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Climate change at the ecosystem scale: A 50-year record in New Hampshire

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Abstract

Observing the full range of climate change impacts at the local scale is difficult. Predicted rates of change are often small relative to interannual variability, and few locations have sufficiently comprehensive long-term records of environmental variables to enable researchers to observe the fine-scale patterns that may be important to understanding the influence of climate change on biological systems at the taxon, community, and ecosystem levels. We examined a 50-year meteorological and hydrological record from the Hubbard Brook Experimental Forest (HBEF) in New Hampshire, an intensively monitored Long-Term Ecological Research site. Temperature trends were most significant among examined climate metrics (ranging from 0.7 to 1.3°C increase over 40-50 year records at 4 temperature stations), while analysis of precipitation and hydrologic data yielded mixed results. Regional records show generally similar trends over the same time period, though longer-term (70-102 year) trends are less dramatic. Taken together, the results from HBEF and the larger regional records indicate that the climate has warmed detectably at HBEF over 50 years, with important consequences for hydrological processes. Understanding its effects on ecosystems will require a diversity of metrics and concurrent ecological observations at a range of sites, as well as a recognition that ecosystems have existed in a directionally changing climate for decades, and are not necessarily in equilibrium with the current climate.

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1 Introduction

To better understand the effects of past and future climate change on ecosystems, we must first examine trends in changing climatic variables and their effects on physical and biological processes critical to ecosystem functioning. Climate models cannot resolve climate-ecosystem interactions at all biologically or ecologically relevant scales, given the complexity of real-world landscapes and their dependence on spatially variable histories of disturbance. Some predictions made at the regional scale may not represent trajectories at specific well-studied locations, (conversely, no single location is likely to adequately represent the region). Therefore, existing empirical data are of critical importance, as they allow us to examine interactions among diverse climate variables in real-world ecosystems and thus better understand the effects of climate change on specific taxa, communities, and ecosystem processes, including inputs and outputs of water, nutrients, and carbon.

New England has experienced significant climatic change over the 20th Century, and given its long history of climate observations provides a good study system for understanding the relationship between local and regional trends. For example, mean annual temperature in New England and New York increased about 1.1°C over the 20th Century (Trombulak and Wolfson 2004). In New Hampshire, warming appears to have been greater in the southern part of the state than in the north (Keim *et al.* 2003; Trombulak and Wolfson 2004), and total annual precipitation has increased by approximately 10mm per decade over the 20th Century in the northeast (Hayhoe *et al.* 2007). Winters have warmed more than summers, consistent with the globally observed pattern (Lugina *et al.* 2004), and the fraction of precipitation falling as snow has decreased (Huntington *et al.* 2004). Biologically relevant seasonal transitions are also changing, from earlier snowmelt (Hodgkins and Dudley 2006) to a longer frost-free season (Easterling 2002). Leaf-out is occurring earlier throughout the northern hemisphere (Schwartz *et al.* 2006), as is flowering across many taxa in the northeast (Primack *et al.* 2004; Houle 2007), and growing season is lengthening significantly (Kunkel *et al.* 2004, Wolfe *et al.*, 2005; Richardson *et al.*, 2006). Similarly, long-term records show trends toward earlier ice-out dates on lakes and rivers across the northeast (Hodgkins *et al.* 2002; Huntington *et al.* 2003), and

reduced duration of snowcover (Burakowski *et al.* 2008).

Over the 21st Century, models predict a dramatic warming trend in the northeastern United States, with modest increases in total precipitation. Christensen *et al.* (2007) averaged 21 models to predict approximately 3.5-4°C of warming and a 5-10% increase in precipitation in New England over the 21st century, with greater changes in the winter than in the summer. Easterling *et al.* (2000) and Schär *et al.* (2004) report evidence of increases in the frequency of extreme climate events globally and climate modeling indicates a likely increase in the frequency and intensity of extreme events at the extremes of previously established frequency distributions (including drought) in the northeastern U.S. (Wehner, 2004, Tebaldi *et al.*, 2006; Hayhoe *et al.*, 2007).

1.1 Question and Hypotheses

Here, we test whether the 50-year climate record at Hubbard Brook Experimental Forest matches modeled changes for the 21st century - a warmer, wetter, more variable climate. We also hypothesize that trends in directly measured climate data (air temperature, precipitation) will be reflected by trends in ecologically relevant derived variables (e.g. degree day metrics) and measurements of variables that are driven by both temperature and precipitation (e.g. evapotranspiration, soil frost). If the existing 50-year record of climatic change qualitatively matches expected future change, there may be important lessons to be learned from the changes observed in the existing long-term ecosystem process datasets of the Hubbard Brook Ecosystem Study. These expected trends include:

- Increased mean annual temperature
- Increased mean seasonal temperatures (summer and winter)
- Increased maximum and minimum annual temperatures
- Increased annual total growing degree-days
- Increased winter thawing degree-days
- Increased length of the frost-free growing season
- Increased total annual precipitation
- Increased length of periods without measurable precipitation

- Increased frequency of days with intense precipitation
- Increased frequency of extreme high and low streamflow
- Increased total annual evapotranspiration
- Earlier spring thaw streamflow conditions
- Decreased duration of snowpack
- Decreased maximum snow pack water content
- Increased soil frost (as a consequence of reduced snowpack)

While a variety of studies, including those cited above, have examined many of these variables in distributed regional and national data sets, we take a different approach, examining a comprehensive hydrometeorologic data set from a single study site, which may provide insight into interactions among processes that averaged data from regional networks would obscure. We ask whether the hypothesized and regionally-observed changes are detectable in the 50-year hydrological and meteorological records from one intensively monitored forest ecosystem, and compare the trends observed at the local scale to other local stations with similar data sets to determine whether trends are representative regionally. While problems with long-term changes in some climate data networks have been resolved (e.g. the Historic Climatology Network maintained by NOAA; see Keim *et al.* 2003), comparison with highly complete records from a small number of stable sites in the region with a documented lack of changes in land-cover, instrumentation, and methodology may provide additional confidence in observed trends. The long-term maintenance of high-quality hydrological and meteorological records across a single, large and intensively studied ecological research site provides a unique opportunity to investigate the influence of concurrent changes in multiple dimensions of climate on ecosystem processes.

1.2 Site Description

The Hubbard Brook Experimental Forest (HBEF) was established in 1955 for the study of forest hydrology. The 3160 ha forest (43°56'N, 71°45'W) is located in the White Mountain region of New Hampshire (Fig. 1a). Native peoples never lived in the valley, so selective cutting circa 1900 represents the only

major historic disturbance. Today the forest is dominated by American beech (*Fagus grandifolia* Ehrh.), sugar maple (*Acer saccharum* Marsh.), and yellow birch (*Betula alleghaniensis* Britt.), transitioning to balsam fir (*Abies balsamea* L.) and red spruce (*Picea rubens* Sarg.) at the ridge tops. Eastern hemlock (*Tsuga canadensis* L.) is important at the lowest elevations (Schwarz *et al.* 2001). Existing research on forest processes (Likens and Bormann 1995) and high-quality long-term climate records (Bailey *et al.* 2003) make HBEF an ideal place to study the effects of a continuously changing climate on ecosystem structure and function in the northern hardwood forest. Recent work, for example, has focused on snowpack and soil frost (Groffman *et al.* 2001; Campbell *et al.* 2010), the response of canopy structure and composition to a severe ice storm (Rhoads *et al.* 2002; Weeks *et al.* 2009), climatic drivers of variation in tree phenology (Richardson *et al.* 2006) and winter injury in red spruce (Hawley *et al.* 2006).

2 Data and Methods

Meteorological data are collected continuously at the HBEF using standard methods and simple mechanical instruments with proven reliability in harsh environmental conditions. All instruments are visited weekly ensuring that the instruments are well maintained and that problems, when they do occur, are short-lived. The data-collection network currently includes 24 precipitation collectors (those used in this study are mapped in Fig. 1a). Hygrothermographs housed in standard shelters (Fig. 1b) are co-located with seven of the precipitation collectors (Bailey *et al.* 2003). A realtime network with electronic probes and radio transmission has recently been established, to eventually replace historic measurement techniques. The network is currently in a period of calibration and observation with co-located instruments for data quality assurance. Data used here are from original instruments that have undergone consistent and routine calibration and data processing procedures, and span from the start of each record in the 1950's or 1960's through 2005. The headquarters site (G22) has a more dynamic recent land-use history (building expansion, road paving, and parking lot expansion) than the other studied locations, which may affect the integrity of the record. There are nine gauged watersheds; continuous stream-height measurements are made in a stilling well attached to a V-notch weir at the

bottom of each watershed; each weir has been calibrated to measured flow rates (Bailey *et al.* 2003). As a general rule, we selected the longest available records that were not affected by experimental manipulation (e.g. forest cutting) for trend analysis.

2.1 Temperature

We analyzed temperature records from the four longest-running meteorological stations at HBEF (Table 1, Fig. 1a). For each station, we calculated the annual mean temperature, based on daily means (reported as the mean of the daily minimum and daily maximum temperatures). Annual means were calculated on a calendar-year basis. Seasonal means were calculated on a calendar-month basis (DJF, MAM, JJA, SON), so that the mean winter value of one year includes December of the previous year. Annual minimum and maximum temperatures were also determined.

We also derived secondary temperature-based climate metrics, including the annual total number of growing degree days (GDD; the annual sum of the difference between each day's mean temperature and a base temperature of 4°C; Richardson *et al.* 2006) in each calendar year, thawing degree-days (TDD; from a base temperature of 0°C) for the periods December-March and January-February, as well as the dates of the first and last frost each year.

2.2 Hydrological Cycle

We examined precipitation data from W3, a southwest-facing watershed used as the hydrologic reference at HBEF, and W7, which has the longest record among the north-facing watersheds (Table 1). Precipitation for each watershed is calculated as a Thiessen-polygon weighted mean of precipitation collected at rain gauges within and near each watershed (Fig. 1a; Bailey *et al.* 2003). We summed precipitation on a calendar year basis. Though we did not have a specific directional-change hypothesis to test (Hayhoe *et al.* 2007), we also tabulated summer precipitation (June through August), the period when evapotranspirative flux is greatest, and major ecosystem processes (e.g., photosynthesis, soil respiration) are most likely to be water-limited. We also analyzed the shorter precipitation record at G22, which receives less precipitation than the experimental watersheds due to its lower elevation and topographic position.

The daily precipitation record allowed us to analyze the timing of precipitation in addition to the annual

total. For each calendar year, we tabulated the number of days with ≥ 50 mm in total precipitation, and determined the length of the longest period in each year with no recorded daily rainfall greater than 1 mm.

Total annual streamflow data from W3 were analyzed on a calendar year basis between 1959 and 2005, and from W7 between 1966 and 2005. For each calendar year, we calculated the number of days with ≥ 50 mm in total streamflow (the convention at HBEF is to express streamflow volume divided by the area of the contributing watershed, so that it can be directly compared to precipitation data in mm), and length of the longest period during which daily streamflow did not exceed 0.1 mm day⁻¹.

We analyzed the timing of spring melt-influenced streamflow conditions following the center-of-volume date methodology used by Hodgkins *et al.* (2003). For each year, we calculated the total January-May streamflow, and identified the date at which half the total volume had passed the weir.

Having both precipitation (water input) and streamflow (water output) for the same watershed allows us to estimate evapotranspiration, an output which can otherwise be measured only using eddy covariance methods (which are unsuitable for the steep topography at HBEF). This method assumes that ecosystem water storage (soil water and snowpack) is the same at the beginning and end of each measured period. High interannual variation in snowpack water content and persistence is the reason a June 1 water year is traditionally used at HBEF. Over many years any errors in storage must approximately cancel, making this method appropriate for examining long-term trends. We calculated evapotranspiration at W3 and W7 for water years beginning June 1.

2.3 Snowpack and Soil Frost

Snow course transects are located near several rain gauges in the watersheds, with the longest continuous record at G2 (Fig. 1; Table 1). Snow is sampled for depth and water content with a Mount Rose snow tube each week during the winter. Snow surveys begin as soon as there is 6" (~15 cm) of snow that is expected to remain all winter, and continue as long as there are patches of snow to measure. Each week, 10 measurements of the snow pack are made every two meters along a transect, and are averaged. An entry of 0 indicates no snow at any of the 10 sampling points on the transect; when snow

is patchy, zero values are included in the average. We analyzed the maximum observed snow water content recorded each winter, as well as the period of snow cover. Data on the presence or absence of soil frost are also taken each time snow is measured; we analyzed the maximum soil frost occurrence observed each winter.

2.4 Regional and Global Data

To determine how consistent HBEF data were with records taken elsewhere in the region, and to provide a longer-term context for interpreting trends, we examined long-term temperature records from other weather stations in the White Mountain region (Fig. 1a; Table 1). We included temperature data from three sites: Mt. Washington (47 km NE of HBEF), the highest peak in the northeastern United States; Pinkham Notch (50 km NE of HBEF), a mid-elevation location; and Hanover, (50 km SW of HBEF), which has an exceptionally long and complete temperature record. We calculated annual and seasonal mean temperatures for each station using the same methods we applied to the HBEF data. However, as these datasets were less complete than those from HBEF (Table 1), we excluded any year with >10 days missing data. In years with ≤ 10 days missing, missing maximum and minimum daily temperatures were linearly interpolated before calculating daily mean temperatures.

Additionally, we examined temperature records in the context of two broader data sets. As an indicator of regionally integrated trends, we extracted annual and seasonal mean data from 1901-2002 for the grid cell centered on 43.75°N, 72.75°W from a global dataset (Mitchell and Jones 2005). These values were calculated using a weighted mean of all stations located within 1200 km of the pixel. We also examined the global annual and seasonal mean temperature record for the 30-60°N latitude band (Lugina *et al.* 2004). To facilitate comparison, all temperature data were normalized to represent deviations from the 1991-1996 mean, the longest recent period for which all records were complete.

The analysis of spring streamflow center-of-volume date was performed on 101 years of gage data (1904-2004) from the Pemigewasset River at Plymouth (USGS-01076500, 43°46'N 71°41'W), which drains a 1610 km² watershed, including the HBEF (Table 1).

2.5 Statistics

We used a non-parametric Mann-Kendall test to detect trends (Helsel and Hirsch 1992). This test does not require normally distributed data, but does assume (as does ordinary least-squares regression) there is no significant autocorrelation in the time series. We confirmed this with a Durbin-Watson test, which indicated that model residuals were not significantly autocorrelated for any time series (though any long-term persistence is not accounted for, resulting in potentially smaller-than-justified p-values; Cohn and Lins 2005). The rate of change in each variable was determined with the *Sen slope*, a non-parametric estimate of the rate of change in a variable over time. It is attractive for this kind of analysis because it requires fewer assumptions than, for example, least-squares regression, and is relatively insensitive to outliers. The *Sen slope* is the median slope among all pairs of points in the dataset (Gilbert 1987). A custom SAS program (Winkler 2004) was used for the Mann-Kendall and *Sen slope* analyses. All analyses were done both for the entire length of the record, and also for the period 1966-2005, which allows direct comparison among all HBEF records, some of which began as late as 1966. Trends for both periods are expressed per decade. Where trends were small in magnitude and there were a large number of data ties among years (e.g. for metrics that were counts of days with given conditions), confidence intervals were often not meaningful and are not reported. Due to a large number of years with no observed soil frost, the soil frost record was analyzed using linear least-squares regression, despite its acknowledged limitations. Trends with $p < 0.10$ were considered significant, though trends with $p < 0.05$ were tallied separately.

Temperature at G1 and streamflow for W3 were analyzed using a frequency distribution approach. These are the longest-term records of their respective types that are not subject to experimental manipulation (forest cutting or fertilization). Temperature data were classed into 5°C bins and the daily streamflow (in mm/day) was classed into 20 bins on a logarithmic scale. We compared the frequency distribution of the 20-year period at the beginning of the record (1958-1977) to that of the 20-year period at the end of the record (1986-2005). Differences between the two time periods are expressed as percent change in frequency in each class.

3 Results

3.1 Temperature

The four HBEF temperature records examined are highly correlated in mean annual temperature (Fig. 2a). The offset among these sites is largely a factor of elevation; the lapse rate calculated using G1 and G6 is $-5.2^{\circ}\text{C}/1000\text{ m}$ and shows no detectable trend over time. Three of the four stations showed significant increases in mean annual temperature over the entire available record; trends ranged from $0.13 - 0.32^{\circ}\text{C}/\text{decade}$ (Fig. 2a; Table 2). When only the 1966-2005 record was examined, three of four were still significant, and the mean trend was $0.21 \pm 0.08^{\circ}\text{C}/\text{decade}$ (95% CI). Over both time periods, G14 had the strongest trend, while G1 and G6 had the weakest. Over the last 40 years of the record, trends observed at HBEF are similar to those seen elsewhere in the region and globally (Fig. 2). Pearson's correlation coefficients (r) between G1 and regional records range from 0.83 (Pinkham Notch) to 0.88 (Mt. Washington), while correlation with the global record ($r=0.52$) is poorer. Observed changes in mean annual temperature at HBEF ($0.13-0.32$ per decade in the 40-50 years leading up to 2005) are decidedly greater than the longer-term regional data we examined (Table 2), and are also greater than the $\sim 0.04-0.1^{\circ}\text{C}/\text{decade}$ trends observed in NH from 1931-2000 (Keim *et al.* 2003) and the $\sim 0.12^{\circ}\text{C}/\text{decade}$ trend observed over the full 20th century (Trombulak and Wolfson 2004) using statewide networks of stations.

Examined individually, records of mean seasonal temperatures often look very different from mean annual temperatures (Table 2). At HBEF, mean winter temperatures (G1 stdev for 1996-2005= 1.80°C) are more variable than are summer temperatures (G1 stdev for 1996-2005= 0.74°C). Trends for mean winter and mean summer temperatures are each significant at three of the four stations examined (Table 2), and rates of winter warming over the past 40 years ($0.31-0.45^{\circ}\text{C}/\text{decade}$) are non-significantly greater than rates of summer warming ($0.02-0.33^{\circ}\text{C}/\text{decade}$) at all sites except G22 at the headquarters building. This difference is consistent in pattern with the findings of Burakowski *et al.* (2008) who documented a similar rate of winter warming ($0.43^{\circ}\text{C}/\text{decade}$) in the northeast over a similar time period (1965-2005). Hayhoe *et al.* (2007) also observed greater warming in winter than summer in

New England from 1970-2000, though the increases were smaller in magnitude.

Of the four temperature records at HBEF, only G22 showed a significant increase in maximum annual temperature ($0.52^{\circ}\text{C}/\text{decade}$ between 1957 and 2005, $p<0.01$). This could be the result of increasing the extent of paving in the immediate surrounding area. Minimum annual temperature only increased significantly at G14 ($1.24^{\circ}\text{C}/\text{decade}$ between 1965 and 2005, $p=0.03$). Overall temperature extremes are a stochastic metric, and it is not surprising that few statistically significant trends were observed.

Annual growing degree days (GDD) increased significantly in three of four HBEF temperature records examined. The mean slope was 32.6 GDD/decade between 1966 and 2005. Thawing degree days (TDD) have increased significantly at G1 both in January-February and December-March (Fig. 3). When the 1956-2005 record is examined, the slope is $+2.9$ TDD/decade for January and February alone, and 10.5 TDD/decade for December-March. The length of the frost-free growing season has increased significantly at G1, by 3.75 days/decade between 1956 and 2005 ($p=0.03$). This result is driven both by a later date of the first autumn frost (1.9 days/decade, $p=0.04$) as well as a non-significantly earlier date for the last spring frost (1.6 days/decade, $p=0.15$).

3.2 Precipitation and Hydrology

No significant trend was observed for total annual precipitation at either W3 or W7, though the nonsignificant trends were both positive (note that a prolonged drought occurred in the early 1960s, but was over by the start of the W7 record in 1966). Summer precipitation showed a significant trend (17.1 mm/yr/decade) in W3 between 1958 and 2005, but the trend was not significant between 1966 and 2005, or at W7. There were no significant changes in the length of each year's longest precipitation-free period, and the frequency of days with >50 mm of precipitation increased significantly at W7 but not at W3.

The frequency of days with low streamflow (<0.1 mm daily) at W3 decreased significantly between 1958 and 2005 (-8.7 days/yr/decade, $p<0.01$). A very small but significant increase in the frequency of days with intense precipitation (>50 mm daily) occurred at W7. Evapotranspiration decreased significantly at W3 (-10.9 mm/yr/decade, $p=0.01$) but not at W7.

Spring streamflow center-of-volume date has become significantly earlier at W3 (2.1 days earlier per decade over 1958-2005) and at W7 (3.1 days earlier per decade over 1966-2005), though the trend was not significant in the 1966-2005 record from W3 (Table 1; Fig. 4). The mean spring center-of-volume date from these two first-order watersheds correlates remarkably well with those from the Pemigewasset River at Plymouth ($r=0.94$), despite the fact that the watersheds differ in area by a factor of more than three orders of magnitude and in elevation range by a factor of >5 (Table 1). Spring center-of-volume date on the Pemigewasset has not changed significantly over the entire 101-year records, but its trend since 1966 is similar to those at HBEF, (1.9 days/decade earlier).

3.3 Snowpack and Soil Frost

The observed trend in date of first measurable snowpack each season, 1.7 days later /decade was not significant, while the date of last measurable snowpack in spring has become significantly earlier at a rate of 2.5 days/decade, for a significant net reduction by 4.2 days/decade in the snowpack duration ($p = 0.06$). Regionally, Burakowski *et al.* (2008) found a similar reduction in snowpack duration (3.6 days/decade), as well as significantly reduced total snowfall, which we did not examine. Maximum snowpack water content at G2, the only long-term record of its kind at HBEF, decreased significantly between winters 1956 and 2006, with a slope of -10.5mm/decade (about -5%/decade; $p=0.08$).

Soil frost coverage is expected to be negatively correlated with snowpack depth and water content, but is highly variable both spatially and interannually. No soil frost was observed at G2 between 1965 and 1969, though the sampling intensity during these years is not well documented (we did not include these years in the trend analysis). While widespread frost was not observed at HBEF before 1970, it is not uncommon now (Likens and Bormann 1995; Campbell *et al.* 2010).

3.4 Frequency Analyses – Temperature and Runoff

Frequency distribution of daily mean temperature at G1 (Fig. 5a) and daily total streamflow at W3 (Fig. 5b) shifted observably over the record. In particular, days with a mean temperature of $-25\pm 2.5^{\circ}\text{C}$ were 39% less frequent from 1986–2005 compared to 1958–1977, while days of $25\pm 2.5^{\circ}\text{C}$ were 5% more frequent. The temperature with the greatest

increase was $5\pm 2.5^{\circ}\text{C}$, which was 10% more frequent. This temperature occurs most commonly in early spring and mid-autumn, times when organisms are transitioning physiologically and are thus particularly sensitive to swings above and below freezing. The pattern of streamflow at W3 also changed, with low rates becoming less common and high rates becoming more common although it is important to note that this pattern is likely to have been influenced by a severe drought that affected the northeast in 1963-5.

3.5 Pattern of change

Out of 18 metrics examined over the past 40-50 years, 13 have changed significantly ($p<0.10$) in the hypothesized direction in at least one location at HBEF (of which 11 had $p<0.05$), while two (evapotranspiration and frequency of low streamflow) have changed significantly in the opposite direction (Table 2). These metrics are not statistically independent (we did not include our one observationally non-independent metric, evapotranspiration, in the count), but demonstrate that a wide variety of climate metrics yield similar qualitative results. Observed trends depend somewhat on the length of record analyzed: measured temperature and temperature-derived trends have larger slopes and greater significance when examined for the period 1966-2005 than for longer time periods. While this result matches with the general observation that global temperatures began rising rapidly in the 1970's (Folland *et al.* 2001), it is also true that the 1966 start date we selected to compare the maximum length of record across all sites examined (Table 1), anchors the beginning of our time series in an anomalously cold period for the region (Fig 2b). On the other hand, the late-1950's were considerably warmer than either the preceding 50 years or the following 30 years globally and at other regional stations (Fig 2b), and only around 1980 does the start of a clear rising trend become obvious in the longer-term data, which provide important context for interpreting the trends observed in the 40-50 year records examined at HBEF.

4 Discussion

4.1 Climate shifts

Changes in frequency distributions of temperature and streamflow (Fig. 5) show that the width of the probability density functions for these two variables have remained approximately constant over the past

50 years at HBEF, while the means of those distributions have shifted. Put simply, the climate as a whole is becoming warmer, and streamflow is increasing, but the (admittedly variable) extremes do not appear to be changing faster than the means. This finding, while similar to those reported by Frich *et al.* (2002), who examined 10 climatic parameters at the global scale over the past 50 years, contrasts somewhat with the more mixed findings of larger-scale analyses of the frequency of extreme events in some climate metrics (Schär *et al.* 2004; Tiebaldi *et al.* 2006; Alexander *et al.* 2006), though differences among the metrics examined make such comparisons difficult.

4.2 Changes in the timing of seasonal transitions

Greater warming in winter and spring than summer has been widely observed in the northern hemisphere (Liu *et al.* 2007) and can lead to large changes in the timing of ecologically relevant seasonal transitions. For example, streamflow conditions indicative of spring snowmelt advanced 8-12 days over the ~50-year record, similar to the 1-2 week advance observed regionally by Hodgkins *et al.* (2003). On the other hand, the 18-day advance in ice-out on Mirror Lake (adjacent to HBEF) since the late 1960's (Likens 2009) appears quite dramatic in comparison to the longer record presented by Hodgkins *et al.* (2002). Along with earlier snowpack melting at G2 (12 days earlier since 1955), these results point to a significantly earlier transition to spring thaw conditions characterized by the availability of liquid water and soil temperatures permitting root activity. The end of the snowpack likely represents a major "tipping point" in the seasonal transition, when surface soil temperature is released from control by the thermal insulation and high albedo of existing snowpack, as well as latent heat loss of the melting snowpack. Following this transition, soils temperatures generally warm rapidly, but are also subject to more dramatic fluctuation, with potential consequences for physical and physiological disturbance to shallow roots and ground-level vegetation.

Similar changes appear to be occurring in the fall as well. Easterling (2002) found that since 1948 the first frost has occurred about 2-3 days later, while the date of last frost has occurred about 5 days earlier throughout the northeast. The HBEF data from G1 for the period 1956-2005 indicate a much larger shift, with first frost occurring 10 days later ($p=0.05$).

Changes in seasonal timing can affect the growing season utilized by plants. Global data suggest that the onset of the northern hemisphere spring advanced approximately 1 week from the 1970s to 1990s (Keeling *et al.* 1996), and similar trends are found in studies of budburst and flowering at multiple scales (8 days earlier over 100 years in Boston, Primack *et al.* 2004; 5 to 6 days earlier over 35 years across the US, Schwartz and Reiter 2000). Tree phenology data are collected at HBEF and correlate well with spring temperatures (Richardson *et al.* 2006). While the 16-year record is insufficient for time-series analysis, they allowed the calibration of a model that estimated a slow trend towards earlier spring (0.2 days advance per decade) for the period 1957-2004 (Richardson *et al.* 2006).

4.3 Changes in precipitation and streamflow

While there are some broad trends in the hydrologic record over the last 50 years at HBEF, few are statistically significant (Table 2). This is probably due both to interannual and decadal-scale variation which are large relative to the magnitude of any long-term trend. For example, annual precipitation at W3 ranges from 979-1793 mm, and the observed increase (~34 mm annual total per decade) was non-significant, but at least generally consistent with the range of trends observed by Lins and Slack (1999), who showed significant increases in streamflow across most of the continental US on streams with multi-decadal records. Streamflow is highly correlated with precipitation, ($r=0.96$ at W3 on a water-year basis).

At HBEF, trends in evapotranspiration were mixed, with a significant decrease at W3 over the full record, and a non-significant increase at W7 since 1966; trends at W3 and W7 did not differ significantly during this period. This result is also unexpected, as warmer and longer growing seasons would be expected to increase, rather than decrease, total annual ET (Huntington *et al.* 2009). However, a complicating factor is that the forest canopy at HBEF has increased in structural complexity (i.e. spatial variation in maximum height) since the 1950's, due to the aging of the forest, as well as a number of disturbance events (e.g. the 1998 ice storm). Alternatively, the observed decrease in ET may be related to trends in unexamined climatic factors, such as cloudiness, relative humidity, and windiness, for which the records at HBEF are not complete enough to justify trend analysis. Changes in stomatal density and

regulation related to forest age and structure, atmospheric CO₂ concentrations, nutrient availability, and acid deposition in northern hardwood species are other potential explanations for changes in ET, and are the focus of much current research at HBEF and throughout the region.

Lins and Slack (2005) and McCabe and Wolock (2002) observed increases in the lower end of the flow-volume distribution at both the national and regional scales, which they interpret as due to increased precipitation in the warm (low-flow) months, while high flows have generally not changed significantly. At W3, we observed an unexpected decrease in the frequency of days with low streamflow (<0.1 mm), precisely the transient conditions in which evapotranspiration might be expected to be water-limited rather than energy-limited, *sensu* Budyko (1974). We did not observe a significant change in the maximum duration of precipitation-free periods which might be expected to contribute such temporary water limitation. These trends (presumed increases in soil water availability as reflected by reduced low-streamflow events, along with decreases in apparent annual evapotranspiration) appear to run counter to each other, unless the factor driving the change in the system is a decrease in evapotranspirative demand. Another un-investigated pathway that may help explain the change in evapotranspiration in light of observed winter warming is snowpack sublimation. It is important to note, however, that none of these trends are significant in the 40-year record (1966-2005), only in the 50-year record, which includes the unusually dry period of the early 1960's.

4.4 Vegetation community implications

Species distributions worldwide have been observed to respond to 20th century climate change (Parmesan and Yohe 2003). Iverson and Prasad (2002) show the potential for dramatic climate-driven shifts in forest species composition in New England by the year 2100, though range expansions are unlikely to keep up with the expected rate of climate change, and species shifts may be delayed, abrupt, and asynchronous across the landscape (Mohan *et al.* 2009). The 1°C of warming in the 50-year temperature record at HBEF is equivalent to a drop in elevation of 150-200 m, roughly the elevation range covered by the south-facing experimental watersheds, across which forest composition ranges from northern hardwoods at lower elevations to boreal spruce-fir at higher elevations. Presettlement

forest records (Vadeboncoeur *et al.* 2012) suggest that the hardwood/boreal ecotone has moved ~60 m upslope in ~200 years, due in part to the decline of red spruce, which has been attributed climate change (Hamburg and Cogbill 1988; Beckage *et al.* 2008). Further upslope encroachment of hardwoods seems likely; on low ranges such as those at Hubbard Brook, spruce-fir forests may be extirpated completely as has been previously suggested (Iverson and Prasad, 2002). Lund and Livingston (1999) suggest that spruce damage is brought on by cold temperatures following midwinter thaws that result in the foliage dehardening, rendering it susceptible to frost damage. Hawley *et al.* (2006) found significant damage in 2003 on W6 at HBEF, in a year when the temperature reached -26°C at G6 only 8 days after a significant thaw. Spruce cold tolerance is also decreased in areas of high N-deposition and low soil Ca (Hawley *et al.* 2006). The combined effect of increased winter thawing (Fig. 3), and a decrease in the frequency of extreme cold temperatures (Fig. 5a) on red spruce in the future remains unclear, and may depend on future trends in acid deposition.

Other tree species may also be affected by changing patterns of winter temperatures, particularly with regard to specific thresholds and extremes. Changes in snowpack depth and duration, and associated changes in soil frost, seem particularly likely to affect young size classes of trees and evergreen forest herbs. Borque *et al.* (2005) suggest that yellow birch suffers winter damage when the minimum temperature falls below -4°C after experiencing 50 growing degree days in the spring. At G1, at an elevation where yellow birch is abundant at HBEF, 1981 was the year with the greatest potential for yellow birch damage according to these criteria. In fact 1981 saw a major regional dieback of birches and sugar maple (Auclair 2005). However, systemic dieback of yellow birch has not been reported at HBEF, and no trend in the frequency of such events since 1956 is evident in the data from G1. To the extent that spring phenology is determined by accumulated degree days rather than by photoperiod, an earlier and more variable spring might lead to more frequent spring-freeze defoliation events (e.g. Gu *et al.* 2008); indeed a hard frost in May 2010 following budburst occurred at Hubbard Brook and elsewhere in the northeast, with important effects on carbon balance in several canopy species (Hufkens *et al.* 2011). Whether such events become more frequent is dependent on the

rate at which frost-sensitive phenological stages advance relative to the advance in the latest occurrence of damaging cold each spring (Scheifinger *et al.* 2003); similar interactions may occur with soil freeze-thaw events and root activity. Continued increases in atmospheric CO₂ concentration may also mitigate stresses caused by extreme or unusually-timed extreme temperatures (Wayne *et al.* 1998).

Insects as well as trees may respond to changes in winter climate. The hemlock wooly adelgid, an introduced insect, has decimated hemlock forests in southern New England, but has not yet spread to northern New England. Temperatures of -25°C are >98% lethal to hemlock wooly adelgids (Skinner *et al.* 2003), yet at G22, at the elevation where hemlock is most important at HBEF, wintertime temperature minima have dropped below -25°C in only four of the last 10 years, compared with eight of the first ten years of the record. If such winter warming continues, the adelgid can be expected to expand its range northward toward HBEF in years with warm winters (Evans and Gregoire 2007).

4.5 Biogeochemical implications

The effect of a changing climate on ecosystem-level C cycling is likely to be mixed. Campbell *et al.* (2009) modeled increased primary production due to a longer growing season. However, if severe storms become more frequent (Easterling *et al.* 2000), or if winter dieback increases or new pathogens invade, there could be a net loss of standing live biomass and a resulting increase in litterfall, which would be rapidly respired particularly under a warmer, wetter climate (Rustad *et al.* 2001; White and Nemani 2003). Reduced snowpack may or may not increase soil frost in light of warmer winter temperatures (Campbell *et al.* 2010), perhaps affecting fine root turnover (Groffman *et al.* 2001). Soil frost and temperature-driven increases in N mineralization and nitrification may increase nitrate leaching (Groffman *et al.* 2001; Campbell *et al.* 2009), transferring N from terrestrial to aquatic ecosystems, though the long-term trajectory of anthropogenic N loading on these ecosystems remains uncertain and will interact with the effects of a changing climate on ecosystem processes.

Hollinger *et al.* (2004) showed that warm spring and autumn temperatures correlated with increased production, while hot summers correlated with decreased production, so the end result of disproportionate winter and transition-season

warming might be an increase in ecosystem C storage. Warm springs that lead to early budburst may lead to both immediate and time-lagged effects on ecosystem C exchange, and increases in production may be greater for deciduous species (Richardson *et al.* 2009). Huntington (2005) proposed that the Ca status of forested sites in Maine may decline with increased production over a longer growing season, increased base cation leaching from greater rainfall, as well as shifts from spruce and fir to more Ca-demanding hardwoods. Indeed, over the longer term, the indirect effects of climate changes on biogeochemical cycles via species shifts (e.g. spruce and hemlock giving way to deciduous species) will be important to consider.

4.6 Conclusions

Ecologically meaningful changes in the climate at HBEF have occurred over the past 50 years, and predicted changes are likely to further alter the forest's community ecology and biogeochemistry. A complete understanding of these changes requires long-term studies at the ecosystem scale where the multiple interacting climatic variables are monitored alongside ecological observations and other changes (e.g. CO₂ and O₃ concentrations, atmospheric inputs of acidity and nutrients). Inherent in a great deal of ecological research is the assumption that the system is in equilibrium. However, we show here that significant changes in multiple climatic variables have occurred over the 50-year history of ecological research at HBEF. The non-equilibrium status of the current forest ecosystem with respect to climate and other perturbations is an important consideration when examining responses to disturbance, or parameterizing models describing future ecosystem function.

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Table 1. Characteristics of the meteorological stations at HBEF and the surrounding region.

<i>Data from HBEF</i>				
TEMPERATURE				
Station	Elev. (m)	Aspect	Record	Completeness
G1	490	SSE	1956-2005	>99.99%
G6	750	SSE	1961-2005	99.96%
G14	730	N	1965-2005	99.83%
G22	250	--	1957-2005	>99.99%
SNOW DEPTH AND SOIL FROST				
Station	Elev. (m)	Aspect	Record	
G2	560	SSE	winters 1955-6 thru 2005-6; six winters' soil frost data missing	
PRECIPITATION AND STREAMFLOW				
Station	Elev. (m)	Aspect	Watershed	Record
W3	530-730	SW	42 ha	1958-2005
W7	620-900	N	77 ha	1965-2005
<i>Data from other regional stations</i>				
TEMPERATURE				
Station	Source	Elev. (m)	Record	Completeness
Hanover	NCDC	180	1893-2005	98.73%
Pinkham Notch	NCDC	610	1930-2005	98.81%
Mt. Washington	NCDC	1910	1948-2005	99.68%
RIVER FLOW				
Station	Source	Elev. (m)	Watershed	Record
Plymouth	USGS	140-1600	1610 km ²	1904-2004

Table 2: Climatic metrics examined at HBEF and regionally, with predicted direction of change (H) identified prior to the analysis. Trends are reported for the full length of each record, and also for only the most recent 40 years of data. Trends and 90% confidence interval limits are reported per decade; “M” indicates that missing data precluded the calculation of a confidence interval. “T” indicates a lack of valid confidence intervals in data sets with a large number of ties and near-zero trends. Slopes in bold differ significantly from 0 ($p < 0.10$; asterisks indicate $p < 0.05$). Shaded cells indicate trends in the hypothesized direction.

	H	years	units	full record Trend (units/decade)		90% CI	1966-2005 Trend (units/decade)		90% CI
MEAN ANNUAL TEMPERATURE									
G1	+	50	°C	0.13*	0.03 – 0.26		0.13	-0.01 – 0.32	
G6	+	45	°C	0.13	-0.02 – 0.27		0.19*	0.01 – 0.37	
G14	+	41	°C	0.32*	0.17 – 0.48		0.28	0.14 – 0.46	
G22	+	49	°C	0.24*	0.16 – 0.34		0.26	0.15 – 0.41	
Hanover	+	102	°C	0.12*	0.09 – 0.15		0.29*		M
Mt. Washington	+	56	°C	0.06	-0.03 – 0.16		0.26*		M
Pinkham Notch	+	69	°C	-0.04	-0.11 – 0.02		0.25*		M
Regional (Mitchell and Jones 2005)	+	102	°C	0.14*	0.10 – 0.18		0.23*		M
Global 30-60°N (Lugina et al. 2004)	+	117	°C	0.07*	0.05 – 0.08		0.29*		M
MEAN SUMMER TEMPERATURE									
G1	+	50	°C	0.12	-0.02 – 0.27		0.02	-0.24 – 0.23	
G6	+	45	°C	0.23*	0.08 – 0.40		0.21	-0.01 – 0.40	
G14	+	41	°C	0.25*	0.07 – 0.43		0.20	0.02 – 0.39	
G22	+	49	°C	0.37*	0.25 – 0.50		0.33*	0.17 – 0.49	
MEAN WINTER TEMPERATURE									
G1	+	50	°C	0.38*	0.11 – 0.69		0.45	0.04 – 0.87	
G6	+	44	°C	0.27	-0.01 – 0.56		0.26	-0.08 – 0.62	
G14	+	40	°C	0.45*	0.09 – 0.81		0.45*	0.09 – 0.81	
G22	+	48	°C	0.36*	0.15 – 0.61		0.31*	0.07 – 0.62	
ANNUAL MAXIMUM TEMPERATURE									
G1	+	50	°C	0		T	0		T
G6	+	45	°C	0		T	0		T
G14	+	41	°C	0		T	0		T
G22	+	49	°C	0.53*	0.00 – 0.90		0.71*	0.00 – 1.25	
ANNUAL MINIMUM TEMPERATURE									
G1	+	50	°C	0		T	0		T
G6	+	45	°C	0		T	0		T
G14	+	41	°C	1.25*	0.00 – 1.91		1.00*	0.00 – 1.74	
G22	+	49	°C	0.27	0.00 – 0.86		0		T
GROWING DEGREE DAYS									
G1	+	50	°C*days/yr	20.5	-5.6 – 4.8		12.7	-18.6 – 43.3	
G6	+	45	°C*days/yr	28.1	4.6 – 52.4		30.9*	6.4 – 65.2	
G14	+	41	°C*days/yr	42.4*	17.9 – 73.0		35.4*	13.8 – 64.8	
G22	+	49	°C*days/yr	54.2*	30.5 – 78.0		51.3*	25.3 – 84.3	
THAWING DEGREE DAYS (G1)									
January and February	+	50	°C*days/yr	2.9*	5.3 – 16.4		1.96	-0.77 – 5.00	
December - March	+	49	°C*days/yr	10.5*	4.6 – 52.4		8.82*	1.37 – 16.78	

Table 2 continued

	H	years	units	full record		1966-2005		
				Trend (units/decade)	90% CI limits	Trend (units/decade)	90% CI limits	
FROST-FREE SEASON (G1)								
Last spring frost	-	50	date	-1.58	-3.33 – 0.00	0.62	-1.94 – 3.21	
First fall frost	+	50	date	1.92*	0.32 – 3.33	0	T	
Frost-free season	+	50	days/yr	3.75*	0.81 – 6.57	0	T	
TOTAL ANNUAL PRECIPITATION								
W3	+	48	mm/yr	34.2	-3.0 – 69.4	7.0	-34.0 – 64.9	
W7	+	40	mm/yr	--	--	16.5	-35.4 – 67.8	
SUMMER PRECIPITATION								
W3		48	mm/yr	17.1*	2.0 – 30.0	2.8	-15.7 – 25.0	
W7		40	mm/yr	--	--	10.0	-12.1 – 33.5	
LONGEST DRY PERIOD								
W3	+	48	days	0	T	0	T	
W7	+	40	days	--	--	0	T	
INTENSE PRECIPITATION								
Days of precipitation > 50mm W3	+	48	days/yr	0	T	0	T	
Days of precipitation > 50mm W7	+	40	days/yr	--	--	0.33	0.00 – 0.79	
LOW STREAMFLOW								
Days with streamflow <0.1mm at W3	+	48	days/yr	-8.74*	-14.78 – -4.26	-2.50	-8.86 – 4.69	
Days with streamflow <0.1mm at W7	+	40	days/yr	--	--	-1.18	-8.39 – 4.58	
HIGH STREAMFLOW								
Days with streamflow >50 mm at W3	+	48	days/yr	0	T	0	T	
Days with streamflow >50 mm at W7	+	40	days/yr	--	--	0	T	
EVAPOTRANSPIRATION								
W3	+	47	mm/yr	-10.9	-19.6 – -3.5	-4.7	-14.2 – 4.0	
W7	+	40	mm/yr	--	--	3.1	-3.6 – 10.0	
TIMING OF SPRING RUNOFF								
W3	-	48	date	-2.07*	-3.33 – -0.35	-1.89	-3.27 – 0.61	
W7	-	40	date	--	--	-3.06*	-5.23 – -1.04	
Pemigewasset R. at Plymouth	-	101	date	-0.29	-0.83 – -0.24	-1.92	M	
SNOWPACK DURATION (G2)								
First measurable snowpack	+	51	date	1.67	-1.08 – 4.48	3.28	-0.48 – 7.82	
Last measurable snowpack	-	51	date	-2.50*	-4.27 – -0.83	-5.00*	-7.33 – -2.11	
Snowpack duration	-	51	days/yr	-4.00	-7.39 – 0.00	-8.60*	-12.18 – -4.36	
MAXIMUM SNOWPACK								
G2	-	51	mm water	-10.5	-21.9 – 0.0	-16.7	-31.0 – -0.9	
MAXIMUM SOIL FROST †								
G2	+	45	% area	8.8*	2.1–15.6	3.1	-8.1–14.3	

†Soil frost statistics are from least-squares linear regression; the non-parametric method was inappropriate due to a large number of ties resulting from years with no soil frost observed.

Fig. 1. (a) Location of hydrometeorologic stations used in this study. (b) Gage 1 at HBEF. Temperature and precipitation data have been collected continually at this location since 1955.

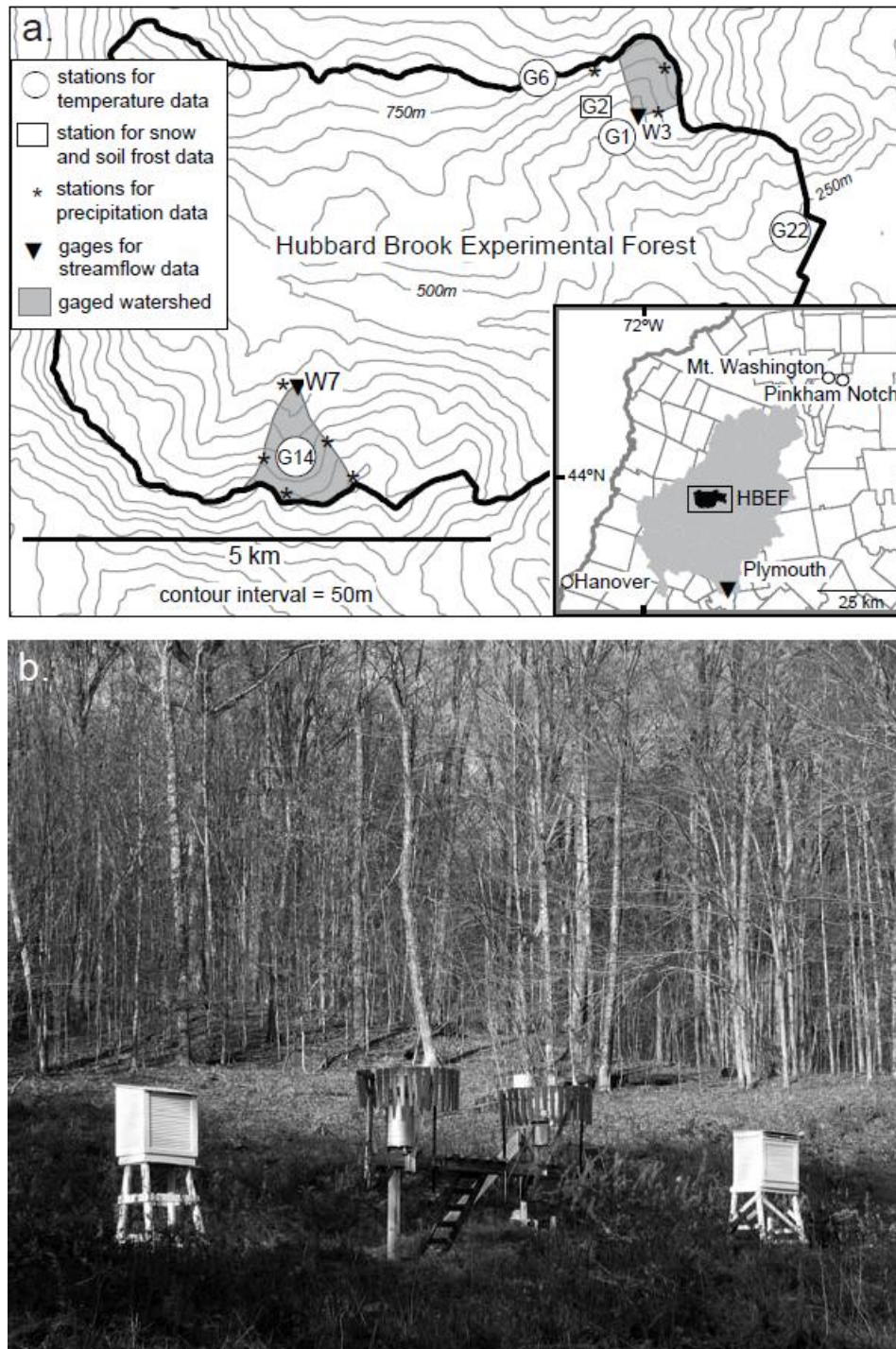


Fig. 2. (a) Mean annual temperature at the four longest-operating stations at HBEF. Differences in temperature among stations are attributable to elevation and aspect (Table 1). All four show significant warming trends. **(b)** Mean annual temperature at G1 station shows a pattern similar to integrated regional data (Mitchell and Jones 2005) and global data (Lugina *et al.* 2004).

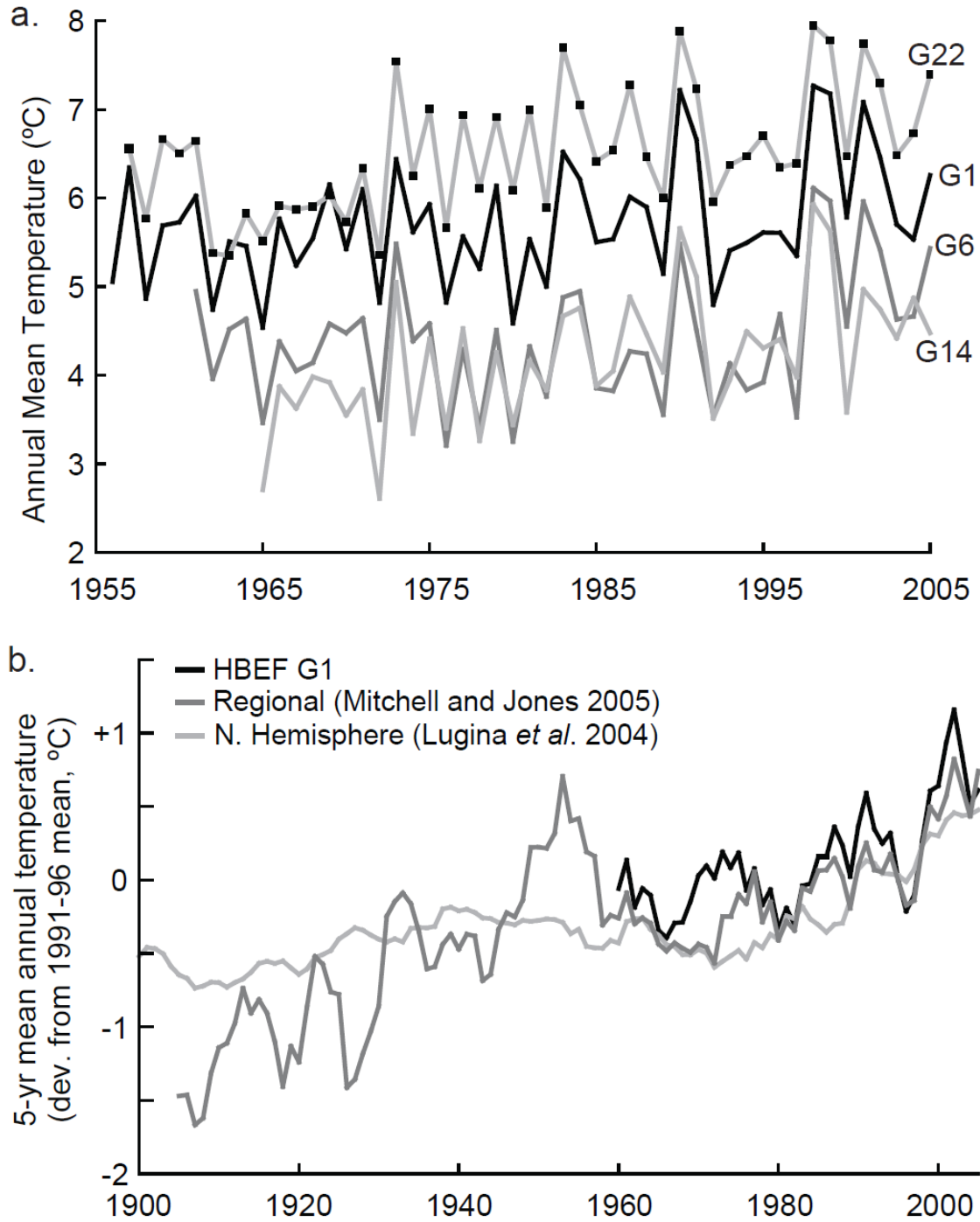


Fig. 3. Thawing degree days at station G1, have doubled for both the mid-winter months (January and February) as well as the entire winter (December to March).

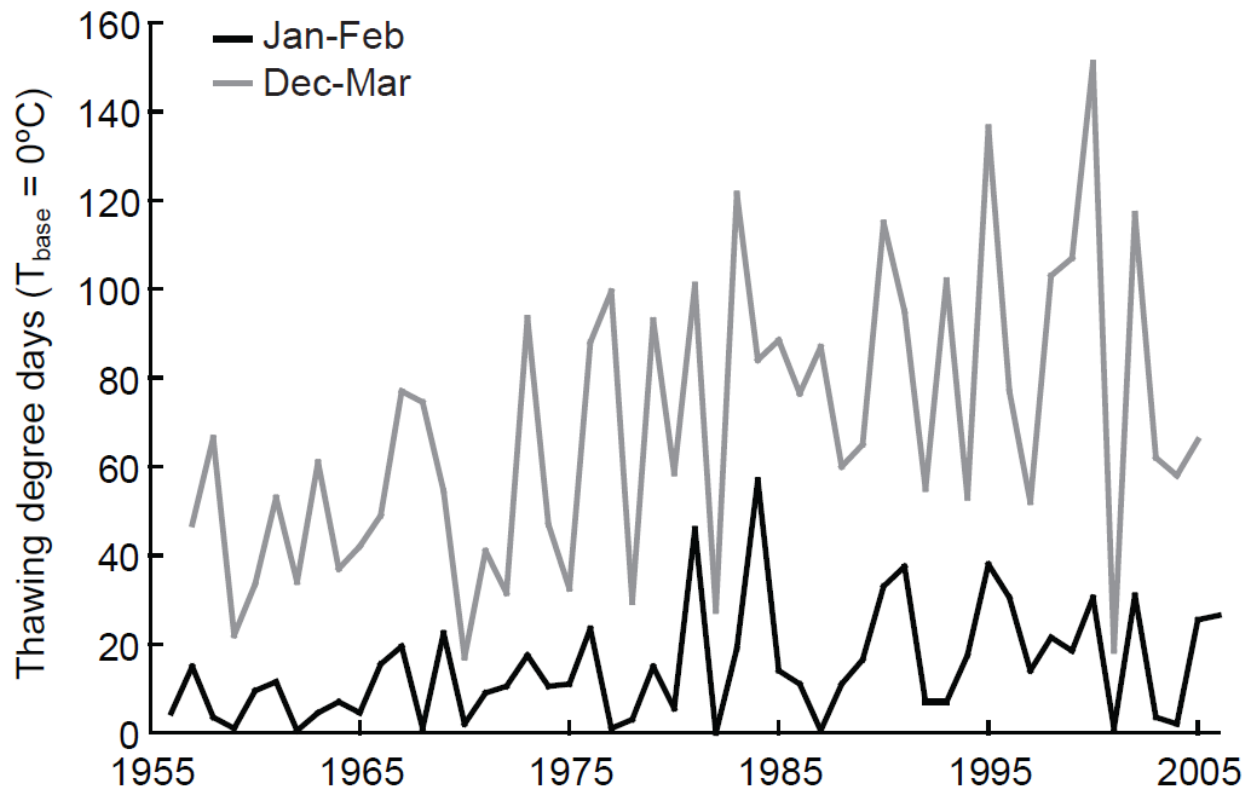


Fig. 4. Date of spring center-of-volume stream flow at W3 (south-facing) and W7 (north-facing). This date is a good indicator of the timing of spring thaw conditions, and occurs 10-12 days earlier now than at the start of the record.

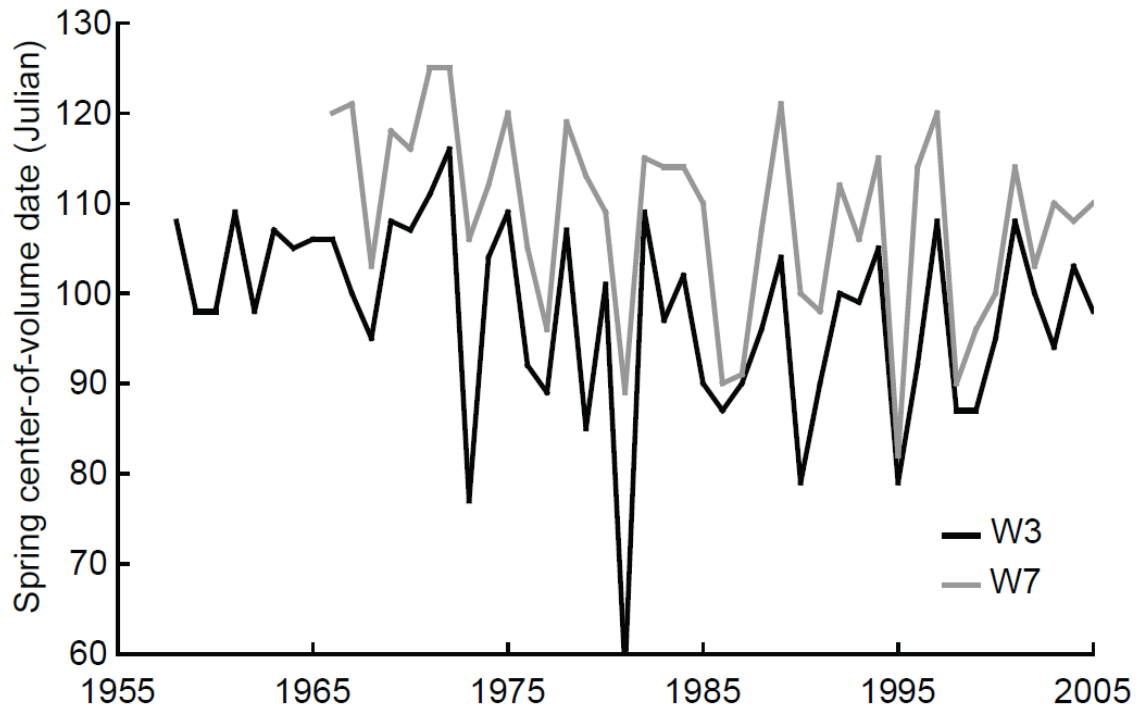


Fig. 5. Frequency analysis of (a) daily mean temperature at G1 and (b) daily streamflow at W3. The frequency of days in each bin between 1986 and 2005 is expressed as a percentage change from its frequency between 1958 and 1977. Under the null hypothesis of no climate change, we would expect small, non-systematic deviations from the 0% change line. Extreme cold days have become far less frequent, while extreme hot days have become slightly more frequent. Days of extreme low flow have become far less frequent, while extreme high flows have become more frequent.

