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An isoline separating relatively warm from relatively cool wintertime forest surface temperatures for the southeastern United States

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ABSTRACT

Forest-oriented climate mitigation policies promote forestation as a means to increase uptake of atmospheric carbon to counteract global warming. Some have pointed out that a carbon-centric forest policy may be overstated because it discounts biophysical aspects of the influence of forests on climate. In extra-tropical regions, many climate models have shown that forests tend to be warmer than grasslands and croplands because forest albedos tend to be lower than non-forest albedos. A lower forest albedo results in higher absorption of solar radiation and increased sensible warming that is not offset by the cooling effects of carbon uptake in extra-tropical regions. However, comparison of forest warming potential in the context of climate models is based on a coarse classification system of tropical, temperate, and boreal. There is considerable variation in climate within the broad latitudinal zonation of tropical, temperate, and boreal, and the relationship between biophysical (albedo) and biogeochemical (carbon uptake) mechanisms may not be constant within these broad zones. We compared wintertime forest and non-forest surface temperatures for the southeastern United States and found that forest surface temperatures shifted from being warmer than non-forest surface temperatures north of approximately 36°N to cooler south of 36°N. Our results suggest that the biophysical aspects of forests' influence on climate reinforce the biogeochemical aspects of forests' influence on climate south of 36°N. South of 36°N, both biophysical and biogeochemical properties of forests appear to support forestation as a climate mitigation policy. We also provide some quantitative evidence that evergreen forests tend to have cooler wintertime surface temperatures than deciduous forests that may be attributable to greater evapotranspiration rates.

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1. Introduction

There is uncertainty regarding the influence of extra-tropical forests on climate (Bala et al., 2007; Bonan, 2008). Most studies examining the role of mid- and high-latitude forests have found that removal of forests at these latitudes would promote cooling, with relatively few finding that such forest removal would promote warming (Wickham et al., 2013). Although many factors influence forest versus non-forest surface temperatures (e.g., surface roughness, frictional resistance to transpiration), the competing influences of albedo and carbon uptake appear to be the main factors producing the uncertainty (Bala et al., 2007; Betts, 2001; Betts et al., 2007; Davin and De Noblet-Ducoudré, 2010; Defries et al., 2002). Forests tend to be dark and therefore absorb more of the sun's radiation than surrounding fields. In the absence of photosynthetic activity (i.e., mid- and high-latitude winters) the lower albedo of forests leads to warming that is not offset by carbon uptake and the cooling effects of transpiration. The net effect of the interplay between albedo and seasonal photosynthetic activity is that mid- and high-latitude forests tend to be warmer than surrounding fields such that deforestation would promote cooling at high latitudes and have little or no effect at mid-latitudes (Bala et al., 2007; Bonan, 2008).

The uncertainty regarding the influence of extra-tropical forests on climate affects forest-oriented policies adopted to mitigate climate change. The United Nations Framework Convention on Climate Change (UNFCCC), the Intergovernmental Panel on Climate Change (IPCC), and other organizations promote forest management as a means to mitigate climate change impacts (Nabuurs et al., 2007; Phelps et al., 2010; US EPA, 2005). The main objective of such policies is to increase carbon stocks through the avoidance of deforestation and forest degradation and the promotion of forestation. Some have pointed out that such policies need to account for biophysical influences of forests on climate (Anderson et al., 2011; Betts et al., 2007; Jackson et al., 2008). We use the term forestation as a convenient integrator of afforestation (planting trees where they have not been historically) and reforestation (planting trees where they once were). Both restoration practices are promoted by UNFCCC, IPCC, and other organizations.

Tropical, temperate, and boreal are broad generalizations of the types of climates that occur within each latitudinal zone. Considerable variability exists within each latitudinal zone, and this variability may influence the degree to which forests act as warming or cooling agents.

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Within the temperate zone, uncovering geographic variation in the warmth of forests relative to croplands, pastures, grasslands, and shrublands can be used to inform forest-oriented climate mitigation policies by showing those locations where the biophysical properties of forests reinforce the cooling effects of carbon uptake.

The study is undertaken in the southeastern United States, where forests occur within subtropical, warm temperate, humid temperate, and humid continental climates (Trewartha, 1961). Comparison of forest and non-forest temperature patterns is focused on the winter season. In previous studies covering the continental United States, forest surface temperatures tended to be cooler than non-forest surface temperatures annually and in all seasons except winter (Wickham et al., 2012, 2013). Further, these analyses revealed a geographic gradient in wintertime surface temperatures such that the forests switched from being relatively warmer to relatively cooler than non-forest as latitude decreased. The main objective of this research is to map more precisely where that switch occurs. Mapping where that switch occurs will identify a region within the continental United States where forest surface temperatures are cooler than non-forest surface temperatures throughout the entire year.

Analysis of forest and non-forest temperature patterns is based on surface temperatures rather than near-surface air temperatures. Surface temperature is often derived from satellites based on the thermal emission of surface features, whereas near-surface air temperature is measured with a thermometer at ~1.5 m above the ground (Jin et al., 1997). Surface temperature is a key parameter in the surface energy budget. Surface temperature is needed to calculate all parameters on the right-hand side of the surface energy budget equation (emitted longwave radiation, sensible heat, latent heat, and conduction).

2. Methods

The research was undertaken in the southeastern United States. Southeastern was defined as south of approximately the 38°N parallel and east of the 95°W meridian. The southern and eastern boundaries were the Gulf of Mexico and Atlantic Ocean. The northern boundary includes the cities of St. Louis Missouri, Lexington Kentucky, and Richmond Virginia, and the western boundary includes the eastern margins of Texas, Oklahoma, and Kansas. The study area is approximately 20% of the conterminous United States.

Surface temperatures were from the Moderate Resolution Imaging Spectroradiometer (MODIS). We used the MODIS-Aqua Version 5, 8-day composite (MYD11A2). Version 5 includes the latest updates and refinements to the MODIS land surface temperature (LST) data, and differences between MODIS LST and measured values at 47 validation sites were less than 1 K (Wan, 2008). The MODIS LST data include daytime and nighttime surface temperatures at 1 km^2 spatial resolution. We used the MODIS-AQUA (afternoon overpass) rather than the MODIS-TERRA (morning overpass) data so that our analyses were based on observations for the warmer part of the day. The wintertime surface temperature data were collected for the years 2007 through 2012 (inclusive), and winter was defined as December, January, and February. We calculated the six-year seasonal mean for each pixel using both daytime and nighttime temperatures. We used both daytime and nighttime surface temperatures because daytime and nighttime profiles change as land cover changes (Wickham et al, 2012). Omission of nighttime surface temperatures can lead to bias in the comparison of surface temperature across land cover types (Lee et al., 2011). Pixels with fewer than six observations for a year were discarded, and averages were not computed for discarded pixels.

Land cover was from the National Land Cover Database (NLCD) 2001 (Homer et al., 2007). NLCD is a remotely sensed product derived from the Landsat satellite series. NLCD 2001 includes a 16-class data set of generalized land cover classes (www.mrlc.gov/nlcd01_leg.php) resolved at the Landsat native pixel size of 30 m × 30 m (see Supplemental Material, Fig. S1). We simplified the NLCD land cover into forest and non-forest. We defined forest as the deciduous forest, evergreen forest, mixed forest, and woody wetlands classes; and we defined non-forest as cropland, pasture, grassland, and shrubland. Urban, water, and emergent wetland classes were not included in the non-forest class.

To reconcile the differences in spatial resolution between the MODIS and NLCD data, we used moving window spatial convolution to measure the amount of forest and non-forest surrounding each MODIS 1 km^2 pixel (Riitters et al., 2000, 2002). We used the spatial convolution results to define forest and non-forest MODIS pixels as those pixels that were at least 75% forest or non-forest. We chose the 75% threshold as a compromise between homogeneity and sufficient sample size. Fewer MODIS pixels could be classified as forest or non-forest as the threshold increased.

To examine the relationship between wintertime surface temperatures and land cover, the study area was divided into 318 tiles that were 100 km (east–west) by 50 km (north–south). The 50 km north–south distance was chosen to minimize the impact of latitude on the relationship between surface temperatures and land cover. Fifty (50) km spans approximately 0.45° of latitude. Each tile was then subdivided into 200 cells that were 5 km × 5 km to control for spatial correlation in MODIS surface temperatures. Per-class (forest, non-forest) surface temperature means were then calculated for each 25 km^2 cell within each tile (Fig. 1). At a 1 km^2 spatial resolution, there are potentially 25 MODIS pixels within each 25 km^2 cell and 5000 MODIS pixels within each tile. Assuming a uniform spatial distribution, each tile would be comprised of 200 forest and non-forest means that were based on 12–13 MODIS observations per class within each 25 km^2 cell. Comparison of forest and non-forest surface temperatures was restricted to tiles that had at least 10 observations (25 km^2 cells) for each class.

Comparison of forest and non-forest surface temperature means per tile was based on a *T*-test ($\alpha = 0.05$). The *T*-tests were informed by regressions of surface temperature versus elevation on a per-class basis (Supplemental Material, Table S1). Scatterplots were examined to find ranges, where possible, over which elevation was less influential when the goodness-of-fit estimates for surface temperature versus elevation were high ($R^2 \geq 0.25$). The use of restricted elevation ranges occurred most commonly in the Appalachian and Ozark Mountains. The use of restricted elevation ranges because of an elevated R^2 was not always necessary. Even when R^2 values are high (e.g., ≥ 0.5), residuals can still be sufficiently large that the use of a restricted elevation range would not produce meaningful changes in surface temperature means. R-square values were commonly less than 0.25.

Studies have suggested that non-growing season transpiration may be higher in evergreen forests than deciduous forests in the southeast (Stoy et al., 2006; Juang et al., 2007). The cooling effect of transpiration may be reflected in the surface temperatures such that values for deciduous forests tend to be warmer than those for evergreen forests in our study area. We conducted a statistical comparison for 10 tiles (see Fig. 1) in which there were sufficient observations for each forest type to determine if wintertime surface temperatures were higher for deciduous forests than evergreen forests. For the comparison, we selected the surface temperature observation within each 25 km^2 cell that had the highest proportion of either deciduous or evergreen forest (above the 75% minimum used as the threshold for homogeneity) within the 1 km^2 area representing the MODIS pixel, and compared the results by tile using a *T*-test ($\alpha = 0.05$). Statistical comparison for most tiles was not possible because of the strong geographic pattern of forest types (see Fig. S1). Tiles tended to be dominated by either deciduous or evergreen forest such that sufficient sample sizes of each were not available in most tiles.

The evergreen versus deciduous forest surface temperature comparisons were supported by comparison of evapotranspiration rates. We used 1 km^2 MODIS-based (MOD16) data (Mu et al., 2011) to compare evapotranspiration rates between evergreen and deciduous forest for the 10 tiles in which surface temperature comparisons were undertaken. The MOD16 data were compiled for the same time period

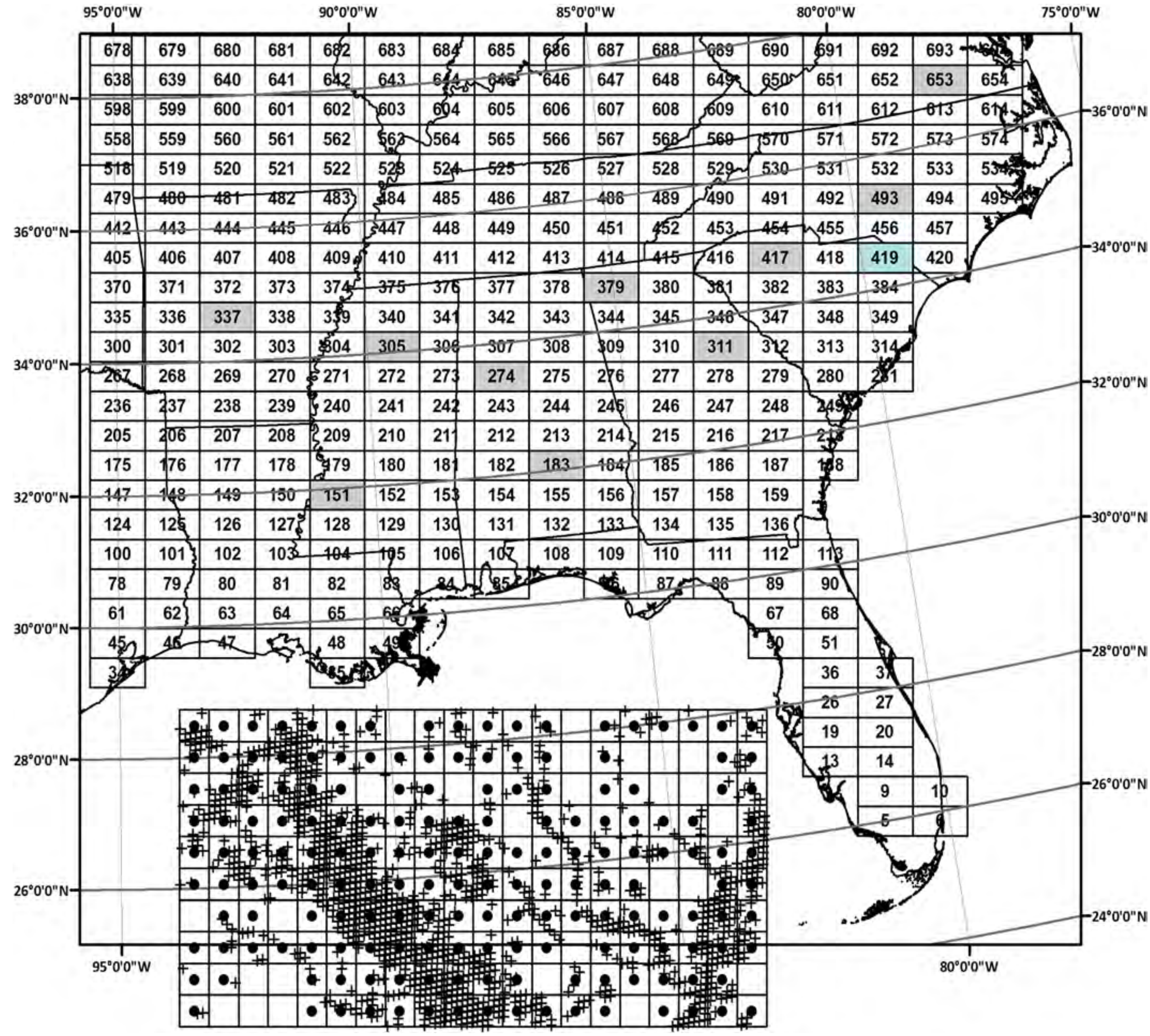

Fig. 1. Study area. The inset (Tile 419) shows the spatial distribution of MODIS observations assigned to the forest class (+) and the associated 25 km^2 cells for which a mean (per cell) forest surface temperature could be calculated (•). Results for tiles shaded in gray are reported in Table 1.

(2007–2012; inclusive) to estimate the average total wintertime evapotranspiration.

3. Results

The 36°N parallel appears to be an approximate boundary between relatively warm and relatively cool wintertime forest surface temperatures (Fig. 2). North of 36°N, wintertime forest surface temperatures tended to be significantly warmer that non-forest surface temperatures, and south of 36°N, wintertime forest surface temperatures tended to be significantly cooler that non-forest surface temperatures. Tiles 181 (~32°N) and 453 (~35°N) were the only locations south of 36°N where wintertime forest surface temperatures were significantly warmer than non-forest surface temperatures. The mean forest minus non-forest differences for these tiles were 0.1 °C for tile 181 and 0.3 for tile 453.

The transition between relatively warm and relatively cool wintertime forest surface temperatures was abrupt. There was not a "smooth" pattern of significantly warmer to statistically equivalent to significantly cooler wintertime forest surface temperatures from north to south. Along the 36°N parallel, tiles where forest was significantly warmer often had neighbors where forest was significantly cooler. The abrupt transitions had both north–south and east–west orientations. For example, mean forest minus non-forest differences changed from 0.66 to −0.24 moving from tile 525 south to tile 486 and from 0.27 to −0.38 moving from tile 526 east to tile 527.

Of the 318 tiles in the study area, 73 had statistically equivalent forest and non-forest surface temperatures. These tiles were concentrated south of the 36°N parallel and were not organized into a distinct spatial pattern. Overall, forest surface temperatures were slightly warmer in these 73 tiles, with 60% of them having slightly warmer forest surface temperatures.

As expected, there was a north–south gradient in the magnitude of the difference between forest and non-forest surface temperatures (Fig. 3). Mean forest minus non-forest surface temperature differences ranged from 1.19 °C for tile 639 (north of Springfield, MO) to

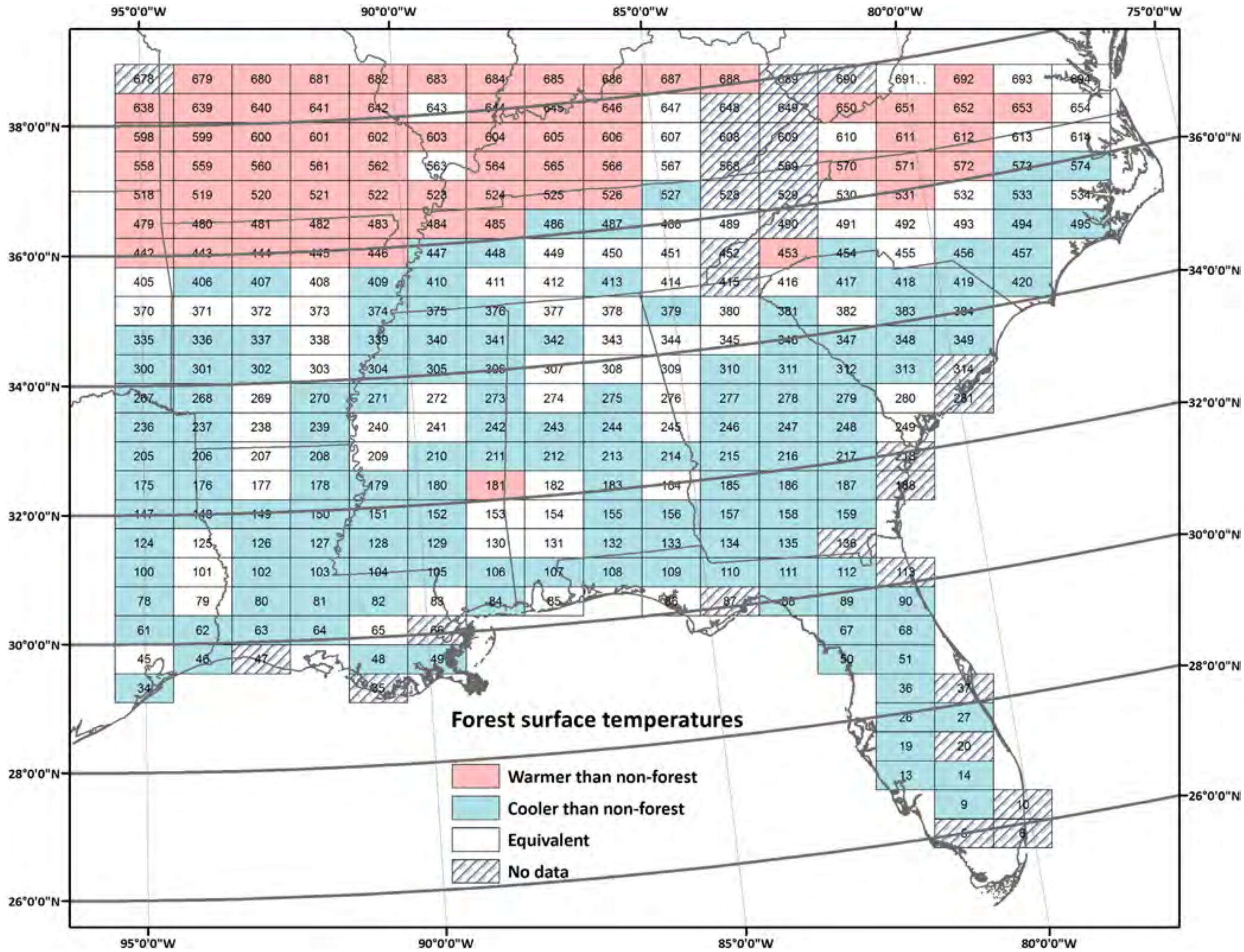

Fig. 2. Comparison of wintertime forest and non-forest surface temperatures. Cells displayed as no data had less than 10 observations (25 km^2 cells) for either the forest or non-forest class.

−0.98 °C for tile 51 (Orlando, FL). Overall, ±0.1 °C was an approximate difference threshold for statistical significance. Moving northward from 36°N, forest minus non-forest surface temperature differences tended to increase from 0.1 °C to 0.75 °C, and moving southward from 36°N, forest minus non-forest surface temperature differences tended to decrease from −0.1 °C to −0.50 °C (see also Table S1). The mean forest minus non-forest difference for the set of tiles with statistically significant warmer forest surface temperatures was 0.40 °C and the mean forest minus non-forest difference for the set of tiles with statistically significant cooler forest surface temperatures was −0.27 °C.

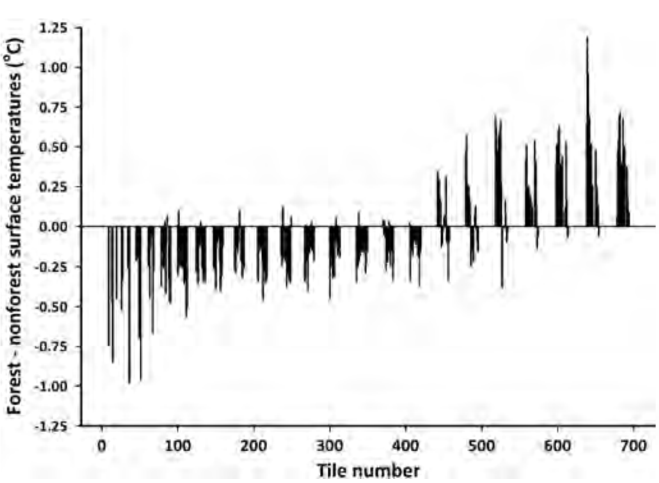

Fig. 3. Differences in mean forest minus non-forest surface temperatures by tile.

In part because of the spatial pattern of inter-tile variability in surface temperatures (e.g., Fig. 2), intra-tile variability in forest minus non-forest surface temperatures was examined for 12 tiles (685, 682, 604, 653, 479, 448, 382, 374, 370, 236, 181, 135) by comparing the surface temperatures for the 25 km^2 cells within a tile that had forest and non-forest observations. Even within tiles where forest surface temperatures were significantly warmer than non-forest surface temperatures, there was a substantial fraction of 25 km^2 cells where forest surface temperatures were cooler than non-forest surface temperatures (Fig. 4). As expected, the proportion of 25 km^2 cells within a tile that had cooler forest surface temperatures increased as latitude decreased. For example, only 25% of the 25 km^2 cells in tile 653 had cooler forest surface temperatures, whereas 86% of the 25 km^2 cells in tile 382 had cooler forest surface temperatures. In the anomalous tile (181), ~43% of the 25 km^2 cells had cooler forest surface temperatures.

Evergreen forest surface temperatures were significantly cooler than deciduous forest surface temperatures in 6 of the 10 tiles for which there were sufficient observations of each forest type (Table 1). The tiles for which evergreen and deciduous forests surface temperatures were statistically equivalent (653, 417, 379, 311) tended to be at more northerly latitudes, and the results for two of the tiles may be attributable to a small sample sizes for deciduous forests (311) or evergreen

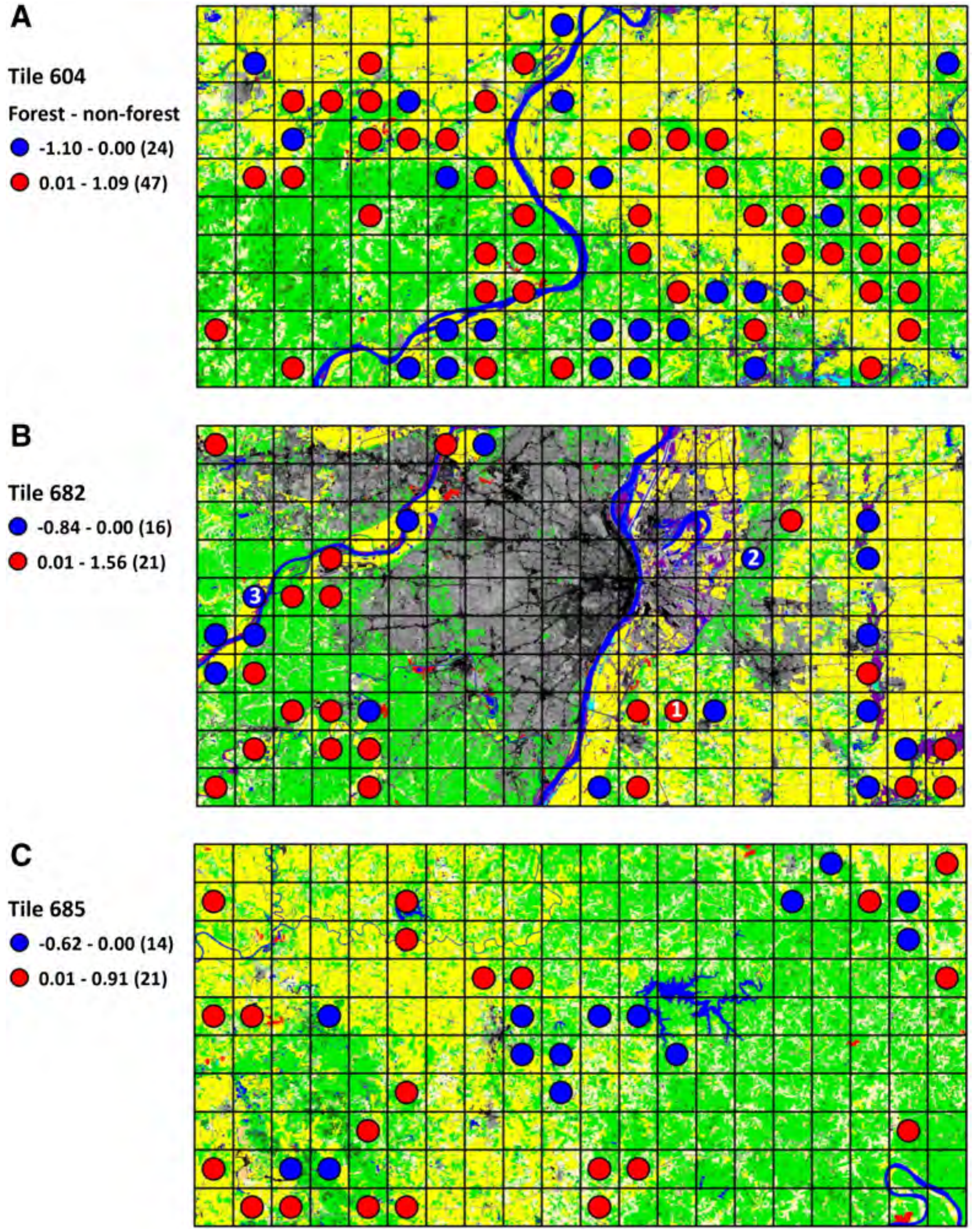

Fig. 4. Intra-tile variability in surface temperatures (°C). Cells labeled 1, 2, and 3 in panel B are used in Table 2. The proportion of deciduous and evergreen forest (respectively) by tile is 0.36, 0.03 (604); 0.23, 0.00 (682); and 0.47, 0.02 (685).

forests (379). Deciduous forest surface temperatures were less than non-forest surface temperatures in 7 of the 10 tiles, and evergreen forest temperatures were less than non-forest surface temperatures in 9 of the 10 tiles (results not shown and based on arithmetic differences only). Surface temperatures were cooler for non-forest than deciduous forest in tiles 653, 493, and 274. Forest versus non-forest comparisons for these tiles (see Fig. 2) were either statistically equivalent (493, 274) or non-forest surface temperatures were significantly cooler (653). Of the ten tiles examined, only 653 had non-forest surface temperatures that were cooler than evergreen forest surface temperatures.

Differences in evapotranspiration rates between evergreen and deciduous forest were generally consistent with the surface temperature differences (Table 1). Evergreen forest had significantly greater evapotranspiration than deciduous forest in those tiles where deciduous forest was significantly warmer than evergreen forest (151, 183, 274, 493), and evergreen forest evapotranspiration was statistically equivalent to deciduous forest evapotranspiration in tiles where the surface temperature differences were not significant (311, 379, 417). In the remaining tiles, surface temperature differences were statistically different and evapotranspiration was statistically equivalent (305, 337) or vice versa (653).

4. Discussion

Biophysical properties of forests, through their influence on the surface energy budget, affect climate. South of the 36°N in the southeastern United States, forest wintertime surface temperatures tended to be cooler than surrounding croplands, pastures, grasslands, and shrublands. The relatively cool forest surface temperatures occurred during a season of minimal photosynthetic activity and therefore the cooling effect of transpiration was not as strong a contributor as it would be in the spring, summer, and fall. Our results suggest that the

Table 1
Differences in mean wintertime surface temperature and evapotranspiration between evergreen forest (EF) and deciduous forest (DF) by tile. Values in parentheses are the number of observations. Significance ($\alpha = 0.05$) is denoted using " > " and " = " signs (e.g., DF > EF is significant and DF = EF is not significant).

			Surface temperature (°C)			Evapotranspiration (mm)		
Tile	Latitude	Longitude	EF	DF	Significance	EF	DF	Significance
653	36.92°N	77.83°W	5.56 (78)	5.63 (40)	DF = EF	66.2 (78)	58.8 (46)	EF > DF
493	35.32°N	79.34°W	7.01 (62)	7.32 (31)	DF > EF	71.6	61.8	EF > DF
417	34.73°N	81.70°W	7.50 (40)	7.57 (69)	DF = EF	71.4 (69)	71.7 (40)	EF = DF
337	34.66°N	92.77°W	6.73 (86)	7.05 (79)	DF > EF	70.2 (86)	70.4 (75)	EF = DF
379	34.64°N	85.08°W	6.34 (12)	6.49 (74)	DF = EF	65.8 (12)	67.0 (73)	EF = DF
305	34.08°N	89.51°W	7.06 (52)	7.25 (35)	DF > EF	82.5 (52)	80.6 (33)	EF = DF
311	33.53°N	83.02°W	8.27 (124)	8.22 (8)	DF = EF	76.7 (123)	75.3 (8)	EF = DF
274	33.49°N	87.39°W	7.71 (57)	7.94 (65)	DF > EF	73.4 (57)	69.5 (62)	EF > DF
183	32.07°N	86.47°W	9.28 (52)	9.42 (38)	DF > EF	89.6 (52)	82.8 (28)	EF > DF
151	31.90°N	90.75°W	8.99 (25)	9.24 (66)	DF > EF	93.3 (25)	86.4 (66)	EF > DF

Table 2
Forest and non-forest albedo estimates from surface temperatures for Tile 682 and three 25 km^2 cells within Tile 682. Column headings are surface temperature (T_s) emitted longwave radiation (L↑), sensible heat (H), absorbed solar radiation (α) and albedo (r).

Class	T_s	L↑	H	α	r
Tile 682					
Forest	1.43	314.2	−62.3	0.82	0.18
Non-forest	0.71	311.0	−79.7	0.65	0.35
Cell 1					
Forest	0.92	311.9	−74.7	0.70	0.30
Non-forest	0.75	311.1	−78.8	0.66	0.34
Cell 2					
Forest	0.95	312.0	−73.9	0.70	0.30
Non-forest	0.99	312.2	−73.0	0.71	0.29
Cell 3					
Forest	0.81	311.4	−77.3	0.67	0.33
Non-forest	0.90	311.8	−75.1	0.69	0.31

Notes: Emitted longwave radiation was solved using $L\uparrow = \varepsilon\sigma(T_s + 273.15)^4$, where ε is emissivity, σ is the Stefan–Boltzmann constant, and T_s is surface temperature. Emissivity was set to 0.975 for both forest and non-forest and the Stefan–Boltzmann constant is 5.67×10^{-8}. Sensible heat was solved using the equation $H = (-\rho Cp(T_a - T_s)) / r_H$, where ρ is the density of air, Cp is the heat capacity of air, T_a is the air temperature, and r_H is the transfer resistance. Values for ρ, Cp, T_a, and r_H were set to 1.2, 1010, 4(°C), and 50, respectively. S↓ and L↓ were set to 125 and 150, respectively, and are representatives of the cloudy and cool conditions of St. Louis winters. The value used for T_a (4 °C) is the long-term January daytime mean for St. Louis. Values for all constants and incoming solar (S↓) and longwave (L↓) radiation were taken from Bonan (2002).

southeastern United States, south of the 36°N, is a region where forestation has unambiguous climate mitigation potential because biophysical factors should reinforce the cooling effects of carbon uptake.

The climate mitigation potential of forestation may extend beyond the southeastern region to larger portions of the United States. Forest surface temperatures were found to be cooler than cropland surface temperatures in the spring, summer, fall, and annually for most locations in the conterminous United States (Wickham et al., 2012), and spatially extensive forests tended to have cooler surface temperatures than forests with limited areal extent (Wickham et al., 2013). Montenegro et al. (2009) examined forestation potential and found that the cooling effects of carbon uptake outpaced the warming effects of lower albedo within mid- and high-latitude environments. Zhao and Jackson (2014) found that the conversion of cropland to forest reduced mean annual surface temperatures in North America between 20°N and 60°N.

Although there are other biophysical properties that affect the surface radiation budget (e.g., frictional resistance to transpiration, surface roughness), albedo is regarded as the dominant factor (Bala et al., 2007; Betts, 2001; Betts et al., 2007; Davin and De Noblet-Ducoudré, 2010; Defries et al., 2002). The effects of the other biophysical factors tend to be local, whereas the effect of albedo tends to be global (Davin and De Noblet-Ducoudré, 2010). We can use the basic equation for the surface energy budget to back-calculate albedo using the surface temperatures reported here and available literature values (Bonan, 2002). The surface energy budget equation for the back-calculation of albedo is:

$$\alpha = (L\uparrow + H + LE + G - L\downarrow)/S\downarrow, \text{ where} \quad (1)$$

L↑ and L↓ are the emitted and incoming longwave radiation, respectively, H is the sensible heat, LE is the latent heat, G is the conduction, S↓ is the incoming solar radiation, and α is $(1 - r)$ where r is the albedo. Back-calculation of albedo is reported for Tile 682 and three 25 km^2 cells within the tile (Table 2). By using tile 682, which includes St. Louis, Missouri, we can ignore the contributions of latent heat and ground conduction since the long-term average daily maximum and minimum January temperatures are 4 °C and −4 °C, respectively. Non-forest albedo is greater than forest albedo by approximately 0.2 for the tile, but there is considerable variation in albedo differences within the tile. The back-calculated albedo values for forest are higher than the non-forest albedo values in the two 25 km^2 cells for which forest surface temperatures are lower than non-forest surface temperatures.

Although only illustrative, the variability in back-calculated albedo values is consistent with the IPCC's assessment that our understanding of the role of albedo as a climate driver is low to moderate (IPCC, 2007). Our illustrative albedo estimates are also consistent with mid-March (leaf off) Landsat TM albedo estimates for the Washington, DC area (Shuai et al., 2011). For the mid-March Landsat-based data, median albedo values for cropland and deciduous forest are different by only 0.035 and the upper quartile of the deciduous forest albedos and lower quartile of cropland albedos are approximately equivalent (Table S2). The statistical pattern of albedo in Table S2 is consistent with the spatial pattern of albedo in Fig. 4B (and Table 2). One would expect to find some locations with similar cropland and forest albedos (Fig. 4B) if their statistical distributions were overlapping.

Albedo is conceptually simple (percentage of the sun's radiation that is reflected), but it is difficult to measure precisely because it is sensitive to a wide array of environmental conditions. Albedo changes throughout the day and from season to season due to changes in sun angle and cloud cover. Albedo is affected by soil color and soil wetness (Bonan, 1997), and forest composition affects albedo because the bark of some trees are lighter than others (Jackson et al., 2008). At mid- and high latitudes, the amount, timing, and persistence of snow will affect albedo seasonally and from one year to the next (Wang and Davidson, 2007). The albedos of grassland, deciduous forest, and open Jack pine (evergreen) stands in Canada were found to be essentially equivalent during periods of snow cover and strongly similar during snow-free periods (Davidson and Wang, 2004).

Shadowing and light trapping are the main mechanisms producing lower forest albedos relative to herbaceous vegetation (Davidson and Wang, 2004; Dickinson, 1983). The amount of shade in a forest stand is proportional to the height and width of the tree boles (Dickinson, 1983). Relative to open areas, direct beam shortwave radiation is attenuated under forest canopies such that the two main radiation components are diffuse shortwave and an enhanced longwave component emitted from the canopy (Link et al., 2004; Pomeroy et al., 2009; Sicart et al., 2004; Schelker et al., 2013). Despite the enhanced longwave component, there tends to be a net decline in total radiation under a forest canopy (Link et al., 2004) that results in lower sub-canopy air temperatures relative to above canopy air temperatures during winter (Pomeroy et al., 2009). Attenuation of shortwave radiation also tends to result in sub-canopy air temperatures that are cooler than surrounding fields (Link and Marks, 1999), and although there are many complicating factors (Musselman et al., 2012), snow tends to persist longer under a forest canopy than in an open area (Link and Marks, 1999; Schelker et al., 2013).

It does not appear that snow cover was an important determinant of albedo or surface temperatures in this study. The maximum number of days of detected snow cover was 93, approximately 16% of the total number of days in the study (Fig. S2). Only ~0.6% of the study area had at least of one month (>30 days) of snow cover during the 18 months covered in this study. The number of days in which snow was detected was five or less for many of the northern tier of cells in which forest surface temperatures were warmer than non-forest surface temperatures (western Kentucky and southeastern Missouri), suggesting that snow was not a significant determinant of the relative differences in surface temperatures for these tiles.

The correspondence between evergreen and deciduous forest surface temperatures and evapotranspiration (Table 1) suggests that cooler surface temperatures in evergreen forests relative to deciduous forests is at least partly attributable to greater evapotranspiration rates in evergreen forests during the winter. The differences between forest and non-forest surface temperatures (Fig. 3) may have been slightly less than realized if deciduous forest was more prevalent in the southern and eastern margins of our study area (Fig. S3). The generally higher wintertime evapotranspiration rates in evergreen forests found here is consistent with field-based studies in North Carolina (Stoy et al., 2006; Juang et al., 2007).

In previous research, we found evidence of a geographic gradient in which wintertime forest surface temperatures were warmer than non-forest surface temperatures at the northern margins of the continental United States and cooler at the southern margins (Wickham et al., 2012). In this research, we have quantified and mapped where the reversal occurs. South of ~36°N forest surface temperatures switch from being warmer than non-forest to cooler than non-forest. The results are based on a geographic gradient that used analysis units smaller than 0.5° of latitude and high resolution land-cover (0.0009 km^2/pixel) and surface temperature data (1 km^2/pixel). Our results add geographic and quantitative precision to the broad-scale, qualitative analysis by Anderson et al. (2011), who also suggested that the southeastern United States was a region where the biophysical properties of forests reinforce the cooling effects of carbon uptake.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gloplacha.2014.05.012.

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