# Climatic Consequences of Leaf Presence in the Eastern United States

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ABSTRACT

At the time of leaf emergence in deciduous forests, markedly enhanced evapotranspiration leads to a rapid drop in the Bowen ratio. A small fraction of this surface flux alteration converges into the boundary layer, and this can be detected in the mean temperature and humidity daily increments at the surface. A simple technique is presented for identifying this response in surface climate data and extracting time series for the date of spring onset and for the ‘‘spring intensity,’’ a measure of surface energy budget partition change in spring. A tendency Bowen ratio $B'$ is found from changes in the daily increment of temperature and humidity in multidecadal averages. The spring date determined using this criterion for stations along the U.S. east coast corresponds to the date at which the normalized difference vegetation index (NDVI) reaches 80% of its seasonal maximum. Northward movement of the vernal front is similar to that obtained using Hopkins' empirical rule; it is linearly related to leaf emergence and flowering dates from the North American lilac phenology network. Spring intensity increases northward; the states from Virginia north exhibit distinctly higher values. There has been a trend in the most recent decades toward earlier spring dates, except for regions in Virginia and North Carolina. The same analyses performed using the small subset of stations with longer-term records indicate that a trend toward an earlier spring date is confined to recent decades. An inverse relationship between the spring date and spring average temperature was found for the Midwest but is inadequate for the northeast. Spring intensity has generally increased in northeastern North America throughout the twentieth century. However, large oscillations with an approximate 20-yr period distinguish the northeastern United States from the Midwest, indicating that the intensity of spring is not a simple function of spring temperature or of forest cover fraction.

## 1. Introduction

A striking land surface change occurs annually in the eastern North America with springtime leaf emergence in widespread deciduous forests. Mean temperature and humidity at a characteristic station (Fig. 1) indicate that the warming rate subsides sometime after day 100, roughly at the time of the smallest annual values of daily minimum relative humidity. This occurs near the date when the leaves emerge from dominant tree species in the region. The daily averaged surface energy balance (Fig. 2) reflects the fact that over the course of 10–14 days in spring sensible heat flux drops rapidly. The partition of net radiation shifts in favor of latent heat flux, most of which is transpiration (Moore et al. 1996a). Canopy resistance to water vapor transfer ($r_c$) drops during the same period in a manner directly proportional to the leaf area index (LAI) or normalized difference vegetation index (NDVI) during leaf emergence (Sakai et al. 1997). A detectable shift in daily temperature range occurs at approximately the same time (Schwartz 1996). An increase in the daily increment of specific humidity and a decrease in the increment in temperature indicate that a measurable portion of the surface flux alteration converges into the boundary layer. Sellers (1965, p. 106) showed a similar shift in Bowen ratio associated with spring in Wisconsin, but his budget approach involved more assumptions and used only monthly data, not suitable for present purposes. In springtime, a large modification of surface fluxes yields small but measurable temperature and humidity responses. The seasonal change illustrates the extent to which large land surface cover changes in a region may affect its climate.

The location and timing of the surface flux alteration over a wide area depends on the phenology of the dominant tree species. Leaf emergence in trees determines the northward speed of the vernal front, or ‘‘green wave,’’ as it is seen in satellite images. Interannual variability in spring date ($d_s$) has been related to heat sums, though other factors, such as chilling requirements, are important for some species (Schwartz 1997; White et al. 1997; Goulden et al. 1996).

The annual spring change in land surface mirrors a long-term shift in land-use pattern. With the abandonment of agriculture in the northeast, the forest cover fraction today is similar to what it was in the early

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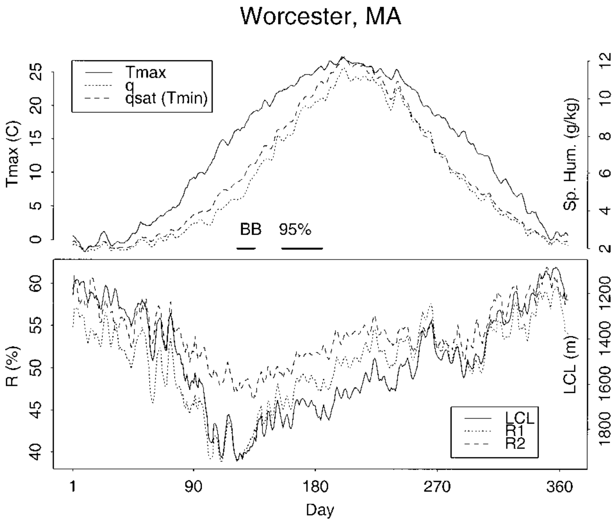


FIG. 1. 1946–94 daily averages for the Worcester, MA, climate station, smoothed within a 10-day running mean after averaging. (top) Daily maximum temperature ($T_{max}$, solid), daily average specific humidity ($q$, dotted), specific humidity found as the saturation specific humidity of the daily minimum temperature [$q_{sat}(T_{min})$, dashed]. Period marked ''BB'' represents the range of budbreak dates and ''95%'' the range of 95% of full leaf for dominant tree species during the 1990s at Harvard Forest, MA. (bottom) Relative humidity *R1* found using average $q$ and $T_{max}$ (dotted); and *R2* using $T_{max}$ and $q_{sat}(T_{min})$ (dashed); lifting condensation level [LCL($T_{max}$, $q$), right axis] at average station pressure.

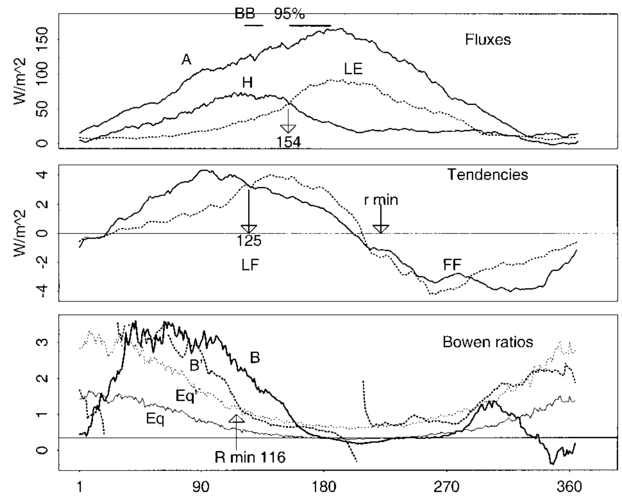


FIG. 2. (top) Daily averaged fluxes above the canopy at Harvard Forest for 1991–98. Available energy $A = (-Q^* - Q_g)$, sensible heat flux $H$, and latent heat flux LE (dotted). Here ''BB'' and ''95%'' are as indicated in Fig. 1, and $H$ = LE at day 154. (center) 1946–94 daily averaged and smoothed temperature tendency ($C_p \partial T/\partial t$, solid) and humidity tendency ($L \partial q/\partial t$, dotted). Tendencies are equal at day 125. Here ''r min'' represents the average date of the annual minimum canopy resistance at Harvard Forest (Sakai et al. 1997). Here ''LF'' and ''FF'' represent climatological last and first 0°C freeze dates. (bottom) Bowen ratios $B$ and $B'$ and equilibrium Bowen ratios $Eq'$ and $Eq$ as described in the text. Date of average annual minimum relative humidity ''R min 116'' indicated.

eighteenth century (Fig. 3), though the forest composition and age is quite different.

Our aim is to quantify changes in temperature and humidity in the lower atmosphere that have been occasioned by the increased presence of leaves, either seasonally or through long-term land-cover changes. We present some historical background (section 2) and describe our data sources (section 3). The approach is to identify spring date and intensity (a measure of lower atmospheric energy partition modification) from changes in daily trends in temperature and humidity alone (section 4), separating measures of vegetative state from measures of the climate response. In section 5 the northward progress of spring is compared with the venerable empirical rule presented by Hopkins (1918). For the satellite era, we also calibrate our spring date results by comparison with NDVI and with dates for leaf emergence and bloom from the lilac phenology network, available from the early 1960s. Using a network of 789 surface stations, we identify spatial patterns and temporal trends in spring date and intensity over the last four decades in the northeastern United States (section 6). We then examine similar results for a much smaller sample of stations with data for the entire twentieth century (section 7). Long-term differences in the rate of change in spring date and intensity in the reforesting eastern United States are compared with those observed at stations in the Midwest, where forest cover fraction remained small throughout.

## 2. Background

### a. Vegetation and climate

To link either seasonal or long-term vegetation changes to climate, three kinds of data have been used: (a)

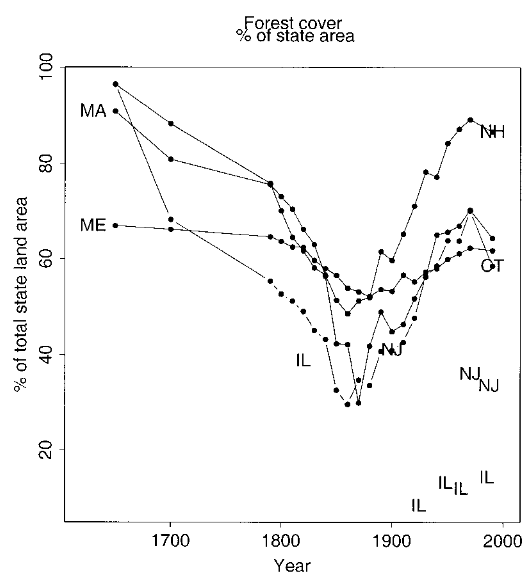


FIG. 3. Estimated fraction of the percentage of state land area in forest. States are given by their two-letter postal abbreviations. Data assembled by Foster and colleagues at Harvard Forest except for New Jersey (Russell 1997) and Illinois (Iverson et al. 1989).

plant phenological data (Schwartz 1998), (b) satellite-acquired measures of vegetation, such as NDVI, and (c) surface climate data. Phenological data are extremely valuable in climate studies. In the early parts of the century flowering and leafing dates were regularly reported as part of climatic data in the United States and in Europe (e.g., Jeffree 1960; Hopkins 1918, 1938). Long time series are unavailable, since this practice was discontinued in North America early in the twentieth century. The most extensive direct North American phenological data come from a 110-station network (Schwartz and Marotz 1986). Caprio (1993) assembled data for a similar network in the western United States. Budbreak, leaf emergence, and flowering dates are available for lilac and a few other species, but few stations have continuous data dating from earlier than the 1960s (Schwartz and Karl 1990). Similar data exist from phenological gardens in Europe (Menzel and Fabian 1999).

Satellite indicators of surface greenness, such as the NDVI, have provided detailed spatial coverage of the vegetation presence since 1978 (*TIROS-N*). The NDVI has been related to ground truth at a small number of sites, for which empirical relationships with LAI, $r_c$, albedo, and other surface biophysical properties have been established (Hall et al. 1995; Sakai et al. 1997). The clear limitation in either direct or satellite-based measures of leaf emergence is that they are available for 30 yr at best.

### *b. Identifying the date of spring onset and its northward velocity*

The simplest and most venerable estimate of the speed of northward march of the vernal front in the eastern United States is Hopkins' Law (Hopkins 1918; Reader et al. 1974). Hopkins drew on information from scores of farmers to develop a purely geographical rule: the onset of spring is delayed 4 days for every degree of latitude, a day and a quarter for each degree of longitude east of the origin, and a day for every 100 feet in elevation. Hopkins' Law, or another simple empirical method describing the northward march of spring can provide useful standards against which more sophisticated methods can be judged, much as simple persistence is the standard against which weather forecasts are measured.

Schwartz (1997) used the lilac network phenological state to guide development of statistical models to understand how leaf emergence alters surface temperature and humidity (sections 2a and 2b). Leaf emergence date has been modeled by others using schemes based primarily on estimating local heat sums, though other factors such as chilling requirements and solar radiation totals have also been considered (e.g., Hunter and Lechowicz 1992). White et al. (1997) presented a more complex model for use in climate models. This model predicts leaf emergence, incorporating existing phenological data and surface climate data into empirical multivariate functions. Validation of this model was compromised by the lack of long-term phenology and satellite-based vegetation index records. White et al. tested their model with only 3 yr of satellite and ground truth data, insufficient to assess interannual variability.

### *c. Is spring getting earlier?*

Some evidence is accumulating that Northern Hemisphere spring is advancing. Lilac phenology network stations with longer records in the eastern parts of North America appear to indicate a trend toward earlier leaf emergence since 1965 (Schwartz 1994). Menzel and Fabian (1999) report on budbreak dates for a variety of species found in the European phenological garden network. Results indicate that spring phenological events in Europe have advanced by 6 days during the last 30 yr. Schwartz and Reiter (1999) used a spring index model to identify a 6-day advance in spring date over the last 30 yr in North America. Using multiyear NDVI images, Myneni et al. (1997) detected spring earlier by 8 days in the North American boreal forest above 55°N from 1981 to 1991.

There are limitations to these findings. Satellite-based measurements are limited necessarily to the most recent decades. Using a single indicator species in a phenological network may limit the applicability of the results to the ecosystem, since the selected species may not behave as do the dominant participants in growing season evapotranspiration. This limitation was addressed in part by Schwartz (1997), who used three indicator species. Lilac is an understory species that leafs out earlier than do dominant North American deciduous trees. For example, Oglesby and Smith (1995) showed that while certain spring wildflowers in upstate New York have a tendency for earlier blooming throughout the twentieth century, a larger number showed no trend. Significant regional deviations from continental trends can occur. For example, European stations in the Balkan region indicated a later spring onset (Menzel and Fabian 1999).

Schwartz (1998) and Schwartz and Marotz (1986) developed an empirical model for spring index to link the phenological characteristics of lilac and honeysuckle species to climatic data. They inferred phenological spring onset from climate records since 1900. When using this approach, one mixes climatic and phenological data at the outset. The advantage is that results are then tied empirically to leaf state for given species, but the disadvantage is that the final index is a mixture of climatic and phenological data. The calibrating climate station sample includes very few stations in the northeast, comprising primarily stations in the Midwest (M. D. Schwartz 1998, personal communication). Schwartz confirmed that temperate North American spring appears to be arriving earlier between 1978 and 1990. However, he cautioned that this feature is less striking when one notes the wide variability of spring dates observed during the course of the century.

### *d. Does changing forest cover fraction have climatic consequences?*

Foster (1992) showed that forest cover fraction in New England (Fig. 3) is recovering to levels last seen at the time of first European settlement. Forest cover fraction in Massachusetts, for example, reached its minimum near 30% in the midnineteenth century, recovering to >60% after the 1980s, and slightly declining subsequently. It seems reasonable that there is at least a limited analogy between the climatic effects of annual leaf emergence and the long-term effects of afforestation.

While the relationship between changes in surface fluxes and surface temperature is clear on the seasonal scale, the impact of reforestation on climatic trends is open to question. If more forest cover leads to enhanced regional evapotranspiration, naive expectation would be that replacing pasture with forest in the eastern United States might lead to cooler temperatures than otherwise would have occurred. The springtime enhancement of transpiration directs available energy from sensible to latent heat. If more forest cover leads to increased evapotranspiration (ET), the intensity of the spring transition may be altered. Part of a supposed long-term warming trend might be masked by increased ET in the region. A shift to earlier spring dates, a possible consequence of an overall warming climate, could introduce a cool bias in average temperatures on a given date. A shift in the onset of biological spring could then be similar in importance to the shift in astronomical spring that Thomson (1995) found. He was able to detect springtime warming as a result of the shift in the equinox, owing to the precession of the perihelion of the earth's orbit.

Model studies of regions where deforestation occurs indicate that large changes to surface type might lead to small, but important changes in climate. In the Amazon, for example, the global climate model (GCM) study by Lean et al. (1996) indicated that a large deforested tropical region can be from 0.1 to 2.3 K warmer than the original, depending on canopy and soil moisture parameterizations. Bonan (1997) and Xue et al. (1996) presented model studies of the effect of vegetation on North American climate. In contrast to the findings of Foster (1992), Bonan argued that ''Land use practices have replaced much of the natural . . . forests of the Eastern United States with crops.'' Bonan reported that that introducing fewer, shorter trees to the northeast led to 1.1°C cooling there. On the other hand, Xue et al. (1996) found that replacing selected areas of eastern North America with trees rather than crops led to 1°C cooling in the east. We note that the resolution of these models is necessarily quite low at present. Though model studies of climate sensitivity are essential to determine how land cover can affect climate, their results rely on model accuracy of vegetation characteristics, including long-term successional changes of forested landscapes.

The presence of more active trees should also shift the regional hydrological balance, based on the assumption that forests are more efficient at returning precipitation to the atmosphere than would be a mix of forests and low vegetation areas. Kittredge (1948) noted, however, that there were published arguments for and against enhanced precipitation over midlatitude forests as compared to meadows. Karl and Knight (1998) recently found an 11%–20% increasing trend in moderate rainfall events in the eastern United States during the twentieth century.

Actually demonstrating these plausible arguments has been difficult. If ET were enhanced owing to increasing forest cover, the regional water balance could be shifted, for example, in favor of reduced streamflow. Most, but not all, studies of catchments for which trees have been replaced with grass generally indicate that forested areas exhibit relatively reduced runoff, indicating enhanced ET and interception in forested areas (e.g., Greenwood 1992). Greenwood and others note that the more deeply rooted forests do not want for soil moisture well into the growing season. Thus, evapotranspiration rates after leaf emergence in forested areas remain high even during drought, supporting the suggestion that the intensity of the spring transition might be enhanced during regional reforestation.

## 3. Data and processing

Turbulent heat and water vapor fluxes and the long- and shortwave components of the surface radiation budget at a 30-m research tower at the Harvard Forest (42°36′N, 72°14′W, ''HF,'' Fig. 4, inset) have been archived since 1991. Details are presented in Moore et al. (1996a) and Sakai et al. (1997). Climatic data for 54 first-order stations ($T$, $T_{min}$, $T_{dew}$) and 789 cooperative observer climate stations ($T_{max}$, $T_{min}$, Fig. 4) were extracted from the National Climate Data Center Summary of the Day and Surface Airways datasets (EarthInfo, Inc.). The dewpoint used for the first-order stations is the daily average. North American biweekly composite NDVI data [U.S. Geological Survey (USGS) EROS Data Center, Sioux Falls, SD] were obtained for the period 1990–98. This dataset is based on 1-km resolution data. The more numerous cooperative observer stations do not report a humidity variable directly. To obtain specific humidity for use in the calculation (2), we associate the daily averaged specific humidity with the saturation specific humidity of the minimum temperature [$q_{sat}(T_{min})$] (Hayden 1998). We tested this rule on the average data from the 54 first-order climate stations here and found it satisfactory. This method leads to slight overestimates in specific humidity (e.g., Fig. 1), but daily tendencies are only slightly affected. A

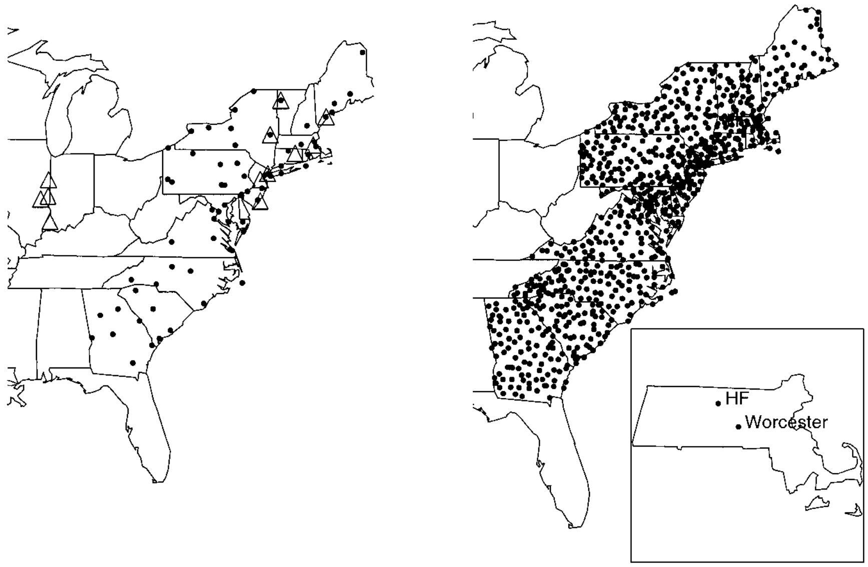


FIG. 4. (left) Locations of first-order climate stations used here. Triangles are stations with longer-term records referred to in section 7 and Table 1. (right) Locations of Cooperative Observer stations used. Inset: State of Massachusetts with location of Harvard Forest (HF) and Worcester indicated.

detailed case study presented in section 4 focuses on climatic data from nearby Worcester, Massachusetts (42°17′N, 71°48′W, Fig. 4, inset right). Studies of longer-term trends (section 7) were done using a subset of eight stations in the northeast and three comparison stations in the Midwest extracted from the U.S. Historical Climate Network (Karl et al. 1990; http://www.ncdc.noaa.gov/ol/climate/research/ushcn/ushcn.html). Station locations are shown in Fig. 4 (left) and detailed locations are presented in Table 1.

## 4. Energy flux convergence and springtime surface temperature and humidity changes

### *a. Development*

Rapid seasonal changes in surface fluxes (Fig. 2, top) provoked by springtime leaf emergence can change the average temperature and humidity as a portion of the altered surface fluxes converge locally. At Harvard Forest, 52% of the surface turbulent sensible heat flux and 42% of the latent heat fluxes occur during a 3-h period around local noon (Sakai et al. 1999, submitted to *J. Appl. Meteor.*). During these convective conditions, above a surface superadiabatic layer there is a buoyantly mixed convective boundary layer 1–2 km thick, idealized here as a mixed layer of thickness $h$. The seasonal trend in $h$ exhibits a springtime maximum in concert with a maximum in sensible heat flux $H$ that precedes a growing season enhancement for boundary layer forced cumulus clouds (Freedman et al. 2001). Forced cumulus frequency is enhanced during the growing season in the northeast (Freedman et al. 2001), a fact thatis hinted at by the rapid lowering of the surface-based lifting condensation level (Fig. 1). Freedman et al. present details of this phenomenon.

In any given year, frontal passage might obscure $T$ and $q$ changes associated with leaf emergence. However, the timing of such events is not fixed to calendar date, and the effect should be spread out over the spring. The changes resulting from leaf emergence can thus appear in long-term averages, but only if averages are taken over many years.

Mixed layer ($h$) average budgets are

$$h\rho L\frac{\partial q_m}{\partial t} = \mathrm{LE}_o - (\mathrm{LE}_h - \mathrm{ADV}_{qm} - \mathrm{SRC}_q)$$

$$h\rho C_p\frac{\partial \theta_m}{\partial t} = H_o - (H_h - \mathrm{ADV}_{\theta m} - \mathrm{SRC}_\theta), \qquad (1)$$

TABLE 1. Selected long-term climate stations.

| Station | Latitude (°N) | Longitude (°W) | Begin year | End year |
|---|---|---|---|---|
| Northeast | | | | |
| Albany, NY | 42.75 | 73.80 | 1874 | 1996 |
| Atlantic City, NJ | 39.38 | 74.43 | 1874 | 1987 |
| Blue Hill, MA | 42.22 | 71.12 | 1897 | 1996 |
| Burlington, VT | 44.47 | 73.15 | 1892 | 1996 |
| Central Park, NY | 40.78 | 73.97 | 1876 | 1987 |
| New Brunswick, NJ | 40.47 | 74.43 | 1893 | 1996 |
| Portland, ME | 43.65 | 70.30 | 1875 | 1996 |
| Storrs, CT | 41.8 | 72.25 | 1888 | 1996 |
| Midwest | | | | |
| Charleston, IL | 39.48 | 88.17 | 1897 | 1987 |
| Hoopeston, IL | 40.47 | 87.67 | 1903 | 1987 |
| Paris, IL | 39.63 | 87.70 | 1894 | 1987 |
| Princeton, IN | 38.35 | 87.58 | 1899 | 1987 |

where $o$ and $h$ subscripts denote surface and mixed layer top, respectively, subscript $m$ refers to a mixed layer average, $\rho$ is the air density, $L$ is the latent heat of vaporization, $C_p$ is the specific heat of air at constant pressure, $\mathrm{ADV}_{xm}$ represents the mixed layer average advective contribution ($x = q$ or $\theta$), $H$ represents the sensible heat flux, and LE the latent heat flux. Any of the terms in the budget equations can be important, depending on the timescale of interest. The diabatic flux divergence term ($\mathrm{SRC}_\theta$, primarily radiative flux divergence) is important to the layer budget, though the mixed layer moisture source ($\mathrm{SRC}_q$) is probably negligible in long-term averages. Only a very small shift in the convergence of the surface fluxes is needed to explain the observed springtime shifts in surface $T$ and $q$. In model simulation, the common approach is to estimate each of the budget terms in brackets in (1), altering parameterizations if tendencies depart too far from observations. Finding mixed layer flux divergence observationally has been done only for limited periods of intensive observation (e.g., Betts 1992). Necessarily empirical parameterizations based on limited intensive observations used in the simulations can compromise a model's ability to track subtle climatic trends. The approach used here, amenable to the analysis of many years of climate data, is to analyze the tendencies for $q$ and $\theta$. This requires careful filtering and averaging of long data series, procedures discussed in more detail in the appendix.

BOWEN RATIOS

Our goal is to determine spring onset using data from a large number of ordinary climate stations alone. We do this by tracking the seasonal course of the tendency Bowen ratio $B'$ and the flux Bowen ratio $B$, defined:

$$B' \equiv \frac{c_p \dfrac{\partial \theta_m}{\partial t} h}{L \dfrac{\partial q}{\partial t} h}; \qquad B \equiv \frac{H_o}{\mathrm{LE}_o}. \tag{2}$$

Periods for which $B' \approx B$ do occur (see below) implying that the bracketed terms in (1) are either negligible or are proportional to the surface flux. Here $H_h$ (strictly speaking the buoyancy flux) in convectively mixed layers is known to be a small negative fraction ($\approx -¼$) of the surface flux (e.g., Garratt 1992, p. 155), but no such relationship would necessarily apply to the humidity budget (1). Advective effects dominate the boundary layer budgets only just before and during frontal passages, less than 25% of the total time (Freedman 2000). Frontal passages, not being fixed by day of year, are thus not likely to dominate the long-term average budgets.

We identify the onset of spring using one of two related criteria. First, we seek a springtime minimum in the afternoon relative humidity. Second, we search for the first date at which $B'$ falls below one. These two criteria are related but not identical (see below). Potential advantages of using these methods to determine spring date include: (i) results should be less sensitive to changes to station location or instrumentation, since a ratio of daily differences is used in calculation; (ii) $B'$ exaggerates the differences seen in spring, since $q$ and $T$ changes are of opposite sign; (iii) the result is independent of changes in mixed layer thickness; (iv) information from large numbers of surface climate stations can be used; (v) determination of the spring date is explicitly connected to the lower atmosphere heat and moisture budgets, through (1); (vi) results do not depend on statistical relationships between spatially and/or temporally disparate climate and phenological data sets. The independent estimate for the date of spring ($d_s$) can then be compared with NDVI or phenology measurements. Disadvantages include: (i) long time series are required to reduce the effects of individual strong springtime synoptic disturbances on the average daily increments. This technique is not useful for studying year-to-year variability; (ii) in areas for which the transition is weak, it can be difficult to determine the $B' = 1$ date accurately; and (iii) we have assumed that mean advection does not contribute significantly to the budgets.

### *b. Case study for Worcester, Massachusetts*

We examine tendencies of the 46-yr (1949–94) average daily maximum temperature $T_{\max}$ and daily averaged $q$ at Worcester, Massachusetts (Fig. 2, center), chosen as surrogates for mixed layer values ($\theta_m$, $q_m$). Using the daily average temperature, or its common approximation $(T_{\max} + T_{\min})/2$ would place undue emphasis on nocturnal surface layer conditions, possibly more unrepresentative of the mixed layer above, and mix two independent measurements made at standard climate stations. Flux data for a similar number of years is not available anywhere. We can compare the tendencies with $H$ and LE averaged for 6 yr of observation at the Harvard Forest (Fig. 2, top). It is plausible that the flux averages made over a short period might come closer to a long-term mean more rapidly than would the tendencies, because the fluxes represent a more purely local observation.

The transition from spring to summer divides naturally into several periods. The $C_p\, \partial T_{\max}/\partial t$ maximum (an inflection point in $C_p T_{\max}$) occurs in early spring, roughly at the time of maximum daily increment in mean surface sensible heat flux. Before this time, the flux and the tendency are proportional, with about 5% of the surface sensible heat flux converging into the mixed layer. Freedman and Fitzjarrald (2001) formed composites describing boundary layer recovery for several days following cold frontal passage in the middle of the growing season in the northeast. The estimated temperature and humidity boundary layer convergence from the time of minimum just after cold frontal passage to maximum $q$ 5 days later is 53% for the surface LE flux and 49% for the surface sensible heat flux. However, if the reference is taken to conditions before the frontal passage, there is essentially negligible tendency in $q$ or $T$ over this period, as would be expected for midsummer conditions from the climatological average in Fig. 2.

In early spring, $H$ remains close to its annual peak for several weeks, dropping only as the vegetation begins transpiring (after day 100). After this date, the daily $q$ increment increases sharply, perhaps an indication of early understory transpiration (Moore et al. 1996b), a finding consistent with the date of first leaf from lilac phenology studies (Schwartz 1997). The $C_pT_{\max}$ and $Lq$ tendencies are about equal at day 125 ($B' = 1$), what we take to be the spring date, $d_s$. At Worcester, this also coincides approximately with the date of climatological last frost (0°C, ''LF'' in Fig. 3; Lautzenheiser 1969). Our estimate of $d_s$ also compares well with the date for initial budbreak for the most common deciduous trees at Harvard Forest (''BB,'' Figs. 1, 2; J. O'Keefe 1998, personal communication). By day 154 the flux Bowen ratio $B \rightarrow 1$. In early spring $L\partial q/\partial t \approx 5\%$ of LE, a period that ends with a dramatic upward shift in LE at about the time of O'Keefe's estimate of 95% leaf emergence.

An alternate criterion to describe the initial phase of spring is the date of the afternoon mean relative humidity minimum. At Worcester, $R$, a relative humidity based on $T_{\max}$ and $q$, daily average specific humidity, goes through a minimum near the time of leaf emergence (Fig. 1), at day 116. This is only a few days before the point at which $B'$ first equals 1 (day 125, Fig. 2). The relative timing of the two events is not fixed geographically (see below), since the former depends on the springtime value of $R$. At extremes of $R$, $\partial R/\partial t = 0$. Noting that $R \equiv q/q_s(T_{\max})$, and following the chain rule of elementary calculus, at these points $L\partial q/\partial t = R\varepsilon(C_p\partial T_{\max}/\partial t)$. It follows that $B' = 1/(R\varepsilon)$, where $\varepsilon = (L/C_p)(\partial q_{\mathrm{sat}}/\partial T)$, with $q_{\mathrm{sat}}(T, p)$ the saturation specific humidity at given temperature and pressure. Schwartz (1994) found a similar minimum in $R$ when composing surface climate data according to a spring date based on his spring index. This index is a statistical model linking surface temperature records to observed lilac leaf state, and was used as a substitute for actual first-leaf data.

Here $B = 1/\varepsilon$ is the ''equilibrium'' flux Bowen ratio (Monteith and Unsworth 1990, p. 196). Recall that only 5% of the springtime surface flux converges into the boundary layer. This feature of the average tendencies means that the notion that the evaporation proceeds at near to the equilibrium rate because fluxes are converging beneath a rigid lid (e.g., McNaughton 1976) is not relevant. If the Priestley–Taylor approximation were valid, $B \approx 1/(\alpha_{\mathrm{PT}}\varepsilon)$, where $\alpha_{\mathrm{PT}}$ is an empirical coefficient, often taken to be approximately 1.25. In midsummer, the smoothed daily rate of change of $q$ passes through zero, and $B'$ becomes indefinite near day 200. At Worcester $B'$ has several extremes during the summer and thus is constrained to reflect average values of $1/(R\varepsilon)$ (Fig. 2, bottom panel). It is plausible to suppose that $R$ in the PBL is controlled by the moderating influence of clouds, so that $\alpha_{\mathrm{PT}}$ and the mean $R$ are connected. Composites of case studies of this effect are presented in Freedman and Fitzjarrald (2001).

By midsummer, LE describes a broad peak determined largely by net radiation available, beginning to drop in mid-August, a time when canopy resistance at Harvard Forest reaches its annual minimum (Sakai et al. 1997). By the time of average first frost (''FF,'' Fig. 2), the tendencies are again equal. Thereafter, deciduous tree leaf fall first reduces and then frost eliminates the role of regional transpiration. The annual cycle is neatly described by the courses of $B$ and $B'$ (Fig. 2, bottom panel): (a) early spring $B' \approx B$, up to day 90; (b) transition, from day 90 to the time at which again $B' \approx B$, about day 170; (c) summer equilibrium, up to about day 270, and (d) autumn on to the end of the year.

To conclude the case study, we find:

(a) When many years of data are available, reliable estimates of the temperature and specific humidity tendencies can be found using only data from standard climate stations.

(b) Time series of tendencies showing distinct features of the annual cycle are modified by the presence of transpiring vegetation. The condition $B' = 1$ tracks the beginning of the temperature and humidity change associated with leaf emergence in the regional eastern deciduous forest around our case study station. This condition is preceded by a distinct minimum in $R$ (Fig. 1). We assert that the same criteria can be used to track the consequences of leaf emergence at other locations for which only climatic data are available. In autumn a second peak in $B$ (the ''brown nose'') occurs as leaves senesce and drop.

(c) On average less than 5% of the surface heat and water vapor fluxes converge into the lower atmosphere during springtime. In other words, only 4 W m$^{-2}$ in each component is sufficient to warm and moisten a New England site from spring to summer conditions.

(d) In the core of the summer, surface fluxes in the region approach an approximate equilibrium condition, with $B/B_{\mathrm{eq}}$ ($\alpha_{\mathrm{PT}}$, the Priestley–Taylor factor) equal to 1 at days 185 and 240, rising to a peak of approximately 1.5 in between.

The $B' < 1$ criterion was used to determine spring onset using the maximum number of years of $T$ and $q$ data for 54 first-order stations along the United States east coast. The clear latitudinal trend (not shown) in the predicted spring date indicated that using (2) with climatological data can provide an estimate on the arrival of spring. For all the climate stations studied, we found a monotonic decrease in $B'$ in spring similar to that shown for the example (Fig. 2). The day of the year for which $B'$ first equals 1 is a unique date, representing equipartition of boundary layer flux convergence into

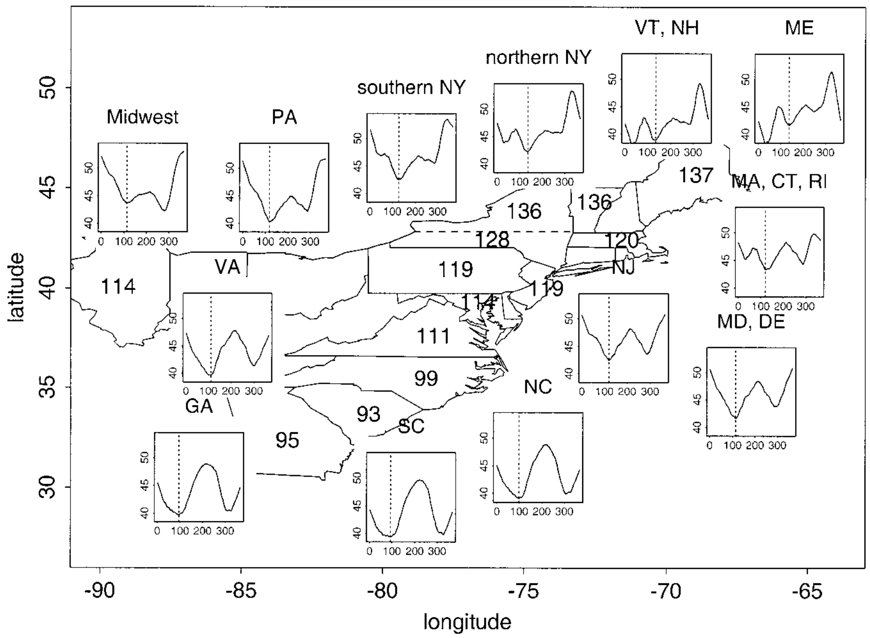


FIG. 5. Average relative humidity $R$ by day of year for eastern coastal states. Stations with at least 30 yr of data are included. ''Midwest'' refers to the average of the three Illinois stations and the one Indiana station listed in the table. Vertical dashed line represents the day of the springtime minimum of $R$, which is repeated on the map at each state center. Negative longitude equals west longitude from Greenwich meridian.

$H$ and LE. We identify it here as the spring date $d_s$. For a small number of stations, spurious results were obtained for signals nearly tangent to the $B' = 1$ line. To avoid this problem, we found the spring date by varying the criterion from 0.8 to 1.2 at intervals of 0.01, accepting the median value as the spring date. In the overwhelming number of cases, the consensus result was the same as that obtained using the $B' = 1$ criterion alone. Also, we found that when averages over several decades are considered, a clearly defined minimum in $R$ occurs within a week of the $B' = 1$ condition (see below).

## 5. Mean conditions in spring

### *a. Seasonal patterns of R along the East Coast*

There is a springtime minimum in the relative humidity $R$ evident in all stations. Calculated using $T_{max}$ and the average daily $q$, $R$ approximates the minimum surface relative humidity for the day. The related LCL (e.g., Fig. 1) is a close approximation to actual forced cumulus cloud base (Fitzjarrald and Moore 1994; Freedman et al. 2001). When daily averages are made over many years and assembled by state (Fig. 5), systematic changes in the annual cycle are evident. The spring minimum varies from day 95 in Georgia to day 137 in Maine. In the most northern states (Maine, Vermont, New Hampshire, New York) the spring $R$ minimum follows an earlier one, possibly indicative of the ''January thaw'' (Wahl 1952; Hayden 1976). This minimum appears only in the relative humidity; it is not present in the corresponding specific humidity series.

Going toward the south, the amplitude of this early minimum diminishes and disappears. In Georgia and the Carolinas, the annual cycle in $R$ is approximately sinusoidal with a 6-month period. The sharpness of the spring minimum in $R$ is largest in the northern regions. A second local minimum occurs in the autumn. This is consistent with the increase in flux Bowen ratio as transpiration is reduced at the end of the growing season (e.g., Fig. 2, top panel).

The patterns of $R$ are consistent with the idea that the springtime climatic change is more intense in regions with a large proportion of deciduous trees. The saturation vapor pressure curve is highly nonlinear, so $R$ time series combine the springtime effects of enhanced LE and decreased $H$. For this reason, we elected to describe spring onset using the linear $B' = 1$ condition.

### *b. Spring date ($d_s$)*

There are too few first-order stations (which have direct dewpoint measurements) to allow us to make maps of spring date with sufficient horizontal resolution to describe regional variation. Spring date was found for each of the 789 cooperative observer stations, linearly interpolated onto a grid, and a contour map produced (Fig. 6). The $d_s$ thus calculated ranges from day 89 in central Georgia to days in the 150s in northern

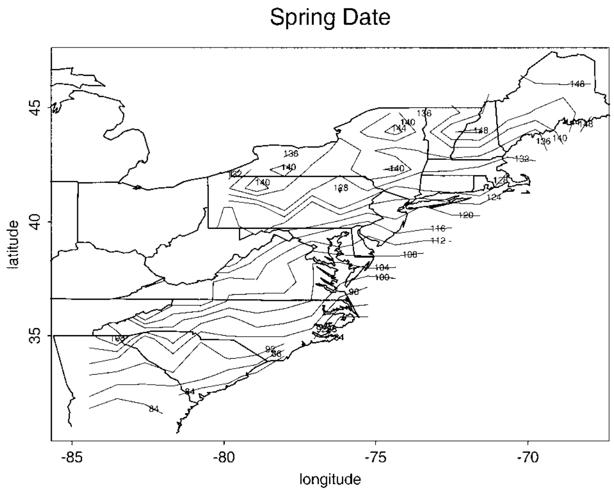


FIG. 6. Date of spring $d_s$ determined from the $B' = 1$ criterion described in text.

Maine. The vernal front does not progress uniformly, something noted previously by naturalists (e.g., Teale 1951). There is rapid motion through the Carolinas and Virginia, but progress is much slower in eastern Pennsylvania and southern New York State, an indication of the delaying effect of altitude that Hopkins had noticed. Sloping contours along the coast attest to a moderating maritime influence.

We developed animations to visualize the relative evolution of the $T_{max}$, $T_{min}$, and $d_s$ contours. The animations can be seen at http://sequoia.asrc.cestm.albany.edu/jrgroup/animations.html. The spring date most closely follows the $T_{min} = 6.5°C$ contour and the vernal front moves north. This is evident in the annual course of $T_{min}$ relative to $d_s$ for the entire 789 stations. For the stations considered: $T_{min}(d_s) = 6.6°C \pm 1.0°C$, normally distributed with a narrow range 3°–9°C, and $T_{max}(d_s) = 20.3°C \pm 1.7°C$, skewed positively with a range 12°–25°C. Because of the marine influence on coastal stations, contours of $T_{max}$ move eastward from the Hudson valley across Massachusetts, for example, even as the $d_s$ contour moves inland from the coast.

Spring date found from (3) (the first day for which $B' < 1$) is related to the date when NDVI at each spot reaches at least 80% of its maximum value for a given spring (Fig. 7, left). North American biweekly composite NDVI data (USGS, EROS Data Center) were used to get greenness values for the pixels centered about each of the 789 second-order stations for years 1991–98. The mean spring date ($d_s$) for each location was computed over these 8 yr. Though it comes as no surprise that spring arrives earlier in Georgia than it does in Maine, the quantitative results of the comparison (Fig. 7) are nonetheless remarkable. This is because(a) NDVI only are presented for biweekly composites that can represent irregular intervals at individual stations; and(b) such a small number of years of data are included in the average. Forty-one urban and isolated coastal stations that exhibited very low NDVI and small seasonal variation were excluded from the analysis, though their exclusion had little effect on the resulting linear regression.

The present results on average spring date do not depart greatly from Hopkins' 1918 rule for rate of northeastward motion of the vernal front (Hopkins 1918, also cited in Reader et al. 1974). Spring date based on Hopkins' Law was determined from the latitude, longitude, and elevation [DEM, USGS (EROS Data Center) Advanced Very High Resolution Radiometer companion disk] for each location. Day 80 was taken as the start date for the southernmost station used (30.7°N, 82.1°W),

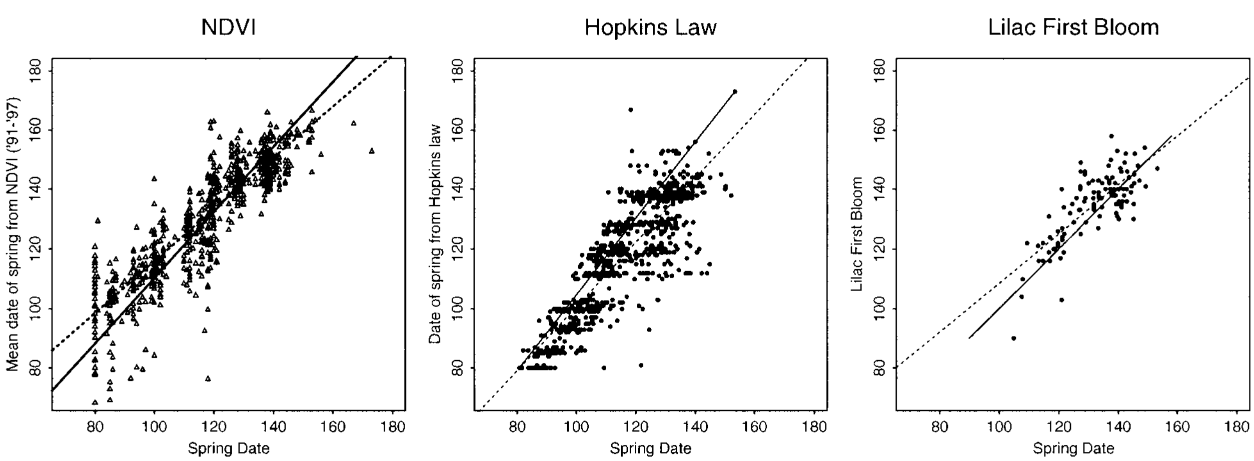


FIG. 7. Comparisons of spring date $d_s$ with other estimates. (Left) Spring date $d_s$ vs day of year at which the pixels nearest to the station reach 80% of their annual maximum, NDVI 1991–97. Two regression lines are shown: (a) line forced through the origin (solid), $d_{NDVI} = 0.93d_s$, $r^2 = 0.99$; and (b) Robust regression (dashed), $d_{NDVI} = 0.83d_s + 13$, $r^2 = 0.74$. The mean difference between the 80% of full NDVI date and $d_s$ is 11.8 days. (center) Spring date $d_s$ vs day of year for spring arrival as predicted using Hopkins Law ($d_{HOP} = 1.1d_s - 7.1$, $r^2 = 0.74$). Solid line represents equality, the dotted line represents a least-squares fit. (right) Spring date $d_s$ vs the date of lilac first bloom ($d_{LIL} = 0.8d_s + 26.4$, $r^2 = 0.59$). Solid line represents equality, dotted line represents a least squares fit.

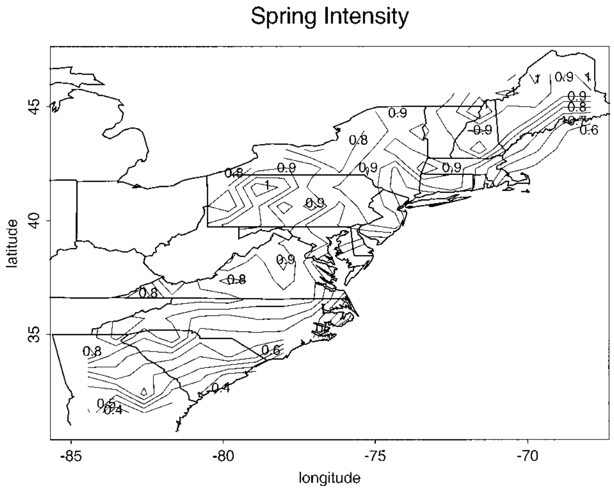


FIG. 8. Spring intensity $I_s$ determined from the $B' = 1$ criterion described in text. Contour lines are for 0.05 intensity intervals.

in southern Georgia (Fig. 7, center). Though many fewer stations are available from the lilac phenology network and they are available for a shorter period, values of $d_s$ compare reasonably well with the date of first bloom of lilac (Fig. 7, right).

### *c. Spring intensity $I_s$*

Spring intensity $I_s$ is defined as the difference between average tendency Bowen ratios for a window centered around the spring date, $d_s$:

$$I_s = \overline{B'[(d_s - w):d_s]} - \overline{B'[d_s:(d_s + w)]}, \qquad (3)$$

where the overbar indicates averaging. At individual stations the spring intensity ($I_s$) increases approximately linearly with the length of the averaging window ($w$), for $w$ between 5 and 40 days. Here $B'$ decreases at a constant rate for averaging periods centered at $d_s$ up to approximately 80 days. Therefore, the scale of the index depends on the window chosen, but the spatial and temporal patterns do not. Such behavior is observed for stations in the entire network, indicating that $I_s$ is a robust index, one that does not depend critically on the choice of $w$ as long as it is chosen within the linear range (5–40 days). For larger $w$, northern and southern stations respond differently to increases in window size. Here $I_s$ varies smoothly from $\approx$0.4 in Georgia, increasing to slightly $>$1 from Pennsylvania north (Fig. 8). Increased intensity is apparent at higher altitude and more inland sites, such as the Appalachian Mountains in Virginia. This pattern of $I_s$ is consistent with the popular notion that springtime arrives more abruptly in the north than in the south (e.g., Teale 1951).

Spring intensity $I_s$ is larger in areas dominated by hardwood species, in the north and in the higher altitudes in the south. Figure 8 indicates a lower spring intensity along the coasts in the Carolinas and Virginia, where conifers predominate (see Barrett 1962, Figs. 1–2). The shift from $I_s \approx 0.6$ to $I_s > 0.8$ occurs near the North Carolina–Virginia border. Forest inventory statistics indicate that the fraction of tree biomass in hardwood species (U.S. Department of Agriculture, http://rredc.nrel.gov/biomass/forest/forest_br/) changes from 55% in Georgia and South Carolina to 79% in Virginia and 95% in West Virginia. Only New Jersey of the northern states has this fraction $<$70%. Another aspect of the effect of tree type on climate are the distinct shapes in the seasonal course of $R$ presented in Fig. 5.

## 6. Patterns of change in recent decades

Data from a 556-station subset of cooperative network stations were used to determine the spatial pattern of changes in recent decades. The station subset was chosen such that selected stations have at least 40 yr of data. There was no geographical bias in the final dataset. We calculated the changes in spring date ($\Delta d_s$) and spring intensity ($\Delta I_s$) between the periods 1956–75 and 1976–95. Spring date is 6–8 days earlier in the northeast for the nominal period 1965–85 than in the previous 20 yr (Fig. 9a). In southern areas, the trend is not as pronounced. For large areas of Virginia and the Carolinas, the spring date is later by up to 8 days. This finding is not dependent on bias from an aberrant station. Ten of eleven stations in central Virginia, for example, show spring later in a range of 3–10 days. One station (Charlottesville Observatory, $\Delta d_s = -1.6$ days) shows a contrary trend toward earlier spring. However, this station is sometimes removed from climatic normals because of station siting (P. Michaels 1998, personal communication). The tendency toward later spring in central Virginia also appears if spring date is determined from the spring minimum in relative humidity $R$ (figure not shown). Excepting the Piedmont areas of Virginia and the Carolinas, the spring intensity trend (Fig. 9b) is broadly negative, representing a 10%–15% reduction.

Reductions in the intensity of spring of about 10% are found all along the East Coast. The largest differences occur in eastern areas of Massachusetts and Connecticut. In these regions the forests have not continued to expand in the last 20 yr and have been older on average. Murty et al. (1996) note that older forests transpire at a reduced rate when compared to their younger selves. We speculate that these factors might explain this reduction in intensity, but leave detailed consideration for future studies.

Since the partition of net radiation shifts from sensible heat into latent heat with leaf emergence, an earlier spring date does not lead to a warmer average spring. On the contrary, at any given date, earlier spring is slightly cooler and more humid. Let $m_-$ and $m_+$ (K day$^{-1}$) represent the daily temperature tendency for periods 15 days before and after $d_s$, respectively, and similar tendencies for specific humidity are denoted by $m_{q-}$ and $m_{q+}$ (q kg$^{-1}$ day$^{-1}$). Linear estimates of the biases in the average temperature and specific humidity at the

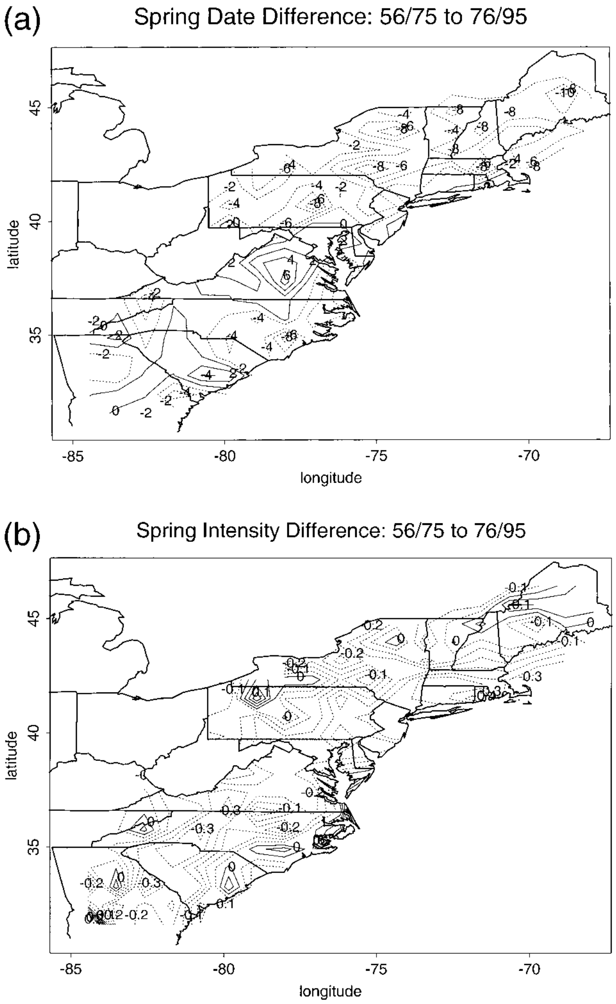


FIG. 9. Average difference (a) in spring date $d_s$ between the periods 1956–75 and 1976–95 and (b) in spring intensity $I_s$ between the periods 1956–75 and 1976–95. In (b) contour lines are for 0.05 intensity intervals.

original spring date are $\delta T_{ds} = \Delta d_s(m_- - m_+)$ and $\delta q_{ds} = \Delta d_s(m_q - m_{q+})$. For the case study for Worcester, Massachusetts (Fig. 2) with spring date 6 days earlier in recent decades (indicated in Fig. 9a), the tendencies around spring date are $(m, m_+, m_q, m_{q+}) = (0.222, 0.186, 0.068, 0.086)$. This leads to $\delta T_{ds} \approx -0.2°C$ and $\delta q_{ds} \approx 1.3$ g kg$^{-1}$. The maximum range of spring date differences indicated in Fig. 9a is 12 days. The linear analysis thus indicates that maximum distortions of up to 0.4°C and 2.6 g kg$^{-1}$ at springtime can occur.

## 7. Trends during the twentieth century

Are the changes in recent decades part of an even longer-term trend? Were sufficient daily data available, maps similar to Figs. 6 and 9a could have been produced for the entire century. However, we located only a limited number of selected stations in the northeast and in the midwest with many years of daily $T_{max}$ and $T_{min}$ records (Historical Daily Climate Data Network, Karl et al. 1990). Three stations in Illinois and one in Indiana serve as regions with little or no reforestation for comparison to the northeastern stations. (Refer to Table 1 and Fig. 3c for locations.) In most cases, a century or more of data for each station is available. We used the tendency Bowen ratio criterion to study the evolution of the spring date and intensity throughout many decades. Since humidity records are unavailable, $\partial q/\partial t$ in the $B'$ calculation (3) was inferred using daily increments in $q_{sat}(T_{min})$ as described earlier. Spring date and intensity were smoothed using a moving 20-yr window.

### *a. Spring date*

Though there is considerable variability in the course of spring date for the long-term stations (Fig. 10), some patterns emerge. Most of the stations in the northeast show a period of the earliest spring dates just before 1920. Only during the last 30 yr, up to the late 1980s, is there evidence of spring getting earlier. These trends are not evident, however, in our subset of Midwestern stations. When $d_s$ is split regionally (Fig. 11), it is clear that the midwestern and New Jersey stations are similar.

Unfortunately, our sample of long-term stations is small. To get a qualitative idea about the representativeness of each station, we found the interpolated values of the fields used to make difference maps (Fig. 9) for the long-term stations. These values (shown as dots in Figs. 10 and 12) indicate, perhaps unsurprisingly, that the least representative long-term sites for $d_s$ are Central Park, New York, and Atlantic City, New Jersey. Note, however, that the less urban coastal station Portland, Maine, not only has $d_s$ quite similar to inland Albany, New York, but also appears to be quite representative of the surrounding region. This might result from more inland characteristics of the climate station site. Though there is a weak tendency toward earlier spring in the overall data, only the northeastern stations show a sharp shift since the early 1960s. There is no striking trend in $d_s$ that corresponds to the overall period of warming in spring since the 1920s minimum.

We compared the present approach of estimating spring date with a more common method of linking spring to a heat sum. For example, Goulden et al. (1996) related leaf emergence at Harvard Forest to accumulated heating degree-days after day 100, taking spring to be the date at which 300 degree-days are accumulated. When smoothed over many years, this degree-day spring date resembles a scaled and inverted version of the mean spring temperature for the northeast shown in Fig. 11. The range of predicted spring date using the heat-sum approach and smoothing is much smaller than when using the present approach. Time series of the spring date found using this heat sum approach for all the stations in Table 1 were presented by Fitzjarrald et al. (1998). For example, at Albany this range spans only

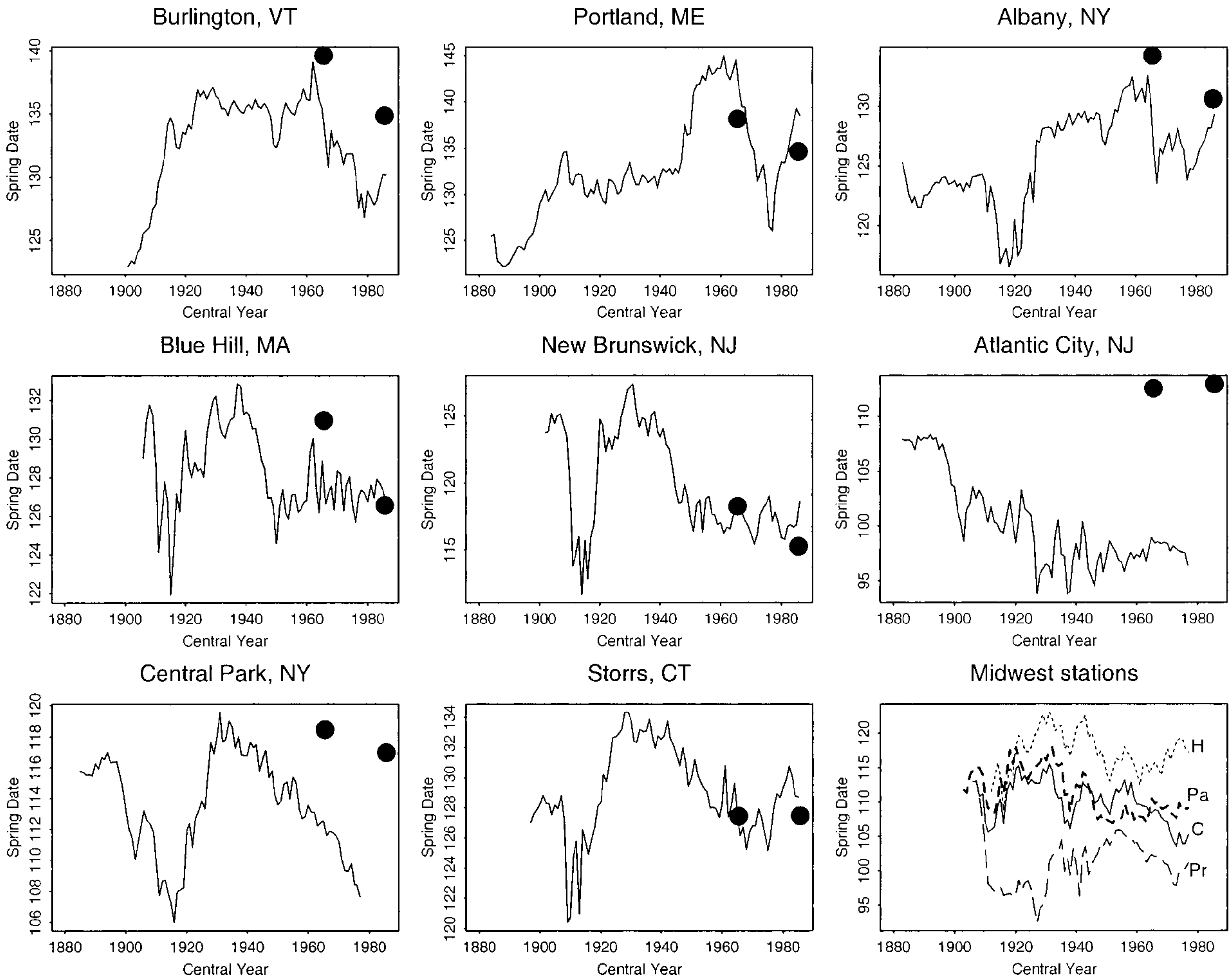


Fig. 10. Time series of spring dates $d_s$ for stations listed in the table. Curves were found using a moving 20-yr averaging window. Black dots represent the averaged $d_s$ for the grid point closest to the individual stations. Panel at lowest right represents data from the Midwest stations: Paris, Illinois, Waterworks (Pa); Hoopeston, Illinois (H); Charleston, Illinois (C); and Princeton, Indiana (Pr).

between days 119 and 121 during the twentieth century. For this reason, we did not pursue the heat sum approach further. To be useful to describe spring onset along the entire East Coast, a geographically based heat-sum zero reference day would have to be specified. By focusing on the consequences of enhanced transpiration, we argue that our method to identify the spring date is more climatologically relevant. However, we recognize that ours is a limited diagnostic rather than a prognostic approach.

### b. *Intensity of spring*

In contrast to the decrease seen in the most recent decades (Fig. 9b), there is an overall increase in the spring intensity among the stations during most of the twentieth century. Spring intensity has generally increased in eastern North America throughout the twentieth century (Fig. 12). However, large oscillations with an approximate 20-yr period distinguish the northeastern United States from the Midwest, indicating that the intensity of spring is not a simple function of spring temperature or of forest cover fraction.

Though there are too few stations to preclude doubt, the longer time series indicate that conclusions based on data from the last 40 yr would get the longer-term picture exactly backward. The trend toward earlier spring in recent decades seen in Fig. 9 is not unique in the century—spring was just as early in the period 1910–20. While it is reasonable to expect the spring transition to be more intense when there are more deciduous trees present, neither forest cover fraction or spring temperature describes long period oscillations in spring intensity that appear in the northeast but not in the Midwest (Fig. 13). The increase in spring intensity registered in Illinois indicates, however, that the overall warming tendency in the eastern United States is relevant.

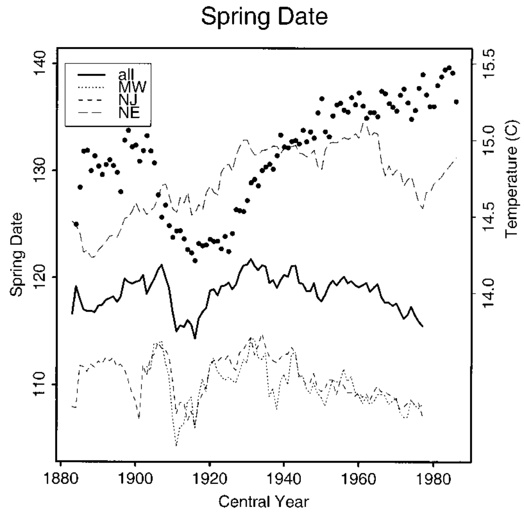


FIG. 11. Average spring dates by region. Solid line: all stations; long dashes: New England stations; short dashes: Midwestern stations; dotted line: New Jersey stations (including Central Park, New York City, for graphical purposes only). Dots (axis on right) presents the springtime (day 110–125) air temperature for the northeastern stations using a moving 20-yr averaging window.

## 8. Summary and conclusions

Using the simple technique outlined here one can identify the average surface climatic response to leaf emergence through reference to surface climatic data alone. The approach presented here links earlier studies of plant phenology with the boundary layer heat and water vapor budgets. A case study presented here illustrates that the boundary layer modification that accompanies spring leaf emergence in deciduous forests can be accomplished by converging only 5% of the surface fluxes into the overlying mixed layer. Date of spring for stations along the United States East Coast determined as detailed here corresponds to the date at which NDVI reaches 80% of its seasonal maximum. We made regressions on several stages of lilac phenology. Our $d_s$ is most highly correlated with the ''full bloom'' dates reported in the lilac phenology network (Schwartz 1996). The northward speed of the vernal front is just as closely correlated with Hopkins' (1918) empirical rule as it is with the lilac characteristics.

Returning to the questions posed in the introduction:

• *Is spring getting earlier?* Results indicate a tendency toward spring arriving 4–6 days earlier in eastern North America since the mid-1960s. The largest shifts in spring date are in the northern states. Except for regions in Virginia and North Carolina, there has been a trend in the most recent two decades toward spring arriving 4–6 days earlier. That spring appears to be occurring later in Virginia is difficult to explain. Freedman et al. (2001) demonstrate that the mid-Atlantic states exhibit the largest springtime occurrence of boundary layer cumulus cover fraction. We hypothesize that increasing trends in low cloudiness, perhaps a result of slow changes in ocean–land temperature contrast, might account for this feature.

• *Does changing forest cover fraction have climatic consequences?* Data from a small number of long-term stations indicates that the trend toward earlier spring dates in recent decades has not persisted long. The intensity of spring—a measure of how much the surface climate is altered in spring—shows an overall upward trend among stations over the twentieth century. Spring intensity is larger in the north than it is in the south or the Midwest. In the longer term, spring intensity in the northeast exhibits oscillations that have an approximate 20-yr period.

We suspect that the main climatic consequence of reforestation is in the degree of boundary layer modification that accompanies leaf emergence rather that in a simple trend in leaf emergence date. The recent reversal of trend could possibly be related to the effects that might limit early spring evapotranspiration in the northeast, such as logging and overall forest aging. Refining and testing this hypothesis, beyond our current scope, is left for continuing work. Properly assessing this relationship must await careful analysis of much more long-term data. Spring intensity was found to decrease in the northeast during the last two decades. Results for a small subset of stations with longer-term records indicate that spring intensity has increased throughout the twentieth century, though there have been large oscillations in the northeast. In contrast, the trend toward earlier spring is confined to recent decades.

Our findings support Schwartz's (1998) contention that the satellite record is much too short for definitive conclusions to be made about spring date. Those who make claims that there is unambiguous evidence of a secular change in spring onset based on satellite measurements may be premature.

In future work, many more long-term stations with daily data must be included in the analysis. This will allow us to determine the spatial pattern of changes in spring date and spring intensity during the reforestation of the eastern United States. Such information can then be used to assess the suitability of vegetation modules in large-scale models of surface–atmosphere interaction. We are pursuing the idea that changing sea surface temperatures offshore from the Mid-Atlantic States might account for the anomalous delay of spring seen during the last 30 yr.

Continuing studies also include looking at wide range of watershed responses. Parallel case studies of frontal sequences in spring and summer have demonstrated a link between leaf appearance and low cloudiness (Freedman and Fitzjarrald 2001). We will report later ongoing studies of the changing springtime thermal response as measured from satellites.

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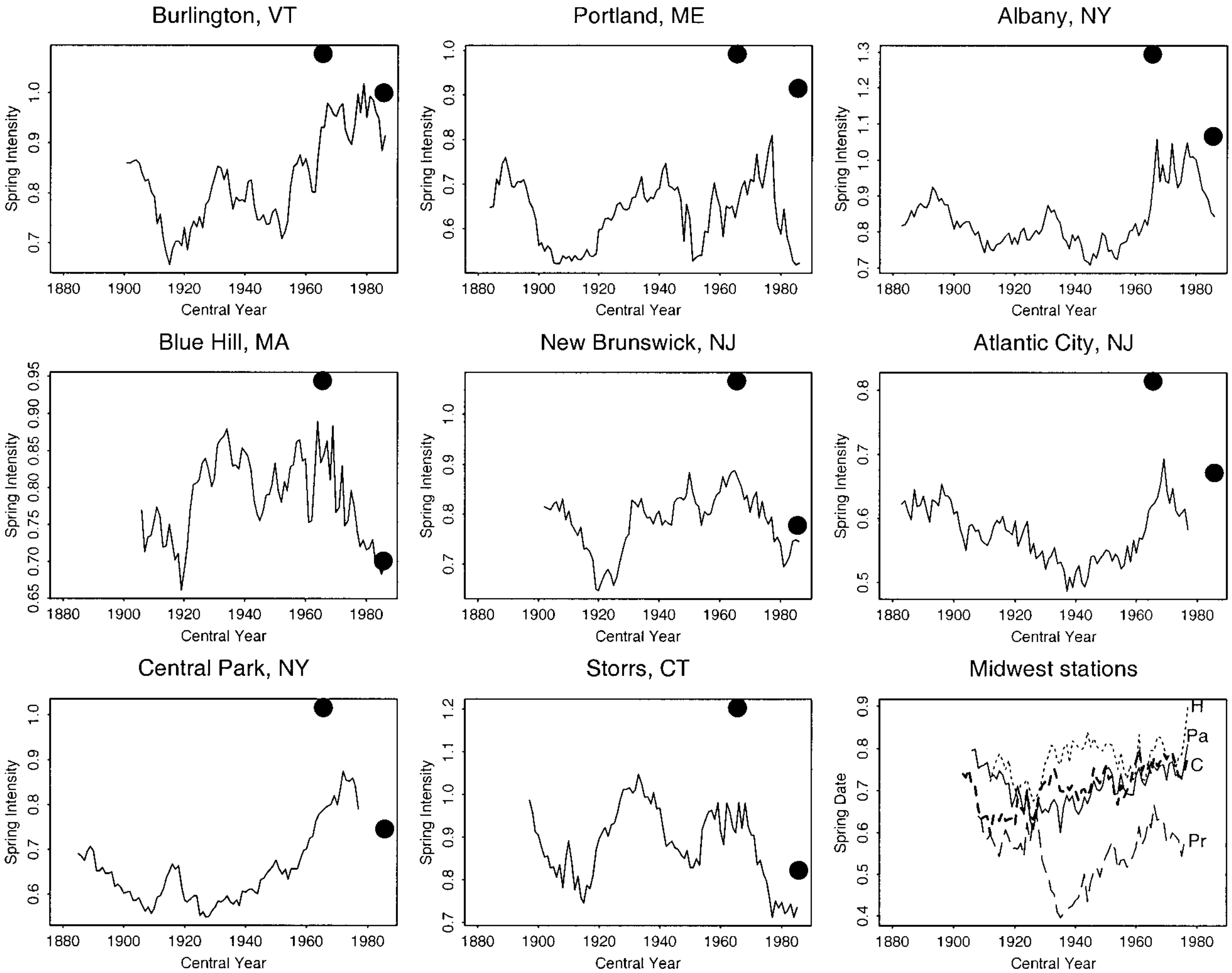


FIG. 12. As in Fig. 10 but for spring intensity $I_s$.

the Atmospheric Sciences Research Center (ASRC), University at Albany, SUNY. O. Acevedo acknowledges support for graduate studies at the University at Albany from CNPq, the Brazilian Science Foundation, Graduate Studies Scholarship 200619/96-9. Observations at Harvard Forest by ASRC are supported by the Northeast Regional Center of NIGEC, supported by DOE, under subcontract 901214 from Harvard University. We are grateful to Mark Schwartz, who provided data for Fig. 7c and offered useful information and many helpful comments during this work. We benefited from numerous discussions over the years with John O'Keefe at the Harvard Forest, who also collected and provided the phenology data referred to in Figs. 1 and 2. Our flux measurements at Harvard Forest have been done in collaboration with J. W. Munger and S. C. Wofsy of Harvard University. Audrey Plotkin and David Foster of Harvard Forest kindly allowed us access to their assembled forest cover data for New England, which appears in Figs. 3 and 13.

## APPENDIX

### Determining Tendencies

Careful evaluation of the time derivatives in (1) and (2) is required for our approach to be useful. Any technique to infer the time derivative from data exaggerates noisy high frequencies, and careful low-pass filtering is required (e.g., Hamming 1977, p. 42). We tested three approaches to determine the time derivatives on the 46-yr data series for Worcester, Massachusetts.

First, we applied the nearly equal ripple low-pass derivative filter (Kaiser and Reed 1977) on the series of $T_{\max}q(T_{\mathrm{dew}})$ and $q(T_{\min})$. Missing data were filled by linear interpolation. The filter applied here has an extremely sharp cutoff at 0.2 day$^{-1}$. Resultant series for the time derivative were then averaged by day of year, and $B'$ in (2) were then calculated.

The second approach was to smooth the 46-yr time series of the three primary variables using a locally weighted regression smoothing function ("*supsmu*" as

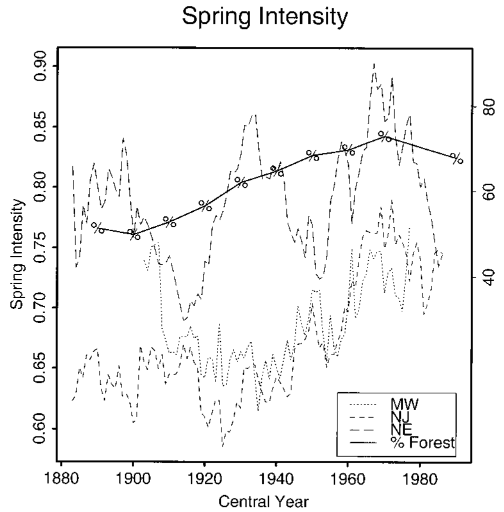


FIG. 13. As in Fig. 11 but for spring intensity $I_s$. Solid line with dots (axis on right) shows percentage of New England states in forest, as in Fig. 3.

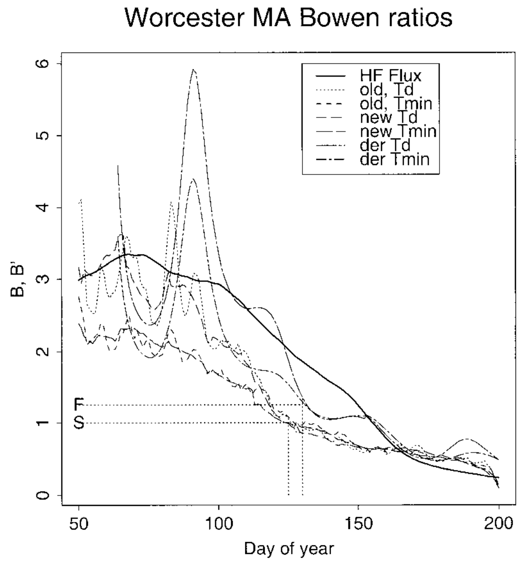


FIG. A1. Time series of Bowen ratios. The flux Bowen ratio $B$ (solid line) and tendency Bowen ratio $B'$ found using different techniques to solve for daily tendencies. Two methods of using the *supsmu* filter are used as described in the text. ''Old'' signifies finding the average over many years before smoothing to find the time derivative; ''new'' signifies finding the derivative before averaging by day of year; ''der'' refers to use of derivative filter; ''Td'' indicates that specific humidity is found using the actual dewpoint measurement; ''$T_{\min}$'' indicates that specific humidity was found using $q_{\mathrm{sat}}(T_{\min})$, as discussed in text. The dotted line marked ''S'' shows how $d_s$ is found using the old *supsmu* method described in the text, with the $B' = 1$ criterion. The dotted line marked ''S'' shows how $d_s$ is found using the Fourier derivative filter method described in the text, with the $B' = 1.25$ criterion.

implemented in Splus®; Mathsoft 1995, 7–44). The smooth version is found by finding the $k$ nearest neighbors (the *span*) and calculating the largest distance $\Delta x_o$ between a given point $x_o$ and other points in the span neighborhood. Then the function assigns weights to each point in the neighborhood using a tricube weighting function based on the fractional distance from a center point. For $u = |x_o - x_i|/\Delta x_o$, the weighting $W(u) = (1 - u^3)^3$, for $0 < u < 1$, otherwise $W(u) = 0$. The final smooth curve results from a weighted least-squares fit to the data. The function *supsmu* was applied here with a 30-day smoothing interval. After averaging the resultant series by day of year, a 15-day centered running mean smoother was applied before finding $B'$ in (2).

The third approach also uses the *supsmu* function. Here we simply formed day-of-year averages for a specific number of years. The resulting time series of $T$ and $q$ were smoothed twice using a locally weighted regression smoothing function. The smoothing span interval was taken to be 0.01 of the data record (3.6 days) during the first smoothing pass and 0.1 (36 days) during the second pass. Time derivatives used for the calculation of $B'$ were then found after the second smoothing by forward differences. For a small number of stations, $B'$ exhibited small oscillations near one. To minimize the effect of the arbitrary threshold ($B' < 1$), we found the median of the first day for which the Bowen ratio was smaller than $T$, where $T$ was varied between 0.7 and 1.5 in many steps. Doubtless, other kinds of derivative or smoothing filters could be used to accomplish the same task, with results similar to those presented here. A comparison evaluation of $B'$ at Worcester, Massachusetts, using the three smoothing techniques and two methods of determining specific humidity indicates that roughly comparable results for $d_s$ can be found (Fig. A1, ''S''). To obtain numerically equivalent results among the methods, it would be necessary to use a higher $B' \rightarrow 1.25$ criterion when using the Fourier filter methods (Fig. A1, ''F''). We are searching for relatively abrupt shifts in tendencies, and found that the strict Fourier filtering led to results that did not resemble the original time series. We deemed the *supsmu* procedures superior, and these are used in this paper. No important difference arises from the order of averaging in time or finding the derivative.

By including increasingly more years in the average, we found that the estimates of $d_s$ found in this way converge to a well-defined mean value after about 20 yr (Fig. A2). The spring intensity is taken to be the magnitude of the jump in $B'$, defined by (3), section 5.

It is somewhat difficult to determine the accuracy of our determination of the $d_s$ estimate. Analyzing variance in a traditional manner to find our error estimates is somewhat artificial. With the filters, we removed relatively high-frequency variance, presuming that such fluctuations reflect short-term modifications associated with local conditions or with the passage of synoptic disturbances. Agreement of the several smoothing approaches shown in Fig. A1 suggests that $\Delta B' \approx 0.1$ and $\partial B'/\partial t \approx 0.2$ day$^{-1}$, leading to an estimated error $\Delta d_s \approx 5$ days $\Delta B'/B'$, where $\Delta B'$ is the standard error of $B'$

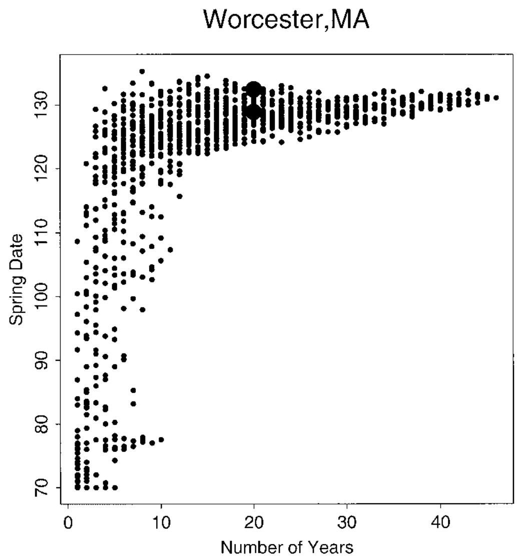


FIG. A2. Spring date (ordinate) as determined using the smoothing filter for different averaging periods (abscissa), using surface climate data from Worcester, MA, for the period 1949–94. Large dots represent the two 20-yr periods being compared in Fig. 9.

at $d_s$, is 0.25, indicating $\Delta d_s \approx 12$ days. These are large uncertainties, but they are in keeping with the scatter found when independent spring measures are fitted by simple linear regression (Fig. 7). When the 789 stations (Fig. 4) are used to make the $d_s$ estimates for the spatial averages shown in Fig. 6 and the difference in Fig. 9a, each grid point represents on average 5.6 stations. If we lump these stations together and assume that time and spatial averages are equivalent, then standard error of the $B'$ estimates would drop by a factor of $(1/(\sqrt{5.6}) \approx 0.4$. This would bring the uncertainty in $d_s$ to 2.5 days, and differences greater than about 5 days in Fig. 9a are believable.

Our method is somewhat arbitrary but it is objective. There is good agreement among our spring date obtained using these tendency estimates and independent measures of spring and autumn leaf state (Fig. 2). The fact that the resulting fields of $d_s$ and $I_s$ can be successfully contoured and interpreted (Figs. 5, 6, and 8) further strengthens our confidence in the approach.

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