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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^{−2} 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^{−1} while soil moisture is available. In contrast, the more sensitive vegetation response reduces its transpiration by more than 1 mm day^{−1} 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 CO₂ 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 CO₂ 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 CO₂, 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 C₄ 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 CO₂ 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_s 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 CO₂, 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 CO₂ 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 CO₂ (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 CO₂ at top of the ABL and the mixed-layer CO₂ mole fraction. The latter is affected by all boundary fluxes of CO₂, 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 (γ_θ) (Ek and Holtslag, 2004; Williams et al., 2011).

Fig. 1. Interactions of the vegetated land-ABL system, based on the studies of H9 and H14. Each arrow represents the impact of an increase in variable “x” on another variable “y”. Boundary and initial conditions explored in this study are presented in squared boxes. Model variables (encircled) are from the surface to the top of the ABL: the surface conductance (g_s), net surface shortwave radiation (SW_{net}), sensible heat flux (SH), latent heat flux (LE), net primary production (NPP), heterotrophic respiration (R_{het}), the internal to external CO_2 concentration ratio (c_i/c), the surface vapor pressure deficit (VPD), the ABL potential temperature (θ), specific humidity (q), CO_2 mole fraction (c), and the ABL height (h). Changes in h are related to changes in ABL volume (Vol.) and entrainment flux (Entr.) and these are separated when divergent.

As a result of the coupling of the carbon, water and heat cycles at the surface (g_s) and at the top of the ABL (h), it is necessary to consider the L-A system as a whole to understand variations in specific parts of the carbon, water or heat cycles.

2.2. Description of the MXL-A- g_s model

We focus on the case of a well-grown short vegetation surface that interacts with a convective ABL, in the mid-latitudes. We choose to model the interactions of the coupled land-atmosphere system with a simple process-based modeling framework: the MXL-A- g_s model. This model consists of a zero-order model for the convective ABL (Tennekes, 1973), coupled with a simple radiation scheme using a net sky transmissivity (Stull, 1988), a land-surface scheme using a stomatal conductance approach (Jacobs et al., 1996), and a modified two-layer force-restore soil model (Duykerke, 1991; Noilhan and Planton, 1989). We introduce hereafter the equations of the MXL-A- g_s model that describe the ABL dynamics, the mixed-layer CO_2 mole fraction, and the net surface flux of CO_2 . In addition, we refer to Appendices A and B for the complementary equations describing the heat and water cycles and the full surface conductance formulation.

2.2.1. Atmospheric convective boundary layer

The mixed-layer model (MXL) calculates the diurnal evolution of the atmospheric state variables and CO_2 in a convective boundary layer. The mixed-layer state variables (also called “bulk” variables) do not vary with height due to the intense convective turbulent mixing. First, under convective conditions the diurnal evolution of the ABL height (h) can be described as follows:

$$\frac{dh}{dt} = w_e + w_s, \quad (1)$$

$$w_e = -\frac{(w'\theta'_v)_e}{\Delta\theta_v} = \beta_{\theta_v} \frac{(w'\theta'_v)_s}{\Delta\theta_v}, \quad (2)$$

with w_e the entrainment vertical velocity, w_s the negative vertical subsidence velocity to include the effect of large-scale subsidence motions, β_{θ_v} the ratio between the entrainment turbulent flux of virtual heat ($(w'\theta'_v)_e$) and the surface turbulent flux of virtual heat ($(w'\theta'_v)_s$), and $\Delta\theta_v$ the virtual potential temperature jump at the inversion height, as we assume that the inversion layer is infinitesimally thin. With this formulation, entrainment is thus parameterized as function of $(w'\theta'_v)_s$, and turbulent motions do not need to be resolved.

To complete Eqs. (1) and (2), we need a budget equation for the ABL virtual potential temperature (θ_v) and for the evolution of the virtual potential temperature jump at the inversion height ($\Delta\theta_v$), which we present in Appendix A. The diurnal evolution of all ABL state variables (i.e. the potential temperature θ , the specific humidity q , and the wind speed u) and of the atmospheric compounds (e.g. CO_2) can be expressed as a function of their boundary fluxes (i.e. the surface, entrainment and advection fluxes) and of the ABL volume. For the CO_2 mole fraction (c), the physical interpretation is as follows if we neglect the horizontal advection of CO_2 :

$$\frac{dc}{dt} = \frac{1}{h} \times \left(\underbrace{\overline{w'c_s}}_{\text{Surface turbulent flux of CO}_2} - \underbrace{\overline{w'c_e}}_{\text{Entrainment turbulent flux of CO}_2} \right), \quad (3)$$

$$= \frac{1}{h} \times \left(\frac{\text{NEE}}{\rho_{\text{air}}} + w_e \Delta c \right),$$

with NEE the net ecosystem exchange in $\text{mg}_{\text{CO}_2} \text{ m}^{-2} \text{ s}^{-1}$, ρ_{air} the density of the air, and Δc the CO_2 jump at the inversion height. Note that h controls the dilution of the CO_2 mole fraction in Eq. (3). The gradient Δc evolves in time due to the growth of the ABL into the free troposphere and to the evolution of the mixed-layer CO_2 . This expression reads:

$$\frac{d\Delta c}{dt} = \gamma_c w_e - \frac{dc}{dt}, \quad (4)$$

with γ_c the free-troposphere lapse rate of CO_2 . At Eq. (3), we represent the entrainment as the entrainment velocity (see Eq. (2)) times the instantaneous gradient of CO_2 at top of the ABL (see Eq. (4)). We refer to Appendix A for similar atmospheric equations to Eqs. (3) and (4) for θ and q .

2.2.2. Land-surface

The A- g_s model expresses the net surface fluxes of heat (SH), water (LE) and CO_2 (NEE) as the temperature, moisture and CO_2 gradients times their corresponding surface conductance, except for NEE where we have to add the contribution of the heterotrophic respiration (R_{het}):

$$\text{NEE} = \overbrace{\rho_{\text{air}} g_{s,c} (c_i - c)}^{\text{NPP}} + R_{\text{het}}, \quad (5)$$

$$g_{s,c} = \frac{1}{r_a + r_c}, \quad (6)$$

with $g_{s,c}$ the surface conductance for CO_2 transfer, c_i the CO_2 concentration in the plant sub-stomatal cavities, r_a the aerodynamic resistance, r_c the canopy stomatal resistance to CO_2 transfer. R_{het} is calculated as the product of a reference respiration constant with a soil water-stress factor and an exponential function of the surface temperature. We refer to Appendix A for the complementary equations for the surface heat and moisture fluxes.

We describe the aerodynamic resistance as an inverse function of wind speed and a drag coefficient for heat, which is a function of the surface roughness and atmospheric stability. This formulation is issued from the Monin–Obukhov similarity theory. At the origin of the r_c formulation, Jacobs et al. (1996) expresses the net leaf assimilation A_n (i.e. the net flux of CO_2 through the stomata of a leaf) as the sum of gross leaf assimilation (A_g) and leaf dark respiration (R_{dark}):

$$A_n = \frac{1}{r_{l,c}} (c - c_i) = \beta A_g^* - R_{\text{dark}}, \quad (7)$$

with $r_{l,c}$ the leaf stomatal resistance to CO_2 transfer, β the plant water stress factor and A_g^* the unstressed gross leaf assimilation. R_{dark} is parameterized as 11% A_g^* . Note that A_n is not equal to NPP

as (a) the CO_2 still has to transfer through the leaf boundary-layer (i.e. experience r_a) and (b) the CO_2 assimilation has to be upscaled to the canopy level, before we obtain NPP. Jacobs et al. (1996) uses an empirical response function of A_g^* to light, CO_2 and temperature, which allows us to rewrite Eq. (7) as follows:

$$A_g^* = (A_m + R_{\text{dark}}) \left[1 - \exp \left(- \frac{\text{LUE } K_x \text{ PAR}_t e^{-K_x L}}{(A_m + R_{\text{dark}})} \right) \right], \quad (8)$$

with A_m the primary productivity under high light conditions, LUE the light-use efficiency constant, K_x the light extinction coefficient, PAR_t the photosynthetically active radiation at an instant t , and L the integrated LAI at a specific depth within the canopy (we integrate from 0 at top of the canopy to the total LAI at bottom). A_m is a function of the internal CO_2 concentration c_i , of the CO_2 compensation point Γ , of the maximum productivity $A_{m,\text{max}}$, and of the mesophyll conductance g_m , all later three parameters being dependent on the leaf temperature (T_{sk}). Jacobs et al. (1996) also found that the c_i/c ratio is a function of the vapor pressure deficit (VPD) at the leaf surface:

$$\frac{c_i - \Gamma}{c - \Gamma} = f_0 \left(1 - \frac{\text{VPD}}{\text{VPD}_0} \right) + f_{\text{min}} \frac{\text{VPD}}{\text{VPD}_0}, \quad (9)$$

with f_0 and f_{min} respectively the maximum and minimum values of $(c_i - \Gamma)/(c - \Gamma)$, and VPD_0 the value of VPD at which the stomata close. Combining Eqs. (7)–(9), the leaf stomatal resistance for CO_2 transfer $r_{l,c}$ can be rewritten as:

$$\frac{1}{r_{l,c}} = g_{\text{min},c} + \frac{a_1 A_g^* \beta}{(c - \Gamma)(1 + \text{VPD}/\text{VPD}_*)}, \quad (10)$$

with $g_{\text{min},c}$ the cuticular conductance for CO_2 transfer. a_1 is $1/(1 - f_0)$ and VPD_* is $\text{VPD}_0/(a_1 - 1)$. To upscale $r_{l,c}$ from the leaf to the canopy level (i.e. to obtain r_c), we integrate Eq. (10) over LAI (see the full derivation in Appendix B). Note that our model can simulate both the C_3 and C_4 photosynthetic pathways by adapting the photosynthesis constants (see Ronda et al., 2001; Jacobs et al., 1996). In A- g_s , the C_3 and C_4 pathways differ on how they respond to light, temperature and VPD. Here we use the parametrization for the C_4 pathway to represent a maize case. In the end, and to summarize, the A- g_s formulation from Jacobs et al. (1996) expresses r_c as a function of PAR, T_{sk} , the c_i/c_a ratio, VPD, the plant water stress (β , our main point of investigation in this study) and LAI. Note that in the rest of the manuscript, we analyze the variations of g_s (i.e. the surface conductance for water transfer) which is equal to $g_{s,c}/1.6$. The variations in g_s thus reflects the variations in both r_a and r_c .

2.2.3. Soil water budget

We use the soil water budget from the two-layer force-restore model of Noilhan and Planton (1989). It describes water flow as follows:

$$\frac{dw_1}{dt} = \underbrace{\frac{C_1}{\rho_w d_1} \left(Pr - \frac{LE_{\text{evap}}}{L_v} \right)}_{\text{input-output atmosphere}} - \underbrace{\frac{C_2}{\tau} (w_1 - w_{1eq})}_{\text{diffusion into the ground}}, \quad (11)$$

$$0 \leq w_1 \leq \text{wsat},$$

$$\frac{dw_2}{dt} = \frac{1}{\rho_w d_2} \left(Pr - \frac{LE_{\text{transpi}}}{L_v} \right), \quad 0 < w_2 \leq \text{wsat}, \quad (12)$$

with w_1 and w_2 the volumetric soil moisture content of the first and second soil layers, ρ_w the density of liquid water, d_1 and d_2 the depths of the first and second soil layers, Pr the input of water from precipitation, LE_{evap} the output of water from soil evaporation, LE_{transpi} the output of water from plant transpiration, L_v the latent heat of vaporization, C_1 and C_2 are two coefficients related to the

Fig. 2. Range of soil water-stress (β) response functions modeled in this study. The soil water content is expressed on the x-axis with the soil moisture index (SMI, see Eq. (13)), thus wilting point and field capacity are respectively represented as the $\text{SMI} = 0$ and $\text{SMI} = 1$ dashed lines. Below wilting point, plants are incapable of taking water up from the soil. Above field capacity, the maximum amount of water is being stored in the soil pores. Plant water-stress occurs in between those two points.

Clapp and Hornberger parameterization (Clapp and Hornberger, 1978), w_{1eq} the soil moisture content at which the capillary forces balance the gravity forces and no diffusion of water occurs to the second layer, and τ the time-scale of that return to equilibrium.

The force-restore model neglects the diffusion of water between the second layer and the deeper – non-modeled – soil layers, and the diffusion of water from the second soil layer to the upper layer by capillarity. All soil evaporation is done from the first soil layer (first 10 cm), and all plant transpiration is taken from the second layer (lower 1.5 m). We thus neglect the roots water extraction from the upper superficial layer. We refer to Combe et al. (2015) for the constants of the soil model used in our study.

2.2.4. Updated plant water-stress parameterization

As mentioned in Section 2.2.2, in the A- g_s model, the plant water-stress factor (β) is directly multiplied with the leaf gross assimilation A_g^* . The standard β function of MXL-A- g_s is the linear one, and is equal to the soil moisture index (SMI) of the second soil layer:

$$\text{SMI} = \frac{w_2 - w_{wp}}{w_{fc} - w_{wp}}, \quad (13)$$

with w_{wp} and w_{fc} the volumetric water contents at wilting point and field capacity. To vary the curvature (C_β) of this response function we implemented a new exponential function suggested by Ronda et al. (2001):

$$\beta = \frac{1 - \exp(-P(C_\beta) \times \text{SMI})}{1 - \exp(-P(C_\beta))},$$

$$\text{with } P(C_\beta) = \begin{cases} 6.4 \times C_\beta & \text{for } 0 \leq C_\beta < 25\%, \\ 7.6 \times C_\beta - 0.3 & \text{for } 25 \leq C_\beta < 50\%, \\ 2^{(3.66 \times C_\beta + 0.34)} - 1 & \text{for } 50 \leq C_\beta \leq 100\%. \end{cases} \quad (14)$$

With Eq. (14), we vary the curvature of the response function (C_β) from 0 (sensitive) to 100% (insensitive) to obtain the range of β response functions plotted in Fig. 2. This range encompasses a large variety of β response functions currently used in land-surface models (e.g. see Powell et al., 2013). This range also corresponds to one half of the observed water-stress response spectrum, namely the anisohydric plant species. Anisohydric species do not close their stomata much under soil moisture depletion and thus cannot maintain their leaf water potential as they continue to transpire. If maize has been reported as an isohydric species by several studies (see e.g. Tardieu and Simonneau, 1998; Jones, 1998), Domec and Johnson

(2012) have shown that iso- and anisohydric behaviors could be displayed by the same species under differing environmental conditions. Note that at a fixed $\text{SMI} < 1$, we obtain a higher β value and thus higher g_s with the non-linear anisohydric responses.

2.3. Diagnostic variables

We calculate two supplementary diagnostic variables to help analyze the state of the diurnal L-A system. Firstly, we calculate the ratio of the entrainment to surface CO_2 tendencies (R_c), which quantifies how much of the total daytime CO_2 tendency is driven by entrainment, in comparison to the surface NEE:

$$R_c = \frac{\int_{t=6:00 \text{ UTC}}^{t=18:00 \text{ UTC}} (-\bar{w}'c'_e/h)dt}{\int_{t=6:00 \text{ UTC}}^{t=18:00 \text{ UTC}} (\bar{w}'c'_s/h)dt}. \quad (15)$$

A ratio of 1 means that the surface and ABL-top processes contributed equally to the daytime mixed-layer CO_2 tendency.

Secondly, we calculate the evaporative fraction (EF) at 14:00 UTC, which quantifies the partitioning of energy at the surface going into evapotranspiration:

$$\text{EF} = \frac{\text{LE}}{\text{LE} + \text{SH}}. \quad (16)$$

We calculate the 14:00 UTC value as the surface fluxes of water and heat are high and the ABL is fully convective.

2.4. Sensitivity analysis

In order to quantify the interactions between the land surface and the ABL, we design a sensitivity analysis with our MXL-A- g_s model. We use a recently published control case (Combe et al., 2015) that is based on observations of the soil, plants and ABL, over a maize field grown in Wageningen (the Netherlands) on 4 August 2007 (Jans et al., 2010). Using this comprehensive set of observations to design our control case gives us confidence that we are representing a realistic situation. The main initial and boundary conditions of our control case can be found in Table C.1, in Appendix. We first focus on the influence of the type of plant water-stress response function on the two key coupling points presented in Section 2.1. For this, we perform a series of sensitivity experiment where we simultaneously vary SMI from 0 to 1 and C_β from 0 to 100%, performing 100×100 cases. We then analyze the response of g_s , R_c , EF and h to these combined variations of SMI and C_β . The results of this series of experiments are presented in Section 3.1.

We then analyze the impact of the plant water-stress response on the time evolution of a dry spell (i.e. here: a 21-day continuous soil moisture depletion). We start two sensitivity experiments with the control case of Combe et al. (2015), only changing the curvature of the plant water stress response function. We use two extreme plant water stress responses: our most sensitive response ($C_\beta = 0\%$) and our most insensitive response ($C_\beta = 100\%$). Over a 21-day period, we transfer the soil moisture depletion caused by evapotranspiration (LE) from one day to the next, while we reinitialize all other variables identically to the control case. This is a similar approach to Daly et al. (2004), except we do not represent the nocturnal boundary-layer and nighttime surface fluxes. The results of this analysis are presented in Section 3.2.

Finally, because our previous experiment does not include variations in the daily atmospheric initial conditions that occur during a dry spell, we assess the impact of atmospheric model errors on the carbon and heat cycles. We define two new base cases that are identical to the one of Combe et al. (2015) except for their SMI ($= 0.2$) and the curvature of their plant water stress response function ($C_\beta = 0\%$ and 100%). These values generate the largest differences in plant water stress factor β (see Fig. 2) and surface coupling

Fig. 3. Diurnal cycles of c and θ for the dry soil ($\text{SMI} = 0.2$) base cases of the sensitive (DS) and insensitive (DI) vegetated surfaces.

Table 1
Explored model errors.

Variable [units]	Initial value	Change
w_s [s^{-1}]	0.7×10^{-5}	$+3.3 \times 10^{-5}$
γ_θ [K km^{-1}]	8	+3
cc [%]	22.5	−22.5
θ_0 [K]	286	+4

g_s between our two plant water stress responses. We illustrate the two generated base cases in Fig. 3, which we respectively name DS (dry soil – sensitive to water stress) and DI (dry soil – insensitive to water stress). Note that the diurnal cycles of CO_2 and temperature presented in Fig. 3 reach two characteristic values: a diurnal minimum for CO_2 ($c_{\min}$) and a diurnal maximum for temperature ($\theta_{\max}$). These are the two characteristic values that we hereafter focus our analysis on.

We select four initial and boundary conditions that vary from day-to-day during a dry spell: the subsidence velocity (w_s), the free-troposphere temperature lapse rate (γ_θ), the early-morning ABL potential temperature (θ_0), and the cloud cover (cc , see Fig. 1). For each of these variables we define a model error (see Table 1). Note that for θ_0 we create a set of consistent initial conditions for each run: (a) we conserve the temperature differences between the soil layers and the atmosphere, effectively shifting the entire temperature profile, and (b) we conserve the initial VPD of the base case (94%) by adapting the initial q . We subsequently apply all our

defined model errors first one-by-one, then all at once on our two base cases (DS and DI) and analyze their impact on $c_{\min}$ and $\theta_{\max}$. The result of these experiments is presented in Section 3.3.

3. Results

3.1. Coupling points

The atmospheric budget of CO_2 can switch from being dominated by surface fluxes to dominated by entrainment depending on the stomata aperture, which is controlled by the soil moisture level and the vegetation sensitivity to water stress. This is illustrated in Fig. 4a, which presents the response of the surface coupling point (g_s) and of the mixed-layer CO_2 budget (through R_c) to the assumed curvature of the plant water-stress response function (C_β on the y-axis) and soil moisture level (SMI on the x-axis). If we first focus on soil moisture at a given water-stress response ($C_\beta = 0$), we find two opposite regimes of the CO_2 budget indicated by two dots marked WS (wet soil – sensitive to water stress) and DS (dry soil – sensitive to water stress). The non-stressed vegetation of the WS case has a g_s of 12 mm s^{-1} which enables high photosynthesis rates and a large contribution of NEE to the CO_2 budget ($R_c = 0.5$) while the stressed vegetation of the DS case has a g_s of 3 mm s^{-1} , which almost fully inhibits photosynthesis and reduces the coupling between the vegetation and the atmospheric CO_2 budget ($R_c = 3$). This reduced coupling of the ABL carbon budget from the surface fluxes is important for the interpretation of observed mixed-layer CO_2 mole fractions under drought conditions as it is often assumed that CO_2 measurements in the ABL predominantly reflect vegetation carbon exchange – which is clearly not true in this example. Further reducing soil moisture below $\text{SMI} = 0.2$ would lead to an exponential R_c increase and even weaker surface coupling.

However, we can again increase the vegetation influence on ABL CO_2 at low soil moisture by changing the assumed vegetation sensitivity to water stress. In Fig. 4a, the point marked DI (dry soil – insensitive to water stress) is a simulation at the same soil moisture level as the DS point. It maintains high conductance, high carbon assimilation and evaporation rates, and thus conserves a strong influence of surface carbon exchange on atmospheric CO_2 . The R_c -value at this point is even below 0.5, while conductance remains close to 15 mm s^{-1} confirming that downregulation of photosynthesis has not yet started, in agreement with the $\text{SMI} = 0.2$ and $\beta = 0.95$ point in Fig. 2. We tentatively define the threshold of

Fig. 4. Values of key variables in the coupled land–atmosphere system as a function of soil moisture index (SMI , x-axis) and of curvature of the water-stress function (C_β , y-axis). Shown are (left) diurnal average g_s in gray scale with R_c in contours and (right) ABL height in gray scale and evaporative fraction in contours. We highlight where $R_c = 3$ with a thick dashed line, which corresponds roughly where $g_s = 3 \text{ mm s}^{-1}$. Three specific points are indicated: a wet soil case simulated with the sensitive water-stress response function (WS), a dry soil case simulated with the sensitive water-stress response function (DS), and a dry soil case simulated with an insensitive water-stress response function (DI).

Fig. 5. The effect of soil moisture depletion over a 21-day dry period under two different plant water-stress representations. Results using the sensitive response are shown with a gray background, while the results with an insensitive response are shown with a white background. Both simulations start at SMI = 0.55. Variables shown are (a) LE and SMI, (b) c_{min} and R_c , (c) SH and θ_{max} , (d) g_s and h . The surface coupling threshold for the CO₂ budget at $g_s = 3 \text{ mm s}^{-1}$ is shown by the thick dashed line in panel (b).

$g_s = 3 \text{ mm s}^{-1}$ shown as the thick dashed line in Fig. 4a as the point where we enter a very weakly coupled regime for CO₂ in the ABL. We note that this threshold shifts toward very low values of SMI if C_β (and thus β) increases in our analysis, and a similar effect occurs if we apply the linear β function on the mesophyll conductance instead of on the gross primary production, as done by Calvet et al. (1998) and as suggested by Keenan et al. (2010) (not shown here). This will be discussed further in Section 4.

The weakly coupled regime of surface CO₂ is always correlated with high ABL growth and high entrainment of dry and warm air, which will ultimately feed back on surface conditions. This is illustrated in Fig. 4b that presents the response of the coupling point at the top of the ABL (h) and of the evaporative fraction (EF) to the variation of the curvature of the water-stress response function (y -axis) and SMI (x -axis). We first see the complementarity of Fig. 4a and b, and how low g_s switches the surface energy budget toward the release of sensible heat. From the WS to the DS case, the 10 mm s^{-1} drop in g_s causes EF to decrease from 0.74 to 0.39, and thus the ABL height to increase from 1131 m to 1494 m. Then from the DS to the DI case, the 12 mm s^{-1} increase in g_s reverts EF and h back to 0.76 and 1120 m. These substantial differences in ABL height and surface energy balances are relevant for both numerical weather prediction and climate modeling, as simulated temperatures near the surface as well as cloud formation and precipitation rates now depend on plant water-stress response.

We find that water-stress sensitive and insensitive vegetated surfaces can behave identically from a carbon, water, and heat exchange point of view, despite a large difference in their hydrological state. In Fig. 4a, the vegetation of the DI case has the same g_s ($12\text{--}15 \text{ mm s}^{-1}$) as the vegetation of the WS case. In Fig. 4b this leads them to have a similar surface energy budget (EF ≈ 0.75) and ABL growth ($h \approx 1130 \text{ m}$). Specifically for coupled land-surface simulations such as done in climate models, it is important to realize that in addition to their effects on local conditions mentioned above, the differing sensitivities ultimately feed back on the larger scale atmospheric state. This shows that the choice of plant water stress representation in land-surface models could be an important area of model improvement.

3.2. Temporal evolution of a dry spell

We find that vegetation insensitive to water-stress depletes soil moisture faster than the sensitive vegetation because it does not downregulate g_s and LE until the soil moisture is very close to wilting point. Even on the first day when SMI = 0.55 the sensitive response downregulates LE to 300 W m^{-2} whereas the insensitive response keeps evapotranspiring at potential level (LE = 400 W m^{-2}) (Fig. 5). The accumulated amount of water transpired from the deeper soil layer – where the maize roots are located – over the 21-day period is much greater for the insensitive response (78 mm) than for the sensitive response (57 mm). As a result, the SMI for the insensitive response drops below wilting point and is much lower than for the sensitive response. The insensitive response is similar to observed C₃ and C₄ grasses over Europe, while the sensitive response is more conservative of water resources as observed over forests (Wolf et al., 2013; Teuling et al., 2010).

Our simulations illustrate how vegetation insensitive to water-stress delays atmospheric warming. Fig. 5c and d show this vegetation conserves a high $g_s > 14 \text{ mm s}^{-1}$, thus a low maximum SH $\approx 100 \text{ W m}^{-2}$ and a low θ_{max} of 298 K over the first 14 days. Between days 14 and 19 in Fig. 5c it switches suddenly to a low $g_s < 1 \text{ mm s}^{-1}$, and subsequently a high maximum SH flux of 400 W m^{-2} and a high θ_{max} of 303 K. In contrast, vegetation sensitive to water-stress evolves gradually toward a warmer atmospheric state over the 21 days, but without ever reaching the same range of temperature increase (from 299 to 302 K). Our finding is thus in agreement with Teuling et al. (2010), van Heerwaarden and Teuling (2014), who showed that during the European 2003 heat-wave forests downregulated their transpiration sooner than grass and were showing increases in air temperature earlier in time. Our result shows that the choice of water stress representation in a land-surface model is important to simulate the build-up of heat waves, and is an additional factor to take into account in heat waves studies like Miralles et al. (2014).

Finally, when we look at the mixed-layer CO₂ mole fraction, we find the insensitive water-stress response consistently generates

lower CO₂ mole fractions than the more sensitive response over the first 15 days. After day 17, this difference reverses as the insensitive response abruptly decreases g_s , reducing the coupling between land and atmosphere for the carbon cycle. In Fig. 5b we find the diurnal minimum CO₂ mole fraction for the sensitive response evolves linearly from 354 to 362 ppm, whereas the diurnal minimum CO₂ mole fraction for the insensitive response switches abruptly from 341 to 368 ppm between day 12 and 19. As a consequence, it reaches the weakly coupled CO₂ regime that we tentatively defined in Section 3.1 at $g_s = 3 \text{ mm s}^{-1}$. Any model interpretation of observed CO₂ mole fractions under water-stressed conditions thus needs to consider the type of water stress response to explain (a) the temporal evolution pattern of CO₂ mole fractions, and (b) the relative importance of the entrainment process and of large-scale conditions in the ABL CO₂ budget.

3.3. Sensitivity to model errors

The long-term temperature and CO₂ increases from Section 3.2 can be regarded as a lower-bound estimate during such drought, as we have currently only carried over the effects on soil-moisture from one day to the next. Additional cumulative effects can be expected for example from increasing early-morning temperatures (which increases g_s , VPD and R_{het} , see Fig. 1), reduced absolute humidity (which increases VPD), reduced cloud cover (which increases PAR). However, these cumulative changes cannot be modeled readily with our framework, as we cannot simulate night-time conditions, meso-scale air circulation or cloud formation. But we can assess the effects of possible errors in these variables (i.e., a deviation from their true state) that can occur in our simulations or in other models that do not correctly capture the dynamics of the land (e.g. LAI), or atmosphere (e.g. meso-scale air circulation), or their interaction.

Errors in simulated atmospheric conditions can have significant impacts on the ABL CO₂ mole fraction and ABL temperature on a single day. We show this in Fig. 6 where we present the change in the daily minimum c (y-axis) and in the daily maximum θ (x-axis) resulting from prescribed errors in the atmospheric initial conditions of the DI and DS base runs. All prescribed errors listed in Table 1 cause a positive or negative change in c_{min} larger than 1 ppm, and an always positive change in θ_{max} larger than 0.5 K independent of the assumed plant water stress response. For

Fig. 6. Contrasted response of the ABL CO₂ and temperature to atmospheric model errors for two contrasted plant sensitivity to water stress, under severe water deficit (SMI = 0.2). In Fig. 3, the sensitive (DS) and insensitive (DI) base cases reached two different c_{min} (364 ppm for DS and 343 ppm for DI) and θ_{max} (301 K for the DS and 297 K for DI) due to their contrasted responses to water stress. We present here the deviations from these base cases c_{min} and θ_{max} , after applying changes to the runs initial conditions (see Table 1).

example, overestimates in simulated early morning temperature of the mixed-layer (+4 K θ_0) translate to almost identical overestimates in the daytime maximum temperature (+3 to +3.5 K θ_{max}). Similar positive impacts on the maximum temperature are seen from overestimates in large-scale subsidence rates (+1 to +2 K) or temperature lapse-rates (+0.5 to +1 K), as qualitatively illustrated in our conceptual Fig. 1. Our examples thus clearly illustrate that simulated conditions in the atmospheric boundary layer during dry spells are very sensitive to both its boundary and initial conditions. Their errors, as well as the simulated coupling strengths thus determine the accuracy of a simulated heatwave.

Moreover, depending on the assumed vegetation sensitivity to water-stress the effect of such model errors on the CO₂ mole fraction and ABL temperature can double, or half, or even reverse sign. For instance in Fig. 6, a switch from a sensitive to an insensitive plant response to water stress causes the impact of an error in simulated subsidence rates to be twice as large on c_{min} , but half as large in θ_{max} . Also overestimates in simulated γ_θ causes generates a 2 ppm higher c_{min} when assuming the sensitive response or a 4 ppm lower c_{min} when assuming the insensitive response. The key difference when switching from the sensitive to the insensitive water-stress response cases is the increase in surface coupling and the subsequent 4 K lower temperature regime of the ABL (see Fig. 3). The assumed plant water-stress sensitivity in the model puts us on opposite sides of the optimum temperature for photosynthesis (298 K): while the leaf temperature of the DI base case is consistently below this optimum throughout the day, the one of the DS base case is consistently above this optimum. The extra warming then generated by each model error (from +0.5 to +3.5 K for c , γ_θ , w_s , and θ_0) causes the insensitive vegetation to increase its assimilation rate and the sensitive vegetation to decrease its assimilation rate, generating consistently higher c_{min} for the sensitive cases compared to their corresponding insensitive cases. Our results highlight the importance of the representation of plant-water stress not just for the temporal evolution of the mean atmospheric conditions as in Section 3.2, but also for directing the impact of errors in a coupled model for carbon, water, and energy. We will further discuss this in Section 4.

Finally, we see that simultaneous errors in variables associated with the development of a dry spell add non-linearly to affect the CO₂ minimum and ABL temperature. In Fig. 6 we find that a combination of reduced cloud cover with an increased subsidence rate, free-troposphere temperature lapse rate, and early-morning temperature of the boundary layer, can amount to an error in θ_{max} of 4.6 to 5.2 K and in c_{min} of -27 to +2 ppm. Interestingly, the small effect on CO₂ in the sensitive vegetation case is not equal to the sum of the individual model error effects, and this is also true for the total temperature error. This especially illustrates the interdependence of the effects triggered by the combined model errors (see Fig. 1). Their combination reflects a model where all drivers during a drought are overestimated but in the sensitive vegetation response CO₂ is hardly affected, while in the insensitive case the modeled daytime CO₂ minimum is too low by almost 30 ppm. The implications of this finding will be discussed in the next section.

4. Discussion

The modularity of the atmospheric model (MXL) has allowed its users to add many processes to explore their importance for the diurnal simulation of the ABL temperature, moisture and CO₂. This includes a number of processes not considered in this study, such as the effect of shear-driven turbulence (Pino et al., 2006), residual boundary layers (Ouwensloot et al., 2012), shallow-cumulus formation and surface shading (van Stratum et al., 2014; Vilà-Guerau de Arellano et al., 2012), large-scale advection of heat (Miralles et al.,

2014) and heating of the ABL by aerosols (Barbaro et al., 2014). Guided by the observations, we chose here to focus on the most important processes for a convective boundary-layer when coupled to a vegetation model that includes carbon exchange. On the atmospheric side of our framework this includes entrainment, subsidence, boundary-layer growth, free-tropospheric conditions, and horizontal heat advection. In our land-surface model (A-g_s) this includes the coupling of the carbon and water cycles through stomatal conductance. Its formulation has been validated for a number of surfaces with low vegetation such as maize (Combe et al., 2015), C₃ and C₄ grass, soybean (Ronda et al., 2001), and grapevine (Jacobs et al., 1996). We specifically note that to extend it to high vegetation (i.e. forests), the disturbance of the turbulence by the canopy should be included in the model. Such extension for the MXL model is being done at the moment, following the principles of Harman and Finnigan (2007). To additionally include the effect of meso-scale circulations, nighttime turbulent exchange, topography, or cloud dynamics requires a more explicit 3-D modeling framework, which would add considerable complexity to the study. We simply represent those as large-scale forcings by imposing vertical and horizontal advection of heat and moisture (Pietersen et al., 2015).

The range of water-stress responses and soil moisture index explored here is representative of a specific range of observed situations. A soil moisture index of 0.2 is a typical water deficit observed during summer in Southern Europe (e.g. southern France and Portugal as shown in Barbu et al., 2011; Jongen et al., 2011) that can also occur in wetter regions during prolonged droughts. Our findings are thus relevant for those cases, as the replenishment of soil water by precipitation or irrigation will largely diminish our contrasted plant water stress responses effect on the stomatal conductance. Moreover, note that here we are exploring the spectrum of water stress responses existing between the linear and the very non-linear insensitive responses. We chose that spectrum as it was representative of a number of land-surface water-stress responses (SiB3, CTESSEL, JULES, CLM, A-g_s, see Boussetta et al., 2013; Powell et al., 2013; Ronda et al., 2001). However, there is a second half of the spectrum, between the linear and the non-linear very sensitive response, the latter being representative of plants that shut their stomata as soon as there is a small soil moisture deficit (Calvet et al., 2004; Tardieu and Simonneau, 1998). By extension of our results, we expect the decoupling of the L–A water and carbon exchange (and the corresponding high L–A heat exchange) to happen at higher SMI and thus earlier in time during dry spells for this very sensitive non-linear response to water stress.

The plant water stress parameterization we investigated here is not the only approach taken in land-surface models. Several studies (Egea et al., 2011; Keenan et al., 2010; Calvet et al., 1998) have suggested that in addition to downregulation of stomatal conductance, the mesophyll conductance and/or the maximum carboxylation rate should also be downregulated. The latter would account for non-stomatal effects, such as the decline in leaf nitrogen content (Grassi and Magnani, 2005) and the hardening of mesophyll cell walls (Keenan et al., 2010). Even in models that use the β -function as employed here, it is not clear which shape to choose because as argued by Joetzjer et al. (2014) the physiological response to water stress itself is not well understood. It will vary between species, and is observed to even be different for different varieties of crops (Zhou et al., 2013; Carmo-Silva et al., 2007; Calvet et al., 2004; Tardieu and Simonneau, 1998). As an alternative, mechanistic approaches were proposed (Verhoef and Egea, 2014; McDowell et al., 2013) that describe plant internal hydraulics. Such models in turn require more complex model input, with variables difficult to acquire (e.g. root density, soil hydraulic conductivity). For both water-stress response approaches, Powell et al. (2013) showed that a full drought response of vegetation should moreover include a change in respiration, LAI, plant mortality, carbon

allocation, root adaptation, and even reproduction strategies (also see van der Molen et al., 2011). This full drought response has to be represented in order for models to be able to simulate the longer-term vegetation–atmosphere interactions, with cumulative effects of water-stress on the state of vegetation. However, note that these cumulative effects do not play a role on the short time scales of our analysis. Hence a first extension of our study could be to repeat our experiments with our coupled MXL-A-g_s model but introducing the mechanistic water-stress approach, to observe which type of sensitivity to water stress and which L–A coupling it can model.

The idea of a coupling strength in climate models has been raised before. For example, Koster (2004) and Dirmeyer et al. (2006) looked at the hydrological cycle to define a coupling strength between the atmosphere and land, specifically using soil moisture and precipitation. The fraction of explained variance in soil moisture by variations in precipitation allowed a quantification of this strength, and the separation of different coupling regimes. Seneviratne et al. (2010) summarizes similar approaches, amongst others for the soil moisture–temperature coupling strength. More similar to our use of a coupling strength, Santanello et al. (2009) analyzed the local land–atmosphere coupling state of heat and water showing that there are three surface energy balance regimes depending on the soil wetness (dry, intermediate, wet). Our study expands this analysis also to the surface carbon fluxes, and through R_c moreover accounts for CO₂ entrainment at the top of the PBL. This coupling strength at the top of the boundary layer is often not considered in coupled land-surface modeling, but forms an important connection to the larger atmospheric scales that control for example the development of heat waves.

Our study of the development of a heat wave is highly simplified due to the impossibility of the model to run through nighttime that led us not to change atmospheric conditions from day-to-day. This simplification can be partly supported by the fact that a heatwave typically develops in a synoptic situation that is relatively stable, with high-pressure, few clouds, strong subsidence, and little precipitation. However, variations in other variables are likely to occur over the 21-day period: heat and moisture advection, temperature lapse rates, and CO₂ in and above the ABL typically also change over synoptic scales (5–10) days (Miralles et al., 2014). These variations are partly offset by the rapid adaptation of the short time scale processes of ABL growth to the long term variations (e.g., subsidence, soil moisture). In our results we therefore specifically isolated the effect of progressive soil moisture depletion on the land–atmosphere coupling and obtained a reasonable soil moisture depletion over the 21 days of 30–55% in SMI. This is on the high side compared to the SMI decrease observed by Wolf et al. (2013) during a 30-day drought in Switzerland (–20% SMI) but can be partially explained by the observed replenishment of soil moisture by precipitation, which is absent in our model. Our findings agree with Joetzjer et al. (2014) that the choice of plant water-stress response can strongly affect the dynamic simulation of soil moisture in a land-surface model.

Although our study is conceptual in nature, its results can be generalized to a large number of other models that use the same principle of coupled photosynthesis–conductance modeling, and moisture-stress induced reductions of leaf assimilation. This includes for example the version of the Integrated Forecast System built by ECMWF used in the Copernicus project (Boussetta et al., 2013). In its coupled land-surface model (CTESSEL) the stomatal conductance, and hence energy and water balance, will respond very similarly to what we have shown here to the atmospheric and surface drivers, and to errors therein. The skill of this numerical weather prediction system therefore is tied partially to the vegetation state and its response to water-stress. A similar argument can be made for Earth System Models used in coupled carbon–climate simulations (Meir et al., 2006; Friedlingstein et al., 2006; Booth

et al., 2012), which instead simulate the long-term fate of vegetation and soil carbon pools based on this L–A coupling. With the tropics as dominant region for carbon–climate feedbacks, Joetzjer et al. (2014) illustrated the effect of different drought response function on the simulation of carbon fluxes in the Amazon in one such model (ISBA). As hypothesized by Berry et al. (2010) and in agreement with our study, large effects on stomatal conductance lead to large variations in evaporation and carbon assimilation. These authors stressed the need to include drought effects on plant mortality, in addition to a better understanding of the mechanisms of drought response as is also advocated in van der Molen et al. (2011).

5. Conclusion

We systematically evaluated the sensitivity of a coupled land–atmosphere system to the plant water–stress response under drought conditions. Using various responses from linear (sensitive) to non-linear (insensitive) within the MXL–A–g_s model, we quantified the differences in the simulated budgets of carbon, water, and heat in the daytime ABL. The strength of our framework is that it includes the essential processes of the diurnal convective ABL coupling with the surface characteristics of low vegetation, similar to larger, more complex, and more widely used land–surface models. We use our conceptual framework to answer three research questions.

Firstly, we find that under low soil moisture conditions the use of a sensitive plant water–stress response in a land–surface model can strongly decrease the diurnal exchange of carbon and water at the surface and increase the exchange of heat. This increases ABL growth and entrainment, contributes to further warming and drying of the ABL, and shifts the dominant processes in the ABL CO₂ budget away from surface CO₂ exchange. Misspecification of this simulated water–stress response can lead to errors in simulated temperatures, a misinterpretation of observed CO₂ mole fractions, but also leads to errors in coupled land–atmosphere simulations with climate models for longer integration times (drought, heat–waves).

Secondly, during the development a dry spell, plants that are insensitive to water stress will delay atmospheric warming at the expense of a faster depletion of soil moisture. This will create a sudden switch in surface coupling strength, CO₂ mole fractions, and temperature when approaching wilting point. On the contrary, plants that are more sensitive to water–stress will immediately contribute to increased atmospheric warming by reducing evaporation to preserve soil water resources. These contrasting vegetation responses are important to model the onset of heat waves and the development of droughts in numerical weather prediction models, which are starting to rely on coupled photosynthesis–evaporation formulations for stomatal conductance.

Thirdly, we find that the nonlinear interactions between water–stress sensitivity, stomatal conductance, and atmospheric dynamics create a wide range of system responses in which errors in initial and boundary conditions for the ABL can affect the ABL budgets of carbon, water, and heat very differently, and even non-linearly. This means that a coupled model's sensitivity to errors in simulated drought conditions can shift strongly due to underlying assumptions. This is especially relevant for models used in numerical weather prediction, and models used to simulate droughts in a future climate.

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Appendix A. Equations for the heat and water cycles

This appendix strictly provides the set of equations necessary to obtain the ABL virtual potential temperature and its jump at the inversion height (see Section 2.2). More information can be found about the mixed-layer theory in Vilà-Guerau de Arellano et al. (2015).

First, the diurnal evolution of the ABL potential temperature (θ) is described with the following equation:

$$\begin{aligned} \frac{d\theta}{dt} &= \frac{1}{h} \times \left(\underbrace{\overline{w'\theta'_s}}_{\text{Surface turbulent flux of temperature}} - \underbrace{\overline{w'\theta'_e}}_{\text{Entrainment, turbulent flux of temperature}} \right) + \underbrace{adv_\theta}_{\text{Horizontal advection of heat}}, \\ &= \frac{1}{h} \times \left(\frac{SH}{\rho_{air}c_p} - w_e\Delta\theta \right) + adv_\theta, \end{aligned} \quad (A.1)$$

with SH the surface sensible heat flux in W m^{−2}, c_p the specific heat of moist air, and $\Delta\theta$ the potential temperature jump at the inversion height. $\Delta\theta$ evolves in time due to the entrainment of free-tropospheric warmer air into the ABL and is expressed as:

$$\frac{d\Delta\theta}{dt} = \gamma_\theta w_e - \frac{d\theta}{dt}, \quad (A.2)$$

with γ_θ the free-troposphere lapse rate of potential temperature. SH is expressed as the temperature gradient between the surface and the ABL times the corresponding surface conductance for heat transfer:

$$SH = \rho_{air}c_p \frac{1}{r_a} (\theta_s - \theta). \quad (A.3)$$

Very similarly to θ , the diurnal evolution of the ABL specific humidity (q) is described with the following equation:

$$\begin{aligned} \frac{dq}{dt} &= \frac{1}{h} \times \left(\underbrace{\overline{w'q'_s}}_{\text{Surface turbulent flux of humidity}} - \underbrace{\overline{w'q'_e}}_{\text{Entrainment turbulent flux of humidity}} \right) + \underbrace{adv_q}_{\text{Horizontal advection of moisture}}, \\ &= \frac{1}{h} \times \left(\frac{LE}{\rho_{air}L_v} + w_e\Delta q \right) + adv_q, \end{aligned} \quad (A.4)$$

with LE the total surface latent heat flux in W m^{−2} and Δq the specific humidity jump at the inversion height. Δq evolves in time due to the entrainment of free-tropospheric dryer air into the ABL and is expressed as:

$$\frac{d\Delta q}{dt} = \gamma_q w_e - \frac{dq}{dt}, \quad (A.5)$$

with γ_q the free-troposphere lapse rate of specific humidity. LE is expressed as the moisture gradient times the corresponding surface conductance for water transfer:

$$LE = \rho_{air}L_v \left(\frac{f_{veg}}{r_a + r_s} + \frac{1 - f_{veg}}{r_a + r_{soil}} \right) (q_{sat}(\theta_s) - q), \quad (A.6)$$

with f_{veg} the vegetated fraction of the ground and r_{soil} the bare soil resistance to water transfer. Note that we assume here no evaporation of liquid water (e.g. dew).

Finally, the mixed-layer virtual potential temperature (θ_v) used in the calculation of the entrainment velocity (w_e , see Eq. (2)) is the mixed-layer potential temperature corrected for the variation of density due to the presence of water vapor in the air. For unsaturated air, θ_v and its jump at the inversion height ($\Delta\theta_v$) are calculated as follows:

$$\theta_v = \theta \times (1 + 0.61q), \quad (A.7)$$

$$\Delta\theta_v = \Delta\theta + 0.61(q\Delta\theta + \theta\Delta q + \Delta\theta\Delta q). \quad (A.8)$$

Appendix B. Derivation of r_c

As explained in Section 2.2.2, the leaf stomatal resistance to CO₂ transfer r_{lc} can be expressed as:

$$\frac{1}{r_{lc}} = g_{min,c} + \frac{a_1 A_g^* \beta}{(c - \Gamma)(1 + VPD/VPD_*)}.$$

Then, to upscale r_{lc} from the leaf to the canopy level (i.e. to obtain r_c), we must integrate this equation over LAI:

$$\begin{aligned} \frac{1}{r_c} &= \int_0^{LAI} \left[\frac{g_{min,w}}{1.6} + \frac{a_1 A_g^* \beta}{(c - \Gamma)(1 + VPD/VPD_*)} \right] dL, \\ &= \frac{g_{min,w}}{1.6} LAI + \frac{a_1 \beta \int_0^{LAI} A_g^* dL}{(c - \Gamma)(1 + VPD/VPD_*)}, \end{aligned} \quad (B.1)$$

with $g_{min,w}$ the cuticular conductance for water transfer, which is related to $g_{min,c}$ by the molecular diffusion constant 1.6. Obtaining r_c then only requires to integrate A_g^* (see Eq. (8)) over LAI:

$$\begin{aligned} \int_0^{LAI} A_g^* dL &= \int_0^{LAI} \left[(A_m + R_{dark}) \left\{ 1 - \exp \left(- \frac{LUE K_x PAR_t e^{-K_x L}}{(A_m + R_{dark})} \right) \right\} \right] dL, \\ &= (A_m + R_{dark}) \left\{ LAI - \int_0^{LAI} \exp(-be^{-K_x L}) dL \right\}, \end{aligned} \quad (B.2)$$

with $b = LUE K_x PAR_t / (A_m + R_{dark})$. We introduce the transformation $y = be^{-K_x L}$. Under this transformation, the boundaries of the integral over the leaf area become $L = 0 \Rightarrow y = b$ and $L = LAI \Rightarrow y = be^{-K_x LAI}$. Thus, Eq. (B.2) can be rewritten as:

$$\begin{aligned} \int_0^{LAI} A_g^* dL &= (A_m + R_{dark}) \left\{ LAI + \frac{1}{K_x} \int_b^{be^{-K_x LAI}} \frac{e^{-y}}{y} dy \right\}, \\ &= (A_m + R_{dark}) \left\{ LAI + \frac{1}{K_x} [E_1(b) - E_1(be^{-K_x LAI})] \right\}, \end{aligned} \quad (B.3)$$

with $E_1(x)$ the integral of $\frac{e^{-y}}{y}$ from x to infinity. The final form of Eq. (B.2) thus becomes:

$$\begin{aligned} \int_0^{LAI} A_g^* dL &= (A_m + R_{dark}) \left\{ LAI + \frac{1}{K_x} \left[E_1 \left(\frac{LUE K_x PAR_t}{(A_m + R_{dark})} \right) \right. \right. \\ &\quad \left. \left. - E_1 \left(\frac{LUE K_x PAR_t}{(A_m + R_{dark})} e^{-K_x LAI} \right) \right] \right\}, \end{aligned} \quad (B.4)$$

which can be used in Eq. (B.1) to calculate r_c .

Appendix C. Model settings of the base case

Table C.1

Table C.1

Model constants, boundary and initial conditions of the control case.

Variable	Value	Units	Description
<i>Constants</i>			
C ₃ /C ₄	C ₄	–	C ₃ or C ₄ photosynthesis
<i>Boundary conditions</i>			
Lat	51.59	°N	Latitude
Lon	5.38	°E	Longitude
Day	216	DOY	Date
p	1022.0	hPa	Atmospheric pressure
w _s	7 × 10 ^{−6}	s ^{−1}	Large-scale divergence
cc	0.225	–	Cloud cover
γ _θ	8 × 10 ^{−3}	K m ^{−1}	Potential temperature lapse rate
γ _q	−5 × 10 ^{−4}	g kg ^{−1} m ^{−1}	Specific humidity lapse rate
γ _c	−10 × 10 ^{−3}	ppm m ^{−1}	CO ₂ mole fraction lapse rate
adv _θ	3.0 × 10 ^{−4}	K s ^{−1}	Horizontal advection flux of heat
adv _q	3.5 × 10 ^{−4}	g kg ^{−1} s ^{−1}	Horizontal advection flux of moisture
α	0.198	–	Surface albedo
LAI	3.5	m ² m ^{−2}	Leaf area index
f _{veg}	0.97	–	Vegetation fraction
<i>Initial conditions</i>			
h ₀	230.0	m	Initial boundary-layer height
θ ₀	286.0	K	Initial potential temperature
Δθ ₀	5.0	K	Initial potential temperature jump
q ₀	8.5	g kg ^{−1}	Initial specific humidity
Δq ₀	−1.0	g kg ^{−1}	Initial specific humidity jump
c ₀	422.0	ppm	Initial CO ₂ mole fraction
Δc ₀	−50.0	ppm	Initial CO ₂ mole fraction jump
SMI ₀	0.55	–	Initial soil moisture index

NB: note that the horizontal advection fluxes are set to zero at 07:30 UTC for moisture and 10:00 UTC for heat.

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