IMPACT OF ATMOSPHERIC SURFACE-LAYER PARAMETERIZATIONS IN THE NEW LAND-SURFACE SCHEME OF THE NCEP MESOSCALE ETA MODEL

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Abstract. We tested three atmospheric surface-layer parameterization schemes (Mellor-Yamada level 2, Paulson, and modified Louis), both in a 1-D mode in the new NCEP land-surface scheme against long-term FIFE and HAPEX observations, and in a coupled 3-D mode with the NCEP mesoscale Eta model. The differences in these three schemes and the resulting surface exchange coefficients do not, in general, lead to significant differences in model simulated surface fluxes, skin temperature, and precipitation, provided the same treatment of roughness length for heat is employed. Rather, the model is more sensitive to the choice of the roughness length for heat. To assess the latter, we also tested two approaches to specify the roughness length for heat: 1) assuming the roughness length for heat is a fixed ratio of the roughness length for momentum, and 2) relating this ratio to the roughness Reynolds number as proposed by Zilitinkevich. Our 1-D column model sensitivity tests suggested that the Zilitinkevich approach can improve the surface heat flux and skin temperature simulations. A long-term test with the NCEP mesoscale Eta model indicated that this approach can also reduce forecast precipitation bias. Based on these simulations, in January 1996 we operationally implemented the Paulson scheme with the new land-surface scheme of the NCEP Eta model, along with the Zilitinkevich formulation to specify the roughness length for heat.

Key words: Surface-layer parameterization, Land-surface process, Roughness length for heat, Soil moisture simulation, Numerical weather prediction

1. Introduction

At the National Center for Environmental Prediction (NCEP/NOAA), formerly the National Meteorological Center (NMC), a series of initiatives has been undertaken (Mitchell, 1994) to, a) upgrade the land-surface physics coupled to the NCEP operational mesoscale Eta model, and b) provide sound mesoscale 4-D data assimilation and forecast products for the research community of the Global Energy and Water Cycle Experiment (GEWEX) Continental-Scale International Project (GCIP). These initiatives include 1) developing, testing, and implementing a modern soil/vegetation/hydrology land-surface model suitable for an operational mesoscale numerical model over a continental domain, and 2) improving the atmospheric surface-layer parameterization in the Eta model to better accommodate this new land-surface package.

In recent forerunner studies, we have addressed the first initiative above, wherein we extended the soil/vegetation model of Mahrt and Pan (1984) and Pan and

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Mahrt (1987) to include more comprehensive canopy resistance and surface runoff treatments and tested the extended model in 1-D uncoupled mode against long-term field observations of FIFE (Chen et al., 1996) and CABAUW (Chen et al., 1997). This new NCEP land-surface model can, in general, reasonably reproduce the observed diurnal variation of surface heat fluxes and surface skin temperature important for mesoscale forecasts. The model is also able to capture the seasonal evolutions in evaporation and soil moisture, which is crucial for soil moisture initialization via long-term 4-D data assimilation systems. By contrast, as shown in Chen et al. (1996), the bucket model, which is conceptually similar to the land-surface model in the formerly operational NCEP Eta model, reduced the soil water storage available for summer by overestimating evaporation during wet (soil recharge) spring periods, and thus underestimated evaporation during summer drying periods.

This new land-surface model was further tested in a coupled mode with the Eta model, and preliminary results showed that it improved the Eta model precipitation forecast skill. In coupled mode, however, the atmospheric surface-layer parameterization is a crucial component influencing the success of a soil/vegetation model, as it serves as the link between the land-surface and the lowest atmospheric level in the host numerical model. Therefore, in the present study, we address the second aforementioned initiative, i.e., improve the atmospheric surface-layer parameterization in the operational NCEP Eta model, which formerly used a derived Mellor-Yamada Level-2 scheme (Łobocki, 1993) and assigned the same roughness length for momentum and heat transfer.

During the last 20 years, many atmospheric surface-layer parameterization schemes have been developed for use in general circulation models (GCMs) and mesoscale models (e.g., Businger et al., 1971; Paulson, 1970; Dyer, 1974; Łobocki, 1993). In these schemes, the stability functions as well as the resulting surface exchange coefficients appear to have different forms. Since these parameterization schemes are applied in atmospheric numerical models, it is desirable to understand their inherent differences and different impact on surface flux calculations and numerical model forecasts. Additionally important, many researchers have suggested that, in the surface-layer parameterization, the roughness length for heat (moisture) must be different from that for momentum (Brutsaert, 1982; Garratt, 1992; Sun and Mahrt, 1995; Mahrt, 1996). Some recent studies (e.g., Braud et al., 1993; Beljaars and Viterbo, 1994) on 1-D uncoupled land-surface model simulations showed that using a value of roughness length for heat smaller than that for momentum could improve the simulated surface fluxes and surface skin temperature compared to observations. Nevertheless, the advantage of employing a different roughness length for heat in a coupled mesoscale model was unclear heretofore.

For the above reasons, our focus in this paper is to examine and evaluate alternative atmospheric surface-layer parameterization schemes with the new NCEP soil/vegetation model, both in coupled and uncoupled mode. Therefore, we test three surface-layer parameterization schemes that have been widely used in GCMs and mesoscale models, and also evaluate different approaches in specifying the roughness length for heat. In these tests, we employ both 1-D uncoupled land-surface model runs and 3-D coupled mesoscale model runs, the latter using the NCEP operational Eta model (Black, 1994; Janjić, 1994). We verify the 1-D land-surface model simulations against FIFE (Sellers et al., 1992; Betts and Ball; 1993) and HAPEX-MOBILHY (André et al., 1986) observations.

In this paper, we do not intend to evaluate these surface-layer parameterizations from a theoretical or experimental viewpoint. Rather, our practical objective is to understand the impact of different surface-layer parameterization schemes and the roughness length formulations on surface heat fluxes and surface skin temperature simulations. In particular, we try to assess these impacts on the accuracy of mesoscale short-range forecasts and to provide some insights on the practical treatment of these problems in numerical weather prediction (NWP) models.

2. Method and Data

2.1. Three surface-layer parameterizations to test

A surface-layer parameterization should provide the surface (bulk) exchange coefficients for momentum, heat, and water vapor used to determine the flux of these quantities between the land-surface and the atmosphere. From the flux-profile relationships based on observations (e.g., Businger et al., 1971; Dyer, 1974), different integrated stability functions are proposed to obtain surface exchange coefficients (e.g., Paulson, 1970). Although these parameterization schemes were intended to fit observations, the integrated stability functions as well as the resulting surface exchange coefficients proposed by various investigators appear to have different forms. According to the review of Yaglom (1977), these discrepancies were generally believed not to be real differences, but rather due to instrument and measurement problems. In trying to eliminate instrumental problems, Högström(1988) conducted a carefully designed experiment to re-evaluate some of these formulations. He found that for unstable conditions, the formulae of Businger et al. (1971) and Dyer (1974) agree with each other to within $\pm 10\%$, but the various formulae still disagree to a greater extent for stable conditions.

To understand their impact on surface heat flux calculations and numerical model forecasts, we select three surface-layer parameterization schemes to test in this study. These schemes include: 1) an implicit scheme based on the Obukhov length (Paulson, 1970; Garratt, 1992, herein called the Paulson scheme), 2) another implicit scheme based on the Obukhov length but derived from the Mellor-Yamada level-2 formulation (Łobocki, 1993, herein called the Mellor-Yamada scheme), which was used in the Eta model (Janjić, 1990, 1994) prior to the outcome of this study, and 3) an explicit scheme based on a bulk Richardson number, which was first developed by Louis (1979) and then modified by Mahrt (1987) for the stable case and by Holtslag and Beljaars (1989) for the unstable case (herein called the modified Louis scheme). These schemes (or variations of them) have been widely used in atmospheric numerical models (e.g., see Pielke, 1984; Garratt, 1992; Beljaars and Viterbo, 1994; Janjić, 1994; Black, 1994; Mascart et al., 1995). A detailed description of these three surface-layer schemes is given in Appendix A.

2.2. Specification of Roughness Length for Heat

To obtain the traditional bulk aerodynamic formulation estimating surface fluxes, one must integrate the Monin-Obukhov similarity theory between a reference height within the surface layer (as the upper boundary) and the roughness length (height) for momentum, heat, and water vapor (as the lower boundary). Commonly in numerical models, the air temperature at the roughness height (aerodynamic temperature) is often replaced by the surface radiative (skin) temperature which can be readily computed from the surface energy budget. The surface skin temperature is, however, 2–6 °C higher (lower) than the inferred aerodynamic temperature under unstable (stable) atmospheric conditions (Sun and Mahrt, 1995). Therefore, the thermal instability is overpredicted (underpredicted) by using the surface radiative temperature for the unstable (stable) case.

As reviewed by Sun and Mahrt (1995), there are four existing ways to remedy problems resulting from using the surface skin temperature as the lower boundary condition. Among them, one solution, which is easy to implement in atmospheric numerical models, is to formulate an empirical relation between the surface aerodynamic temperature and the surface radiative temperature. This is equivalent to using a value of roughness length for heat, z_{0t}, and water vapor, z_{0q}, different from that for momentum z_{0m} (extensive discussions can be found in Brutsaert, 1982, Chapter 5; Garratt, 1992, Chapter 4; Sun and Mahrt, 1995). An additional motivation for such a distinction between z_{0m} and z_{0t} (z_{0q}) is that the heat transfer and momentum transfer are controlled by essentially different mechanisms in the roughness layer. While the momentum flux is induced primarily by a pressure gradient (form drag), the heat (moisture) flux is completely due to the heat conductivity within the viscous sublayer adjacent to the roughness elements (Brutsaert, 1982; Zilitinkevich, 1995). The final motivation for this distinction is to take into account the subgrid-scale variability in the ratio of z_{0m}/z_{0t} (e.g., Beljaars and Holtslag, 1991; Hopwood, 1995; Mahrt, 1996). For instance, based on observations conducted above a typical inhomogeneous and vegetated land surface, Hopwood (1995) suggested that the ratio of z_{0m}/z_{0t} is about 80.

While the idea of differentiating z_{0t} from z_{0m} is supported by a number of observations (Brutsaert, 1982; Beljaars and Holtslag, 1991; Hopwood, 1995; Sun and Mahrt, 1995; Mahrt, 1996), how to specify z_{0t} in the numerical models applied to continental scales is a more challenging problem. A simple approach is to assume that z_{0t} is a fixed fraction of z_{0m}. For example, Garratt (1992) suggested

$z_{0m}/z_{0t} \approx e^2$ for practical application. Based on their 1-D tests against observations over different sites, Braud et al. (1993) and Beljaars and Viterbo (1994) found that using $z_{0m}/z_{0t} = 10$ can improve model simulated surface fluxes and surface skin temperature. However, a specification of the ratio z_{0m}/z_{0t}, according to them, would have to be prescribed as a function of vegetation and soil types. On the other hand, for an aerodynamically smooth flow, z_{0m} is comparable with, or even less than, z_{0t} (e.g., see Garratt, 1992).

Presumably, the more physical approach is to relate the ratio z_{0m}/z_{0t} to the property of flow, since this ratio depends on the difference between the surface radiative temperature and air temperature. Some formulae (e.g., Kubota and Sugita, 1994) define z_{0t} as a function of solar elevation, solar radiation, leaf area index, and canopy height among many others. However, others (e.g., Brutsaert, 1982; Garratt, 1992; Kubota and Sugita, 1994; Zilitinkevich, 1995) suggest parameterizations relating the ratio z_{0m}/z_{0t} to heat fluxes or roughness Reynolds number. It is worth bearing in mind that we are seeking an approach that should be easily implemented in operational NWP models. Thus, in this study, we test a fixed ratio of z_{0m}/z_{0t} as well as a Reynolds number-dependent formulation proposed by Zilitinkevich (1995), due to its simplicity. This formulation is written as:

$$\frac{z_{0m}}{z_{0t}} = \exp(k\ C\ \sqrt{Re^*}) \tag{1}$$

$$Re^* = \frac{u_0^* z_{0m}}{\nu}$$

where k is the von Kármán constant ($k = 0.4$), ν is the kinematic molecular viscosity, Re^* is the roughness Reynolds number, and u_0^* is the surface friction velocity. C is an empirical constant and we will discuss it in Section 3.2.

2.3. Land-Surface Model

We use a multi-layer soil/vegetation model based on the work of Mahrt and Pan (1984) and Pan and Mahrt (1987). This model is extended to include an explicit plant canopy resistance treatment that depends upon not only soil moisture but also on atmospheric conditions such as solar insolation, water vapor pressure deficit, and air temperature (Kim and Ek, 1995; Chen et al., 1996). Chen et al. (1996) demonstrated that incorporation of such a diurnally varied canopy resistance can effectively reduce the overestimation of evaporation during wet periods, retain the water in the soil, and release it to provide enough evaporation during drying periods. This, together with a new runoff formulation based on a subgrid-scale hydrologic approach (Schaake et al., 1996; Chen et al., 1996), significantly improved the model's ability to simulate the long-term evolution in evaporation and soil moisture. More detailed information on this land-surface model can be found in Appendix B.

For the FIFE and HAPEX-HOBILHY uncoupled runs, the model has one canopy layer and utilizes four soil layers, namely: a thin top layer of 10 cm, a second root zone of 10 cm, a deep root zone of 80 cm, and a sub-root zone of 1 m. This configuration is intended to simulate the diurnal, weekly, and seasonal variation in soil moisture. With such a configuration, this model has 10 prognostic variables: soil moisture and temperature in the four layers, water stored on the canopy, and snow stored on the ground.

For the coupled runs with the NCEP Eta model, this land-surface model utilizes two soil layers: a thin top layer of 10 cm and a deep root zone of 190 cm. This latter soil layer configuration matches that used in the NCEP Global Data Assimilation System (GDAS), whose two layer soil moisture and temperature are used to initialize the NCEP mesoscale Eta model.

2.4. Data used

2.4.1. *FIFE Data Set*

The FIFE data set used in this study is a single spatial-mean time series, with a time interval of 30 min, derived by Betts and Ball (1993) by averaging data collected over different stations in the FIFE area of 15 km $\times$ 15 km. The PAM (Portable Automated Mesonet) station time-series data (30-minute averages at about 10 stations) consisted of atmospheric forcing data (spanning 22 May to 16 October 1987) such as wind, air temperature and humidity, precipitation, incoming and reflected solar radiation, net radiation and incoming long-wave radiation, and a radiometric measure of the ground surface temperature.

For validation, this data set also includes the spatial-mean surface sensible heat, latent heat, and soil heat fluxes averaged over 17 selected surface-flux stations. The averaged heat flux observations from the flux stations, only available for four Intensive Field Campaigns (IFCs) (26 May–6 June, 25 June–11 July,6 Aug–21 Aug, and 5 Oct–16 Oct) were extensively used to validate the model simulated fluxes. For validation in-between these IFC periods, we used the longer-term (from 27 May to 16 October) but spatially less representative (two stations) time series of surface fluxes of Smith et al. (1992). As shown by Chen et al. (1996), Smith's sensible and latent heat flux measurements are higher than the IFC area-averaged fluxes, partly due to higher net radiation values over Smith's stations.

2.4.2. *HAPEX-MOBILHY Data Set*

We used the same HAPEX-MOBILHY data set prepared by Drs. Mahfouf and Noilhan and used for PILPS Phase 2(b) (Shao and Henderson-Sellers, 1996). The data were obtained from HAPEX-MOBILHY at Caumont (SAMER No. 3) (Goutorbe, 1991). The set consists of a full year (1986) of the same atmospheric forcing terms as aforementioned for the FIFE data set. The surface energy flux measurements are available for the Intensive Observation Period (IOP) (28 May to 3 July, 1986). Net radiation, soil heat flux, and sensible heat fluxs are directly measured, while latent heat flux is obtained from the surface energy balance.

According to Goutorbe (1991), the accuracy of the flux measurements is about 15% at short time scales and about 10% at large time scales.

One unique aspect of the HAPEX-MOBILHY data set is its one-year continuous soil moisture observations. From the surface down to 1.6 m, the soil moisture is measured weekly at every 0.1 m by neutron sounding probes. Shao and Henderson-Sellers (1996) pointed out that $\pm 10\%$ error margins for soil moisture measurements are reasonable estimates. However, using the soil moisture to validate the land-surface models may be challenged by the water imbalance at the local scale. For instance, Mahfouf (1990) studied the surface water budget using data from several HAPEX-MOBILHY sites and found that at the Caumont station, the water imbalance (i.e., total precipitation – total evaporation – soil moisture change) is -25 mm for the IOP. This large negative imbalance may, according to Mahfouf (1990), indicate an extra supply of water from below or laterally and may also be due to irrigation applied for this soya field. Thus, in order to avoid any serious drift between model and data so that we could use the critical surface fluxes observed during the IOP to validate our model, we forgo full 1-year simulation and instead initialize the model with the observed soil moisture on 28 May 1986 and examine the 7-month simulations during and after the roughly one-month IOP.

3. Results and Discussions

3.1. Tests of Three Surface-Layer Parameterizations

We applied three surface-layer schemes (i.e., the modified Louis scheme, the Mellor-Yamada level 2 scheme, and the Paulson scheme) to the FIFE case with the new NCEP land-surface model. In these first applications, all three schemes assume the ratio $z_{0m}/z_{0t} = 10$ (we will discuss the model sensitivity to this ratio in Section 3.2). In Figure 1, we see that the Mellor-Yamada (M-Y2) and Paulson schemes produce nearly the same C_h coefficients. Although all three schemes produced very similar C_h, the modified Louis scheme generally produced slightly higher (lower) C_h than those calculated by the other two schemes for stable (unstable) conditions. Since the modified Louis scheme by Mahrt (1987) for stable conditions considers the intermittency of turbulence caused by subgrid-scale variability such as slope flow, the decrease in C_h with increasing stability is slower than the other two schemes. The latent and sensible heat fluxes obtained by these three schemes in the NCEP land-surface model for the FIFE and HAPEX-MOBILHY cases were also very similar (an example of the diurnal variation of latent heat fluxes in FIFE is shown in Figure 2).

Turning next to the impact of the surface-layer scheme on the long-term performance of the soil/vegetation model, we compare the 7-month simulated soil water content for the first 1 m depth of soil against HAPEX-MOBILHY observations in Figure 3. Again, the results from the three surface-layer schemes are very close.

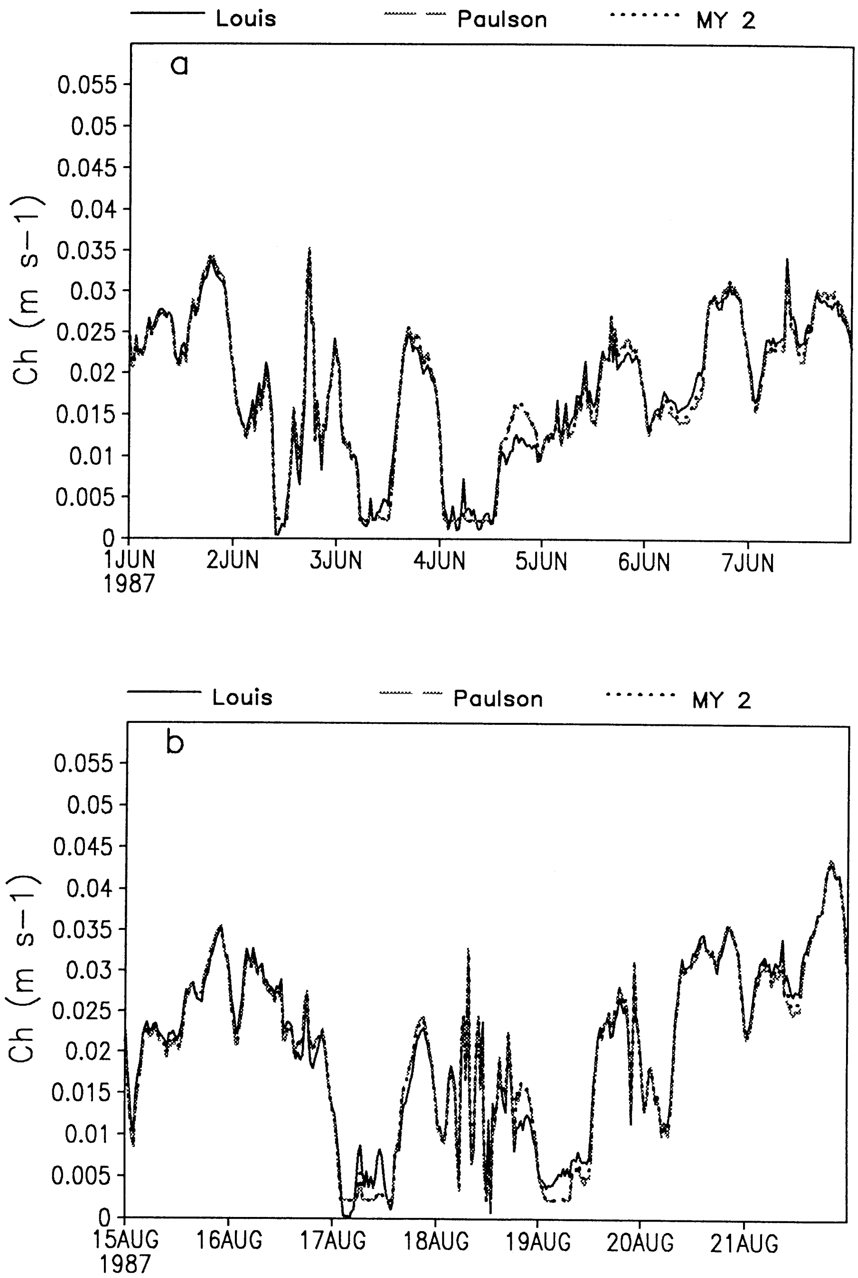


Figure 1. Diurnal variation of surface exchange coefficient for heat (C_h in m s^{-1}) obtained by the modified Louis scheme, the Paulson scheme, and the Mellor-Yamada level 2 scheme respectively for the FIFE case. We assume $z_{0m} = 4.5$ cm and $z_{0m}/z_{0t} = 10$. (a) One week in June, 1987, and (b) one week in August, 1987.

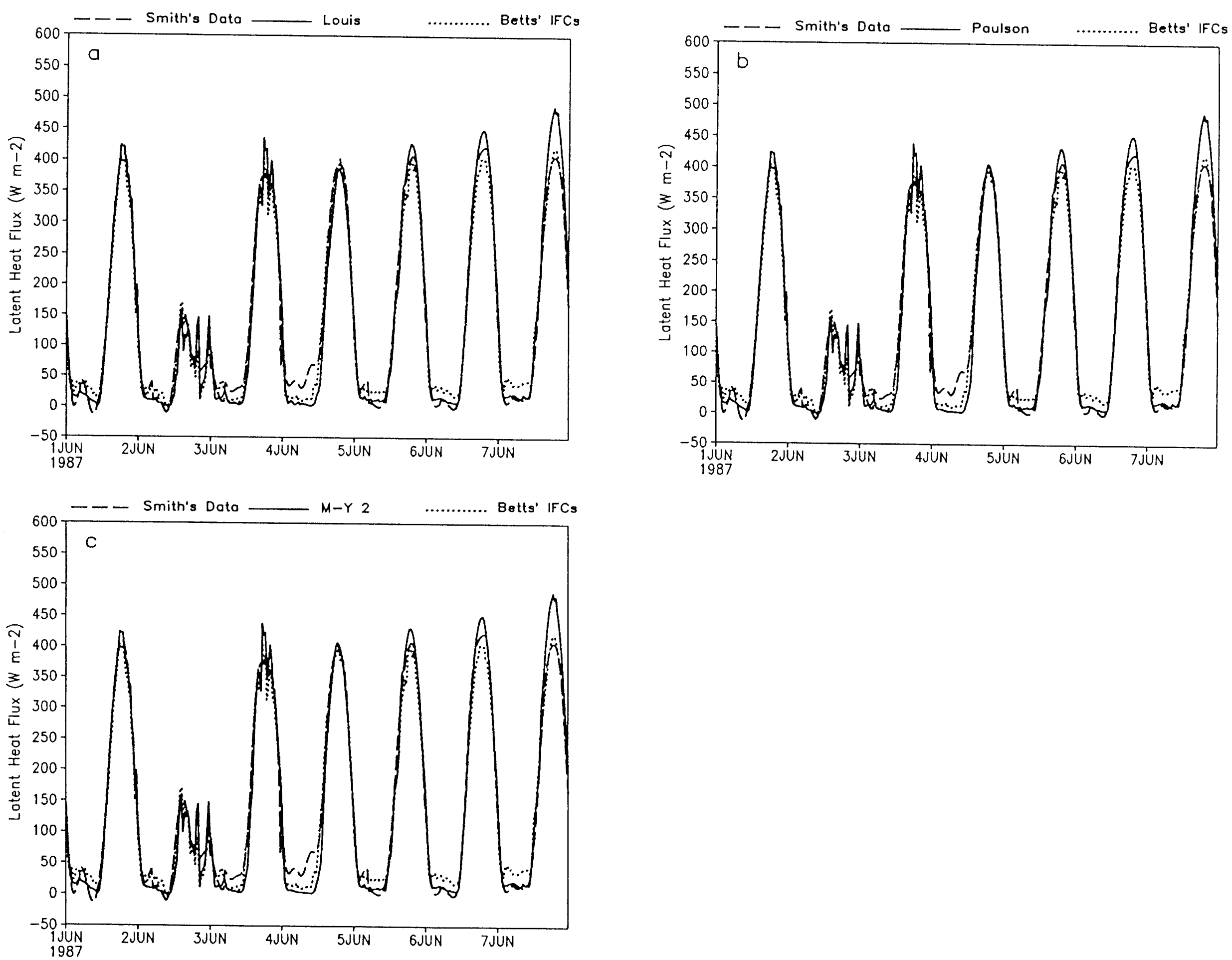


Figure 2. Comparison of the diurnal variation of latent heat flux between simulation and FIFE observations for 1–7 June 1987, using the modified Louis scheme, the Paulson scheme and the Mellor-Yamada Level 2 scheme.

Trying to simulate the long-term evolution in soil moisture is not a simple task even when using observed precipitation. It requires the correct estimation of evaporation and runoff. Among the many factors such as canopy resistance, surface-layer formulation, and runoff which can affect soil moisture simulation (Chen et al., 1996), the vertical distribution of vegetation root density is also important. According to a sensitivity test in which we specified uniformly distributed roots throughout the root zone, the simulated *total* soil water over 1 m depth compared favorably to HAPEX-MOBILHY measurements, but the shallow soil layers (0–0.2 m) were too wet and the deep layer (0.2–1 m) was too dry. Following the PILPS recommendation (Shao and Henderson-Sellers, 1996), we distributed 60% of the roots in the two top 0.2 m soil layers and placed the remaining roots in the deep root zone between 0.2 to 1 m in our 4-layer soil model. This significantly improved the evolution of soil moisture in the three root zone layers (not shown).

As we have thus far shown that the C_h values obtained from the MY-2 and Paulson schemes are very close (see Figure 1), we hereafter focus on the comparison between the Paulson scheme and the modified Louis scheme for the Eta model coupled runs. Next in this study, the mesoscale Eta forecast model, having a horizontal resolution of 80 km and 38 atmospheric vertical levels, was executed with the new NCEP land-surface scheme to produce a 48-h regional forecast covering most of North America. In these simulations, everything is identical except the choice of surface-layer parameterization scheme, wherein the ratio z_{0m}/z_{0t} is still assumed to be 10. Figure 4 displays a comparison of latent heat fluxes between the Eta model using the Paulson scheme and the one using the modified Louis scheme. The results are shown for 18Z 6 June 1987, i.e., the 42-h forecast time starting from 00Z 5 June 1987. Over most of the continental area, the difference between these two schemes is less than 10 W m^{-2}. Similarly close agreement was found in the sensible heat flux and skin temperature (not shown here).

Based on the above 1-D uncoupled tests against observations and 3-D coupled model simulations, these three atmospheric surface layer schemes produced insignificant differences in the surface heat fluxes and skin temperature. We selected the Paulson scheme as the new NCEP Eta model surface layer parameterization. Although the Paulson scheme is computationally more expensive, it is easier to modify its stability functions should more accurate and/or extensive field observations in the future motivate refinements to the stability functions.

3.2. Model sensitivity to z_{0m}/z_{0t}

In the previous 1-D and 3-D model simulations discussed in Section 3.1, we applied the constant ratio of $z_{0m}/z_{0t} = 10$ for roughness length. In this section, we use different values of the ratio z_{0m}/z_{0t} as well as the equation proposed by Zilitinkevich (1995) to calculate this ratio as a function of roughness Reynolds number (see Equation (1)) in the Paulson scheme. In Equation (1), the constant C modulates how strongly the z_{0m}/z_{0t} ratio depends on roughness Reynolds

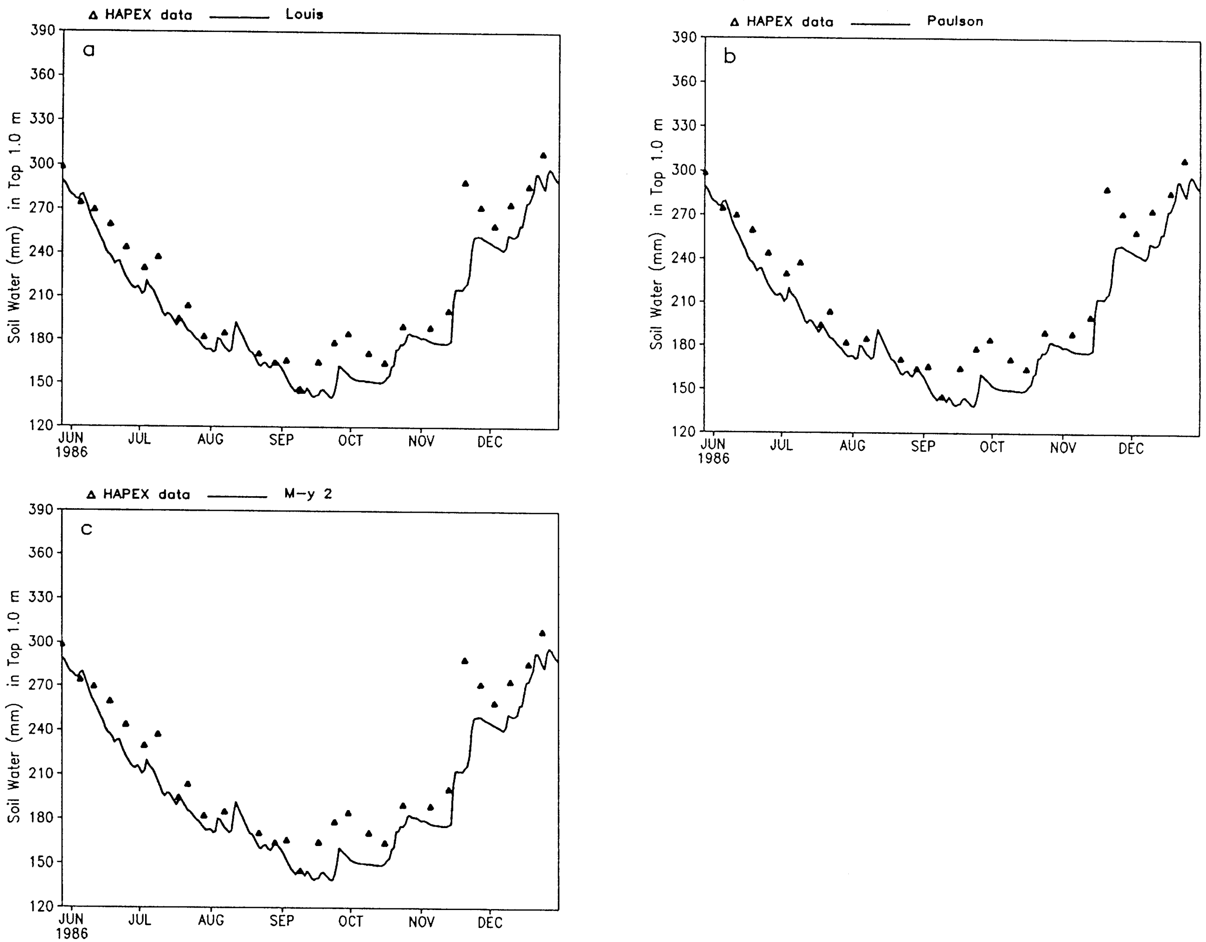


Figure 3. Seasonal evolution (from 28 May to 31 December 1986) of daily averaged water content (in mm) in the first 1 metre depth of soil: a comparison between HAPEX-MOBILHY observations and three simulations.

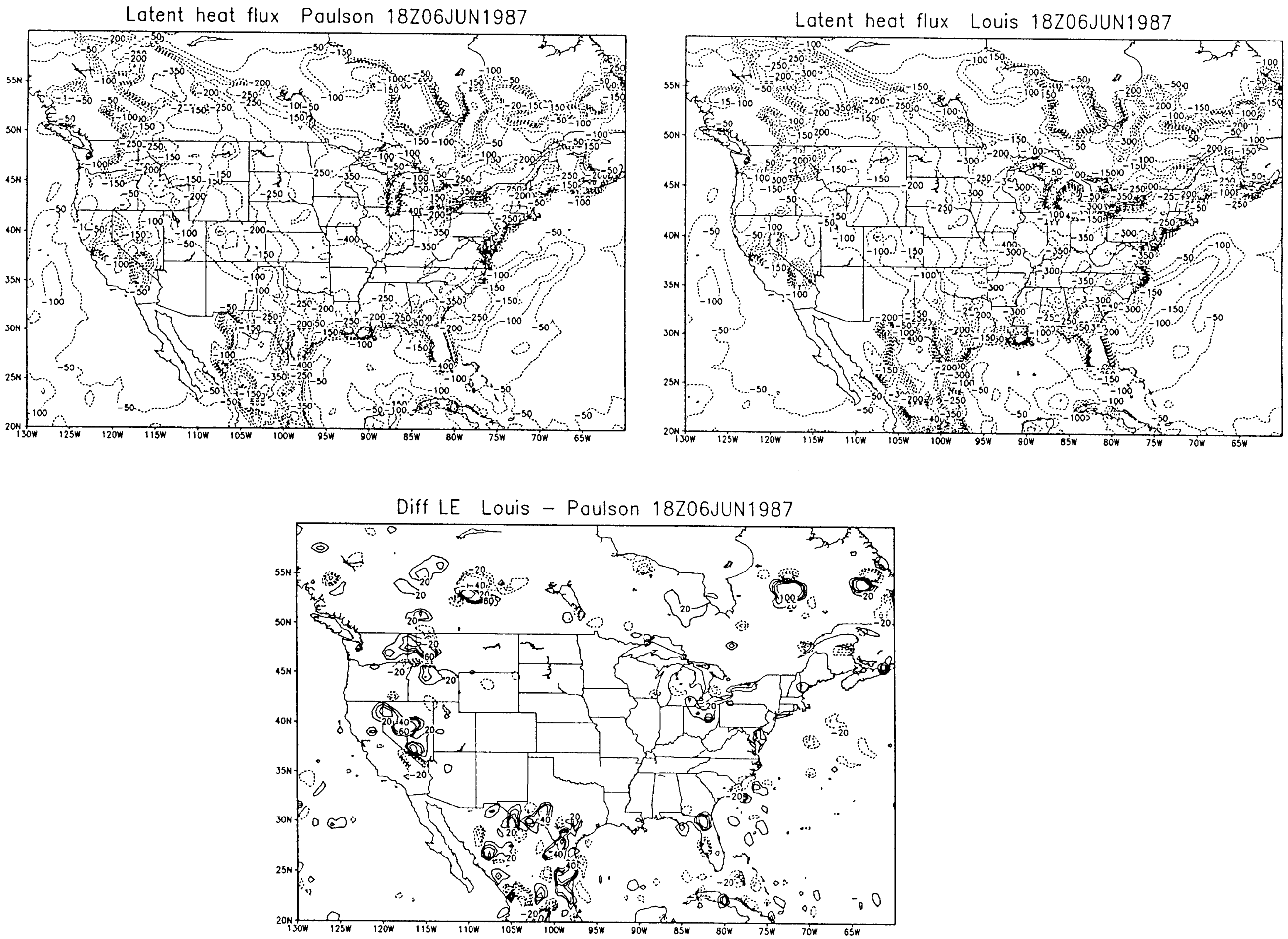


Figure 4. Eta model forecast of latent heat flux (in W m^{-2}, upward is negative) at 18Z, 6 June 1987 (the 42-h forecast time) obtained from two surface layer schemes (the Paulson scheme and the modified Louis scheme).

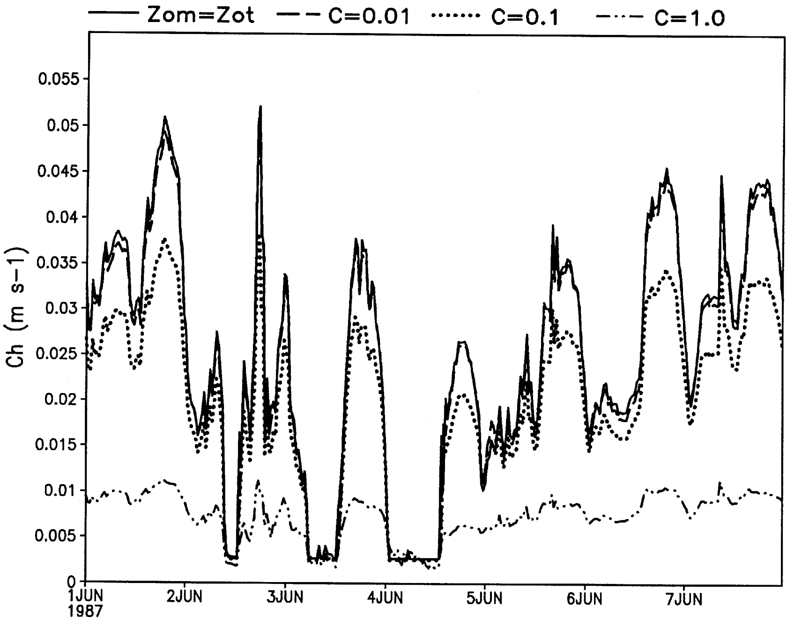


Figure 5. Comparison of diurnal variation of (a) surface exchange coefficient, (b) latent heat flux, (c) sensible heat flux, and (d) surface skin temperature between FIFE observations for 1–7 June 1987 and simulations with the following different z_{0t}: $z_{0m} = z_{0t}$; $C = 0.01$ in Zilitinkevich formulation (refer Equation (1)); $C = 0.1$; and $C = 1.0$. The Paulson scheme is used in these simulations.

number Re^{*}. For instance, given typical Re^{*} values ranging from 10 to 10^4 in the atmosphere, the z_{0m}/z_{0t} ratios obtained in Equation 1 range from 1.01 to 1.49 using $C = 0.01$, from 1.13 to 54.6 using $C = 0.1$, and from 3.54 to 2.35×10^{17} using $C = 1$. For small C values (e.g., $C < 0.01$), z_{0m}/z_{0t} tends to be 1 throughout the physical range of Re^{*}, while large C drastically decreases z_{0t} and, as a result, reduces the surface exchange coefficient C_h under high roughness Reynolds number regimes.

As an example to illustrate the impact of different z_{0m}/z_{0t} ratios on C_h, Figure 5a shows C_h values obtained by using different z_{0m}/z_{0t} ratios in the Paulson scheme applied for the FIFE case. The high C_h values from using $C = 0.01$ are close to those from using $z_{0m} = z_{0t}$. The intermediate C_h values from using $C = 0.1$ are close to those of Figure 1a formerly obtained with the fixed ratio of $z_{0m}/z_{0t} = 10$. The extreme simulation using $C = 1$ reduces C_h values by more than four times as compared to the simulation using $C = 0.01$. The effect of large C values in reducing C_h is more apparent in late July for the FIFE case (not shown) when the soil becomes very dry and the skin temperature is higher than the June case discussed here.

In Figures 5b–d, we compare surface heat fluxes and skin temperature for this FIFE case obtained by using different z_{0m}/z_{0t} ratios in the Paulson scheme. When $C = 0.01$, in most cases, the latent and sensible heat fluxes are not different from

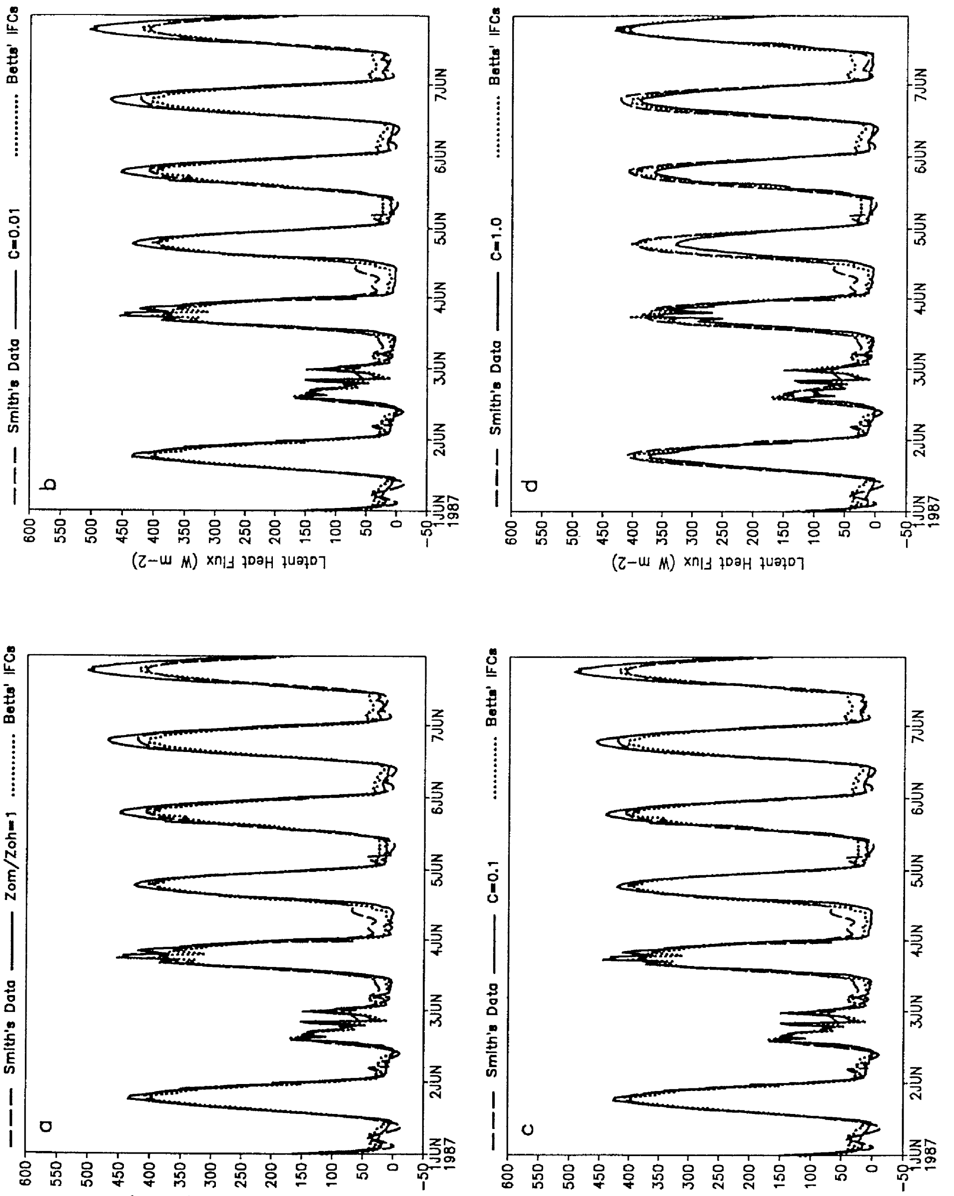

a
b
c
d
Smith's Data
Zom/Zoh=1
C=0.01
C=0.1
C=1.0
Betts' IFCs
Latent Heat Flux (W m-2)
1JUN 1987
2JUN
3JUN
4JUN
5JUN
6JUN
7JUN

Figure 5b.

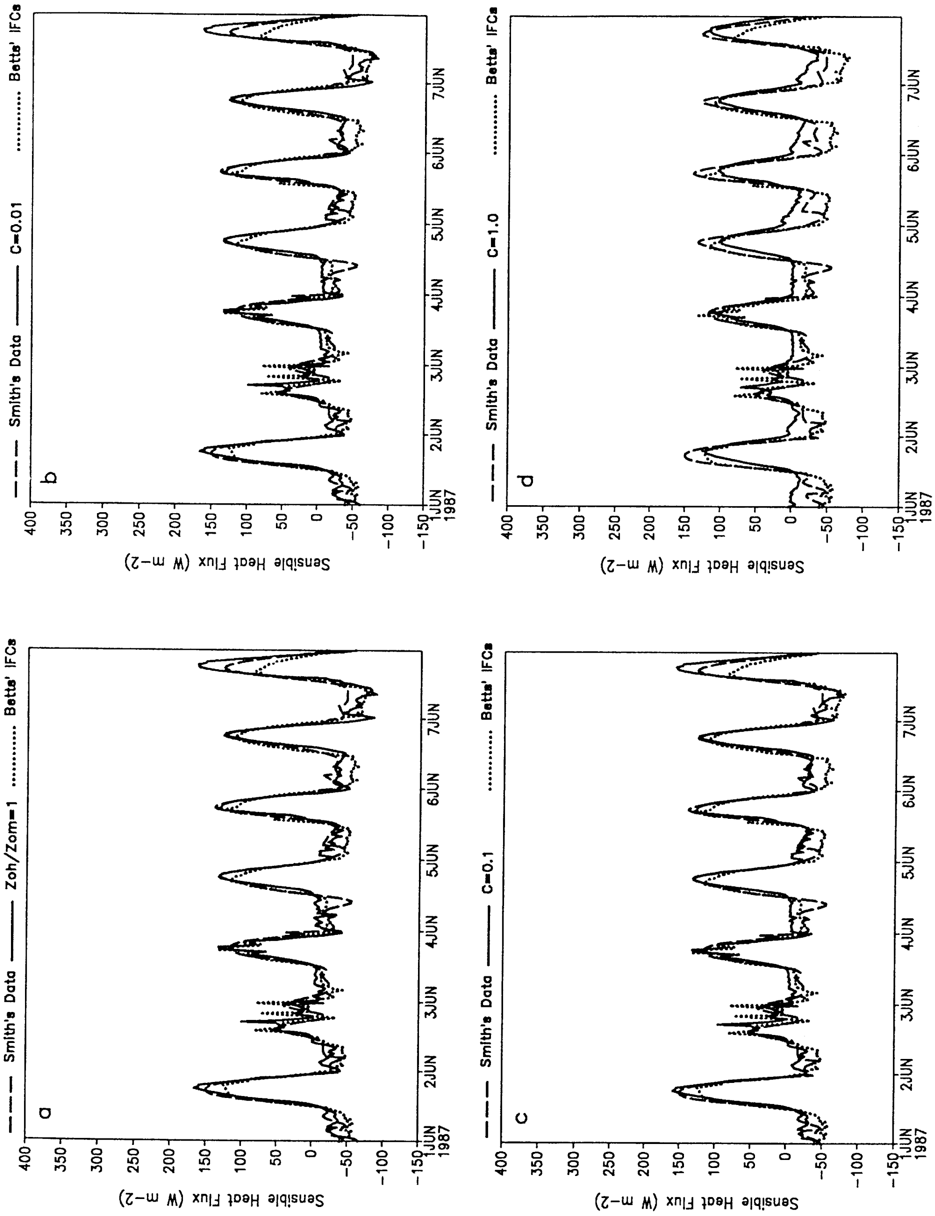
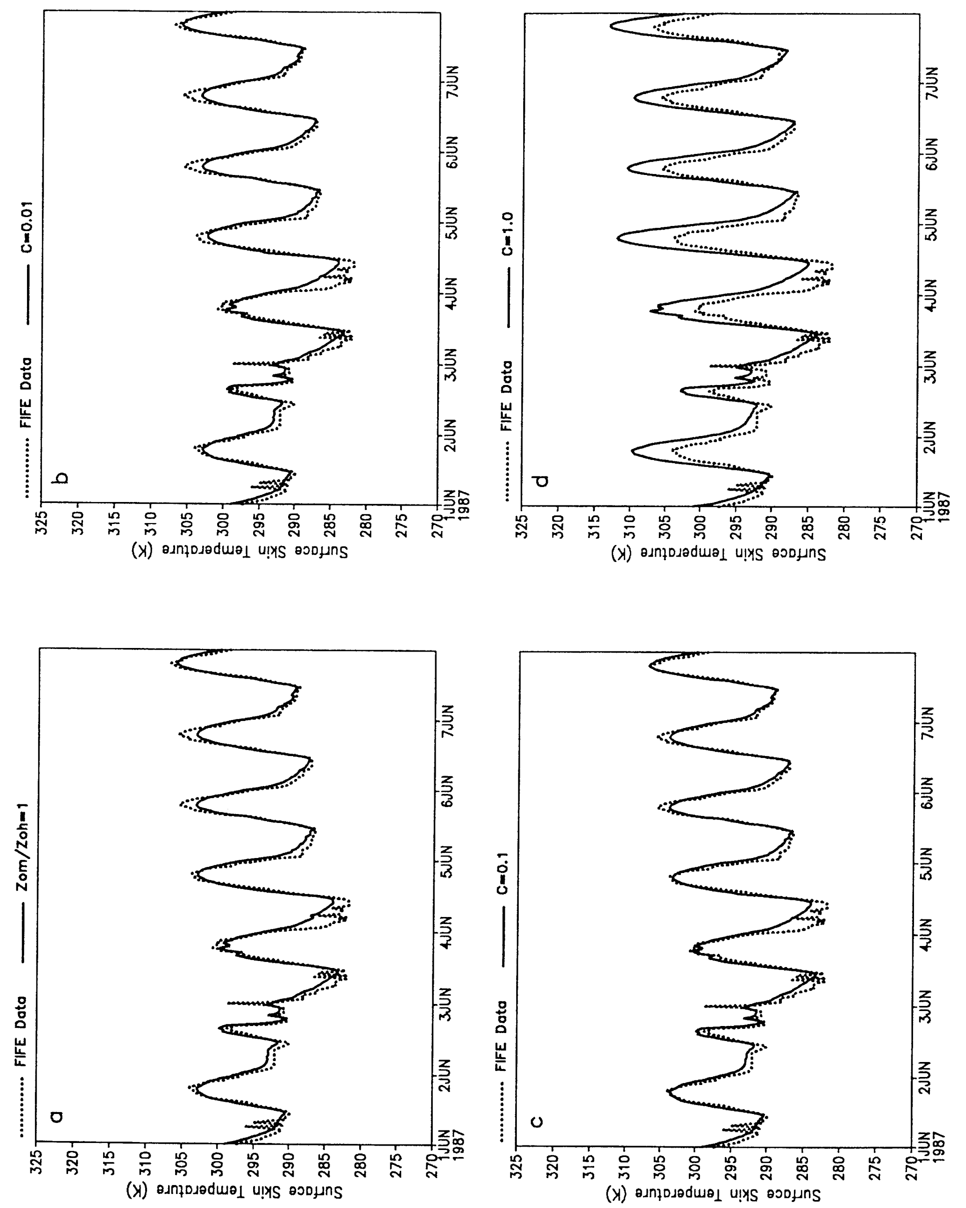

a
Smith's Data
Zoh/Zom=1
Betts' IFCs
b
C=0.01
c
C=0.1
d
C=1.0
Sensible Heat Flux (W m−2)
1JUN 1987
2JUN
3JUN
4JUN
5JUN
6JUN
7JUN

Figure 5c.



a
FIFE Data
Zom/Zoh=1
b
C=0.01
c
C=0.1
d
C=1.0
Surface Skin Temperature (K)
1JUN 1987
2JUN
3JUN
4JUN
5JUN
6JUN
7JUN

Figure 5d.

the results of assuming $z_{0t} = z_{0m}$, consistent with their respectively similar C_h values in Figure 5a. With $C = 1.0$, both the latent heat flux and sensible heat flux are underestimated (Figures 5b,c). This large C resulted in overly high surface skin temperatures as compared to observations (Figure 5d). This arises because, owing to the decrease of z_{0t} and C_h, the land surface is more decoupled from the atmosphere and the surface skin temperature is increasingly the result of balance between surface net radiation and ground heat flux.

For example, compared to the more reasonable simulation using $C = 0.1$, the simulation using $C = 1$ reduced the maximum latent heat fluxes on June 4 by about 100 W m^{-2}, while it increased the skin temperature by about 10 degrees. If the soil is instead relatively dry and the sensible heat flux is dominant(e.g., October in the FIFE case, not shown here), changes in the z_{0m}/z_{0t} ratio do not affect significantly the latent heat flux, but do affect the sensible heat flux by reducing its maximum by about 150 W m^{-2}. In these cases, a large C value can overestimate the maximum skin temperature by more than 10 degrees.

As we see in Figure 5b, a low z_{0m}/z_{0t} ratio overestimated the latent heat flux during early summer in the FIFE case. Interestingly, it underestimated the latent heat flux in late August compared to observations (not shown). This occurred because the soil did not retain enough water during the early summer due to the aforementioned over-evaporation and hence later the surface evaporation is limited by the soil moisture stress during relatively dry periods. From our FIFE and HAPEX-MOBILHY sensitivity tests, we find that using a C value around 0.1 in Equation (1) seems a reasonable choice in that this choice reduces overestimation of surface heat fluxes and at the same time does not produce unduly high surface skin temperature.

These results are consistent with the study of Beljaars and Viterbo (1994). In their 1-D test compared to CABAUW data, they found that using $z_{0m}/z_{0t} = 10$ can considerably improve the problem of winter surface-layer cooling due to the overestimation of evaporation. Other studies, e.g., Braud et al. (1993), showed that a value of the ratio z_{0m}/z_{0t} close to 10 is a good choice for fairly homogeneous areas of bare soil and vegetation. However, investigators suggested that the higher ratios of z_{0m}/z_{0t} should be used for heterogeneous surfaces. Originally, the purpose of using z_{0t} value smaller than z_{0m} was intended to remedy the problem of having to resort to use of the surface skin temperature as the lower boundary condition for integrating Monin-Obukhov similarity theory for the surface layer (e.g., see Sun and Mahrt, 1995). Recently, ideas are emerging to take into account the subgrid-scale variability (e.g., Beljaars and Viterbo, 1994; Hopwood, 1995). For instance, Hopwood (1995) suggested that the ratio of z_{0m}/z_{0t} is about 80 for a typical inhomogeneous and vegetated land surface. Therefore, using a sound z_{0m}/z_{0t} ratio in the atmospheric surface-layer parameterization applied to large-scale and mesoscale numerical models may be an appealing approach to account for the heat fluxes generated by the subgrid-scale variability discussed in various literatures (e.g., Chen and Avissar, 1994). Using a fixed ratio $z_{0m}/z_{0t} = 10$ such as proposed by Beljaars and Viterbo (1994) yields results similar in most cases to using the Zilitinkevich formulation with $C = 0.1$. The latter non-fixed formulation, however, is more preferable in our view, as it can account for regimes of very high Reynolds number and produce very small z_{0t} as supported by observations (e.g., Beljaars and Holtslag, 1991). On the other hand, since the topography strongly modifies the momentum roughness length (e.g., see Hopwood, 1995), it would be interesting to investigate, in the future, the possible topographical effects on z_{0m}/z_{0t}.

Furthermore, in our 3-D coupled simulations with the Eta model, we found that the impact of the z_{0m}/z_{0t} ratio on short-term surface forecasts is not negligible. An example of such impact is illustrated in Figure 6, where the results are obtained from the Eta model (80 km horizontal resolution with 38 vertical levels using the Paulson scheme) with $z_{0m} = z_{0t}$ for one simulation and with $C = 0.1$ in Zilitinkevich's formulation for the other simulation. While in some areas the simulation using Zilitinkevich's formulation with $C = 0.1$ generates higher evaporation, it produces less evaporation in most areas (see Figure 6) and smaller sensible heat flux (not shown here) than the simulation with $z_{0m} = z_{0t}$. The difference in the surface heat flux between these two simulations at 1800Z can be as high as 60 W m^{-2} in some regions.

As a result of its surface heat flux reduction, the $C = 0.1$ case has a higher (slightly lower) surface skin temperature in daytime (nighttime, not shown here). As seen in Figure 6b, in most areas, its surface skin temperature is 2–4 K higher than the simulation with $z_{0m} = z_{0t}$. This difference can reach eight K, however, in dry areas. Interestingly, the 2-metre air temperature at 18Z obtained by Zilitinkevich's formulation is actually 1–2 K lower than the simulation with $z_{0m} = z_{0t}$ (see Figure 6c). Using a smaller z_{0t} generally reduces the mixing in the surface layer and tends to increase the gradient of temperature between the first model layer above the surface and surface skin temperature. Thus the simulation using Zilitinkevich's formulation produces a higher surface skin temperature and at the same time a lower near-surface air temperature than the simulation with $z_{0m} = z_{0t}$.

At NCEP, precipitation verification is an important test criteria to evaluate model development (Mesinger, 1996). Following Mesinger (1996), we use the following equitable threat score T_e and bias score B for evaluation of the precipitation forecast:

$$T_e = \frac{n_{fo} - E(n_{fo})}{n_{fno} + n_{nfo} + n_{fo} - E(n_{fo})}$$

$$E(n_{fo}) = \frac{n_f\, n_o}{N}$$

$$B = \frac{n_f}{n_o}$$

where n_{fo} is the number of grid points at which a precipitation event is forecast and also observed, n_{fno} is the number of points at which the event is forecast but

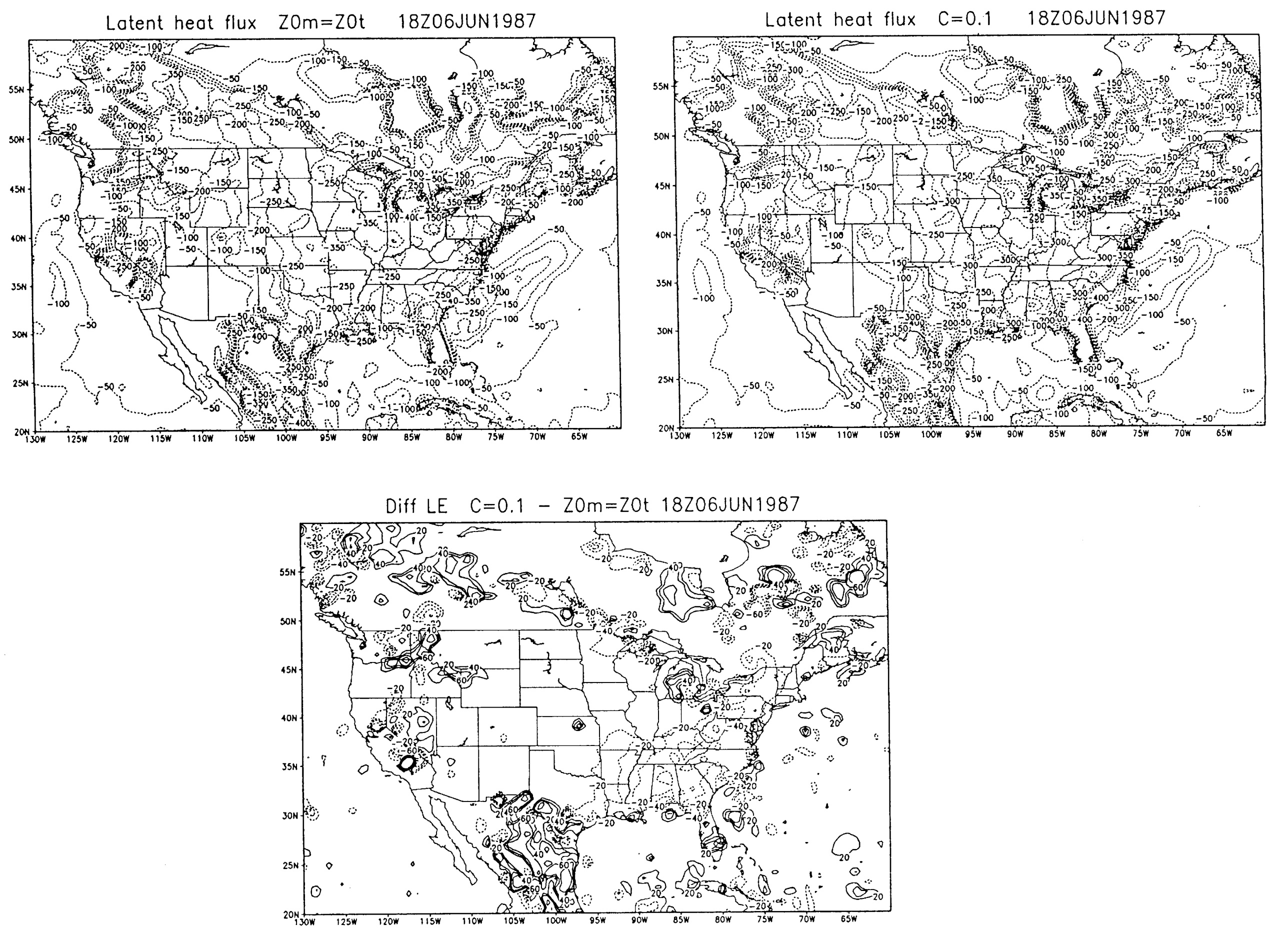


Figure 6. Eta model forecast at 18Z, 6 June 1987 (the 42-h forecast time) obtained with $z_{0m} = z_{0t}$ and $z_{0m} \neq z_{0t}$ (via $C = 0.1$ in Zilitinkevich formulation). (a) Latent heat flux (in W m^{-2}, upward is negative), (b) surface skin temperature, and (c) 2-metre air temperature. The Paulson scheme is used in these simulations.

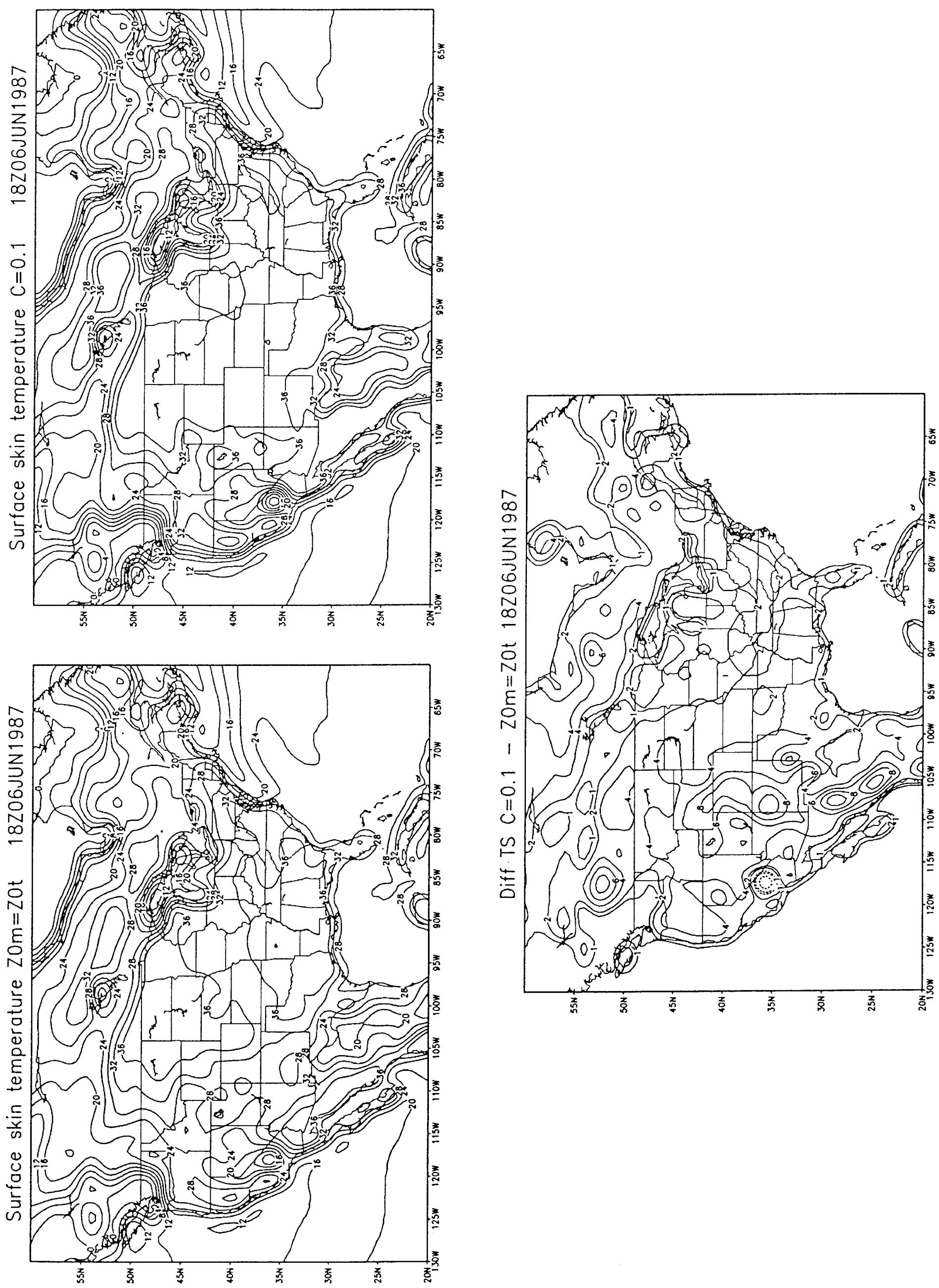

Surface skin temperature C=0.1 18Z06JUN1987
Surface skin temperature Z0m=Z0t 18Z06JUN1987
Diff TS C=0.1 - Z0m=Z0t 18Z06JUN1987

Figure 6b.

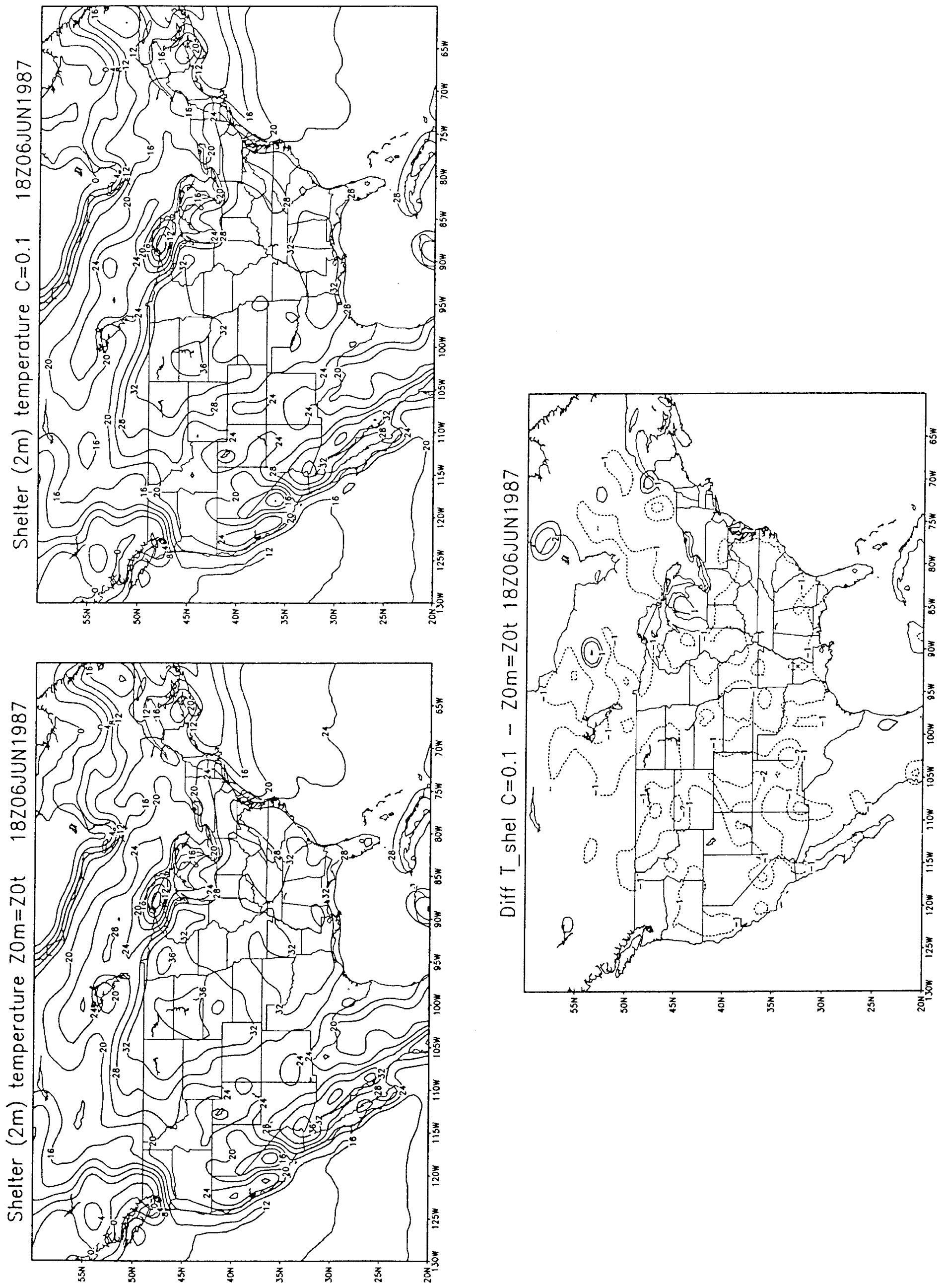

Shelter (2m) temperature C=0.1 18Z06JUN1987
Shelter (2m) temperature Z0m=Z0t 18Z06JUN1987
Diff T_shel C=0.1 - Z0m=Z0t 18Z06JUN1987

Figure 6c.

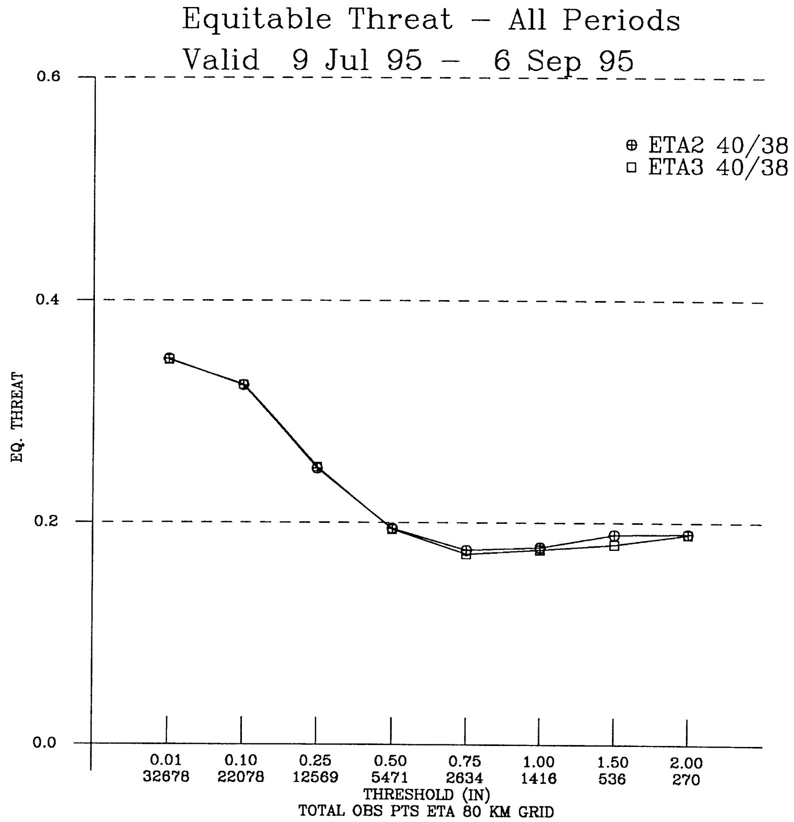


Figure 7. Equitable threat score (a) and bias score (b) for NCEP Eta model forecast from 9 July to 6 September 1995 obtained by ETA2 using the modified Louis scheme with $z_{0m}/z_{0t} = 10$ and ETA3 using the Paulson scheme with $z_{0m} = z_{0t}$. 'All periods' are 24-hr sums from 24, 36, and 48-hr forecasts.

not observed, n_{nfo} is the number of points at which the event not forecast but is observed, and n_f and n_o are the total number of points at which precipitation is forecast and observed, respectively. Also N is the total number of points at which observations (and forecasts) are made. In descriptive terms, $E(n_{fo})$ is the expected number of 'hits' in a random forecast and T_e represents the fractional number of correctly forecast points over and above random.

Figure 7 shows nearly three months (from 9 July to 6 September 1995) of Eta model precipitation forecast scores compared with observations. These Eta model forecasts are made with a 40-km horizontal resolution and 38 vertical levels. The so-called 'ETA2' version used the modified Louis scheme and assumed $z_{0m}/z_{0t} = 10$, while the 'ETA3' version used the Paulson scheme and $z_{0m} = z_{0t}$. According to our sensitivity tests (not shown here), the Eta model using the modified Louis scheme produces precipitation scores that are not noticeably different from those of the Eta model using the Paulson scheme, provided the same z_{0m}/z_{0t} ratio is

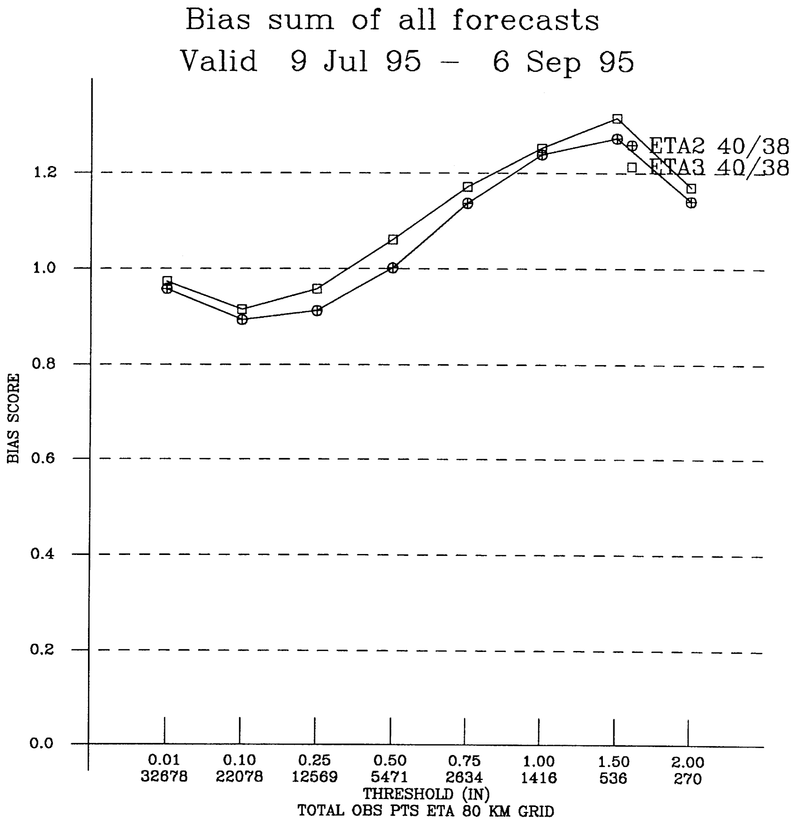


Figure 7b.

used. Thus, the difference between ETA2 and ETA3 could be interpreted as the result of using a different z_{0m}/z_{0t} ratio.

For ETA2 and ETA3, the equitable threat scores T_e are very close at almost each category of precipitation, except for the 1.5-inch category where the ETA2 has a slightly higher score than the ETA3. However, the precipitation bias score, indicating the overall precipitation amount predicted over the continental US compared to observations, displays a systematic difference between the two models: ETA3 has higher precipitation bias than ETA2. The ETA2 used a higher z_{0m}/z_{0t} ratio than ETA3 and, as we previously discussed, reduced the latent heat and sensible heat fluxes. This decrease in surface heat flux is likely the reason for the lower precipitation bias in ETA2.

4. Conclusions

We tested three atmospheric surface-layer parameterization schemes (Mellor-Yamada level 2, Paulson, and modified Louis) both in a 1-D uncoupled mode against long-term observations and in a coupled 3-D mode with the NCEP Eta model. These three parameterization schemes are all based on the Monin-Obukhov similarity theory. Hence, the difference in these three schemes does not, in general, result in significant difference in surface fluxes, skin temperature, and precipitation, provided the same treatment of roughness length for heat (z_{0t}) is used.

Rather than the choice among these three surface-layer schemes, the model is more sensitive to the treatment of roughness length for heat (moisture) in any of the three schemes. The need for separating the roughness length for heat (moisture) z_{0t} from that for momentum z_{0m} arises from the need to use a radiative surface skin temperature.

In the present study, we tested two approaches to specify the z_{0m}/z_{0t} ratio. Our 1-D column model sensitivity tests suggested that the type of equation proposed by Zilitinkevich (1995) for calculating the ratio of z_{0m}/z_{0t} can improve the surface heat flux and skin temperature simulations. The long-term 3-D test with the NCEP mesoscale Eta forecast model indicated that using $C = 0.1$ in this formulation can reduce forecast precipitation bias.

In the light of these 1-D uncoupled and 3-D coupled simulations, we selected the Paulson scheme as the new NCEP Eta model surface-layer parameterization. We implemented the Zilitinkevich formulation with $C = 0.1$ in the NCEP mesoscale Eta model to specify the z_{0m}/z_{0t} ratio. Parameterizing z_{0m}/z_{0t} as a property of flow in atmospheric numerical models seems a practical and appealing approach to account, to some extent, for subgrid-scale variability.

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Appendix A. Three Surface-Layer Parameterizations

Surface fluxes (stress τ, sensible heat flux H, and latent heat flux E) are related to mean properties of the flow through the use of drag and bulk transfer relations (e.g., Garratt, 1993). These bulk relations are written as

$$\tau = \rho C_d |u_a|^2 \tag{A1}$$

$$H = \rho c_p C_h |u_a| (\theta_s - \theta_a) \tag{A2}$$

$$E = \rho C_e |u_a| (q_s - q_a) \tag{A3}$$

where ρ is the air density, and c_p the air heat capacity. The quantities $|u_a|$ (wind speed), θ_a (air potential temperature), and q_a (air specific humidity) are evaluated at the lowest model level, while θ_s and q_s are surface values, with θ_s typically obtained from the surface energy balance. $C_d,\ C_h,$ and C_e are the surface exchange (or bulk transfer) coefficients for momentum, heat, and water vapor, respectively. In this paper, we, like most numerical modelers, assume that C_h is identical to C_e and will focus on the discussion of three C_h formulations tested in this study.

A.1. Formulations based on the Obukhov length

As cited in the main text, two of the three C_h formulations, namely those of Paulson and Mellor-Yamada Level 2, directly use the similarity theory based stability functions, Ψ_m for momentum and Ψ_h for heat. When distinct roughness length values are retained for momentum and heat, then the C_h formulation in terms of Ψ_m and Ψ_h can be expressed as (e.g., Brutsaert, 1982; Garratt, 1992):

$$C_h = \frac{k^2/R}{[\ln(\frac{z}{z_{0m}}) - \Psi_m(\frac{z}{L}) + \Psi_m(\frac{z_{0m}}{L})][\ln(\frac{z}{z_{0t}}) - \Psi_h(\frac{z}{L}) + \Psi_h(\frac{z_{0t}}{L})]} \tag{A4}$$

where z_{0m} and z_{0t} are the roughness lengths for momentum and heat respectively. k is the von Kármán constant, z is the reference height within the surface layer, and R, estimated at 1.0, is the ratio of the exchange coefficients for momentum and heat in the neutral limit. L is the Obukhov length expressed as

$$L = -\frac{\overline{\Theta_v} u_*^3}{kg\overline{w'\Theta_v'}}$$

where Θ_v is the virtual potential temperature, u_* is the friction velocity, and g is the acceleration of gravity.

A.1.1. *Paulson Formulation for* Ψ_m *and* Ψ_h

Paulson (1970) integrated a set of analytical expressions to specify non-dimensionalized wind speed and potential temperature gradients as function of z/L. Following Sun and Mahrt (1995), these stability functions are

$$\Psi_m = \begin{cases} -5\zeta & 0 < \zeta < 1 \\ 2\ln(\dfrac{1+x}{2}) + \ln(\dfrac{1+x^2}{2}) - 2\tan^{-1}(x) + \dfrac{\pi}{2} & -5 < \zeta < 0 \end{cases} \tag{A5}$$

$$\Psi_h = \begin{cases} -5\zeta & 0 < \zeta < 1 \\ 2\ln(\dfrac{1+x^2}{2}) & -5 < \zeta < 0 \end{cases} \tag{A6}$$

where $\zeta = z/L$ and $x = (1 - 16\zeta)^{1/4}$.

A.1.2. *Mellor-Yamada Level-2 Formulation for* Ψ_m *and* Ψ_h

Łobocki(1993) developed a procedure for the derivation of surface-layer relationships from simplified second-order closure models and applied it with the Mellor-Yamada level-2 model to obtain the following stability functions:

$$\Psi_m = \begin{cases} \dfrac{\zeta}{R_{FC}} - 2.076\left(1 - \dfrac{1}{\zeta+1}\right) & 0 \le \zeta < 1 \\ -0.96\ln(1 - 4.5\zeta) & -5 < \zeta < 0 \end{cases} \tag{A7}$$

$$\Psi_h = \begin{cases} \dfrac{\zeta R_{ic}}{R_{FC}^2 \phi_T(0)} - 2.076\left(1 - \exp(-1.2\zeta)\right) & 0 \le \zeta < 1 \\ -0.96\ln(1 - 4.5\zeta) & -5 < \zeta < 0 \end{cases} \tag{A8}$$

where R_{ic} ($R_{ic} = 0.183$) is the critical gradient Richardson number, R_{FC} ($R_{FC} = 0.191$) is the critical flux Richardson number, and $\phi_T(0) = 0.8$ is the dimensionless velocity gradient for neutral conditions.

A.2. Formulation Based on Bulk Richardson Number

To avoid costly iterations during model integration, Louis (1979) introduced a precomputed approximation of the Monin-Obukhov flux-gradient relationships. This approach provides explicit formulae for flux calculations and is applied in many atmospheric mesoscale models (see Pielke, 1984; Mascart et al., 1995). In this study, C_h is parameterized following the modified Louis approach by Mahrt (1987) for the stable case and Holtslag and Beljaars (1989) for the unstable case, wherein

$$C_h = \frac{k^2}{R}\frac{F2(z, z_{0m}, z_{0t}, R_{ib})}{\ln(\dfrac{z}{z_{0m}})\ln(\dfrac{z}{z_{0t}})} \tag{A9}$$

The bulk Richardson number for the surface (R_{ib}) is defined

$$R_{ib} = \frac{g\ z\ (\theta_{av} - \theta_{sv})}{\theta_{av}\ |u_a|^2} \tag{A10}$$

where θ_{av} and θ_{sv} are the virtual potential temperature at the reference height (usually the first model level) and at the surface respectively. We use the specific humidity at the first model level as a first-order approximation to calculate θ_{sv}. The function $F2$ in Equation (A9) is

$$F2 = \begin{cases} e^{-aR_{ib}} & \text{stable} \\ 1 - \dfrac{15R_{ib}}{1 + A(R_{ib}, z, z_{0m}, z_{0t})} & \text{unstable} \end{cases} \tag{A11}$$

where,

$$A = \frac{70.5k^2\sqrt{-R_{ib}\dfrac{z}{z_{0t}}}}{\ln(\dfrac{z}{z_{0m}})\ \ln(\dfrac{z}{z_{0t}})} \tag{A12}$$

and a is a constant currently equal to 1.0 in the model. According to Mahrt (1987), the introduction of the term $e^{(-aR_{ib})}$ for the stable case could include the most important qualitative aspects of subgrid-scale averaging.

Appendix B. New Land-Surface Model in the NCEP Mesoscale Eta model

This land-surface scheme is based on the coupling of the diurnally-dependent Penman potential evaporation approach of Mahrt and Ek (1984), the multi-layer soil model of Mahrt and Pan (1984), and the primitive canopy model of Pan and Mahrt (1987). It is then extended by Chen et al. (1996) to include the modestly complex canopy resistance approach of Noilhan and Planton (1989) and Jacquemin and Noilhan (1990) (hereafter NP89 and JN90, respectively). It has one canopy layer, and the following prognostic variables: soil moisture and temperature in the soil layers; water stored on the canopy; and snow stored on the ground. In the NCEP Eta model, this land-surface model utilizes two soil layers: a thin top layer of 10 cm and a deep root zone of 190 cm. This soil layer configuration is purposely selected in order to use the NCEP Global Data Assimilation System (GDAS) (wherein a similar land-surface scheme with the same soil layer structure is used) soil moisture and temperature to initialize the NCEP mesoscale Eta model.

B.1. Model thermodynamics

The surface skin temperature is determined following Mahrt and Ek (1984) by applying a single linearized surface energy balance equation representing the combined ground/vegetation surface. Accompanying this, the ground heat flux is controlled by the usual diffusion equation for soil temperature (T):

$$C(\Theta)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(K_t(\Theta)\frac{\partial T}{\partial z}\right) \tag{B1}$$

where the volumetric heat capacity C and the thermal conductivity K_t are formulated as functions of volumetric soil water content Θ (fraction of unit soil volume occupied by water). The layer-integrated form of Equation (B1) for the i-th soil layer is:

$$\Delta z_i C_i \frac{\partial T_i}{\partial t} = \left(K_t\frac{\partial T}{\partial z}\right)_{z_{i+1}} - \left(K_t\frac{\partial T}{\partial z}\right)_{z_i}. \tag{B2}$$

The prediction of T_i is performed using the Crank-Nicholson scheme.

B.2. MODEL HYDROLOGY

In the hydrology model, the prognostic equation for the volumetric soil moisture content (Θ) is:

$$\frac{\partial \Theta}{\partial t} = \frac{\partial}{\partial z}\left(D\frac{\partial \Theta}{\partial z}\right) + \frac{\partial K}{\partial z} + F_\Theta \tag{B3}$$

where both the soil water diffusivity D and hydraulic conductivity K are functions of Θ, and F_Θ represents sources and sinks (i.e., precipitation and evaporation) for soil water.

Integrating Equation (B3) over J ($J = 4$ in the uncoupled land-surface model and $J = 2$ in the coupled Eta model) soil layers and expanding F_Θ, we obtain:

$$d_{z_1}\frac{\partial \Theta_1}{\partial t} = -D\left(\frac{\partial \Theta}{\partial z}\right)_{z_1} - K_{z_1} + P_d - R - E_{dir} - E_t \tag{B4}$$

$$d_{z_i}\frac{\partial \Theta_i}{\partial t} = D\left(\frac{\partial \Theta}{\partial z}\right)_{z_{i-1}} - D\left(\frac{\partial \Theta}{\partial z}\right)_{z_i} + K_{z_{i-1}} - K_{z_i} - E_{t_i}; \quad i = 2, J-1 \tag{B5}$$

$$d_{z_J}\frac{\partial \Theta_J}{\partial t} = D\left(\frac{\partial \Theta}{\partial z}\right)_{z_{J-1}} + K_{z_{J-1}} - K_{z_J} \tag{B6}$$

where d_{z_i} is the i-th soil layer thickness. P_d is the precipitation not intercepted by the canopy. R is the surface runoff and specified by the Simple Water Balance model surface runoff formulation, which is a hydrological approach that considers the subgrid-scale variability in precipitation and soil moisture (see Chen et al., 1996, Schaake et al., 1996). E_{t_i} is the canopy transpiration taken by the canopy root in the i-th layer within the root zone layers (the root zone has three layers in the uncoupled land-surface model and two layers in the coupled Eta model, respectively). K_{z_J} is the moisture loss due to 'gravitational' percolation through the bottom of the J-th layer, also named sub-surface runoff or drainage.

The total evaporation E is the sum of the direct evaporation from the top shallow soil layer, E_{dir}, evaporation of precipitation intercepted by the canopy, E_c, and transpiration via canopy and roots, E_t, i.e., $E = E_{dir} + E_c + E_t$.

The direct evaporation from the ground surface is determined by

$$E_{dir} = (1 - \sigma_f)\ \ MIN\left(\left[-D\left(\frac{\partial\Theta}{\partial z}\right)_{z_1} - K_{z_1}\right], E_p\right) \tag{B7}$$

where E_p is the potential evaporation and σ_f is the green vegetation fraction. The potential evaporation E_p is calculated by a Penman-based energy balance approach including a stability-dependent aerodynamic resistance (Mahrt and Ek, 1984).

The wet canopy evaporation is determined by

$$E_c = \sigma_f E_p \left(\frac{W_c}{S}\right)^n \tag{B8}$$

where W_c is the intercepted canopy water content, and S is the maximum allowed W_c capacity, chosen here to be 0.5 mm. $n = 0.5$. This is formulated similarly to NP89 and JN90. The intercepted canopy water budget is governed by

$$\frac{\partial W_c}{\partial t} = \sigma_f P - D - E_c \tag{B9}$$

wherein P is the input total precipitation. If W_c exceeds S, the excess precipitation or drip D reaches the ground (note that $P_d = (1 - \sigma_f)P + D$ in Equation (B4)). The canopy evapotranspiration is determined by

$$E_t = \sigma_f E_p B_c \left[1 - \left(\frac{W_c}{S}\right)^n\right] \tag{B10}$$

where B_c is a function of canopy resistance and is formulated as:

$$B_c = \frac{1 + \dfrac{\Delta}{R_r}}{1 + R_c C_h + \dfrac{\Delta}{R_r}}. \tag{B11}$$

Here C_h is the surface exchange coefficient for heat and moisture, Δ depends on the slope of the saturation specific humidity curve, R_r is a function of surface air temperature, surface pressure, and C_h. Details on C_h, R_r, and Δ are provided by Ek and Mahrt (1991). The canopy resistance R_c is calculated here following the formulation of JN90.

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