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Systematic variation in evapotranspiration trends and drivers across the Northeastern United States

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Abstract

The direction and magnitude of responses of evapotranspiration (ET) to climate change are important to understand, as ET represents a major water and energy flux from terrestrial ecosystems, with consequences that feed back to the climate system. We inferred multidecadal trends in water balance in 11 river basins (1940–2012) and eight smaller watersheds (with records ranging from 18 to 61 years in length) in the Northeastern United States. Trends in river basin actual ET (AET) varied across the region, with an apparent latitudinal pattern: AET increased in the cooler northern part of the region (Maine) but decreased in some warmer regions to the southwest (Pennsylvania–Ohio). Of the four small watersheds with records longer than 45 years, two fit this geographic pattern in AET trends. The differential effects of the warming climate on AET across the region may indicate different mechanisms of change in more- vs. less-energy-limited watersheds, even though annual precipitation greatly exceeds potential ET across the entire region. Correlations between AET and time series of temperature and precipitation also indicate differences in limiting factors for AET across the Northeastern U.S. climate gradient. At many sites across the climate gradient, water-year AET correlated with summer precipitation, implying that water limitation is at least transiently important in some years, whereas correlations with temperature indices were more prominent in northern than southern sites within the region.

KEYWORDS

energy limitation, evapotranspiration, water balance, water limitation

1 | INTRODUCTION

Water vapour flux between the earth surface and the atmosphere via evapotranspiration (ET) is a major component of water and energy balances. Changes in ET have important consequences for the reliability of surface freshwater resources, ecosystem productivity, and soil biogeochemical processes, as well as feedbacks to the global climate system. A general consensus has emerged that anthropogenic climate forcing is likely to intensify the global hydrological cycle (Hobbins, Ramirez, & Brown, 2004; Huntington, 2006; Van Heerwaarden, Vilà-Guerau De Arellano, & Teuling, 2010; Walter, Wilks, Parlange, & Schneider, 2004). However, the complex set of factors controlling ET fluxes challenge the notion that ET might simply increase due to climate warming. Transpiration, which represents the majority of ET in forested landscapes (Jasechko et al., 2013; Zhang et al., 2016), is controlled not only by the atmospheric demand for water but also by soil water availability, physiological traits of vegetation, and the duration of the leaf-on season (Huntington, 2004; Kirschbaum, 2004; Meinzer et al., 2013). The popular Budyko water balance framework models (after Budyko, 1974) are parsimonious in nature, partitioning precipitation run-off and ET (e.g., Jones et al., 2012), where ET is limited either by soil water availability or by atmospheric demand (potential ET, PET). PET is a function of air temperature, atmospheric pressure, wind speed, specific humidity, and solar radiation (Penman, 1948). Thus, studies examining the reasons for ET changes must recognize the potential for offsetting trends among the multiple dimensions of climate change (Donohue, McVicar, & Roderick, 2010), as well as potentially complex interactions of multiple global change drivers on ET.

Some previous analyses of ET trends utilized regional to global networks of evaporation pan observations to provide estimates of PET, with results showing consistent multidecadal declines in pan evaporation (Lawrimore & Peterson, 2000; McVicar et al., 2012; Roderick, Hobbins, & Farquhar, 2009). However, in humid regions, interpreting pan evaporation is complicated because increases in transpiration from the surrounding landscape may sufficiently reduce the vapour pressure deficit (VPD) to depress rates of pan evaporation (Hobbins et al., 2004; Roderick et al., 2009; Van Heerwaarden et al., 2010).

Studies using direct estimates of ET based on flux tower data or water yield from large river basins around the world have reported trends of both increasing ET (Hobbins et al., 2004; Walter et al., 2004; Zeng et al., 2012) and decreasing ET (Jung, Chang, & Risley, 2013; Keenan et al., 2013). Observed ET trends have varied across time periods (Jung et al., 2010; Yao et al., 2012) or sites (Teuling et al., 2009; Walter et al., 2004). Long-term trends and drivers of ET differ considerably among various landscapes, climatic regions, and time (Jones et al., 2012).

In the Northeastern United States, trends in ET and the dominant controls on this trend are not clear. There have been long-term increases in both precipitation (Hayhoe et al., 2007; Keim, Fischer, & Wilson, 2005) and streamflow (Collins, 2009; Hodgkins and Dudley, 2005; McCabe & Wolock, 2002). Like the global and continental scale studies discussed above, water balance analyses at various scales suggest that ET in the northeast might, on balance, be increasing (Huntington & Billmire, 2014; Jung et al., 2013; Kramer et al., 2015; Lu et al., 2015; Walter et al., 2004; but see Campbell, Driscoll, Pourmokhtarian, & Hayhoe, 2011). The combination of increasing precipitation and air temperature (Hayhoe et al., 2007; Kunkel et al., 2013) is expected to enhance ET (Huntington, Richardson, McGuire, & Hayhoe, 2009) relieving both water and energy limitations where and when they occur. These climate-induced changes in ET have been demonstrated with models run using future climate change projections, and the effect is attributed largely to a lengthening of the leaf-on season of deciduous trees (Hayhoe et al., 2007; Pourmokhtarian et al., 2017; Szilagyi, Katul, & Parlange, 2001).

There are a number of observed and predicted changes in drivers of ET (climate and vegetation) that might act to increase or decrease ET at the regional scale. At the first order, a warming climate should increase VPD and PET and, therefore, actual ET (AET) in energy-limited environments. On the other hand, where daily temperature ranges (DTR) have declined due to greater warming at night than during the day, as in the Northeastern United States (Lauritsen & Rogers, 2012), daytime humidity might increase enough to stabilize VPD despite warming temperatures. In fact, there is a substantial negative feedback in this system, because increases in ET tend to reduce both VPD and daytime temperatures (Bounoua et al., 2010; Durre & Wallace, 2001; Kramer et al., 2015). In addition to VPD, PET is also determined by incident solar radiation (e.g., as affected by cloudiness and atmospheric aerosols) as well as by surface wind velocities (Penman, 1948). Time series for these climatic parameters are coarser and considerably less complete than those for temperature, precipitation, and humidity (Dai, Karl, Sun, & Trenberth, 2006; Harris, Jones, Osborn, & Lister, 2014; Pryor et al., 2009; Willett, Reynolds, Stevens, Ormerod, & Jones, 2000). However, in the Northeastern United States, substantial increases in cloudiness and decreases in wind have been observed over the past century (Iacono, 2009; Lauritsen & Rogers, 2012; Pryor et al., 2009; Ukkola & Prentice, 2013), potentially offsetting the effect of warming on PET. Conversely, global decreases in incident radiation due to aerosol pollution in the mid-late 20th century (which would have suppressed ET) appear to have largely reversed themselves since about 1990 (Wild, 2012). The net effect of these changes is difficult to model over the long term at the regional scale with available data but may be detectable in the water balance of long-term monitored watersheds.

In our study, we investigated spatial and temporal patterns of ET in both large and small watersheds across the Northeastern United States. We quantified trends in ET in 11 large watersheds (hereafter "river basins"; >5,000 km^2) and in eight small upland watersheds (<10 km^2), for which there are high-quality observational datasets on precipitation and streamflow allowing calculation of ET at an annual time scale via the water balance method. The advantages of the river basins and small watersheds are complementary. Large watersheds are representative of the overall landscape, and spatially averaged precipitation time series at this scale are relatively insensitive to data gaps and variable record lengths (Daly et al., 2008; Di Luzio, Johnson, Daly, Eischeid, & Arnold, 2008). On the other hand, the small upland watersheds we studied were 100% forested for the full duration of each record, eliminating the effect of land-cover change on water balance, and tend to have thin soils, minimizing the potential for interannual variation in storage.

Our objectives were (a) to determine whether consistent long-term trends in AET exist in large and small watersheds across the Northeastern United States and (b) to determine the extent to which AET variation can be explained by simple metrics of PET or water availability at the scale of individual basins.

2 | STUDY SITES AND HYDROMETEOROLOGICAL DATA

We selected study watersheds across the Northeastern United States (Figure 1, Tables 1 and 2) at two different spatial scales. As with most water balance analyses, we aggregated daily or monthly precipitation and streamflow data (as well as climatological data in subsequent correlation analyses) into water years (WY) rather than calendar years, to reduce the effect of sometimes large interannual variation in storage within the watershed (especially in snowy climates) on January 1. The optimal WY in each catchment was chosen (beginning the first of any month without snowpack under normal conditions) as the WY with

FIGURE 1 Map of the 11 river basins (shaded; gage locations shown) and seven locations with small watersheds (triangles) that we included in the analyses (including two small watersheds at Hubbard Brook). Abbreviations follow Tables 1 and 2. The inset map at the upper left shows the study region within the conterminous United States

the highest correlation between precipitation and streamflow over the full record. This procedure minimizes interannual variation in both storage and AET and improves the sensitivity of tests for long-term trends (Likens, 2013; Senay et al., 2011).

We calculated ET as the difference between precipitation and streamflow on a WY basis, following the general approach of previous studies (Campbell et al., 2011; Huntington & Billmire, 2014; Senay et al., 2011; Walter et al., 2004). We did not consider change in storage in this analysis because it is likely a small term in the overall water balance, especially when analysing the trend in water balance over a sufficiently long time period (Huntington & Billmire, 2014; Kramer et al., 2015; Senay et al., 2011; Sharma & Walter, 2014).

2.1 | Small watershed data

We analysed data from eight small watersheds (~0.1–10 km^2), located at research sites where hydrology and forest ecology have been intensively studied by the U.S. Geological Survey (USGS), the U.S. Forest Service, and research universities. All watersheds were forested and unmanaged for the duration of the hydrologic record, and several serve as reference areas for nearby forest management or biogeochemistry experiments (e.g., Fernandez, Adams, SanClements, & Norton, 2010; Green et al., 2013; Hornbeck, Adams, Corbett, Verry, & Lynch, 1993; Wang, Burns, Yanai, Briggs, & Germain, 2006). Clearly delineated small catchments suitable for such research tend to be located in areas of relatively high topographic relief. Most have one or several precipitation collectors onsite or nearby. Site characteristics and site-specific data sources and data processing are described in Appendix S1 and Table SA1. At research sites with multiple small watersheds, we selected the unmanipulated watershed with the longest record. However, at Hubbard Brook, where there are five unmanipulated watersheds, we selected one each from the north- and south-facing slopes. We analysed data records from these small watersheds ranging from 18 to 61 years in length, though the trend analysis in sites with records <45 years was included only for completeness, because these sites were included in the correlation analysis.

2.2 | River basin data

We analysed hydrologic records from 11 river basins (~5,000–50,000 km^2) gaged by the USGS. These records were selected from the Hydro-Climatic Data Network (HCDN; Falcone, Carlisle, Wolock, & Meador, 2010; Slack & Landwehr, 1992), to maximize geographic coverage of the region and examine the largest basins with sufficiently long and complete records, excluding nested smaller watersheds. Together, these basins cover a sizeable fraction of the total area of the Northeastern United States (Figure 1). We considered the density of observations available to reconstruct suitable basin-scale precipitation data to be a larger potential source of error than the inclusion of major human disturbances such as land-use change, net groundwater removal, impoundments, and engineered interbasin transfers, which

TABLE 1 Descriptive data and AET trends in the 11 river basins

USGS gage #	Code	Watershed and gage location	Area (km^2)	Mean annual T (°C)	Mean P (mm year^{-1})	Mean AET (mm year^{-1})	Best WY	Forest cover (%)	Impervious. (%)	Mean elev (m)	Mean slope (%)	Withdrawals (mm year^{-1})	Dam storage change (mm year^{-1})	AET Sen slope (mm year^{-2})
01034500	PEN	Penobscot R. at West Enfield, ME	16,633	4.20	1,092	432	Jun	81	0.6	265	5	1	0.5	**1.08**
01046500	KEN	Kennebec R. at Bingham, ME	7,032	3.18	1,078	488	Jun	73	0.2	448	8	1	1.2	**0.92**
01059000	AND	Androscoggin R. near Auburn, ME	8,451	4.73	1,167	475	Aug	82	0.9	425	12	6	0.2	0.42
01100000	MER	Merrimack R. at Lowell, MA	12,005	7.18	1,188	548	Jun	73	4.6	249	8	55	1.8	−0.09
01170500	CON	Connecticut R. at Montague City, MA	20,357	5.69	1,165	493	Sep	83	3.0	419	13	34	1.1	0.00
01357500	MOH	Mohawk R. at Cohoes, NY	8,935	6.69	1,158	554	Aug	60	1.8	380	8	30	0.3	0.14
01554000	SUS	Susquehanna R. at Sunbury, PA	47,397	7.84	1,048	514	Jul	67	1.1	449	12	32	0.7	−0.33
03049500	ALL	Allegheny R. at Natrona, PA	29,552	8.02	1,139	520	Jul	70	1.1	488	10	44	1.9	0.24
03105500	BEA	Beaver R. at Wampum, PA	5,789	9.17	1,010	579	May	38	4.7	331	2	89	2.0	**−0.99**
03140500	MUS	Muskingum R. near Coshocton, OH	12,585	9.50	1,006	607	May	39	3.2	334	6	46	0.2	0.17
04249000	OSW	Oswego R. at Lock 7, Oswego, NY	13,209	8.03	1,007	510	May	37	1.9	257	4	87	0.2	0.28

Note. Watershed area, elevation, slope, withdrawals (1995–2006), and storage change (1950–2006) are from Falcone (2010). Mean temperature and precipitation are from PRISM for the period 1971–2000. Forest cover and impervious surface are from NLCD 2011 (Homer et al., 2015). Mean AET and AET trend statistics are for WY 1940–2012. Sen slopes in bold are significant (Kendall $p < 0.05$). Sites are listed from northeast (cooler) to southwest (warmer). AET: actual evapotranspiration; USGS: U.S. Geological Survey; WY: water years.

TABLE 2 Descriptive data for the eight small study watersheds.

Abbrev	Study site (watershed)	Lat	Lon	Gage elev (m)	Max elev (m)	Area (km^2)	Hydrology starts	Best WY	*n* WY	Mean annual T (°C)	Mean P (mm $year^{-1}$)	Mean AET (mm $year^{-1}$)	AET Sen slope (mm $year^{-2}$)
BB	Bear Br. (East)	44.86	−68.10	276	450	0.11	1988	Oct	25	5.9	1401	382	−7.3
HB3	Hubbard Br. Exp. For. (W3)	43.95	−71.72	527	732	0.42	1957	Jun	54	5.3	1353	489	**−1.0**
HB7	Hubbard Br. Exp. For. (W7)	43.92	−71.66	619	899	0.76	1965	Jun	47	4.5	1485	511	**1.2**
SR	Sleepers River (W9)	44.48	−72.16	520	675	0.41	1992	Sep	18	4.3	1333	520	−2.4
HWF	Huntington For. (Archer Cr.)	43.99	−74.25	514	741	1.30	1995	Oct	18	4.8	1170	446	7.7
BSB	Biscuit Br.	42.00	−74.50	633	1128	9.63	1983	May	29	8.5	1642	626	**9.0**
LR	Leading Ridge(W1)	40.67	−77.94	270	500	1.23	1958	May	53	9.5	1070	626	0.4
FEF	Fernow Exp. For.(W4)	39.05	−79.69	744	867	0.39	1951	May	61	9.4	1450	805	**−1.4**

Note. MAT is from PRISM (1980–2009). Sites are listed geographically across the climate gradient, from northeast to southwest. Sen slopes in bold are significant (Kendall $p < 0.05$). AET: actual evapotranspiration; WY: water years.

are well characterized by Falcone et al. (2010). To ensure adequate precipitation network density, we analysed only the period from 1940 to 2012. Although large basins are more subject to human impacts than continually forested, unmanaged headwater catchments, the HCDN watersheds were specifically chosen by the USGS to minimize these effects.

USGS gages report discharge (volume per time), which was converted to annual run-off (depth per time), using the basin area from a delineation by Falcone et al. (2010). Precipitation (P) data were acquired from PRISM 4-km resolution monthly data product (Di Luzio et al., 2008) and averaged across the river basins delineated by Falcone et al. (2010).

2.3 | Mean PET climatology

To place the results from each of the river basins and small watersheds in the context of the regional climate gradient, we calculated a single long-term mean PET estimate for each watershed. Because Falcone et al. (2010) calculated mean PET for the river basins using 1961–1990 average temperatures from PRISM and the Hamon method (Hamon, 1963), we calculated long-term mean PET for the small watersheds in the same way, following Hamon calculations detailed by Federer, Vorosmarty, and Fekete (1998). We used this Hamon PET estimate only to relate any observed trends in ET to a consistent general metric of the energy available for ET along the northeast-southwest axis of our study area and to illustrate the theoretically strong energy limitation of ET in these watersheds on a Budyko plot (Figure 2). Hamon PET was not used directly in the trend analysis.

3 | METHODS

3.1 | Analysis of long-term AET trends

The significance of monotonic trends in AET was assessed in all AET and correlative climate time series using the Mann-Kendall test (Helsel & Hirsch, 2002). The slope and intercept of trends were estimated using the Sen (1968) method. Statistical tests were conducted in R versions 3.2.4-3.4.3 (R Core Team, 2017) using the zyp (Bronaugh & Werner, 2013) and Kendall (McLeod, 2011) packages. Variation in the ET trend across the regional climate gradient was assessed using ordinary least-squares linear regression between the Sen slope of ET and the mean P and PET at each site.

3.2 | Correlation analysis between AET and climatic drivers

We examined several climatic metrics of both evaporative demand and water availability as possible explanatory variables. We used local records for precipitation over the full WY and meteorological summer (June, July, August; JJA) as potential explanatory variables for ET at the WY time scale. We also included summer Palmer Drought Severity Index (PDSI), which is a modelled metric of soil moisture (Szép, Mika, & Dunkel, 2005); average maximum daily temperature (Tmax); and average DTR for the National Oceanic and Atmospheric Administration (NOAA) climate division of each site (NOAA National Centers

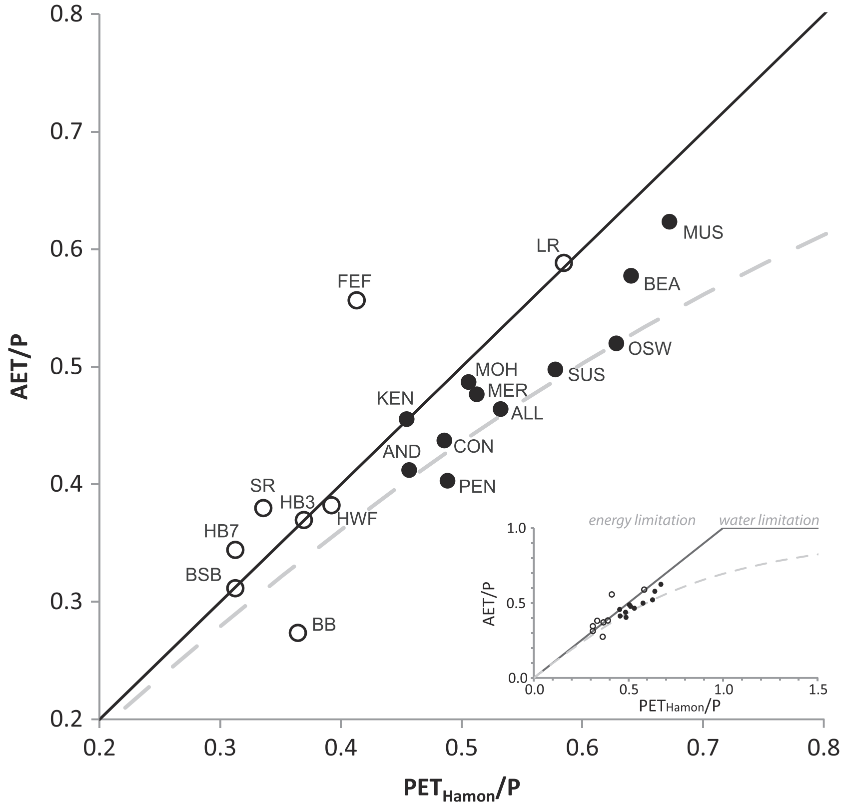

FIGURE 2 All study watersheds are in strongly energy-limited climates. The solid line shows a 1:1 relationship, and the dashed line shows the Budyko curve (Budyko, 1974). Open circles represent the small watersheds; filled circles represent river basins. Abbreviations follow Tables 1 and 2. The inset shows where the sites fall relative to the threshold between energy- and water-limitation of ET (at PET/*P* = 1). AET: actual evapotranspiration; ET: evapotranspiration; PET: potential evapotranspiration

for Environmental Information, 2017). Climate division data were used to reduce the influence of variable record length and data gaps among nearby stations. At Hubbard Brook, where long-term temperature records exist for both the north- and south-facing slopes (Bailey, Hornbeck, Campbell, & Eagar, 2003), we used those to better distinguish the two watersheds. For the river basins, Tmax and DTR means were extracted from gridded monthly PRISM data. We assigned each river basin PDSI values from the climate division of the streamgage.

For the four longest records in the small watersheds (HB3, HB7, FEF, and LR), we also calculated a seasonal average for daily max VPD, which was the variable we could most reliably extract from the available records while minimizing the effect of data gaps and changing data-collection protocols over the study period. We used the hourly temperature and dewpoint data from the nearest station with complete hourly data in the National Climatic Data Center database (https://gis.ncdc.noaa.gov/maps/ncei/cdo/hourly) for the full period of the hydrologic record (Concord, NH; Elkins, WV; and Williamsport, PA, for Hubbard Brook, Fernow, and Leading Ridge, respectively). Because observation times varied (and were sometimes only recorded every 3 hr), we averaged daily VPD at 14:00 EDT +/− 30 min to approximate the peak daily value in a consistent manner for the JJA growing season in all years. The Concord, NH, VPD record was also used for the Connecticut and Merrimack watersheds, and the Williamsport, PA, record was used for the Susquehanna watershed, though the data are only available starting in 1948.

In addition to testing for interannual correlations between AET and temperature and VPD, which are important components of PET, we also included PET estimates from two gridded data products: CRU-TS 3.21 (Harris et al., 2014) and L15 (Livneh et al., 2013, 2015) as correlative time series. The CRU-TS estimate of PET is calculated using a modified Penman-Monteith approach based on the gridded monthly temperature, vapour pressure, and cloudiness data time series for each 0.5° pixel, and assuming a fixed monthly wind climatology. The L15 estimate of PET is based on daily meteorological data (precipitation, maximum and minimum temperature, and mean wind speed) interpolated on a grid of 0.0625° pixels and input into Variable Infiltration Capacity hydrologic model, an energy-balance model that accounts for vegetation characteristics (Liang, Lettenmaier, Wood, & Burges, 1994).

Relationships between ET and correlative climate variables described above were evaluated using Pearson correlations between the time series for ET and the possible explanatory variables available for each study watershed. Variables included energy metrics ($Tmax_{JJA}$, DTR_{JJA}, VPD_{JJA}, PET_{WY}, and PET_{JJA}), water availability metrics (P_{WY} and P_{JJA}), and $PDSI_{JJA}$, which integrates cumulative energy and water metrics. A significance threshold (a) of 0.05 was used, but we also report results with $p < 0.10$ to allow a more complete assessment of the geographic patterns of correlation between AET and various climatic drivers.

4 | RESULTS

4.1 | Long-term AET trends

4.1.1 | AET trends in small watersheds

Four of the small watersheds have hydrological records greater than 45 years in length, whereas four have shorter records ranging from 18 to 29 years. Of the four long records, two (HB3 and FEF) show significant declines of approximately 10% of AET (Table 2; $p = 0.01$ and

0.002, respectively). There was a significant increase in ET at HB7, and no significant trend at LR. Trends are less clear among the shorter records; one site (BSB) showed a significant increase ($p = 0.02$), and the other three showed no significant change. Owing to the length of the time series, these sites are less suited to climate trend analysis and were used here to assess spatial variation in the climatic controls, rather than trends.

4.1.2 | ET trends in river basins

Between 1940 and 2012, AET increased significantly in two watersheds: PEN and KEN (both at the northern limit of our study area; $p = 0.002$ and 0.008, respectively), and decreased significantly in BEA (in the south-western part of the region; Table 1; Figure 3; $p < 0.001$). We found a significant negative relationship between mean annual PET and the rate of AET change since 1940 (Figure 3). The relationship follows:

$$\text{AET Change} = 4.6 - 0.0076(\text{PET})$$

with an r^2 of 0.50 and p of 0.02.

There is no analogous relationship between mean annual P and the rate of AET change.

LOESS smoothed curves of the time series show that there are more complicated patterns than monotonic trends (Figure SA1). Most of the large watersheds show a decline in AET in the early part of the study period (1940 to the 1960s), with mostly increasing trends thereafter (Figure 3). This early decline likely relates to a period of lower-than-average rainfall through the late 1940s to the mid-1960s across the study region (Paulson, Chase, Roberts, & Moody, 1991), culminating in a historically unprecedented drought from 1962 to 1965 (Cook & Jacoby, 1977; Namias, 1966). However, individual site records also display a variety of other short- and long-term dynamics. For example, MOH shows a sudden increase between 1970 and 1990, followed by a decline in AET since then.

4.2 | Correlations with explanatory climate variables

4.2.1 | Correlations with metrics of evaporative demand

Summer maximum temperature correlated significantly and positively with AET in two of the northern small watersheds, BB and HB7, with an equally strong relationship at HWF that was not statistically significant, likely due to the shorter record (Table 3). Among the river basins, MER, SUS, and BEA each showed significant positive relationships between AET and $Tmax_{JJA}$.

No statistically significant relationships were detected between DTR_{JJA} and AET in the small watersheds, but among the river basins, AND showed a negative relationship and MER showed a positive relationship. There were also no significant relationships between VPD_{JJA} and AET at the WY time scale for the four small watersheds for which we had data, though for one of the river basins (MER), a significant positive correlation existed.

Summer and WY gridded PET estimates were generally not significantly positively correlated with observed ET variation across the study watersheds (Table 3). In the small watersheds, the strongest result was at BB, where both summer and WY PET from L15 correlated strongly with AET over the 25-year time series. Better correlations with L15 than with CRU were also seen at HB3, though the correlations are weaker. Among the river basins, there were nearly as many significant negative correlations between AET and PET metrics (see ALL and MUS) as many significant ($p < 0.05$) positive correlations (see MER and BEA). It is worth noting that the PET values in the CRU-TS dataset differ systematically from Hamon PET and also from

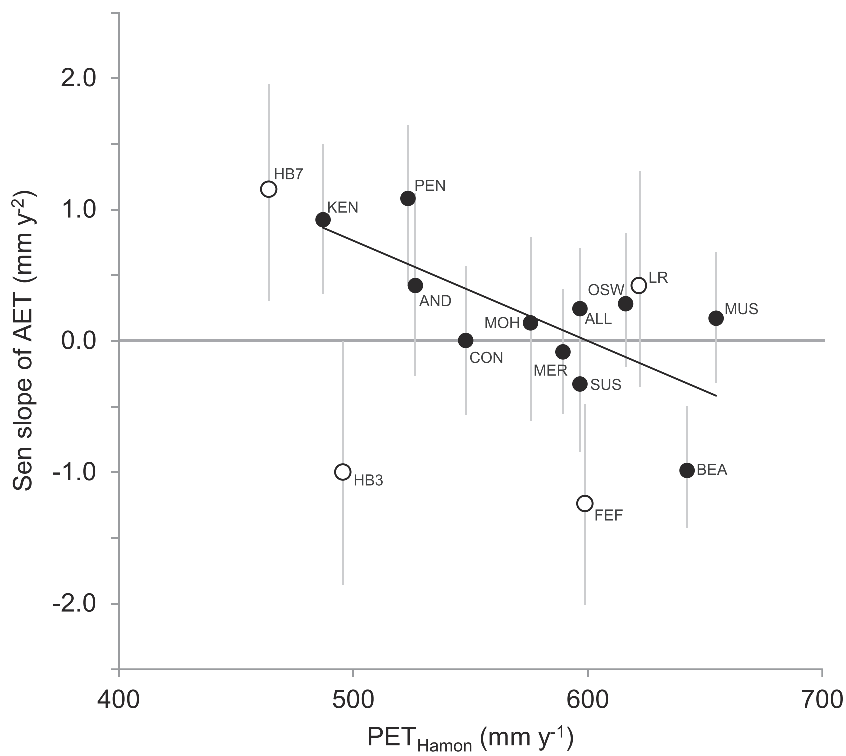

FIGURE 3 The observed trend in AET inferred from water balance is related to mean annual PET. Open circles represent the small watersheds; filled circles represent river basins. Watershed abbreviations follow Tables 1 and 2. Only small watersheds with >45 years of data are included here. Trends are expressed using Sen's monotonic slope estimate and shown with 95% confidence intervals; trends with confidence intervals not overlapping the zero line are significant at $a = 0.05$. The regression line shown, for the large watersheds only, has a slope of −0.0076 and an intercept of 4.6 mm year^{-2}. AET: actual evapotranspiration; PET: potential evapotranspiration

AET inferred from water balance (Figure SA2). PET from the L15 dataset also exceeds observed AET but by substantially less.

4.2.2 | Correlations with metrics of water availability

There were significant positive correlations between ET and both WY and summer precipitation at the three southernmost small watersheds (those with the greatest PET), plus conflicting results from the two watersheds at HB for summer precipitation (one positive, one negative; Table 3). The correlation was equally strong but not significant at SR due to the shorter record. This implies that water availability during the summer at least transiently limits ET during some years at these sites. Seven of the 11 river basins showed significant positive correlations with P_{JJA}, which was the strongest and most consistent correlative variable overall (Table 3).

WY precipitation correlated significantly with ET at the three southernmost small watersheds, and in two of these, it correlated much more strongly than did JJA precipitation. A stronger relationship with P_{WY} than with P_{JJA} was also seen at HB7. One possible explanation for this disjunct geographic pattern might be the longer leaf-on season of both the southernmost (predominantly deciduous) watersheds and the conifer-dominated HB7, relative to the northern deciduous watersheds.

PDSI was substantially less useful than other metrics as a predictor of ET, with a counter-intuitive negative correlation in the Merrimack watershed, the only significant relationship with PDSI across both the river basins and small watersheds (Table 3). Preliminary analyses using the Standardized Precipitation Evapotranspiration Index (Vicente-Serrano, Beguería, & López-Moreno, 2010) instead of PDSI in a subset of watersheds yielded similar results (not shown).

5 | DISCUSSION

5.1 | Direction of ET trends differ across the study region

We found a number of significant trends in AET in both small watersheds and large basins across the study region. These trends are not consistent across the entire study region and in fact differ systematically in direction across the modest climate gradient we examined, with cooler northern watersheds experiencing an increase in AET and warmer watersheds to the southwest seeing a modest decrease (Figure 3, Tables 1 and 2). Increasing AET in the coolest climates likely relates to the alleviation of energy limitation with longer and warmer leaf-on seasons (Cleland, Chuine, Menzel, Mooney, & Schwartz, 2007; Dragoni et al., 2011; Keenan et al., 2014). Declines in AET at the warmer sites are somewhat unexpected given the overall warming across the region during the study period (Table SA2; Hamburg, Vadeboncoeur, Richardson, & Bailey, 2013; Hayhoe et al., 2007; Trombulak & Wolfson, 2004).

TABLE 3 Pearson correlation coefficients (R values) between AET and eight possible explanatory variables across small and large watersheds

	Metrics of energy limitation							Metrics of water limitation		
	Tmax JJA °C	DTR JJA °C	VPD JJA kPa	CRU PET WY mm year^{-1}	L15 PET WY mm year^{-1}	CRU PET JJA mm year^{-1}	L15 PET JJA mm year^{-1}	P WY mm year^{-1}	P JJA mm year^{-1}	PDSI JJA (unitless)
Small watersheds										
BB	0.41**	0.36*	NA	0.13	0.45**	0.26	0.50**	0.13	−0.14	−0.29
HB3	0.02	0.04	0.11	−0.19	0.24*	0.00	0.27*	−0.16	−0.31**	−0.20
HB7	0.38**	0.24	0.17	0.36	0.22	0.34	0.17	0.54**	0.46**	0.35
SR	0.02	0.04	NA	−0.21	0.02	−0.15	0.05	0.15	0.44*	0.08
HWF	0.40*	0.01	NA	0.32	0.16	0.27	0.06	0.20	0.29	0.07
BSB	0.02	−0.08	NA	−0.13	0.08	0.01	0.03	0.70**	0.37**	0.19
LR	−0.21	−0.26*	−0.27*	0.12	−0.05	−0.13	−0.22	0.32**	0.45**	0.09
FEF	0.12	0.23*	−0.09	0.15	0.14	0.19	0.04	0.66**	0.39**	0.13
River basins										
PEN	0.09	−0.11	NA	−0.05	0.06	−0.03	0.08	0.29**	0.19	−0.10*
KEN	0.19	−0.02	NA	−0.05	0.02	0.05	0.17	0.19	0.25**	−0.06*
AND	0.00	−0.24**	NA	0.17*	−0.05	−0.10	−0.13	−0.08	0.44**	NA
MER	0.34**	0.23**	0.28**	−0.06	0.18	0.37**	0.42**	−0.11	−0.04	−0.32**
CON	−0.04	−0.15	−0.23*	0.03	0.01	−0.12	0.00	−0.08	0.47**	0.18*
MOH	−0.13	−0.22*	NA	−0.05	−0.20	−0.17	−0.10	−0.20*	0.38**	NA
SUS	0.26**	0.06	0.18	−0.07	0.13	0.16	0.08	−0.20*	0.32**	NA
ALL	0.15	−0.04	NA	−0.32**	0.19	0.04	0.01	−0.12	0.36**	NA
BEA	0.26**	0.23*	NA	0.08	0.25*	0.24**	0.08	−0.07	0.04	−0.16*
MUS	−0.11	−0.15	NA	−0.09	−0.22*	−0.23*	−0.33**	0.07	0.38**	0.06*
OSW	0.13	−0.14	NA	0.17	−0.14	0.12	0.02	0.04	0.13	−0.12*

Note. Because PDSI is cumulative, it is not examined as a correlate where our WY splits the main growing season. Watershed abbreviations follow Tables 1 and 2. ** indicates that a correlation is significant at $p < 0.05$; * indicates correlation is significant at $p < 0.10$. "NA" indicates analyses that were not conducted at a given location. AET: actual evapotranspiration; DTR: daily temperature ranges; PDSI: Palmer Drought Severity Index; PET: potential evapotranspiration; Tmax: average maximum daily temperature; VPD: vapour pressure deficit; WY: water years.

The observed trends are best interpreted in the context of trends in the climatic variables hypothesized to limit AET. The primary temperature metric we examined, maximum daily temperature for June–August, showed little evidence of warming overall (Table SA2), particularly for the records that start in the early 1940s, an anomalously warm decade in parts of the northern hemisphere (Brönnimann, 2005). However, DTR for June through August declined significantly across most studied watersheds, indicating substantial night time warming during the summer (Table SA2), which generally indicates higher dewpoints and reduced daytime VPD. Regionally, warming has been stronger in the winter and shoulder seasons (spring and fall) than in the summer (Hamburg et al., 2013; Hayhoe et al., 2007), and reductions in DTR have been observed worldwide (Thorne et al., 2016). Increases in precipitation have also been seen throughout the study region, but they are most significant in more northern sites.

Increases in AET have long been a hypothesized consequence of a warming climate in temperate regions where ET is strongly energy-limited. On balance, long-term analyses generally find these increases to some extent. Walter et al. (2004) found a significantly positive area-weighted AET trend across six large basins in the United States, though most basins did not show a significant trend individually. Both nationally and across the northeast, Jung et al. (2013) found that significant increases in AET generally outnumbered decreases, and more specifically in Maine and New Hampshire, Huntington and Billmire (2014) found increases in ET in 16 out of 22 basins. At a smaller spatial scale in eastern Pennsylvania, a 44-year analysis of a single 7-km^2 mixed agricultural-forest watershed also showed a strong increasing AET trend (Lu et al., 2015), which was attributed mostly to increased temperatures and a longer growing season. Kramer et al. (2015) found an increase in AET across most eastern U.S. hydrologic regions, which they attributed in part to increased transpiration during longer growing seasons. Some models and global-scale studies have predicted declines in AET, or smaller gains than would be expected from climate drivers alone, associated with increased water use efficiency as stomatal conductance is reduced in response to increasing atmospheric CO_2 concentrations (Mao et al., 2015; Milly & Dunne, 2016; Szilagyi et al., 2001). Declines in VPD (Seager et al., 2015) could also explain declining AET and might be consistent with relatively stable summer maximum temperatures accompanied by declining DTR (Table SA2). However, the VPD records we examined show no evidence of such a decline.

5.2 | Both precipitation and temperature explain interannual variation

Correlation analysis (Table 3) suggests that both water and energy availability influence the interannual variation in AET across the watersheds we examined. In the small watersheds, there is a geographic pattern in the correlations, with temperature metrics correlating with AET only in the north, and precipitation metrics strongly correlated with AET in the south and only sometimes in the north. These patterns indicate that transient periods of water limitation are more limiting to annual AET than temperature in the southern watersheds. Temperature correlates significantly and more strongly with AET than does precipitation only at three watersheds (BB, MER, and BEA). The apparent greater importance of precipitation than temperature as a control of AET is unexpected in a region where ET is traditionally considered to be energy-limited (i.e., precipitation substantially exceeds PET; Tables 1 and 2; Figure 2). Among the larger river basins, there is no clear geographic pattern in which correlation is strongest, though significant correlations with summer precipitation were more prevalent and stronger than with summer temperature.

The strength of correlations between seasonal or annual climate data and annual AET might be limited by the fact that averages at annual or seasonal scales have limited power to capture variability that is controlled by processes that operate at daily-to-hourly time scales. This may be due to non-linear responses of AET to its controlling variables. For example, both in simulation models of ET (Fatichi & Ivanov, 2014) and in analyses of flux-tower data (Zscheischler et al., 2016), the prevalence of short periods of meteorological conditions favouring high AET rates within each year better explained annual AET than did monthly to annual variables of the type we employed, which are typically available at the scale of large watersheds. These results suggest that higher frequency meteorological data might need to be incorporated into an analysis like ours to account for the importance of short time scales where the controls on AET are most evident. Such an approach may greatly improve our ability to understand the relative importance of drivers of changes in AET, compared with the seasonally averaged approaches used in our analyses.

Interestingly, the simple temperature and precipitation metrics showed the strongest correlations with AET at the WY time scale. More complicated metrics intended to more closely capture the factors limiting AET (VPD, PET, and PDSI) showed fewer, if any, significant relationships with ET. In fact, VPD and PDSI each showed only one significant correlation with $p < 0.05$ among the 19 watersheds examined (Table 3), roughly the type I error rate expected from random chance. We also conducted preliminary analyses on similar related metrics at a subset of sites, including Standardized Precipitation Evapotranspiration Index, mean pressure at nearest station, and solar radiation at nearest station, but none of these analyses improved upon the correlations we found with simple metrics like P_{JJA} and T_{JJA} (data not shown).

Analyses of PET_{WY} and PET_{JJA} from two different gridded data products also showed relatively few significant correlations with AET. There was also remarkably little improvement in these correlations between the coarse-resolution CRU data set and the finer-resolution L15 (but see BB and HB3; Table 3). Contrary to what would be expected in a climate where precipitation greatly exceeds PET on an annual basis, we found that the interannual variation in PET from these gridded datasets explains little of the observed variation in AET.

5.3 | Potential confounding factors in our analyses

Our water balance estimate of AET is not a direct measurement and thus could be influenced by changes in other water budget terms. Changes in the efficiency of run-off generation can arise with land-use change or changes in precipitation intensity, which would make AET appear to change. Similarly, groundwater storage changes are not accounted for in the water balance. Trends in groundwater levels are difficult to quantify, and the evidence is mixed in the study region

(Brutsaert, 2010; Dudley & Hodgkins, 2013; Kramer et al., 2015; Shanley, Chalmers, Mack, Smith, & Harte, 2016), but on balance groundwater levels appear to be lowering over the time period we examined. Nonetheless, the effects of the exclusion of groundwater estimates on the results of any water balance analysis deserve consideration (Sharma & Walter, 2014). A decline in inputs to groundwater would reduce the effect of an increasing AET trend on water balance, so the real trends in AET in the northern river basins might be greater than those we calculated. Using the Variable Infiltration Capacity hydrologic model, Parr and Wang (2014) found increasing run-off ratios, but no AET trend for the Connecticut watershed since 1950. We also found no AET trend for this basin looking only at water balance (Table 1).

We considered the possibility that variation in certain ET drivers could create interannual variation in storage (e.g., greater storage in a year of above-average precipitation), leading to spurious correlations. However, the results do not support the idea that this effect drives the trends. For example, we see significant correlations between summer precipitation and AET across most of our small watersheds (Table 3). This includes those with WY that end in the fall (SR, CON), in which the effect might be expected to be the strongest, and also in WY that end late in the spring (FEF, LR, HB, KEN, and MUS), a time of year in which it is difficult to imagine that storage is highly determined by precipitation in the preceding growing season.

Our examination of two different size classes of watersheds leveraged their complementary advantages. Continually forested small reference watersheds offer a high degree of certainty that land cover did not change over the studied time period and, in some cases, offer very accurate and complete data sets for both precipitation and run-off. There is somewhat more hydrologic uncertainty in the large river basins (e.g., in precipitation interpolations and changes in storage), and land cover has been subject to change. Over the 20th century, Yang et al. (2015) show that east-coast agricultural land declined from 18% to 11%, whereas forest cover increased modestly from 67% to 70%, and impervious surface increased from 1% to 3%. There have also been changes in impoundment, interbasin transfers, and groundwater use over time. The HCDN watersheds were selected to minimize these effects (Slack & Landwehr, 1992), but such changes cannot be eliminated with basins this large in a heavily populated region. On the other hand, large watersheds are more representative of the region in terms of land cover, elevation, and soil types, than are the small watersheds we studied. Despite these differences between the large and small watersheds, we found broadly similar geographic patterns in long-term AET trends (Figure 3) and in correlations between AET and metrics of energy and water limitation (Table 3). This provides a greater level of confidence that our conclusions about trends and drivers are not predominantly driven by factors such as land-cover change or human water use and are also not limited to small, forested, upland catchments.

5.4 | Vegetation mediation of climate drivers of ET

There are a number of ways in which vegetation can complicate relationships between climate and ET. In landscapes with high vegetative cover, the transpiration component of ET dominates (Jasechko et al., 2013), and variation in ET is driven by plant physiology (i.e., regulation of gas exchange via stomatal opening and closure) and influenced by plant functional group differences in productivity, structure, phenology, physiological responses to stress, and access to soil water. Therefore, changes in the atmospheric demand for water may not directly explain changes in ET, as water balance and hydrologic cycling may respond to a number of physiological effects as well. Increasing length of the leaf-on growing season (Dragoni & Rahman, 2012; Richardson et al., 2009; Schwartz & Reiter, 2000) may increase total annual ET. On the other hand, climate change scenarios for this region project more frequent water limitation of forest productivity in spite of modest increases in precipitation, due to less reliable precipitation timing and greater evaporative demand (Douglas & Fairbank, 2011; Hayhoe et al., 2008; Pourmokhtarian et al., 2017; Swain & Hayhoe, 2014; Tang & Beckage, 2010). Recent synthesis efforts indicate that the forests of humid regions like the Northeastern United States may be more sensitive to drought than previously thought (Choat et al., 2012; Pederson et al., 2012; Wright, Williams, Starr, McGee, & Mitchell, 2013), which is supported by our finding of some degree of water limitation of ET in the southern part of our study region.

Studies of flux tower data and carbon isotope ratios in tree rings have shown long-term increases in water-use efficiency (WUE; the ratio carbon assimilation to transpiration; Nobel, 2005) driven largely by increasing atmospheric CO_2 (Franks et al., 2013; Keenan et al., 2013). These trends may be reflected in a decline in AET and increase in river discharge at the global scale (Gedney et al., 2006). However, other factors influencing forest productivity, including changes in climate (Hayhoe et al., 2007), nitrogen deposition (Bowen & Valiela, 2001), and long-term species change (Caldwell et al., 2016) may make it difficult to detect these changes in WUE (Mao et al., 2015). Changes in acid deposition and recovery therefrom may also directly influence WUE over time (Thomas, Spal, Smith, & Nippert, 2013). Indeed, at the regional scale, the lack of a consistent regional decline in AET (Figure 3) implies that the direct CO_2 effect is small relative to changes driven by climate, such as temperature and precipitation, unexamined climatic drivers such as radiation (Dai et al., 2006; Wild, 2012) and wind (McVicar et al., 2012; Pryor et al., 2009), and the negative feedback between AET and VPD (Huntington, 2008). Furthermore, physiological relationships at the leaf level often do not scale linearly to the canopy or regional level (Guerrieri, Lepine, Asbjornsen, Xiao, & Ollinger, 2016; Wullschleger, Gunderson, Hanson, Wilson, & Norby, 2002). To the extent that WUE is increasing, we would expect it to offset the increases in ET that are hypothesized in a warmer climate with a longer leaf-on season, particularly in watersheds dominated by deciduous forests. The direct CO_2 effect on WUE may therefore explain some of the decline in AET seen in the more southern watersheds (and HB3), but it runs counter to the trend observed in the north, where ET is increasing.

6 | CONCLUSIONS

We found that 74-year trends in AET, calculated from hydrologic water balance, varied systematically across a climate gradient in 11 Northeastern U.S. river basins, with increasing AET in the coolest, most energy-limited part of the region, and declining AET in the south. Of the four small watersheds examined with records longer than 45 years, three had significant trends, two of which fit this regional pattern.

Correlation analysis of AET with climate metrics over all 19 watersheds also implied that limitation of ET by summer temperature was greater in the northern part of the study region, whereas at least transient limitation by summer precipitation was more prevalent in the southern part of our study region. Overall, WY AET correlated significantly with summer precipitation in more than half of the watersheds examined. This result is surprising because even at the southern sites, annual precipitation greatly exceeds PET. Understanding how the controls on ET trends vary across the Northeastern United States, where energy is generally more limiting than moisture, is important for predicting future changes in water balance as the climate changes.

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SUPPORTING INFORMATION

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