Issues in Modeling the Impact of Land Use Change on the Global Climate

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The prediction of the impact of human activities on the climate system, in particular the impact of anthropogenic land use change, hinges on the ability to properly model the interaction between the heterogeneous land surface and the atmosphere in global climate models. Heterogeneities in the land surface exist on a wide range of spatial scales and make modeling the land-atmosphere coupling particularly complex. Observational, fine mesh modeling, and now global model evidence exists which suggests that the vertical extent to which these heterogeneities directly affect the boundary layer turbulence is an important consideration in understanding the land-atmosphere interaction. A global climate model that includes a conceptualization of this physical process may improve the ability to assess and predict the effects of land use change on the climate.

General Circulation Model (GCM) simulations have been conducted with an ‘Extended Mosaic’ (EM) technique for the land-atmosphere coupling, which allows the direct impact of surface heterogeneities to extend upwards into the overlying turbulent boundary layer, and with a standard Mosaic technique. This study evaluates the results of these simulations in terms of their implications for the strength of the coupling between land and atmosphere and how a change in the character of the land surface may propagate through the climate system. We conclude that allowing the direct influence of surface heterogeneities to extend in the vertical results in a strengthening of the land atmosphere coupling in many regions.

Key words: Climate change, land use change, modeling.

Introduction

Records of the observed global surface air temperature over the past century indicate a clear warming trend for our planet (IPCC 2001). The by now familiar figure of the temperature anomaly curve since 1850 (Figure 1) shows, for example, that the year 2001 was the second warmest on record. Many studies are reported which evaluate the impact of the temperature trend and the changes in moisture that it brings on regional and local ecosystems and on the ability of the land to be used for agriculture, grazing and ecological evolution of species. These studies commonly use the IPCC simulations under the different scenarios to evaluate the impact of climate change on the local land use and ecology (for example, see Bounoua 1999; Solecki and Olivieri 2004). The focus of this article is the segment of the feedback between the land surface and the climate in which the human-induced land use change impacts the climate. We examine here the modeling techniques that are necessary to represent the contribution of land use change to the observed temperature trend.

There is now evidence that the warming observed over the last 50 years is due in part to anthropogenic causes. Figure 1 shows comparisons between observations and global model simulations of the evolution of the earth’s climate over the period 1860-2000, and the closest agreement between observed and simulated climate is obtained when the simulations include
both natural and human factors. The simulations include the effects of the anthropogenic increase of greenhouse gases in the atmosphere as well as a rudimentary accounting for the effects of anthropogenic land use change. Anthropogenic land use change (LUC) includes urbanization, deforestation, desertification and agricultural practices, and has been shown recently to have an important effect on climate and climate change (Pielke et al. 2002). Urbanization and the associated heat island has been shown to contribute regionally to the warming trend of the last 50 years over the U.S. (Gallo et al. 1999), and the contributions of urbanization and agricultural expansion combined have been shown to account for as much as half of that trend (Kalnay and Cai 2003). Deforestation in the Amazon has been shown to influence both regional and remote climate by inducing local changes in cumulus convection and exciting planetary scale wave activity Werth and Avissar 2002). The change in the Florida Everglades between 1900 and 1990 from wetlands and marsh to agricultural and urban terrain has been shown to be connected with a drier and warmer climate (Pielke et al 2001), and the conversion of terrain in south-western Australia has been shown to be associated with increased formation of cumulus clouds in that region (Ray et al. 2001).

That changes in the land surface may impact the global temperature trend is related to the critical role that the land surface plays in the climate system. The energy and material exchanges that occur at the land surface act to partition the net radiation into surface heating, deep soil heating, sensible and latent heat fluxes, act to redistribute precipitation into evaporation, soil storage, ground water recharge and runoff, and act to regulate biogeochemical cycles such as photosynthesis, transpiration, the nitrogen cycle, and carbon uptake. The fluxes at the land surface are known to significantly influence rainfall, temperature and circulation, from daily to inter-glacial scales (Milly and Dunne 1994; Polcher 1995). The effects of these land-atmosphere exchanges may also extend beyond the regions in which they are initiated by inducing modifications of the large-scale circulation. To understand and predict the behavior of the earth’s climate system and what contributes to climate trends it is therefore important to realistically simulate the land surface-atmosphere exchanges and the range of modes of variability in global climate models (Leetma and Higgins 1999).

High resolution (grid size < 10 km) models covering particular regions around the globe have been used to accurately simulate the land-atmosphere interaction and demonstrate the influence of land use change on regional climate. High-resolution models have also been nested inside global models to study the influence of land use change on remote climate (Hahmann and Dickinson 1997). The computational limitations, however, make this an impractical approach for the global high-resolution coverage needed to model and predict the climate feedbacks and to make longer-term predictions. Many studies of the effects of deforestation and overgrazing have been conducted using coarse resolution global models, but they still produce results that differ from high-resolution models and observations, and the results are highly dependent on the specific model formulation (Koster et al. 2002). Properly capturing (parameterizing) the effects of land use change in global models, then, in which the changes in land use occur on scales smaller than the typical grid scale (~100 km), is crucial to the prediction of climate and climate change, and to the prediction of the effect of LUC on the changing climate.

The summary of existing work that will be presented here is centered on techniques for modeling the interactions between a heterogeneous land surface and the atmosphere in global climate
models. The background section describes theoretical, observational and modeling studies which show that the presence of heterogeneities in the land surface on different scales influences the manner in which the land surface and the atmosphere interact, and in particular that the influence extends vertically well away from the land surface. The background also includes a review of commonly used techniques to model these interactions in global climate models, along with their limitations in the ability to capture the vertical extent of the direct influence of land surface heterogeneities. In the following section called “Extended Mosaic and the Model Blending Height” we present a newly developed technique that has the ability to overcome some of these limitations. We next discuss the impact on global climate simulations of using different modeling techniques, and show the potential for the new technique to capture important climate feedback mechanisms. We conclude with a summary of the implications for modeling the impact of land use change on regional and global climate.

Background

Subgrid Scale Heterogeneities and the Mosaic Approach

The calculation of the land-atmosphere exchange in global climate models, as well as the prediction of the potential impact on climate of land use change, is complicated by the fact that the character of the land surface is highly variable, due to factors such as the variability of vegetation cover, the types of terrain, soil texture and wetness, the amount of cloud cover and precipitation, and the extent of urban areas. The wide range of scales of heterogeneity in the land surface can be seen in the global vegetation maps that have been compiled from visible and other imagery (Figure 2). Figure 2a shows the continental scale heterogeneities in South America, and Figure 2b shows the kilometer (or smaller) scale heterogeneities present in a small urbanized region. The scale of these heterogeneities may be smaller, and in some cases much more so, than the characteristic grid scale in most current General Circulation Models (GCMs) used in climate studies. Almost all the Soil-Vegetation-Atmosphere-Transfer (SVAT) models that are coupled to state of the art regional and global climate models employ some technique to attempt to account for the subgrid-scale heterogeneities. The most often used techniques are ‘dominant’, ‘composite’ and ‘mosaic’ and they are briefly discussed here. Comparisons between these different schemes have served to demonstrate the strengths and weaknesses of the schemes, and also to aid in the understanding of the influence of small-scale land surface heterogeneity on the atmospheric boundary layer and climate.

The earliest of the SVAT formulations for GCMs assumed that the land surface in a GCM grid square can be adequately described by the ‘dominant’ soil and vegetation characteristics from climatology (Dickinson et al. 1986). A single set of parameters is specified in the dominant technique that realistically describes the predominant (largest areal fraction) vegetation and soil type in any GCM grid box. The dominant technique cannot, however, account in any way for the influence of other vegetation and soil types that may exist over significant areas of the grid box. Many of the GCMs that are participating in the Atmospheric Model Intercomparison Project (AMIP) II (Gates 1995) account for the subgrid-scale variability by specifying soil and vegetation parameters that represent a homogeneous ‘composite’ vegetated surface and its underlying soil for each GCM grid square. Among these are the GCMs used at the ECMWF (Viterbo and Beljaars 1995), NCAR (Gates 1995), NCEP (Pan and Mahrt 1987), the Center for
A few of the AMIP II GCMs and some others account for the subgrid-scale heterogeneity using an approach referred to as the ‘mosaic’ approach. In this approach, the heterogeneous grid box is viewed as a ‘mosaic of independent vegetation stands’ (Koster and Suarez 1992), and is characterized by the different vegetation and soil types and their fractional area of coverage in the grid box. Separate surface heat and moisture balance equations are solved for each vegetation type contained within a GCM grid square and the resulting heat and moisture fluxes which describe the coupling to the atmospheric boundary layer are aggregated linearly. The mosaic approach is employed, for example, in the GCMs used at the Goddard Institute for Space Studies (Rosenzweig and Abramopoulos 1997), Laboratoire Météorologie Dynamique (Ducoudre, Laval and Perrier 1993), the Australian Bureau of Meteorology Research Centre (Desborough and Pitman 1998), and the NASA/Goddard Seasonal to Interannual Prediction Project (Koster and Suarez 1992). The turbulent diffusion in the boundary layer and above is then computed based on the grid-averaged surface flux of heat and moisture. The mosaic approach to handling sub-grid scale heterogeneities in a GCM is presented schematically in Figure 3, where a sample GCM grid square containing the ‘tiles’ that describe the mix of surface scene types is shown. In this example, all of the bare soil portions of the grid box are treated as though they are juxtaposed, as are all of the deciduous trees, evergreen trees, and shrubs. Each of these types is assigned a fraction of areal coverage, which is used to compute grid box aggregated fluxes.

Many researchers, using either in situ data or high-resolution models as the basis for comparison, have carried out validations of the mosaic approach for GCMs and comparisons against composite schemes. The results of these evaluations generally show that the mosaic approach is an improvement over a composite or dominant approach, and that mosaic provides a reasonable approximation of the turbulent fluxes at the surface under many conditions. An evaluation of the ability of ‘mosaic’ to capture the observed surface fluxes over a mesoscale-sized area of heterogeneous terrain was conducted by Bringfelt (1999), using in situ data taken in summertime over northern Europe. When the sensible and latent heat fluxes in the simulation were computed by the model using the observed net radiation, the agreement between model and observations was quite good. The mosaic approach captured the observed magnitude and diurnal variation of the surface fluxes, as well as the relationships between fluxes over different terrain types. The study concluded that the mosaic approach provided reasonable estimates of the magnitude and spatial variability of the surface fluxes. An example of a study in which in situ data were used to evaluate comparisons between simulations with a mosaic type scheme and a composite scheme is reported in van den Hurk and Beljaars (1996). The comparisons were performed over GCM grid-sized regions in the U.S. Central Plains and in a Spanish vineyard in summertime. The result for both climate regimes was an improvement in the mosaic simulation estimate of the grid-averaged fluxes and skin temperature relative to the composite simulation.

Limitations of Mosaic and Early Techniques

Observational and modeling studies (Claussen 1995; Mahrt 2000 and articles cited therein) have shown that the direct influence of the surface heterogeneities extends vertically in the
atmosphere up to a level that may be above the surface layer (which is nominally 50 m above the ground) and within the planetary boundary layer (on the order of 1 km above the ground). Early studies of the vertical influence of the land surface heterogeneities were conducted by Wieringa (1986), where he defined the ‘blending height’ as the level inside the planetary boundary layer above which the flow becomes horizontally homogeneous in the absence of other influences. The early studies were valid for small horizontal scales (less than 1 km), and Mason (1988) extended the analysis to include larger horizontal scales. He characterized a relationship between the horizontal scale of variability $L_h$ and the blending height $l_b$. The scaling was based on the assumption that the boundary layer flow is mainly determined by a balance between vertical advection and the divergence of the vertical turbulent flux. Claussen (1995) determined that from Mason’s heuristic model one can conclude that $l_b \sim 0.01 \times L_h$. This estimate agrees with the blending height estimates reported by the early studies, where the horizontal scales of 10 km being considered would imply a blending height of 100 m. For horizontal scales of heterogeneity of the order of 50 to 100 km or larger that characterize part of the globe (see Figure 2), this would imply a blending height of 500 m to 1 km, which is of the order of the planetary boundary layer height. A comprehensive survey of blending height estimates under different atmospheric conditions was presented by Mahrt (1996), where he reported that $l_b$ can be as high as the height of the planetary boundary layer (PBL), or even higher for an unstable atmosphere under the influence of strong surface heating.

Observational evidence exists which supports the scaling arguments that suggest that the blending height may extend well into the PBL. Segal et al. (1989) show, using aircraft measurements of temperature and humidity at four levels in the atmosphere over the boundary between an irrigated and a dry crop area, that the contrast between the wet, cool air over the irrigated area and the dry, warm air over the dry area extends up to at least 440 m above the surface. These observations would suggest that the blending height is somewhere above 440 m. The presence of an “elevated large scale blending height” can also be inferred from lidar measurements of the mixed layer height taken in the vicinity of Nashville, Tennessee, spanning the interface between a forested and an agricultural area. The mixed layer heights over the different terrain with patch sizes of ~30 km were shown to differ by up to 400 m (Angevine et al. 2003).

The existence of a blending height on larger scales is also demonstrated by a study that uses radiosonde data from the 1994 field observations obtained during the Boreal Ecosystem Atmospheric Study (BOREAS) (Salmin, Molod and Huang 2004). The data used in this analysis were from temporally concurrent measurements (daytime or afternoon), taken over a wide enough range of surface types and separated by distances comparable to a GCM grid size. Blending heights were observed over more than 80 percent of the vertical soundings from seven locations that were analyzed, and on average over 50 percent of the observed blending heights were above than 1 km. Figure 4 shows a set of temperature profiles as a function of pressure thickness above the surface for several of the BOREAS stations, including locations in the Northern Study Area (WTH and YYQ), higher elevation locations (WIQ and KEY) and intermediate locations between the Northern and Southern Study Areas (WTH and TPA). The air temperature over stations that are separated by distances of the order of 200 km are different near the surface, and approach each other until they are indistinguishable at the heights indicated by the small dash in the figure. The level at which they are indistinguishable is the blending
height. The collection of profiles shown in the figure demonstrates the range of variability of blending heights observed in this study. Figure 4 (a), (b) and (c) shows a typical daily evolution of the level of the blending height over a selected pair of stations, while panels (d), (e) and (f) are chosen to illustrate times and locations where the blending of air properties above different stations occurs at low, high and intermediate levels in the atmosphere, respectively.

Evidence that the blending height may be of the same order as the planetary boundary layer height has implications for the different techniques used to capture sub-grid scale heterogeneity in a GCM. The composite approach does not allow for the direct propagation of the independent characteristics of each vegetation type into the atmosphere at all. The ‘mosaic’ technique does not account for the vertical influence of the land surface heterogeneities beyond the height of the surface layer. In all the techniques described so far, the direct vertical influence of surface heterogeneities is confined to the surface layer or below. The limits on the vertical influence of surface heterogeneities imposed by the above techniques may well constitute an important limitation to capturing the effectiveness of the communication between the land surface heterogeneity and the atmosphere (Mahrt 2000).

The present review of the behavior of mosaic in comparison to composite or dominant has made it clear that under many conditions a mosaic approach provides a more realistic simulation of the land-atmosphere exchange. Because of the importance of the land surface to the global climate, this implies that a GCM using a mosaic approach would provide better simulations of the mean climate and of the response to a perturbation in forcing. The extended mosaic technique that is presented in Molod, Salmun and Waugh (2003) is designed to allow the level of the blending height to be diagnosed as a result of the behavior of the turbulent boundary layer over each different vegetation type, rather than be chosen a priori. Extended mosaic therefore addresses a limitation of previously existing techniques, and has been shown in Molod, Salmun and Waugh (2004) to have an important impact on the simulated climate in global models. The limitations of the mosaic approach that have been discussed here, and the ability to overcome some of them in the extended mosaic scheme, would imply that a GCM using EM ought to provide improved simulations of climate and of the climate response to a land use change as compared to using the standard mosaic.

‘Extended Mosaic’ and the Model Blending Height

The technique called ‘Extended Mosaic’ (EM) presented in Molod, Salmun and Waugh (2003) is designed to allow the influence of subgrid scale interactions between the land surface and the atmosphere to extend vertically. The sub-grid scale variability of the surface is modeled by computing the surface energy and moisture transfers in the surface layer and soil, as well as the energy and moisture transfers in the turbulent layer above, separately for each tile that makes up the mosaic. The net change in temperature and moisture (and passive tracers) are aggregated to compute a GCM grid average.

In an analogy to the blending height concept, Molod, Salmun and Waugh (2003) define the “model blending height” (MBH) as the model level above the heterogeneous land surface at which it is assumed that the turbulent mixing has homogenized the air mass properties. This is the level above which the flow over the different tiles begins to appear uniform. Each of the
aggregation techniques described above implies a different extent of the vertical influence of surface heterogeneity in the GCM. We can view the ‘composite’ methodology as setting an MBH at the ground and the mosaic approach as setting an MBH at the top of the surface layer. The EM approach is the only approach that has a variable MBH.

The algorithm to compute the MBH was derived based on results of a year-long GCM simulation using EM. Typical profiles of the turbulent fluxes of temperature, momentum and humidity over each “tile” (mosaic ‘patch”) in a grid box are shown in Figure 5. Figure 5a shows the profiles of $\bar{w}' S'$ and $\bar{w}' q'$ at 3 UTC over a grid point located in the northern steppes (58N, 75E). This particular grid box has three tiles, a broadleaf deciduous trees tile, a bare soil tile, and a needleleaf trees tile. The values of the flux emanating from the surface layer (those at the bottom model $\sigma$-level) are very different from tile to tile. We also see that the differences in the turbulent fluxes between the tiles have been propagated upwards, as the fluxes are still distinct aloft. At higher levels, values of $\bar{w}' S'$ approach each other until approximately $\sigma = 0.975$, where the fluxes are almost indistinguishable. At this ‘homogenized’ level, however, the fluxes are still significant. The turbulent boundary layer height is defined in the GCM as the level at which the turbulent fluxes decay below 10 percent of their column maximum value, and this does not occur until $\sigma = 0.95$. Figure 5b shows the turbulent fluxes of momentum for the same location and time, and these are seen to exhibit the same character. This qualitative sense of the location of the MBH was quantified by defining a measure of the surface variability and some threshold of variability below which the fluxes over different tiles may be considered homogenized. The measure of tile-to-tile variability ($V$) at any level is $V = F_{\text{max}} - F_{\text{min}}$, where $F_{\text{max}}$ and $F_{\text{min}}$ are the maximum and minimum values of the turbulent flux among all the tiles respectively at any given level. The measure of surface variability is $V_s = \max (F_{\text{max}} - F_{\text{min}})$, where ‘max’ of the difference refers to the level at which the difference is maximum (usually near the surface). As we ascend in the atmospheric column, $V$ decreases with height relative to its maximum value, $V_s$, and we define the MBH as the level at which it descends below the threshold.

The spatial variability of the MBH and the relationship with the planetary boundary layer height were shown in Molod, Salmun and Waugh (2003) from a year-long segment of a longer simulation performed with the EM technique as implemented in the GEOS-Terra GCM. JJA seasonal mean values of MBH, PBL, and the ratio MBH/PBL, given in pressure thickness above the surface, are shown in Figure 6. The PBL height in the GCM is the level at which the turbulent kinetic energy decays below 10 percent of its maximum value in the air column, and there is therefore no a priori relationship between the MBH and the PBL height. The plots in Fig. 5 show that the MBH is variable, is above the surface layer, and is related to the PBL height but does not follow it exactly. The values of MBH range from 10 mb to near 40 mb (100m to 400m, approximately). Some of the areas where the MBH is high correspond to high PBL values, such as the deserts of Sahara, Gobi, Saudi Arabia and southwestern United States, and regions in southern Africa and just south of the Amazon. An additional band of high MBH values exists at approximately 35°-40° N latitude, stretching across Canada, Europe and Asia. These high MBH values do not correspond to local maxima in the PBL height; rather they correspond in the shape of the contour to areas where there is a change in the nature of the vegetation (see Fig. 2 of Molod and Salmun 2002). In these regions the vegetation changes from a combination of bare soil and dwarf trees to predominantly needleleaf trees. The geographical
locations where the MBH is low are the Amazon, where the variability of the vegetation is small, and the Andes and the Tibetan Plateau, where the altitude and the cold temperatures limit the height of the PBL. There is also a local minimum in MBH in the southern tip of Africa, and another in central South America, just south of the local maximum. The ratio of MBH to PBL values is generally lower than in the northern hemisphere wintertime (see Molod, Salmun and Waugh 2003), with the lowest values (less than 20 percent) corresponding to the highest values of PBL height. Except for Southern Africa, the ratios are higher in the southern hemisphere (winter hemisphere) than in the summer hemisphere.

The evidence presented here about the behavior of the EM scheme, that it allows the formation of a blending height at a level dictated by the local mean flow and turbulence, indicates that when using this scheme a GCM can indeed account for the extension of the direct influence of land surface heterogeneities in the vertical. Allowing the different vegetation types contained within a GCM-sized grid area to exert the proper influence on the grid-scale behavior of the planetary boundary layer is an important component of modeling the climate impacts of land use change. Very often LUC is accompanied by a change in the character of the heterogeneity of the land surface as well as a change in its mean character, because small-scale features are either created (during deforestation or conversion to agricultural use) or destroyed (urbanization). The EM scheme is equipped to simulate this aspect of the climate impact of LUC in a GCM.

**Modeling the Impact on the Climate**

Two 10-year long simulations were performed with the GEOS GCM, and the impact on the resulting climate of using the Extended Mosaic technique rather than a Mosaic technique to couple the heterogeneous land surface and the atmosphere was evaluated by Molod, Salmun and Waugh (2004). The results of these side-by-side simulations showed statistically significant differences on local and global scales, and demonstrated that extending the direct influence of subgrid scale heterogeneities upwards into the boundary layer using EM has the potential to affect the modeled climate. This comparison between two modeling techniques, when viewed as a difference in the amount of subgrid scale variability that is allowed to propagate through the climate system by the different models, gives an indication of the response to a perturbation in the character of the underlying land surface.

The impact on the atmospheric circulation of EM was assessed by examining the ensemble mean stationary wave pattern. This impact is manifest directly in the seasonal mean deviation from the zonal mean meridional transport of heat or momentum, and also in the seasonal mean sea level pressure (SLP) field and 300 mb eddy height (deviation from the zonal mean geopotential height) field. The DJF seasonal mean sea level pressure fields for the EM-M differences and for the EM simulation alone are shown in Figures 7a and 7c, respectively. Similar spatial patterns are seen in the 300 hPa eddy height field, again shown for the EM-M difference and for EM alone, in Figures 7b and 7d. The pattern of alternating troughs and ridges in the DJF EM eddy height field (7), shown by alternating centers of light grey (ridge) and dark grey (trough) beginning with the ridge in the central Pacific and continuing into the northern Pacific and over the North-American continent into the Atlantic, is the Pacific North-American (PNA) pattern of variability, defined originally by Wallace and Gutzler (1981). The mechanism suggested to explain the presence of this stationary pattern in wintertime is the excitation of Rossby waves in
the subtropics and their propagation into higher latitudes (see, for example, Trenberth et al. 1998). The difference field (Figure 6b) has a large negative center in the northern Pacific region where the eddy height has a trough, and a positive center over the North American west coast where the eddy height has a ridge. There is a shadow of this pattern of negative EM-M difference over troughs and positive difference over ridges stretching across the North American continent into the Atlantic. This illustrates the strengthening of the PNA pattern of variability in the EM simulation. This result suggests that the manner in which land surface heterogeneities are modeled may exert an influence on the PNA pattern. The possible influence of the land surface hydrology on the amplitude of the PNA has been reported by Trenberth et al. (1998) in a study based on observations and model results, although no specific mechanism was suggested. Also, Werth and Aivissar (2002), in their study of the teleconnection patterns in European climate generated by Amazonian deforestation, suggest that planetary scale Rossby waves may indeed be excited due to processes that occur at the land surface, related to changes in landscape.

Another important indicator of the extent to which a perturbation in the character of the land surface (through anthropogenic LUC, for instance) will affect the global climate is the strength of the coupling between the land surface and the atmosphere. For this purpose, we regard each year of the 10-year simulations as an independent ensemble member, and following Koster et al. (2002), and Koster et al. (2004) (for the evaporation field for example), we computed the quantity:

\[ \Omega_E = \frac{10\sigma^2_E - \sigma^2_{\bar{E}}}{9\sigma^2_{\bar{E}}} \]

where \( \sigma^2_E \) is the variance of the evaporation E across all ensemble members and all time periods (10 and 12 respectively) and \( \sigma^2_{\bar{E}} \) is the variance of \( \bar{E} \), the ensemble mean. If each ensemble member is identical, then \( \sigma^2_{\bar{E}} = \sigma^2_E \) and \( \Omega_E = 1 \). If all the members of the ensemble are uncorrelated then \( \sigma^2_{\bar{E}} = 10\sigma^2_{\bar{E}} \) and \( \Omega_E = 0 \). Similar considerations can be made for the precipitation field.

If the sea surface temperature alone governs the behavior of the evaporation, then all the time series of evaporation from the different ensemble members would be nearly perfectly correlated and \( (1-\Omega_E) \) would be small. To the extent that natural variability (and atmospheric initial conditions) and the land boundary condition play a role in determining the outcome of the evaporation in each simulation, the time series would be less correlated and \( (1-\Omega_E) \) would be larger. We interpret \( (1-\Omega_E) \) as an indicator of the influence of the land surface on the variability of the evaporation because our results indicate that the magnitude of the natural variability in the EM and in the M ensembles is not substantially different. Global maps of the EM-M difference of \( (1-\Omega_E) \) for evaporation and of \( (1-\Omega_P) \) for precipitation are shown in figure 8. Regions
where the EM-M difference in \((1-\Omega_E)\) is positive, such as the large area in the central United States, indicate regions where the land surface in EM exerts more influence on the variability of the evaporation than in M, and vice-versa. The results of these simulations therefore indicate that in a model using EM the strength of the land-atmosphere coupling has been increased in many regions, from which we understand that EM would simulate an increased response relative to M to a change in the character of the land surface.

**Concluding Remarks**

The issues surrounding the interaction between a heterogeneous land surface and the climate system, in the context of the prediction of the impact of anthropogenic land use change on global warming, have been addressed here. In particular, we are concerned with how the effects of small patches of terrain in a heterogeneous landscape are captured in global models that have a grid size that is substantially larger than the scale of heterogeneities. Anthropogenic land use change, such as the fishbone pattern of Amazon deforestation, introduces changes in the patchiness of the land surface. Capturing the influence of these changes on the grid scale is essential for capturing and predicting the influence of LUC on climate. Due to the non-linearity of the interplay between the land surface and the atmosphere, these small patches are capable of exerting a disproportionate influence on the grid scale boundary layer characteristics. This disproportionate influence is due in part to the vertical extent to which the small patch influences the atmosphere. We have argued that it is critical to the proper modeling of the land surface interaction in global models to capture the vertical influence of the subgrid scale heterogeneities.

We have presented a review of results from simulations performed with a general circulation model (GCM) in which the land surface atmosphere coupling was accomplished using an ‘extended mosaic’ (EM) approach. EM allows the GCM to interactively determine the vertical extent to which subgrid scale heterogeneities directly influence the boundary layer, a process that is not captured in the standard mosaic technique. Climate simulations with EM demonstrated significant differences from climate simulations with the standard mosaic (M). These differences are concentrated in the regions where the land-atmosphere coupling has been shown to be the determining factor in the local climate. The EM simulations also showed a strengthened stationary wave pattern as compared to the standard mosaic, exemplified by the stronger Pacific-North America (PNA) pattern of variability.

A measure of the strength of the land-atmosphere coupling was described and adapted to evaluate the climate as simulated by EM versus M. We showed that the land-atmosphere coupling is stronger over certain regions when using EM. These regions coincide roughly with the ‘hot spots’ identified by Koster et al. 2004, as regions where the atmosphere is strongly influenced by the land surface. Based on the comparisons between EM and M, we conclude that a GCM with the EM technique has the potential to improve our ability to simulate the exchanges between the land surface and the atmosphere as well as improve predictions of the influence of anthropogenic land use change on the climate.

**References**


Figure 1: (a) Time series of global surface temperature as reconstructed by IPCC models. (b) Same as (a) except that the simulations include human impacts only, (c) same as (a) except that simulations include human and natural forcing.
Figure 2: (a) South American 8 km resolution Land Use Map from the data of Defries, et al. (2001), based on AVHRR. The different colors represent different vegetation types, and illustrate continental scale heterogeneities. (b) Land Use Map of Rehoboth Beach, Delaware from the Delaware department of Recreation, illustrating small-scale heterogeneities.
Figure 3: Schematic of the mosaic technique. The jagged shape of the tiles signifies an arbitrary shape, and the juxtaposition of tiles is not relevant. From: Molod, Salmun and Waugh (2003).
Figure 4: Temperature profiles from radiosonde data as a function of pressure thickness above the surface. Data are from different stations in the BOREAS field campaign. (a) June 11, 1994, 5Z profiles from two stations in the Northern Study Area, WTH and YYQ. (b) Same as (a) except at 17Z (mid-day), (c) same as (a) except 23Z (afternoon). (d) July 20, 1994 11Z profiles from KEY and WIQ, (e) Sept 15, 17Z profiles from WTH and TPA, (f) Sept 19 11Z profiles from WTH and TPA. Estimates of the level of the blending height are indicated in each panel.
Figure 5: Typical turbulent flux profiles for the tiles in a grid box. (a) Heat and moisture flux, and b) u-momentum and v-momentum flux for typical values of $V_s$. The units of the horizontal axis indicate the value of the appropriate turbulent flux, where heat flux is in kg m$^{-2}$s$^{-1}$ K, moisture flux is in kg m$^{-2}$s$^{-1}$ dg kg$^{-1}$, and momentum flux is in kg m$^{-1}$s$^{-2}$. The vertical axis is model $\sigma$ level, which can be multiplied by 1000 to approximate pressure in milibars (mb). From: Molod, Salmun and Waugh (2003).
Figure 6: June-July-August averaged Model Blending Height and Planetary Boundary Layer depth. (a) Model Blending Height in mb, (b) Planetary Boundary Layer depth, also in mb, and (c) ratio of MBH to PBL (dimensionless). The contour levels are indicated in the color bar to the right of each panel. From: Molod, Salmun and Waugh (2003).
Figure 7: (a) DJF Ensemble mean EM-M Sea Level Pressure difference in mb. Contour interval is 1 mb, dark grey shading is for differences less than -1 mb, light grey shading for differences greater than 1 mb. (b) DJF Ensemble mean EM-M 300 mb Eddy Height difference in m. Contour interval is 10 m, dark grey shading is for differences less than -10 m, light grey shading for differences greater than 10 m. (c) EM Sea Level Pressure in mb. Contour interval is 4 mb, dark grey shading is for pressures below 988 mb, light grey shading for pressures above 1020 mb. d) Ensemble mean 300 mb eddy height m. Contour-60 m, light grey shading indicates heights greater than 60 m. From: Molod, Salmun and Waugh (2004).
Figure 8: (a) EM-M difference of $(1-\Omega_E)$ and (b) EM-M difference of $(1-\Omega_P)$, measures of the strength of the land-atmosphere coupling. Large positive values, such as seen in the central U.S., indicate a stronger coupling in EM.