + L_{\text{in}} - L_{\text{out}} \tag{29}$$

where S_{in} and S_{out} are the incoming and outgoing components of the short wave radation, and L_{in} and L_{out} the incoming and outgoing component of the long wave radiation.

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The incoming short wave radiation S_{in} is calculated by:

$$S_{in} = S_o T_r \sin(\Psi) \tag{30}$$

where S_o is the constant solar irradiance at the top of the atmosphere, taken at 1365 W m^{-2}, T_r the net sky transmissivity, that takes into account the influence of radiative path length and atmospheric absorption and scattering using:

$$T_r = 0.6 + 0.2\sin(\Psi) \tag{31}$$

Through Ψ, both expressions depend on geographical location, day of the year and time of the day. Ψ is calculated using:

$$\sin(\Psi) = \sin(\phi)\sin(\delta_s) - \cos(\phi)\cos(\delta_s)\cos\left(2\pi\frac{t_{UTC}}{t_d} + \lambda_e\right) \tag{32}$$

where t_{UTC} is the universal time (UTC) and t_d is the diurnal period of 24 h. The latitude ϕ (positive north of Equator) and longitude λ_e (positive east of Greenwich) define the geographic location. Variable δ_s is the solar declination, which is a function of the day number:

$$\delta_s = \Phi_r \cos\left(2\pi\frac{d - d_r}{d_y}\right) \tag{33}$$

where Φ_r is the tilt of the Earth's axis relative to the elliptic, equal to 0.409 rad. The Julian day is represented by d and d_r is 173. The number of days in a year d_y is taken as 365.

The outgoing shortwave radiation depends on the surface albedo:

$$S_{out} = \alpha S_{in} \tag{34}$$

where α is the surface albedo.

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The outgoing long wave radiation is calculated using the the Stefan–Boltzmann Law:

$$L_{out} = \epsilon_s \sigma_{SB} T_s^4 \tag{35}$$

where ϵ_s is the surface emissivity taken equal to 1, σ_{SB} the Stefan–Boltzmann constant
equal to $5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$, and T_s the surface temperature (K).

The incoming long wave radiation is computed using the same expression, but here it uses the temperature at the top of the surface layer T_{sl}. This temperature is acquired by converting the potential temperature of the mixed-layer θ to absolute temperature using the height of surface layer top, which we define as 10 % of the boundary-layer height h. This gives the expression:

$$L_{in} = \epsilon_a \sigma_{SB} T_{sl}^4 \tag{36}$$

where the atmospheric emissivity ϵ_a is 0.8.

2.7 Land surface model

A land surface model (Van Heerwaarden et al., 2009, 2010; Van Heerwaarden, 2011) is included as an option in MXL/MESSy, which enables the interactive calculation of surface heat fluxes. With this inclusion, MXL/MESSy can be used to simultaneously and interactively calculate the exchange of energy (sensible heat flux), water (latent heat flux), wind (momentum flux) and chemical species (emission and deposition) between the land surface and the ABL. In that way, a fully online coupled land surface-ABL chemistry model is obtained, in which all exchanges between the land surface and the ABL at the diurnal time-scale are internal variables of the coupled system. This means that only forcings (drivers external to the system at the appropriate time scales) are prescibed to the model. These includes land surface forcings (e.g. leaf area index (LAI), roughness length, soil moisture) and large-scale meteorological forcings on the ABL (e.g. incoming solar radiation, FT conditions).

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2.7.1 The surface energy balance

The surface energy balance (SEB) forms the basis for the land surface model. It relates net radiation R_n (Eq. 29) to the surface heat fluxes:

$$R_n - G = \text{SH} + \text{LE} \tag{37}$$

where SH is the sensible heat flux, LE is the latent heat flux and G is the ground heat flux.

The sensible and latent heat flux, which together supply the energy to the atmosphere that drives turbulent convection, are a function of land surface and atmospheric properties:

$$\text{SH} = \frac{\rho c_p}{r_a}(\theta_s - \langle\theta\rangle) \tag{38}$$

$$\text{LE} = \frac{\rho L_v}{r_a + r_s}(q_{sat}(T_s) - \langle q\rangle) \tag{39}$$

where r_a is the aerodynamic resistance, r_s is the surface resistance, θ_s and $\langle\theta\rangle$ are the potential temperatures of the surface and the mixed-layer respectively, $q_{sat}(T_s)$ is the saturated specific humidity inside the canopy and $\langle q\rangle$ is the mixed-layer specific humidity.

The Penman–Monteith equation (Monteith, 1965) is then used to relate the evaporation flux to radiation, temperature, humidity, and aerodynamic and surface resistance:

$$\text{LE} = \frac{\frac{dq_{sat}}{dT}(R_n - G) + \frac{\rho c_p}{r_a}\left(q_{sat}(T) - \langle q\rangle\right)}{\frac{dq_{sat}}{dT} + \frac{c_p}{L_v}\left(1 + \frac{r_s}{r_a}\right)}, \tag{40}$$

where $\frac{dq_{sat}}{dT}$, which is the slope of the saturated specific humidity curve with respect to temperature, and q_{sat} are evaluated using the atmospheric mixed-layer temperature at the top of the atmospheric surface layer defined as $z_{sl} = 0.1h$.

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2.7.2 Resistances

The exchange of momentum and heat between the atmosphere with the surface is related to the strength of the turbulence. The aerodynamic resistance r_a, that appears in Eqs. (38) and (39) is the inverse of this turbulent intensity and defined as:

$$r_a = \frac{1}{C_H \langle U \rangle} \tag{41}$$

where C_H is the drag coefficient for heat and $\langle U \rangle$ is the mixed-layer wind speed (Sect. 2.3). It shows that stronger winds lead to lower aerodynamic resistance and enhanced turbulent exchange.

The turbulent exchange also depends on the drag, which is a function of the stability of the surface layer: a growing instability in the surface layer leads to stronger convection and mixing. The drag coefficient is calculated following:

$$C_H = \frac{\kappa^2}{A \cdot B} \tag{42}$$

with:

$$A = \ln\left(\frac{z_{sl}}{z_{0m}}\right) - \Psi_M\left(\frac{z_{sl}}{L}\right) + \Psi_M\left(\frac{z_{0m}}{L}\right)$$

$$B = \ln\left(\frac{z_{sl}}{z_{0h}}\right) - \Psi_H\left(\frac{z_{sl}}{L}\right) + \Psi_H\left(\frac{z_{0h}}{L}\right)$$

where κ is the von Kármán constant, z_{0m} and z_{0h} are the roughness lengths for momentum and heat, respectively, z_{sl} is the depth of the atmospheric surface layer of $0.1h$, L is the Monin–Obukhov length and Ψ_M and Ψ_H are the integrated stability functions for momentum and heat taken from Beljaars (1991). To find the values of L that are required in this function, the following implicit function is solved using a Newton–Raphson

iteration method:

$$Ri_B = \frac{g}{\langle\theta_v\rangle}\frac{z_{sl}\left(\langle\theta_v\rangle - \theta_{vs}\right)}{U^2} \tag{43}$$

$$= \frac{z_{sl}}{L}\frac{\left[\ln\left(\frac{z_{sl}}{z_{0h}}\right) - \Psi_H\left(\frac{z_{sl}}{L}\right) + \Psi_H\left(\frac{z_{0h}}{L}\right)\right]}{\left[\ln\left(\frac{z_{sl}}{z_{0m}}\right) - \Psi_M\left(\frac{z_{sl}}{L}\right) + \Psi_M\left(\frac{z_{0m}}{L}\right)\right]^2}$$

where Ri_B is the bulk Richardson number and θ_{vs} and $\langle\theta_v\rangle$ are the virtual potential temperatures of the surface and the mixed-layer atmosphere respectively.

The calculations of both resistances r_{veg} and r_{soil} follow the method of Jarvis (1976) and employed similarly as in the ECMWF global forecasting model. The vegetation resistance is based on the following multiplicative equation.

$$r_{veg} = \frac{r_{s,min}}{LAI}f_1(S_{in})\,f_2(w_{soil2})\,f_3(VPD)f_4(T) \tag{44}$$

where $r_{s,min}$ is the minimum surface resistance, LAI the leaf area index of the vegetated fraction, f_1 a correction function depending on incoming short wave radiation S_{in}, f_2 a function depending on soil moisture w, f_3 a function depending on vapor pressure deficit (VPD) and f_4 a function depending on temperature T. Of these correction functions, the first three originate from the ECMWF model documentation and the fourth from Noilhan and Planton (1989) and are defined as:

$$\frac{1}{f_1(S_{in})} = \min\left(1, \frac{0.004S_{in} + 0.05}{0.81\,(0.004S_{in} + 1)}\right) \tag{45}$$

$$\frac{1}{f_2(w)} = \frac{w - w_{wilt}}{w_{fc} - w_{wilt}} \tag{46}$$

$$\frac{1}{f_3(VPD)} = \exp\left(g_D VPD\right) \tag{47}$$

$$\frac{1}{f_4(T)} = 1.0 - 0.0016(298.0 - T)^2, \tag{48}$$

where w_{wilt} is the volumetric soil moisture at wilting point, w_{fc} is the volumetric soil moisture at field capacity and g_D is a correction factor for vapor pressure deficit that only plays a role in high vegetation.

The soil resistance depends on the amount of soil moisture in the layer that has direct contact with the atmosphere.

$$r_{soil} = r_{s,\,min}\, f_2(w_{soil1}), \tag{49}$$

where f_2 is calculated following Eq. (46).

2.7.3 Evapotranspiration calculation

The total evapotranspiration consist of three parts: soil evaporation, leaf transpiration and evaporation of liquid water from the leaf surface. The total evapotranspiration is therefore proportional to the vegetated fraction of the land surface:

$$LE_{tot} = c_{veg}(1 - c_{liq})LE_{veg} + c_{veg}c_{liq}LE_{liq} + (1 - c_{veg})LE_{soil}, \tag{50}$$

where LE_{tot} is the evapotranspiration, LE_{veg} the transpiration from vegetation, LE_{soil} the evaporation from the soil and LE_{liq} evaporation of liquid water. The fractions that are used are c_{veg} which is the fraction of the total area that is covered by vegetation and c_{liq}, which is the fraction of the vegetated area that contains liquid water. Since c_{liq} is not constant in time, it is modeled following:

$$c_{liq} = \frac{W_l}{LAI\, W_{max}} \tag{51}$$

where W_{max} is the representative depth of a water layer that can lay on one leaf and W_l the actual water depth. The evolution of W_l is governed by the following equation:

$$\frac{dW_l}{dt} = \frac{LE_{liq}}{\rho_w L_v} \tag{52}$$

where ρ_w is the density of water.

2.7.4 Soil model

A soil model is available for the calculation of soil temperature and moisture changes that occur on diurnal time scales, and that therefore affect the surface heat fluxes. It is a force-restore soil model, based on the model formulation of Noilhan and Planton (1989) with the soil temperature formulation from Duynkerke (1991). The soil model consists of two layers, of which the thin upper layer follows a diurnal cycle. The soil temperature in the upper layer is calculated using the following equation, where the first term is the force term and the second the restore term:

$$\frac{\mathrm{d}T_{\text{soil1}}}{\mathrm{d}t} = C_T G - \frac{2\pi}{\tau}(T_{\text{soil1}} - T_{\text{soil2}}), \tag{53}$$

where C_T is the surface soil/vegetation heat capacity, G the soil heat flux already introduced in the SEB and τ the time constant of one day. The soil heat flux is calculated using

$$G = \Lambda(T_s - T_{\text{soil1}}) \tag{54}$$

where Λ is the thermal conductivity of the skin layer. The heat capacity C_T is calculated following Clapp and Hornberger (1978) using:

$$C_T = C_{T,\text{sat}} \left(\frac{w_{\text{sat}}}{w_2}\right)^{b/2\log 10} \tag{55}$$

where $C_{T,\text{sat}}$ is the soil heat capacity at saturation and b an empirical constant that originates from data fitting.

The evolution of the volumetric water content w is calculated using:

$$\frac{\mathrm{d}w_1}{\mathrm{d}t} = -\frac{C_1}{\rho_w d_1}\frac{\text{LE}_{\text{soil}}}{L_v} - \frac{C_2}{\tau}\left(w_2 - w_{2,\text{eq}}\right) \tag{56}$$

where w_1 and w_2 are the volumetric water content of the first two layers, C_1 and C_2 are two coefficients related to the Clapp and Hornberger parameterization (Clapp and Hornberger, 1978), that are calculated using:

$$C_1 = C_{1,\text{sat}} \left(\frac{w_{\text{sat}}}{w_2} \right)^{b/2+1} \tag{57}$$

$$C_2 = C_{2,\text{ref}} \frac{w_2}{w_{\text{sat}} - w_2} \tag{58}$$

where $C_{1,\text{sat}}$ and $C_{2,\text{ref}}$ are constants taken from Clapp and Hornberger (1978). The water content in equilibrium $w_{\text{g,eq}}$ is:

$$w_{\text{g,eq}} = w_2 - w_{\text{sat}} a \left(\frac{w_2}{w_{\text{sat}}} \right)^p \left(1 - \frac{w_2}{w_{\text{sat}}} \right)^{8p} \tag{59}$$

with a and p as two more fitted constants from Clapp and Hornberger (1978).

2.7.5 Surface fluxes of momentum

Finally, we introduce parametrizations for the momentum fluxes at the surface (Vilà-Guerau de Arellano et al., 2015) and the friction velocity (u_*), which are required to calculate the horizontal wind budget in the mixed layer (Sect. 2.3).

$$u_* = \sqrt{C_M} U \tag{60}$$

$$\overline{u'w'}_s = -C_M U \langle u \rangle \tag{61}$$

$$\overline{v'w'}_s = -C_M U \langle v \rangle \tag{62}$$

where C_M is the drag coefficient for momentum, defined as:

$$C_M = \frac{\kappa^2}{\left[\ln\left(\frac{z_{\text{sl}}}{z_{0m}} \right) - \Psi_M \left(\frac{z_{\text{sl}}}{L} \right) + \Psi_M \left(\frac{z_{0m}}{L} \right) \right]} \tag{63}$$

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3 Implementation in the MESSy structure

For the implementation of MXL, a generic 1-D basemodel is created in MESSy, called VERTICO (VERTICal COlumn), which defines the number of boxes stacked on top of each other and directs the time integration. VERTICO is a de facto 3-D model in which the horizontal resolution has been reduced to a single grid box, to facilitate the submodel coupling within the MESSy framework. This also facilitates the possible development of a column model that includes more vertical levels in both boundary layer and free troposphere.

The current implementation of VERTICO consists of two domains (see Fig. 4): the lower one represents the well-mixed boundary layer during daytime, as represented by the MXL equations, and the upper box contains a simplified description of the free troposphere on top of the ABL. The prognostic equations for the ABL dynamics and the land surface scheme are integrated by an Euler forward solver.

3.1 Coupling with other MESSy submodels

Through the MESSy framework, two types of submodels are coupled: the first type are the generic submodels, which constitute the model infrastructure that is not directly related to actual physical processes, such as time, tracer and data management. The second type are the process submodels, which represent the individual processes that contribute to the evolution of chemical species in the ABL (see Eq. 23 and Fig. 3). Table 1 shows all generic and process submodels that are currently coupled to form MXL/MESSy, and their tasks. In principle, thanks to such an implementation, any MESSy submodel can be used in MXL/MESSy.

The modular nature of the MESSy interface allows full flexibility: processes can easily be switched on and off through switches in a namelist. Emission and deposition of species takes place only in the lower box, while chemistry and gas/particle partitioning take place in both ABL and FT. The chemical production and loss of a species is calculated in the MECCA submodel (Sander et al., 2011), which uses photolysis rates

from JVAL (Sander et al., 2014). For the lower boundary conditions, there are two possibilities. On one hand, MXL/MESSy offers the possibility to use interactive emission (via the ONEMIS, Kerkweg et al., 2006b and MEGAN Guenther et al., 2006 submodels), dry deposition (DDEP Kerkweg et al., 2006a) and land surface parametrisations

5 (Sect. 2.7). In these submodels, land surface-ABL exchange is calculated as function of land surface and ABL characteristics, like stomatal resistance and air temperature. On the other hand, it is possible to prescribe emissions and surface heat fluxes following simplified functions for the users who like to keep full control over the boundary conditions of the model (Sect. 2.5). In that way, MXL/MESSy can for instance be used

10 to evaluate the sensitivity of the chemistry in the ABL to uncertainties in emission estimates. Further, OFFEMIS (Kerkweg et al., 2006b) allows for the extraction of emission data from an emission database. Additionally, the organic aerosol submodel ORACLE (Tsimpidi et al., 2014), allows for the representation of the organic aerosol composition and evolution.

15 **4 DOMINO case study**

To evaluate MXL/MESSy, we revisited the case study of Van Stratum et al. (2012), which is based on observations from the DOMINO campaign in the south of Spain. MXL/MESSy, with initial conditions as in Table A1, represents the dynamics of the boundary layer well, as compared to observations from a tower and radio soundings

20 (Fig. 5). Since the equations and the initial and boundary conditions are equal to those of Van Stratum et al. (2012), the dynamics of $\langle \theta \rangle$, $\langle q \rangle$ and h are identical to those shown in their Fig. 3. As a result of the positive heat fluxes (Fig. 5d), the initial potential temperature inversion is broken after 09:00 LT, and the boundary layer starts to grow rapidly. During this period of strong ABL growth, air from the FT is entrained, which causes

25 (1) a strong increase in the $\langle \theta \rangle$, since warm air from the FT is entrained and (2) a decrease in $\langle q \rangle$ despite the positive latent heat flux, since dry air from the FT is entrained into the ABL. After 13:00 LT, the ABL growth slows down, and the effect of entrainment

on scalars and chemical species becomes smaller. The correct representation of the boundary layer dynamics ensures that we also simulate the effect of ABL dynamics on the evolution of chemical species well (through h and w_e in Eqs. 23 and 24).

Figure 6 shows the diurnal evolution of several gas-phase chemical species, from measurements and from three model runs:

- the MXL/MESSy code with MIM2 chemistry (Taraborrelli et al., 2009), with initial conditions as in Van Stratum et al. (2012) (SameIC),

- the MXL/MESSy code with MIM2 chemistry (Taraborrelli et al., 2009), with initial conditions that gave the best fit with the measurements (BestFit),

- the MXLCH code as in Van Stratum et al. (2012).

The MESSY/MXL SameIC and MXLCH simulations where performed with identical initial conditions as listed in Table A2. In the MXL/MESSy SameIC run we overestimate NO, NO_2 and isoprene and underestimate O_3, HO_2 and H_2O_2. However, these results are influenced by the surface fluxes (emission and deposition) and the mixing ratios in the FT of several species. Since these were not measured, this gives us some degrees of freedom to adjust them, within realistic bounds, to obtain an improved fit of MXL/MESSy with the observations.

Figure 6 also shows the MXL/MESSy BestFit results (with initial conditions as in Table A2). It shows that MXL/MESSy with the MIM2 chemical mechanism is able to reproduce the diurnal dynamics of the main gas-phase chemical species well. While it reproduces the timing and peak values of OH and HO_2 well, the mixing ratios of both radicals are underestimated in the early morning and late afternoon. Also H_2O_2 is underestimated from 10.00 a.m. to the late afternoon, which is related to the underestimation of HO_2.

To give more insight in the diurnal evolution of chemical species, it is useful to analyse the contributions of the different processes to the total tendency (Eq. 23). The modular nature of MXL/MESSy facilitates the evaluation of the contributions of the individual processes, thanks to the diagnostic tools present in the MESSy framework.

In Fig. 7, an example is shown for O_3, based on the optimally tuned (BestFit) simulation. The upper panel shows how the total O_3 tendency is built up from the contributions of chemistry, entrainment and dry deposition (although the latter is set to zero), and the lower panel shows the chemistry term and how gas-phase and photolysis reactions contribute to the loss and production of ozone in the ABL. The contribution of entrainment to the O_3 budget is strongest during the morning when the ABL grows rapidly and a large quantity of ozone-rich air is mixed in into the ABL from the FT. Except for the early morning and late afternoon, the sum of the chemical production and destruction of O_3 is positive, and the net O_3 production peaks around noon when photochemistry is strongest. Note that while we chose to show total production and loss terms for O_3 here, it possible to split up these terms into the contributions of the individual reactions.

5 Summary

We have implemented the MXL model as a new submodel in MESSy, in order to represent the diurnal dynamics of the atmospheric boundary layer and their effect on atmospheric chemistry. The comprehensiveness of MXL/MESSy in representing the processes relevant to atmospheric chemistry in the convective boundary layer, while keeping computational requirements low, makes it an ideal tool for applications in atmospheric chemistry that ask for systematic sensitivity analyses. These include the interpretation of observations from field campaigns and the evaluation of new process parametrizations under ambient conditions, as well as theoretical studies on the coupled land surface-boundary layer-atmospheric chemistry system. Expansion of the model with additional relevant MESSy submodels and conversion into a multi-layer column model is planned for the future: e.g. including the surface layer, entrainment zone, account for imperfect mixing in ABL (e.g. due to clouds, aerosol layers) and the stable nocturnal boundary layer.

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R. H. H. Janssen and A. Pozzer

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Code availability

MXL/MESSy is part of the Modular Earth Submodel System (MESSy), which is continuously further developed and applied by a consortium of institutions. The usage of MESSy and access to the source code is licenced to all affiliates of institutions which are members of the MESSy Consortium. Institutions can be a member of the MESSy Consortium by signing the MESSy Memorandum of Understanding. More information can be found on the MESSy Consortium Website (www.messy-interface.org).

The Supplement related to this article is available online at doi:10.5194/gmdd-7-7197-2014-supplement.

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Boundary layer dynamics with MXL/MESSy

R. H. H. Janssen and A. Pozzer

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Table 1. List of MESSy submodels coupled with MXL.

submodel	description	reference
generic submodels		
BLATHER	standard output to log-files	Jöckel et al. (2010)
CHANNEL	memory and meta-data management and data export	Jöckel et al. (2010)
CONSTANTS	constants shared between submodels	Jöckel et al. (2005)
SWITCH & CONTROL	switch and call individual submodels	Jöckel et al. (2005)
TIMER	time control	Jöckel et al. (2010)
TOOLS	tools shared between submodels	Jöckel et al. (2005)
TRACER	management of data and meta-data of constituents	Jöckel et al. (2008)
process submodels		
DDEP	dry deposition of trace gases and aerosols	Kerkweg et al. (2006a)
JVAL	photolysis rates	Sander et al. (2014)
MECCA	atmospheric chemistry	Sander et al. (2011)
MEGAN	emission of tracers	Guenther et al. (2006)
OFFEMIS (formerly OFFLEM)	prescribed emissions of trace gases and aerosols	Kerkweg et al. (2006b)
ONEMIS (formerly ONLEM)	online calculated emissions of trace gases and aerosols	Kerkweg et al. (2006b)
ORACLE	organic aerosol composition and evolution	Tsimpidi et al. (2014)
PTRAC	define additional prognostic tracers via namelist	Jöckel et al. (2008)
RAD[a]	radiation, heating rates	www.messy-interface.org
SURFACE[a]	surface processes	www.messy-interface.org
TNUDGE[b]	tracer nudging	Kerkweg et al. (2006b)

[a] These submodels are under development and will be implemented in the future. Currently, radiation and land surface properties are calculated as described in Sect. 2.6 and 2.7, respectively.
[b] To be implemented.

Table A1. The initial and boundary conditions in the atmospheric boundary layer (ABL) and free troposphere (FT) as as used in MXL/MESSy and MXLCH. All initial conditions are imposed at 07:00 LT. t is the time elapsed since the start of the simulation (s) and t_d the length of the simulation (s). The subscripts s and e indicate values at the surface and the entrainment zone, respectively.

Property	Value
Initial ABL height h (m)	500
Subsidence rate ω (s^{-1})	5×10^{-6}
Surface sensible heat flux $\overline{w'\theta'}_s$ (K m s^{-1})	$0.22 \sin(\pi t/t_d)$
Entrainment/surface heat flux ratio $\beta = -\overline{w'\theta'}_e/\overline{w'\theta'}_s$ (dimensionless)	0.2
Initial ABL potential temperature $\langle\theta\rangle$ (K)	287
Initial FT potential temperature θ_{FT} (K)	288.5
Potential temperature lapse rate FT γ_θ (K m^{-1})	0.006
Surface latent heat flux $\overline{w'q'}_s$ (g kg^{-1} m s^{-1})	$0.03 \sin(\pi t/t_d)$
Initial ABL specific humidity $\langle q\rangle$ (g kg^{-1})	5.3
Initial FT specific humidity q_{FT} (g kg^{-1})	4.5
Specific humidity lapse rate FT γ_q (g kg^{-1} m^{-1})	−0.0012

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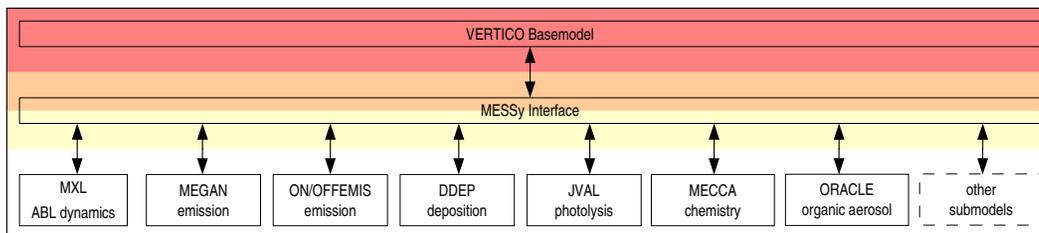Full Screen / Esc

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Table A2. Initial mixing ratio in ABL and FT, and surface emission fluxes of the reactants for the different runs in MXL/MESSy and MXLCH. Species in the reaction mechanism that are not included in this table have zero initial concentrations and zero surface emissions. For O_2 and N_2 we have imposed the values 2×10^8 and 8×10^8 ppb, respectively.

	O_3	NO	NO_2	ISO	CH_4	CO	H_2O_2
Van Stratum et al. (2012)							
Initial mixing ratio (ppb)							
ABL	30.0	0.0	0.6	0.0	1724.0	105.0	0.1
FT	39.0	0.0	0.0	0.0	1724.0	105.0	0.1
Surface emission flux (mg m^{-2} h^{-1})	0.0	0.13	0.0	$0.30\sin\left(\frac{\pi t}{t_d}\right)$	0.0	0.0	0.0
BestFit MXL/MESSy							
Initial mixing ratio (ppb)							
ABL	30.0	0.008	0.65	0.0	1742.0	105.0	0.1
FT	41.0	0.0	0.0	0.0	1742.0	105.0	0.1
Surface emission flux (mg m^{-2} h^{-1})	0.0	0.07	0.0	$0.15\sin\left(\frac{\pi t}{t_d}\right)$	0.0	0.0	0.0
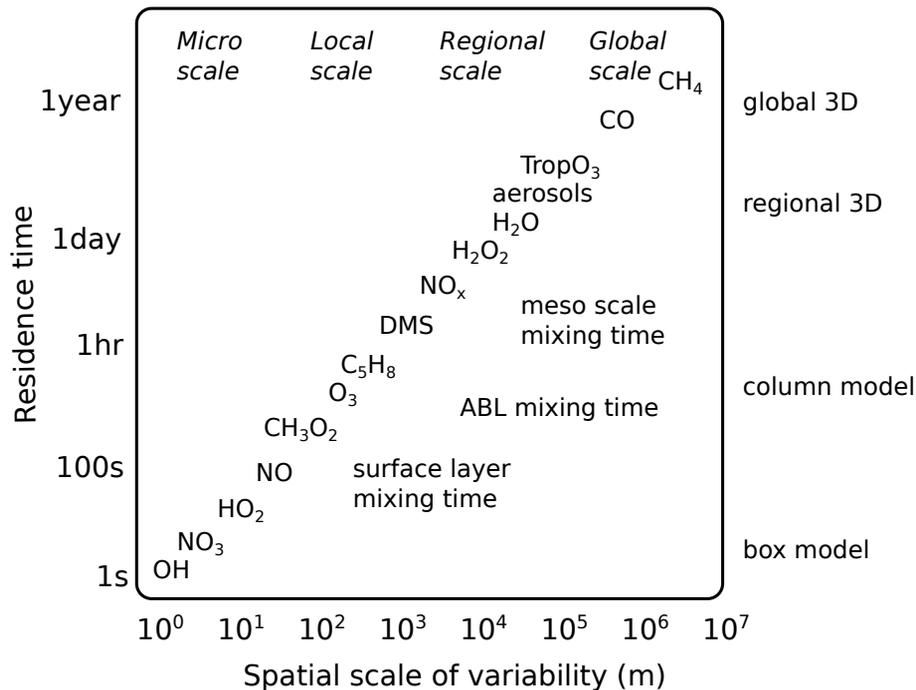
Figure 1. Spatio-temporal scales in atmospheric chemistry, typical species associated with these scales and the type of model that should be used to study their behaviour in the atmosphere.

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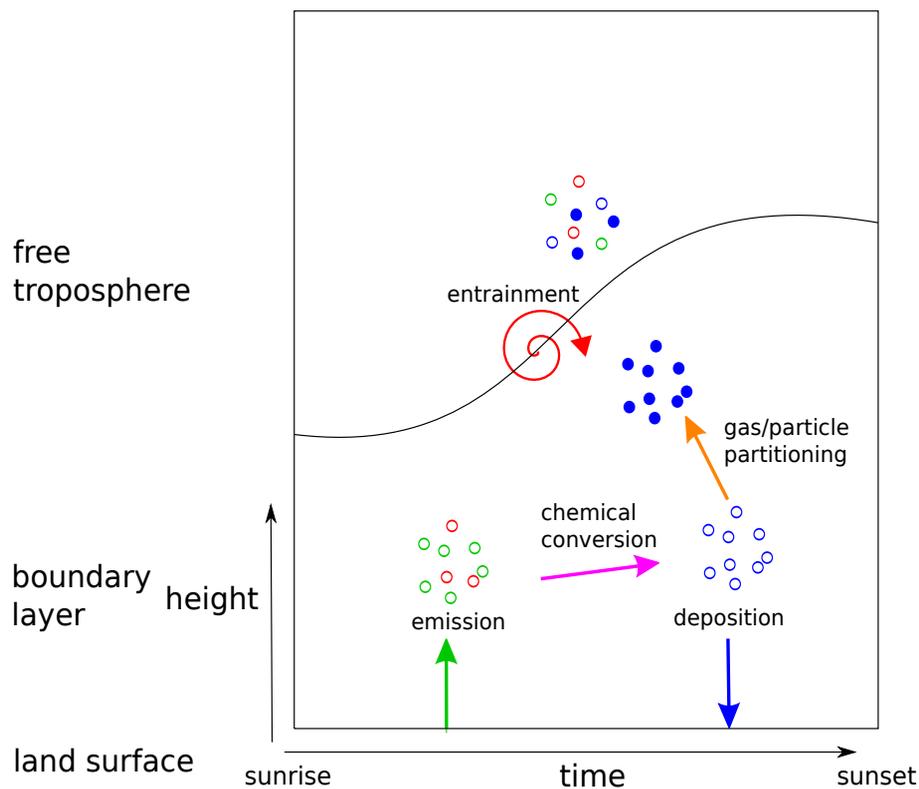Full Screen / Esc

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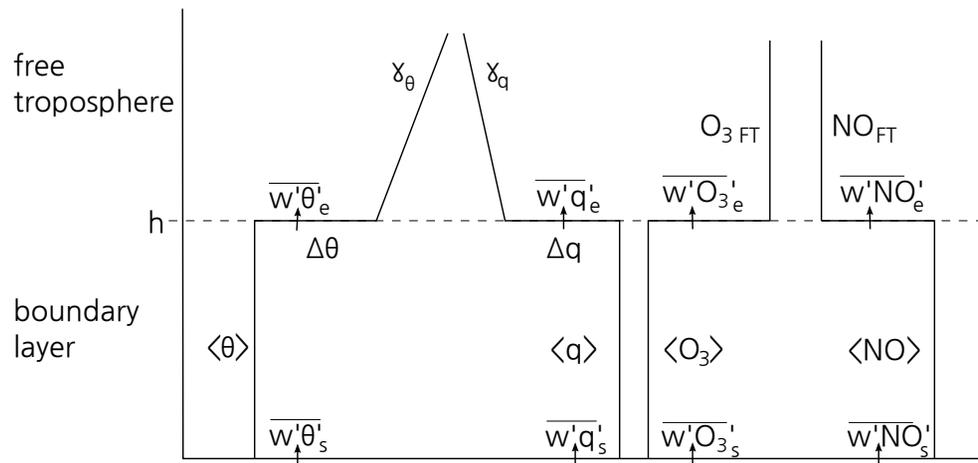
Figure 2. Typical mixed-layer profiles of potential temperature (θ), specific humidity (q) and chemical species (in this case O_3 and NO). Also surface and entrainment fluxes of heat, moisture, O_3 and NO are indicated; the arrows indicate the direction in which the flux is defined positively.

Figure 3. Scheme of the implementation of the VERTICO basemodel in the MESSy structure and the coupling to the MXL and other submodels through the MESSy interface.

Figure 4. Processes relevant to evolution of species concentrations in the boundary layer. Open and closed circles depict gas-phase and aerosol-phase species, respectively.

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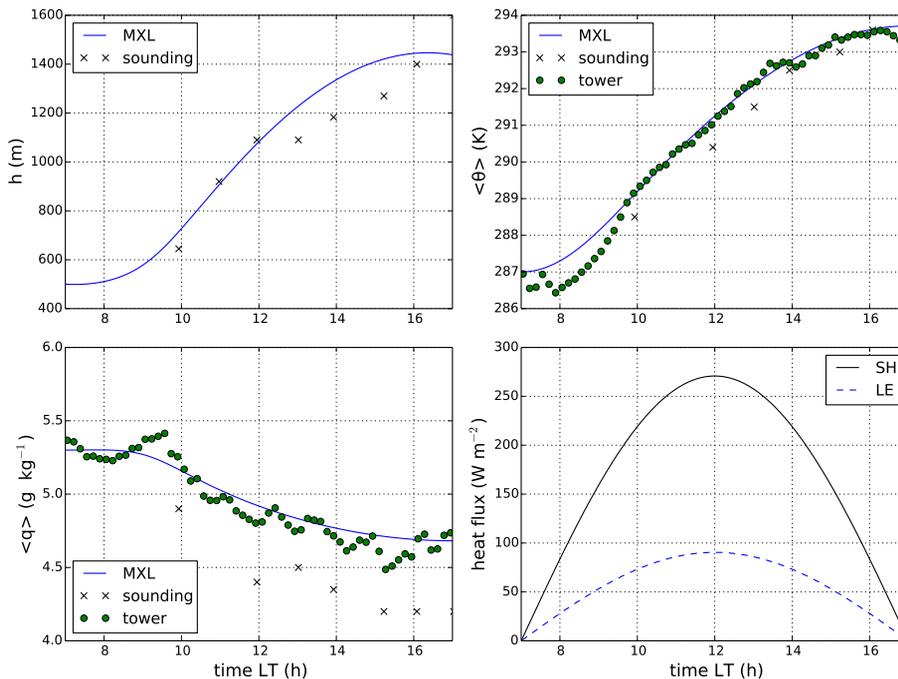
Figure 5. Diurnal evolution of observed and modelled **(a)** mixed-layer height (h), **(b)** mixed-layer potential temperature (θ), **(c)** mixed-layer specific humidity (q) and **(d)** prescribed sensible (SH) and latent (LE) heat fluxes.

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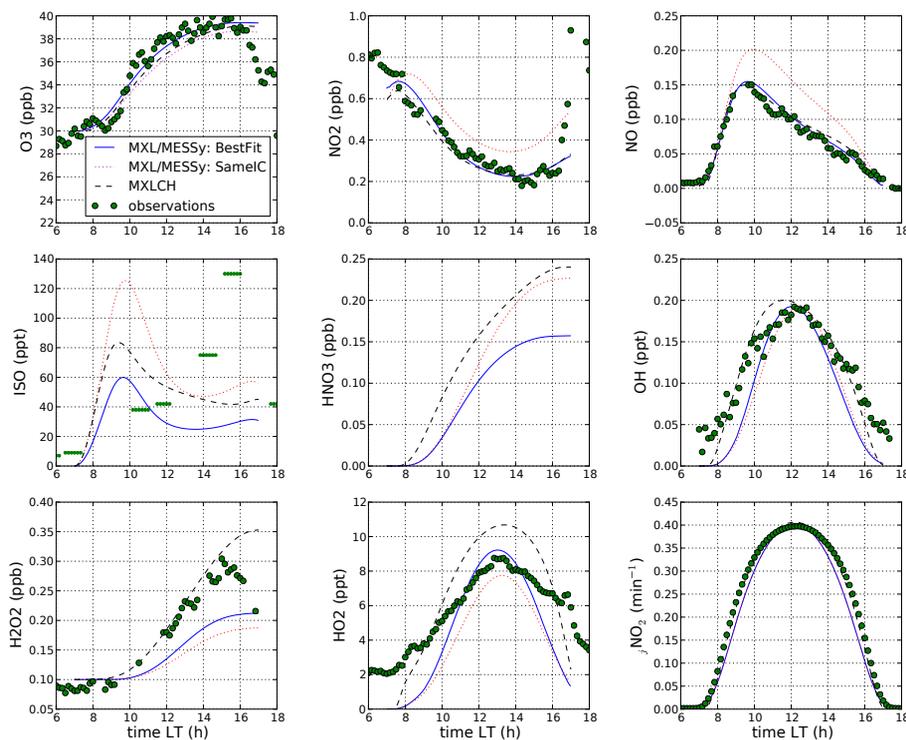
Figure 6. Observed and modelled diurnal evolution of gas-phase chemistry and NO_2 photolysis rate for the DOMINO case, comparing MXL/MESSy and MIM2 chemistry with MXLCH and reduced chemistry. Results for both the SameIC and the BestFit runs with MXL/MESSy are shown.

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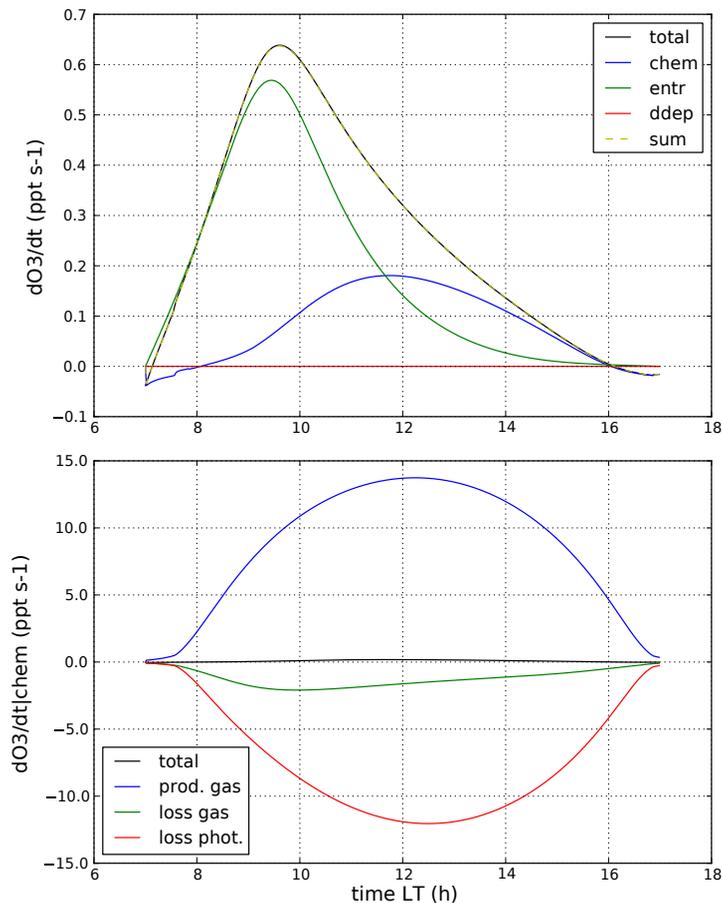Printer-friendly Version

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Figure 7. O_3 budget for the DOMINO case showing the contributions of **(a)** the processes entrainment, chemistry and dry deposition and **(b)** the chemical production and loss with the chemical loss split up in a gas-phase destruction and a photolysis term.