Remote Sensing of Environment 193 (2017) 244–256

Contents lists available at ScienceDirect

Remote Sensing of Environment

journal homepage: www.elsevier.com/locate/rse

Examination of the impacts of vegetation on the correlation between snow water equivalent and passive microwave brightness temperature

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ARTICLE INFO

Article history:
Received 23 October 2016
Received in revised form 18 February 2017
Accepted 9 March 2017
Available online xxxx

Keywords:
Remote sensing of snow
Passive microwave radiometry
Snow water equivalent
Vegetation
Sierra Nevada

ABSTRACT

Snow accumulation, ablation, and runoff in mountainous areas are critical components of the hydrologic cycle, but are poorly known. Passive microwave (PM) measurements are sensitive to snow water equivalent (SWE), even in mountain regions, but vegetation masks the microwave signals and reduces this sensitivity. This study examines how the PM snow signal is affected by the forest density in fourteen basins in the Sierra Nevada, USA, and in a series of sixteen subsets of the Kern basin that have varied vegetation density. 36.5 GHz vertical polarization brightness temperature (T_b) time series for each basin were produced from the spaceborne AMSR-E operational period (water year (WY) 2003 to WY2011). For each basin, the coefficient of determination (R^2) between the annual minimum T_b and the concurrent SWE was calculated to evaluate the sensitivity of the PM to SWE. The relationship between the R^2 values and the forest density was then analyzed to assess how vegetation affect the SWE information in the observed T_b. Mean forest coverage from MODIS was used to represent forest density. The R^2 between the annual minimum T_b and concurrent SWE was >0.6 for three of the basins. Consistent with previous studies, WY2006 demonstrated anomalous T_b values for many basins, apparently due to anomalous warm winter rainfall. Excluding WY2006, R^2 is significantly higher in all basins: eight of fourteen basins have R^2 >0.6. For basins with average elevation >2500 m, SWE correlates well with T_b. The R^2 decreases monotonically with decreasing elevation. Basin elevation and forest cover are highly correlated in the Sierra; a basin elevation of 2500 m generally coincides with forest cover of 20%. A total of 42% of Sierra Nevada has <20% forest cover; this corresponds to an estimated 34% of the total SWE in the Sierra. Thus, SWE and T_b are empirically correlated for over a third of the SWE in the Sierra, typically at high elevations and above treeline.

1. Introduction

Many mountainous areas globally serve as water towers, storing precipitation in winter snowpacks and releasing in spring and summer as snowmelt runoff when ecosystem and agricultural demands are greatest (Viviroli et al., 2007). However, the water resources role of mountainous areas will likely be altered by climate change (Viviroli et al., 2011). Indeed, high-elevation regions are experiencing greater rates of warming than the global average (Pepin et al., 2015). Observation networks are not dense enough in global mountains to fully resolve snow processes, which vary dramatically in space and time. In the Sierra Nevada, for example, automated snow sensors have a density of approximately 1 per 700 km^2, whereas SWE varies at scales on the order of 1 m (Guan et al., 2013). In many other mountain ranges, little or no information on SWE is collected or shared; this limits scientific ability to characterize mountain hydrology globally, at a time of climatic change when such understanding is critical. Satellite remote sensing is attractive, in this regard; however, estimating SWE from spaceborne platforms is still problematic in mountainous areas (Lettenmaier et al., 2015).

Dozier et al. (2016) reviewed remote sensing technologies utilized to study mountain SWE. In mountainous areas, passive microwave (PM) retrievals are plagued by a number of challenges including i) the coarse spatial resolution of PM is often inadequate to resolve the snow spatial variability in mountainous regions, ii) the confounding effect of snow grain size and stratigraphy on the relationship between T_b and SWE (Durand et al., 2008), iii) the influence of forest cover on reducing sensitivity of the top-of-atmosphere T_b to SWE (Vander Jagt et al., 2013), and iv) the so-called saturation effect. The T_b-SWE relationship is usually said to saturate after snow depth exceeds a certain threshold (e.g. 8–80 cm in Chang et al., 1987, 20–25 cm in Mätzler, 1994, 15 cm in Dong et al., 2005, 13 cm in Derksen et al., 2010), beyond which point T_b has been shown to be relatively invariant or increase (e.g.

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http://dx.doi.org/10.1016/j.rse.2017.03.006
0034-4257/

Derksen et al., 2010) with increasing SWE in many cases. Thus, inference of SWE from T_b in mountainous regions where SWE often exceeds the saturation threshold has often been assumed to be infeasible.

Recent work seems to indicate that passive microwave measurements may still have some utility in mountainous areas. Regarding issue i) in the previous paragraph, Vander Jagt et al. (2013) showed that T_b should theoretically be sensitive to the mean SWE within passive microwave footprints, despite its coarse spatial resolution. Regarding ii), models of stratigraphy and grain size have advanced significantly (e.g. Morin et al., 2013), improving the viability of radiance assimilation methodologies (e.g. Durand et al., 2009). Regarding iii), Vander Jagt et al. (2015) showed that spaceborne T_b is expected to be sensitive to SWE in areas with low forest cover, e.g. above treeline in mountainous areas. SWE retrieval in mountainous areas in the presence of forest cover has been demonstrated using a radiance assimilation scheme in theory (i.e. using synthetic observations; Durand and Margulis, 2007) and in practice (Dechant and Moradkhani, 2011). Critical to radiance assimilation in the presence of forest cover is accurate estimates of vegetation transmissivity. Langlois et al. (2011) illustrated an intriguing method for correcting spaceborne T_b for the effect of vegetation in the (non-mountainous) boreal forest of northern Quebec province, Canada; this correction has not yet been attempted in mountainous areas.

Regarding the saturation issue: Li et al. (2015a, 2015b) have highlighted that new snowfall leads to T_b changes for SWE greater than the typical saturation depths, but that these changes do not fit the model utilized by the spectral difference retrieval algorithms. Indeed, synthetic and real radiance assimilation results have illustrated retrieval of SWE in mountainous regions well past the typical saturation SWE values at the point scale (Durand and Margulis, 2006; and Durand et al., 2008) and at basin scale (Durand and Margulis, 2007; Dechant and Moradkhani, 2011; Li et al., 2017). Li et al. (2012) found an intriguing empirical correlation in the largely-unforested upper part of the Kern River watershed (1276 km^2) in the Sierra Nevada. Referred to as L12 hereafter, the study related the minimum T_b of each water year with the SWE at the time of the minimum T_b, and showed high interannual correlation between SWE and T_b, for SWE up to 75 cm, far beyond typically-cited SWE saturation values. Note that the minimum T_b values tended to occur late in the snow season, after the onset of melt-refreeze cycles, where snow is typically wet during the daytime, and refreezes at night, leading to rapid grain growth, and corresponding low T_b. The effect of melt-refreeze cycles on T_b has been documented using in situ measurements (e.g. Mätzler and Wiesmann, 1999; Macelloni et al., 2005), and utilized to characterize snowpack melt state from space (e.g., Ramage and Semmens, 2012). L12 hypothesized that the correlation between SWE and nighttime T_b beyond the saturation depth was due to the spatial variability of snow cover within the passive microwave footprints near the peak snow season. In other words, while the SWE in some places within a footprint may have been large enough to saturate the local microwave radiance, the snow in other places in the footprint may still be shallow and the radiance in these areas are unsaturated. As a result, the PM footprint as a whole can still sense the radiance from the unsaturated areas, even when the average SWE over the footprint has exceeded typical saturation depth. Thus, the relationship between the SWE and T_b involves SWE spatial variability, and variabilities of other factor such as snow grain size. The extent of ice crusts may also affect the overall footprint saturation depth. This hypothesis of SWE variability leading to a large saturation depth for raw PM footprint observations is later proved in Li et al. (2015a), and it agrees with the modeling results in the Upper Kern presented by Li et al. (2015b).

The L12 results suggest ability to utilize T_b for characterizing mountain SWE. Thus, empirical SWE and T_b relationships of the kind explored by L12 are worthy of future exploration. In particular, better understanding of how vegetation modulates these signals is of value for any future attempt at correcting T_b for the effect of forest cover, or radiance assimilation approaches in mountainous areas. While L12 showed nearly linear T_b – SWE relationships for one unforested basin, in this study we expand L12 to examine the other watersheds within the Sierra Nevada. Note that results of L12 (and the relationships developed herein) are not an adequate basis on which to build retrieval algorithms, since e.g. the T_b – SWE relationship changes throughout the course of the winter season (e.g., Li et al., 2015b). Instead, the focus for this study is to explore the effect of forest cover on the T_b – SWE relationships in the Sierra Nevada. Our research questions are: Do linear relationships similar to those shown by L12 for the Kern River exist in other basins? How consistent are the slopes of the linear relationship among the basins? To what extent does forest cover control T_b – SWE relationships?

2. Study area and data

2.1. Study area

The Sierra Nevada is characterized by a north-south elevation gradient, with the southern part significantly higher in elevation. The elevation of the fourteen Sierra Nevada river basins is shown in Fig. 1. The treeline in the Sierra Nevada ranges from 3200 to 3400 m. There is typically greater vegetation density in the northern Sierra (Bales et al., 2006). From Table 1, the maximum elevations range from 2548 m to 4412 m. Rather than studying entire basins, for each of the fourteen basins, we focused on sub-sets of the entire basin for faster PM data processing; we use sub-basins to refer to the spatial areas analyzed throughout the rest of the paper. The sub-basins were selected with the constraints that 1) the selected area be >150 km^2, which is ~1.5 times of the footprint size for the AMSR-E passive microwave observations (refer to Section 2.2); 2) there must be at least one active snow pillow within each sub-basin, and 3) The sub-basin needs to have a range of vegetation density that is similar to the entire basin. The vegetation in the study area is mainly coniferous forest. Short vegetation (e.g. grasses, shrubs) makes up only a limited portion of the total vegetation cover and is often buried by snow in winter; therefore, the focus of this study is on the impact of tall vegetation (forest) cover on the microwave radiance. From Table 1, vegetation cover ranges from 9% (Owens, San Joaquin, Merced) to 49% (Feather). The selected sub-basin areas range from 160 km^2 to 1565 km^2, and the average elevations range from 1513 m to 3170 m.

We selected the Kern basin to study the effect of forest cover on PM. The Kern basin is characterized by a north-south gradient of increasing forest coverage (Fig. 2), which is ideal to explore the response of PM to forest cover.

2.2. Passive microwave observations

AMSR-E was in operation from 2002 to 2011; in this study we processed all the AMSR-E L2A 36.5 GHz v-pol T_b data (Ashcroft and Wentz, 2006) collected during the AMSR-E operational period to derive the brightness temperature time series. The 36.5 GHz frequency tends to be the most sensitive of all the AMSR-E channels to a SWE signal, and v-pol is less sensitive to the presence of ice lenses compared with h-pol (Rees et al., 2010). It was shown to be advantageous to work with the Level-2A (L2A) AMSR-E observations (90.2 km^2 in size) rather than the 25 km × 25 km EASE-Grid T_b; due to the relatively higher resolution of the L2A data, the L2A T_b was three times more sensitive to SWE than was the EASE-Grid T_b (Li et al., 2012). In Li et al. (2012), the experiments were conducted in the Upper Kern basin which was characterized by high elevation (average elevation above 3000 m) and low forest coverage (7% on average). The L2A footprints are elliptical in shape, with minor and major axes of 8.2 and 14 km, respectively. The total area of an L2A footprint is thus 90.2 km^2 in size, which is just 14.4% of 625 km^2 EASE-Grid. We used the basin-scale average method developed by Li et al. (2012) to estimate the average radiance of a watershed from the L2A data. Fig. 3 shows four overpasses of AMSR-E L2A footprints over Mono Basin. Note that the footprint locations and orientations change on each pass. Only the T_b from the nighttime