Roni Avissar

Abstract: This study documents significant atmospheric effects over the continental U.S. caused by human modification of the landscape. Using observations and an atmospheric model, we show here for the first time that diurnal, thermally-induced circulations occur during summer over a 250 x 250 km region in Oklahoma and Kansas. Furthermore, we show that the driving force behind these circulations is the landscape heterogeneity resulting from differential land use patterns, that such atmospheric phenomena are characteristic of surfaces with this type of heterogeneity and not limited to infrequent days when unusual wind or other meteorological conditions prevail, and that the net effect of these motions is significant, not only locally, but also at the regional and global scales.

1. Introduction

Parallel questions currently face those who study climate and climate change: in what ways are humans impacting the climate system, and do scientists and policy makers have the ability to predict future change resulting from this impact? One of the areas in which our understanding of both questions is most limited is that of the effect, on atmospheric processes, of human-induced changes of the landscape.

Variability in land properties occurs at a continuum of spatial scales. As deforestation, shifts in agriculture, and urbanization rise dramatically, natural vegetation patterns are modified, resulting in a complex and heterogeneous patchwork of surface characteristics unique to a given region. The interaction between these landscapes and the overlying atmosphere has a powerful effect on the way heat, moisture, momentum, dust, and pollutants move upward from the surface into the atmosphere as a whole.

The fluxes of heat and moisture from the surface into the adjacent atmosphere are fundamental driving forces of atmospheric motions in the planetary boundary layer (PBL) (Stull 1988), and the response of the atmosphere to this forcing depends strongly on the scale of the landscape variability which determines the distribution of these fluxes. For a domain characterized by horizontal variability in surface sensible heat (SH) and latent heat (LH) fluxes, theory suggests that there are two different atmospheric responses depending on the scale of the variability (Vidale et al. 1997; Roy and Avissar 2000; Avissar and Schmidt 1998; Dalu et al. 1996; Chen and Avissar 1994a; Wang et al. 1998). For landscape heterogeneity with typical scales on the order of 5 km or less (such a land-surface pattern might therefore "look" homogeneous at much larger scales), the atmospheric response consists solely of chaotic motions (turbulence) with eddies that have a scale on the order of the PBL depth. For patterns which consist of areas of similar surface fluxes covering distances greater than this, up to about 100-200 km, it is suggested that the randomly distributed turbulent motions can organize into much larger systems, commonly referred to as mesoscale circulations, oriented according to the underlying landscape heterogeneity and at a similar spatial scale. Observations of the scope necessary to validate these ideas are sparse, but a few studies have identified possible signatures of landscape-induced mesoscale circulations in certain regions and for certain conditions (Rabin et al. 1990; Bougeault et al. 1991; Cutrim et al. 1995; Brown and Arnold 1998). Other observational studies, focusing on regions where such effects might be expected, have either found no evidence for them or found that any resulting circulations are very weak (Mahrt et al. 1994; Doran et al. 1995; Hubbe et al. 1997). Various investigations using numerical models have suggested that these circulations can be intense, may play an influential role in producing and organizing clouds and precipitation, and could have effects that are significant at scales much larger than simply the scale of the systems themselves (Pielke et al. 1991; Segal et al. 1988; Segal et al. 1989; Ookouchi et al. 1984; Mahfouf et al. 1987; Yan and Anthes 1988; Avissar and Pielke 1989; Pinty et al. 1989; Chen and Avissar 1994b; Lynn et al. 1995a; Seth and Giorgi 1996; Avissar and Liu 1996). These earlier model studies have tended to report on the results from highly idealized simulations, thus making it difficult to assess their relevance for insights into real-world processes. More recent work, however, has moved in the direction of better observational constraint of surface characteristics, the incorporation of realistic, synoptic-scale meteorology into the numerical integrations, and the use of more complex modeling systems run at higher spatial resolutions (Vidale et al. 1997; Zhong and Doran 1997,1998). These more recent and realistic treatments of the problem have only sharpened the debate regarding the importance of mesoscale landscape effects in the contexts of climate and climate prediction. The results documented here are one of the first demonstrations that landscape heterogeneities produce mesoscale atmospheric effects of a scope and significance that is not limited to small areas and rare conditions.

2. Methods and Data

Our basic methodology is to perform simulations using a mesoscale model, namely, the Colorado State University (CSU) Regional Atmospheric Modeling System (RAMS) (Pielke et al. 1992), of several case-study days observed at the Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) Cloud and Radiation Testbed (CART) in Oklahoma and Kansas (Stokes and Schwartz 1994) as part of the Global Water and Energy Cycle Experiment (GEWEX) Continental Scale International Project (GCIP) Enhanced Seasonal Observing Period for 1995 (ESOP-95) (Coughlan and Avissar 1996). Because of a lack of suitable observations of the complete atmospheric effects of mesoscale landscape heterogeneity, the only reasonable tool for a comprehensive investigation of this phenomenon is a numerical model.

The ARM/CART covers approximately 300 by 350 km in Oklahoma and Kansas, and this heavily-instrumented site provides frequent measurements of surface SH and LH fluxes, as well as other meteorological quantities, throughout the year (Stokes and Schwartz 1994). The human footprint on this region is readily apparent. During summer, large differences in land use across this site, particularly the contrast between already-harvested winter wheat in the central portion of the domain and actively growing vegetation elsewhere, result in significant gradients in surface fluxes (Fig. 1a) (Zhong and Doran 1998). SH flux contrasts of approximately 200 W m^{-2} exist between areas (patches) with characteristic scales of 20-100 km. The relative roles of evaporation from moist or vegetated surfaces and direct heating of dry surfaces lead to compensation between surface SH and LH flux. As a result, the LH flux field for this domain (not shown) exhibits a pattern that is essentially the inverse of the SH flux shown in Fig. 1a, with similar magnitude. While topography is often an additional and crucial forcing factor for PBL flow, the terrain of the ARM/CART site is relatively flat (Fig. 1b), and removing the influence of topography does not qualitatively affect our simulation results.

The RAMS solves the full, nonlinear equations of motion for the atmosphere, and it includes parameterizations for turbulent fluxes, radiation, convection, and cloud microphysics, as well as a comprehensive soil and vegetation model (see Pielke et al. 1992 for more details). For this study, we take advantage of the nested (telescoping) grid capability of RAMS. We use three such nested grids centered on the CART site, with the innermost (Grid 3) covering 252x252 km with a horizontal resolution of 2 km, the next coarser grid (Grid 2) covering 720x720 km with a resolution of 8 km, and the coarsest grid (Grid 1) encompassing an area 1600x1600 km with a resolution of 32 km (Fig. 1b). This nesting of fine grids within coarse ones is a compromise, forced by existing limitations on computing technology, between simulating the region of interest with a high enough resolution to capture the important scales and processes and simulating a large enough domain to correctly represent all the relevant meteorological forcing. There are 38 vertical levels from the surface to approximately 22 km, with half of these located in the first 3 km. The horizontal area of Grid 3 is comparable to that of a typical Global Climate Model (GCM) grid element, and thus we expect averages over this domain to represent the effects that a GCM should be required to simulate with reasonable fidelity.

We selected six case-study days: July 6, 7, 10, 12, 13, and 21, 1995. These particular days in July were chosen because the CART site was relatively unobscured by any extensive cloud systems of obvious large-scale (non-local) origin, thus making possible the identification of any signature of the local land-surface. By this choice we do not mean to imply that landscape-induced mesoscale effects would necessarily be small on days when larger-scale cloud systems are present over the ARM/CART, simply that their effects on clouds and precipitation would be masked. Further simulations (not shown here) based on observations from other days when overall conditions were much more disturbed still produced strong mesoscale circulations and effects.

The pattern of surface fluxes on all days is similar, though not identical, to the July 13 case shown in Fig. 1a and evolves slowly throughout the month (Zhong and Doran 1998). Each simulation covers 12 hours, from 6 a.m. to 6 p.m. local time (1200 to 2400 UTC). We constrain the simulation for each day with two types of observationally-derived data. For the surface boundary condition, we use gridded surface SH and LH fluxes calculated by using a land-surface model to integrate measured fluxes from a series of locations within the ARM/CART (Doran et al. 1998). These observationally-derived fluxes are specified on Grid 3 and for the portion of Grid 2 which covers the Grid 3 domain. For the remainder of Grid 2, and for Grid 1, the surface fluxes are calculated using the RAMS soil and vegetation model. We initialize each simulation using geopotential height, winds, temperature, and relative humidity from the National Center for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis project (Kalnay et al. 1996). The lateral boundary regions of Grid 1 are nudged during each model timestep toward the reanalyses, which are updated every six hours.

To aid in the interpretation of the model results, we use 2-km resolution visible imagery from the Geostationary Operational Environmental Satellite (GOES-8) and composited rain-gauge/WSR-88D radar rainfall measurements from the Arkansas-Red

River Basin Forecast Center (ABRFC) collected as part of ESOP-95 (more information about these GCIP data sets can be found at www.abrfc.noaa.gov and www.ogp.noaa.gov/mpe/gcip/index.htm). The results we show here illustrate how model simulations, coupled with complementary observations, provide a powerful method for answering complex questions of this type.

3. Results

Fig. 2 shows the simulated mid-PBL vertical velocity (*w*) field at two times, noon and 3:30 p.m. local time, for the July 13 case. At noon (Fig. 2a), relatively weak upward motion of a few cm s^{-1} is located over patches of large SH flux (recall Fig. 1a) with scales on the order of a few tens of km. Over the course of the afternoon, a series of coherent circulations with strong upward motion on the order of 1-2 m s^{-1} gradually evolve, with convergence of air flowing from high-LH (cooler) to high-SH (warmer) areas (Fig. 2b). Over some locations, this circulation penetrates from the surface to a height of more than 3 km. While the model does simulate clouds and precipitation for this case, the model cloud response is much weaker than that of the real atmosphere because of the limitations imposed by the 2 km horizontal resolution. The RAMS microphysics package is able, in principle, to produce sufficient cloud for a given dynamical forcing, but the 2-km horizontal grid spacing of Grid 3 is too coarse to fully resolve cloud-scale motions without a suitable convection scheme. The RAMS convection scheme, however, is only appropriate for significantly larger grid spacing. Therefore, to illustrate the correspondence between the model and the observations more clearly, we only show simulated *w* here. This is a reasonable choice, since updrafts in the PBL provide the direct forcing for convective clouds and precipitation, and therefore, if the model is correctly reproducing the actual dynamics over the ARM/CART, we expect to find any observed clouds and precipitation at the locations where the model produces the strong positive *w* values.

To highlight the impact of the surface variability, rather than simply the mean surface conditions, we performed an additional simulation of the July 13 case, but with the realistic surface SH and LH fluxes replaced with their mean values over the Grid 3 domain, thus eliminating any mesoscale heterogeneity. [We provide some variability in surface fluxes to aid in the generation of convection by imposing a random perturbation at each grid cell of amplitude 10 W m^{-2}. Thus, while heterogeneity is present, it is at the 2-km horizontal scale of a single grid element]. As seen in Fig. 2c, his mean flux case, rather than developing any organized circulation, has a very weak vertical flow field with peak *w* values of a few cm s^{-1}. A finer contour interval would show that the *w* field is characterized by essentially random updrafts and downdrafts at a small horizontal and vertical scale. This contrast in atmospheric response between homogeneous and heterogeneous surfaces is in agreement with some previous theoretical studies, as noted above (Vidale et al. 1997; Roy and Avissar 2000; Avissar and Schmidt 1998; Dalu et al. 1996).

That the simulated circulation pattern forced by the landscape heterogeneity is undoubtedly a real-world feature is demonstrated by the excellent correspondence between the simulated *w* field and the satellite cloud image (Fig. 2d), and composited rainfall measurements (Fig. 3). The clouds and precipitation are oriented along the linear, N-S convergence lines of strong upward motion. The model is able to capture not only a snapshot agreement but correctly reproduces the evolution in time of the convergence field as diagnosed by successive hourly rainfall composites (Fig. 3); in the observations, the broadly V-shaped precipitation zone moves northward and dissipates over the course of the late afternoon, and the model correctly captures this behavior.

The satellite observations indicate that two other of the remaining five case study days (July 6, 21) also have clouds, and a comparison between the *w* field and the satellite imagery for these days also shows good agreement between the location and orientation of the clouds and the patterns of upward motion (Fig. 4). This agreement provides a high level of confidence in the ability of the model to correctly represent the essential physics of the organization of the atmospheric flow by the landscape pattern and therefore allows us to use the model results to illuminate details, over the full 65,000 km^2 Grid 3 domain, of the evolution, intensity, and transport properties of the circulations for which there are no suitable observations.

That the cloud pattern is different among these 3 days, and that indeed there are no clouds on the remaining three (July 7, 10, 12; not shown), illustrates another fundamental point: a basically similar pattern of surface forcing can interact with different ambient meteorology to produce different atmospheric responses. Recalling Fig. 1a, it is clear that there are many different "choices" of surface flux patch patterns with, for example, an approximately 50-km patch scale. The induced circulations on each of the six days are comparably strong, as summarized in Table 1 (the peak *w* values are all large, on the order of 1 m s^{-1} or greater), and in each case they cover a large fraction of the domain and are properly classifiable as mesoscale. Nevertheless, they take on different orientations and peak locations in response to differences in large-scale meteorology.

Additionally, for a given circulation pattern and intensity, whether and how much cloud formation occurs depends strongly on the ambient moisture profile. The contrast between the July 12 and 21 cases is instructive. While the peak *w* value for July 21 is the lowest of any of the days (Table 1), and the value for July 12 the highest (i.e., the dynamical driving force for cloud formation is the strongest), convective clouds form on the 21st and not the 12th simply because the 12th is a much drier day, and therefore even a very strong circulation is unable to produce clouds. It is important to keep this issue in mind when considering global impacts of land-use changes on clouds and precipitation. For example, we point out that overall conditions are relatively dry over OK/KS in July compared to some other regions of the globe with substantial observed mesoscale surface heterogeneity (e.g., Amazonia), and much stronger cloud and precipitation responses might be expected elsewhere given similar intensities of landscape/atmosphere effects.

To demonstrate the important role these circulations play in transporting heat and moisture vertically, we calculated the three-dimensional SH and LH fluxes by both the resolved motions in the model, or mesoscale fluxes (i.e., vertical SH and LH fluxes due to motions with scales larger than the 2-km model grid spacing), and the subgrid-scale, parameterized turbulent fluxes. These mesoscale fluxes are one measure of the intensity of the landscape-induced circulations, and the peak domain-averaged mesoscale LH fluxes for each case day are summarized in Table 1.

Fig. 5 shows mesoscale, turbulent, and total (mesoscale + turbulent) LH flux, averaged over the Grid 3 domain at 4 p.m. local time (the peak mesoscale fluxes occur in the late afternoon, approximately three hours later than the peak surface SH flux), for both the mean and variable surface flux July 13 simulations (recall Fig. 2b,c). We only show mesoscale LH flux because it tends to be larger than SH flux for the cases considered here. The key result is that mesoscale landscape heterogeneity significantly alters the magnitude and shape of the vertical flux profile for given mean levels of surface fluxes and given meteorological conditions. Shifting some of the vertical transport from small (turbulent) scales to larger (mesoscale) scales produces substantially more total LH transport (on the order of 200 W m^{-2} in late afternoon) in the upper PBL, with correspondingly significant implications for production of clouds and precipitation as well as for the energy available to drive the overall atmospheric circulation. The physical explanation for this process is that the mesoscale circulations result in enhanced *vertical* moisture transport because of their ability to transport moisture *horizontally*, from areas where surface LH is large, but the PBL is consequently shallow, to areas where surface LH flux is small but surface SH flux is large, and where therefore the PBL is deep and the moisture can penetrate higher. As a result, over portions of the domain, the mesoscale fluxes in the mid- and upper PBL are much larger than the turbulent fluxes (often reaching into the thousands of W m^{-2}), and this is reflected in the domain averages. The second, more subtle, factor is that, in addition to this direct transport by the mesoscale circulation, the changes in air mass movement resulting from the mesoscale flow additionally change the turbulent tranport (see also Vidale et al. 1997), as illustrated in Fig 5b by the larger values of turbulent LH flux above 1700 m for the heterogeneous surface case. All scales of motion are therefore influenced by the surface heterogeneity.

Some model and analytical studies have indicated that, except under very weak large-scale background wind conditions (e.g., on the order of 2-3 m s^{-1} or less), landscape-induced mesoscale flows are inhibited, and the corresponding mesoscale fluxes are therefore also very weak (Avissar and Schmidt 1998; Wang et al. 1996; Liu et al. 1999); hence the most common argument against their climatological importance, i.e., the percentage of time these landscape heterogeneity effects are significant is expected to be small (Zhong and Doran 1997, 1998). This view has been challenged recently (Vidale et al. 1997; Wang et al. 1998), and our results demonstrate for the first time for a range of cases that increasing large-scale wind does not necessarily inhibit these mesoscale circulations.

Fig. 6 shows vertical-longitude cross-sections of simulated *u* and *v* at 8 a.m., noon, and 4 p.m. local time along 36.8 N latitude for the July 13 case. This figure illustrates two major points. First, the wind velocities are not particularly small, approaching 10 m s^{-1} or greater in the upper portions of the PBL at various points during the day. Second, one can clearly see the evolution of the mesoscale circulations by 4 p.m., visible here primarily as zonal convergence and divergence cells in the PBL, and one can also clearly see that this evolution is essentially decoupled from the evolution of the large-scale background flow in the free troposphere. This behavior is typical for all the other days studied, all with essentially comparable mean winds (Fig. 7), and we find no particular relationship between overall synoptic-scale wind speed and the viability or intensity of the landscape-forced mesoscale circulations. Indeed, the day with the highest mean winds in the lowest part of the atmosphere (July 7; Fig. 7) also has one of the highest mean mesoscale LH fluxes (Table 1).

Instead of having an inhibiting effect, the synoptic-scale wind plays a key role in orienting, steering, and advecting the landscape-induced mesoscale circulations. The differences in mesoscale circulation patterns among the different case days, and the northward movement of the precipitation system on July 13 (recall Fig. 3) result from this interaction between the mesoscale and the synoptic-scale. This finding has important implications for our understanding of the wider area and longer time impacts of landscape heterogeneity on the atmosphere. First, since these mesoscale effects can occur under conditions of moderate and strong synoptic-scale winds, they probably are important more often (climatologically) than previously assumed. Second, the advection of mesoscale cloud and precipitation systems away from the local area whose surface conditions generated the circulation complicates the problem of determining radiative and hydrological feedbacks between the land-surface and the atmosphere.

4. Conclusions

The results shown here demonstrate that, by changing the scale and properties of the naturally occurring land-surface elements, human influences can significantly affect local weather and climate, at least in the short term. Since both natural and human-modified heterogeneous landscapes are ubiquitous around the globe (e.g., snow/soil, land/water, pasture/forest, city/country), such atmospheric effects must influence global climate today. However, the extent of this influence, and the longer-term impacts of accelerating land use changes on future regional and global climate, are currently unknown. At present, our only tools for accurate climate prediction are GCMs, and the efficacy of such models for adequate risk assessment depends most importantly on how well they simulate the various climate processes, rather than any particular climate state. A good representation of present-day climate is worthless from a prediction standpoint without confidence in the model's ability to respond realistically to evolving climatic boundary conditions. Since most GCMs used for climate prediction can only explicitly resolve scales of hundreds of km, much of the challenge of model improvement is in finding appropriate parameterizations of processes which occur at scales smaller than this. For heterogeneous land surfaces, both the underlying heterogeneity and the atmospheric response, in the form of turbulent and mesoscale motions, are subgrid-scale processes in GCMs, and their effects, if significant, must be accounted for. While current GCMs employ parameterizations for subgrid-scale flow, these parameterizations are based solely on observations and theories of turbulence processes over homogeneous landscapes, and consequently, important mesoscale effects like the ones described in this study are completely ignored. Thus, we are unable to correctly account for the effects of landscape heterogeneity in our simulations of present climate, and we are ill-equipped to predict the full consequences of future land-use changes. To address this, a few preliminary GCM parameterizations of mesoscale fluxes have already been proposed (Liu et al. 1999; Avissar and Chen 1993; Arola 1999; Zeng and Pielke 1995; Lynn et al. 1995b), and the results shown here should stimulate further efforts to implement realistic representations of these processes in state-of-the-art climate models with the aim of addressing current problems in, for example, simulating continental shallow convection (Berbery et al. 1997), correctly representing albedo and hydrological feedbacks between the surface and the atmosphere (Berbery et al. 1997; Betts et al. 1996), and improving simulations of future climate change.

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Case	Max w (m s^{-1})	Max Domain-Mean Mesoscale LH Flux (W m^{-2})
July 6	1.59	133.0
July 7	2.61	167.6
July 10	1.64	111.3
July 12	2.88	66.0
July 13	2.77	177.5
July 21	0.91	56.8

Table 1. For each of the six case study simulations, shown are maximum, mid-PBL updraft velocity (w) for the given day and maximum domain-averaged vertical mesoscale LH flux. The domain averages are over the entire Grid 3 area. For all cases, the maximum w and mesoscale LH flux occurred in the mid- to upper PBL during the late afternoon.

Figure Captions

1. (a) Grid 3 surface SH flux (W m^{-2}) at 1 p.m. local time (1900 UTC) on July 13, 1995. (b) The full RAMS simulation domain (Grid 1) showing surface topographic height (m) and the locations of the nested Grids 2 and 3.

2. (a) RAMS simulated w (m s^{-1}) at 1100 m altitude at noon local time (1800 UTC) on July 13, 1995. (b) RAMS simulated w (m s^{-1}) at 1100 m altitude at 3:30 p.m. local time (2130 UTC) on July 13, 1995. (c) RAMS simulated w (m s-1) at 1100 altitude at 3:30 p.m. local time (2130 UTC) on July 13, 1995. For this simulation, the Grid 3 surface SH and LH fluxes were replaced by their domain-averaged values. (d) GOES-8 satellite visible image at 3:15 p.m. local time (2115 UTC) on July 13, 1995.

3. (a) RAMS simulated w (m s^{-1}) at 1100 m altitude at 3 p.m. local time (2100 UTC) on July 13, 1995. (b) ABRFC accumulated precipitation composite for 3 p.m. to 4 p.m. local time (2100-2200 UTC) on July 13, 1995. (c) RAMS simulated w (m s^{-1}) at 1100 m altitude at 4 p.m. local time (2200 UTC) on July 13, 1995. (d) ABRFC accumulated precipitation composite for 4 p.m. to 5 p.m. local time (2200-2300 UTC) on July 13, 1995. (e) RAMS simulated w (m s^{-1}) at 1100 m altitude at 5 p.m. local time (2300 UTC) on July 13, 1995. (f) ABRFC accumulated precipitation composite for 5 p.m. to 6 p.m. local time (2300-0000 UTC) on July 13, 1995. (g) RAMS simulated w (m s^{-1}) at 1100 m altitude at 6 p.m. local time (0000 UTC) on July 13, 1995. (h) ABRFC accumulated precipitation composite for 6 p.m. to 7 p.m. local time (0000-0100 UTC) on July 13, 1995.

4. (a) RAMS simulated w (m s^{-1}) at 1100 m altitude at 3:30 p.m. local time (2130 UTC) on July 6, 1995. (b) RAMS simulated w (m s^{-1}) at 1100 m altitude at 3:30 p.m. local time (2130 UTC) on July 21, 1995. (c) GOES-8 satellite visible image at 3:15 p.m. local time (2115 UTC) on July 6, 1995. (d) GOES-8 satellite visible image at 3:15 p.m. local time (2115 UTC) on July 21, 1995.

5. (a) Grid 3 averaged mesoscale LH flux (W m^{-2}) at 4 p.m. local time (2200 UTC) on July 13, 1995. The solid curve is for the realistic surface flux run, and the dashed curve is for the mean surface flux run. (b) Same as (a) but for Grid 3 averaged turbulent LH flux. (c) Same as (a) but for Grid 3 averaged total (mesoscale + turbulent) LH flux. (d) The difference in total LH flux between the runs with mean and realistic surface flux.

6. (a) Altitude-longitude cross-section of RAMS simulated u (m s^{-1}) at 8 a.m. local time (1400 UTC) for July 13, 1995. (b) Altitude-longitude cross-section of RAMS simulated v (m s^{-1}) at 8 a.m. local time (1400 UTC) for July 13, 1995. (c) Same as (a) but for noon local time (1800 UTC). (d) Same as (b) but for noon local time (1800 UTC). (e) Same as (a) but for 4 p.m. local time (2200 UTC). (f) Same as (b) but for 4 p.m. local time (2200 UTC).

7. (a) Grid 3 averaged wind speed (m s^{-1}) at 8 a.m. local time (1400 UTC) for each of the six case study days. (b) Grid 3 averaged wind speed (m s^{-1}) at noon local time (1800

UTC) for each of the six case study days. We show early morning and noon wind here because it is illustrative of the large-scale wind conditions existing prior to the development of any landscape-generated mesoscale circulations. The winds later in the day include the effects of these circulations, and thus cannot properly be considered to reflect the ambient synoptic-scale environment in which such circulations must form.