



# Plant water-stress parameterization determines the strength of land–atmosphere coupling



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## ABSTRACT

Land-surface models used in studies of the atmosphere and vegetation during droughts usually include an underlying parameterization that describes the response of plants to water stress. Here, we show that different formulations of this parameterization can lead to significant differences in the coupling strength (i.e. the magnitude of the carbon and water exchange) between the land surface and the atmospheric boundary layer (ABL). We use a numerical model that couples the daytime surface fluxes typical for low vegetation to the dynamics of a convective ABL, to systematically investigate a range of plant water-stress responses. We find that under dry soil conditions, changing from a sensitive to an insensitive vegetation response to water stress has the same impact on the land–atmosphere (L–A) coupling as a strong increase in soil moisture content. The insensitive vegetation allows stomata to remain open for transpiration (+150 W m<sup>-2</sup> compared to the sensitive one), which cools the atmosphere (–3.5 K) and limits the ABL growth (–500 m). During the progressive development of a dry spell, the insensitive response will first dampen atmospheric heating because the vegetation continues to transpire a maximum of 4.6 mm day<sup>-1</sup> while soil moisture is available. In contrast, the more sensitive vegetation response reduces its transpiration by more than 1 mm day<sup>-1</sup> to prevent soil moisture depletion. But when soil moisture comes close to wilting point, the insensitive vegetation will suddenly close its stomata causing a switch to a L–A coupling regime dominated by sensible heat exchange. We find that in both cases, progressive soil moisture depletion contributes to further atmospheric warming up to 6 K, reduced photosynthesis up to 89%, and CO<sub>2</sub> enrichment up to 30 ppm, but the full impact is strongly delayed for the insensitive vegetation. Then, when we analyze the impact of a deviation of the modeled large-scale boundary conditions (e.g. subsidence, cloud cover, free-troposphere lapse rates, etc.) from their true state during a drought, we find that the two coupled systems (with a sensitive or insensitive vegetation) respond much differently to the generated atmospheric warming, this due to the difference in the basic surface coupling regime (coupled vs. uncoupled). This is of importance for the simulation of heat waves and meteorological droughts, as well as carbon-climate projections, as we show the predictive skill of coupled models is tied to the underlying vegetation response to water stress.

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## 1. Introduction

Diurnal land–atmosphere (L–A) interactions have been shown to impact cloud formation (Vilà-Guerau de Arellano et al., 2012; Vilà-Guerau de Arellano et al., 2014; Ek and Holtslag, 2004), precipitation (Santanello et al., 2013), as well as the build up of heat waves and droughts (Miralles et al., 2014; Teuling et al., 2010). Over

vegetated surfaces, the diurnal cycles of carbon, water and energy are coupled at the surface through stomatal control (Berry et al., 2010; Leuning et al., 1995; Collatz et al., 1991; Jarvis, 1976) and at the top of the atmospheric boundary layer (ABL) through the entrainment of air from the free troposphere (McGrath-Spangler and Denning, 2010; van Heerwaarden et al., 2009). The exchange of carbon for example affects both the CO<sub>2</sub> mole fractions (Combe et al., 2015; Pino et al., 2012) and the amount of carbon stored in vegetation. Little attention has been given so far to the impact of conditions in the free troposphere and upper ABL on surface carbon exchange. However, through the vegetation response to atmospheric conditions, important variables such as the net

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primary production (NPP) or the surface water-use efficiency can be under strong atmospheric control.

In the diurnal L–A system, upper-atmosphere and surface processes typically together determine conditions in the ABL. At the surface both plant phenology (Richardson et al., 2013; Peñuelas et al., 2009) and soil moisture (Seneviratne et al., 2010; Koster, 2004) have been shown to be key drivers of the surface fluxes of water and CO<sub>2</sub>, as well as the surface energy balance. Among these surface drivers, the impact of plant water-stress response on the strength of the L–A coupling remains an open question. Observations show that water-stress responses can differ between plant species and varieties (Calvet et al., 2004; Tardieu and Simonneau, 1998) and various plant water-stress parameterizations are now used in land-surface models. But there is little observational evidence to support their realism over the wide range of conditions they are applied to (Powell et al., 2013).

While a few attempts to mechanistically represent plant water stress have been made (Verhoef and Egea, 2014), the majority of land-surface models use a simple parameterization of plant water stress. They most often scale down net assimilation, stomatal and/or mesophyll conductance, or other photosynthesis parameters such as the maximum carboxylation rate. The scaling factor depends on the levels of soil moisture relative to field capacity and wilting point, and various response curve shapes are used ranging from linear (e.g. the CTESSEL and JULES models in Boussetta et al., 2013; Best et al., 2011) to highly non-linear (e.g. SiB3 and ORCHIDEE in Baker et al., 2008; Krinner et al., 2005). Observational studies suggest that the linear response might not be a realistic assumption for C<sub>4</sub> crops like maize (Verhoef and Egea, 2014), and crop models often use a strongly non-linear response to water stress by downregulating photosynthesis with the ratio of actual to potential transpiration (e.g. GECROS, WOFOST, SUCROS in Yin and van Laar, 2005; Van Ittersum et al., 2003). Powell et al. (2013) showed that the shape of the water-stress response function can make large differences for the simulation of diurnal and seasonal surface CO<sub>2</sub> fluxes under dry soil conditions. van der Molen et al. (2011) and Combe et al. (2015) also speculated about such an impact on modeled L–A interactions. An extensive exploration of the effects of plant water-stress parameterizations on the coupled L–A system has yet to be performed.

The primary aim of our study is therefore to systematically assess the impact of differing plant water-stress parameterizations on the strength of the L–A coupling. We perform a sensitivity analysis of the coupled L–A system using a diurnal L–A modeling framework, called the MXL–A–g<sub>s</sub> model, as a continuation of the studies of Combe et al. (2015), van Heerwaarden and Teuling (2014), and van Heerwaarden et al. (2009). In this work we introduce an adjustable plant water-stress function into the model, allowing us to explore a wide range of water stress responses. Our model represents the daytime surface fluxes of carbon, water, and energy coupled to the dynamics of a convective boundary layer. Its strength is to include the essential diurnal processes of the L–A in a concise manner. With this system we address three research questions:

1. What is the impact of changing the plant water-stress response function from a sensitive to an insensitive formulation on the simulated atmospheric boundary-layer of a coupled land–atmosphere (L–A) system?
2. How does this choice of plant water-stress response function affect the development of a dry spell over time?
3. How are the interactions of the diurnal L–A system, as well as its sensitivity to model errors, affected by the choice of plant water-stress response function?

We base our sensitivity analysis on a control case that represents a grown maize crop field during a sunny summer day in the

Netherlands. This control case has been validated with observations and discussed in Combe et al. (2015), and is generally representative of short vegetation. In Section 3.1, we first modify the conventional representation of soil water stress (i.e. the linear response) in our model, and explore the impact of a range of other, non-linear water-stress response curves on the mixed-layer budgets of CO<sub>2</sub>, water, and heat. We next turn to the development of a dry spell in Section 3.2, and investigate the impact of two different plant water-stress responses on the coupled L–A system, during the dynamic soil drying of a three-week period. Finally, we show in Section 3.3 that the impact of errors in the simulated early-morning temperatures, cloud cover conditions and large-scale air motions can be large, small, and even of opposite sign depending on the choice of water-stress response function. The implications of these findings are discussed in Section 4.

## 2. Research strategy

### 2.1. Conceptual view of the land–atmosphere system

Fig. 1 presents a schematic view of a coupled land–ABL system. This figure was extended to include the carbon cycle from the work of van Heerwaarden and Teuling (2014) and van Heerwaarden et al. (2009) (hereafter H14 and H9), who focused on the water and heat cycles only. Fig. 1a represents a well-watered short vegetation surface coupled with a convective ABL under no subsidence. Fig. 1b then shows the changing interactions under large-scale subsidence.

In H9, the authors presented three negative feedback loops that regulate evapotranspiration (LE) under well-watered conditions: a heating feedback, a drying feedback and a moistening feedback (see the blue shaded part of Fig. 1). We refer to H9 for their full description. This description of the feedbacks excluded the possibility for plants to regulate their transpiration flux. In H14, the authors thus added the concept of an adaptable surface conductance ( $g_s$ ), which could additionally modify LE (see the brown shaded part of Fig. 1a). Knowing that this representation of the water and heat cycles is more correct for vegetated lands, we expanded this picture to carbon dioxide as it is controlled by many of the same processes, as shown in Fig. 1a.

From H14, we first added the net surface flux of CO<sub>2</sub> or net ecosystem exchange (NEE). This variable is determined by the net primary production (NPP) from plants and the heterotrophic respiration ( $R_{het}$ ). While NPP is controlled both by the canopy conductance ( $g_s$ ) and by the stomatal demand for CO<sub>2</sub> ( $c_i/c$  ratio),  $R_{het}$  is known to be a function of soil temperature (Karhu et al., 2014; Davidson and Janssens, 2006). In that way,  $g_s$  acts as a coupling point for the carbon cycle, the water, and the heat cycles previously described by H9 and H14 (see Fig. 1a). This coupling point is directly controlled by surface drivers such as the soil moisture index (SMI) and the leaf area index (LAI) (Ronda et al., 2001).

From H14, we then added the entrainment flux of CO<sub>2</sub> at top of the ABL and the mixed-layer CO<sub>2</sub> mole fraction. The latter is affected by all boundary fluxes of CO<sub>2</sub>, as well as the volume dilution due to the growth of the ABL during the day (Pino et al., 2012). Both the entrainment velocity ( $w_e$ ) and volume dilution are related to changes in  $h$ , which is determined by the amount of energy that is partitioned into sensible heat at the surface and by subsidence (see Eqs. (1 and 2)). In that way,  $h$  acts as the second coupling point of the L–A system as it joins the carbon, water and heat water cycles at top of the ABL (see Fig. 1a). This second coupling point is directly influenced by the free-tropospheric processes and ABL-top conditions, such as subsidence ( $w_s$ ) and the free-troposphere temperature lapse rate ( $\gamma_\theta$ ) (Ek and Holtlag, 2004; Williams et al., 2011).

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