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IMPACT OF SOIL MOISTURE ON BOUNDARY-LAYER CLOUD DEVELOPMENT

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1. INTRODUCTION

We study the daytime land-atmosphere interaction using model with an atmospheric boundary-layer (ABL) scheme coupled with a land-surface (LS) scheme using observations taken on 31 May 1978 at Cabauw, Netherlands. In a previous study (Holtslag et al 1995) it was found that in coupled (LS-ABL) model simulations using a simple LS scheme did not accurately represent surface fluxes and coupled atmospheric boundary-layer development. Using a more sophisticated LS scheme in the study here allows the land-atmosphere system the freedom to respond interactively with the ABL where a many processes and important feedback mechanisms are represented (Figure 1). Results indicate that in coupled land-atmosphere simulations, realistic daytime surface fluxes and atmospheric profiles are produced, even in the presence of ABL clouds.

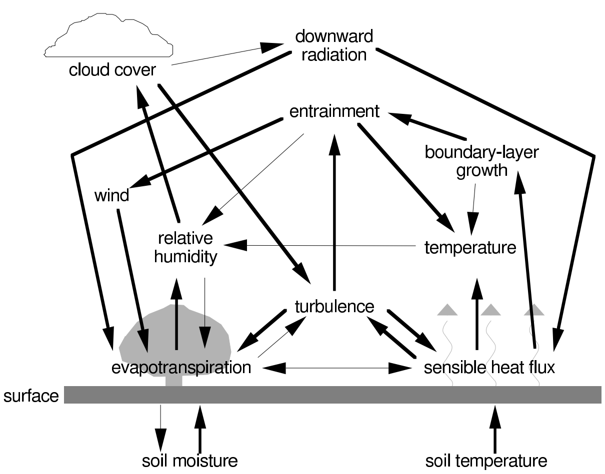

Figure 1. Important interactions between the land-surface and atmospheric boundary layer for conditions of daytime surface heating. Thick arrows indicate the direction of feedbacks, which are normally positive (leading to an increase of the recipient variable), while thin arrows indicate negative feedbacks. Two consecutive negative feedbacks make a positive one. Note the many positive and negative feedback loops, which may lead to increased or decreased relative humidity and cloud cover.

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Subsequently, the role of soil moisture in the development of ABL clouds is explored via model runs, analytical development, and with observational (Cabauw) data in terms of a relative humidity (RH) tendency equation at the ABL top which involves a number of land-atmosphere interactions. It is shown that the effect of soil moisture is to increase ABL-top RH tendency and thus potential for ABL cloud formation (confirming intuition), but only if the stability above the ABL is not too weak (and given sufficient initial RH in the ABL and air above the ABL that is not too dry). Alternately, for weak stability above the ABL, drier soils yield a greater ABL-top RH tendency and thus potential for ABL cloud formation (somewhat counter-intuitive), where in this case soil moisture acts to limit the increase of ABL-top RH, and that the largest values of ABL-top RH tendency are achieved not over moist soils, but rather over dry soils.

2. LAND-SURFACE – ATMOSPHERIC BOUNDARY LAYER MODEL EVALUATION AT CABAUW

We begin by representing the soil-vegetation system in offline model runs for Cabauw (Netherlands) for a case study day (31 May 1978) using a LS-only model driven by observed atmospheric forcing using existing formulations *without tuning* model parameters (Ek and Holtslag 2002a). We follow this with ABL-only model runs (driven by observed surface fluxes) and then coupled LS-ABL model runs (Ek and Holtslag 2002b). Results indicate that in all systems: LS-only, ABL-only and coupled LS-ABL model runs, realistic daytime surface fluxes and atmospheric profiles including ABL clouds are produced and compare well with observations using our LS-ABL model with updated or alternative, *but un-tuned*, parameterizations. Both land-surface and ABL model runs yielded encouraging results operating separately, and interactively when coupled together, even in the presence of model-predicted ABL clouds. This suggests that in this coupled land-atmosphere system, processes are well-represented by our LS-ABL model. Model parameterization updates include a change to the boundary-layer depth formulation, and modifications in land-surface formulations related to the parameterization of canopy conductance at Cabauw (Beljaars and Bosveld 1997), soil heat flux formulation, and plant root density distribution alternatives.

3. SOIL MOISTURE IMPACT ON ABL CLOUDS

3.1 *Model sensitivity tests*

Establishing that our coupled LS-ABL model represents the interactive land-atmosphere system at Cabauw rather well, we now more fully explore the interaction of the land-surface with the ABL and the effect on boundary-layer cloud development. In order to examine the role of soil moisture, we make a series of model runs (a 'reference' set) where we change the soil moisture from quite dry to quite wet. Initial conditions and forcing are the same as in our previous coupled model runs, except now we vary soil moisture from below the wilting point (quite dry) to near saturation (quite wet). We note that for the various model runs, as we decrease the initial soil moisture from intermediate soil moisture values (close to observations, volumetric soil moisture ≈ 0.43) to below the wilting point, ABL cloud cover decreases to zero (Figure 2), a somewhat intuitive result.

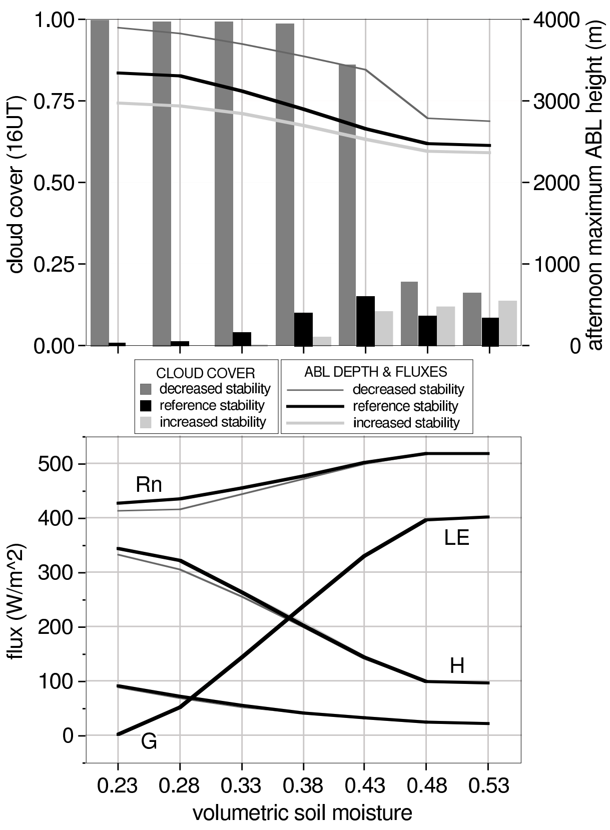

Figure 2. Impact of variation in volumetric soil moisture for different sets of model runs (reference, and increased and decreased stability above the ABL) on ABL depth and cloud cover (top), and components of the surface energy budget (bottom).

However, as we increase the initial soil moisture from intermediate soil moisture values to near saturation, ABL cloud cover decreases slightly, a somewhat counter-intuitive result. Certainly there are a number of processes that account for this behavior, i.e. interactions between the land-surface, atmospheric boundary layer (including ABL clouds), free atmosphere, and initial ABL conditions.

Before attempting an explanation of this response, we also examine the role of atmospheric stability ('capping inversion') above the ABL in land-surface interaction with the evolving boundary layer since above-ABL stability has a strong influence on boundary-layer growth. We make two additional sets of model runs as above, except now we prescribe one set with *increased* atmospheric stability above the observed afternoon boundary-layer top (compared with the reference set of model runs above), and another set with *decreased* atmospheric stability. We then examine the resulting afternoon ABL depth and fractional cloud cover and the mid-day surface energy budget as it changes with changing prescribed initial soil moisture (Figure 2).

The set of model runs with stronger atmospheric stability have a shallower ABL depth than the reference set and less cloud cover for drier soils, with increasing cloud cover for model runs with increased soil moisture (Figure 2). However, in great contrast, the set of model runs with weaker atmospheric stability above the ABL have a deeper ABL depth (as one would expect) and yet a much greater cloud cover for drier soils, with decreasing cloud cover for increasing soil moisture. In the next section, we will attempt to explain this result in terms of a tendency equation for relative humidity at the ABL top.

3.2 *ABL-top RH tendency*

The role of soil moisture in ABL cloud development involves a complex interaction of surface and atmospheric processes (see Figure 1). Ek and Mahrt (1994) examined the daytime evolution of ABL-top relative humidity which is expected to control ABL cloud development. They showed that the relative humidity tendency at the ABL top involves a number of competing mechanisms, with relative humidity directly *increasing* due to surface evaporation and due to ABL growth (ABL-top temperature decreases), and relative humidity directly *decreasing* due to surface sensible heat flux and due to entrainment of warm and dry air into the ABL from above. The *indirect* role of surface evaporation is to reduce surface heating, thereby competing with ABL growth, though diminishing ABL-top warm- and dry-air entrainment.

To further understand the role of soil moisture and other factors on ABL cloud development, we extend the work of Ek and Mahrt (1994) (with an update by Chang and Ek 1996) and examine a useful new equation for relative humidity (RH) tendency at the ABL top (see Ek and Holtslag 2002b for the full development):

$$\partial RH/\partial t = (R_n - G) / (\rho L_v h q_s) [e_f + x(1-e_f)]. \qquad (1)$$

Here R_n-G is available energy at the surface (R_n is net radiation and G is soil heat flux), ρ is air density, L_v is latent heat, h is ABL depth, and q_s is saturation specific humidity just below the ABL top. The surface evaporative fraction (of surface energy available for evaporation), e_f, is defined as:

$$e_f = LE / (R_n\text{-}G), \qquad (2)$$

where LE is the surface moisture flux. Furthermore, x reflects the direct effects of non-evaporative processes on relative humidity tendency, where x is given by:

$$x = L_v/c_p\ (1+C_\theta)\ [\ \Delta q/(h\ \gamma_\theta) + RH(c_2/\gamma_\theta) - c_1)\ ]\ , \qquad (3)$$

where c_p is specific heat, C_θ is the ratio of surface to ABL-top sensible heat flux, Δq is the specific humidity drop above the ABL (negative), γ_θ is the potential temperature lapse rate above the ABL, and c_1,c_2 are functions of surface pressure, temperature and pressure at the ABL top, and constants (see Ek and Holtslag 2002b). x consists of three terms (each multiplied by $L_v/c_p\ (1+C_\theta)$): ABL-top dry-air entrainment ($\Delta q\ /(h\ \gamma_\theta)$, a negative contribution), boundary-layer growth ($RH\ c_2/\gamma_\theta$), a positive contribution), and boundary layer heating through surface warming and ABL-top warm-air entrainment ($RH\ c_1$, a negative contribution).

From Eq. 1 we see that the relative humidity tendency is proportional to available energy and inversely proportional to ABL depth and temperature (via saturation specific humidity), while the sign of the relative humidity tendency is determined by the sign of $e_f + x(1\text{-}e_f)$. Examining Eq. 1, it is apparent that the direct role of e_f is to increase the ABL-top relative humidity, while the indirect role of surface evaporation (via reduced surface heating, and diminished ABL growth and entrainment) is found in the expression $x(1\text{-}e_f)$. Figure 3 shows how $e_f + x(1\text{-}e_f)$ depends on e_f versus x, where $e_f + x(1\text{-}e_f)$ is simply the relative humidity tendency, $\partial RH/\partial t$, normalized by the available energy term, $(R_n\text{-}G) / (\rho\ L_v\ h\ q_s)$.

For the case where $x<1$, $\partial RH/\partial t$ increases as the evaporative fraction (e_f) *increases*, confirming intuition. (For the range $0<x<1$, $\partial RH/\partial t>0$ and increases with increasing e_f, while for $x<0$, $\partial RH/\partial t>0$ only when e_f exceeds some threshold value which increases for increasingly negative values of x). Here soil moisture acts to increase ABL-top relative humidity tendency and thus increases the probability of ABL cloud initiation given a sufficient initial ABL relative humidity. This is the case when the soil is sufficiently moist for unrestricted surface evaporation, and the environment above the ABL is not too dry (Δq small) and atmospheric stability (γ_θ) is not too weak.

For the case where $x>1$, $\partial RH/\partial t$ increases as e_f *decreases* (which is somewhat counter-intuitive) so that here soil moisture acts to limit the increase of ABL-top relative humidity and thus decreases the probability of ABL cloud initiation. This is the case when the environment above the ABL is again not too dry (Δq small) but atmospheric stability (γ_θ) is rather weak, so for drier soils, surface evaporation is lower with boundary-layer growth less restricted than with moister soils. Note that the largest values of $\partial RH/\partial t$ are achieved for $x>1$ suggesting that the greatest potential for ABL cloud initiation is not over moist soils, but rather over dry soils with weak stability (and air not too dry above the ABL).

Before we proceed, we note that the outcome of Eqs. 1–3 (as presented in Figure 3) agrees well with the output of the coupled model (confirmed by more than a thousand runs), as long as $h/\text{-}L>5$ (ABL depth/Obukhov length) which is required for the assumption of mixed-layer conditions (see Holtslag and Nieuwstadt 1986).

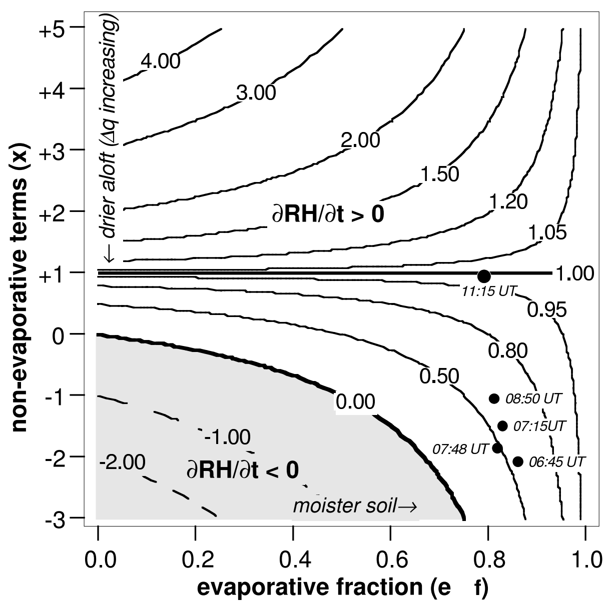

Figure 3. Relative humidity tendency equation (normalized by the available energy term), $e_f + x(1\text{-}e_f)$, as a function of evaporative fraction (e_f) versus non-evaporative processes (x) with Cabauw observed values and times indicated (dots).

3.3 *Discussion*

We can examine the various ABL-top relative humidity tendency terms in Eqs. 1-3 for Cabauw data during periods of positive surface fluxes and when $h/\text{-}L>5$ (Figure 3). From mid-morning until mid-day, the dry-air entrainment term decreases greatly with time because of increasing ABL depth and a somewhat steady value of dry air above the ABL (despite decreasing atmospheric stability just above the growing ABL), while the ABL growth term increases greatly as the atmospheric stability decreases. During this same time period the ABL warming term diminishes only modestly, and the evaporative fraction increases only slightly. The effect of soil moisture is then to increase the ABL-top relative humidity (*x<1*), except during the mid-day rapid ABL growth when the effect of soil moisture only modestly increases ABL-top relative humidity (x<≈1). We note that the ABL-top relative humidity increased sufficiently for ABL clouds (both modeled and observed) to form by mid-to-late afternoon.

We now focus on the rapid ABL growth period (e.g. 11:15 UT at Cabauw), during or after which ABL clouds are generally initiated, and examine the effect of changing evaporative fraction and atmospheric stability on the relative humidity tendency. Using the initial soil moisture values near those observed at Cabauw, note that as with the Cabauw observations, *x<≈1* for the reference set model run as well. For a drier soil in this case, normalized relative humidity tendency decreases slightly with ABL cloud cover also decreasing. In a deeper growing boundary layer due to larger surface sensible heat flux, a larger *h* yields a smaller actual relative humidity tendency (see Eqs. 1-3), and less cloud cover (Figure 2). Here stronger warm- and dry-air entrainment negates the effect of ABL-top cooling on the increase of ABL-top relative humidity. For a moister soil, normalized relative humidity tendency increases slightly, though with the a shallower ABL depth the actual relative humidity tendency is less with subsequently less cloud cover. In this case the greatest relative humidity tendency (and thus cloud cover) occurs for intermediate soil moisture. This is in agreement with our assessment of the role of soil moisture on ABL cloud development based on the development in the previous section.

For the set of model runs with increased (stronger) atmospheric stability, ABL depth is shallower (as one would expect), and since *x<1* there is a decrease in ABL-top relative humidity tendency and thus less cloud cover for drier soils (Figure 2), with increasing cloud cover for increasing soil moisture (*x≈0*). In contrast, for the set of model runs with decreased (weaker) atmospheric stability, ABL depth is deeper (as one would expect), and yet since *x>1* there is an increase in ABL-top relative humidity tendency and thus more cloud cover for drier soils, with decreasing cloud cover for increasing soil moisture (x>>1). Note that the largest values of ABL-top RH tendency and thus ABL cloud cover are achieved for a small evaporation fraction (lower soil moisture) with weak stability (x>>1), as was suggested in the relative humidity tendency development in the previous section.

These findings are qualitatively consistent with Ek and Mahrt (1994) for HAPEX-MOBILHY data (summer 1986, southwest France) which found that a day with strong atmospheric stability above the ABL and a large observed evaporative fraction (via higher soil moisture) gave a similar mid-day relative humidity at the ABL top as a case 9 days later with weaker atmospheric stability and soil moisture that had decreased by 20%.

4. SUMMARY

The role of soil moisture on ABL cloud development was explored in terms of a new ABL-top relative humidity tendency equation, where a number of land-surface and atmospheric processes interact. It was shown with good agreement between model runs, an analytical development, and finally analysis of Cabauw data, that the effect of soil moisture is to increase ABL-top relative humidity tendency and thus the potential for ABL cloud formation (given a sufficient initial ABL relative humidity, and air above the ABL is not too dry) only if the stability above the boundary layer is not too weak. On the other hand, for weak stability above the boundary layer, drier soils yield a greater ABL-top relative humidity tendency and thus cloud cover. There is great interest in the study of land-atmosphere interaction and a large number of data sets from many field programs representing diverse geophysical locations with which to study these interactions. The new relative humidity tendency equation presented here may provide a useful quantitative framework for future land-surface - ABL interaction studies in the formation of ABL clouds.

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