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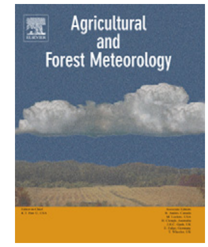

Persistent effects of fire-induced vegetation change on energy partitioning and evapotranspiration in ponderosa pine forests

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ABSTRACT

We compared energy fluxes between a site converted from ponderosa pine (*Pinus ponderosa*) forest to sparse grassland by a severe wildfire 10 years ago and a nearby, unburned forest. We used eddy covariance and associated instruments to measure total radiation, net radiation, albedo, and fluxes of energy into latent heat, sensible heat, and the soil. Total radiation, vapor pressure deficit, and air temperature were similar for each site. Compared to the unburned site, net radiation efficiency (net radiation/total radiation) was 30% lower and albedo 30% higher at the burned site. The magnitude of sensible and latent heats varied seasonally at both sites. Sensible heat was the major component of the energy balance in cold or dry seasons, whereas latent heat was the major component in the warm and wet season. Soil heat flux was the smallest in magnitude of the measured energy fluxes. Compared with the unburned forest, the burn-created grassland generally had lower sensible and latent heats, but greater soil heat flux for both soil cooling in winter and warming in summer. The grassland had similar maximum air temperature as the forest, and warmer surface soil temperature during the summer. Thus, the lower albedo and greater sensible heat of the forest did not produce a warmer site compared with the grassland, apparently because of the cooling effect of greater latent heat in the forest. Our results suggest only small changes in site air temperature, but larger changes in site surface soil temperature by shifts from forest to grassland caused by severe fire in northern Arizona ponderosa pine forests.

1. Introduction

Local climate may be strongly influenced by interactions between large-scale atmospheric circulation and fine-scale spatial variation in land cover (e.g., Diffenbaugh et al., 2005; Went et al., 2007; Bonan, 2008). Changes in land-cover, such as caused by severe wildfire, can act as climate-forcing events by altering vegetation, surface color, albedo, emissivity, and the partitioning of radiant energy between latent and sensible heat fluxes (Thompson et al., 2004; Feddema et al., 2005; Jin and Roy, 2005; Beringer et al., 2007). While partitioning of energy between latent and sensible heat fluxes is thought to be

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an important influence on land-atmosphere coupling of energy exchange (Gu et al., 2006), our understanding of how disturbance modifies components of energy balance is limited by lack of data for most vegetation types (Bonan, 2008).

In the southwestern U.S., wildfire is a common and potentially severe disturbance in forests dominated by ponderosa pine. Over a century of fire suppression and heavy livestock grazing have altered the structure of ponderosa pine forests and disrupted the natural low-intensity surface fire regime throughout the region (Cooper, 1960; Covington et al., 1994; Swetnam and Baisan, 1996; Fulé et al., 1997). Ecosystem structure has shifted from open, savanna-like forests to dense stands of small trees with little understory (Covington et al., 1994; Swetnam and Baisan, 1996; Fulé et al., 1997). Stand-replacing wildfire is now considered inevitable over large areas of southwestern ponderosa pine forests given current high fuel loadings (Covington et al., 1994; Fulé et al., 2004; Moore et al., 2004) and recent climate warming (Brown et al., 2004; Westerling et al., 2006). Large forest fires are increasing in frequency in ponderosa pine forests of the southwestern U.S. (Swetnam and Betancourt, 1998; Westerling et al., 2006) and impact a significant amount of forest area. For example, recent estimates of the area of ponderosa pine forests burned in Arizona and New Mexico over all land ownerships based on the differenced normalized burn ratio indicate that 0.33 million ha of 2.3 million ha of this forest type (O'Brien, 2002, 2003) burned between 1999 and 2007 with 0.034 million ha in the high severity class and 0.07 million ha in the moderate severity class (C. McHugh, USDA Rocky Mountain Research Station, unpublished data).

Vegetation shifts between forests and grasslands or crops can accelerate or slow climate warming. While forests slow climate warming due to carbon sequestration, they may accelerate climate warming compared with other vegetation types due to their lower albedo that produces greater site net radiation and sensible heat (Gibbard et al., 2005; Bonan, 2008). Yet, this potential effect of forests on site energy balance is moderated by the cooling effect of latent heat. The impact of shifts between forests and grasslands on interactions among albedo and energy balance components, and hence forcings and feedbacks on climate are poorly understood for temperate forests in part because of a lack of empirical data for such forests (Bonan, 2008). In this study, we examined effects on the energy balance of the vegetation change from ponderosa pine forest to grassland caused by stand-replacing wildfire in northern Arizona. Over a one-year period, we compared the energy fluxes at a site burned in a stand-replacing wildfire ten years before our study to a similar nearby site that was not burned. We evaluate the hypothesis produced by model simulations of global-scale replacements of forests by grasslands (Gibbard et al., 2005) – grasslands are cooler than forests in the same region – for the change from forest to grassland that commonly occurs when dense ponderosa pine forests burn in the southwestern U.S.

2. Methods

2.1. *Study sites*

Our study compares two sites within the region of northern Arizona that is dominated by ponderosa pine: an unmanaged, undisturbed forest and a forest that burned in a stand-replacing wildfire during the summer of 1996. The sites are located near Flagstaff, AZ, USA, are 35 km apart, and have similar climatic and edaphic conditions (Table 1). The climate of the area is characterized by cold winters, and irregular and moderate precipitation (610 mm, 1977–2007 average, Western Regional Climatic Center, http://www.wrcc.dri.edu/index.html) concentrated as snow during the winter and as rain during the July and August monsoon season. The spring and fall seasons between these precipitation periods are generally dry (Sheppard et al., 2002). The long-term average length of the frost-free season is 94 days.

The unburned site is located in the Northern Arizona University Centennial Forest (35°5′20.5″N, 111°45′43.33″W, elevation 2180 m), and represents a typical stand of ponderosa pine in northern Arizona that has not been disturbed by tree harvest, thinning, or fire for decades, but has been indirectly managed through the policy of fire suppression. Seasonal maximum leaf area index (LAI; projected area) during the study averaged 2.3 $m^2\,m^{-2}$, tree basal area averaged 30 $m^2\,ha^{-1}$, and average tree age was 87 years (Table 1; Dore et al., 2008). The forest is dominated by ponderosa pine in the overstory, with a bunchgrass-dominated understory that includes *Festuca arizonica*, *Elymus elymoides*, *Boutelua gracilis*, and *Blepharoneuron tricholepis*. The seasonal maximum LAI of the understory vegetation was 0.06 $m^2\,m^{-2}$ (Dore et al., 2008).

Table 1 – Mean stand and soil characteristics of the unburned and burned sites (±one standard error). For the burned site, estimates of the pre-fire characteristics were made from adjacent, unburned areas (Dore et al., 2008).

Characteristics	Unit	Unburned	Burned
Total leaf area index	$m^2\,m^{-2}$	2.30 (±0.38)	0.6/before fire 2.4 (±0.45)
Understory leaf area index	$m^2\,m^{-2}$	0.06 (±0.02)	0.60 (±0.17)
Tree density	no. ha^{-1}	853 (±189)	0/before fire 343 (±49)
Basal area	$m^2\,ha^{-1}$	30 (±4.7)	0/before fire 31 (±6)
Canopy height	m	18 m	<0.5 m
Soil type	USDA Soil Taxonomic subgroup	Complex of Mollic Eutroboralfs and Typic Argiborolls	Mollic Eutroboralf
Depth of A horizon	cm	0–5	0–7
Bulk density A horizon	$Mg\,m^{-3}$	1.15	1.01
Depth of B horizon	cm	5–15	7–15
Bulk density B horizon	$Mg\,m^{-3}$	1.15	1.21

Soil characteristics of the site are described in Table 1. Slope at the site is about 2–3%.

The burned site is located on the Coconino National Forest in the 10,500 ha area burned by the Horseshoe-Hochderffer Fires of 1996 (35°26′43.43″N, 111°46′18.64″W, elevation 2270 m). The site is covered by grasses, shrubs, and boles from trees killed by the fire. Before the fire, the burned site had stand characteristics similar to the unburned site (Table 1). Since the 1996 fire, no post-fire management, such as salvage logging or tree planting, has occurred at the site, and the site contains little natural tree regeneration. The vegetation at the site has a single-storied structure, and includes a variety of grasses (e.g., *Bromus tectorum* and *Elymus repens*), shrubs (e.g., *Ceanothus fendleri*), and forbs (e.g., *Oxytropis lambertii*, *Verbascum thapsus*, *Linaria dalmatica*, and *Circium wheeleri*), with a seasonal maximum LAI of 0.6 m^2 m^{-2}. Soil characteristics are described in Table 1. Slope is about 3%. Average ground cover at the burned site, based on cover estimates from four 0.5 m^2 subplots in each of five plots located in the footprint of the eddy covariance measurements, was: rock 18.75% (SE = 2.59), grass 27.0% (SE = 3.72), soil 12.5% (SE = 3.23), litter 16.75% (SE = 2.35), forbs 13.25% (SE = 1.87), coarse woody debris 9.25% (SE = 2.29), and shrubs 2.50% (SE = 1.89).

2.2. Eddy covariance system

We estimated the flux of latent (LE) and sensible (*H*) heats using the eddy covariance technique (Baldocchi et al., 1996) using high frequency (20 Hz) measurements of CO_2, concentration of water vapor, and air temperature above the vegetation canopy of each site. Air was sampled 10 cm from the sonic anemometer (CSAT-3D, Campbell Scientific, Logan, UT, USA) through a 1 μm Teflon filter. A diaphragm pump (KNF mod. N-89, Trenton, NJ, USA) aspired air at 10 l min^{-1} from the intake to the closed-path infra-red gas analyzer (LiCor, LI-7000, Lincoln, NE, USA) through 4 and 9 m (unburned and burned, respectively) of 4-mm diameter Teflon PFA tubing. We used a 27-m tower to install the sensors and to sample air 5 m above the canopy (height of 23 m) at the unburned site. The infrared gas analyzer is located near the top of the tower at the same height of the sampled air at the unburned site. A pole was used to elevate the instruments and sample air 4 m above the ground at the burned site. Power was provided by deep-cycle batteries charged by solar panels.

At the burned site the fetch of homogeneous vegetation in 2006 was about 1 km in the SW–SE and NE directions, and 0.2 km in the E and W directions. The prevailing wind direction at the burned site was from the SE to SW (60% of the time), and the average daytime 70% cumulative footprint (Schmid, 1997) during 2006 was 176 (±3) m. At the unburned site the fetch of homogeneous vegetation in 2006 extended for 1 km from the tower in N and S directions, and 0.5 km from the tower in E and W directions. The prevailing wind direction at the unburned site was from the NE–E (48% of the time) during night and S–SW (47% of the time) during day. The daily average footprint in 2006 at the unburned site was 271 (±4) m.

We acquired and processed the raw data using software designed by Giovanni Manca (Centro di Ecologia Alpina, Italy) following the methods reported by Aubinet et al. (2000). This software has been shown to provide similar values as those produced using the Ameriflux standard (Aubinet et al., 2000). The software incorporated corrections related to the data coordinate rotation, linear detrending, and corrections for flux losses due to sensor separation, time lag, and tube attenuations. Average daytime corrections for 2006 were less than 3.5% for *H* at both sites, and 5% for LE at the unburned site and 8% at the burned site. Thirty-minute flux data were flagged for quality applying the steady state and integral turbulence characteristic test (CarboEurope Spoleto QA/QC Workshop for Eddy Covariance Measurements http://www.geo.uni-bayreuth.de/mikrometeorologie/QC_Workshop/QA_QC_012.f90). Additional parameters used to assess data quality were the variance in scalars, precipitation and number of bad readings in the 30 min intervals.

We gap-filled missing or bad quality LE and *H* data using relationships between Rn and LE, and between Rn and *H*, respectively (Aubinet et al., 2002). The relationship was determined for the seven-day period previous to the gap. The percentage of missing data in 2006 was 0.1% for *H* and 10% for LE at the burned site, and 14% for *H* and 17% for LE at the unburned site. The additional quality assessment of fluxes increased the gap-filled contribution to 10% for *H* and 21% for LE at the burned site, and 26% for *H* and 32% for LE at the unburned site. The percentage of gap-filled data in our study is similar to the 100 site average of 25% for *H* and 31% for LE in Falge et al. (2001).

Storage of fluxes was calculated at the unburned site using a profile that measured CO_2, H_2O and temperature at heights above ground of 1, 8, 16 m in addition to 23 m at the top of the tower. Only the 4 m measurement height was used to calculate storage fluxes at the burned site. Storage fluxes were calculated using the difference in CO_2, H_2O and temperature measured by the profile in 30 min intervals (Morgenstern et al., 2004; Humphreys et al., 2006). Additional information about the eddy measurement system and protocol is in Dore et al. (2008).

2.3. Meteorological sensors

We used dataloggers (CR1000 and CR10X, Campbell Scientific) and multiplexers (AM16/32, Campbell Scientific) to collect data every 30 s. Meteorological instruments were installed at the same height as the eddy covariance system. Data were stored as 30-min means. Collected meteorological data included total and diffuse photosynthetic active radiation (BF3, Delta T, Cambridge, UK), relative humidity, air temperature, atmospheric pressure, wind speed and direction, and precipitation (WXT510, Vaisala, Helsinki, Finland), and incoming and outgoing short- and long-wave radiations and albedo (CNR1, Kipp & Zonen, Delft, The Netherlands).

2.4. Soil measurements

We measured soil heat flux (*G*) at a 8 cm depth in the mineral soil by adding soil heat flux density (W m^{-2}) measured with a soil heat flux sensor (HFP01SC, Hukseflux, Delft, The Netherlands) to soil heat storage (*S*). Soil heat storage (Campbell and Norman, 2000) was estimated from soil temperature (TCAV thermocouples, Campbell Scientific) averaged over a depth of 2–6 cm and volumetric soil water content (measured with two dielectric probes; ECH2O-EC20, Decagon, Pullman, WA, USA) to estimate the soil heat capacity of moist soil. Soil heat flux sensors, thermocouples, and dielectric probes were installed in one representative location within 5 m of the instrument tower at each site. Our estimates of G should be interpreted with caution because of the lack of spatial replication of the soil sensors.

We measured soil temperature (model 107 probes, Campbell Scientific) and soil water content (ECH2O-EC20, Decagon) at depths of 20, 100, 200, and 500 mm in the mineral soil at one representative location near the instrument tower at each site. We calibrated the dielectric probes used to measure soil water content at two different soil depths at each site corresponding to the A and B soil horizons. The calibration consisted of developing equations for soil water content as a function of the sensor voltage output and temperature based on data obtained from reconstructed soil layers equilibrated at different known water contents and temperatures under laboratory conditions (e.g., Bosch, 2004; McMichael and Lascano, 2003; Plauborg et al., 2005). The output voltage from the probes was corrected for variation in temperature with the following equation:

$$V_{corrt} = V + (20 - Ts)(a + bV - cV^2) \quad (1)$$

where V_{corrt} is the output voltage (mV) corrected to 20 °C (the temperature used to determine the relationship between volumetric soil water content and sensor output), V is the output voltage (mV) from the probe, Ts is the soil temperature (°C), and a, b, and c are the regression coefficients. Calibration results from the A and B horizons were similar; thus, only the calibration results from the two sites pooled over soil horizons are presented. Regression coefficients for the unburned site were $a = -10.63$, $b = 0.039$ and $c = -2.77 \times 10^{-5}$, and for the burned site were $a = -12.4$, $b = 0.047$, and $c = -3.72 \times 10^{-5}$. Volumetric soil water content was calculated from the corrected output voltage as:

$$\theta v = V_{corrt}B + A \quad (2)$$

where θv is the volumetric water content ($m^3\ m^{-3}$), and A and B are the regression coefficients (burned site: $A = -0.258$, $B = 7.34 \times 10^{-4}$, $r^2 = 0.95$; unburned site: $A = -0.2623$, $B = 7.07 \times 10^{-4}$, $r^2 = 0.90$).

We estimated water storage (mm) in four soil layers (0–60 mm, 60–150 mm, 150–450 mm, and 450–650 mm) from the vertical profile of water content sensors by assuming that the water content measured at each sample depth (20, 100, 200, 500 mm) was the average for each layer. Water storage to a depth of 650 mm was estimated by summing storage over all layers.

2.5. Data analysis

Data were compared between sites on rain-free days. These days were selected based on having at least 40 half-hour, rain-free values from the total of 48 possible day and night half-hour values in order to have net radiation measured when the sensor was dry. We compared energy fluxes between sites as weekly averages; only weeks with at least three days of data were included in the analysis (Dolman et al., 2002; Kumagai et al., 2004; Kotani and Sugita, 2005). We used all data, including gap-filled data, to compare evapotranspiration ($mm\ d^{-1}$) between sites. We compared albedo between sites only when total solar radiation (Rt) exceeded 100 $W\ m^{-2}$, and only with data collected between 9:00 and 18:00 h to avoid low sun angles (Domingo et al., 2000; Li et al., 2006). We used Sigma plot (ver. 10.0, Systat Software, San Jose, CA, USA) and JMP (ver. 5.1, SAS Institute, Cary, NC, USA) for regression and correlation analyses and used t-test to compare regression coefficients between sites.

3. Results

3.1. Environmental conditions

Both sites had low precipitation between January and March, moderate precipitation in April, little precipitation in May and June, a wet period between July and October, and little precipitation thereafter (Fig. 1A). Annual precipitation was 35% higher at the unburned site (696 mm) than the burned site (517 mm) primarily because of larger rain events during July and August (Fig. 1A).

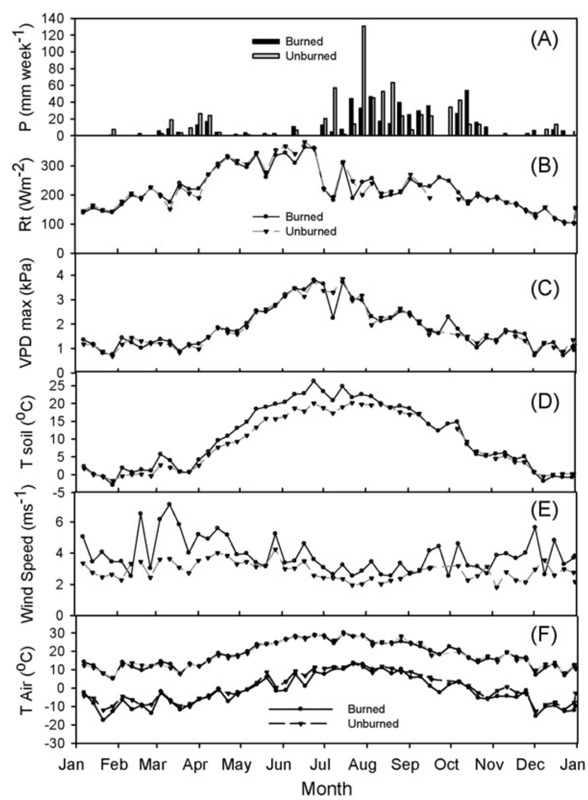

Fig. 1 – Weekly values of climatic variables at the burned and unburned sites for calendar year 2006: (A) weekly total precipitation (P); (B) average weekly total solar radiation (Rt); (C) average weekly maximum vapor pressure deficit (VPD); (D) average weekly soil temperature at a mineral soil depth of 20 mm (Tsoil); (E) average weekly wind speed; and (F) average weekly maximum and minimum air temperatures.

Total solar radiation (Rt) at both sites increased from winter to mid-summer, and then decreased during the rainy monsoon season in July to September due to increased cloud cover (Fig. 1B). Total solar radiation was similar for the two sites on most dates. Maximum vapor pressure deficit (VPD) was also similar at the two sites, with seasonal variation in maximum VPD similar to air temperature (Ta) (Fig. 1F) and Rt (Fig. 1C). Soil temperature (20-mm mineral soil depth) was consistently higher at the burned site than the unburned site from April through September, with a maximum difference of about 7 °C in June (Fig. 1D). Wind speed was consistently higher at the burned site than the unburned site (Fig. 1E). Maximum daily air temperature 4 m above the plant canopy was almost identical at the two sites, whereas the minimum daily air temperature was 3–4 °C lower at the burned site than the unburned site on several dates (Fig. 1F).

3.2. *Net radiation and albedo*

The two sites had similar Rt on most dates (Fig. 1B), but radiation efficiency, the net radiation (Rn) produced per unit of incoming total solar radiation (Iziomon and Mayer, 2002; Chambers et al., 2005), was consistently higher at the unburned than the burned site (Fig. 2). Radiation efficiency at the unburned site was 0.50–0.70 for much of the summer, while radiation efficiency at the burned site was 0.30–0.50.

Albedo was consistently greater at the burned than the unburned site (Fig. 3). During the summer, albedo ranged between 0.13 and 0.15 at the burned site, and between 0.09 and 0.10 at the unburned site. Temporal variation in albedo was strongly associated with variation in snow cover, thus the highest albedo at both sites occurred in late March and December (Fig. 3). Albedo was lowest during the rainy season in August (Fig. 3). Albedo in the summer decreased significantly as surface soil (20 mm depth) water content increased, and the slope of the decrease was similar ($P > 0.05$) for each site (Fig. 4).

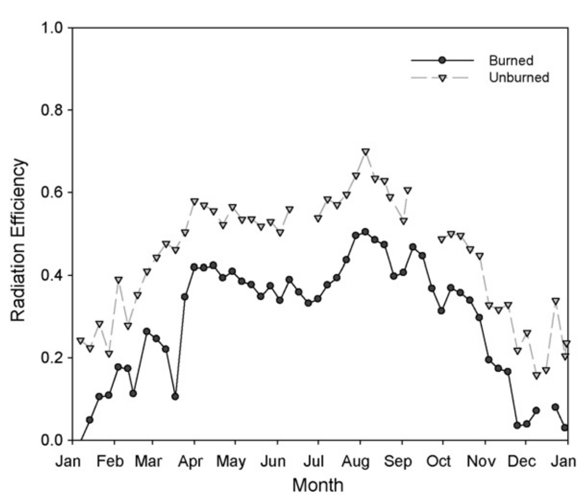

Fig. 2 – Average weekly radiation efficiency, defined as net radiation divided by total radiation, for the burned and unburned sites during calendar year 2006.

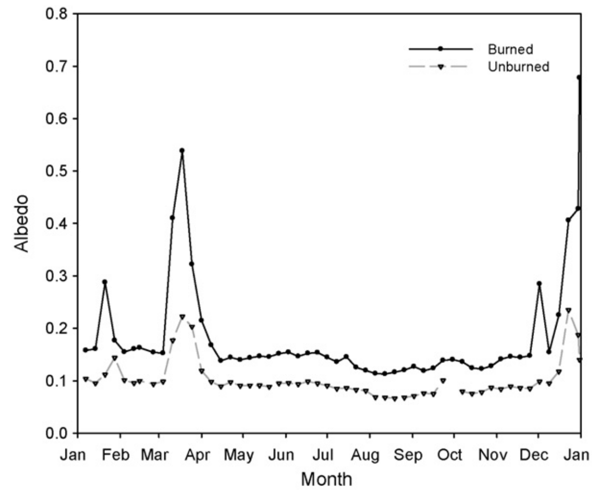

Fig. 3 – Average weekly daytime (9:00–18:00 h) surface albedo for the burned and unburned sites during calendar year 2006.

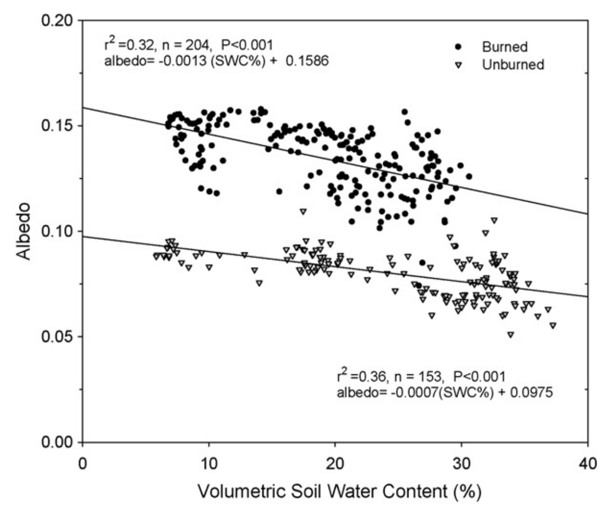

Fig. 4 – Relationship between average daily volumetric soil water content at 20 mm depth and average daily albedo for the burned and unburned sites for all days during the summer in calendar year 2006.

3.3. *Energy flux partitioning*

Fig. 5 shows average weekly energy fluxes when at least three days in the week were rain free. During winter, maximum Rn was about 2 MJ m^{-2} d^{-1} at the burned site and 3 MJ m^{-2} d^{-1} at the unburned site (Fig. 5A). During the summer, maximum Rn was about 11 MJ m^{-2} d^{-1} at the burned site and about 16 MJ m^{-2} d^{-1} at the unburned site (Fig. 5A). Most variation in average weekly Rn was explained by the summed values of LE, *H*, and *G* at both sites with little evidence of consistent over- or under-estimation of the energy closure. The linear regression of Rn versus the sum of LE, *H*, and *G* had an r^2 of 0.99 ($n = 33$ weeks, $P < 0.001$) at the unburned site, and 0.82 ($n = 26$ weeks, $P < 0.001$) at the burned site.

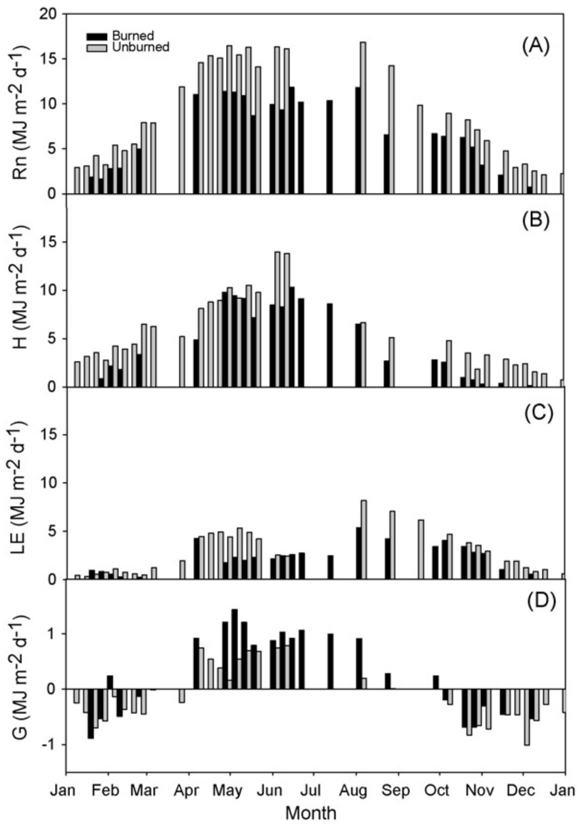

Fig. 5 – Average weekly energy fluxes for the burned and unburned sites during calendar year 2006. Fluxes are: net radiation (Rn, panel A), sensible heat (*H*, panel B), latent heat (LE, panel C), and soil heat flux (G, panel D). Rainy days were excluded from the weekly averages of LE, *H*, and G. Note the different y-axis scale for G in panel D.

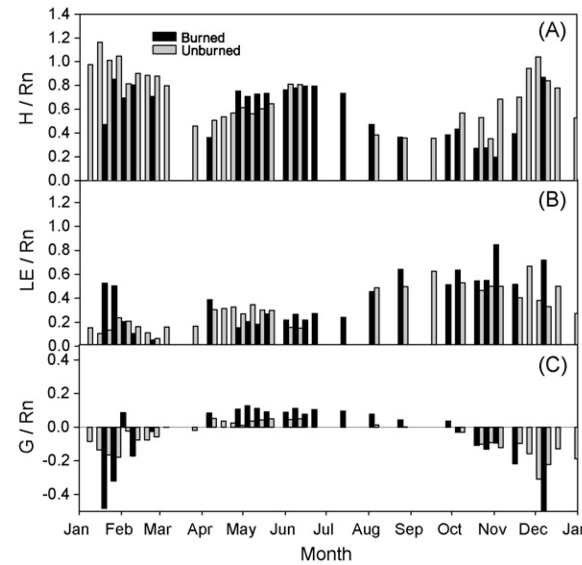

Fig. 6 – Ratios of sensible heat (*H*, panel A), latent heat (LE, panel B), and soil heat flux (G, panel C) to net radiation (Rn) at the burned and unburned sites in calendar year 2006. Note the different y-axis scale for G in panel C.

Sensible heat (*H*) was the largest energy flux at both sites (Fig. 5B). Temporal variation in *H* was strongly associated with variation in Rn for the unburned ($r^2 = 0.84$, $n = 33$, $P < 0.01$) and burned sites ($r^2 = 0.81$, $n = 26$, $P < 0.001$). Sensible heat was typically greater at the unburned than the burned (Fig. 5B) site.

Latent heat (LE) was the second largest energy flux on most dates (Fig. 5C). Temporal variation in LE was positively correlated with variation in Rn only at the unburned site ($r^2 = 0.33$, $n = 33$, $P = 0.007$). Latent heat was consistently greater at the unburned than the burned site between April and October (Fig. 5C).

The relative importance of LE to the energy balance differed between sites depending on the season. In mid winter (January and February), LE was a greater proportion of the energy balance at the burned than the unburned site (Fig. 6B). In the dry month of May, LE was a greater proportion of the energy balance at the unburned than the burned site (Fig. 6B). The proportional contribution of LE to the energy balance was similar for both sites in the wetter months of August, September, and October based on a limited number of weeks of data. Similar to the earlier winter months of January and February, LE was a greater proportion of the energy balance at the burned site than the unburned site in November and December (Fig. 6B). Temporal variation in the relative importance of *H* to the energy balance at each site (Fig. 6A) was about the opposite of the pattern for LE (Fig. 6B).

During the dry and warm months of April, May, and June, *G* was positive at both sites, indicating soil warming (Fig. 5D). The wetter months of August and September were associated with a decrease in *G* at both sites based on a small number of weeks of data. In contrast, *G* was negative during the winter months at both sites, indicating soil cooling (Fig. 5D). Soil heat flux was a greater component of the energy balance at the burned site than the unburned site for periods of both soil cooling and warming (Fig. 6C).

3.4. *Evapotranspiration*

Evapotranspiration (ET) estimated by eddy covariance was low at both sites during winter (Fig. 7A). Evapotranspiration increased similarly at both sites during March and April. Between mid-April and December, ET was generally greater at the unburned than the burned site, with the largest differences in May and July (Fig. 7A). Exceptions to this pattern occurred in June at the end of the seasonal drought (Fig. 7C) and in late-September and early October at the end of the rainy season (Fig. 7C), when ET was similar at both sites (Fig. 7A). Maximum ET was 2.4 mm d^{-1} at the unburned site (August) and 2.0 mm d^{-1} at the burned site (September).

Soil water storage was more temporally dynamic at the unburned site than the burned site (Fig. 7B) due to both greater precipitation during the summer (Fig. 7C) and higher ET at the unburned site (Fig. 7A). Precipitation during March and April increased soil water storage (650 mm soil depth) from 40 mm to 100 mm at the unburned site, and from about 60 mm to 110 mm at the unburned site (Fig. 7B). Consistent with the higher ET at the unburned site (Fig. 7A), soil water was depleted faster at the unburned than at the burned site during

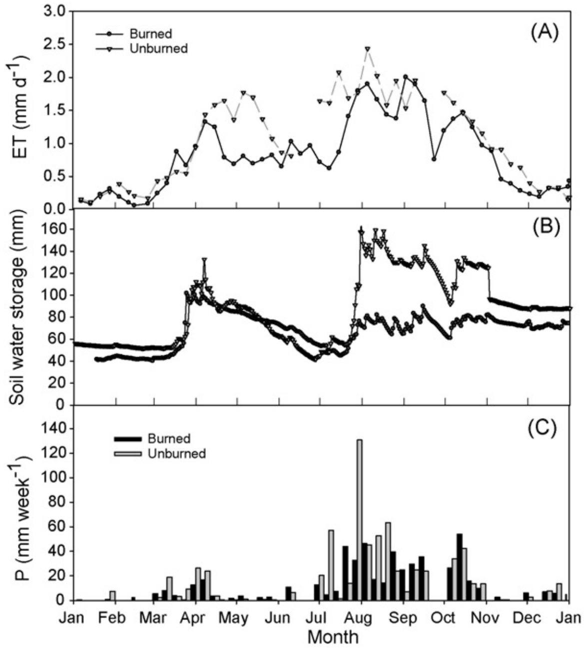

Fig. 7 – Weekly evapotranspiration (ET) calculated from the eddy covariance system (A), soil water storage (mm) to a depth 650 mm (B), and weekly total precipitation (C) for the burned and unburned sites in calendar year 2006.

the dry months of May and June (Fig. 7B). The greater precipitation in July and August at the unburned site (Fig. 7C) increased soil water storage more than at the burned site (Fig. 7B). Similar to May and June, depletion of soil water was faster at the unburned than the burned site in late-September and October following summer rains, and in November following October rains (Fig. 7B).

4. Discussion

Multiple lines of evidence suggest that the two sites in our study were similar prior to the fire that burned one site in 1996. First, tree basal area and LAI were almost identical at the two sites prior to the fire (Table 1). Second, soils at both sites are Mollic Eutroboralfs (or a very similar soil type, Typic Argiborolls), with similar thicknesses of A and B horizons. Third, vapor pressure deficit, maximum air temperature, and total radiation were similar for each site during our measurements in 2006. Thus, we believe that our reported differences in energy balance between sites were caused by the fire and associated vegetation change from forest to grassland, rather than differences that existed before the fire.

The wildfire that burned at our study site in 1996 caused pronounced changes in stand structure, vegetation, and ground cover. The fire killed most overstory trees over much of the 10,500 ha burned area, and converted the vegetation from a dense pine forest to sparse cover by herbaceous plants and a few shrubs. Tree regeneration has not occurred since the fire burned 10 years ago, and likely will not occur for many decades because of limited seed source, climatic constraints on seedling establishment caused by the characteristic spring drought of the region (Savage et al., 1996), and competition between tree seedlings and herbaceous plants (Long and Wagner, 1992). The fire also altered the ground cover by increasing coarse woody debris by almost four fold (Dore et al., 2008). The structural impacts of wildfire at our study site are typical of contemporary wildfires that burn during dry, windy conditions in late spring and early summer in dense ponderosa pine stands in the southwestern U.S. (e.g., Finney et al., 2005). Similar effects of severe fire on stand structure and vegetation have been reported for other forest types (Chambers et al., 2005; Amiro et al., 2006).

Fire-induced reductions in plant biomass and leaf area, and increases in bare soil and light-colored coarse woody debris (Dore et al., 2008), increased the albedo at the burned site. Greater albedo at the burned site reduced conversion of incoming short-wave radiation into net radiation compared to the unburned site. For example, in mid summer about 40% of total radiation was converted into net radiation at the burned sited compared with about 60% at the unburned site (Fig. 2). Thus, the fire-induced vegetation change reduced the amount of energy available for heating the air (*H*), soil (*G*), and evaporating water (LE).

Seasonal variation in albedo was pronounced and the temporal pattern was similar at both sites (Fig. 3). Albedo peaked when the ground was snow covered, and quickly decreased as snow melted. Temporal variation in albedo in the summer was associated with variation in soil water content (Fig. 4). The decrease in albedo as soil water content increased was likely due to the darkening of soil and plant surfaces (Domingo et al., 2000; Li et al., 2006) and an increase in understory leaf area in the wet season (Thompson et al., 2004) based on results from other ecosystem types.

The relative importance of LE, *H*, and *G* in the energy balance varied seasonally at both sites (Figs. 5 and 6). Sensible heat was the most important component during the dry months of January to July when it accounted for 50–90% of Rn depending on date and site. In contrast, LE increased in importance during the wetter months of August to December when it often accounted for 50–70% of Rn. Soil heat flux was the smallest component of Rn at both sites and usually was less than 20% of Rn. Temporal variation in *G* consisted of soil cooling in the winter and warming in the spring and summer.

Differences between sites in the importance of LE and *H* to Rn were not consistent temporally (Fig. 6) and appeared to depend on water and temperature constraints on plant growth and transpiration. For example, in January and February LE was a more important component of Rn at the burned site than the unburned site. Energy flux into LE in winter at the unburned site was likely constrained by low-temperature limitations on stomatal conductance of ponderosa pine (Gaylord et al., 2007) and less energy available to directly evaporate soil water compared with the direct exposure of soil to Rt at the burned site. Warmer and wetter conditions in April and May increased LE at the unburned site, likely via increased transpiration by the forest canopy, and LE became a more important component of Rn at the unburned than the burned site. Low soil moisture content and high VPD in June to July likely reduced canopy stomatal conductance (e.g., Simonin et al., 2006; Gaylord et al., 2007) and reduced LE more at the higher-LAI unburned site than the burned site, resulting in a more similar proportional contribution of LE to Rn at each site during this time period. The relative importance of LE to Rn was similar for each site in the wetter months of August to October when heavy rains provided ample water for both soil evaporation and plant transpiration. Similar to the pattern for January and February, cold weather in November and December was associated with a decrease in relative importance of LE to Rn at the unburned site likely due to stomatal closure in the forest canopy. The surprisingly high importance of LE to the energy balance at the sparsely vegetated burned site (peak projected LAI = 0.6) in the wet months can be explained by a high rate of soil evaporation in moist, fine-textured soils in northern Arizona (Springer et al., 2006).

Our results suggest that conversion of vegetation from forest to grassland by severe fire in ponderosa pine forests reduces the magnitude of *H* and LE. Despite temporal variation in the proportion of Rn partitioned to *H* and LE between sites, absolute values of *H* and LE were usually greater at the unburned forested site than at the burned grassland site (Fig. 5). Reduction of *H* and LE that occurs when southwestern ponderosa pine forests are converted to grassland by severe forest fires may reduce convection and modify regional precipitation (Hoffmann and Jackson, 2000; Beringer et al., 2003, 2005; Jin and Roy, 2005).

Was the lower ET of the burned site compared with the unburned site due to a difference in the amount of water available for ET or to other differences between the sites caused directly by the conversion of forest to grassland by the fire? Whereas soil water storage often differed between sites, precipitation in March and April recharged water storage to similar levels at both sites (Fig. 7B). Evapotranspiration and depletion of soil water storage were lower at the burned than the unburned site in May and June following this recharge event. This comparison suggests that lower ET at the burned site was due to factors other than water availability, such as lower Rn (driven by higher albedo) and lower LAI at the burned site than the unburned site. Our finding that a large reduction in tree leaf area caused by severe fire reduced ET, and consequently increased soil water storage, is consistent with several experimental studies in other northern Arizona ponderosa pine forests that substantially reduced tree leaf area by thinning (Kolb et al., 1998; Feeney et al., 1998; Simonin et al., 2006; Hart et al., 2006); large reductions in tree leaf area increased soil water content during dry periods when plant demand for water was high.

The change in vegetation from forest to grassland caused by the fire altered the magnitude and temporal variation of partitioning of Rn to G and warmed the site surface. The open, unshaded condition of the burned site resulted in greater absolute (Fig. 5) and relative (Fig. 6) partitioning of Rn to *G* compared with the unburned site for both winter cooling and summer warming of the soil. Consequently, the surface soil of the burned site was slightly cooler in winter (about 1 °C) and much warmer (3–7 °C) in summer compared with the unburned site (Fig. 1D). These results suggest that vegetation change from forest to grassland caused by severe fire in pine forests of northern Arizona may have long-term effects on soil processes that strongly depend on soil temperature, such as evaporation (Morecroft et al., 1998; Suleimann and Ritchie, 2003) and CO_2 efflux (Hart, 2006; Hart et al., 2006; Sullivan et al., 2008).

We evaluated the hypothesis that conversion of forest to grassland caused by the severe fire would result in a cooler site because the cooling effect of higher albedo of the grassland would be more important than the warming effect of lower ET of the grassland. Whereas this hypothesis was supported by model simulations at the global scale (Gibbard et al., 2005), it was not completely supported by our empirical data for two sites in northern Arizona. Consistent with the hypothesis, conversion from forest to grassland increased albedo and reduced Rn. Inconsistent with the hypothesis, the higher albedo at the grassland was associated with warmer surface soil in the summer because *G* was a more important component of the energy balance at the grassland compared with the forest. Maximum air temperature above the plant canopy was similar at each site, suggesting that the warming effect of higher Rn and *H* at the forest site was offset by the cooling effect of higher LE. Hence, our results suggest only small changes in air temperature, but larger changes in surface soil temperature by shifts from forest to grassland at our study sites.

5. Conclusions

Severe fire in ponderosa pine forests of northern Arizona often shifts the dominant vegetation from forest with high LAI to sparse grassland with low LAI. Our study suggests that this vegetation shift changes energy and water balances for at least one decade after burning. Lower use of energy and water for ET in grasslands created by severe fire can increase soil water content and potentially the amount of water available for aquifer recharge compared to forests, at least in landscapes with low slope and soil erosion such as our study sites. The fire-created grasslands have a higher albedo and lower Rn than unburned forests, but also have warmer surface soil in the summer because low LAI limits the cooling effect of LE. Given the increasing size and severity of wildfires in the western U.S. (Westerling et al., 2006) and the expected reduction in snowpack in this region (Hamlet et al., 2005), surface soil warming due to wildfire that we describe here should become more pronounced in coming decades.

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