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Long-Term Soil Moisture Patterns in a Northern Minnesota Forest

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Forest hydrological and biogeochemical processes are highly dependent on soil water. At the Marcell Experimental Forest, seasonal patterns of soil moisture have been monitored at three forested locations since 1966. This unique, long-term data set was used to analyze seasonal trends in soil moisture as well as the influence of time-lagged precipitation and modified Thornthwaite modeled potential evapotranspiration (PET) on seasonal soil moisture at three depths (0–15, 76–107, and 198–229 cm). Despite no change in precipitation during the 45-yr record, mean annual soil moisture from 0 to 228.6 cm has declined ($p < 0.05$). Precipitation minus PET was found to account for >50% of the variability in seasonal soil moisture ($p < 0.001$). Our findings suggest that further increases in mean annual temperature and evapotranspiration may lead to decreases in soil moisture. These decreases could negatively impact forested ecosystems in northern Minnesota.

Abbreviations: CCP, critical climate period; MEF, Marcell Experimental Forest; PET, potential evapotranspiration.

Soil moisture influences ecosystem processes, including soil pedogenesis, the type and abundance of flora and fauna, decomposition, and nutrient availability (Rodriguez-Iturbe, 2000; Porporato et al., 2002, 2004; Eamus, 2003; Jenerette and Lal, 2005). In forested ecosystems, extreme high or low soil moisture conditions can lead to decreased photosynthesis (Chaves et al., 2002) and root (Kuhns et al., 1985) and tree growth (Hinckley et al., 1979), changes in phenology (Borchert, 1994), and increased susceptibility to diseases and pathogens (Desprez-Loustau et al., 2006). Soil moisture is so vital to hydrological, biological, and biogeochemical processes that the European Space Agency designated it to be an essential climate variable (Wagner et al., 2012). Soil moisture is also critical to climate forecasting and can impact management decisions regarding future climate scenarios, flood and drought mitigation, and land management policy (Adams et al., 1991; Eltahir, 1998; Norbiato et al., 2008). Despite its important role in forested ecosystems, soil moisture is rarely measured, especially with respect to depth in the soil profile and time.

Although in situ measurements of soil moisture are becoming more prevalent, current understanding of in situ soil moisture dynamics is limited in time, space, and depth (Baker et al., 1979; Passioura, 1982; Adams et al., 1991; Hollinger and Isard, 1994; Stephens, 1995; Western and Grayson, 1998; Robock et al., 2000; Rodriguez-Iturbe, 2000; Kirkham, 2004; Dorigo et al., 2011). Specifically, long-term records of in situ soil moisture are rare (Adams et al., 1991). The current understanding of soil moisture dynamics is limited in time (Robock et al., 2000),

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to agricultural ecosystems (Hawley et al., 1983; Robock et al., 2000), or to the effects of disturbance events (Adams et al., 1991; Robertson et al., 1993; Guo et al., 2002). In many cases, the understanding of available soil water is based on theoretical models (Huang et al., 1996; Nijssen and Lettenmaier, 2001; Wagner et al., 2003). While modeling approaches are useful, their accuracy can be variable due to the complexity of upscaling from individual plant–water use relationships to larger, heterogeneous landscapes.

Understanding soil moisture dynamics may become increasingly important under changing climates. In the north-central United States, temperatures are expected to increase while precipitation is expected to become more variable (Christensen et al., 2007). The effects of climate change on soil moisture will depend on the timing and severity of the changes. Globally, soil moisture has increased despite increases in temperature, suggesting that increased precipitation will offset increased plant water demand with warmer temperatures (Robock et al., 2000). Regional studies of soil moisture in northern forests have shown increases in soil moisture with time (Vinnikov et al., 1996; Groffman et al., 2012), and some studies have already shown an increase in plant water demand and C sequestration due to increasing temperatures, especially in northern forests (Pastor and Post, 1988; Hyvönen et al., 2007). In light of anticipated climatic changes, long-term, spatially resolved measurements of in situ available soil water, weather (e.g., temperature, precipitation, and humidity), and plant water use are needed to better understand the mechanisms behind complex forested ecosystems. Increased knowledge about these mechanisms can help us to better understand the resilience and resistance of these ecosystems under increased water stress.

To quantify the role of changing climatic regimes on soil moisture in northern hardwood forests, we examined 45 yr of in situ soil moisture measurements from the Marcell Experimental Forest in north-central Minnesota. This record of soil moisture is among the longest ongoing, continuous records of soil moisture currently available. The relationship between available soil water and climate forcings (e.g., precipitation, temperature, evapotranspiration) was analyzed using a critical climate period (CCP) analysis (Craine et al., 2009). The CCP can be defined as the window of time during which a climatic variable explains the maximum variation in a response, in this case, available soil water. The CCP has been used to assess the relationship between climate and grass culm production (Craine et al., 2010), bison weights (Craine et al., 2009), and grassland productivity (Craine et al., 2012). The CCP has not previously been used to explore the relationship between climate variables, but it allows for general knowledge of how changes in climate variability that occur during one season may influence the available soil water and thus ecosystem processes following a climatic event. Specifically, this study sought to: (i) investigate the trends in climatic patterns, including available soil water, from 1966 to 2011 at the Marcell Experimental Forest; and (ii) identify the CCP that influences available soil water in these systems.

STUDY SITE

The Marcell Experimental Forest (MEF) (47.5° N, 93.5° W) is located on the eastern edge of the Chippewa National Forest in north-central Minnesota (Fig. 1). Established by the U.S. Forest Service in the 1960s, early research focused on the hydrology of peatlands. Meteorological and hydrological monitoring began in 1961 (Sebestyen et al., 2011; Verry et al., 2011b). The 1100-ha forest is divided into six research watersheds, each of which contains a peatland bog or fen and a surrounding upland ecosystem. This study focused on the two control watersheds, S2 and S5, that have not been disturbed since establishment of the experimental forest (Table 1).

The climate at the MEF is continental, with cold, dry winters and warm, moist summers. Mean annual precipitation from 1961 to 2012 was 78.0 cm, with approximately one-third of the precipitation occurring as snowfall and the remainder as rain (Sebestyen et al., 2011). The mean annual temperature from 1961 to 2012 was −15.1°C in January and 18.9°C in July. Aspen (*Populus tremuloides* Michx. and *Populus grandidentata* Michx.) dominates the upland landscape, with smaller populations of red pine (*Pinus resinosa* Aiton) and mixed hardwoods (*Tilia americana* L., *Acer saccharum* Marshall, and *Acer rubrum* L.) also common in the uplands. Upland soils are predominately deep glacial tills. Water drains through mineral soils on low-elevation ridges (approximately 20-m relief) through peatlands to ephemeral streams or the regional groundwater system (Sebestyen et al., 2011; Verry et al., 2011a).

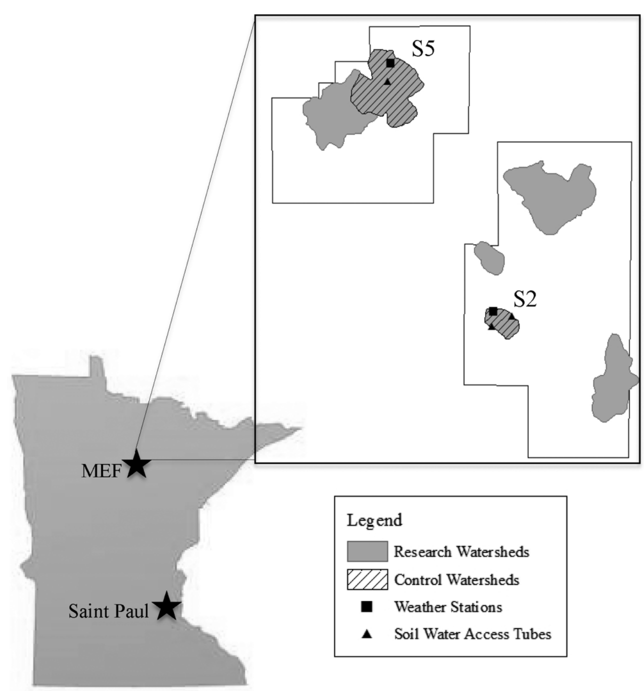

Fig. 1. The Marcell Experimental Forest (MEF) is located in north-central Minnesota and consists of six research watersheds, two of which are control watersheds that have not undergone any management.

Table 1. Characteristics of upland forest management treatments for two established research watersheds at the Marcell Experimental Forest.

Watershed	Total area	Max. elevation	Outlet elevation
	ha	m	m
S2	9.7	430	420
S5	52.6	438	422

METHODS
Climatic Trends

Air temperature at the S2 watershed has been recorded since 1961, while the temperature at the S5 watershed has been recorded since 1962. Daily maximum and minimum temperatures were recorded using Belfort Model 594-1 hygrothermographs (Belfort Instruments) and were then averaged to calculate mean daily air temperatures at the two meteorological stations (Rosenberg et al., 1983). Daily precipitation data have been collected since 1961. The stations are equipped with Belfort Universal recording precipitation gauges (Belfort Instruments); the S2 watershed was updated with an ETI NOAH IV digital rain gauge (ETI Instrument Systems) in 2009. Snow water equivalent was calculated by first equipping the rain gauges with antifreeze to melt snow and then obtaining weekly precipitation measurements using a temperature-compensated spring scale (Sebestyen et al., 2011).

To evaluate trends in climatic data from 1966 to 2012 at the MEF, a nonparametric Mann–Kendall trend test (Mann, 1945) was fitted to temperature and precipitation records. This test is more robust for trend detection than linear regression because it does not require normality of the data set and the method is insensitive to missing data. The mean annual temperature was calculated by averaging mean daily temperatures between the S2 and S5 watersheds and then averaging this value across years. Mann–Kendall trends were then assessed on mean annual temperature, mean maximum annual temperature, mean minimum annual temperature, and mean seasonal temperature. Seasons were defined as winter (January, February, and March), spring (April, May, and June), summer (July, August, and September), and fall (October, November, and December) and are consistent with seasons previously defined at the MEF (Sebestyen et al., 2011).

Total annual precipitation was calculated by averaging total daily precipitation between Watersheds S2 and S5 and then summing the daily precipitation measurements across the calendar year. Mann–Kendall trends were then assessed on total annual precipitation and total seasonal precipitation. Again, seasons were defined as winter (January, February, and March), spring (April, May, and June), summer (July, August, and September), and fall (October, November, and December). Many studies have found an increase in the severity of droughts and floods with time that are not always reflected in changes in total annual precipitation (Alexander et al., 2006; Dai et al., 1998; Easterling et al., 2000; Meehl and Tebaldi, 2004). To evaluate whether changes in the size, frequency, and severity of rainfall events has changed with time, trend tests were performed on mean annual number of days since the last rainfall (used as an index of drought), annual maximum intensity of rainfall (mm h^{-1}), mean annual total volume of rainfall per storm, and frequency of rainfall events.

Additionally, we applied a Mann–Kendall trend test to modeled potential evapotranspiration (PET) because PET can account for up to 65 to 66% of annual precipitation at the MEF (Brooks et al., 2011). Daily PET was calculated using a modified Thornthwaite equation (Pereira and Pruitt, 2004; Thornthwaite, 1948), which corrected for temperatures greater than 26°C and photoperiods greater or less than 12 h. Daily PET calculations were averaged to obtain seasonal and yearly estimates of PET.

Soil Moisture Trends

Soil moisture has been measured since 1966 at three sites within the two control watersheds (Table 2). Measurements typically occurred three times per year, once each at leaf-out, tree senescence, and before soil freeze. These times roughly fell in May, September, and November, but exact measurement days and months varied depending on the year, weather, and length of the growing season. Seasonal snow cover and frozen soils prevented measurements during winter months (January–March). Data were measured using the neutron probe technique (Brakensiek et al., 1979) with a Troxler Model 105 depth moisture gauge before 1990 and a Series 4300 gauge from 1990 to 2012. Data were measured as soil moisture percentage and were subsequently converted to centimeters of available soil water per sampling depth. Soil moisture was measured in 30.4-cm increments from 15.2 cm to the depth of the probe (Table 2). Soil moisture was measured gravimetrically for 0 to 15 cm using a drying oven (Gardner, 1986) because measuring moisture in near-surface soil horizons using the neutron probe technique can lead to spurious measurements due to neutrons escaping from the soil surface (Bell et al., 1987). The neutron probe collected the volumetric water content (θ_v) of the soil, which was subsequently converted to centimeters of water per horizon by multiplying θ_v by the sampling

Table 2. Characteristics of three historical soil water monitoring locations at the Marcell Experimental Forest.

Site	Record start date	Probe depth	Soil type	Soil texture†	Drainage class	Cover type	Stand age‡	Slope
		m					yr	%
S2-E	Oct. 1967	2.3	Haplic Glossudalf	fine sandy loam	well drained	aspen	96	1–8
S2-S	Apr. 1968	3.2	Haplic Glossudalf	fine sandy loam	well drained	aspen	96	1–8
S5	Sept. 1966	2.3	Typic Udipsamment	fine sandy loam	poorly drained	aspen	92	1–10

† Source: Soil Survey Staff (2013).

‡ As of 2013.

depth. Values were then converted to centimeters of available soil water by subtracting the soil permanent wilting point (defined as θ_v at −1500 kPa). The soil field capacity and permanent wilting point for each sampling depth were determined at the time of access tube installation.

A Mann–Kendall trend test was used to assess shifts in soil moisture with time. Because there was interannual variation in the sample timing of available soil water measurements, only measurements collected during May, September, and November were used in the analysis; this created minor gaps in the data set. Total available soil water from each sampling depth was aggregated to obtain total site available water to a depth of 228.6 cm. Data were analyzed at the sites and individual watershed levels and as the averaged of the two watersheds. To obtain the averaged watershed data, total available soil water was averaged for all months of measurement and the two watersheds to obtain one annual soil moisture value. Relationships were similar across all sites, and subsequent results display averaged data for the control watersheds.

Differences between sites, depths, and months were analyzed using a Tukey–Kramer test (Tukey, 1949; Kramer, 1956) for differences in means. The two control watersheds were included in this analysis, and differences between sites were analyzed at three depth increments (0–15.2, 76.2–106.7, and 198.1–228.6 cm) and 3 mo (May, September, and November). These depths were chosen to represent upper, middle, and lower soil horizons. Analysis included individual Tukey–Kramer tests for each depth increment during each month.

Critical Climate Period

To investigate the relationship between available soil water and climate forcing using CCP, precipitation measurements were taken from the MEF precipitation gauge located nearest to each soil water sampling site. Precipitation the day before the soil moisture measurement date was used as a starting point for analysis. Precipitation was iteratively summed each day back to a period of 750 d. The analysis was performed for each of three depths (0–15.2, 76.2–106.7, and 198.1–228.6 cm) and for 3 mo (May, September, and November) at each depth. Analysis was repeated across all three sampling locations. The optimized CCP was then calculated by minimizing the variation and deviation from the maximum R^2 across all three sites (within the two watersheds). The CCP was calculated for the summed daily precipitation, daily PET, and daily precipitation minus PET. For all climate variables, an optimized CCP was determined for the soil depth and month of measurement. All analyses were computed using SAS Version 9.3 (SAS Institute).

RESULTS

Climatic Trends

Mean annual temperature at the MEF has increased by 0.5°C per decade since 1966 ($p < 0.001$; Fig. 2). Winter temperatures account for much of the annual increase (0.7°C per decade January–March), with smaller yet statistically significant increases occurring during the spring (0.3°C per decade April–June), summer (0.3°C per decade July–September), and fall (0.04°C per decade October–December) (Fig. 2). Mean annual minimum and maximum temperatures followed similar patterns throughout the 45-yr study period.

From 1966 to 2012, we detected no statistically significant trends in total annual precipitation at the MEF (Fig. 3). Additionally, seasonal precipitation, frequency of rainfall events, volume of rainfall, and number of consecutive days without rainfall have not changed during the 45-yr study period.

Despite no changes in precipitation, annual available soil water at the MEF has been declining since 1966. Mean annual available soil water from 0.0 to 228.6 cm has decreased at a rate of 0.8 cm per decade ($p < 0.03$) (Fig. 3). The strongest decline has been in May available soil water (1.3 cm per decade, $p < 0.0001$), while no statistically significant changes in September or November available soil water were detected (Fig. 3). Some of the largest decreases in mean annual available soil water occurred between 45.7 and 259.0 cm in the soil profile (Table 3). No statistically significant changes in mean annual available soil water were found in deeper soils.

While the rate of change in modeled mean annual PET (1966–2012) is not statistically significant, the variability around the mean annual PET has increased during the same time period (Fig. 4). Significant increases in PET were found for the summer (July, August, and September: 1.2 cm per decade, $p = 0.02$) and spring (April, May, and June: 1.4 cm per decade, $p = 0.09$). September PET has been increasing at a rate of 1.2 cm per decade ($p = 0.05$), while no changes were found in May and November PET (data not presented).

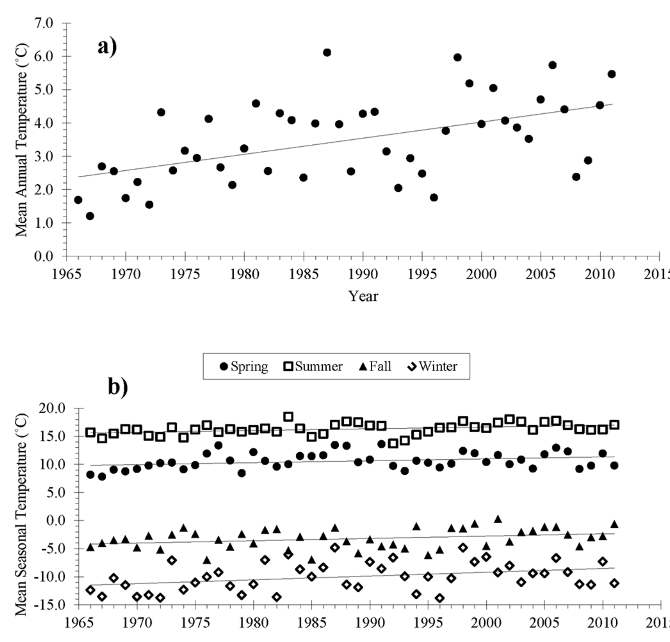

Fig. 2. Mean (a) annual and (b) seasonal air temperature at the Marcell Experimental Forest from 1966 to 2010. Mean annual air temperature has significantly increased since the start of measurement ($p < 0.001$). Summer air temperature has increased since 1966 ($p < 0.001$), as have winter, spring, and fall air temperature ($p < 0.05$). Lines represent statistically significant linear regressions.

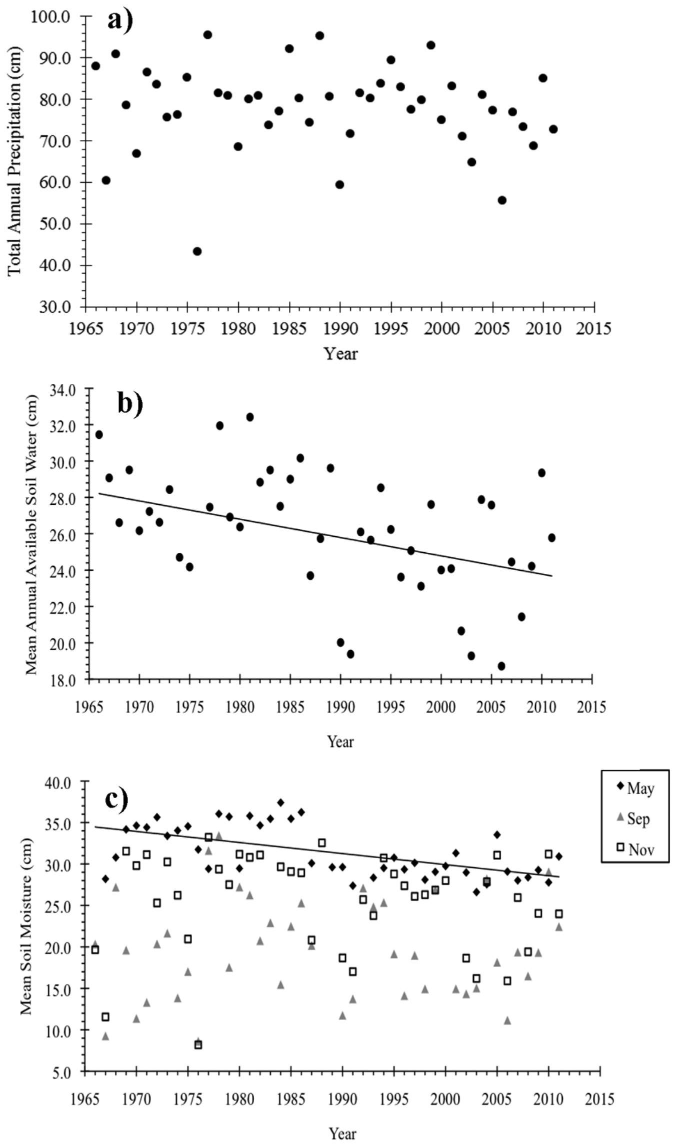

Fig. 3. Mean (a) annual precipitation, (b) annual available soil water, and (c) seasonal available soil water at the Marcell Experimental Forest. Mean annual soil water has been decreasing at a rate of 0.08 cm yr^{-1} (p < 0.03) and May mean available soil water has been decreasing at a rate of 0.13 cm yr^{-1} (p < 0.0001). Lines represent statistically significant linear regressions.

Table 3. Linear regression variables for mean annual soil moisture against year for each sampling depth at Marcell Experimental Forest Watersheds S2 and S5.

Soil depth	R^2	p value	α	β
cm				
0.0–15.2	0.39	<0.01	0.024	−45.85
15.2–45.7	0.02	0.912	−0.001	6.192
45.7–76.2	0.61	<0.0001	0.028	−52.15
76.2–106.6	0.46	<0.01	−0.018	38.77
106.6–137.1	0.17	<0.01	−0.014	30.78
137.1–167.6	0.49	<0.001	−0.018	38.77
167.6–198.1	0.32	0.030	−0.010	22.79
198.1–228.6	0.69	<0.0001	−0.040	82.63
228.6–259.0	0.85	<0.0001	−0.072	145.5
259.0–289.5	0.00	0.776	0.001	−0.745
289.5–320.0	0.00	0.982	0.000	2.147

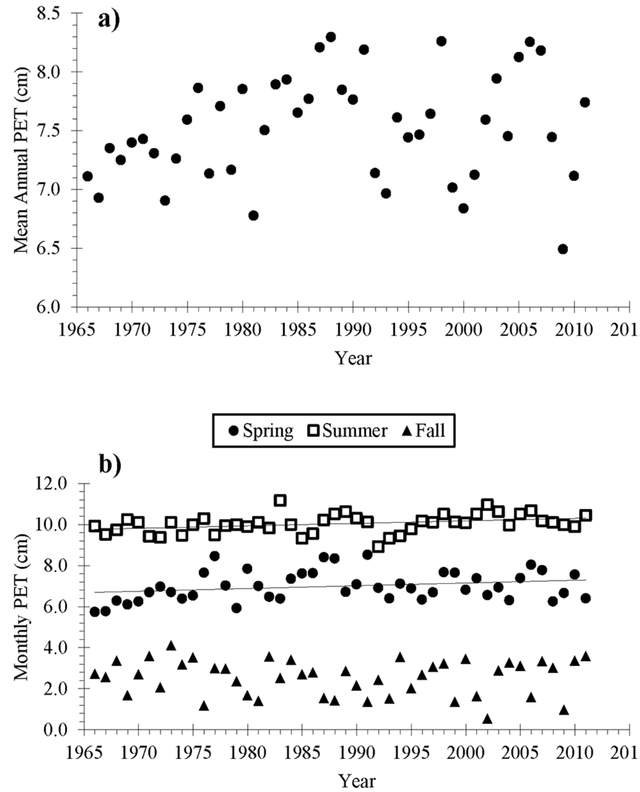

Fig. 4. Mean (a) annual and (b) seasonal potential evapotranspiration (PET). Annual PET has been increasing, albeit not statistically significantly, since 1966. Mean spring and summer PET have been increasing since 1966. Lines represent statistically significant linear regressions.

Seasonal Trends in Soil Moisture

Collected soil moisture data show a marked seasonal pattern in available soil water that is consistent across all three sampling sites (Fig. 5). The three monthly data points differentiate the spring dry-down period, when evapotranspiration begins and starts to deplete available soil water, as well as the fall recharge phase, when evapotranspiration slows down and precipitation increases the available soil water. High interannual variability among sites is present within all months of recorded available soil water (Fig. 5).

Critical Climate Period

At the MEF, precipitation minus PET (summed across a period of days before the available soil water measurements) was found to explain the variability in mean available soil water more than precipitation or PET alone (Table 4).

The relationship between the summed precipitation minus PET and available soil water was found to vary greatly among months, depths, and sites (Fig. 6). Summed precipitation minus PET was found to explain as much as 72% of the variation in available soil water in the upper soil layers and 56% of the variation in deeper soil layers. Summed precipitation minus PET had difficulties explaining the variation in May available soil water, particularly as depth in the soil profile increased. Overall, relationships between the summed precipitation minus PET and available soil water were strongest in September and November.

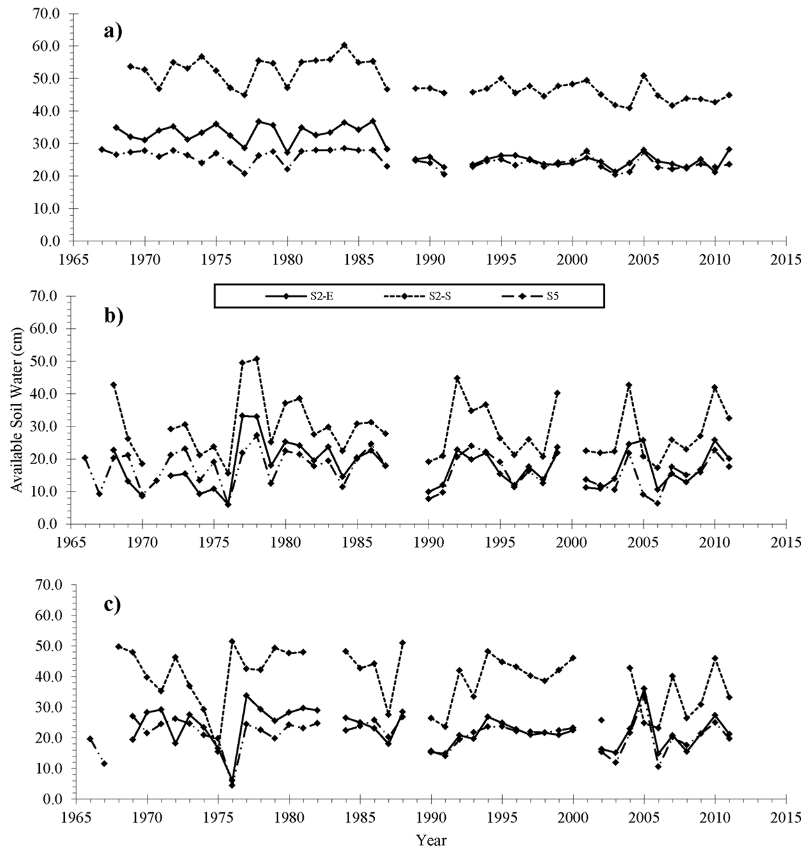

Fig. 5. Available soil water (0–229-cm depth) for three sampling sites during (a) May, (b) September, and (c) November at the Marcell Experimental Forest, 1966 to 2011.

Table 4. Optimized critical climate period (CCP) duration for each sampling depth and month of measurement. Optimized CCP maximizes the variation in available soil water that can be explained by a climate variable (precipitation, potential evapotranspiration [PET], and precipitation–PET) for an optimized number of days before the sampling date.

Depth	Month	Precipitation		PET		Precipitation–PET	
		Optimized CCP	R^2	Optimized CCP	R^2	Optimized CCP	R^2
cm		d		d		d	
	May	44	0.226	4	0.450	31	0.452
0–15	Sept.	22	0.265	145	0.368	22	0.723
	Nov.	84	0.257	44	0.331	84	0.416
	May	30	0.123	18	0.313	312	0.254
76–107	Sept.	117	0.428	51	0.338	77	0.692
	Nov.	109	0.385	107	0.160	109	0.721
	May	500	0.240	60	0.269	319	0.233
198–229	Sept.	116	0.313	660	0.306	77	0.557
	Nov.	147	0.260	295	0.275	109	0.552

DISCUSSION

Declining Soil Moisture

Despite no changes in annual or seasonal precipitation from 1966 to 2012, available soil water at the MEF has decreased significantly since 1966, specifically in May. In northern Minnesota, May soil moisture is a function of precipitation and the preceding snowmelt, as well as antecedent available soil water from the previous fall. While our data suggest that the total precipitation inputs to the system may be unchanging, the timing of these inputs is critical. The observed increase in winter and spring temperatures probably results in snow melting earlier in the season. As a result, snowmelt, runoff, and available soil water levels would peak in April as opposed to May. Additional data are needed to test this hypothesis. If seasonal snowpack is melting earlier in the season, evapotranspiration may be beginning earlier, leading to an additional depletion in May soil moisture. The day of the snowmelt centroid (the day when the accumulated stream flow exceeds 50% of the total stream flow) is currently 10 to 27 d earlier than it was in the 1960s (Sebestyen et al., 2011). Peak snowmelt at the MEF occurred in the first week of May in 1962 and has become increasingly earlier. In 2011, snowmelt occurred in the first and third weeks of April for Watersheds S5 and S2, respectively (Sebestyen et al., 2011). The growing season length, defined as the number of days between spring and fall frosts (Skaggs and Baker, 1985), has been increasing at the MEF. The MEF mean growing season length from 1961 to 1970 was 94 d, while the mean growing season length from 2000 to 2010 was 129 d (unpublished data).

Soil Moisture Patterns

The three annual measurements of soil moisture suggest that available soil water at the MEF follows anticipated patterns, with a spring draw-down and subsequent fall recharge of available soil water (Baker et al., 1979; Grayson et al., 1997; Tromp-van Meerveld and McDonnell, 2006). Additional measurements would be needed to determine if these patterns are consistent throughout the year. Annual soil water can vary greatly across months, years, and with depth. Before the growing season, soils are recharged via spring snowmelt (Sebestyen et al., 2011). Evapotranspiration and a lack of summer precipitation result in declining available soil water from May until August. Soils begin to recharge in the fall with the cessation of evapotranspiration. Available soil water usually falls below field capacity during the late summer months (average field capacity to a depth of 259.0 cm across the three sites is 31.9 cm).

Critical Climate Period

An analysis of the CCP of available soil water suggests that soil moisture is not simply a function of the weather conditions (i.e., precipitation, temperature, PET) on the day of soil moisture measurement but is instead a signature of precipitation minus PET from the 1 to 10 mo before the time of measurement. Precipitation minus PET takes into account precipitation, evapotranspiration, and temperature, all of which are important in determining soil water availability (Robertson et al., 1993). In

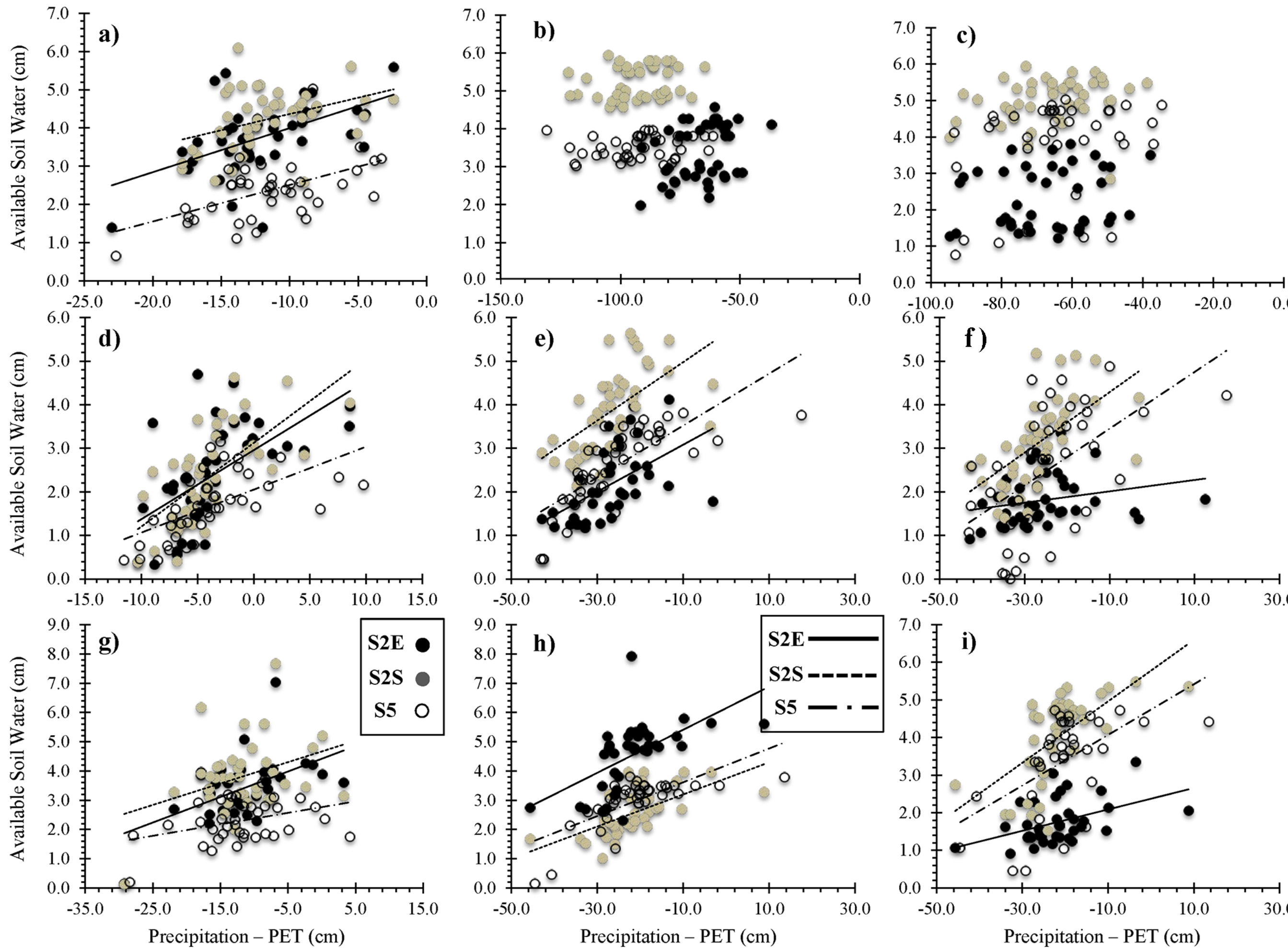

Fig. 6. Precipitation–potential evapotranspiration (PET) summed across the optimized critical climate period (CCP) duration vs. available soil water for three control watersheds at the Marcell Experimental Forest: (a) 0- to 15-cm sampling depth, May measurements, optimized CCP of 31 d; (b) 76- to 107-cm sampling depth, May measurements, optimized CCP of 312 d; (c) 198- to 229-cm sampling depth, May measurements, optimized CCP of 319 d; (d) 0- to 15-cm sampling depth, September measurements, optimized CCP of 22 d; (e) 76- to 107-cm sampling depth, September measurements, optimized CCP of 77 d; (f) 198- to 229-cm sampling depth, September measurements, optimized CCP of 77 d; (g) 0- to 15-cm sampling depth, November measurements, optimized CCP of 84 d; (e) 76- to 107-cm sampling depth, November measurements, optimized CCP of 109 d; and (f) 198- to 229-cm sampling depth, November measurements, optimized CCP of 109 d. Lines are linear relationships significant at $p < 0.01$.

general, the relationship between precipitation minus PET and available soil water in May is weaker than the relationships for September and November. This is probably due to (i) the influx of water to the soil profile via spring snowmelt, which was not included as a variable in precipitation minus PET, and (ii) the timing of May available soil water measurements. Most of the May measurements occurred before deciduous vegetation has leafed out and would be transpiring water to full capacity, yet the modeled PET that was included in precipitation minus PET did not take into account this short period of time during which PET is possible but trees have not yet started transpiring. In general, optimized CCP (the number of days before available soil water measurement in which the summed precipitation minus PET accounted for the highest variability in soil moisture) increased with depth. Water that enters the deeper soil horizons must first infiltrate through upper horizons, leading to time lags between precipitation events and deep soil moisture signatures.

Our results suggest that the timing of climate events and interannual variability has lingering impacts on available soil moisture. For instance, a spring drought may significantly influence the available soil moisture in the upper soil horizons for the current spring but may also impact late-summer available soil moisture in the middle soil horizons, as well as deep soil moisture in the fall. Droughts have effects lasting the entire year, even though precipitation may have rebounded from drought conditions. Analyses such as CCP are beneficial, but complex soil–atmosphere feedbacks not considered by the analysis may further enhance the variability in soil moisture (Eltahir, 1998; Schar et al., 1999; D'Odorico and Porporato, 2004; Koster et al., 2004).

Trends showing a decrease in available soil water at the MEF contrast with studies conducted in the boreal forests of Russia, which has a similar climate to the MEF. These studies have found increases in available soil water despite increases in temperature (Vinnikov et al., 1996; Robock et al., 2000). However, no comparative records of long-term available soil water are known to exist in boreal peatland ecosystems. Results from global soil moisture are similar to those found in Russia and suggest that increases in precipitation offset increases in evaporative demand (Robock et al., 2000). However, we have found that an increase in mean annual temperature does significantly affect soil moisture even when the total annual precipitation remains steady. This finding may have profound implications for available soil water, especially because global temperatures are expected to rise by 1.1 to 6.4°C by 2100 (National Research Council, 2010). Our results are especially important considering that most climate models project decreasing or static summer (June, July, and August) precipitation combined with increasing temperatures across the contiguous United States (Christensen et al., 2007). Such a scenario could result in drastic decreases in available soil water with time, leading to shifts in species, decreases in decomposition rates, and lower water tables.

CONCLUSIONS

We have presented a unique long-term record of seasonal soil moisture that spans 45 yr across three sites and reaches a depth of at least 2.3 m. Our results show that long-term available soil water in a northern forest has been decreasing at three sites despite no significant changes in precipitation with time and that increases in temperature and PET may be accounting for changes in available soil water. In a CCP analysis, precipitation minus PET accounted for greater variability in the available soil water than precipitation or PET alone. These results suggest that temperature and precipitation inputs combine to influence available soil water. Temperature at the MEF is predicted to continue to rise, which may contribute to even greater declines in available soil moisture.

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