

An Observational and Modeling Study of Impacts of Bark Beetle–Caused Tree Mortality on Surface Energy and Hydrological Cycles

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ABSTRACT

Bark beetle outbreaks have killed billions of trees and affected millions of hectares of forest during recent decades. The objective of this study was to quantify responses of surface energy and hydrologic fluxes 2–3 yr following a spruce beetle outbreak using measurements and modeling. The authors used observations at the Rocky Mountains Glacier Lakes Ecosystem Experiments Site (GLEES), where beetles killed 85% of the basal area of spruce from 2005–07 (prebeetle) to 2009/10 (postbeetle). Observations showed increased albedo following tree mortality, more reflected solar radiation, and less net radiation, but these postoutbreak radiation changes are smaller than or comparable to their annual preoutbreak variability. The dominant signals from observations were a large reduction (27%) in summer daytime evaporation and a large increase (25%) in sensible heat fluxes. Numerous Noah LSM with multiparameterization options (Noah-MP) simulations incorporating beetle-caused tree mortality effects were conducted to assess their impact on the surface hydrological cycle components that were not directly observed. Model results revealed substantial seasonal variations: more spring snowmelt and runoff, less spring–summer transpiration, and drier soil in summer and fall. This modeled trend is similar to observed runoff changes in harvested forests where reduced forest density resulted in more spring snowmelt and annual water yields. Model results showed that snow albedo changes due to increased litter cover beneath killed trees altered the seasonal pattern of simulated snowmelt and snow water equivalent, but these changes are small compared to the effect of leaf loss. This study highlights the need to include the transient effects of forest disturbances in modeling land–atmosphere interactions and their potential impacts on regional weather and climate.

1. Introduction

Over the past decade, high temperatures and severe droughts combined with historic forest management practices have resulted in increased levels of bark beetle

(BET)-caused tree mortality across the western mountain regions of North America (Fig. 1; Meddens et al. 2012). Under current climate change projections and assuming management approaches remain the same, future increases in air temperature and drought will likely result in more frequent beetle attacks (Bentz et al. 2010; Hicke et al. 2006). Beetles introduce blue-stain fungus into the tree vascular system, which interrupts the flow of water to the crown, thereby killing trees. This study assesses effects of bark beetle–caused tree mortality on surface energy and hydrologic budgets for the first few years following initial attack while the tree is dying, but needles are still on.

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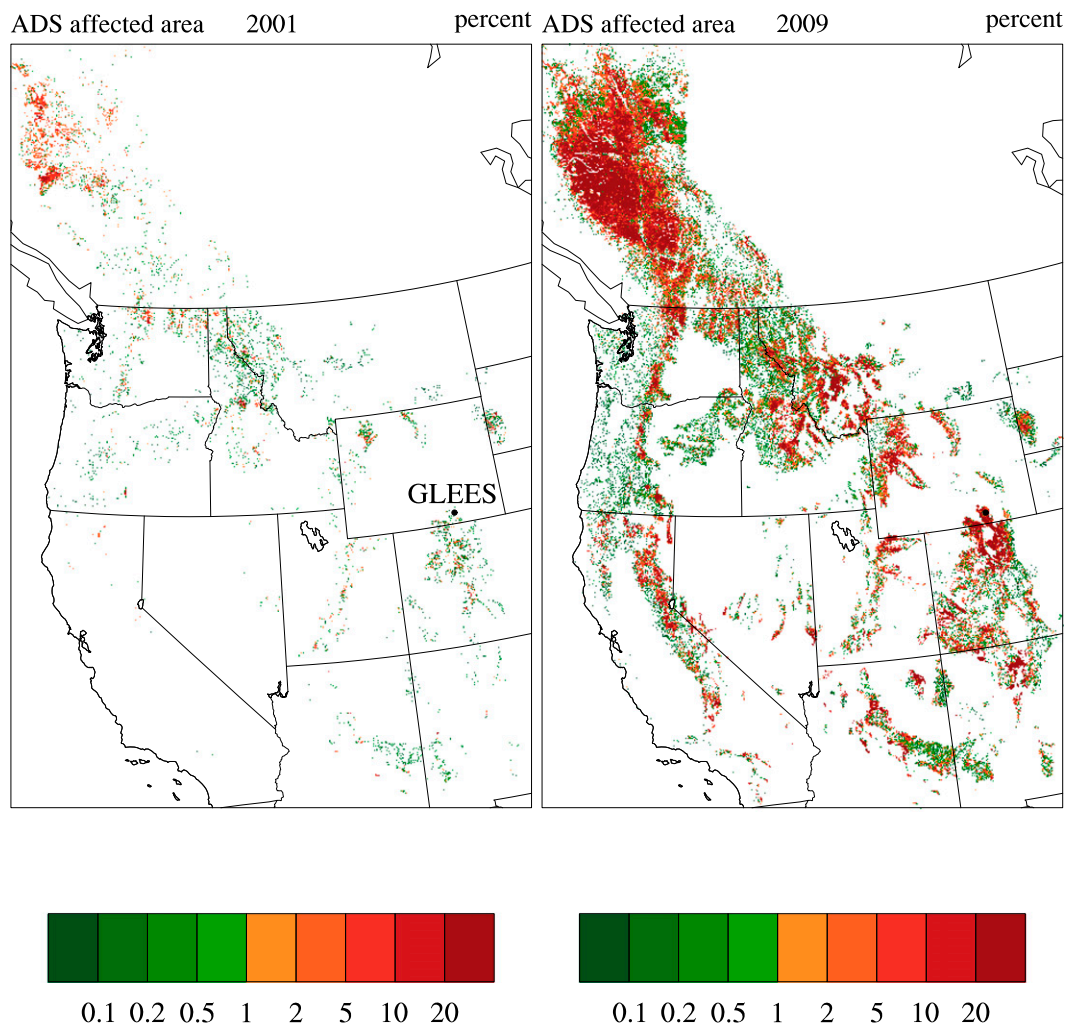


FIG. 1. Percent area of 1 km^2 grid cells killed by bark beetle from Meddens et al. (2012) based on the aerial detection survey (ADS) in the western United States and British Columbia shown for (left) 2001 and (right) the cumulative affected area from 2001 to 2009. Also shown is the approximate location of GLEES.

Insect outbreaks alter multiple forest ecosystem processes. Recognized biogeochemical effects of beetle-induced tree mortality include the conversion of severely infested forests from a small net carbon sink to a large net carbon source (Kurz et al. 2008), increased monoterpene emissions and concentrations of secondary organic aerosols (Berg et al. 2013), and potential impacts on biogenic volatile organic compound (BVOC) emissions (Duhl et al. 2013).

Biogeophysical impacts include changes to leaf area index (LAI), evapotranspiration, and surface albedo (ALB; Bright et al. 2013). Observations indicate that beetle-caused mortality modifies forest structure (e.g., reducing LAI; Coops et al. 2009; Goetz et al. 2012) and radiative properties (e.g., albedo and net radiation; O'Halloran et al. 2012). Such modifications inevitably change the forest-atmosphere exchange of heat and

moisture and surface hydrological cycles. Boon (2009) found a prolonged snowpack for the beetle-killed tree stands in British Columbia, Canada. In contrast, Pugh and Small (2012) noted a faster snowmelt in the beetle-killed forests in the headwaters of the Colorado River. A recent modeling study by Wiedinmyer et al. (2012) revealed the potentially significant influences of beetle-killed forests on regional weather, especially increasing surface sensible heat fluxes and surface temperature, consistent with the findings of Maness et al. (2013).

Although impacts of forest harvesting on hydrological variables have been documented (MacDonald and Stednick 2003; Guillemette et al. 2005), there have been few investigations that systematically assessed the effects of beetle-caused mortality on the surface energy and hydrological cycles. Boon (2012) pointed out the need for further modeling studies to examine the

influences of snowpack changes in postdisturbance forests on runoff generation. Adams et al. (2012) hypothesized that tree mortality could directly change canopy interception and evaporation and indirectly modify watershed hydrological cycle components such as runoff and streamflow. Pugh and Small (2013) indicated a reduced interception of snowfall for beetle-killed lodgepole stands. Modeling results by Mikkelsen et al. (2013) showed increased runoff at hillslope scales, but Guardiola-Claramonte et al. (2011) found decreased streamflow in a large catchment. Moreover, the review article by Edburg et al. (2012) pointed out a daunting challenge to bridge biophysical processes between tree-stand scales and regional scales.

The main objective of this study was to develop, through analysis of Glacier Lakes Ecosystem Experiments Site (GLEES) observations and land surface model simulations, a synthesis of beetle-caused mortality impacts on surface-energy and hydrological components at tree-stand scales. First, we analyzed surface-energy observations obtained from the GLEES tower. Second, we evaluated pre- and postbeetle Noah land surface model (LSM) with multiparameterization options (Noah-MP) simulations of the surface-energy parameters for GLEES against observations. Finally, we explored, through numerous Noah-MP sensitivity simulations, the impact of LAI and snow-surface albedo change following beetle-caused tree mortality on the surface energy and the hydrologic cycle.

2. Observations and the Noah-MP LSM

We selected GLEES for this study because it collects comprehensive, continuous, high-quality measurements of various terms in the surface-energy budget as part of the AmeriFlux network (<http://ameriflux.ornl.gov/>) and has been affected by beetle outbreak. GLEES, a 575-ha research watershed at an elevation of 3200–3400 m in the Snowy Range of southeastern Wyoming, has a long history of alpine vegetation research and meteorological monitoring (Musselman 1994). The subalpine forest basal area is composed of approximately 80% Engelmann spruce (*Picea engelmannii*) and 20% subalpine fir (*Abies lasiocarpa*), and the canopy reaches 18-m high.

GLEES has recently experienced a spruce beetle (*Dendroctonus rufipennis*) epidemic. In Frank et al. (2014), dendrochronological techniques were used to detect the timing of spruce beetle attacks. Though some attacks occurred as early as 2003, the epidemic peaked in 2008, while some new attacks occurred well after then. When spruce beetles attack these trees, they expose the plant to blue-stain fungi that occlude the xylem, cause hydraulic failure, and restrict gas exchange. Remarkably,

Engelmann spruce can survive several years in this condition. Unlike bark beetle attacks in pine trees, the spruce forest does not experience a red phase; rather, trees eventually drop their green needles and die. Moderate Resolution Imaging Spectroradiometer (MODIS) LAI revealed that needle drop and mortality peaked at GLEES in 2010, 2 yr after the peak in attacks (Frank et al. 2014). By 2010, 85% of the forested basal area had been impacted by spruce beetle (Speckman et al. 2015).

Thus, in our analysis, 2009 and 2010 are characterized as the postbeetle era. The 23-m-tall GLEES AmeriFlux scaffold tower is located at an elevation of 3190 m (41.3665°N, 106.2400°W) where ecosystem fluxes and ambient meteorology have been measured since October 2004. Ecosystem fluxes of carbon dioxide, water vapor, sensible heat, and momentum were obtained by applying the eddy covariance technique (Lee et al. 2004) to high-frequency measurements from a sonic anemometer (SATI/3Vx, Applied Technologies, Inc., Longmont, Colorado) and an open-path infrared gas analyzer (IRGA; LI-7500, Li-Cor, Inc., Lincoln, Nebraska). Four-component radiation fluxes are measured separately for shortwave radiation (PSP, Eppley Laboratory, Newport, Rhode Island) and longwave radiation (PIR, Eppley Laboratory). The effective footprint sampled by the scaffold is an area that extends about 0.75 km due west $\pm 60^\circ$ (Frank et al. 2014), with the footprint of the radiometers about $1/20$ the area, but also capturing a mixture of live and dead trees. Pre-epidemic measurements by Bradford et al. (2008) at 43 plots at GLEES showed a mean projected LAI of $7.7 \text{ m}^2 \text{ m}^{-2}$, but by 2010 the LAI declined to about $2 \text{ m}^2 \text{ m}^{-2}$ at the site (P. J. Fornwalt 2014, Rocky Mountain Research Station, U.S. Forest Service, personal communication).

The half-hour micrometeorological flux measurements at the GLEES AmeriFlux site are more comprehensively described by Frank et al. (2014). Data processes not described by Frank et al. (2014) include gap-filling techniques for the flux data. In total, 23% (minimum of 17% in 2006, maximum of 29% in 2010) of the sensible H and latent heat LE flux data that were missing from the dataset were filled based on biweekly (or seasonally for larger gaps) regressions. Daytime H and LE gaps were predicted from photosynthetic photon flux density, while nighttime gaps for both were predicted from the mean. Data were not screened for minimum, though only 5% of the dataset included measurements associated with friction velocity u_* below 0.2 m s^{-1} , the empirically derived GLEES threshold (Frank et al. 2014).

Noah-MP is a land surface model using multiple options for key land–atmosphere interaction processes (Niu et al. 2011). It contains a vegetation canopy submodel defined by a canopy top and bottom, needle dimensions

and density, crown radius, and radiometric properties. The snow interception in Noah-MP allows for both liquid water and ice to be present on the vegetation canopy and accounts for loading and unloading of snowfall, melting of intercepted snow and refreezing of the meltwater, frost–sublimation, and dew–evaporation. The canopy submodel employs a two-stream radiation transfer approach within the canopy along with shading effects necessary to achieve proper surface energy and water transfer processes, including undercanopy snow processes (Niu and Yang 2004). In addition, Noah-MP calculates vegetation transpiration using a Ball–Berry photosynthesis-based stomatal resistance with a dynamic vegetation model that allocates carbon to various parts of vegetation (leaf, stem, wood, and root) and soil-carbon pools (fast and slow).

Noah-MP uses a multilayer physically based snow model (Yang and Niu 2003; Niu et al. 2011), and its total number of layers is variable (up to three) depending on the total snow depth. Prognostic variables for each snow layer include partial volumes of liquid water and ice, snow density (or thickness), and temperature. Snow density is solved through three types of compaction processes: destructive metamorphism of new snow; snow load or overburden; and changes in snow structure due to melt–freeze cycles plus changes in crystals due to liquid water, following Anderson (1976). The rate of melting (or refreezing) is assessed from the energy excess (or deficit) resulting from adjusting the layer temperature to the freezing point. A stability correction to the undercanopy turbulent transfer is introduced to account for the strong stable condition of the warmer canopy overlying the snow surface during the melting season. The water flow out of the bottom of the snowpack is then available for infiltration into the soil and runoff.

The Noah-MP model uses a simple TOPMODEL-based runoff scheme with a simple groundwater model (Niu et al. 2005) to compute surface runoff and groundwater discharge. Surface runoff is mainly saturation-excess runoff, that is, the water (sum of rainfall, dew, and snowmelt) incident on the fractional saturated area of a model grid cell. Noah-MP uses a parameter to define the decrease of wind speed from the top of the canopy to the ground. The higher the parameter, the faster the wind decreases through the canopy, and the lower the exchange coefficient between the canopy and ground. This version of Noah-MP uses a constant wind parameter ($a = 3$) independent of wind speed. There are formulations of canopy wind parameters that depend on LAI (e.g., Goudriaan 1977) that indicate that the wind parameter would vary from $a = 2$ at LAI = 2 to $a = 4$ at LAI = 7. Therefore, Noah-MP does not have a variable wind parameter depending on LAI and a midrange value is used. Without considering stability corrections

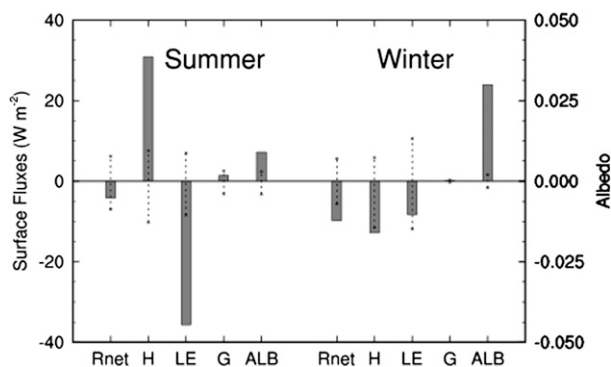


FIG. 2. Changes in Rnet, H , LE, G , and ALB based on measurements at GLEES for summer and winter. Vertical bars represent differences in surface-energy budget terms between postbeetle 2009/10 averaged values and prebeetle 2005–07 averaged values, and the dashed lines are the max and min annual departures during prebeetle era from their 2005–07 mean values.

to the wind profile, this limitation could introduce a factor of 2 difference in below-canopy exchange coefficient when using the assumed $a = 3$.

In this study, we used hourly GLEES meteorological (surface temperature, pressure, humidity, and precipitation) and radiation (downward shortwave and longwave) observations as forcing conditions to perform offline Noah-MP model simulations from October 2004 to December 2010. The first-year simulation was mainly for soil-state variables to spin up. Simulations for 2006/07 were used for validating the model and to represent natural (pre-epidemic) forest conditions, and simulations for 2009/10 were used to simulate beetle-caused mortality effects. First, a control simulation (termed as the CTL simulation) was conducted for the entire period (2004–10) without considering beetle-caused mortality effects. Then, the simulation BET was conducted in which Noah-MP parameters were modified to capture the known first-order influences of beetle-caused mortality by 1) increasing canopy reflectance (infrared reflectance is increased from 0.14 to 0.18 and visible reflectance is increased from 0.07 to 0.09) based on site observations of daytime net radiation Rnet reduction (Fig. 2) and 2) increasing tree canopy resistance to shut off transpiration, as shown in Fig. 2, namely, reducing transpiration by limiting photosynthesis through reduction of the maximum rate of carboxylation from 50 to $5 \mu\text{mol m}^{-2} \text{s}^{-1}$. Additionally, the Ball–Berry formulation for conversion of photosynthesis A to canopy resistance r_s [$1/r_s = mf(A) + b$] was modified to effectively cease canopy transpiration (m was reduced from the default evergreen needleleaf value of 6 to 1 and b was reduced from 2000 to 1). Additional sensitivity simulations were conducted to assess uncertainties in canopy

TABLE 1. Daytime surface-energy budget components averaged for summer (June–August) and winter (December–February) using the GLEES data. Also included are averaged winter and summer surface albedo (ALB) and PAR. Precipitation (P) is accumulated for a whole year and for spring–summer (March–August). The years 2005–07 (2009/10) represent prebeetle (postbeetle) conditions.

	Year	Daytime					P (mm)	PAR ($\mu\text{mol m}^{-2} \text{s}^{-1}$)	
		Rnet (W m^{-2})	H (W m^{-2})	LE (W m^{-2})	G (W m^{-2})	ALB			
Summer	2005	366.4	135.4	126.6	−22.7	0.08	1046	852	Whole year
	2006	358.6	122.7	142.0	−28.4	0.07	1031	839	
	2007	353.2	117.7	132.2	−26.5	0.07	1051	843	
	2009	357.8	154.1	100.2	−26.9	0.07	1416	841	
	2010	352.7	158.3	95.8	−22.1	0.09	1632	840	
Mean	2005–07	359.4	125.3	133.6	−25.9	0.07	1043	845	
	2009/10	355.2	156.2	98.0	−24.5	0.08	1524	841	
Winter	2005	164.2	97.5	66.9	1.1	0.15	394	990	Mar–Aug
	2006	169.7	97.4	79.9	1.4	0.15	380	984	
	2007	175.4	114.7	57.5	1.6	0.15	460	963	
	2009	165.6	88.3	68.9	1.5	0.16	576	976	
	2010	154.3	92.4	50.7	1.5	0.19	675	963	
Mean	2005–07	169.8	103.2	68.1	1.4	0.15	411	979	
	2009/10	160.0	90.4	59.8	1.5	0.18	625	969	

leaf loss and snow-surface albedo and are discussed in sections 3c and 3d.

3. Impacts of beetle-caused mortality on surface energy and hydrological cycles

a. Observation analysis

The quality of nighttime flux-tower data is questionable, especially for GLEES located at complex terrain. Therefore, we focused our analysis on daytime GLEES observation data. In our analysis, the daytime period was defined when the incoming solar radiation is larger than 20 W m^{-2} . Caution must be exercised in interpreting whether the beetle-caused tree mortality is solely responsible for recent changes in some surface-energy-flux components. For instance, summer daytime Rnet decrease is within its annual variation before infestation (Fig. 2) and change in ground heat flux G is generally less than its annual variations.

As shown in Table 1 and Fig. 2, postbeetle surface albedo slightly increases as a result of needle color change and needle loss, and its summer increase is less than that for winter, possibly as a result of increased snow-cover exposure caused by the reduced leaf area after a beetle attack. Such a seasonal trend and increased summer value (~ 0.01) are consistent with O'Halloran et al. (2012), but the winter change (0.03) is less than their reported value of 0.04–0.06. Daytime net radiation reduction ($\sim 1\%$ for summer and $\sim 6\%$ for winter) is primarily due to higher albedo. The greatest changes in surface-energy budget components are found for the summer daytime H and LE, which increased by 25% and decreased by 27%, respectively, and the

smallest changes are found in ground heat fluxes. Daytime winter H decreased after infestation, implying stronger downward transfer of heat from the atmosphere to the tree canopy and surface.

In winters following a beetle attack, smaller daytime Rnet that results from less solar energy absorbed by the canopy surface should reduce snowmelt rate. However, reduced sensible heat fluxes from the surface (i.e., more heat transferred from the warmer atmosphere to the forests) may speed up snow melting. The relative strength of these two factors in different climate regimes and for different tree stands plays a major role in explaining why snow depletion is faster (Pugh and Small 2012) or slower (Boon 2009) after beetle-caused mortality. This issue is explored through model-sensitivity simulations in sections 3c and 3d. The observed lower winter latent heat flux following beetle-caused tree mortality suggests less snow sublimation at this site.

The most prominent and robust signal is a large increase in summer daytime sensible heat flux and a slightly larger reduction in latent heat flux (i.e., evaporation). As a result, the combination of higher sensible heat flux and lower latent heat flux substantially increases the summer Bowen ratio (ratio of sensible heat flux to latent heat flux; Figs. 3a–c), especially in July and August. Decreased evaporation following tree die-off events at tree-to-stand scales is perhaps the most recognized effect of beetle-caused tree mortality on energy and water processes (Edburg et al. 2011; Hubbard et al. 2013; Maness et al. 2013). There is no observed reduction in precipitation and photosynthetically active radiation (PAR; Table 1) for 2009 and 2010; in fact, the observed annual and spring–summer precipitation for

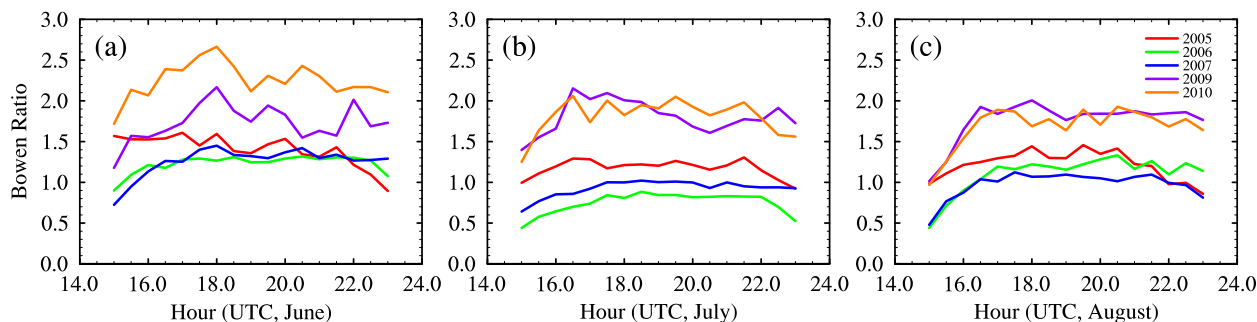


FIG. 3. Observed daytime (1500–2300 UTC, 0800–1600 LT) Bowen ratio (H/LE) averaged for (a) June, (b) July, and (c) August. Years 2005–07 represent prebeetle era and 2009/10 represents postbeetle era.

those two years is greater than that in the three preinfestation years. Thus, we can reasonably attribute these H and LE changes to the beetle-caused mortality that stops plant photosynthesis process and transpiration.

b. Evaluation of Noah-MP model simulations

Using the Noah-MP model simulations allows us to assess other tree mortality-caused changes in surface energy and hydrological terms that are not directly observed. A clear first step is to evaluate the capability of the model in simulating forest surface-energy exchange before (i.e., the CTL simulation) and after beetle mortality (i.e., the BET simulation) against observations. Although the 2006/07 CTL simulation underestimates winter (December–February) latent heat fluxes, its simulated diurnal cycle of surface heat fluxes and the summer (June–August) Bowen ratio closely track observations (see Figs. 4a–c). Table 2 summarizes the verification statistics of simulated net radiation, sensible heat, latent heat, and ground heat fluxes for summer when the forest–atmosphere interactions reach their peak. Following the method of Willmott (1981), the index of agreement (IOA) is computed as $IOA = 1 - \sum_{i=1}^n (M_i - O_i)^2 / \sum_{i=1}^n (|M_i - \bar{O}| + |O_i - \bar{O}|)^2$, where M and O are modeled and observed values. IOA is used

as a standardized measure of the degree of model prediction error and varies between 0 (no agreement) and 1 (perfect match). The model RMSEs ($\sim 50 \text{ W m}^{-2}$) are similar to those reported previously [e.g., see Table 5 of Chen et al. (2007)], and the bias ($< 21 \text{ W m}^{-2}$) and IOA (> 0.9 for R_{net} , H , and LE ; Pielke and Pearce 1994) are generally good.

We then evaluated the Noah-MP model simulation with beetle-caused tree mortality effects (BET). Figures 5a–e show that the Noah-MP simulation incorporating the major tree mortality effects (i.e., ceasing transpiration and increasing surface reflectance in the beetle-affected trees) increases summer sensible heat fluxes' maxima by approximately 30 W m^{-2} . While it may have underestimated latent heat fluxes and hence overestimated the Bowen ratio, the simulation with postoutbreak effects (BET) is generally closer to postbeetle-caused mortality observations than the simulation assuming natural forests (CTL). However, including beetle effects in Noah-MP does not significantly impact winter surface-energy fluxes (Figs. 5a,c).

c. Simulated impact of vegetation cover change on surface energy and hydrologic cycle

It is known that in the first few years following initial beetle attack, the forest fractional coverage does not

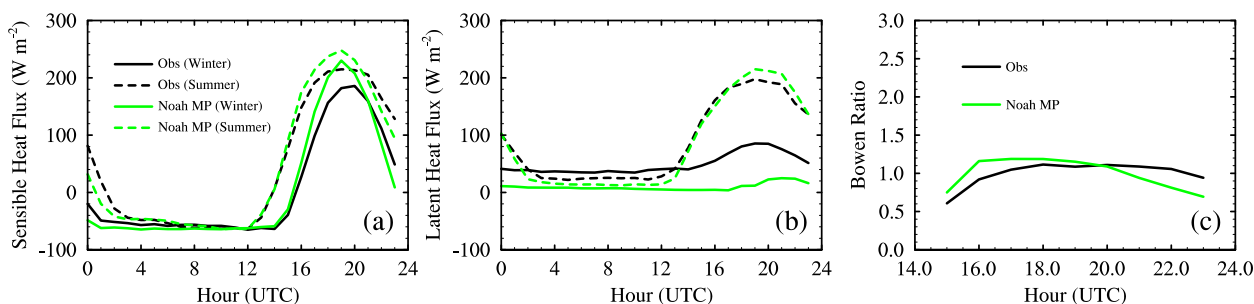


FIG. 4. Observed [black solid (winter) and dashed (summer)] and Noah-MP modeled summer (June–August, green dashed) and winter (December–February, green solid) averaged diurnal cycle of (a) sensible heat fluxes, (b) latent heat fluxes, and (c) observed (black) and modeled summer (green) averaged daytime (1500–2300 UTC, 0800–1600 LT) Bowen ratio for 2006/07 (prebeetle). Noah-MP simulation assumes a pristine needleleaf.

TABLE 2. RMSE, mean bias, and IOA of Noah-MP simulated Rnet, H , LE, and G (W m^{-2}) averaged for June–August for preoutbreak periods (2005–07).

	RMSE				Bias				IOA			
	Rnet	H	LE	G	Rnet	H	LE	G	Rnet	H	LE	G
2005	51.59	59.73	45.11	39.79	−5.76	0.38	−2.56	−5.76	0.99	0.96	0.92	0.75
2006	50.49	50.84	42.05	52.44	−20.97	−1.23	−11.35	−15.82	0.99	0.96	0.94	0.74
2007	50.80	56.53	53.26	30.41	−20.57	0.13	4.16	−3.27	0.99	0.95	0.90	0.85

change significantly; however, the green LAI is reduced (Bright et al. 2013). To explore the uncertainty in the exact amount of LAI loss depending on tree species, we conducted two simulations to assess the effect of LAI: 1) LAI/2, in which the observed prebeetle LAI was reduced by half, and 2) LAI/4, where the original value was reduced to a quarter, similar to that reported for 2010 by P. J. Fornwalt (2014, Rocky Mountain Research Station, U.S. Forest Service, unpublished data; Table 3).

Simulations with lower LAI slightly decreased winter surface-energy fluxes (Figs. 5a,c) as compared with BET, but substantially (especially LAI/4) reduced summer H and increased LE (Figs. 5b,d). For instance, the extreme case of LAI/4 decreased the summer H peak values in BET by $\sim 50 \text{ W m}^{-2}$ and increased LE by roughly the same value. Consequently, Bowen ratios were substantially lower with reduced LAI (Fig. 5e). In general, simulations with LAI reduced by half (LAI/2) produce surface heat fluxes and Bowen ratios that agree better with observations taken during postbeetle periods.

Figure 6 shows the shortwave and longwave radiation transferred between tree canopy and ground–snow surface calculated by Noah-MP. Reducing LAI permitted more penetration of solar radiation from the tree canopy to the ground surface (Figs. 6a,b). The same effect (i.e., more openness in forest canopies) increased the overall albedo because of more exposure of snow surface and bare ground under the canopy, so the amount of reflected radiation increased as well (Figs. 6c,d). In the wintertime, the increased subcanopy solar radiation is nearly compensated by increased reflected solar radiation at the surface due to high snow albedo, so the net gain of surface solar energy is small. In contrast, the summer maximum subcanopy downward solar fluxes is substantially increased (e.g., increased by $\sim 200 \text{ W m}^{-2}$ in LAI/4 compared to CTL and BET), while the reflected solar radiation is only increased by $\sim 20 \text{ W m}^{-2}$, which led to a large increase in summer net solar radiation at the surface. Regarding the subcanopy longwave radiation component, lower LAI slightly decreased the longwave radiation from the canopy to the ground surface (Figs. 6e,f), but increased the upward longwave radiation at the surface, especially in the summertime because of higher subcanopy

solar radiation transmission and hence higher ground surface temperature (Fig. 6h).

Changes in subcanopy net radiation modify the surface hydrologic cycle (Wild and Liepert 2010). In the BET simulation, higher surface albedo allows more solar radiation reflected in summer and hence less net radiation at the surface ($\sim 10\%$ for the annual total amount; Figs. 7a, 8a). Note that in Figs. 7 and 8, all energy flux terms are converted into water equivalent (mm) using the formulation from Stull (1988). For instance, $\text{LH} (\text{mm s}^{-1}) = \text{LE} (\text{W m}^{-2}) / \text{Lv}$, where $\text{Lv} (\text{J kg}^{-1}) = (2501 - 0.00237 \times \text{Ta}) \times 10^6$ (Ta is air temperature in Celsius). Note that these conversions are done with the latent heat of vaporization, but the same amount of energy such as Rnet can produce more (less) snowmelt (snow sublimation) than evaporation. Thus, these converted water-equivalent quantities from energy terms only provide a rough estimate for the sake of comparisons.

Overall, the greatest difference between CTL and BET is the total evaporation (i.e., LE), which from May to October is decreased by $\sim 22\%$ of the annual total amount in BET compared to the CTL simulations (Figs. 7c, 8c), mainly due to the cessation of evapotranspiration during the vegetation growing season. The decrease of latent heat fluxes is greater than the amount of decreased net radiation, and therefore, the annual total amount of sensible heat flux increases by $\sim 16\%$ (Fig. 8b). Less evapotranspiration in BET causes soils to retain more moisture in the spring growing season and results in wetter soils in summer and fall (Fig. 8e), leading to an increase of runoff in BET by $\sim 11\%$ (Fig. 8d). The other significant difference between CTL and BET is the reduction in Rnet (Fig. 8a), which is balanced by a large reduction in LE (Fig. 8c) and a modest increase in H (Fig. 8b).

In the wintertime, because the net increase of subcanopy solar radiation is less than the reduction of longwave radiation in the Noah-MP simulations with LAI loss, which results in a slight decrease in net radiation at the snow surface. Pugh and Small (2013) observed that the interception of snowfall was reduced from 41% for living stands to 18% for beetle-killed lodgepole stands. Combining reduced net radiation with more canopy openness (i.e., less interception of snowfall

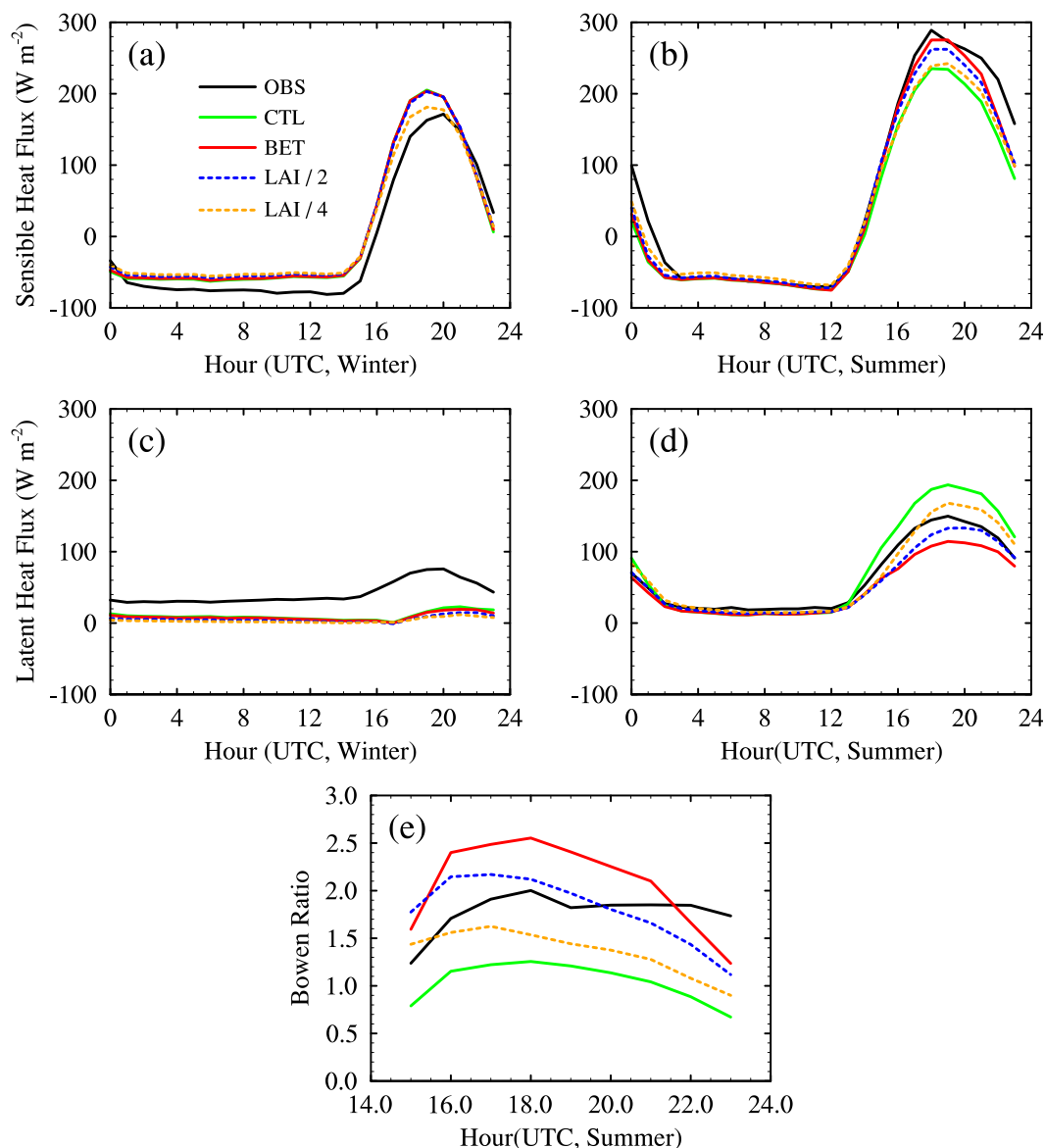


FIG. 5. Observed and Noah-MP modeled summer- and winter-averaged diurnal cycle of surface heat fluxes for 2009/10: (a) winter sensible heat fluxes, (b) summer sensible heat fluxes, (c) winter latent heat fluxes, (d) summer latent heat fluxes, and (e) observed and modeled summer-averaged daytime (1500–2300 UTC, 0800–1600 LT) Bowen ratio. Lines show observations (black), Noah-MP simulation assuming a pristine needleleaf tree (CTL, green), Noah-MP simulation with the beetle-caused tree mortality effects (BET, red), LAI/2 (as in BET, but for reduced LAI = 3.75, blue), and LAI/4 (as in BET, but for reduced LAI = 1.88, orange).

by canopy), simulations with LAI reduction favor snow accumulation. For instance, the peak snow water equivalent (SWE) in LAI/4 in early April substantially increases by roughly 50 mm when compared to the CTL and BET simulations (Fig. 7h). However, large increase of spring (April and May) subcanopy solar radiation in LAI/2 and LAI/4 are used to melt more snow (Figs. 7g, 8g), which resulted in high snow melting rate and earlier snow disappearance by about 10 days in LAI/4 as compared to CTL and BET (Fig. 7h).

Compared to the BET simulation, reducing LAI decreases the spring net radiation and increases the summer net radiation (by approximately 10 mm month^{-1} ; Fig. 7a), but the opposite pattern is found in sensible heat flux (Fig. 7b). The simulations incorporating different degrees of LAI loss (LAI/2 and LAI/4) produce less evaporation in March–May but more evaporation in June–August (by approximately 10 mm month^{-1} ; Figs. 7c, 8c), resulting in drier soil in the summer and fall (Figs. 7e, 8e). Reducing the LAI decreases winter snow sublimation

TABLE 3. Noah-MP model-sensitivity experiments with regards to LAI and snow albedo. More details are provided in section 3b.

Experiment	Description
LAI/2	As in BET, but reducing LAI to half of LAI in BET (LAI = 3.75).
LAI/4	As in BET, but reducing LAI to one fourth of LAI in BET (LAI = 1.88).
LAI/2S5	As in LAI/2, but reducing snow-surface albedo by 5%.
LAI/4S5	As in LAI/4, but reducing snow-surface albedo by 5%.
LAI/2S10	As in LAI/2, but reducing snow-surface albedo by 10%.
LAI/4S10	As in LAI/4, but reducing snow-surface albedo by 10%.

(Fig. 7f), which is consistent with observations from GLEES following the loss of canopy (J. Frank 2013, unpublished data). All things considered, the LAI-loss effect does not change the annual total net radiation but reduces the total evaporation by $\sim 10\%$ (Figs. 8a,c). The uncertainty in LAI loss only affects the seasonal variability of evaporation (Fig. 7c) but not its total amount (Fig. 8c). The total runoff in LAI/4 increases by about 5% compared to BET (Fig. 8d) due to less intercepted precipitation by canopy, decreased sublimation, and more snow accumulated on the ground; however, the snow season is shortened by about 10 days (Fig. 7h). Reductions in total evaporation and increased runoff are expected following tree mortality at tree and stand scales (e.g., Edburg et al. 2012; Hubbard et al. 2013) and are consistent with the modeling results of Mikkelsen et al. (2013).

d. Impact of snow-surface albedo change on surface energy and hydrologic cycle

Thus far, we have examined the possible range of effects of LAI loss on surface energy and hydrological cycles. In this section, we focus on investigating the combined effects of LAI loss and snow-surface albedo following beetle mortality. It is reported that, on average, snow-surface albedo under killed trees is lower by 5% compared to that under pristine trees because of the increased litter cover (Pugh and Small 2012). Again, it is difficult to quantify the exact amount of shedding needles (and hence its influences on snow albedo) for a specific site because of a lack of observations. Using the studies of Pugh and Small (2012) and Bright et al. (2013) as guidance to address this uncertainty issue, we conducted two sensitivity tests for each scenario of LAI loss: with 5% reduction (simulations LAI/2S5 and LAI/4S5) and with 10% reduction in snow-surface albedo (simulations LAI/2S10 and LAI/4S10), as described in Table 3.

Decreasing snow-surface albedo modifies both winter and summer energy fluxes (Figs. 9a–d) because snowpack exists until June at this high-elevation site (Fig. 7h). However, the changes in surface-energy fluxes in those simulations with litter cover were less than 10 W m^{-2} and less than those by changing LAI (see Fig. 5). In general, simulations with lower snow-surface albedo produced higher sensible and latent heat flux in early summer while snowpack still exists, but had negligible influence on winter latent heat fluxes (Fig. 9c). Lower snow-surface albedo with more leaf loss (i.e., in simulations LAI/4S5 and LAI/4S10) tended to increase summer Bowen ratios (Fig. 9e) because of relatively higher summer sensible heat fluxes. It is worth noting that the combination of lower snow-surface albedo and LAI had greater impacts on surface-energy budgets, which is certainly related to subcanopy radiation processes.

Although lower snow-surface albedo (for given LAI) did not significantly alter the subcanopy solar radiation reaching the ground–snow surface (Figs. 10a,b), it reduced the maximum radiation reflected from ground to canopy by approximately $5\text{--}14 \text{ W m}^{-2}$ in winter and by $3\text{--}6 \text{ W m}^{-2}$ in summer (Figs. 10c,d). As a result, in simulations with lower albedo, the snow–ground surface absorbed more solar energy, had higher surface temperature, and produced slightly higher ($<3 \text{ W m}^{-2}$) upward longwave radiation (Figs. 10g,h).

Moreover, in simulations with thinner forest canopy, the loss in net longwave radiation was less than the increase in net solar radiation at the ground surface, resulting in a net gain of surface and undercanopy radiation, especially for the first half year (Figs. 11a,f). This is the primary reason for increased snowmelt amount in spring (April–May; Fig. 11g) and earlier snow ablation (Fig. 11h), which, in turn, produced wetter soil in spring (Fig. 11e). That led to increased runoff, especially in spring (Fig. 11d). Because the precipitation input was identical in all simulations, the earlier snowmelt shortened the snow season (Figs. 11g,h), which decreased soil water storage and runoff in June and subsequent summer months (Figs. 11d,e). Because of little differences in net radiation for the second half year, the decrease of soil water storage limited evaporation from July and therefore enhanced sensible heat fluxes. Note that the soil was wetter in spring and drier in summer and fall in the lower snow-surface albedo simulations, so the annual soil storage change was quite small. However, the enhanced soil-moisture depletion in summer months potentially increases drought, enhances ecosystem impacts, and augments wildfire frequency in high-elevation forested regions (Molotch et al. 2009; Morton et al. 2013).

The annual total amount of net radiation, sensible heat flux, runoff, sublimation, and snowmelt (in water

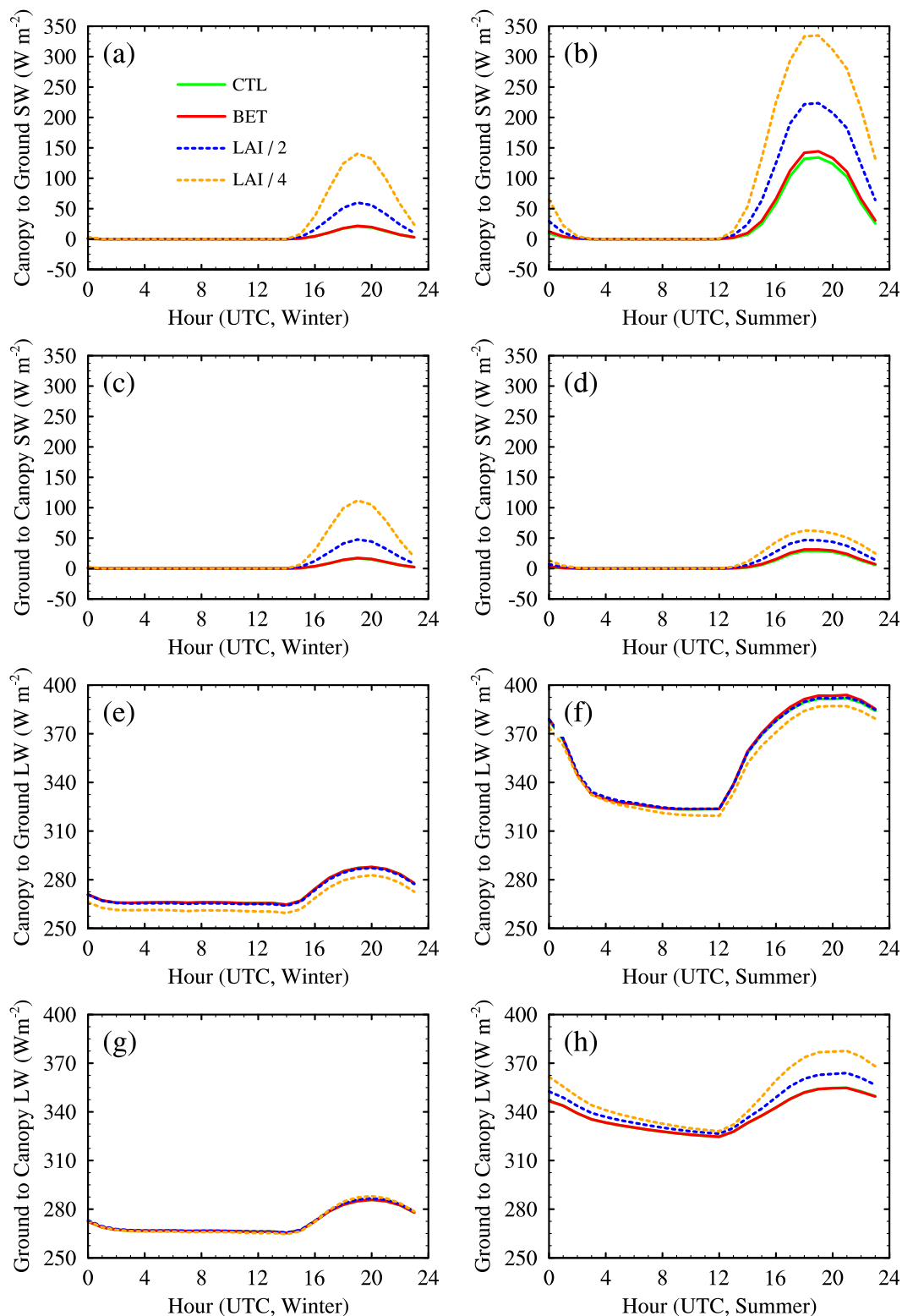


FIG. 6. Noah-MP modeled (a),(c),(e),(g) winter- and (b),(d),(f),(h) summer-averaged diurnal cycle of undercanopy budget terms for 2009/10: (from top to bottom) SW from canopy to ground and from ground to canopy, and likewise for LW. Colors as in Fig. 5.

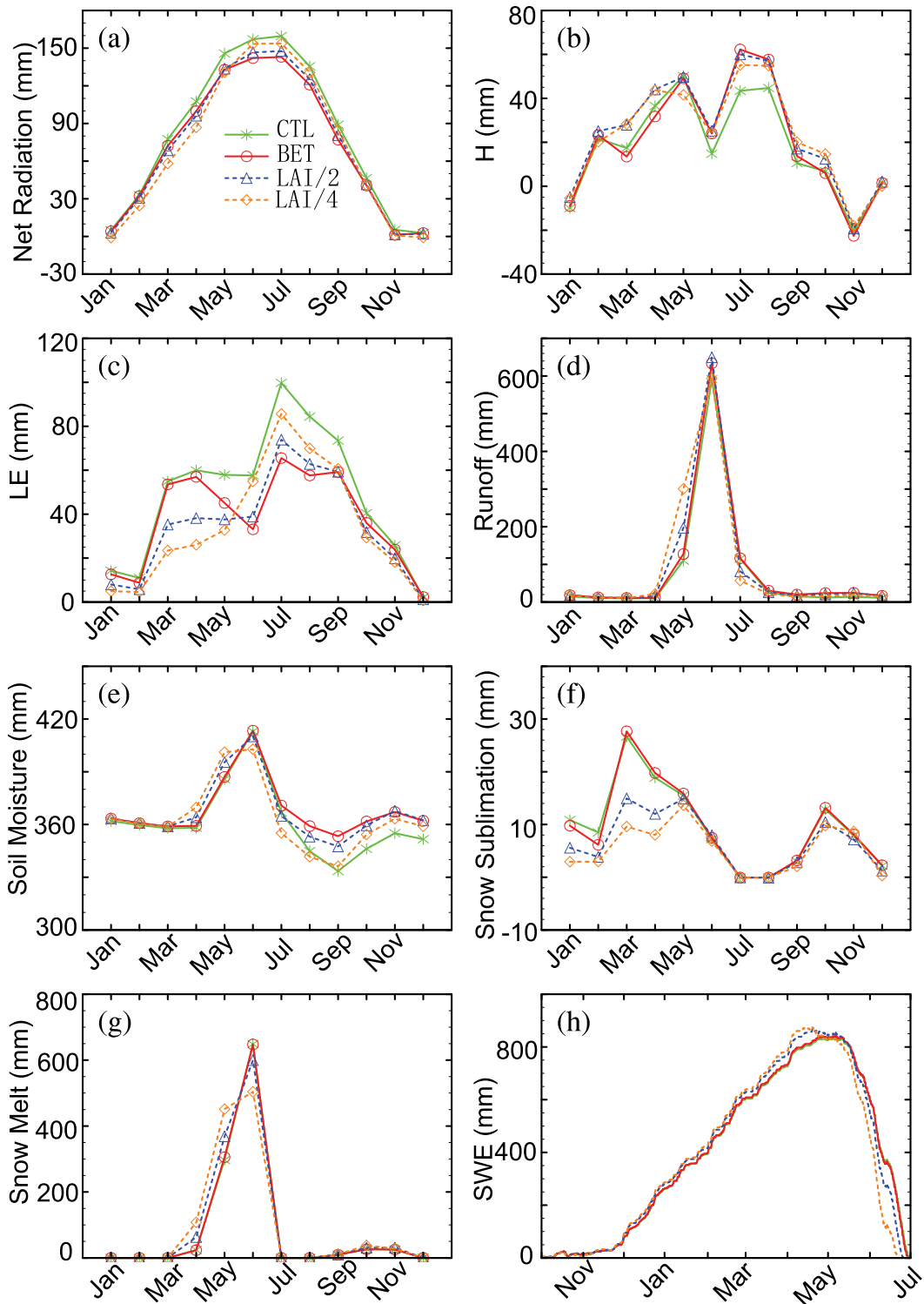


FIG. 7. Simulated hydrological cycles averaged for 2009/10 by Noah-MP with beetle-caused tree mortality (BET) and the range of possible LAI changes (LAI/2 and LAI/4): (a) monthly total net radiation, (b) monthly total sensible heat flux, (c) monthly total evaporation, (d) monthly total runoff, (e) monthly averaged soil moisture in the first 1-m soil depth, (f) monthly total snow sublimation, (g) monthly total snowmelt, and (h) daily mean SWE. Note that the x axis in (h) is different from others and that it ranges from October to June. Colors as in Fig. 5.

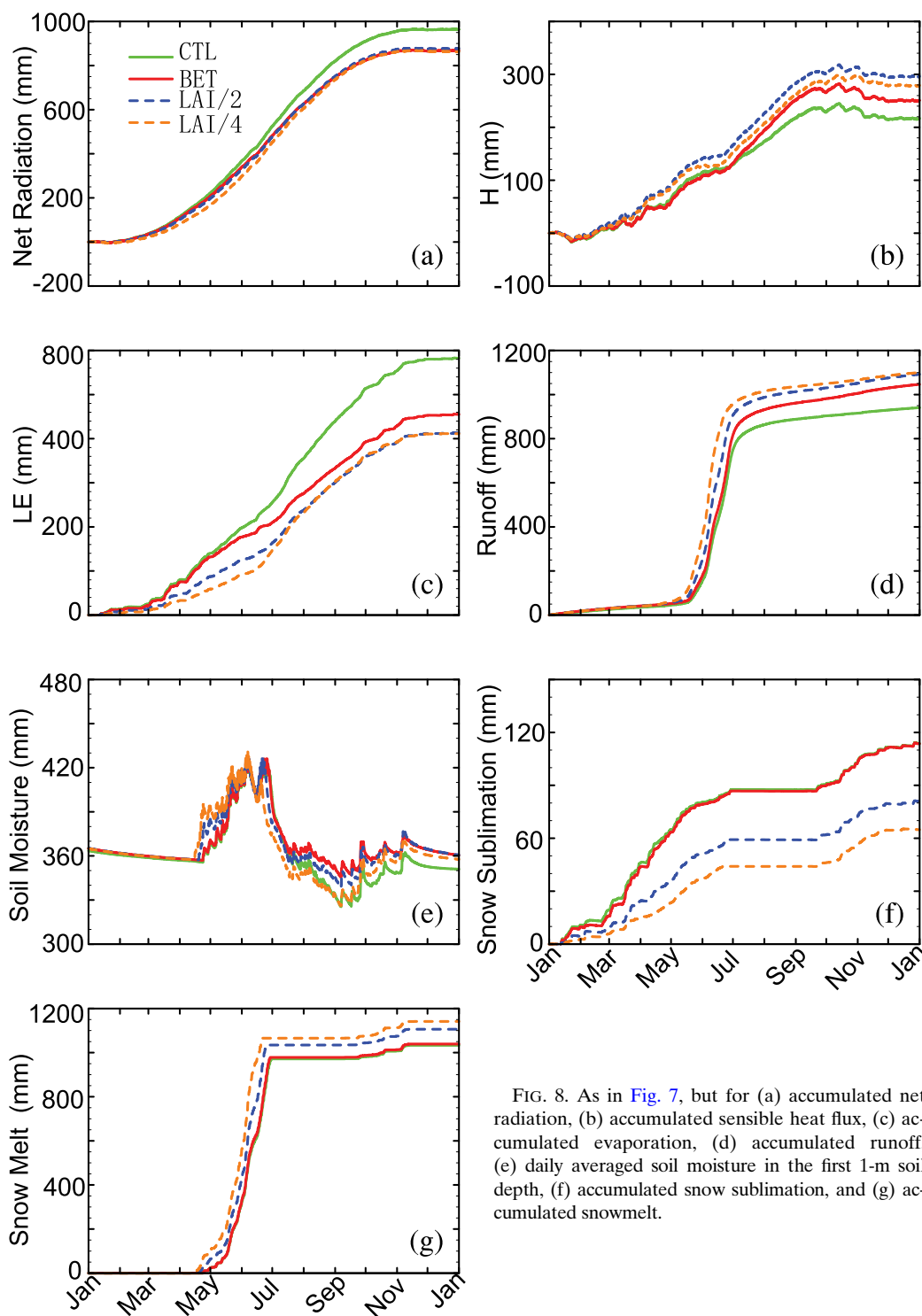


FIG. 8. As in Fig. 7, but for (a) accumulated net radiation, (b) accumulated sensible heat flux, (c) accumulated evaporation, (d) accumulated runoff, (e) daily averaged soil moisture in the first 1-m soil depth, (f) accumulated snow sublimation, and (g) accumulated snowmelt.

equivalent) from all Noah-MP simulations are summarized in Fig. 12 to synthesize various model-sensitivity results. Compared to the control simulation, model simulations with different degrees of beetle-mortality effects (from BET to LAI/4S10) reduced the overall

net radiation (as energy input to the canopy–snow–ground continuum) by 80–101 mm. This R_{net} reduction roughly translates into a reduction in evaporation by 166–171 mm, which is largely compensated by an increase in sensible heat fluxes (35–85 mm). Regarding

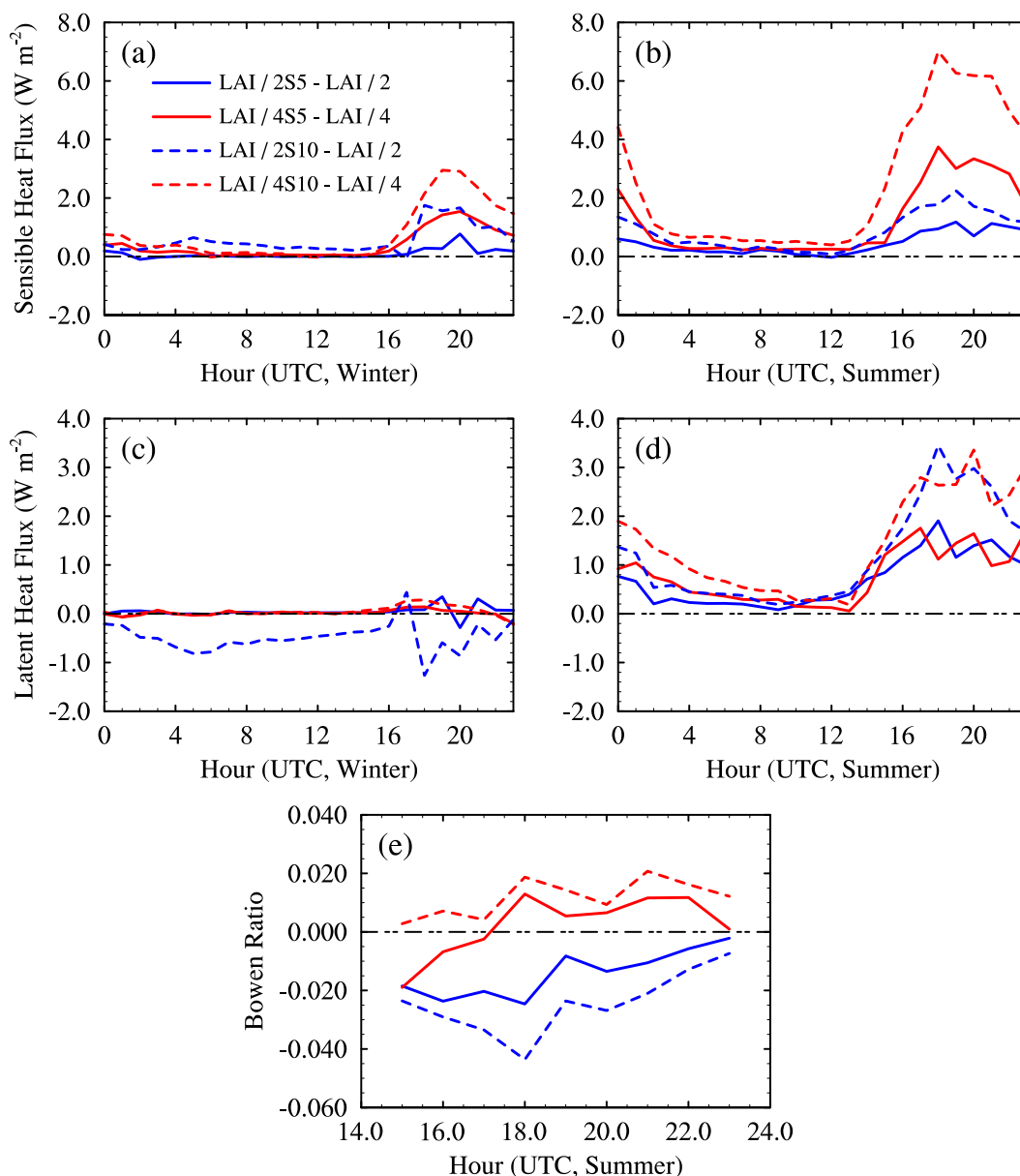


FIG. 9. Differences in diurnal cycle of surface fluxes and Bowen ratios between various Noah-MP sensitivity simulations (Table 2), averaged for (a),(c) winter and (b),(d),(e) summer during 2009/10 (postbeetle). Bowen ratios are calculated only for summer daytime (1500–2300 UTC, 0800–1600 LT). For instance, LAI/2S5 – LAI/2 is the difference between the simulation LAI/2S5 and the simulation LAI/2.

the water cycle components, decreased evaporation led to more runoff (increased by 105–160 mm). Additional sensitivity simulations with varying LAI and snow albedo did not produce significant differences in the overall Rnet, evaporation, and sensible heat fluxes (Figs. 12a,c,d). However, they substantially modified the undercanopy net radiation, and induced large variability in simulated sublimation and snowmelt amount among these simulations. For instance, a greater solar radiation transmitted through the less dense canopy in

LAI/2 and LAI/4 linearly increased the undercanopy net radiation in BET by ~100–200 mm, which was largely reflected in the increased snowmelt (74 mm in LAI/2 and 109 mm in LAI/4). Increasing snow-surface albedo only modestly compensated the effects of leaf loss by increasing slightly the undercanopy net radiation and sensible heat fluxes, but did not significantly change the snow sublimation and melt.

Although Boon (2012) found that large-scale runoff response may be affected by the distribution of

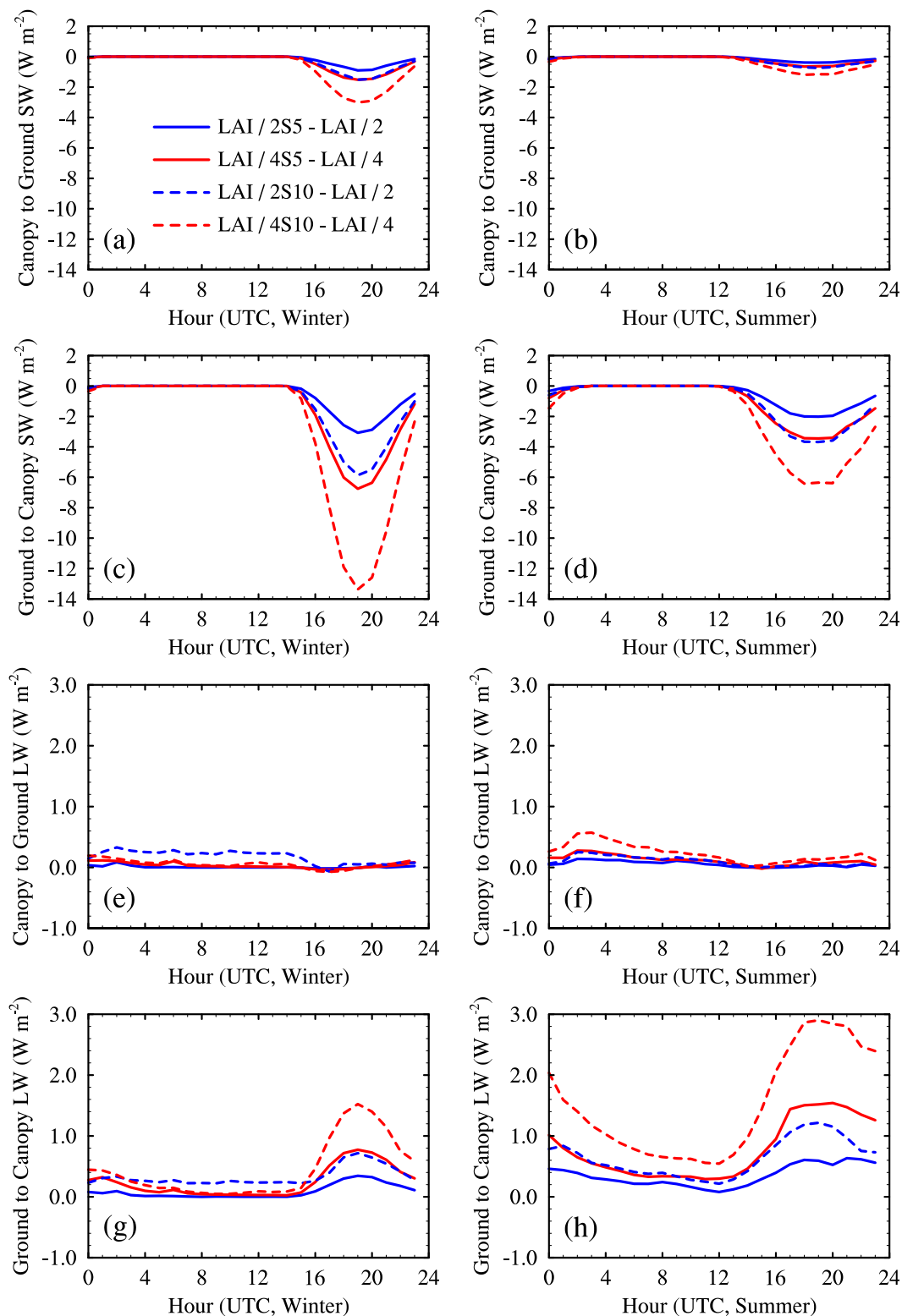


FIG. 10. As in Fig. 9, but for the diurnal cycle of subcanopy radiation budget terms: (a),(b) shortwave radiation (W m^{-2}) from forest canopy to ground; (c),(d) shortwave radiation (W m^{-2}) from ground to forest canopy; (e),(f) longwave radiation (W m^{-2}) from forest canopy to ground; and (g),(h) longwave radiation (W m^{-2}) from ground to forest canopy.

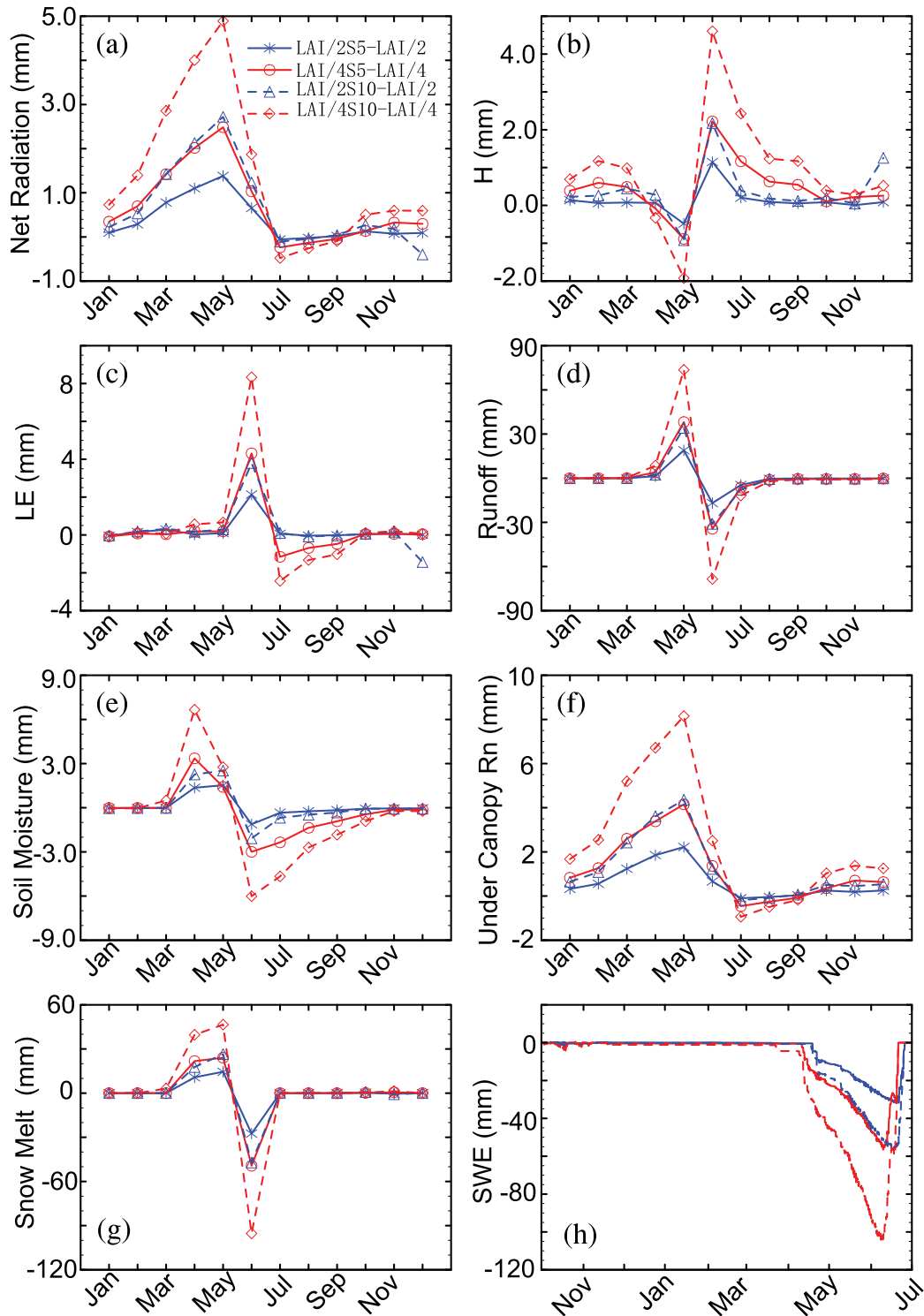


FIG. 11. Differences in simulated energy and hydrological components (mm) between Noah-MP snow albedo sensitivity simulations (Table 2), averaged for winter and summer during 2009/10 (postbeetle): (a) monthly total net radiation, (b) monthly total sensible heat flux, (c) monthly total evaporation, (d) monthly total runoff, (e) monthly averaged 0–1-m total soil moisture, (f) monthly total under canopy net radiation, (g) monthly total snowmelt, and (h) daily SWE. Note that the date in (h) is different from the others and that it ranges from October to June.

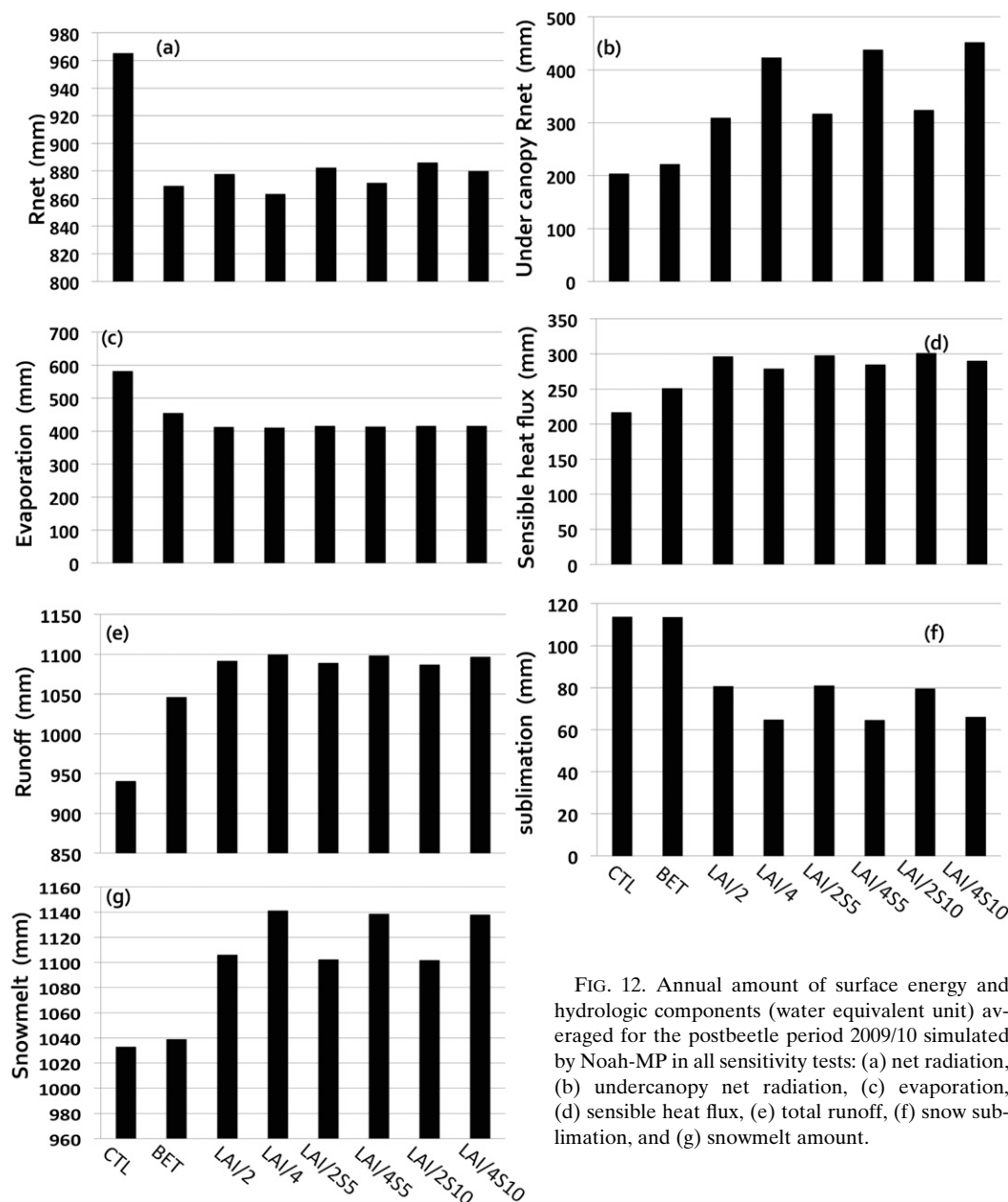


FIG. 12. Annual amount of surface energy and hydrologic components (water equivalent unit) averaged for the postbeetle period 2009/10 simulated by Noah-MP in all sensitivity tests: (a) net radiation, (b) undercanopy net radiation, (c) evaporation, (d) sensible heat flux, (e) total runoff, (f) snow sublimation, and (g) snowmelt amount.

tree-stand types across a watershed, previous studies did not provide extensive evidence of runoff changes in beetle-infested watersheds. Instead, the trend discussed in this paper is similar to the modeling study of Mikkelsen et al. (2013) and to those found in harvested forests where reduced forest density and cover results in more spring snowmelt and annual water yields (MacDonald and Stednick 2003). Their report also indicated that the increase in water yield over harvested forests will decrease as the forest regrows, but it may take about 60 yr to return to their predisturbance levels in spruce–fir and lodgepole pine forests because of their slow recovery rates.

4. Concluding remarks

This study endeavored to synthesize surface energy and hydrological impacts of beetle-caused tree mortality at stand scales. Observations obtained at GLEES from 2005 to 2007 (prebeetle) and from 2009 to 2010 (postbeetle) were combined with Noah-MP model simulations to assess such impacts.

Observations showed that the most dominant signals following spruce beetle attack were a large reduction in summer daytime evaporation (~27%), a large increase in sensible heat flux (~25%), and a large increase of

Bowen ratio that is nearly double its value prior to beetle-caused mortality.

Simulation results with the Noah-MP model revealed the following:

- By representing the effect of tree mortality on transpiration and surface albedo, the Noah-MP model reasonably captured the observed changes following beetle-caused tree mortality. It shows an $\sim 11\%$ increase in runoff and $\sim 22\%$ reduction in evaporation. Further consideration of leaf loss slightly improved simulated postbeetle surface fluxes and Bowen ratio as compared with observations.
- Considering a thinner canopy in the model further amplified changes in the surface hydrological cycle: greater leaf loss resulted in greater snowmelt, a shortened snow season (by about 10 days), and more runoff. However, the response was not linear in that reducing LAI beyond a certain degree did not substantially change results compared to the simulation using LAI = 3.75.
- Incorporating tree mortality effects enhanced simulated seasonal variations in surface energy and water budgets. Less canopy interception led to greater winter snowpack and more snowmelt and runoff in spring, resulting in more arid soil in summer and fall.
- Several competing factors affect modeled surface energy and water budgets. Lower Rnet tended to increase snow accumulation, but the combination of higher nighttime downward sensible heat fluxes and less longwave radiation losses decreased snow accumulation. Moreover, decreased snow albedo due to litter cover under beetle-killed trees only slightly compensated the effects of leaf loss, not substantially affecting water budgets.

Our study reveals important impacts of spruce beetle-caused tree mortality on the surface energy and hydrological cycles at stand scales. However, generalization to other beetle-infested areas may be difficult because of variations in the fractional coverage and severity of beetle infestations, elevation, tree species, and amount of precipitation and incoming solar energy. The next step is to expand the current tree-stand-scale research to process studies based on observations obtained over disturbed watersheds or at regional scales, as indicated by Edburg et al. (2012). Moreover, this study highlights the need for considering the transient effects of forest disturbances in modeling land-atmosphere interactions and their potential impacts on regional weather and climate.

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