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Spatial and temporal evapotranspiration trends after wildfire in semi-arid landscapes

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ABSTRACT

In recent years climate change and other anthropogenic factors have contributed to increased wildfire frequency and size in western United States forests. This research focuses on the evaluation of spatial and temporal changes in evapotranspiration (ET) following the 2011 Las Conchas Fire in New Mexico (USA) using the Operational Simplified Surface Energy Balance Model (SSEBop ET). Evapotranspiration is coupled with soil burn severity and analyzed for 16 watersheds for water years 2001–2014. An average annual decrease of 120 mm of ET is observed within the regions affected by the Las Conchas Fire, and conifers were converted to grassland a year after the fire. On average, the post-fire annual ET in high, moderate, and low burn severity is lower than pre-fire ET by approximately 103–352 mm, 97–304 mm, and 91–268 mm, respectively. The ratio of post-fire evapotranspiration to precipitation (ET/P) is statistically different from pre-fire conditions ($\alpha = 0.05$) in nine of the watersheds. The largest decrease in ET is approximately 13–57 mm per month and is most prominent during the summer (April to September). The observed decrease in ET contributes to our understanding of changes in water yield following wildfires, which is of interest for accurately modeling and predicting hydrologic processes in semi-arid landscapes.

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1. Introduction

Forests cover approximately 42 million km^2 (or 30% of the land surface) in tropical, temperate, and boreal lands (Bonan, 2008) and are critical regulators in the terrestrial water balance through evapotranspiration (ET). Evapotranspiration is an energy-water flux that redistributes water from the soil and plant surfaces (evaporation) and vegetation (transpiration) into the atmosphere (Allen et al., 2005) and is sensitive to vegetation characteristics and climate. In temperate eucalypt forests and semiarid forests, annual ET accounts for 70–80% and 85–100% of precipitation, respectively (Vertessy et al., 2001; Yaseef et al., 2010; Mitchell et al., 2012). While in loblolly pine (*Pinus* taeda) plantations in the southeastern U.S. and ponderosa pine (*Pinus ponderosa*), approximately 70% (Sun et al., 2002) and 85%, respectively, of annual precipitation contributes to ET (Montes-Helu et al., 2009). Evapotranspiration influences climate through energy and water exchange, making forests essential for ecological and watershed services such as flood control, water quality, and carbon and nitrogen storage.

Disturbances in forests, such as wildfires, floods, landslides, and clearcutting can promote healthy tree density, herbaceous and shrub production, water availability and runoff, nutrient availability, soil characteristics, and wildlife habitat (Covington and Moore, 1994). However, global climate changes, largely due to anthropogenic activities, have led to variations in temperature and precipitation regimes, altering natural wildfire patterns, drought, and other forest disturbances (Dale et al., 2001). In the western U.S., climate change contributes to longer dry summer seasons and drier vegetation, and extended wildfire durations and seasons, which subsequently leads to more frequent and intense wildfires. (Dale et al., 2001; Westerling et al., 2006). Invasive species can also impact ecosystem processes and change vegetation composition, which impacts fuel properties and fire regimes such as intensity, frequency, and seasonality of wildfires (Brooks et al., 2004). Following wildfires, lower ET is expected from the loss of foliar volumes and transpiration processes. Also, the replacement of deep-rooted high profile trees by low profile grasses may contribute to lower ET (Neary et al., 2005). The reduced ET means less precipitation is converted into vapor, and thus more water generally is available for streamflow. Given the large role of ET in the water budget, a disturbance can shift the partitioning between dominant water fluxes such as infiltration and surface runoff. Thus, accurate estimates of

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ET are necessary for climate studies, ecosystem modeling, and agriculture and hydrologic applications, especially in regions experiencing acute or chronic changes in land cover.

Post-fire watershed processes in various ecosystems have been documented, including altered sediment yield (Keller et al., 1997; Lane et al., 2006), soil moisture (Ebel, 2013), peak streamflow (Veenhuis, 2000), base flow (Kinoshita and Hogue, 2011), snow accumulation and melt patterns (Neary et al., 2005; Micheletty et al., 2014) and surface water yield (Kinoshita and Hogue, 2015). Bart (2016) noted a 134% increase of streamflow during the first year after fire in 12 watersheds in southern California. Similarly, Kinoshita and Hogue (2015) observed an increase in low streamflow volume (118%–1090%) following the 2003 Old Fire in southern California. The enhanced baseflow in the chaparral dominated watersheds was attributed to the reduction of basin-wide transpiration rates due to the loss of vegetation (Kinoshita and Hogue 2011; 2015). In New Mexico, Wine and Cadol (2016) observed various post-fire streamflow responses in three semi-arid watersheds that were impacted by monsoon storms from 1982 to 2014. The runoff in the Gila watershed (92% burned) was 12–16 times greater than unburned watersheds for 3–5 post-fire years (Wine and Cadol, 2016). The reduced transpiration and increased infiltration-excess overland flow following monsoon storms contributed to increased streamflow. Some watersheds did not show an increased post-fire water yield, which was attributed to the inaccuracy of runoff measurements, increased transpiration by remaining plants, and increased evaporation as a result of reduced tree canopy (Wine and Cadol, 2016). Similarly, Feikema et al. (2013) did not observe an increase of initial post-fire streamflow due to reduced canopy cover, which highlights the importance of antecedent soil moisture conditions. During a drought, low rainfall can result in soil water deficits. For example, when soil moisture deficits are very large or very small (Feikema et al., 2013; Bart, 2016), surface runoff may not be impacted by changes in evapotranspiration. Benavides-Solorio and MacDonald (2001) noted that low runoff may be due to high antecedent soil moisture or dry soils, with a higher capacity for infiltration, even in high burn severity regions. Based on the limited literature, Wine and Cadol (2016) noted the importance of continuing research on the impacts of wildfire on evapotranspiration to further understand post-fire hydrologic and vegetation processes and improve the prediction of streamflow responses.

There is limited research focused on post-wildfire ET. Post-fire ET studies by Montes-Helu et al. (2009) and Dore et al. (2010) observed CO_2 and ET using eddy covariance data applicable for an area of 176 (±3) m (burned) and 271 (±4) m (unburned) during 2006 in Arizona after a severe fire in 1996. After the fire, the ponderosa pine became primarily grassland and shrub with 40% vegetation coverage, 50% bare soil, and 10% logs and snags (Montes-Helu et al., 2009). Dore et al. (2010) noted that post-fire ET was lower for 10–11 years after the fire by 76–116 mm per year. Another field-based study in a mixed eucalypt forest located in the state of Victoria, Australia, observed that ET in high and moderately burned forest are 41% and 3% lower, respectively, over the first two post-fire years (Nolan et al., 2014). Utilizing remote sensing techniques, Sánchez et al. (2015) applied a two-source energy balance approach on two sets (6–7 years and 11 years after the fire) of satellite-based scenes (a total of 15 Landsat 5 Thematic Mapper and Enhanced Thematic Mapper Plus images) to estimate latent heat flux (heat associated with surface evaporation) after a wildfire in Spain. Sánchez et al. (2015) noted a decrease of approximately 1 mm per day of ET in unburned pine forest, 6–7 years after fire. This decrease was attributed to altered vegetation structure, land surface temperature, and albedo (Sánchez et al., 2015). Dore et al. (2010) and Nolan et al. (2014) provided important insight into point-scale evaporative processes after wildfire, while Sánchez et al. (2015) provided a higher spatial resolution study with limited temporal resolution.

This research builds upon previous studies by incorporating remote sensing products to quantify post-fire evapotranspiration at higher temporal and spatial scales. Evapotranspiration before and after the 2011 Las Conchas Fire in New Mexico, United States was investigated utilizing the Operational Simplified Surface Energy Balance Model (SSEBop). Pre- and post-fire ET patterns were compared with spatially distributed precipitation and soil burn severity for water years (WY) 2001 to 2014 (October 1, 2000 to September 31, 2014). This research addresses the following questions, which also provides the structural sub-headings used in the following Methods, Results, and Discussions sections:

I. How well does SSEBop ET represent pre- and post- fire processes?
II. How does annual and seasonal ET change after wildfire?
III. How does pre- and post- fire ET vary spatially and temporally with respect to soil burn severity?

2. Data and study site

2.1. Spatial actual evapotranspiration model

Single source energy balance models such as the Surface Energy Balance Algorithm for Land (SEBAL; Bastiaanssen et al., 1998) and Mapping Evapotranspiration at high Resolution with Internalized Calibration (METRIC; Allen et al., 2007) require manually selected hot and cold reference pixels to compare climatic conditions for one satellite image. The Simplified Surface Energy Balance Index (SSEB-I) and Operational Simplified Surface Energy Balance Model (SSEBop; Senay et al., 2013) omit the calculation of sensible heat flux to reduce the complexity of determining actual ET. SSEBop also assumes that each pixel has pre-defined hot and cold reference values, given that reference points are stationary or the differences are small in relation to the accuracy level obtained through varying boundary conditions (Senay et al., 2013). This approach requires a pre-defined satellite-derived temperature difference (dT), which is based on net radiation, air resistance, Normalized Difference Vegetation Index (NDVI), and air temperature (2–4) (Senay et al., 2013). The ET fraction (ETF) is calculated from hot, cold, and observed surface temperatures (5). Finally, SSEBop actual ET (ET_a) is estimated by multiplying the Penman-Montieth reference ET (ET_o (1)) and the ETF (6). The Penman-Montieth (Monteith, 1965) equation is expressed as:

$$ET_o(\mathrm{mm/day}) = \frac{\Delta(R_n - G) + \left[\rho_a C_p \frac{(e_s - e_a)}{r_a}\right]}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{1}$$

where R_n = clear sky net radiation (W m^{-2}); G = soil heat flux (W m^{-2}); $(e_s - e_a)$ = the vapor pressure deficit of the air (Pa); ρ_a = mean air density (kg m^{-3}); c_p = specific heat of the air (J kg^{-1} K^{-1}); = slope of the saturation vapor pressure temperature relationship (Pa K^{-1}), γ = psychrometric constant (~66 Pa K^{-1}), r_s and r_a are the (bulk) surface and aerodynamic resistances (s m^{-1}). The SSEBop ET algorithm consists of (Senay et al., 2013):

$$T_h = T_c + dT \tag{2}$$

$$dT = \frac{R_n \times r_{ah}}{\rho_a \times C_p} \tag{3}$$

$$T_c = T_h \times c \tag{4}$$

$$ETF = \frac{T_h - T_s}{T_h - T_c} \tag{5}$$

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