Simulated and Observed Surface Energy Fluxes and Resulting Playa Breezes during the MATERHORN Field Campaigns

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ABSTRACT

Weather Research and Forecasting (WRF) Model simulations of the autumn 2012 and spring 2013 Mountain Terrain Atmospheric Modeling and Observations Program (MATERHORN) field campaigns are validated against observations of components of the surface energy balance (SEB) collected over contrasting desert-shrub and playa land surfaces of the Great Salt Lake Desert in northwestern Utah. Over the desert shrub, a large underprediction of sensible heat flux and an overprediction of ground heat flux occurred during the autumn campaign when the model-analyzed soil moisture was considerably higher than the measured soil moisture. Simulations that incorporate in situ measurements of soil moisture into the land surface analyses and use a modified parameterization for soil thermal conductivity greatly reduce these errors over the desert shrub but exacerbate the overprediction of latent heat flux over the playa. The Noah land surface model coupled to WRF does not capture the many unusual playa land surface processes, and simulations that incorporate satellite-derived albedo and reduce the saturation vapor pressure over the playa only marginally improve the forecasts of the SEB components. Nevertheless, the forecast of the 2-m temperature difference between the playa and desert shrub improves, which increases the strength of the daytime off-playa breeze. The stronger off-playa breeze, however, does not substantially reduce the mean absolute errors in overall 10-m wind speed and direction. This work highlights some deficiencies of the Noah land surface model over two common arid land surfaces and demonstrates the importance of accurate land surface analyses over a dryland region.

1. Introduction

The variability of regional land surface characteristics in mesoscale numerical weather prediction (NWP) models has a potentially strong influence on near-surface forecasts. Some sources of land surface variability, such as coastlines and topographic features, are easily represented in NWP models, but other more subtle land surface characteristics (e.g., albedo, emissivity, roughness length, soil porosity, soil texture, and soil moisture) are more difficult to specify and parameterize and can exhibit significant spatial variability and—for characteristics like soil moisture, albedo, and emissivity—temporal variability (e.g., Chen and Dudhia 2001; Ek et al. 2003; Malek 2003). Land surface characteristics affect near-surface forecasts of temperature, moisture, and momentum by changing the relative importance of components of the surface energy balance (SEB), with the net radiation $R_n$ partitioned into surface sensible heat flux $H$, latent heat flux $LE$, and ground heat flux $G$:

$$R_n = H + LE + G,$$

where downwelling $R_n$ and $G$ and upwelling $H$ and $LE$ are defined as positive. Near-surface forecasts are affected not only by the local partitioning among these components but also by regional differences, which can drive mesoscale circulations (e.g., Segal and Arritt 1992) and influence cloud development, precipitation, and atmospheric stability (Stull 1988).
The partitioning of $R_n$ into $G$, $H$, and $LE$ is strongly influenced by the near-surface (0–10 cm) soil moisture and, in some regions, root-zone soil moisture (e.g., Ookouchi et al. 1984; Banta and Gannon 1995; Sun and Bosilovich 1996). Soil moisture affects 1) the ratio of $H$ to LE (i.e., the Bowen ratio; Bowen 1926) through evapotranspiration, 2) $G$ because water has a higher thermal conductivity than the air it replaces (Cosenza et al. 2003), and 3) $R_n$ by altering the surface albedo (Idso et al. 1975). Higher soil moisture typically causes a greater percentage of $R_n$ to be partitioned into LE and $G$ and a lesser percentage to be partitioned into $H$. As a result, the diurnal temperature cycle has a lower amplitude over moist soil than over dry soil under otherwise identical conditions.

Comparison of forecasts and observations of the SEB components offers significant potential to identify land surface deficiencies in NWP models (e.g., Hu et al. 2010; Steeneveld et al. 2008; Svensson et al. 2011; Aas et al. 2015). For example, Aas et al. (2015) compared simulated and observed SEB components over the Svalbard Archipelago in the Arctic Ocean north of Europe, identifying overpredictions of $R_n$ and the Bowen ratio, which they attributed to an underprediction of cloud cover and soil moisture, respectively. A major source of uncertainty with such validation studies, however, is that SEB observations do not close (i.e., $R_n > H + LE + G$) because of the presence of a residual error term (Foken 2008).

This study focuses on the SEB components and associated thermally driven playa breeze in the Dugway Proving Ground (DPG) region of the Great Salt Lake Desert of northern Utah. This region of complex terrain is characterized by two distinct land surfaces: playa and desert shrub [Fig. 1; Weather Research and Forecasting (WRF) Model shrubland is synonymous with desert shrub], which are found in dryland regions around the world (Warner 2004). The playa is a flat, salt-encrusted, vegetation-free, clay surface, and the top of the underlying water table is close to the surface. The adjacent desert shrub is sparsely vegetated, with underlying silt loam and loam soils. The playa has a higher albedo, higher soil thermal conductivity, less vegetation, and higher soil moisture than the desert shrub, leading to differences in the SEB and to temperature gradients between the two surfaces (Rife et al. 2002). In general, the playa is cooler during the day and warmer at night than the surrounding desert-shrub area. Malek (2003) investigated the playa surface and its SEB and discovered distinctive characteristics in comparison with nonplaya land surfaces, such as an early-morning maximum in $LE$, rehydration of the topsoil at night, a shallow water table 0–60 cm below the surface, and a nearly order-of-magnitude difference between the potential and actual mean daily evaporation.

The diurnally modulating temperature gradient between the playa and desert shrub can lead to an off-playa breeze during the day and an on-playa breeze at night (Physick and Tapper 1990; Rife et al. 2002). Similar mesoscale circulations occur elsewhere over soil moisture and vegetation gradients (e.g., Ookouchi et al. 1984; Avissar and Pielke 1989; Fast and McCorcle 1991; Segal and Arritt 1992). During quiescent largescale conditions, observational and numerical modeling studies have found that the daytime off-playa
breeze (sometimes called a salt breeze) extends to as high as $\sim 1000$ m AGL and has 10-m wind speeds of 3–4 m s$^{-1}$ (e.g., Davis et al. 1999; Rife et al. 2002; Knievel et al. 2007). These numerical studies required artificial forcing to produce a realistic off-playa breeze, however. To be specific, Davis et al. (1999) used a soil-dependent thermal inertia instead of a land surface model (LSM), and Rife et al. (2002) increased soil temperature initial conditions by as much as 5°C over the playa.

This paper validates forecasts from the WRF Model using observations of SEB components collected during the Mountain Terrain Atmospheric Modeling and Observations Program (MATERHORN) field campaigns (Fernando et al. 2015). Massey et al. (2014, 2016) incorporated soil moisture measurements and a modified soil thermal conductivity parameterization in the Noah LSM over this region and noted improved temperature forecasts. This study extends that work by examining the accuracy of SEB component forecasts over desert shrub and playa, highlighting remaining WRF and Noah LSM deficiencies over these two common dryland land surfaces. We further examine the accuracy of the 10-m wind forecasts.

2. Data and methods

a. Surface energy balance stations

This study uses data collected during the MATERHORN field campaigns at DPG from 25 September to 25 October 2012 (MATERHORN-Fall) and from 1 May to 31 May 2013 (MATERHORN-Spring; Fernando et al. 2015). MATERHORN-Fall was characterized by predominantly quiescent, fair-weather conditions with only 8.7 mm of precipitation at the DPG National Weather Service cooperative observer Program site. MATERHORN-Spring was characterized by stronger synoptic forcing and 19.4 mm of precipitation at that site. Our analysis concentrates on 20 days during MATERHORN-Fall (1–20 October 2012) and 30 days during MATERHORN-Spring (2–31 May 2013) with available surface flux observations from at least one extended flux site (EFS) located over the playa (EFS-Playa) or desert shrub (EFS-DS) (see Fig. 1 for locations and Fig. 2 for photographs). During MATERHORN-Fall, EFS-Playa and EFS-DS recorded at least 12 h of data for 18 and 14 nonconsecutive days, respectively, and both sites recorded data simultaneously during 12 nonconsecutive days (Table 1). Therefore, the validation considers different MATERHORN-Fall days at EFS-DS and EFS-Playa. During MATERHORN-Spring, data are available continuously at both sites.

Given the large effect that clouds have on the SEB components, following Massey et al. (2016), we delineate mostly clear and mostly cloudy days during the MATERHORN periods using atmospheric transmittance, defined as

$$\text{Transmittance} = \frac{\sum_{t=0000LST}^{2330LST} \ SW_{\text{sfc}}(t)}{\sum_{t=0000LST}^{2330LST} \ SW_{\text{toa}}(t)},$$

with $\ SW_{\text{sfc}}(t)$ being the observed downwelling short-wave (SW) radiation at the surface at time $t$ and $\ SW_{\text{toa}}(t)$ being the theoretical downwelling top-of-atmosphere SW calculated from

$$\ SW_{\text{toa}}(t) = S_0 (a/r)^2 \sin(\phi),$$
where $S_0$ is the solar constant (approximated to be 1370 W m$^{-2}$), $a$ is the annual mean distance between the sun and Earth, $r$ is the daily mean distance, and $\phi$ is the solar elevation angle calculated following Reda and Andreas (2004). Transmittance was calculated for each day at EFS-DS and EFS-Playa, and days with a mean transmittance of $0.65$ [see Massey et al. (2016) for the rationale] were designated as “mostly clear,” with all other days being designated as “mostly cloudy.” Only three days were defined as mostly cloudy during MATERHORN-Fall (Table 1), and 11 days were defined as mostly cloudy during MATERHORN-Spring.

### 1) MEASUREMENT OF SEB COMPONENTS

Sensible heat flux $H$ was calculated at 2 m AGL from sonic anemometer and sonic temperature measurements, and LE was calculated at 10 m from infrared gas analyzer measurements. The fluxes were calculated using 5-min averaging times and were quality controlled using the Utah Turbulence in Environmental Studies processing and analysis code (UTESpac; Jensen et al. 2016). The 2-m $H$ and 10-m LE were treated as proxies for surface fluxes. Jensen et al. (2016) revealed that 5-min flux averaging was sufficient to capture the vast majority of the turbulent flux, but uncertainty due to inherent instrumentation errors lowers confidence in the flux comparisons (Mauder and Foken 2006).

Ground heat flux $G$ was calculated as the sum of 1) the average heat flux from two heat flux plates (Hukseflux Thermal Sensors B.V. model HFP01SC) at a 5-cm depth separated horizontally by approximately 1 m and 2) the change in the heat storage in the 0–5-cm soil layer. The heat storage was calculated using measurements of the 5-cm thermal heat capacity from thermal property sensors (Hukseflux model TP01) and 1-, 2.5-, and 5-cm soil temperature from thermocouples (Omega Engineering, Inc.).

The individual SW and longwave (LW) radiation components of the surface radiation balance were measured with Kipp and Zonen B.V. up- and down-facing CMP21 pyranometers and CGR4 pyrgeometers, respectively, mounted at 2 m AGL on a sawhorse-type structure (Fig. 2). Net radiation $R_n$ was calculated by subtracting the outgoing from the incoming SW and LW components, which were all defined as positive.

### 2) RESIDUAL CORRECTION

We calculated a residual term $Res$ as

$$Res = R_n - H - LE - G.$$

Foken (2008) argued that the data quality and measurement accuracy of the instruments used in this study have improved sufficiently to rule out measurement quality as a major contributor to the magnitude of $Res$. He hypothesized that the eddy covariance turbulent measurements cannot capture the larger eddies and thus underestimate $H$ and LE. One approach to close the SEB, which we utilize, is to assume that the Bowen ratio of the measured small eddies is the same as the Bowen ratio of the larger eddies that are not measured. Therefore, we distributed $Res$ to $H$ and LE according to the Bowen ratio. This approach was used in previous studies (e.g., Twine et al. 2000) and was recommended by Foken (2008) as the best available method. Ruppert et al. (2006) found that the similarity of the Bowen ratio between large and small eddies varies by the size of the eddies and the time of day, limiting the utility of this approach.

Given the high Bowen ratios at both sites during MATERHORN-Fall and MATERHORN-Spring, the majority of $Res$ is transferred to $H$ (Fig. 3). For mostly clear days at EFS-DS for MATERHORN-Fall, $Res$ is approximately 18% of $R_n$ at night (i.e., when $R_n < 0$) and approximately 25% of $R_n$ during the day (i.e., when $R_n > 0$), and nearly all of $Res$ is transferred to $H$ (Fig. 3a). $Res$ is lower at EFS-Playa than at EFS-DS during MATERHORN-Fall, but nearly all of $Res$ is also transferred to $H$ (Fig. 3b). For mostly clear days at EFS-DS for MATERHORN-Spring, $Res$ is near 0 at night but as high as 30% during the afternoon, and approximately 80% of $Res$ gets transferred to $H$, with the remainder going to LE (Fig. 3c). At EFS-Playa, $Res$ is also

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near 0 at night, and approximately 85% of $R_n$ gets transferred to $H$ during the day (Fig. 3d).

b. Surface temperature and wind observations

Near-surface 2-m temperature and 10-m wind forecasts, which are diagnosed from the WRF-Model half-$\eta$ and skin-level fields using similarity theory, were validated against 2-m temperature and 10-m wind observations from 27 surface automated meteorological stations (SAMS) with a site elevation within 100 m of the corresponding WRF elevation. SAMS are located primarily in lowland areas in and around DPG (see Fig. 1 for locations), and observations represent 5- or 15-min averages depending on location. Although no formal quality control was performed, missing and obviously erroneous observations were removed. We also excluded observations that are less than 1 m s$^{-1}$ because such light winds are near the threshold velocity of most propeller anemometers. Bias errors (BE) and mean absolute errors (MAE) were calculated for 2-m temperature, 10-m wind speed, and 10-m wind direction. Wind speed BE and MAE were calculated using scalar wind speed differences, whereas wind direction MAE are calculated from $u$ and $v$ wind components.

c. WRF modeling

We ran three sets of 30-h WRF simulations during the MATERHORN periods that have SEB data available (i.e., 1–20 October 2012 and 2–31 May 2013). Simulations were initialized daily at 1800 UTC (1100 LST), but
only the 6–29.5-h forecasts valid at 0000–2330 UTC (1700–1630 LST) were evaluated so as to allow for a 6-h spinup period. The three sets of simulations are a control simulation (hereinafter “Control”), a simulation with modified soil moisture and a modified soil thermal conductivity parameterization (hereinafter “SM”), and a simulation similar to SM but with modified albedo and saturation vapor pressure over the playa (hereinafter “SM-Albedo”).

The domains, physics, land-use categories, and soil categories follow that of the operational WRF-based Four-Dimensional Weather (4DWX) system as run at DPG (4DWX-DPG; Liu et al. 2008). We used WRF, version 3.5.1, with 30-, 10-, and 3.3-km one-way nested domains centered over DPG and 36 half-η levels, with the lowest half-η level at −15 m AGL. The vertical spacing varied from ~30 m near the surface to ~1250 m in the upper troposphere and lower stratosphere. The physics packages included the Rapid Radiative Transfer Model longwave radiation parameterization (Mlawer et al. 1997), the Dudhia shortwave radiation parameterization (Dudhia 1989), the Noah LSM (Chen and Dudhia 2001), the Yonsei University PBL parameterization (Hong et al. 2006), explicit sixth-order numerical diffusion (Knievel et al. 2007), and the Kain–Fritsch cumulus parameterization (Kain 2004).

We used 0.5° Global Forecast System (GFS) analyses for initial atmospheric and land surface analyses, as well as lateral boundary conditions, as is done in the operational 4DWX-DPG system. Following 4DWX-DPG, we used updated land-cover, soil texture–class, and terrain-elevation datasets that are based on the 33-category National Land Cover Database (Fry et al. 2011), which has an additional playa land-cover and soil texture class. We also changed the Great Salt Lake (GSL) surface temperature to a climatological value (Crosman and Horel 2009) and reduced the saturation vapor pressure over the GSL following Steenburgh et al. (2000) to account for the effect of the GSL brine on surface evaporation.

Control simulations were based on the configuration above; as in Massey et al. (2014), the SM simulations adjusted the 5- and 25-cm soil moisture analyses using in situ soil moisture measurements and replaced the default soil thermal conductivity parameterization with that proposed by McCumber and Pielke (1981) for silt loam and sandy loam soils. In the SM 3.3-km domain, we applied the mean measured soil moisture from in situ stations over loam, sandy loam, silt loam, and silty clay loam to the geographical area defined by those soil texture classes. Observations of volumetric soil moisture, which align with the soil moisture analyses, come from the Texas A&M University North American Soil Moisture Database (NASMD; http://soilmoisture.tamu.edu), which harmonizes and quality controls several in situ soil moisture observing platforms. We considered only stations from the U.S. Department of Agriculture’s Soil Climate Analysis Network (SCAN; Schaefer et al. 2007) and from the global positioning system (GPS) network. SCAN stations measure soil moisture at depths of 5.1, 10.2, 20.3, 50.8, and 101.6 cm, but the Noah LSM is configured with depths centered at 5, 25, 70, and 150 cm, and therefore we considered only the 5.1- and 20.3-cm SCAN levels relative to the 5- and 25-cm Noah LSM levels, respectively, for initialization and validation. The GPS network measures 5-cm soil moisture from L-band radiation (1.57742 and 1.22760 GHz) from GPS satellites reflected off the land surface (Larson et al. 2008). Playa is the only soil texture class that is not represented by a NASMD station, but manual observations were available near EFS-Playa during nine MATERHORN-Spring intensive observing periods (IOPs) and three MATERHORN-Fall IOPs at 5 and 25 cm. Therefore, we applied the playa soil moisture measured during the IOP nearest the initialization time to the playa soil texture class. Soil moisture observations may fall outside the allowable soil moisture range in the Noah LSM but are corrected within the LSM on the first model time step.

For the SM 10-km domain and 10-km-domain footprint on the 30-km domain, we followed Massey et al. (2016) and used 27 SCAN and 15 GPS stations to bias correct the 5- and 25-cm soil moisture analyses at initialization using the mean difference between the observations and the corresponding GFS soil moisture values. This approach was used on the outer domains in lieu of the SM 3.3-km-domain approach because the latter does not account for the large-scale soil moisture gradients that are present across the Intermountain West. Massey et al. (2016) bias corrected soil moisture analyses in all of their domains and noted nighttime and daytime temperature improvement.

The resulting 3.3-km-domain 5-cm soil moisture differences between Control and SM are pronounced during MATERHORN-Fall (Fig. 4a). In Control, the EFS-DS and EFS-Playa soil moistures that are based on GFS soil moisture analyses are similar (on average within 0.01 m³ m⁻³), but SM soil moisture that is based on observations is on average 0.05 m³ m⁻³ lower at EFS-DS and 0.07 m³ m⁻³ higher at EFS-Playa than that in Control. During MATERHORN-Spring, Control again has similar soil moisture at EFS-DS and EFS-Playa (Fig. 4b). The use of measured soil moisture in SM
results in only a 0.01 m$^3$ m$^{-3}$ decrease at EFS-DS but a mean 0.13 m$^3$ m$^{-3}$ increase at EFS-Playa.

In SM-Albedo simulations, we kept the SM soil moisture changes but also changed the albedo and the saturation vapor pressure over the playa. WRF currently calculates an albedo on the basis of land-use classification, snow depth, and green-vegetation fraction, but SM-Albedo used a derived multiday albedo product from the Moderate Resolution Imaging Spectroradiometer (MODIS) that is based on atmospherically corrected surface reflectance observations (MCD43B3; Schaaf et al. 2002). MCD43B3 has two broadband albedos produced every 8 days using data from the previous 16 days at 500-m resolution, but we only used the shortwave white-sky albedo since Liu et al. (2009) found that it most closely matches pyranometer field measurements. We linearly interpolated the albedo measurements closest to the initialization time to the 3.3-km domain and filled missing data, which can occur because of persistent cloud cover, with the nearest available albedo. Meng et al. (2014) included a MODIS-derived albedo in their WRF simulations and obtained improved temperature forecasts.

The resulting mean albedo differences between Control and SM-Albedo for MATERHORN-Fall and MATERHORN-Spring are shown in Fig. 5. During MATERHORN-Fall, the mean Control albedo features little contrast between the playa and the surrounding lower-elevation desert-shrub region (Fig. 5a). In SM-Albedo, the mean albedo is higher over the playa and lower over the surrounding low-elevation desert-shrub region, with greater spatial variability (cf. Figs. 5a and 5b). These differences are also evident for MATERHORN-Spring when the playa albedo is even higher (Figs. 5c,d).

In SM-Albedo, we also reduced the saturation vapor pressure over the playa because the very high salinity of the playa has a strong impact on the evaporation rate and LE, but this effect is not parameterized in the Noah LSM. Salinity reduces the osmotic potential at the soil surface and increases the resistance to water vapor diffusion as a result of the occasional formation of a salt crust (Fujimaki et al. 2006). The complex parameterizations that capture these effects (e.g., Gowing et al. 2006) cannot easily be incorporated into the Noah LSM. Onton and Steenburgh (2001) reduced the saturation vapor pressure over the GSL by 30% and 6% over the north and south arms of the lake, respectively, because the two arms have different degrees of salinity. We reduced the playa saturation vapor pressure by 44%, which is the mean reduction for saturated GSL brine samples between $-10^\circ$ and $40^\circ$C (Dickson et al. 1965). This is a simplistic approach that assumes that the playa soil water is fully saturated with brine identical to the GSL brine and that playa soil water evaporation behaves identically to open-water evaporation. The effects of a reduced osmotic potential and salt crust, which would further reduce evaporation and LE, were ignored.

3. Results

a. Validation of SEB components

1) MATERHORN-FALL

A comparison of mean SEB observations with the corresponding Control forecasts at EFS-DS on mostly clear
days during MATERHORN-Fall shows that Control captures the incoming SW but slightly overpredicts the outgoing SW (Fig. 6a). The latter is because the mean WRF-derived albedo in Control is 0.03 higher than the mean observed albedo (Table 2; the latter is based on the ratio of daily integrated outgoing and incoming SW). Control also underpredicts the incoming LW, especially during the day (i.e., when incoming SW > 0), but the cause of this underprediction is not known (Fig. 6b). The outgoing LW is underpredicted during the day and overpredicted at night (i.e., when incoming SW = 0), suggesting an underprediction of the skin-temperature diurnal range. The SW and LW errors lead to a stronger negative $R_n$ at night and a weaker positive $R_n$ during the day (Fig. 6c).
The largest errors in Control arise in the $H$ and $G$ forecasts. The daytime $H$ maximum in Control is underpredicted by 62 W m$^{-2}$ and occurs 30 min later than observed (Fig. 6d). The magnitude of $G$ is overpredicted by as much as 49 W m$^{-2}$ during the day and 43 W m$^{-2}$ at night (Fig. 6e). The LE reaches a maximum of only 16 W m$^{-2}$ in the observations, which is slightly overpredicted (Fig. 6f).

The SM and SM-Albedo simulations produce nearly identical forecasts of the surface fluxes at EFS-DS that are much improved relative to Control (Fig. 6). The incoming LW and SW, as well as the outgoing SW, are
nearly identical to Control, but the outgoing LW has an improved diurnal cycle in SM and SM-Albedo when compared with the Control (Fig. 4b). The improved outgoing LW improves $R_n$ at night but leads to a slight underprediction of $R_n$ during the day (Fig. 4c).

The most substantial improvement relative to Control occurs in the $H$ and $G$ forecasts. The SM and SM-Albedo $H$ maxima are $43 \text{ W m}^{-2}$ higher than that in Control, although still underpredicted by $19 \text{ W m}^{-2}$, and are in phase with the observed maximum (Fig. 6d). The daytime $G$ maxima improve to a $17 \text{ W m}^{-2}$ under-prediction, and the nighttime upwelling $G$ also improves but remains overpredicted (Fig. 6e). The LE maximum is also closer to observed than is that of Control, but it exhibits a slight $3 \text{ W m}^{-2}$ underprediction (Fig. 6f). Overall, these results indicate that the Noah LSM, when driven with measured soil moisture and a modified soil thermal conductivity parameterization, captures the partitioning among the SEB components well at EFS-DS. Also, the nearly identical SM and SM-Albedo forecasts suggest that the assimilation of MODIS-derived albedo has little effect.

In contrast to the results at EFS-DS, Control provides a closer match to the surface flux observations at EFS-Playa (Fig. 7). The mean WRF-derived albedo in Control is the same as the mean albedo derived from SW observations at EFS-Playa (0.30; Table 2), resulting in nearly perfect forecasts of incoming and outgoing SW (Fig. 7a). The incoming LW remains underpredicted (cf. Figs. 6b and 7b), and the outgoing LW is close to observations at night but is underpredicted during the day (Fig. 7b). Nevertheless, these errors are relatively small, and $R_n$ is well captured (Fig. 7c). The largest forecast errors of the SEB components in Control at EFS-Playa are in $H$ and LE. The Control daytime $H$ maximum is underpredicted by $25 \text{ W m}^{-2}$ and occurs 1.5 h later than is observed (Fig. 7d). The daytime LE maximum is overpredicted by $32 \text{ W m}^{-2}$ and occurs 2.5 h later than is observed (Fig. 7f). Rehydration of the near-surface soil at night likely leads to the observed early-morning maximum, with LE decreasing during the day because of drying of the soil surface (Malek 2003). Ground flux $G$ is captured well during the day, but nighttime upwelling $G$ is overpredicted by as much as $23 \text{ W m}^{-2}$ (Fig. 7e).

The SM and SM-Albedo simulations produce worse forecasts of the SEB components at EFS-Playa when compared with Control, but the differences are small (Fig. 7). The daytime outgoing LW underprediction worsens slightly (Fig. 7b), although $R_n$ remains close to observations (Fig. 7c). The daytime $H$ underprediction also worsens, especially in SM (Fig. 7d), consistent with an increase in LE, which is overpredicted by as much as $74 \text{ W m}^{-2}$ in SM and $52 \text{ W m}^{-2}$ in SM-Albedo (Fig. 7f). In both SM and SM-Albedo, the LE maximum occurs 2.5 h later than is observed. These results indicate that the use of measured soil moisture over the playa actually degrades the forecast of the SEB components. The saturation vapor pressure adjustment in SM-Albedo reduces evaporation and improves the $H$ and LE forecasts, but, because the Noah LSM does not reduce the osmotic potential or parameterize the effects of a possible salt crust that would limit water vapor diffusion, LE remains overpredicted. Also, since the Noah LSM does not account for the uniquely high playa water table and associated rehydration of the topsoil at night (Malek 2003), LE remains out of phase with the observations. In summary, SM and SM-Albedo improve the SEB at EFS-DS but not at EFS-Playa during MATERHORN-Fall.

2) MATERHORN-SPRING

MATERHORN-Spring features stronger synoptic forcing (Fernando et al. 2015), wetter soils (e.g., Fig. 4), and higher playa albedos (e.g., Fig. 5) than does MATERHORN-Fall. During mostly clear days at EFS-DS, incoming and outgoing SW are slightly overpredicted by Control (Fig. 8a), and the incoming LW is underpredicted (Fig. 8b). The overprediction of the outgoing SW occurs because the mean WRF-derived albedo in Control is higher than the mean albedo derived from SW observations by 0.03 (Table 2). Similar to EFS-DS and EFS-Playa during MATERHORN-Fall, the outgoing LW in Control is underpredicted during the day, consistent with an underprediction of daytime skin temperature, and the incoming LW is underpredicted (cf. Figs. 6b, 7b, and 8b). These errors collectively contribute to a slight overprediction of the magnitude of the positive $R_n$ during the day and negative $R_n$ at night (Fig. 8c).

**Table 2.** Mean albedos at EFS-DS and EFS-Playa during MATERHORN-Fall and MATERHORN-Spring.

<table>
<thead>
<tr>
<th>Period</th>
<th>Location</th>
<th>WRF-derived albedo</th>
<th>MODIS-derived albedo</th>
<th>Obs albedo</th>
</tr>
</thead>
<tbody>
<tr>
<td>MATERHORN-Fall</td>
<td>EFS-DS</td>
<td>0.30</td>
<td>0.30</td>
<td>0.27</td>
</tr>
<tr>
<td></td>
<td>EFS-Playa</td>
<td>0.30</td>
<td>0.30</td>
<td>0.30</td>
</tr>
<tr>
<td>MATERHORN-Spring</td>
<td>EFS-DS</td>
<td>0.26</td>
<td>0.29</td>
<td>0.24</td>
</tr>
<tr>
<td></td>
<td>EFS-Playa</td>
<td>0.30</td>
<td>0.36</td>
<td>0.33</td>
</tr>
</tbody>
</table>
Unlike MATERHORN-Fall, \( H \) (Fig. 8d), \( G \) (Fig. 8e), and \( LE \) (Fig. 8f) are all captured well by Control at EFS-DS during MATERHORN-Spring. Also, SM and SM-Albedo have only slightly lower near-surface soil moisture than does Control at EFS-DS (i.e., Fig. 4), making the forecasts of the individual SEB components very similar among the three simulations. The slightly lower soil moisture does reduce the daytime \( G \) maximum by 18 W m\(^{-2}\) in SM and 24 W m\(^{-2}\) in SM-Albedo (Fig. 8e). Also, SM-Albedo has a mean MODIS-derived albedo that is 0.02 higher than the mean WRF-derived albedo in SM and 0.05 higher than the mean albedo derived from SW observations (Table 2), which results in higher outgoing SW (Fig. 8a), slightly lower daytime outgoing
LW (Fig. 8b), and lower $R_n$ (Fig. 8c) relative to Control. Overall, little to no improvement in forecasts of SEB components occurs in SM and SM-Albedo.

On mostly clear days at EFS-Playa for MATERHORN-Spring, the largest errors in forecasts of Control SEB components are an LE and $R_n$ overprediction (Fig. 9). The LE is overpredicted by as much as 49 W m$^{-2}$ and has a clear diurnal maximum that is not present in the observations (Fig. 9f). Incoming SW is slightly overpredicted by Control (Fig. 9a), whereas both outgoing and incoming LW are underpredicted (Fig. 9b). These errors result in a daytime $R_n$ overprediction by as much as 92 W m$^{-2}$ (Fig. 9c). Fluxes $H$ (Fig. 9d) and $G$ (Fig. 9e) are close to observations.
The SM simulation has much higher soil moisture at EFS-Playa than does Control (i.e., Fig. 4b), which exacerbates the daytime LE and \( R_n \) overprediction and also produces a daytime \( H \) and outgoing LW underprediction (Fig. 9). The SM LE maximum is 202 W m\(^{-2}\), which is 160 W m\(^{-2}\) greater than the observed maximum (Fig. 9e), and \( H \) is underpredicted by 75 W m\(^{-2}\) (Fig. 9d).

The mean MODIS-derived albedo in SM-Albedo is 0.36 at EFS-Playa, which is 0.06 higher than the mean WRF-derived albedos in SM and Control and 0.03 higher than the mean albedo derived from SW observations (Table 2). The higher albedo in SM-Albedo results in an outgoing SW overprediction (Fig. 9a) but also brings the daytime \( R_n \) closer to observations (Fig. 9c).
MODIS-derived albedo and saturation vapor pressure modifications also reduce the LE maximum by 51 W m$^{-2}$ relative to SM, but the LE maximum remains overpredicted by 110 W m$^{-2}$, which is 61 W m$^{-2}$ greater than in Control (Fig. 9f).

Results from this section show how SM and SM-Albedo improve SEB-component forecasts substantially at EFS-DS during MATERHORN-Fall and have little effect during MATERHORN-Spring when near-surface soil moisture is higher. On the other hand, SM and SM-Albedo degrade the forecasts of SEB components slightly at EFS-Playa during MATERHORN-Fall and substantially during MATERHORN-Spring.

b. Contrast in SEB component and 2-m temperature between EFS-DS and EFS-Playa

Differences in the SEB components between EFS-DS and EFS-Playa, especially in $H$, can produce spatial temperature gradients. Here, we examine how well WRF captures the $H$ and temperature contrasts between these two land surfaces during mostly clear days when observations were available at EFS-DS and EFS-Playa.

On mostly clear days during MATERHORN-Fall, the observed $H$ is higher at EFS-DS than at EFS-Playa, exceeding 57 W m$^{-2}$ in the late afternoon (Fig. 10a). Control greatly underpredicts this $H$ difference, largely because of the $H$ underprediction at EFS-DS (Fig. 6d). The underprediction of the $H$ difference contributes to an underprediction of the mean daytime temperature difference between seven SAMS sites over the desert shrub (SAMS-DS) and seven SAMS sites over the playa (SAMS-Playa) that is as large as 1.7°C (Fig. 10b; see Fig. 1 for SAMS-DS and SAMS-Playa locations). At night, the observed difference in EFS-DS and EFS-Playa 2-m temperature is as large as $-4.2$°C, but Control predicts a maximum 2-m temperature difference of only $-0.7$°C.

With greatly improved EFS-DS $H$ forecasts but degraded EFS-Playa $H$ forecasts, SM and SM-Albedo overpredict the daytime $H$ difference between the two sites (Fig. 10a) by as much as 52 W m$^{-2}$. The daytime SAMS-DS and SAMS-Playa 2-m temperature difference improves but remains underpredicted by as much as $0.7$°C. At night, the 2-m temperature difference improves substantially in SM and, especially, in SM-Albedo, but the nighttime temperature difference remains underpredicted.

Mostly clear days during MATERHORN-Spring have a larger observed $H$ difference between EFS-DS and EFS-Playa than during MATERHORN-Fall (cf. Figs. 10a and 11a). Control underpredicts the $H$ difference, but to a lesser degree than during MATERHORN-Fall, and underpredicts the 2-m temperature difference (Fig. 11b). The SM and SM-Albedo simulations greatly overpredict the $H$ difference maximum by as much as 102 and 72 W m$^{-2}$, respectively, but the daytime 2-m temperature difference is slightly underpredicted.

Figures 10 and 11 also illustrate how $H$ differences between EFS-DS and EFS-Playa cannot fully explain the difference in 2-m temperatures. For example, SM and SM-Albedo overpredict the daytime $H$ difference...
but underpredict the 2-m temperature difference. The SM also has a higher daytime $H$ difference but lower daytime 2-m temperature difference when compared with SM-Albedo. The cause of these discrepancies was not immediately obvious. In summary, SM-Albedo and, especially, SM increase the daytime $H$ difference between EFS-DS and EFS-Playa relative to those in Control, resulting in an overpredicted $H$ difference but an underpredicted, yet improved, daytime 2-m temperature difference.

c. Playa breezes

The temperature differences between the desert shrub and playa contribute to diurnal boundary layer circulations, such as playa breezes, that develop over the region under quiescent large-scale conditions (Rife et al. 2002). To examine the fidelity of the simulations in generating these circulations, we consider only mostly clear days that are quiescent. Days with a mean SAMS 1400 LST wind speed of less than 5 m s$^{-1}$ are assumed to be quiescent, which eliminates two MATERHORN-Fall and three MATERHORN-Spring mostly clear days from validation. We also only compare SM-Albedo with Control given the slight 2-m temperature difference improvement in SM-Albedo relative to SM. During MATERHORN-Fall, there is a mean 10-m diffusive northerly off-playa breeze at 1400 LST in Control (Fig. 12a). SAMS observations generally support this flow regime, especially over eastern DPG, where most SAMS are located. Overall, the Control wind speed and direction MAEs at 1400 LST are 0.97 m s$^{-1}$ and 42.9°, respectively, with a wind speed BE of $-0.27$ m s$^{-1}$ (Figs. 13a–c). In general, MATERHORN-Fall wind direction and speed MAEs are lowest during this afternoon period.

At 0500 LST, Control produces a large-scale northwesterly flow over much of the playa, with weak and variable flow over eastern DPG (Fig. 12d). These flows contrast with SAMS observations, which show weaker flow over much of the playa and a stronger down-valley, on-playa breeze over eastern DPG. Overall, the Control 0500 LST wind speed and direction MAEs are 1.39 m s$^{-1}$ and 69.8°, respectively, with a wind speed BE of $-0.29$ m s$^{-1}$ (Figs. 13a–c). These are higher than at 1400 LST, consistent with poorer model performance overnight and in the early-morning hours.

With a stronger 2-m temperature contrast between the desert shrub and playa, SM-Albedo produces a stronger off-playa breeze at 1400 LST than does Control (cf. Figs. 12a and 12b). SM-Albedo winds appear close to SAMS observations along the playa boundary in eastern DPG but are stronger than observations in southeastern DPG. Wind speed BEs improve and increase from $-0.27$ to 0.12 m s$^{-1}$ in SM-Albedo, which is statistically significant at the 95% level (Fig. 13c), but the wind speed and direction MAE improvement is not statistically significant relative to Control (Figs. 13a,b). Other daytime hours do have statistically significant wind direction MAE improvement (Fig. 13b). Daily mean wind speed MAE improvement is concentrated at stations located along and near the playa boundary in eastern DPG and over the playa, with higher MAEs at other stations (Fig. 14a). Overall, the stronger afternoon
off-playa breeze in SM-Albedo slightly improves both wind speed BE and wind direction MAE but does not improve the wind speed MAE because wind speeds improve at some stations but are overpredicted at others.

At 0500 LST, flow differences between Control and SM-Albedo are nearly indiscernible (cf. Figs. 12c and 12d), with the exception of a slightly weaker off-playa breeze along the eastern DPG playa boundary. The wind direction MAE improvement relative to Control is not statistically significant (Fig. 13b), although improvements are significant later in the morning. There is also a statistically significant wind speed MAE improvement, but the improvement is only 0.09 m s$^{-1}$, and other nighttime hours do not have statistically significant improvement (Fig. 13a). Thus, the strength
of the daytime off-playa breeze is the largest difference between Control and SM-Albedo during MATERHORN-Fall.

During MATERHORN-Spring, there is a mean diffusent off-playa breeze at 1400 LST (Fig. 15a) but no overall northerly flow component as during MATERHORN-Fall (Fig. 12a). Over eastern DPG, Control has a weaker northerly wind component than do the observations, which contributes to the large wind speed and direction MAEs of 1.51 m s$^{-1}$ and 52.7°, respectively (Figs. 13d,e), and large negative wind speed bias of $-0.38$ m s$^{-1}$ (Fig. 13f). Other daytime hours have similar errors. At 0500 LST during MATERHORN-Spring, Control produces a large-scale northwesterly flow that is supported by observations over much of the playa but is not supported over the lower elevations of eastern DPG because observations show a southeasterly on-playa breeze (Fig. 15c). The Control 0500 LST wind speed MAE is near its diurnal minimum at 1.42 m s$^{-1}$, the wind speed MAE is near its diurnal maximum at 64.3°, and the wind speed BE is $-0.45$ m s$^{-1}$ and negative throughout its diurnal cycle (Figs. 13d,e).

SM-Albedo increases the 1400 LST MATERHORN-Spring off-playa breeze relative to Control, but the largest differences occur away from eastern DPG (cf. Figs. 15a and 15b). This change, however, does not significantly improve wind speed or direction MAEs at 1400 LST or any other hour at the 95% level when compared with Control (Figs. 13d,e). Wind speed MAEs improve at stations located along the playa boundary in eastern DPG but worsen away from the boundary and over the playa (Fig. 14b). The greatest improvement is the increase of the wind speed BE at 1400 LST from $-0.38$ m s$^{-1}$ in Control to 0.10 m s$^{-1}$, which is statistically significant. Other daytime hours also have statistically significant wind speed BE improvement. At 0500 LST, however, there are no discernable flow differences between Control and SM-Albedo (cf. Figs. 15c and 15d), and there are no statistically significant nighttime error improvements (Figs. 13d–f).

In summary, the strength of the daytime off-playa breeze increases in SM-Albedo relative to that from Control, causing the daytime negative wind speed bias to become slightly positive, but wind speed and direction MAEs only marginally improve during MATERHORN-Fall, and the improvement is only occasionally statistically significant. This discrepancy is the result of lower MAEs only occurring over certain regions, like the playa boundary in eastern DPG, with higher MAEs elsewhere.

4. Conclusions

Because of the heterogeneity and complexity of land surface processes, numerical weather models, such as the Weather Research and Forecasting Model, may fail to accurately represent the forcing at the surface. In this study, WRF simulations of the autumn 2012 (1–20 October 2012; MATERHORN-Fall) and spring 2013
Mountain Terrain Atmospheric Modeling and Observations Program field campaigns are validated against observations of the components of the surface energy balance, collected over playa and desert-shrub land surfaces at the Dugway Proving Ground in northwestern Utah. MATERHORN-Spring features wetter soils and stronger synoptic forcing than does MATERHORN-Fall, and the EFS-Playa site has considerably wetter soils and a higher albedo than does EFS-DS, allowing for the validation of the SEB components of two contrasting periods at two contrasting land surfaces.

The largest sensible heat flux and ground heat flux errors in our Control WRF simulations that use a standard Noah land surface model and are initialized with Global Forecast System soil moisture analyses occur during MATERHORN-Fall at EFS-DS. The daytime $H$ is greatly underpredicted, and the $G$ and latent heat flux are greatly overpredicted. Simulations that incorporate soil moisture measurements into the land surface analyses and use the McCumber and Pielke (1981) soil thermal conductivity parameterization over silt loam and sandy loam soils significantly reduce these errors, illustrating the importance of accurate land surface analyses and parameterizations for this region.

At EFS-Playa during MATERHORN-Fall and MATERHORN-Spring, the largest errors in Control are an overprediction of LE and underprediction of $H$. The measured soil moisture is higher than the analyzed soil moisture at EFS-Playa, and when the measured soil moisture is assimilated into the WRF it exacerbates the LE overprediction in the SM simulations. Simulations that also include observed albedo from a Moderate Resolution Imaging Spectroradiometer product and a reduced saturation vapor pressure over the playa only marginally improve these errors at EFS-Playa in the SM-Albedo simulations. Playa land surface processes are unusual, and the playa LE is likely overpredicted because the Noah LSM does not account for the effects of a salt crust, high water table, or reduced osmotic potential on evaporation.

The resulting $H$ forecast improvement at EFS-DS and $H$ forecast deterioration at EFS-Playa between SM-Albedo and Control increase the $H$ difference between EFS-DS and EFS-Playa. Although the $H$ difference becomes overpredicted relative to observations, the 2-m temperature difference between EFS-DS and EFS-Playa improves but remains underpredicted. The improved 2-m temperature difference results in a stronger daytime off-playa breeze in SM-Albedo, but nighttime

Fig. 14. Change in mean daily wind speed MAE (m s$^{-1}$) at each SAMS from Control to SM-Albedo during (a) MATERHORN-Fall and (b) MATERHORN-Spring. Also shown are background WRF land use (color filled; see Fig. 1 for categories) and terrain (gray contours every 150 m).
wind speed and direction differences between Control and SM-Albedo are minimal. Improvement in wind speed mean absolute error is concentrated over the playa during MATERHORN-Fall and along the playa boundary in eastern DPG during both periods. The stronger daytime off-playa breeze may be penetrating too far from the playa.

This work highlights WRF and Noah LSM deficiencies over two common dryland land surfaces. Massey et al. (2014, 2016) showed large temperature forecast improvement when land surface analyses incorporate soil moisture measurements and a modified soil thermal conductivity parameterization, and these results illustrate how these changes also improve forecasts of the components of the SEB over the desert shrub but degrade forecasts over the playa. More work is needed to refine land surface analyses using observations or remote sensing datasets such as the NASA Soil Moisture Active Passive (SMAP) mission. Land surface parameterizations also need to be improved, especially...
for playa, for the benefits of improved land surface analyses to be fully realized. Sophisticated playa land surface parameterizations exist (e.g., Gowing et al. 2006) but have not yet been adapted for the Noah LSM.

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