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1. INTRODUCTION

In this paper we discuss the impact of the coupling between the land-surface, boundary-layer (BL) parameterizations and rainfall on two different time and space scales: a local diurnal timescale and a regional seasonal timescale. The catastrophic flooding in the continental mid-West of the US in the summer of 1993 focused our attention on the coupling between soil moisture and rainfall on long-time scales, just as Cycle 48 of the ECMWF model was implemented. The development of cycle 48 of the ECMWF model was based extensively on the use of local test data-sets, including FIFE-1987 (Betts et al, 1993, Beljaars and Betts, 1992; Beljaars 1993). It included four predictive layers for soil moisture, as well as other changes discussed briefly below. We discovered a large sensitivity of monthly and seasonal precipitation over the US on initial soil moisture. Because of the profound importance of this on long-term rainfall prediction, we are presenting some preliminary results and a brief discussion of some of the feedbacks involved. This research has been further stimulated by the planning needs of the GCIP and BOREAS field programs, which address respectively the interaction between the surface and atmosphere for the Mississippi basin, and a cross section of the Canadian boreal forest between Saskatchewan and Manitoba.

We shall first illustrate the local diurnal feedback using the FIFE data, and a simple mixed boundary layer (BL) model. These show the dependence of the surface diurnal cycle of \( \theta, q \) and \( \theta_e \) on soil moisture and BL entrainment. Greater soil moisture leads to a higher afternoon maximum of \( \theta_e \), as does reduced entrainment. The diurnal cycle of \( \theta_e \) is in turn linked to precipitation and cloudiness. We then show an idealised mixed layer model comparison, which illustrates how drier soil moisture and increased BL entrainment have similar impacts on the BL diurnal evolution, of \( \theta \) and \( q \). The partition of the net surface radiation into sensible and latent heat fluxes does not affect the surface \( \theta_e \) flux, but it does impact both entrainment of low \( \theta_e \) air and BL-depth.

We then discuss the larger regional scale feed-

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fluxes and BL-top entrainment fluxes. Increasing these
entrainment fluxes (the coupling with the free
atmosphere) which increases the warming and drying
of the ABL, is difficult to distinguish from the effect of
low soil moisture, which shifts the partition of the
available energy from latent to sensible
heating. The surface fluxes and entrainment, however,
have different impacts on the $\theta_e$ balance of the BL.
The surface $\theta_e$ flux is related to the net radiation ($R_{Net}$)
at the surface (which is primarily affected by season
and cloudiness), so that for a given $R_{Net}$, changes in
soil moisture do not affect the surface $\theta_e$ flux.
Entrainment introduces low $\theta_e$ air into the BL.

For the low surface Bowen ratio average on the right,
$\theta$ and $q$ rise during the day, and reach a maximum in
the afternoon (near 1300 L or 1915 Z), before the
surface starts to cool. $\theta_e$ reaches a maximum of about
364K (see dashed isopleths) at this time. Moving left
across Fig 1, each subsequent average (with higher
surface Bowen ratio) reaches a warmer drier afternoon
maximum, but at a lower $\theta_e$. With the driest soil, for
which the surface Bowen ratio maximum is 0.89, the
maximum temperature is the highest, but the BL is
driest, and the $\theta_e$ maximum is the lowest, only
353K (reached at the surface before local noon). The
primary reason for this trend of maximum $\theta_e$ is not the
surface $\theta_e$ flux. The maximum net radiation minus
ground heat flux does decrease from 562 to 513 Wm$^{-2}$
with increasing surface Bowen ratio (Table 1), because
the net longwave (out) increases with the warmer
surface and drier atmosphere above: so there is
therefore a small decrease in the surface $\theta_e$ flux.
However there is a large increase in the entrainment of
dry (low $\theta_e$) air into the BL, driven by the increase in
surface heat flux with increasing surface Bowen ratio.
The simple mixed layer model in the next section will
confirm this analysis quantitatively.

Note that the initial condition near sunrise for
each average is also shifted, presumably because on
successive days of high (or low) evaporation, the BL
moistens (or dries); but there is little difference in the
morning temperature minimum, which is controlled by
the radiative balances over the whole day (not
discussed here).

2.3 Mixed layer model

An idealized set of calculations based on the
mixed layer equations from Betts (1992) will quantify
these local interactions. We ignore horizontal
advection as a climatological simplification. The
inversion base virtual heat flux is related to the surface
virtual heat flux using an entrainment coefficient $A_e$
as a closure [Betts, 1973; Carson, 1973; Tennekes,
1973]

$$F_{ib} = -A_e F_{ib}$$  \(1\)

Here $F$ denotes a heat flux in watts per square meter.
The local changes of mixed layer $\theta_e$ and $q_e$ are given
in terms of the flux difference between the surface
(subscript $s$) and the inversion (subscript $i$)

$$C_p \frac{\partial \theta_e}{\partial t} = \frac{g(F_{es} - F_{ei})}{\Delta p}$$  \(2a\)

$$L \frac{\partial q_e}{\partial t} = \frac{g(F_{qs} - F_{qi})}{\Delta p}$$  \(2b\)

where $\Delta p$, (defined positive) is the pressure depth of
the mixed layer. Defining surface and inversion level
Bowen ratios ($\theta_s$ and $g_s$), and using the closure (1) gives
(see Betts (1992))
For Each BR: ◊ 0.35 □ 0.54 ◯ 0.78 + 0.89 △ 350

Figure 1: Daytime (θ,q) path for FIFE 2-m averages from 1145Z to 2345Z for 4 surface Bowen ratios. θe lines are dashed.

Figure 2: Mixed layer model solution as function of surface Bowen ratio. Entrainment parameter Ar=0.4.

Figure 3: As fig 2, starting with Fig 1 initial states of (θ,q) just after sunrise.

Figure 4: As Fig 2 with lower entrainment parameter Ar=0.2.
where $\beta_v = -0.07$ is the slope of the dry virtual adiabat.

The leading terms in (3a) and (3b) are just the surface fluxes, which warm and moisten the ABL. The second pair of terms, proportional to $A_r$, are the entrainment fluxes of typically warm, dry air at the inversion. If we introduce the surface energy balance

$$\text{RNet-Grnd} = F_{se} + F_{si}$$

we can solve (3) and (4), given $\Delta p_i$. A simple solution for the growth of BL depth, $\Delta l_p;$, can be found by assuming the inversion strength ($\Delta \theta$) at BL top to remain constant

$$\Delta \theta = \theta^* - \theta_m = \text{constant}$$

where $\theta^*$ (just above inversion) increases as the BL deepens. Then

$$\frac{\partial \theta_m}{\partial t} = \frac{\partial \theta^*}{\partial t} - \Gamma \cdot \frac{\partial \Delta p_i}{\partial t}$$

where $\Gamma^* = -(\partial \theta / \partial p)^*$, just above the BL top inversion (we neglect the small (and partly canceling) effects of subsidence and radiative cooling on changing $\theta^*$ and $\Delta p_i$). Combining (3a), and (3b) gives $\partial \theta_m / \partial t$ for the mixed layer as

$$\frac{C_p}{\partial \theta_m} \frac{\partial \theta_m}{\partial t} - g \frac{F_{se}}{\Delta p_i} \left[ 1 + A_r (\beta_v - \beta_1) \right]$$

where $F_{se} = F_{se} + F_{si}$ if we approximate $\Delta \theta \approx 1$. This second constant in (7), $\beta_v = -1$ comes from the slope of the wet virtual adiabat (Betts, 1992).

Eqs. (3a), (3b) and (6) can be integrated from an initial morning condition (we used $\Delta p_i = 10$ mb), to give the profile of $(\theta_m, q_m)$ during the day. Note that the slope $\partial \theta_m / \partial q_m$ is only a function of $A_r$, $\beta_v$, and $\beta_1$. If these are constant, so is $\partial \theta_m / \partial q_m$.

Fig 2 shows the numerical integration from a fixed initial condition at sunrise for the four surface Bowen ratios, $\beta_j$ in Table 1, with all other parameters identical. A 30-min time-step was used. Table 2 shows the simple analytic functions for the surface heating and the diurnal dependence of $\beta_j$, $\beta_1$, which we used to approximate the FIFE July and August BL data in Betts and Ball, 1993. A 12-hr daylight period was used, and time after sunrise is denoted by $t$.

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**Table 2: Idealized Model Parameters**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>RNet-Grnd</td>
<td>540(1 - (t - 6)^2) W m^{-2}</td>
</tr>
<tr>
<td>$\beta_f$</td>
<td></td>
</tr>
<tr>
<td>$\beta_{(noon)}$</td>
<td>0.4</td>
</tr>
<tr>
<td>$\beta_{(6.5)}$</td>
<td>-0.6(1 - (6.5/12)^0.7)</td>
</tr>
<tr>
<td>$\beta_{(6.5)}$</td>
<td>-0.6(1 - (6.5/12)^0.7)</td>
</tr>
<tr>
<td>$\Gamma$</td>
<td>5 K/100 mb</td>
</tr>
</tbody>
</table>

Fig 2 shows a similar dependence on Bowen ratio as Fig 1. With a higher surface Bowen ratio, the afternoon state is warmer and drier with a lower $\theta_c$. Fig 3, (where we have integrated from the same morning conditions of $\theta$, $q$ as the Fig 1 data), matches Fig 1 better. The effect of the observed surface variation of net surface flux (Table 1) is very small (< 0.5 K in $\theta$), and we ignore it here. Fig 3 shows the mixed layer path of $\theta$, $q$, while Fig 1 shows that of $\theta$, $q$, at about 2m in the superadiabatic layer; nonetheless they are very similar. Together Figs 2, and 3 suggest that about half the afternoon $\theta$ difference in Fig 1 is due to the difference of surface Bowen ratio alone, and the rest to the different mixing ratio at sunrise.

Earlier ECMWF comparisons with the FIFE data (Betts et al, 1993, Fig 12) had shown a similar impact of soil moisture on the local daytime cycle of $(\theta, q)$. The development of cycle 48 discussed in the next section shows that there is a major regional feedback of soil moisture on precipitation on larger timescales.

### 2.4 Effect of Entrainment parameter

Fig 4 is the same as Fig 2, but the entrainment parameter $A_r$ has been reduced from 0.4 (the value found in FIFE) to 0.2 (the value used in many BL models). The reduced entrainment has a large impact on the afternoon $(\theta, q)$ state. Indeed it is clear that the effects of higher $A_r$ may be difficult to distinguish from lower $A_r$. For example the final state is similar for $\beta_j = 0.5$, $A_r = 0.2$ or $\beta_j = 0.35$, $A_r = 0.4$. Our FIFE analysis (Betts et al 1992, Betts and Ball, 1993) concluded that $A_r = 0.4$, but more coupled field studies of surface fluxes and BL entrainment are needed.

### 3. ECMWF FORECASTS FOR JULY 1993: REGIONAL INTERACTION

The ECMWF model cycle 48 (Beljaars, 1993) introduced major changes to the land surface parameterization, as well as a new boundary layer parameterization and changes to the air-sea interaction formulation, following Miller et al, 1992. The changes over land were prompted by comparisons with field data (Betts et al, 1993; Beljaars and Betts, 1992). The changes most relevant to this paper were the
introduction of four prognostic layers for the land surface hydrology, a new skin temperature and greater entrainment at the top of the unstable daytime BL.

The final "E-suite" testing of the ECMWF model cycle 48 started on July 1, 1993 and was followed by operational implementation in August. In retrospect, this was fortuitous in view of the subsequent catastrophic flooding in the Mississippi basin in July; because comparison of the old and new cycles of the model showed the impact of the new land surface parameterization in cycle 48. Furthermore, global initialization of the soil moisture was a key issue in starting the new model cycle; as it showed the sensitivity of forecasts to initial soil moisture. An attempt was made first to derive a global climate for soil moisture to initialize the new operational model by running the new model for four years at T-63 resolution, and averaging the four annual cycles. However, it was clear that this model "soil climate" was too dry, and the precipitation over land was too low in this 4-year run. It is believed that this results at least in part from long term interactions with errors in the incoming net radiation at the surface (probably coming in turn from too little cloudiness over land). Since this spring and summer were abnormally wet over parts of the central US of Europe, initializing the forecast model with too dry soil moisture fields produced large positive errors in the daytime temperature maxima over the summer continents. So a series of experiments were run with higher soil moisture fields using 1992 data, which showed a remarkable seasonal impact of soil moisture on precipitation. Specifying initially moist soils on May 1, 1992 gave almost double the summer (3-mo) precipitation over the central US (in 120 day T-63 runs), in comparison with forecasts initialized with the dry "climate" soil moisture. Consequently the soil moisture in vegetated areas was increased to field capacity to start the new operational model cycle 48 in August, 1993.

This prompted us to explore the impact of soil moisture on the July 1993 rainfall over the central US. Since cycle 48 analyses were available from July 1, 1993, we ran two one month forecasts from this date at T-106 resolution starting with the same atmospheric data, but different soil moisture. The "wet" forecast started with vegetated soils at field capacity. In the model, this gives "unstressed" evaporation by the vegetation. The second "dry" forecast had soil moisture set so that the vegetative resistance to evaporation was 4x the unstressed value. The impact on the July mean precipitation was remarkable in the central US. Fig 5 is the forecast (T2A) from the wet soil condition; Fig 6 (T3L) is from the dry soil condition. The monthly mean precipitation for the central US (in mm/day) from the initially wet soil simulation is roughly double that from the initially dry soil. Fig 7 shows the difference field between the two simulations: showing a mean increase of precipitation of 4mm day$^{-1}$ over the central US, and decrease to the South over Texas. Fig 8 shows the actual July rainfall in inches from the NOAA Climate Analysis Center. The similarity of this precipitation pattern over the central US and Texas to the July simulation from wet soils (Fig 5) is striking (4 mm/day = 5 inches/mo), although the high precipitation in the monthly "forecast" (from a global model at T-106 resolution) does not extend as far south into Kansas and Missouri as was observed. The precipitation decrease over Alabama and Georgia, shown in Fig 7, mirrors the abnormally low rainfall in July in that region shown in Fig 8, although the "forecast" precipitation still exceeds that observed.

The increase in precipitation has not come from simply evaporating the initial soil moisture, since the differences in soil moisture are maintained for the month (not shown). The mechanism probably involves the higher BL $\theta_v$ over moist soils (shown in the previous section), the increase in convection and precipitation (tied to this higher moist adiabat) heating the atmosphere over a large enough horizontal scale to induce increased moisture convergence. This large-scale feedback is being studied further.

3. CONCLUSIONS

The result shown in Fig 7 is very significant, as it suggests that the coupling of BL equilibrium and precipitation with soil moisture has a long timescale, which may yield some long-term predictability in precipitation over the summer continents, provided we have adequate analyses of soil moisture. These were not available for the start of the ECMWF model cycle 48, (so the model was initialized with soil moisture fields that were partly assumed); but it is hoped that over the next fall, winter and spring the precipitation generated by the model initialization will generate and maintain realistic soil moisture fields.

Acknowledgements

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References

Betts, A.K., Non-precipitation cumulus convection and its parameterization, J. G. R., 97, 18523-18532.
Fig 5  30-day average precipitation (mm/day) from ECMWF Cycle 48 T-106 forecast from July 1, 1993, using wet July 1 soil moisture, corresponding to unstressed evaporation. Contours are at 1, 2, 4, 8 mm/day.

Fig 6  As Fig 5 for forecast initialised with dry soil moisture, corresponding to stressed evaporation.
Fig 7 Difference (Fig 5-Fig 6) showing increase in monthly mean precipitation resulting from increase in initial soil moisture.

Fig 8 NOAA Climate Analysis Centre map of observed July precipitation in inches.


