Large-eddy simulation over heterogeneous terrain with remotely sensed land surface conditions

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Abstract. A framework is developed to explore the coupling between the land and the atmospheric boundary layer using three-dimensional turbulence simulation over remotely sensed land surface images. The coupled set of equations is integrated with boundary conditions from day 221 of the Monsoon ’90 experiment, and analysis is conducted to quantify the transmission of surface heterogeneity information into the atmospheric boundary layer (ABL). The large-eddy simulation (LES) model incorporates radiant energy availability; spatial fields of remotely observed surface cover, temperature, and moisture; and the ability to account for the separate contributions of soil and vegetation (i.e., two sources) to the mass and energy exchanges. This effort reflects a merging of active lines of research: the use of remotely sensed land surface properties to study water and energy fluxes and the use of LES to study impacts of surface variability on ABL processes. Analysis of the results reveals (1) that the combination of remotely sensed data and LES (in the absence of free parameters) yields regionally averaged land surface fluxes and ranges of spatial variability in the fluxes that compare well to similar measures from a network of flux measuring stations, (2) that the correlation between time-averaged surface and air temperatures is dependent on the length scale of the surface features, (3) that the horizontal standard deviation of mean air temperature decreased logarithmically with height in the atmospheric surface layer; and (4) that the mean air temperature contains spatial variability induced preferentially from variations in surface temperature occurring at scales \( \approx 500 \text{–} 1000 \text{ m} \). Hence the feedback strength between the land and the atmosphere is shown to be scale-dependent for the range of length scales (i.e., \( \approx O(10 \text{ km}) \)) studied here.

1. Introduction

Land surface fluxes of energy and mass develop over heterogeneous landscapes in response to the relative differences between surface properties and the overlying air properties. The air properties are influenced to some extent by the underlying land surface heterogeneity through land-atmosphere feedbacks. Therefore the accurate prediction of spatial fields of water and energy fluxes over heterogeneous land surfaces is predicated on an understanding of these feedback effects.

Additional progress is required in our understanding of how regional-scale surface fluxes develop over heterogeneous land before mesoscale flux predictions will be considered generally robust [e.g., Andre et al., 1990; Rodriguez-Camino and Avissar, 1999]. A review of the important issues surrounding the dynamic response of the atmosphere to the land surface is given by Avissar [1995].

The large-eddy simulation (LES) technique has emerged as a useful tool for exploring impacts of land surface heterogeneity on the atmospheric boundary layer (ABL). Most studies addressing land surface heterogeneity with LES have described surface boundary conditions as predefined fluxes with artificial variability [e.g., Hadfield et al., 1991, 1992; Avissar et al., 1998; Avissar and Schmidt, 1998; Cai, 1999] or with spatial variability defined to match the flux fields estimated from an experiment over a particular site [e.g., Hechtel et al., 1990; Eastman et al., 1998]. The analyses included in these studies have addressed potential impacts of heterogeneous land surface fluxes on the spatially averaged vertical profiles of velocity and scalar statistics in the ABL. The published results tended to observe no significant impacts to the regionally averaged profiles. For example, Hechtel et al. [1990] found no significant impacts to mean ABL profiles over the heterogeneous terrain of a field site near Chickasha, Oklahoma. Hadfield et al. [1991, 1992] found that the ABL impacts created over patterned surface heat flux fields were only significant in the absence of a strong mean wind. Avissar et al. [1998] explored the ABL development over the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) site by contrasting a spatially homogeneous (mean) surface heat flux boundary condition to a case with random variability and noted only minor changes to the mean profiles. However, they con-
jectured that a more “organized contrast of heat fluxes may have a significant impact on the [convective boundary layer].” Their surface flux field was purely random and uncorrelated in space. Left unanswered is the question of how the surface heterogeneity affects ABL heterogeneity, as opposed to affecting mean profiles, and how the surface and air properties, in turn, affect the flux fields that develop over a region with heterogeneous surface properties.

A few studies have specified heterogeneous surface states and allowed the fluxes to develop internally with LES [e.g., Albertson, 1996; Albertson and Parlange, 1999a, 1999b; Holson et al., 1999]. In a study of the neutral ABL, Albertson and Parlange [1999a] demonstrated that asymmetries in the flow field’s response to rough-to-smooth transitions, as compared to smooth-to-rough transitions, result in net increases in regional-scale shear stress for decreasing length scale of surface roughness patches. In an LES-based study of blending layers in the convective boundary layer (CBL), Albertson and Parlange [1999b] revealed that mean scalar concentrations blend at lower heights than do the fluxes of the scalars over heterogeneous terrain. They also showed that temperature inhomogeneities persist higher into the ABL than humidity inhomogeneities, because of the active role of temperature on vertical motion. However, these applications employed simple geometric surface features (e.g., rectangles) as proxies for real spatial heterogeneity. To our knowledge, there has yet to be an LES study that both used measured state variables as boundary conditions over a heterogeneous land surface and computed the surface fluxes dynamically, with inclusion of feedback effects on the air properties.

The objectives of this paper are (1) to develop a framework to fully explore the coupled land-ABL system with LES over remotely sensed land surface images and (2) to use the framework to quantify the transmission of surface heterogeneity information into the ABL for a test day of the Monsoon ’90 experiment. Hence we extend our existing LES model to incorporate surface energy availability; spatial fields of remotely observed surface cover, temperature, and moisture; and the ability to account for the separate contributions of soil and vegetation to the mass and energy exchanges. We use this new system to study conditions from day 221 of the Monsoon ’90 field campaign.

In effect, this paper reflects a merging of active lines of research using remotely sensed land surface properties to study water and energy fluxes and the use of LES to study impacts of surface variability on atmospheric boundary layer (ABL) processes. This results in a framework that allows detailed isolation and study of the land-atmosphere feedback pathways between real land surface features and the turbulent ABL.

2. Methods

This effort combines remotely sensed land surface conditions with a large-eddy simulation code and uses this system to explore the coupling of land surface and atmospheric boundary layer processes in a semiarid region. The data are taken from the Monsoon ’90 experiment [Kustas and Goodrich, 1994]. The boundary layer dynamics aspects of the LES code were described previously [Albertson, 1996; Albertson and Parlange, 1999a, 1999b]. In this paper the LES code is modified to include a “two-source” model of surface exchange, where there is an explicit treatment of the separate contributions from bare soil and vegetation to the surface fluxes of water and energy within each surface grid cell, an important feature for semiarid region flux modeling. The portion of the two-source surface model algorithm using radiometric surface temperature observations is based on Norman et al. [1995], and that portion using microwave-derived surface soil moisture is based on Kustas et al. [1998]. The equations are modified to tailor the formulations to the transient nature of the LES framework. In this section we describe the LES concept in brief form, the two-source land surface model, the experimental conditions, the remotely sensed data set, and the protocol for integrating the surface data and the LES to explore the land-atmosphere feedback pathways.

2.1. LES of Boundary Layer Turbulence

The dynamics and transport in the turbulent atmospheric boundary layer are represented by the Navier-Stokes set of equations and related equations describing mass and energy conservation under thermal stratification [e.g., Brutsaert, 1982; Stull, 1988]. Turbulent dynamics are inherently three-dimensional (consider the role of vortex stretching in the spectral cascade), and the scales of motion in the ABL range from \( O(10^{-3} \text{ m}) \) to \( O(10^{3} \text{ m}) \). Current computational capabilities are inadequate to integrate these equations numerically in space and time on a mesh capable of resolving this full range of scales [McComb, 1990]. Hence applications require a reduction in the degrees of freedom being simulated explicitly and the introduction of some model to account for the effects of the remaining degrees of freedom [Lesieur and Metais, 1996].

For several decades, turbulence modeling was conducted primarily using the “Reynolds-averaged” framework, where the equations of motion are averaged, such that the mean field is derived subject to a closure model that approximates the complete set of turbulence effects [Tennekes and Lumley, 1972; Katul and Albertson, 1998, 1999]. The LES technique is fundamentally different. With LES the governing equations are spatially filtered to the scale of a numerical mesh, rather than fully averaged. This filter separates the turbulent field into “resolved” and “subfilter” components

\[
V = \tilde{V} + V',
\]

where \( \tilde{V} \) represents some turbulent field (e.g., a velocity component or scalar concentration) on \( (x, y, z, t) \), \( \tilde{V} \) is its filtered form, still varying on \( (x, y, z, t) \), and \( V' \) is the subfilter or subgrid scale (SGS) component, defined as the difference between \( \tilde{V} \) and \( \tilde{V} \). (See Albertson and Parlange [1999a] for a more complete description of the filtering concept.) Note that here the filtered fields (e.g., \( \tilde{V}(x, y, z, t) \) contain all the large-scale turbulent dynamics, which depend uniquely on the particular boundary conditions, while the subfilter portion contains the small-scale turbulent motion which is more universal and rendered isotropic through the turbulent cascade process. Hence the study of interaction between heterogeneous land surfaces and the ABL is well matched to the LES technique. The filtered Navier-Stokes equations can simulate the effects of the boundary conditions on the large-scale turbulence, which is responsible for the bulk of the water and energy transport, and a subgrid model can approximate the effects of the unresolved eddies on the resolved scales.

The filtered equations that simulate the ABL turbulence in the LES are

\[
\partial_t \tilde{\mu}_i + \tilde{u}_i (\partial \tilde{\mu}_i - \partial \tilde{\mu}) = -\partial \tilde{p} + F_i \delta_{i1} - \partial_j \tau_{ij} + \tilde{B} \delta_{ij},
\]
\[
\frac{\partial \hat{\theta}}{\partial t} + \hat{u}_i \frac{\partial \hat{\theta}}{\partial x_i} = -\frac{\partial \tau_{ij}^s}{\partial x_j},
\]
\[
\frac{\partial \hat{q}}{\partial t} + \hat{u}_i \frac{\partial \hat{q}}{\partial x_i} = -\frac{\partial \tau_{ij}^q}{\partial x_j},
\]  

where \( \hat{u}_i \) is the resolved velocity component in the \( x_i \) direction (\( i = 1 \) (longitudinal), 2 (transverse), 3 (vertical)), \( \hat{\theta} \) is a composite pressure term (including the trace of the subgrid stress tensor and \( x_i \) direction gradient of the turbulent kinetic energy), \( \tau_{ij}^s \) is the mean streamwise pressure gradient, \( \tau_{ij}^q \) is the subgrid stress tensor, \( \hat{\theta} \) is the buoyancy parameter, \( \hat{\theta}_i \) is the time derivative, \( \delta_{ij} \) is a spatial derivative with respect to direction \( x_i \), \( \delta_{ij} \) is the Kronecker delta, \( \hat{\theta} \) is the filtered potential temperature field, \( \hat{q} \) is the filtered humidity field, \( \tau_{ij}^s \) and \( \tau_{ij}^q \) are the subgrid fluxes of temperature and water vapor, respectively, in the \( x_j \) direction, and summation is implied on repeated subscripts.

The turbulent pressure field is computed from a Poisson equation, derived from application of the continuity equation to the divergence of the momentum equation, thus ensuring that the velocity field remains divergence free. The effect of the subgrid stresses on the resolved eddy motion is accounted for by a local, subgrid eddy viscosity acting on the resolved instantaneous gradients [Smagorinsky, 1963], such that the eddy viscosity varies in space and time with the strain rates. The subgrid scalar fluxes (\( \tau_{ij}^s \) and \( \tau_{ij}^q \)) are approximated from a local eddy-diffusivity model using the eddy viscosity modified by the Schmidt number (\( Sc \)). Characteristics of the simulated turbulence in the interior of the flow are relatively insensitive to the SGS model, but flow characteristics near the surface may, in fact, depend on the SGS model formulation. Recent and current efforts suggest that these SGS models may soon become even more robust [Porte-Agel et al., 1998].

The dynamic temperature and humidity fields induce density-driven vertical accelerations in (2) via

\[
\beta(x, y, z, t) \left\{ \left( \frac{\hat{\theta}_i(x, y, z, t) - \langle \hat{\theta}(z, t) \rangle}{\langle \hat{\theta}(z, t) \rangle} \right) \right\},
\]

where \( \hat{\theta}_i \) is the virtual potential temperature (with temperature in K, water vapor concentration in g kg\(^{-1}\)), and the constant (0.61) clears units, \( \hat{\theta} \) is the acceleration due to gravity, and the \( \langle \hat{\theta}(z, t) \rangle \) operator defines a horizontal average. The solution of (2)-(4) is subject to the specification of appropriate boundary conditions. The horizontal directions have periodic boundary conditions, which is a necessary approach in turbulence simulation because it is critical to have fully developed turbulence on both the inflow and exit faces of the domain. When simulating flow over heterogeneous surfaces, such as here, this condition is equivalent to having a regional landscape comprised of repeated versions of the field underlying the model domain.

The top boundary of the computational domain is positioned well above the top of the boundary layer. The top condition is one of vanishing vertical gradients of all primitive variables and no vertical flow across the boundary (i.e., \( u_3 = 0 \)). A numerical sponge is added to the vertical momentum equation near the upper boundary to dampen vertical motions as they approach the no-flow upper boundary [Nieuwstadt et al., 1991; Albertson and Parlange, 1999b]. The effects of this dampening are limited to a small layer located well above the top of the ABL and hence isolated from the important boundary layer dynamics.

The bottom boundary is, in general, far more critical to the structure of the turbulence than the sides and top boundary. In this particular case of studying the impacts of land heterogeneity the treatment of the bottom boundary conditions is of primary importance. Typically, LES codes include surface fluxes that are specified a priori [e.g., Avisser and Schmidt, 1998]. Such an approach instills a one-way form of land-atmosphere interactions, since the changes to air properties engendered by the water and heat exchange are not fed back on the flux processes (as the fluxes are not dependent on ABL processes). We present here a model for capturing both real heterogeneous surface conditions and the important feedback pathways between the land and the atmosphere by way of their mutual controls on the exchange rates.

### 2.2. Two-Source Surface Model

The space-time distributions of water, heat, and momentum exchange rates between the land and the ABL are computed by a combination of energy and aerodynamic considerations and by accounting for the separate contributions from bare soil and vegetation within each “model grid cell” (i.e., a two-source model along the surface). The spatial fields of land surface properties are taken from remote sensing platforms. Visible imagery is used to define the primary variable of vegetation cover, which is combined with field measurements to define spatial maps of albedo, leaf characteristics, zero plane displacement, and roughness lengths. Multispectral imagery is used to define the fractional vegetation cover of each pixel via a semiempirical relationship between normalized difference vegetation index (NDVI) and the fractional cover [Choudhury et al., 1994; Carlson and Ripley, 1997]. Thermal imagery provides the radiometric surface temperatures (\( T_s \)), and passive microwave remotely sensed images are used to map soil moisture status (\( w \)). The incoming shortwave radiative flux density is assumed constant over the region and is specified to the model. Details of the data set for the particular simulation of Monsoon ’90 are described below.

A relatively simple two-source model which addresses the main factors of importance here has been used to generate surface flux maps using thermal infrared data [Kustas and Humes, 1996; Schmugge et al., 1998] and microwave-derived surface soil moisture [Kustas et al., 1998]. The model was designed to use input data primarily from satellite observations; hence several simplifying assumptions about energy partitioning between the soil and vegetation reduce both computational time and input data required to characterize surface properties. The inputs include an estimate of fractional vegetative cover, canopy height, leaf width, surface temperature, solar radiation, wind speed, and air temperature. The two-source approach has been shown to be more robust than a related single-source approach for these types of semiarid land cover [Zhan et al., 1996].

Topography is not explicitly included in the LES. The effect of topography on surface fluxes is predominately an indirect control via influences on soil moisture [Peters-Lidard et al., 1997]. This effect will be implicitly included here, as measured soil moisture fields are prescribed as boundary conditions.

An algorithmic description of the version of the two-source surface flux model used here in the LES is included in Appendix A.
2.3. Data Set

The data set used by the model was collected during the Monsoon '90 field experiment conducted in the Walnut Gulch Experimental Watershed (31.5°N 110°W) maintained by the Southwest Watershed Research Center in Tucson, Arizona [Kustas and Goodrich, 1994]. The primary campaign of the experiment was conducted during a 2-week period in the summer rainy season, a time at which the vegetation was very active. This semiarid rangeland environment supports desert steppe and grassland communities, both of which are contained in the watershed. The vegetation cover is highly variable ranging from <10% in some areas to nearly full cover in the riparian areas. Average elevation of the watershed is nominally 1500 m. The watershed is mildly hilly with ridge to valley heights of the order of 10 m for the west and central portion of the watershed and reaching 15–20 m for the eastern half. Typical distance between ridges is ~500 m.

Meteorological and energy flux (METFLUX) data were collected at eight sites within the watershed covering the main vegetation biomes. In the original data analysis, sensible heat flux $H$ was computed using the variance technique [e.g., Tallman, 1972; Albertson et al., 1995] with measurements at 4 m, and latent heat flux LE was solved as a residual with measurements of $R_n$ and $G$ (i.e., $LE = R_n - G - H$), where $G$ is the soil heat flux. At several sites the variance method was calibrated and tested with one-dimensional (1-D) eddy correlation data collected with a 1-D sonic anemometer, fast response thermocouple, and Krypton hygrometer located at ~2 m above the surface [Stannard et al., 1994; Kustas et al., 1994a]. Kustas et al. [1994a] found that differences between the calculated fluxes and those derived by 1-D eddy correlation measurements of $H$ and LE were within 20% on average. Measurements of wind speed and air temperature at METFLUX sites were made at a nominal height of 4 m. All data were averaged over 20-min intervals.

The radio soundings were conducted within the watershed boundary. Measurements were made of dry and wet bulb temperatures and pressure from the sonde, while elevation and azimuth angles were monitored from a theodolite. These data were used to compute temperature, humidity, and wind speed and direction as a function of height. For a summary of the measurements, see Hipps et al. [1994].

The thermal infrared data used in this analysis were acquired with the NS001 sensor mounted in a NASA C-130 aircraft having a pixel resolution of approximately 6 m. Humes et al. [1997] evaluated the overall quality of radiometric temperatures derived from the NS001 multispectral scanner data by correcting for atmospheric effects and comparing the resulting temperature with simultaneously acquired, ground-based remote sensing measurements over two large target areas. The radiative transfer code used for the analysis was Lowtran, Version 7 [Kniezys et al., 1988]. The atmospheric corrections computed with Lowtran7 were applied on a pixel-by-pixel basis to the entire image. Adjustments were made to the atmospheric corrections based on the scan angle across the image. The results suggest that the aircraft-based radiometric temperatures generally agree with the ground-based temperature measurements to within ~1°–2°C. For more details, see Humes et al. [1997].

Passive microwave brightness temperature images having a pixel resolution of ~200 m were also collected from the NASA C-130 aircraft using the push broom microwave radiometer and were converted to near-surface soil moisture maps by Schmugge et al. [1994]. A Landsat thematic mapper (TM) image collected soon after the Monsoon '90 experiment was used to create a land use and vegetation index map for the whole 150-km$^2$ basin [Kustas and Goodrich, 1994; Kustas and Humes, 1996].

The Landsat TM scene used to estimate the red and near-infrared (NIR) reflectance factors was acquired during the postmonsoon season on September 9, 1990, day of year (DOY) 252. Comparison of ground-based observations of NIR and Red reflectance during the study period and DOY 252 from Moran et al. [1994] indicated little change in their values, which suggests only minor differences in vegetation conditions would have existed. The comparison of the NIR and Red reflectance factors from the Landsat TM image with ground observations showed very good agreement.

The Landsat TM scene was also used in classifying each pixel at the required resolution as either grass-dominated, shrub-dominated, bare soil, or riparian vegetation. With each classification, values of canopy height, momentum roughness length, and leaf width were assigned. Values of $z_{can}$, the momentum roughness length, for the different classifications were guided by the analysis of Kustas et al. [1994a] and Menenti and Ritchie [1994]. Vegetation height estimates were guided by the observations of Wetz et al. [1994] and Stannard et al. [1994], and estimates of leaf width are from M. A. Weltz (personal communication, 1994). See Kustas and Humes [1996] and Schmugge et al. [1998] for more details.

During the 2-week campaign a number of missions were flown covering a wide range of surface soil moisture conditions from uniformly dry to variably wet conditions. The image data acquired on August 9, 1990, DOY 221 during 10:00–10:30 A.M. mountain standard time (MST) were selected for this analysis. During the aircraft overpasses, there were clear-sky conditions within the study area. The surface soil moisture (0–5 cm) was spatially variable because of several recent rainfall events [Humes et al., 1997; Schmugge et al., 1994]. Average wind speed was approximately 6 m s$^{-1}$, and potential air temperature at 100-m elevation was approximately 23°C. The average surface sensible heat flux was about 150 W m$^{-2}$, and the average surface latent heat flux was about 220 W m$^{-2}$, as computed from the METFLUX network. Daytime average Bowen ratios over the METFLUX measurement footprints varied from ~0.5 to slightly larger than 1 indicating significant variability in partitioning of available energy into latent and sensible heat flux [Kustas et al., 1994b]. This significant variability in the heat fluxes is also predicted from surface energy balance models using remotely sensed surface temperature or soil moisture but with uniform meteorological inputs [Humes et al., 1997; Schmugge et al., 1998; Kustas et al., 1998].

The 20-min average of measured incoming solar radiation was equal to 807 W m$^{-2}$, based on the METFLUX network data set at a time closest to the aircraft overpass time (i.e., ~10:10 A.M. MST). Averages of mixed-layer wind speed and potential temperature and free atmospheric lapse rate derived from the two free soundings bracketing the time of the aircraft overpass were employed in initializing the vertical structure of the LES model simulations. The average lapse rate above the capping inversion was approximately 5 K km$^{-1}$ for both soundings.

2.4. Numerical Methods

The basic structure of the numerical routines has been described before [Albertson, 1996; Albertson and Parlange, 1999a,
The three-dimensional turbulent fields of the three velocity components \((u_1, u_2, \text{and } u_3)\), air temperature \((\theta)\), and water vapor \((q)\) evolve in the simulated ABL through the numerical integration of the filtered Navier-Stokes equations, augmented by the dynamic boundary fluxes provided by the remotely sensed surface states and the two-source flux model. The surface flux fields adjust at each time step as they also depend on the distribution of the instantaneous air properties. Turbulent fluxes above the surface, internal to the flow, are readily computed as the inner product of velocity and scalar fields. A 120-min time averaging of these individual and joint fields was performed during the LES run. We focus here on the characteristics of these time-averaged fields and their relationships to one another that evolve through the fully coupled and three-dimensional simulations.

The time-averaged surface flux fields are presented alongside the remotely sensed surface temperature, soil moisture, and normalized difference vegetation index (NDVI) boundary conditions in Plate 1 (top). The bottom row of fields in Plate 1 shows the time-averaged available energy \(R_n\), sensible heat flux \((H)\), and latent heat flux \((LE)\) fields that developed over the land surface. We note that the regions of high \(R_n\) appear to be influenced more by the lower albedo of the vegetated regions than by the temperatures of the lower fractional cover regions. The sensible heat flux field is most strongly affected by the surface temperatures. The latent heat flux field is, as expected, strongly coupled to the NDVI, given that this is an area with active transpiration from the vegetated regions and very low soil evaporation from the more sparse regions. The actual relationship that develops between surface states and surface sensible heat flux is not as obvious as the surface exchange equations might suggest at first glance (see (A6) and (A10) of Appendix A), because in this fully coupled approach the air properties are dynamic and potentially affected by the underlying surface states, thus introducing nonlinearity into the surface flux relationships [e.g., Mahrt et al., 1994; Kustas and Humes, 1997]. The extent of this impact in this study is a topic addressed below.

The values of the correlation coefficients that were observed between the surface states and time-averaged surface flux fields are reported in Table 1. As expected, there is a strong correlation between surface temperature and sensible heat flux, with the departure from unity due to the effects of the fractional cover variability, wind field variability, and the interaction between air temperature and surface temperature. The nature of this semiarid landscape, as encountered on day 221 with plentiful root zone soil water and reduced soil surface moisture (for bare soil evaporation), yields a near-perfect correlation between NDVI (as related to fractional cover) and LE. Such a high correlation would not exist if the vegetation were under stress or water-limited conditions. However, the field data suggest adequate root zone moisture was available, resulting in a significant correlation between vegetation index and relative water use [Kustas et al., 1993]. Moreover, soil-vegetation-atmosphere model simulations indicate adequate moisture availability for plant transpiration [Fierzinger et al., 1998]. These qualitative points serve mainly as justification for the LE modeling approach that we used, which clearly injects a high correlation between NDVI and LE.

The air temperature and specific humidity and the vertical fluxes of latent and sensible heat were averaged through both
time and the horizontal directions for presentation as mean vertical profiles in Figure 1a. The air potential temperature and specific humidity profiles are tracking accurately the measurements from the field experiment. The spatial average of surface fluxes $H$ (150 w m$^{-2}$) and LE (205 w m$^{-2}$) compare extremely well (within 10%) to the observed values from the field experiment [Kustas et al., 1994b]. Note that absolutely no tuning of parameters was performed; this is a strong statement.

**Plate 1.** (top) Remotely sensed surface boundary conditions: radiometric surface temperature ($T_s$ (°C)), soil moisture (wetness), and normalized difference vegetation index (NDVI) over the study region (Monsoon '90 day 221). (bottom) Net radiation field ($R_n$ (W m$^{-2}$)), derived by large-eddy simulation (LES), surface sensible heat flux ($H$ (w m$^{-2}$)), and surface latent heat flux (LE (w m$^{-2}$)). Note that only one third of the y extent of the domain is shown here for illustration purposes.
about the accuracy of the combined two-source surface model and LES code. The temperature lapse rate in the capping inversion was initialized as 5°C km⁻¹ to correspond with the experimental conditions. Previous model runs used to explore the dynamical effects of the capping inversion lapse rate revealed a strong relationship between the strength of the lapse rate and (1) the magnitude of the negative sensible heat flux near the base of the inversion and (2) the shape of the vertical latent heat flux, with a steeper decline of LE with +z through the mixed layer for the stronger capping inversion lapse rate and a more gently decreasing LE with height for the modest lapse rate of 5°C km⁻¹. These results are attributed to the stronger lapse rates impeding penetrative convection and damping out this downward flux of relatively warmer, drier air into the top of the ABL. The magnitude of the entrainment flux of sensible heat corresponds reasonably well to Tennekes's [1973] rule of thumb that the entrainment flux is near 20% of the surface flux (H). The profile of the latent heat flux derived from this coupled model has a shape characteristic of that found in field experiments [Brutsaert and Kustas, 1987; Betts et al., 1992].

In Figure 1b we compare the range of land surface flux values simulated over the study site to the range of values measured from the eight METFLUX stations in the region. The LES and measurements compare quite well both with respect to regionally averaged fluxes and the ranges of variability of the fluxes over the landscape.

In addition to the excellent performance with respect to mean profiles the internal dynamics of the turbulence exhibits

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<td>H</td>
<td>0.70</td>
<td>-0.48</td>
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<td>LE</td>
<td>-0.59</td>
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Table 1. Correlation Coefficients Between Surface States/Properties and the Time-Averaged Surface Fluxes From the Coupled Model

Figure 1a. Time-averaged and horizontally averaged vertical profiles of potential air temperature (θ), sensible heat flux (H), specific humidity (q), and latent heat flux (LE) from the 120-min averaging period over the Monsoon '90 site. The surface fluxes are noted as the surface intercepts on the H and LE panels.
necessary conditions, such as the power spectrum of the $u_1$ fluctuations, as shown in Figure 2. There is close resemblance of the simulated turbulent cascade (i.e., slope in mid to small scales) to the theoretical inertial subrange slope of $k^{-5/3}$ [Kolmogorov, 1941], as referenced by the offset line with a slope of $k^{-5/3}$. This is a critical requirement for applications seeking to simulate the generation of surface flux eddies and explore their fate in the context of turbulent dynamics over heterogeneous terrain. The cascade from large eddies, which are generated through the influence of the boundary conditions, down to smaller eddies, which are dissipated by the subgrid viscosity, can only occur through a three-dimensional process, as vorticity in any arbitrary two-dimensional plane is stretched in the third dimension [Frisch, 1995, p. 241]. Through the cascade processes the effects of the boundary conditions are progressively lost as the turbulence becomes homogenized and rendered isotropic in the mid to small scales, as necessary to recover Kolmogorov’s slope of $k^{-5/3}$. An accurate simulation of this is necessary to elucidate potential scale preferences in the strength of land-atmosphere feedbacks, as addressed later in this section.

Hence we conclude at this point that the LES is providing a simulated ABL that is responding realistically to the heterogeneous surface conditions, with the feedbacks included naturally. We now perform a posterior analysis of the time-averaged results to explore relationships between the resulting surface fluxes and ABL properties and the underlying boundary conditions provided by the remotely sensed surface data sets.

In Figure 3 we show the relationship over the land surface between land cover, as described by NDVI, and the LES-derived Bowen ratio ($\beta = H/LE$), which is a simple measure of the preference for the surface to partition received energy as sensible heat to the atmosphere rather than latent heat through evapotranspiration (note each point on the plot corresponds to a surface grid cell hence $N = 64^2 = 4096$). The range of values derived by the LES is realistic in consideration of the field-scale-averaged values observed from the METFLUX stations (i.e., 0.5–1.0) [Kustas et al., 1994b]. Note that the spatial averaging inherent to the field measurements will reduce the range encountered, with high and low values canceling to some extent. It is evident that the spatial distribution of vegetation (as seen by NDVI) is interacting with the surface energy balance physics and the turbulent air flow to create spatial patterns in the convective forcing of the ABL.

The first evidence of interaction between surface states and mean ABL properties is visible in Figure 4a, showing the positive correlation between surface temperatures and time-average air temperatures in the lower ABL. Each air temperature value is taken from $z = 7$ m and a distance about 100 m downwind of the corresponding surface temperature value (i.e., denoted $T_{a,lag1}$) to account for the advective impact of the streamwise velocity field on the correlations between surface and air properties (more on this below). Two additional and important conclusions can be drawn from this scatterplot: (1) the variability in surface temperature (range on horizontal axis) maps into a reduced spatial variability in time-averaged air temperature (range on vertical axis), and (2) the relationship between these time-averaged state variables is weak to moderate (in correlation), which can be attributed to the added effects of other surface states and properties (e.g., fractional cover, roughness, and soil moisture) and the possible influence that the spatial length scale of surface variability has on the degree of influence that surface temperature exerts on air temperature. These factors must be considered when assessing the nonlinearity of the relationship between surface heat flux and surface temperature. These results are in line with observational results linking spatial variability in air temperature to surface temperature: Mahrt et al. [1994] observed air temperature variability of 1–1.5°C at about 30 m above a surface containing variability of about 20°C. Humes et al. [1997] observed air temperature variability between eight METFLUX stations of just $<1^\circ$C at a height of 4 m during Monsoon ’90. We extend this analysis in Figure 4b using spatially aggregated

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**Figure 1b.** Density functions of land surface fluxes with vertical lines that define the minimum and maximum values observed among the eight meteorological and energy flux (METFLUX) stations during the time period corresponding to the simulations.
(averaged) surface and air temperatures. The square grid cells are averaged to 2 × 2 cells, 4 × 4 cells, . . . , 16 × 16 cells, and a correlation coefficient is computed between these aggregated surface and air temperatures at each scale of aggregation. It appears that the small-scale surface features that do not translate into variations in air temperature exist primarily below a length scale of 500–1000 m. Averaging out these small-scale features and preserving the large-scale features progressively

Figure 2. Power spectral density plot of streamwise velocity field from LES turbulence simulation (\(E_u\)) versus wave number (\(k\) (rad m\(^{-1}\))) shown in circles. The theoretical scaling law of Kolmogorov [1941] for the inertial subrange (i.e., \(k^{-5/3}\)) is shown by a solid line segment offset from the computed spectrum.

Figure 3. Scatterplot of the time-averaged Bowen ratio (\(H/LE\)) computed for each surface grid cell in the LES against corresponding surface NDVI values. The curvilinear relationship shown here is linear when presented as \(\ln (H/LE)\) versus NDVI.
improves the correlation between the variables up to this threshold length scale.

To further explore the propagation of surface heterogeneity effects into the ABL, we measure the correlation between surface temperature at position \((x, y)\) with mean air temperature at position \((x + r, y, z)\). The correlation functions are integrated over \(x\) and \(y\) and indexed on arguments \(r\) (lag) and \(z\) (distance above surface),

\[
\rho_{r, z}(r, z) = \frac{\langle (T_r(x, y) - \langle T_r \rangle) \theta(x + r, y, z) - \langle \theta(z) \rangle \rangle}{\langle (T_r(x, y) - \langle T_r \rangle)^2 \rangle^{1/2} \langle (\theta(x, y, z) - \langle \theta(z) \rangle)^2 \rangle^{1/2}},
\]

where the tilde denotes time averages and the angular brackets denote horizontal averages over the full spatial extent of the modeling domain (i.e., study site). Hence the parameter \(z\) in the correlation function determines the height in the ABL for the air temperature field, and \(r\) determines the horizontal shift under consideration. Note that because of the periodic boundary conditions, the sample size does not decrease in the averaging operator of the numerator of (8) for increasing \(r\); the field simply “wraps around.”

The derived function \(\rho_{r, z}(r, z)\) is plotted against \(r\) and \(z\) in the top of Plate 2. Note the propagation of the surface information into the air properties and how this correlation moves downwind and decays in strength with height. It is apparent that the maximum correlation is about 70% and that the ridge of maximum land-atmosphere correlation propagates upward at a rate approximately one tenth of its downwind propagation rate. As a general feature, this ratio will change with the relative intensity of convective to mechanical turbulence (see the blending layer discussion of Albertson and Parlange [1999b] and the internal boundary layer review by Brutsaert [1982, chapter 7]). The general decrease in magnitude and downwind shift of the peak correlation with increasing \(z\) is clear in the bottom graph of Plate 2, where we plot \(\rho_{T, \theta}\) versus \(r\) for three distinct values of \(z\).

The strong correlations do not necessarily imply a large range of variability in the time-averaged air temperatures sampled over the region but rather imply that the variability that arises reflects, in a geometric sense, the variability in surface temperature. The relative strength of the spatial variability in air temperature is apparent in Figure 5, where the spatial standard deviation in mean air temperature at height \(z\) \((\sigma_{\theta}(z))\), normalized to the standard deviation of surface temperatures \((\sigma_T)\), is plotted against height into the ABL. Hence we see that the spatial variability of time-averaged air temperature for the Monsoon ’90 day 221 at a height of 7 m is less than 6% of the surface temperature variability. Further analysis reveals that the variability decays linearly with the log of the height in the bottom 100 m of the ABL. The strong decay with height is in qualitative agreement with tethersonde observations collected over both wet and dry areas during Monsoon ’90 [Hippes et al., 1994]. This analysis provides a clear measure of the nonlinearity of the relationship between surface heat flux and surface temperature through the impact of surface temperature variability on air temperature variability. In essence, this shows how closely the time-averaged air properties on average reflect the surface properties. This work can serve as a guide and error bound for efforts to use the spatial average (or single-site value) of air temperature to compute spatial fields of surface fluxes from remotely sensed land surface data. We now turn to a further analysis of the role of the length scale of surface temperature variations in the strength of this interaction between surface and air temperatures, as suggested in Figure 4b.

The energy spectrum of the spatial field of surface temperature is a presentation of the spatial variance of surface temperature decomposed into contributing length scales. The area under this spectrum is the spatial variance of the surface temperature. In Figure 6a we present the spectrum of surface temperature and the spectrum of (time averaged) air temperature at a height of 7 m. In both cases these are spectra of spatial fields, such that the horizontal axis is wave number \((\text{rad} m^{-1})\), which is inversely proportional to wavelength. Note that in addition to having a smaller variance than the surface temperature, the spectrum of the time-mean air temperature decays faster for increasing wave number (i.e., decreasing wavelength); thus it has less of its variance in the small spatial scales than does the surface temperature. This is examined in Figure 6b, where we normalize the square root of the air temperature energy at each wave number by the square root of the surface temperature energy at that wave number. We take the square root of the energy for dimensional consistency with the standard deviations of Figure 5. Note in Figure 6a a marked spatial-scale dependence of the relationship between surface temperature and air temperature, showing that the large-scale surface features express themselves in the air temperature, while the small spatial-scale surface temperature variability does not. This is clear evidence that the strength of land-atmosphere feedback processes are not scale-invariant for the range of
length scales studied here. The processes at work here are distinct, and spectrally separated, from those active as mesoscale feedbacks such as regional circulations [e.g., Mahrt and Ek, 1993; Giorgi and Avisar, 1997].

4. Conclusions

A new framework was presented for studying the interaction of the land and the atmosphere over heterogeneous regions. Remotely sensed land surface properties and state variables were incorporated as boundary conditions to a three-dimensional transient simulation of ABL turbulence, using the large-eddy simulation technique. This coupled system was used to study conditions from day 221 of the Monsoon '90 field campaign [Kustas and Goodrich, 1994].

The development of a two-source surface flux model in the LES provides a robust means to capture the relative contributions of bare soil and vegetation fluxes to the subgrid exchange of water and heat at the land surface. This is an important feature to include when using radiometric surface temperature and remotely sensed soil moisture for energy flux partitioning in regions containing a wide range in fractional vegetation cover conditions [Norman et al., 1995]. The surface model incorporates energy availability and spatial fields of remotely observed surface cover, temperature, and moisture.

Horizontally averaged and time-averaged ABL profiles of
Figure 5. The spatial (horizontal) standard deviation of the time-averaged air temperature normalized by the standard deviation of the radiometric surface temperature (times 100%) plotted versus height into the ABL.

Figure 6. (a) The energy spectrum of the radiometric surface temperature (solid line) and the energy spectrum of the time-averaged air temperature (circles) taken from the first plane of nodes above the land surface and plotted against wave number ($k$ (rad m$^{-1}$)). Increasing wave number denotes decreasing wavelength of surface variability. (b) The square root of the air temperature energy divided by the square root of the surface temperature energy at each wave number to show the scalewise coupling of surface temperature variability and time-averaged air temperature spatial variability.
temperature, specific humidity, sensible heat flux, and latent heat flux were well simulated by the model using the remotely sensed surface data. The entrainment processes were captured by the LES, providing a potentially important process not captured by lower-dimensional models of land-ABL interaction. Modeled land surface fluxes compared well to both the regional means and ranges of variations of the measured fluxes from the METFLEX stations.

The spectral representation of the turbulent cascade in the LES was shown to follow closely Kolmogorov’s [1941] inertial subrange scaling \((k^{-5/3})\) for scales approaching the mesh spacing. This is evidence of realism in the turbulent dynamics. The coupled LES–remote sensing model produced turbulent air fields that reflected the combined effects of the underlying surface states and the ABL transport and mixing. Hence the ultimate relationships that arise between forcing and the underlying surface states and the ABL transport and mixing. Hence the feedback strength between the produced preferentially from the large spatial-scale variations in surface temperature. Hence the feedback strength between the

\[ f(x, y) = 1 - \left( \frac{\text{NDVI}_{\text{max}} - \text{NDVI}(x, y)}{\text{NDVI}_{\text{max}} - \text{NDVI}_{\text{min}}} \right)^{0.625}, \]  

(A1)

where \(\text{NDVI}_{\text{max}}\) and \(\text{NDVI}_{\text{min}}\) are the maximum and minimum NDVI values, respectively, in the study site [Gillies et al., 1997]. Correspondingly, the fraction of a grid cell covered by bare soil is \((1 - f_s)\).

The leaf area index (LAI), as used to estimate radiative fractionation between soil and vegetation, is estimated after Norman et al. [1995] as

\[ \text{LAI}(x, y) = -2 \ln (1 - f_s). \]  

(A2)

The net radiation at each surface grid cell is

\[ R_s(x, y) = R_s^0(1 - \alpha(x, y)) + \varepsilon(R_s^0 - \sigma T_s(x, y))^4, \]  

(A3)

where \(R_s^0\) is the incoming shortwave radiation (specified for the study site and time), \(\varepsilon\) is the surface emissivity, \(R_s^0\) is the downwelling longwave radiation as computed from air temperature and humidity using Brutsaert’s [1975] clear-sky model, and \(\sigma\) is the Stefan-Boltzmann constant. The net radiation is partitioned between the two sources, soil and vegetation, as [Norman et al., 1995]

\[ R_s^{\text{ev}}(x, y) = R_s(x, y)[1 - \exp (-0.6 \text{ LAI}(x, y))]. \]  

(A4)

The latent heat flux from the vegetation is computed by the Priestly and Taylor [1972] equilibrium evaporation rate taken in the context of the fraction of radiation captured by the canopy, i.e.,

\[ \text{LE}^{\text{ev}}(x, y) = 1.26 \left[ \frac{\Delta}{\Delta + \gamma} \right] R_s^{\text{ev}}(x, y), \]  

(A5)

where \(\Delta\) is the slope of the saturation vapor pressure curve and \(\gamma\) is the psychrometric constant. Justification for such a formulation is discussed by Norman et al. [1995].

The sensible heat flux from the plant canopy is determined as the residual in the vegetation energy balance, i.e.,

\[ H^{\text{ev}}(x, y) = R_s^{\text{ev}}(x, y) - \text{LE}^{\text{ev}}(x, y). \]  

(A6)

Note from (A1)–(A6) that the latent and sensible heat fluxes from the canopy fraction of each grid cell are computed solely on energy principles. The fluxes from the bare soil fraction are computed on the basis of aerodynamic exchange equations, as derived from turbulent diffusion principles.

Appendix A

Here the two-source model used for computing surface fluxes in time and space over the lower boundary in the LES is described algorithmically. This model is related to that developed by Norman et al. [1995] for radiometric temperature observations and by Kustas et al. [1998] for remotely sensed soil moisture, but it is tailored to the LES framework.

The land surface in the LES is treated as a hypothetical flat plane, described by grid cells which are partially covered by vegetation and partially covered by soil. The surface cover over the study site is described by the input fields of momentum roughness length \((z_0)\), normalized difference vegetation index (NDVI), albedo \((\alpha)\), zero-plane displacement height \((d_0)\), canopy height \((h_c)\), and general leaf width \((l_w)\). The surface state at the time of the study is described by fields of radiometric surface temperature \((T_s)\) and relative saturation of soil moisture \((w)\).

A1. Vegetation Fluxes

From the NDVI an estimate of the area fractional cover of vegetation on each surface grid cell is derived from Choudhury et al. [1994]
\[ T_{eq}^{\text{soil}}(x, y) = T_{eq}(x, y) \]
\[ + \frac{H_{eq}^{\text{soil}}(x, y) [\ln (z/z_o) - \Psi_n] [\ln (z/z_{oh}) - \Psi_h]}{k'U(x, y, z)}, \]
(A8)

where \( z_o \) and \( z_{oh} \) are the momentum and scalar roughness lengths, respectively, \( \Psi_n \) and \( \Psi_h \) are the stability correction functions for the shapes of the velocity and temperature profiles, respectively, in an unstably stratified boundary layer, \( k \) (0.4) is von Karman’s constant, and \( U \) is the instantaneous wind speed in the LES just above the surface grid cell of interest (at height \( z \)). In addition to spatial dependence the heat flux, velocity, and temperatures of (A8) fluctuate in time as well.

The sensible heat flux from the soil fraction of each grid cell is computed on aerodynamic exchange principles as
\[ H^{\text{soil}}(x, y) = \rho L C_s(v T_{eq}^{\text{soil}}(x, y) - T_{eq}(x, y)), \]
(A9)

where \( \rho \) is the air density, \( C_s \) is the specific heat capacity, \( C_h \) is an aerodynamic heat exchange term (that includes the local instantaneous wind speed, the local roughness length, and the effects of stability in the context of Monin-Obukhov similarity theory, with a reduction term (or resistance) to account for the sheltering effect of the vegetation [Norman et al., 1995]), and \( T_{eq} \) is the instantaneous air temperature over the cell of interest.

The latent heat flux from the soil is computed similarly,
\[ LE^{\text{soil}}(x, y) = \rho L C_e(h q^* (x, y) - q_e(x, y)), \]
(A10)

where \( L_e \) is the latent heat of vaporization, \( C_e \) is analogous to \( C_h \), except that it includes an additional reduction (or resistance) to account for the constraint on diffusion of water vapor through the soil pores, \( h \) is an estimated surface relative humidity term (a function of \( w \)). This formulation follows that of Kustas et al. [1998] which was developed to use microwave-derived surface soil moisture.

For completeness we compute a soil heat flux as
\[ G^{\text{soil}}(x, y) = R^{\text{soil}}_n(x, y) - H^{\text{soil}}(x, y) - LE^{\text{soil}}(x, y). \]
(A11)

### A3. Total Grid Cell Fluxes

The total instantaneous sensible and latent heat fluxes are taken at each time step as the sum of the two sources contributions, i.e.,
\[ H(x, y) = H^{\text{soil}}(x, y) + H^{\text{land}}(x, y), \]
(A12)
\[ LE(x, y) = LE^{\text{soil}}(x, y) + LE^{\text{land}}(x, y), \]
(A13)

where the time dependence has been omitted throughout this presentation for compactness.

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