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Spatial and temporal variability of ozone deposition

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Abstract. Soil moisture and ozone deposition velocity under continental climate conditions were estimated using a newly developed algorithm. The relationship between soil moisture and deposition velocity was investigated and analysed. These results emphasize the importance of a sophisticated parameterization of soil moisture in surface-atmosphere interaction processes.

1 Introduction

In the last few years, many researches have pointed out the differences between concentration- and flux-based indices that can be applied for the characterization of effective ozone load (Musselman et al., 2006; Paoletti and Manning, 2007). New indices have been introduced which can more effectively describe the actual destructive effects of ozone. These indices can be estimated using deposition models. In such models, the ozone flux is controlled by ozone concentration and deposition velocity. For the purpose of calculating the deposition velocity, a high resolution deposition model was developed and tested over a continental region (Lagzi et al., 2004; 2006; Mészáros et al., 2006, 2009a). Previous investigations and sensitivity analysis (Mészáros et al., 2009a, b) have shown that in the summer period, the soil moisture could be a crucial stress factor in the deposition processes. Therefore a newly developed, more detailed water-balance module was adapted for use in our deposition model.

The main goal of this study is to present the temporal and spatial variability of ozone deposition velocity under continental climate conditions and to reveal the relationship of the deposition velocity with soil moisture.

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2 Deposition model

Ozone deposition velocity was estimated using the resistance method. In this model, the deposition velocity is defined as the inverse of the sum of the atmospheric and surface resistances:

$$v_d = (R_a + R_b + R_c)^{-1}, \tag{1}$$

where R_a, R_b, and R_c are the aerodynamic resistance, the quasi-laminar boundary layer resistance, and the canopy resistance, respectively. The canopy resistance is parameterized using stomatal, cuticular and surface resistance terms and depends on both meteorological and soil data and physiological plant characteristics. The ozone flux through the stomata can be depressed or sometimes fully blocked by high or low temperature, high vapour pressure deficit, and low soil moisture. Details of the deposition model are described in Mészáros et al. (2009a).

In this study, the daytime (12:00 UTC) deposition velocity was calculated for three summer periods (from 1 April to 30 September 1998 and 2007, and from 1 April to 31 August 1999).

The input meteorological datasets in 0.1×0.15 degrees spatial resolution were taken from the ALADIN meso-scale limited area numerical weather prediction model. These data were interpolated to a finer spatial resolution grid (0.025×0.0375 degrees, about 2.5×2.5 km). According to the Land Use Categories (LUC) used in ALADIN model, eight different vegetation types (grass, agricultural land, orchard, mixed agricultural land and forest, coniferous forest, deciduous forest, mixed forest and moorland) in addition to water and built-up areas were distinguished. Calculations were performed for seven soil types (sand, sandy loam, loam, clay loam, clay, peat and coarse frame). Root-zone soil moisture was estimated by a prognostic bucket model with a daily time step.

3 The water-balance module

The soil moisture is estimated in a bucket which depth is taken to be the root zone depth (z_r), the lateral movement and motion of the water into and from the lower layers are neglected. A root zone depth of 1 m was chosen for all soil types except for peat and coarse frame which were set at 0.2 m in the model. The soil moisture (θ) of the following time step (i+1th day) is calculated by the actual (i-th) daily values (Mintz and Walker, 1993):

$$\theta_{i+1} = \theta_i + (P_i - I_i) - ET_i, \tag{2}$$

where P, I and ET are the precipitation, the interception and the evapotranspiration, respectively.

The soil moisture is determined by the volumetric water-holding properties of the soil.

In the water-balance module, the evapotranspiration (ET) is calculated as the sum of evaporation of the bare soil (E_{dir}) and the transpiration of the vegetation (E_t):

$$ET = E_{dir} + E_t, \tag{3}$$

$$E_{dir} = E_p\,(1 - \mathrm{veg})\,\beta_1, \quad \text{and} \quad E_t = E_p veg \beta_2, \tag{4}$$

where E_p is the potential evapotranspiration, which is calculated by Penman-based approach, veg is the percentage distribution of vegetation, which ranges between 0 and 1, β_1 and β_2 are functions of the soil moisture and canopy resistance, respectively. E_p and β functions are calculated after Chen and Dudhia (2001), the soil parameters using for the estimation of β_1, as wilting point, and field capacity soil moisture content are parameterized after Ács (2003). The ratio of vegetation coverage (veg) for every cell was determined based on dataset of ALADIN model. For every vegetation type in a given cell the same vegetation fraction value was assumed.

The precipitation, which reaches the soil of a vegetated surface, is reduced by the amount of water intercepted by the canopy (I). The amount of the intercepted water on wet days (if there is precipitation) is estimated by the following relation:

$$I = S_m \mathrm{LAI} \tag{5}$$

where LAI is the leaf area index [$\mathrm{m^2\,m^{-2}}$], S_m is the maximum water storage capacity per unit leaf area index (0.2 mm in the model). The upper limit of the amount of the intercepted water is the daily precipitation. Because the water balance model is in a daily time step, it is assumed that the intercepted water is totally evaporated during the day. Therefore, the evaporation of the wet canopy is equal to the daily interception.

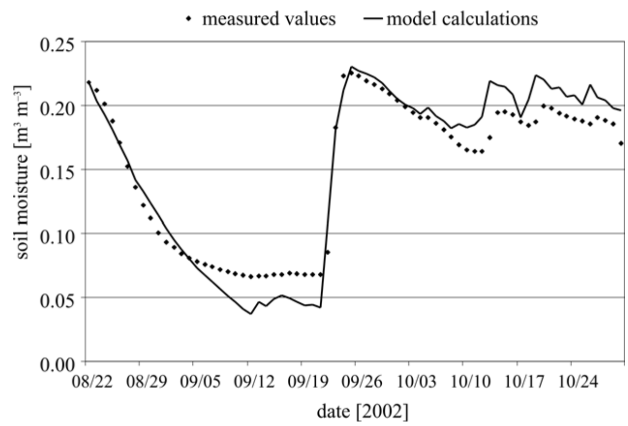

Figure 1. Measured and modelled soil moisture in the upper soil layer in Bugacpuszta.

4 Results

4.1 Comparison with measurements

The calculations of the water-balance module were verified with measurements in Bugacpuszta, Hungary (φ=46°40′ N, λ=19°33′ E, h=110 m). Here the soil type is sand and the land use category is grass. Measurements were carried out by a Campbell CS615 TDR sensor in the upper layer of soil. Measured data were available from 22 August to 30 October 2002. Daily average values were calculated from measurements at 12:00 and 00:00 UTC. Root-zone depth in the model was chosen to 0.25 m.

Figure 1 represents the calculated and the measured soil moisture data from the above mentioned period. The model underestimated the soil moisture in the dryer period (September) and slightly overestimated it in wet period (October). The reason of these discrepancies may be due to neglecting of horizontal and vertical water movements outside of the thin bucket. However there is a good correlation between measured and modelled values and the dynamics of soil moisture (quick growth after precipitation, and the degree of soil desiccation) can be traced with the model.

4.2 Model results

Figure 2 shows the spatial distribution of monthly averages of soil moisture and 12:00 UTC ozone deposition velocity values in July 1998, 1999 and 2007. Soil moisture patterns following the spatial distribution of soil types, and the different weather situations can cause differences in soil moisture among each year for the same period. In 1998 and 2007, the summer was very hot, although high monthly precipitation was observed in the summer of 1998 and 1999, while 2007 was a dryer year. Therefore due to the higher evapotranspiration and lower amount of precipitation, the soil moisture in July was the lowest in 2007. The soil water deficiency can

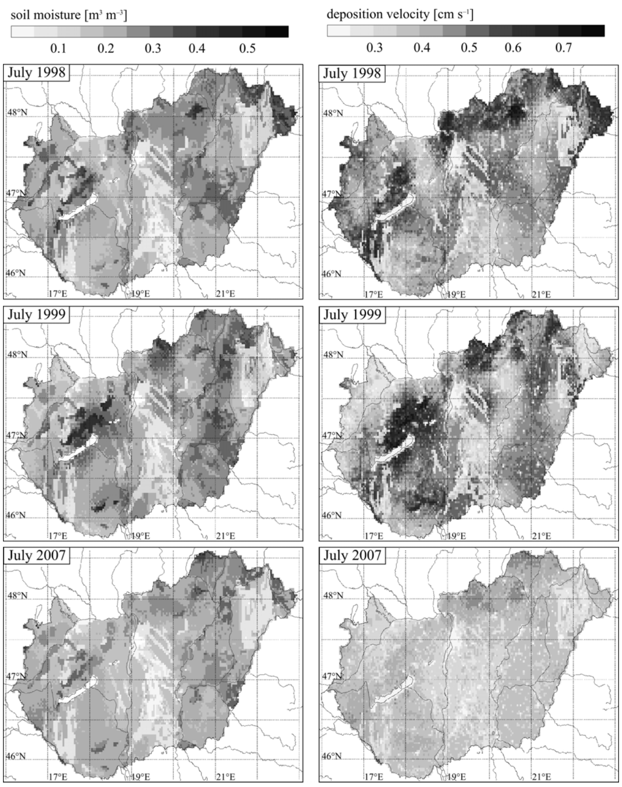

Figure 2. Soil moisture and ozone deposition velocity fields in July of 1998, 1999 and 2007.

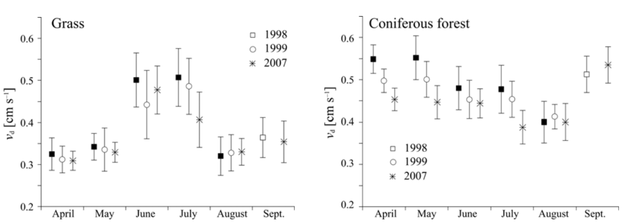

Figure 3. Deposition velocities (average and standard deviation) over grass and coniferous forest.

strongly reduce the stomatal conductance and so the ozone deposition through it. Therefore, there is a good correlation between the spatial and temporal distribution of soil wetness and deposition velocity fields.

Temporal variability of 12:00 UTC ozone deposition velocity during three summer periods are presented in Fig. 3 for grass and for coniferous forest. According to different plant physiology and characteristics, there are significant differences among deposition velocities in each month even as over each surface type. Due to the plant growth (increasing leaf area index), and the optimal environmental conditions for vegetation (higher temperature together with sufficient soil water content), generally higher values occur in June and July for grass. However, for coniferous forest, higher deposition velocity values were obtained in spring. In this case there are no significant changes in leaf area indices between each period, at the same time the lower temperature in spring is more favourable for the stomatal uptake of this type of vegetation. Decreasing soil water content (due to the warmer period of the year without precipitation) in August (or in July in 2007) decreased the deposition velocities in all cases. In contrast of this, in September, the values were raised because the soil water content was increased again.

These results emphasize the important effects of soil moisture in the surface-atmosphere interactions, especially in deposition processes.

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