MAINE
AUDUBON

20 Gilsland Farm Road
Falmouth, Maine 04105
207-781-2330
www.maineaudubon.org

June 1, 2010

Ms. Marcia Spencer-Famous
Land Use Regulation Commission
22 State House Station
Augusta, ME 04333-0022

Re: TransCanada's Kibby Expansion Wind Power Project Development Permit Application DP 4860

Dear Marcia:

Please find enclosed the Consolidated Parties' Post-Hearing Rebuttal. This filing includes:

- David Publicover Post-Hearing Rebuttal Testimony
- David Publicover Attachment B – Tang and Beckage 2010
- David Publicover Attachment C – Spear 1989
- David Publicover Attachment D – Seidel et al 2009
- Susan Gallo Post-Hearing Rebuttal Testimony
- Catherine Johnson Post-Hearing Rebuttal Testimony

I am sending this to you, Juliet Browne and Bob Weingarten electronically. We are also overnighting the originals plus 2 copies plus 12 disks to you and overnighting Juliet Browne and Bob Weingarten a hard copy and a disk.

Thank you for your consideration.

Sincerely,

Jenn Burns Gray

Jennifer Burns Gray
Staff Attorney and Advocate

STATE OF MAINE
LAND USE REGULATION COMMISSION

TransCanada Maine Wind Development) **Development Permit DP 4860)** **Kibby and Chain of Ponds Townships,)** **Franklin County)**	**POST-HEARING REBUTTAL** **TESTIMONY OF** **CATHERINE B. JOHNSON**

The Consolidated Parties offer the following in response to the May 24, 2010 post-hearing submission of Jean Vissering:

Vissering provided six points of post-hearing rebuttal, almost all which are a recitation of her pre-filed testimony with no real new information. The Consolidated Parties (CP) respond to three of her points.

Vissering's point 1. Views from the Public Lands Unit –

Vissering's rebuttal essentially restates her case that if the proposed Project is not readily visible from the terrestrial portion of the Public Lands Unit, it passes the State's scenic criteria. This defies common sense. When the State's Public Lands Unit is located on the shores of a Great Pond, and the primary use of that Public Lands Unit is for boating on the Great Pond, the Great Pond is a crucial feature of the Public Lands Unit, even if it is water and not land. Impacts to the scenic character of the waters within the Public Lands Unit are as significant to the Public Lands Unit as the impacts to the land portion of the Public Lands Unit.

In any event, as we noted in our testimony and Vissering acknowledges, the State is the rightful owner of public waters, regardless of the existence of the Public Lands Unit. The Chain of Ponds are, in themselves, a recognized outstanding scenic resource of statewide significance. Because the adverse visual impact of the proposed Project is significant from Chain of Ponds, and the public purposefully travels to the region specifically to boat on Chain of Ponds, Vissering's post-hearing rebuttal adds neither new information nor in any way rebuts the Consolidated Parties testimony that there would be significant and undue adverse scenic impacts to those utilizing the surface waters, particularly Long Pond and Bag Ponds, the Ponds with the most remote characteristics because the influence of Route 27 and other human factors is little to none. The Project as proposed would impact 31% - almost a third of the entire length of Chain of Ponds – and, as noted by Jim Palmer, would be "collectively dominant" of the scenic character from Long and Bag Ponds.

Vissering's point 3. Elimination of southern seven turbines –

Vissering attempts to suggest that the Consolidated Parties acceptance of some visual impacts from the northern eight turbines implies that an even larger cumulative impact from a full proposed build out is one and the same. The Consolidated Parties agree that some visual impact from the remaining eight northern turbines would still exist. But as we noted, their visual distance is greater and the severity and degree of the impact would be considerably less. And the impacts from the Project on the most visually sensitive parts of Long and Bag Pond, though not entirely eliminated, would be cut roughly in half if the seven southern turbines were eliminated, and would be at a greater distance. The seven most southern turbines greatly increase the visual impact on the Ponds, much more than Vissering's slight of words "*At the eastern end of Long Pond, the towers* ***<u>begin</u>*** *to slip behind Sisk Mountain in any case.*" These seven turbines would still be very prominent and closer to a viewer from Long Pond, and would be eye catching to say the least. And distance and viewer expectation are clear components of the Wind Power Siting Act on visibility.

Vissering's recently submitted Exhibit A also shows, in fact, that the impacts remaining after removal of the seven southern turbines would actually be less than the impacts we estimated in NRCM's Prefiled Testimony, Attachment C2 (which was based on the only information available prior to the hearing.) While the length of Chain of Ponds from which the remaining eight turbines would be visible is the same in both exhibits (Vissering estimates 17%; NRCM estimated roughly 10%; but the actual areas shown on the maps are identical), Vissering's Exhibit A shows that the actual <u>number</u> of turbines that would be visible from various location in Long and Bag Pond would be less than NRCM estimated (1 – 3 turbines or 3 - 6 turbines[1] rather than NRCM's estimated 6 – 9 turbines.) This provides further evidence that removal of the southern seven turbines would greatly decrease the adverse scenic impacts on Long and Bag Ponds.

Vissering's point 4. Scenic Impact Standards –

Vissering provides interesting but convoluted logic that the debate on visual impacts should mostly focus on scenic resources identified and documented through a public process. The designation of Chain of Ponds through the Wildlands Lakes Assessment and the BPL Public Lands Management Planning Process meet that criterion. She references the Wind Power Siting Act's reasonable guidance to consider visual impacts including "distance, viewer expectations, duration of view, etc." Is not 31% of

---

[1] Note that Vissering's testimony includes "3" turbines in both categories, so it is unclear in what color areas from which three turbines would be visible are shown on Vissering's Exhibit A map.

the length of the waters of Chain of Ponds, where visitors would be looking directly at, at close range, a reasonably large visual impact of the Project? If 31% would not constitute a major visual impact, particularly in an environment with little other human visual impact, what percentage would? Given the extensive public processes for the Wildlands Lakes Assessment, that resulted in an outstanding scenic designation for Chain of Ponds, and BPL's Public Lands Management Planning Process that resulted in a Plan highlighting the "highly scenic" and "wild and scenic" character of the Chain of Ponds region[2], Vissering's suggestion that LURC rely on those public processes leads to the conclusion that her finding of no undue adverse impact on scenic resources is not supported by the facts of those public processes.

Summary –

Nothing in Vissering's May 25, 2010 submittal changes our conclusion that the southern seven turbines cause an undue adverse scenic impact on a variety of scenic resources of state and national significance and related uses. No permit should be granted for these southern seven turbines.

June 1, 2010.

---

[2] Comments by Maine Bureau of Parks and Lands Chief of Planning, Kathy Eickenberg, May 12, 2010; BPL's Flagstaff Region Management Plan (including Chain of Ponds), p. 92, p. 31.

June 1, 2010

VERIFICATION

Catherine B. Johnson
Signature of Witness: Catherine B. Johnson

June 1, 2010

Before me appeared Catherine B. Johnson, who being duly sworn, did testify that the foregoing testimony was true and correct to the best of her knowledge and belief.
State of Maine

Kennebec County

Leisa Dennett
NOTARY PUBLIC

LEISA DENNETT
Notary Public, Maine
My Commission Expires November 2, 2014

STATE OF MAINE
DEPARTMENT OF CONSERVATION
LAND USE REGULATION COMMISSION
IN THE MATTER OF

TRANSCANADA MAINE WIND
DEVELOPMENT INC.

KIBBY EXPANSION WIND POWER
PROJECT, DP4860

KIBBY AND CHAIN OF PONDS TWPS.
FRANKLIN COUNTY

POST HEARING
REBUTTAL
TESTIMONY OF
SUSAN GALLO
ON BEHALF OF
MAINE AUDUBON

The Consolidated Parties offer the following in response to the Post-Hearing Testimony submitted by Dana Valleau on behalf of TransCanada.

1. Data Regarding Bicknell's Thrush Use of Regenerating Clear Cuts in the Project Area.

We acknowledge that there was one incident of Bicknell's thrush observed in a regenerating clearcut in the Kibby One project area (as described on page 1 of D. Valleau's post-hearing testimony), and do not disagree that Bicknell's thrush might occasionally be found in a high-elevation regenerating clearcut. However, it is likely that regenerating clearcuts would provide lower quality habitat compared to naturally disturbed forests, and there is ample evidence in the scientific literature showing that lower quality habitat often attracts singing males with little or no chance of successful breeding. Additionally, to take one observation in the western mountains and jump to classifying 90,000 acres of regenerating clearcuts across the state of Maine as potential Bicknell's thrush habitat is ecologically unsound and misleading. There is no documentation in the scientific literature, by the staff at the Maine Department of Inland Fisheries and Wildlife (MDIFW), or by scientists at the Vermont Center for Ecostudies

(VCE) of Bicknell's thrush breeding *successfully* in regenerating clearcuts in Maine. There are also no current studies by either MDIFW or VCE underway to further evaluate Bicknell's thrush use of regenerating clearcuts in Maine, largely we believe due to lack of support for the idea that this forest type offers any significant amount of habitat for Bicknell's thrush.

2. Mapping of "Core" Breeding Areas.

As context, we point out that the Consolidated Parties made multiple verbal and written requests to TransCanada for the Bicknell's thrush data starting in the summer of 2009 and continuing through the late spring of 2010, but every single request was ignored. Then, just a few days before the hearing, we were sent a report, Breeding Bird Survey Report for the Sisk Mountain Wind Power Project (Folsom and Evers November 2009), with Bicknell's thrush data. Why such a critical report was omitted from the application is odd particularly since every other wildlife survey report was included in full. Furthermore, neither of the authors was made available for cross examination. Mr. Valleau is correct in his post-hearing comments that we addressed the Folsom Evers report for the first time at the hearing. We simply had no other option.

The spot-mapping technique used by the applicant is difficult to undertake in the best of circumstances. In stunted spruce fir habitat, it is all but impossible and, paired with the unusual habits of Bicknell's thrush (multiple males with no territories mating with multiple females who do defend territories), even more so. Spot-mapping results therefore must be interpreted with caution, taking into account the actual time spent on surveys (unknown because no documentation of hours spent per plot was provided) and the difficulty for the researchers to physically move around in this habitat. There is little doubt that a string of wind turbines and their associated road system through the center of what TransCanada has delineated as "core" habitat would be the worst case scenario for Bicknell's thrush, as this is where the applicant's surveys documented high numbers of Bicknell's thrush. But for the applicant to go to the edge of where those spot-mapped observations were made and then say that there are no impacts beyond this outer boundary of observations is unwarranted and does not take into account the limits of the data that were collected.

This is especially important for a bird like Bicknell's thrush that uses dynamic habitat, habitat that will move around the landscape over time as local disturbances create new "core" habitat and old "core" habitat grows into unsuitable habitat once again. The applicant has consistently failed to address the fact that Bicknell's thrush habitat shifts across the landscape, which is of particular importance when it comes to interpreting observation data for Bicknell's thrush in the project area.

The circles that were used in my direct presentation at the hearing were an *illustration* of how Bicknell's thrush data might be interpreted. Given the extremely limited time to discuss complex ecological concept of habitat, we wanted to demonstrate the narrow way that TransCanada had interpreted the data, and how impacts could likely be much greater than those articulated by the applicant. In reality Bicknell's thrush, or for that matter any wildlife species, will never have a perfectly circular territory. Had the illustration been made with an oval or an irregular polygon or any other shape of the given area, the message would have been the same. Some areas were not surveyed for Bicknell's thrush or covered by spot-mapping efforts. It is very reasonable to interpret the data in a way that significantly increases the area of habitat for Bicknell's thrush impacted by this project. In fact, the assertion by TransCanada that the project will *only* impact eight acres of Bicknell's thrush habitat is extremely restrictive and does not take into account any edge effects, completely ignoring the likelihood that Bicknell's thrush habitat will be affected beyond the edge of the actual footprint of the turbine pads and roads.

3. Maine IF&W Comments on 2007 Endangered/Threatened Species Process and Habitat Impacts in Redington Wind Power Project.

Special Concern is an administrative listing by MDIFW of species that are believed to be vulnerable and could easily become threatened or endangered because of restricted distribution, low or declining numbers, specialized habitat needs or limits, or other factors. They include species suspected of being threatened or endangered or likely to become so, but for which insufficient data are available (http://www.maine.gov/ifw/wildlife/species/endangered_species/listing_categories.htm).

The state and federal threatened and endangered species listing processes are the last means of defense for protecting rare species. These processes do a good job at catching species on the brink of extinction on two different scales, within state boundaries and within the country as a whole. But, endangered and threatened species listing is not the right tool for species that are restricted regionally to multiple states, like Bicknell's thrush. It's like saying a screwdriver is the only tool you'll use for any project. It works well if most of your projects use screws, but when you get to a project that has nails, it just isn't going to do the job. Lack of listing should in no way diminish the priority of conserving Bicknell's thrush on its breeding grounds.

VERIFICATION

Susan M Gallo

Signature of Witness: Susan M. Gallo

Before me appeared Susan M. Gallo, who, being duly sworn, did testify that the foregoing testimony was true and correct to the best of her knowledge and belief.

NOTARY PUBLIC

attorney at law
bar # 008398

STATE OF MAINE
LAND USE REGULATION COMMISSION

TransCanada Maine Wind Development) **Development Permit DP4860)** **Kibby and Chain of Ponds Townships,)** **Franklin County)**	**POST-HEARING REBUTTAL** **TESTIMONY OF** **DR. DAVID PUBLICOVER** **APPALACHIAN MOUNTAIN CLUB**

The Consolidated Parties offer the following in response to the May 24, 2010 post-hearing submission of Dana Valleau.

In the post-hearing May 24th, 2010 filing by Dana Valleau (at Section 4. *Timber harvesting activities affecting subalpine forest*) for TransCanada, the Applicant continues their attempt to minimize the ecological importance of the documented subalpine forest (Fir-Heartleaved Birch Subalpine Forest) occurrence on Sisk Mountain and the severity of impact from this Project. It is an attempt to partially salvage Dr. Hudson's three false premises, specifically:

i. that the Fir-Heartleaved Birch Subalpine Forest is not particularly rare,
ii. that the occurrence on Sisk Mountain is not particularly important, and
iii. that the impact of project development on this occurrence would not be significant.

As the Consolidated Parties outlined during the hearings, Dr. Hudson's assumption that this community is not rare regionally and that additional areas exist in eastern Canada and other parts of New England is misleading. Occurrences of this community documented by the Maine Natural Areas Program occupy only 0.20% of the state's landscape. Dr. Hudson claims that other undocumented examples exist in Maine, which at a maximum would increase the total land cover of this community to 0.24% of the state. He also suggested it is common in Canada, while ignoring the fact the Canadian examples do not necessarily have the same species composition or disturbance dynamics due to their more northern latitude, much lower elevation, and not being subject to the geophysical environment (such as the high wind regime) found on Maine's mountains. In his post-hearing filing, Mr. Valleau correctly makes the case that this rare habitat is at risk from logging activities. The same is also true for the low-elevation occurrences of this community in Canada.

Regardless of the Canadian examples, LURC's legal responsibility extends to the resources and their status within its jurisdiction. Otherwise it would mean that LURC should ignore its responsibility to protect any rare resources that might be more common elsewhere, even if they are officially recognized as rare or endangered in Maine. Alpine habitat in Maine is not less significant because there is more of it in the White Mountains of New Hampshire. LURC must base its decision on the value of this particular occurrence with a Maine legal context.

Dr. Hudson's second false premise that the occurrence of this habitat type on Sisk (358 acres) is not particularly important ignores that it is the eleventh largest of nineteen documented occurrences in the state. And of the statewide documented 40,000 acres of this community, 86% is located at just five sites – Mount Katahdin, the Mahoosuc Range, Bigelow Mountain, Redington/Crocker and Baker/Lily Bay. Occurrences outside the largest mountain ranges may contribute a relatively smaller part of the total acreage of this community, but contribute to its presence and diversity across the landscape. The seven smallest sites documented by MNAP are but 10 to 50% of the size of the Sisk occurrence. Size is used as a criterion in the Maine Natural Areas Program Element Occurrence (EO) Rank for rare communities and larger occurrences typically have greater habitat diversity, greater resistance to being destroyed, and greater resilience to recover from stresses.

Dr. Hudson also suggested, based on elevational data and not field documentation, that another fifteen undocumented areas of this community may exist in the state. What Dr. Hudson failed to acknowledge in his testimony (but what was confirmed during cross-examination), is that eight of these possible but unconfirmed additional occurrences are smaller than Sisk (see Attachment A). No more than six of these additional areas would meet MNAP's size threshold of 750 acres for an A rank (e.g., exemplary), and none would meet the landscape context threshold of being surrounded by at least 1,000 acres of undisturbed habitat. In addition, some of the larger ones suggested by Dr. Hudson have been adversely impacted by recent logging activities. Extensive clearcutting above 3,000 feet on White Cap and Caribou Mountain is evident in 10-year-old satellite imagery, and Dr. Vickery's presentation at the May 12 hearing showed recent clearcutting above 3,000 feet on Cow Ridge. Thus it is likely that very few of the fifteen potential additional occurrences of this community listed by Dr. Hudson would qualify for an overall 'A' rank. Dr. Hudson's testimony does not change Sisk's relative position in the middle of the size scale or diminish the fact that it was rated as a good quality occurrence by MNAP and described as undisturbed and natural.

To Dr. Hudson's third point, the map included as Exhibit B with his pre-filed testimony shows the extent of impact on this community from the proposed Kibby Expansion project. It indicates that nearly 30% of the extent of this community would be eliminated or indirectly impacted by the project. This cannot be considered a "minimal" impact by any reasonable standard. Furthermore, his estimate is conservatively low. The estimate for "Habitat area adjacent to project footprint" is based on a 50' wide buffer around the footprint, and is a measure of the extent of the remaining part of the community subject to edge effects.

While MNAP uses the same buffer in their estimate of indirect effects, under cross-examination during the May 12 hearing Molly Docherty of MNAP agreed that this estimate was most likely conservative. A variety of other sources indicate that indirect edge effects are more extensive:

- Maine's Beginning With Habitat program uses a minimum buffer of 250' around development in its delineation of Undeveloped Habitat Blocks[1]. This buffer is used around primary gravel loggings roads, even if there is no development along them (though no buffer is placed around smaller logging roads). For example, this buffer has been placed around the Greenville (KI) Road in the eastern part of Greenville[2], even though this road has a narrower travel surface and less extensive cut and fill than the proposed summit road and has no development along it.
- The Unfragmented Forest Block layer developed for New Hampshire's Wildlife Action Plan by The Nature Conservancy uses a 100 meter (330 feet) buffer along primary gravel forest roads.
- The Interior Forest data layer developed for Massachusetts' Comprehensive Wildlife Conservation Strategy also uses a 100 meter buffer along minor roads.

From these sources it is clear that conservation professionals (including state wildlife agencies) consider a buffer of 250 to 330 feet to be more appropriate to separate interior from edge forest. Use of this more widely accepted measure of edge would significantly change the

---

[1] See http://www.beginningwithhabitat.org/the_maps/map3-undev_habitat.html.
[2] See http://megis2.dafs.maine.gov/ifwpdf/Greenville/Map3.pdf.

assessment of how much of this community would be impacted by the project. And in this circumstance the use of a larger impact zone is justified, since the primary edge is created by the summit road extending for over a mile on the steep upper slope along the western edge of the community, enhancing the degrading impact of the prevailing strong westerly winds.

Based on Hudson's Exhibit B map, we estimate that there are about 6700' of project edge along the western and southern side of the northern (73 acre) contiguous remaining habitat block, and 2500' along the northern side of the southern (183 acre) contiguous block. If an edge impact zone of 150' were used (which is still conservative compared to the standard used in state wildlife action plans), an additional 21 acres of the rare subalpine community would be impacted. If a 250' zone were used (consistent with state wildlife action plans) an additional 42 acres would be impacted. This brings the total direct and indirect impact of the project to about 40% of the extent of the community, and reduces the size of the northern contiguous remaining block from 73 to 42 acres. This is a very significant level of impact on a significant rare natural resource, which we believe clearly rises to the level of an "undue adverse" impact.

Mr. Valleau's post-hearing filing only strengthens the case that this rare habitat is at risk, including from regulated logging activities. The occurrence of timber harvesting within this community does not justify further permanent destruction of this community type, in fact it makes it more imperative that remaining undisturbed occurrences (such as are found on Sisk) should be protected. Using past impacts in other areas to justify additional cumulative impact on this habitat would be a very slippery slope, and contrary to the intent of LURC's third principle goal to "*Maintain the natural character of certain areas within the jurisdiction having significant natural values…*" We also note that Mr. Valleau's summary description of the MBPL harvest permit (Exhibit E) as an example of a sanctioned impact to this community is ignores the critical context; this particular harvest application indicates that the area had been harvested about 20 years previously. Sisk is still an undisturbed and natural site. Furthermore, a timber harvest (which is a temporary impact) in a previously disturbed area is in no way comparable to the permanent loss, fragmentation and indirect impact of 40% of an undisturbed natural rare natural community area that this project would cause.

Finally, in regard to the current and future extent of this community and the significance of the occurrence at Sisk - because this important habitat faces risks from timber harvests and wind power development, it strengthens the reason why larger intact, undisturbed examples like

Sisk should be protected as an important and well-recognized part of the state's climate change strategy. Protecting habitats that will have an important role in allowing the region's species to adapt to future climate change is as much needed as is wind power. We include as attachments copies of three peer-reviewed papers cited in Dr. Publicover's pre-filed testimony on the potential decline in spruce-fir forest due to expected future warming and the greater resistance of high-elevation forests to these changes, which make these subalpine forests particularly valuable as potential refugia for now commercially important softwood species at lower elevations that could be dependent on this habitat for their regional survival in the near future.

- Attachment B: Tang, G. and B. Beckage. 2010. Projecting the distribution of forests in New England in response to climate change. Diversity and Distributions 16: 144-158.

  This paper uses climate and vegetation modeling to project that areas capable of supporting spruce-fir forest in Maine are likely to contract to the higher and cooler parts of the western mountain region by the late 21st century, even under a relatively conservative assumptions about increases in atmospheric CO_2.

- Attachment C: Spear, R.W. 1989. Late-Quaternary history of high-elevation vegetation in the White Mountains of New Hampshire. Ecological Monographs 59: 125-151.

  This paper demonstrated that the distribution of subalpine coniferous forest in the White Mountains has been very stable since their post-glacial establishment about 9,000 years ago, even in the face of significant climate shifts (including a much warmer period from 9,000 to 5,000 years ago) that led to large changes in low-elevation vegetation (including a significant decline in low-elevation spruce-fir forest).

- Attachment D: Seidel, T.M., D.M. Weihrauch, K.D. Kimball, A.A.P. Pszenny, R. Soboleski, E. Crete and G. Murray. 2010. Evidence of climate change declines with elevation based on temperature and snow records from 1930s to 2006 on Mount Washington, New Hampshire, USA. Arctic, Antarctic and Alpine Research 41: 362-372.

  This paper, using the best long-term and only high-elevation climate data set from New England, shows that indicators of climate warming declines with elevation and is statistically insignificant at the higher elevations in New England. This helps explain why the region's sub-alpine forest and alpine areas survived past warming trends and served as refugia for low elevation spruce-fir forest, a role they may need to again fulfill in the next century.

Dated June 1, 2010

VERIFICATION

Signature of Witness: David A. Publicover

June 1, 2010

Before me appeared David A. Publicover, who, being duly sworn, did testify that the foregoing testimony was true and correct to the best of his knowledge and belief.

State of New Hampshire
Coos County

Denise M. Horne, Notary Public
My Commission Expires
April 4, 2012

Denise M Horne
NOTARY PUBLIC

ATTACHMENT A

List of documented occurrences of Fir-Heartleaved Birch Subalpine Forest natural community in Maine (from Maine Natural Areas Program), amended to include potential undocumented occurrences listed by Dr. Hudson (shown in italics).

Subalpine Fir Forest records in Maine

Also known as Fir - heart-leaved birch subalpine forest

EO Rank	Survey Site	Counties	Last Observed	Acres	Acres Above 3000'(1)
A	MT KATAHDIN	Piscataquis	2004	18127	
A	Mahoosuc Range	Oxford	2008-09-18	8701	
A	The Bigelows	Franklin, Somerset	2005	3071	
?	*White Cap*				*2901*
A	REDINGTON POND RANGE	Franklin	2006-09-13	2400	
B	BAKER MOUNTAIN TO LILY BAY MOUNTAIN	Piscataquis	2007	2289	
?	*Boundary Peak*				*1629*
?	*Caribou*				*1605*
?	*Cow Ridge*				*1533*
B	BALDPATE MTN	Oxford	2005	1408	
?	*Unnamed (D-series)*				*1152*
?	*Snow*				*1040*
C	BARREN MOUNTAIN	Piscataquis	2007	890	
B	BIG SPENCER MOUNTAIN	Piscataquis	2006-08-10	871	
B	KIBBY MOUNTAIN	Franklin	2006-9-11	614	
?	*Tumbledown (northern Boundary Mountains)*				*492*
B	POPLAR RIDGE	Franklin	1996-09-18	365	
B	**Sisk Mountain**	**Franklin**	**2009**	**358**	
B	Black Nubble	Franklin	2007-07-25	316	
?	*Boil*				*272*
?	*Smart*				*215*
B	CENTER MTN	Piscataquis	2007	172	
?	*Pisgah*				*166*
E	Sugarloaf Mountain	Franklin	1996-08-21	152	
?	*Merrill*				*152*
?	*No. 6*				*139*
?	*Onion Hill*				*127*
B	Sabbathday Pond Area	Franklin	2008-06-18	97	
?	*No. 5*				*85*
C	MOUNT BLUE	Franklin	1999-12-03	52	
E	WHITECAP MOUNTAIN	Piscataquis	1996-10-01	40	
C	CHAIRBACK AREA COLUMBUS MOUNTAIN	Piscataquis	2007	36	
C	BIG SQUAW MOUNTAIN	Piscataquis	2006	35	
?	*Moose*				*5*

(1) Acreage determined from USGS 30-meter Digital Elevation Model data. This likely overestimates the size of the subalpine forest community occurrences. The undocumented occurrences total about 11,500 acres, whereas Dr. Hudson estimated their total extent at 8,000 acres. We note that Sisk Mountain has 557 acres above 3,000' by this assessment, whereas the community occurrence encompasses only 358 acres.

Diversity and Distributions, (Diversity Distrib.) (2010) **16**, 144–158

BIODIVERSITY RESEARCH

Projecting the distribution of forests in New England in response to climate change

Guoping Tang and Brian Beckage*

Department of Plant Biology, University of Vermont, Marsh Life Science Building, Burlington, VT 05405, USA

*Correspondence: Brian Beckage, Department of Plant Biology, University of Vermont, Marsh Life Science Building, Burlington, VT 05405, USA.
E-mail: brian.beckage@uvm.edu

ABSTRACT

Aim To project the distribution of three major forest types in the northeastern USA in response to expected climate change.

Location The New England region of the United States.

Methods We modelled the potential distribution of boreal conifer, northern deciduous hardwood and mixed oak–hickory forests using the process-based BIOME4 vegetation model parameterized for regional forests under historic and projected future climate conditions. Projections of future climate were derived from three general circulation models forced by three global warming scenarios that span the range of likely anthropogenic greenhouse gas emissions.

Results Annual temperature in New England is projected to increase by 2.2–3.3 °C by 2041–70 and by 3.0–5.2 °C by 2071–99 with corresponding increases in precipitation of 4.7–9.5% and 6.4–11.4%, respectively. We project that regional warming will result in the loss of 71–100% of boreal conifer forest in New England by the late 21st century. The range of mixed oak–hickory forests will shift northward by 1.0–2.1 latitudinal degrees (*c.* 100–200 km) and will increase in area by 149–431% by the end of the 21st century. Northern deciduous hardwoods are expected to decrease in area by 26% and move upslope by 76 m on average. The upslope movement of the northern deciduous hardwoods and the increase in oak–hickory forests coincide with an approximate 556 m upslope retreat of the boreal conifer forest by 2071–99. In our simulations, rising atmospheric CO_2 concentrations reduce the losses of boreal conifer forest in New England from expected losses based on climatic change alone.

Main conclusion Projected climate warming in the 21st century is likely to cause the extensive loss of boreal conifer forests, reduce the extent of northern hardwood deciduous forests, and result in large increases of mixed oak–hickory forest in New England.

Keywords
BIOME4, climate change, global circulation model, species shifts, tree distribution, vegetation model.

INTRODUCTION

Global climate is projected to warm by 1.8–6.4 °C this century relative to the 1980–99 mean global temperature in response to continued anthropogenic emissions of greenhouse gases (IPCC 2007; Bates *et al.*, 2008). While terrestrial vegetation is expected to shift poleward and to higher elevations in response to projected warming (Parmesan & Yohe, 2003; IPCC, 2007), more precise projections are necessary for planning for the conservation of biological diversity and increasing the resilience of forest ecosystems (e.g. Farnsworth & Ogurcak, 2006). Projections of forest response to global warming have been largely based on statistical models that map observed species distributions with respect to climate and then re-project these distributions under future climatic conditions (e.g. Iverson & Prasad, 2001; McKenny *et al.*, 2007). This modelling approach assumes a constant relationship between forest distribution and climate and does not account for the direct effects of CO_2

DOI:10.1111/j.1472-4642.2009.00628.x
www.blackwellpublishing.com/ddi

on plant performance. An alternative approach is to simulate plant distributions using process-based models (PBMs) (e.g. Kaplan *et al.*, 2003; Sitch *et al.*, 2003). PBMs simulate fundamental physiological processes that determine vegetation growth, e.g. plant photosynthesis and respiration, as mediated by climate and other environmental conditions, providing an alternative process-based method for projecting vegetation distribution in response to climate change. PBMs have been used to model the distribution of vegetation at the global scale (e.g. Sitch *et al.*, 2003), but this coarse resolution introduces substantial uncertainties in regional projections related to both the lack of detailed parameterization of regional plant functional types (PFTs) and the coarseness of the environmental data employed (e.g. Tang & Bartlein, 2008; Willis & Bhagwat, 2009). Studies that utilize PFTs parameterized for regional vegetation and that downscale general circulation model (GCM) projections to regional geographical scales can improve model projections of future vegetation distribution relative to global model runs.

In this study, we modelled the future distribution of forests across New England in response to projected climate change. Forests in New England are broadly characterized by boreal, northern hardwood and mixed oak–hickory forest types with distributions that have historically corresponded to climate gradients in addition to edaphic conditions and disturbance (e.g. Foster *et al.*, 1998, 2002; Parshall *et al.*, 2003). We used a process-based vegetation model BIOME4 (Kaplan *et al.*, 2003) to simulate the future distribution of these primary forest types under nine future climate change scenarios (CCS). Each CCS is based on three GCM runs, i.e. HadCM3 (Gordon *et al.*, 2000; Pope *et al.*, 2000), CGCM3.1 (Kim *et al.*, 2002, 2003) and ECHAM5 (Jungclaus *et al.*, 2005), driven by three IPCC (International Panel on Climate Change) SRES (Special Report on Emission Scenarios) storylines, i.e. B1, A1B and A2, that bracket the range of likely climate trajectories. Our objective was to determine the potential magnitude of spatial displacement of the three widespread forest types of New England by mid- and late 21st century. Our analysis provides the public and policy makers with additional information on potential future changes in vegetation distribution in New England.

METHODS

Vegetation

We modelled the distribution of three dominant forest types in the New England landscape: boreal conifers (e.g. *Abies balsamea*, *Picea mariana* and *Picea rubens*), northern deciduous hardwoods (e.g. *Acer saccharum*, *Fagus grandifolia* and *Betula alleghaniensis*) and mixed oak–hickory forests (e.g. *Quercus alba*, *Quercus velutina*, *Carya glabra* and *Carya ovata*) (Fuller *et al.*, 1998). The distribution of these forest types reflects climatic conditions, in addition to landscape disturbance, historical land use and soil condition (e.g. Foster *et al.*, 1998, 2002; Parshall *et al.*, 2003): Boreal conifer forests are currently widespread at higher elevations and in northern regions of New England, northern deciduous hardwoods are mainly distributed in the cooler central uplands, and mixed oak–hickory forests are found at lower elevations and more southerly regions. These forest types are generally associated with late successional stages in forest development (McLachlan *et al.*, 2000; Woods, 2000). While earlier successional stages may be present in some regions recovering from natural or anthropogenic disturbance (Fuller *et al.*, 1998), we do not evaluate the effects of historic land use in this study, and removed from this analysis all land-cover types that are currently subject to substantial human use. Although the current relationship between climate and vegetation distribution in New England is partly obscured by human activities and ecological succession following land abandonment (Hall *et al.*, 2002; Parshall *et al.*, 2003), historical evidence points to a strong relationship between climate and vegetation distribution: Rising temperatures in 14,600 yr BP coincided with increases in spruce populations following deglaciation, and subsequent warm and dry conditions after 11,600 yr BP corresponded to the replacement of spruce by pine populations (Shuman *et al.*, 2004). Climate was also a likely driver of the mid-Holocene decline of eastern hemlock as well as in changes in abundance of oaks and chestnut in New England (Shuman *et al.*, 2004; Foster *et al.*, 2006). Recent shifts of northern hardwoods to higher elevations formerly occupied by boreal forests have also been attributed to regional warming over the last century (Beckage *et al.*, 2008). Climate-driven PBMs thus provide a sound basis for projecting vegetation responses to future climate change.

Model

We modelled the distribution of boreal conifers, northern deciduous hardwoods and mixed oak–hickory forests in New England using BIOME4 (version 2b1) (Kaplan *et al.*, 2003). BIOME4 is a physiological, process-based vegetation model that simulates the equilibrium distribution of terrestrial vegetation in response to climate, soils and atmospheric CO_2 concentration. BIOME4, in its unmodified form, simulates global vegetation as mixtures of 13 PFTs. A PFT is defined as a group of plants with similar traits and environmental requirements as defined by physiological and environmental parameters within BIOME4. Biomes are then defined by mixtures of PFTs. Although BIOME4 was designed as a global vegetation model, it has also been successfully modified to simulate the response of regional vegetation to climate change in a number of studies (e.g. Diffenbaugh *et al.*, 2003; Song *et al.*, 2005; Kaplan & New, 2006; Ravindranath *et al.*, 2006) in a similar manner to our study. We have defined three PFTs that correspond to vegetation types – boreal conifer, northern deciduous hardwood and mixed oak–hickory forests – rather than to individual species, because our objective was to project general shifts in forest types and the species within these forest types have somewhat similar physiological and bioclimatic attributes.

Abies balsamea, *Picea mariana* and *Picea rubens*, for example, are all shade-tolerant species with low to medium drought tolerance (Burns & Honkala, 1990). PFT-related parameters were based on previously published literature and were further defined by analysing the climatic features of major species' distribution of each PFT using tree range maps (available at http://esp.cr.usgs.gov/data/atlas/little/) in combination with climate data from our baseline period, described below. Our main PFT parameters are given in Table S1.

We used BIOME4 to model the current (i.e. for the baseline period of 1961–90) and future (i.e. 2041–70 and 2071–99) distributions of forest types across New England under several emission scenarios. In addition, we tested the sensitivity of vegetation predictions for the 2041–70 period to changes in either precipitation or atmospheric CO_2 concentration: We alternatively held precipitation or atmospheric CO_2 concentration to baseline levels while allowing other climatic metrics to vary with climate projections. For the fixed-precipitation experiment, we kept monthly precipitation in the 2041–70 period the same as in the baseline simulation while allowing other input data to reach projected levels. For the fixed-CO_2 experiment, we held atmospheric CO_2 concentrations at the baseline simulation level (333 p.p.m.) while allowing other input data to reach levels projected for 2041–70. Simulations from these two experiments were then compared to simulations that used all projected data for the 2041–70 period, respectively.

Model data

We created climatologies for running BIOME4 using both PRISM (Daly *et al.*, 2000, 2002) and CRU CL 2.0 (New *et al.*, 2002) data sets with a 1961–90 baseline period (hereafter referred to as the 1976 period) for calculation of 'current' climatology. We chose this baseline period because (1) mean-monthly sunshine data required for BIOME4 are not available for other periods (e.g. 1971–2000), and (2) the 30 years of climate record for 1961–90 is immediately prior to the time of vegetation observations (1992–93) used to test our model. A 30-year climate window has been shown to be effective for simulating vegetation response to climate (Tang *et al.*, 2009). Annual atmospheric CO_2 concentration for the baseline period simulation was set at 333 p.p.m. (Schlesinger & Malyshev, 2001). The CRU CL 2.0 data set is on a 10 arc-minute global land grid while the PRISM data set used in this study is at a 2.5 arc-minute scale. We extracted the monthly percentage sunshine (%) from the CRU CL 2.0 data set but derived mean-monthly temperature and precipitation from the PRISM data set. We interpolated all climate metrics to a 30 arc-second resolution using the Shuttle Radar Topography Mission (SRTM) 30 arc-second near-globe digital elevation data (Farr & Kobrick, 2000; Rosen *et al.*, 2000). Soil data were derived from soil survey data for New England, obtained from the United States Department of Agriculture, NRCS Natural Resources Conservation Service (http://soildatamart.nrcs.usda.gov). We detail our method of downscaling below.

We downscaled the 2.5 arc-minute elevation PRISM data to a 30 arc-second resolution using bilinear interpolation. We first fitted a regression model to the PRISM 2.5 arc-minute data that treats climatic value at each grid cell as a function of its latitude, longitude and elevation to estimate the local lapse rates of temperature and precipitation. The calculated local lapse rates were then used to interpolate the PRISM data to a finer 30 arc-second resolution by considering the elevation differences between PRISM points and targets from the SRTM 30 arc-second elevation data. These adjusted climatic values for PRISM points were bilinearly interpolated to obtain the value of a climate variable at a target point. The CRU sunshine data were downscaled by bilinear interpolation using the same approach.

We derived nine future CCSs for New England from HadCM3, CGCM3.1 and ECHAM5 model runs driven by SRES storylines B1, A1B and A2 for the 21st century (Nakicenovic *et al.*, 2000). These scenarios describe future potential economic and societal trajectories that result in different levels of greenhouse gas emissions and associated climate change (IPCC, 2007). The B1 scenario is characterized by environmental and social consciousness, sustainable development, and low energy use. Global population rises to 9 billion by 2050 before declining to 7 billion by 2100. The same population growth trends not only characterize the A1B storyline but also include rapid economic development, which reduces differences between industrialized and developing regions, and very high energy use that comes from both carbon and non-carbon emitting sources. The A2 scenario reflects large population growth (i.e. 15 billion people by 2100), slow technological change, continued disparity between industrialized and developing portions of the world, and high energy use. The A1B results in medium levels of GHG emissions compared to relatively higher levels in the A2 and relatively lower levels in the B1 storylines.

Future climate normals of mean-monthly temperature and precipitation were calculated using each of three GCMs and storylines relative to simulated climate normals for 1961–90. These simulated normals for 1961–90 were subtracted from future simulated climates, resulting in projected change (anomalies) in climatic conditions. These monthly series of anomalies (for temperature) or ratios (for precipitation) were bilinearly interpolated onto the SRTM 30 arc-second grid (from ≥ 1.875° by 1.875°), then added to the downscaled baseline mean-monthly climatologies of climate variables derived from the PRISM data set. We calculated projected 30-year mean-monthly climatologies for two periods: 2041–70 (referred to as 2055 hereafter) and 2071–99 (referred to as 2085 hereafter). The climatologies for these two periods were used to project the future distribution of forests in New England using BIOME4. Future climate normals of mean-monthly percentage sunshine data were derived from GCM simulations of monthly percentage cloud-cover (%) based on historical regression coefficients between two climate variables. The atmospheric CO_2 concentrations under the B1, A1B and A2 storyline were set at 487, 544 and 549 p.p.m. for the 2055, and 568, 657 and 724 p.p.m. for the 2085 periods, respectively (Nakicenovic *et al.*, 2000).

Model tests

We tested our BIOME4 simulations by comparing the projected baseline vegetation to the 1992 National Land Cover Data (NLCD 1992, http://landcover.usgs.gov/us_map.php). The NLCD 1992 data were derived from Landsat Thematic Mapper satellite data at 30-m spatial resolution and classified into 21 land-cover types for the United States (Kelly & White, 1993; Vogelmann *et al.*, 1998a,b). We adjusted the scale of the 30-m NLCD 1992 for New England to coincide with our model grid cells at a 30-arc-second spatial resolution. The land-cover type in each regridded cell was defined as the modal land-cover type in a 30 × 30 grid cell window. We did not use other satellite-based data such as the 1-km global land-cover characteristic data (Loveland *et al.*, 2000) or the 1-km global land-cover classification data (Hansen *et al.*, 2000) because the land-cover classifications and finer resolution of the NLCD 1992 data were more suitable for validating modelled vegetation for New England.

We excluded land-cover types either dominated by human activities such as pasture, crops, residential and urban, or having low spatial coverage, including areas dominated by wetlands, shrubs and grasslands. We also did not use mixed forest, e.g. areas dominated by trees where neither deciduous nor evergreen species represent more than 75% of the cover present, to test model results because each grid cell in our simulation was assigned a single PFT, corresponding to the PFT with the highest net primary production (NPP) for that cell. We used a simplified set of two vegetative cover classifications, i.e. deciduous forest and evergreen forest, to test model results. Deciduous forest was classified as an area where 75% or more of the tree species shed foliage simultaneously in response to seasonal change, while the evergreen forest type was where 75% or more of the tree species maintain their leaves all year. The mismatch between simulated vegetation classes and the simplified two tree-cover classifications from NLCD 1992 forced us to combine two of our modelled categories, i.e. mixed oak–hickory and northern deciduous hardwoods into one category of ‘deciduous forest’ for comparison. In summary, our boreal conifer PFT corresponds to the evergreen classification, and our combination of mixed oak–hickory and northern deciduous hardwoods correspond to the deciduous classification. The use of these broader forest classifications could result in an overestimation of our model accuracy.

We assessed the efficacy of the model predictions by evaluating (1) the model’s accuracy or probability of assignment to correct forest type, μ_f, i.e. the probability that predicted vegetation corresponds to the classification in the NLCD 1992 data set, (2) the producer’s accuracy or the probability of assignment to incorrect forest type, ϕ_f, that refers to the probability that the NLCD 1992 data will be correctly simulated by BIOME4, and (3) the Kappa statistic, a scalar that summarizes the goodness-of-fit while accounting for chance agreement. We evaluated these metrics of model fit by first constructing a two dimensional error matrix **F** corresponding to the observed and predicted coverages for the boreal and deciduous forest cover types. For each forest type *f* in **F**, the model’s accuracy and the producer’s accuracy are given by:

$$\begin{cases} \mu_f = (n^f_{\text{pred}} \cap n^f_{\text{obs}}) \Big/ n^f_{\text{pred}} \\ \varphi_f = (n^f_{\text{pred}} \cap n^f_{\text{obs}}) \Big/ n^f_{\text{obs}} \end{cases}$$

where n^f_{pred} and n^f_{obs} are the number of predicted and observed cells of vegetation type *f*, respectively. The overall accuracy (μ) of model prediction across forest types is expressed as:

$$\mu = \sum_{\mathbf{F}} \left(n^f_{\text{pred}} \cap n^f_{\text{obs}} \right) \Big/ \sum_{\mathbf{F}} \left(n^f_{\text{obs}} \right).$$

The overall Kappa statistic (κ) between two compared maps is given by:

$$\kappa = (\mu - \sum_{\mathbf{F}} \mu_f \varphi_f)/(1 - \sum_{\mathbf{F}} \mu_f \varphi_f).$$

Values of the Kappa statistic > 0.75 indicate very good-to-excellent agreement, values between 0.40 and 0.75 indicate fair-to-good agreement, and values of 0.40 or less indicate poor agreement (Landis & Koch, 1977; Monserud & Leemans, 1992).

BIOME4 simulates potential natural vegetation whereas vegetation across much of New England has been directly influenced by human activities. We therefore also used comparatively undisturbed subregions of New England to test the performance of our model. The three subregions used for model tests were White Mountain National Forest in New Hampshire (bounding box: 71.812° to 71.041° W and from 43.894° to 44.347° N), Acadia National Park in Maine (bounding box: 69.159° to 68.01° W and from 44.007° to 44.498° N), and north-west Maine (bounding box: 70.308° to 68.592° W and from 45.950° to 47.484° N). The geographical delineation of the subregions above was based on two considerations: (1) a region was large enough to contain a significant area of at least two simulated PFTs, and (2) the region was relatively unpopulated and thus comparatively free of recent anthropogenic disturbance.

In addition to validating our model projections against NLCD 1992, we compared simulated leaf area index (LAI) and simulated annual NPP in each of our PFTs with measured or reported data for similar forests. We calculated the mean, minimum and maximum of simulated LAI and annual NPP in each PFT and compared these values to corresponding field measurements as an additional test of the model’s ability to simulate vegetation for New England. Such comparisons offer an additional avenue for assessing the goodness-of-fit of modelled vegetation when other forest cover data are not available.

RESULTS

Projected climate change

Temperature is projected to increase in the 21st century under all of our CCSs (Fig. 1). Annual mean temperature in New England is projected to increase by 2.2 to 3.3 °C by 2055 and from 3.0 to 5.2 °C by 2085, relative to the 1961–90 baseline annual mean temperature (5.9 °C) across emission scenarios for all three GCMs (Fig. 1a–c). Average increases in annual mean temperature by 2055 across all three GCMs were 2.4 °C for the B1, 3.2 °C for the A1B, and 2.9 °C for the A2 storylines (Fig. 2a,e,f). For 2085, the average increases were 3.2 °C for the B1, 4.4 °C for the A1B and 4.8 °C for the A2 storylines (Fig. 2b,f,j). The annual rates of temperature increase were 0.02, 0.03 and 0.03 °C year^{-1} under emission storylines B1, A1B and A2 respectively for the HadCM3 GCM. Projected warming ranged from 2.2 °C under the ECHAM5 B1 scenario to 5.2 °C under the CGCM3.1 A2 scenario and was relatively uniform across New England (Fig. 2a–j).

Annual precipitation in New England is expected to increase by 4.7–9.5% by 2055 and by 6.4–11.4% by 2085 (Table 1), but trajectories of precipitation change are more variable across years and scenarios than for temperature (Fig. 1). Annual precipitation increases consistently over the years 2003–99 for the A1B and, to a lesser degree, the A2 storylines, but tends to decrease after 2060 for the B1 storyline (Fig. 1e,f). The most consistent increases in precipitation occurred in the ECHAM5 GCM, with estimated annual precipitation increases of 1.0, 1.0 and 0.95 mm per year for the B1, A1B and A2 scenarios, respectively (Fig. 1e). Changes in annual precipitation are also spatially more variable across New England than for temperature (Fig. 2). For example, the magnitude of increase in 2055 is lower (< 6.2%) in southern New England and higher (> 6.2%) in middle and northern New England under the B1 scenario (Fig. 2c). The greatest increase in annual precipitation occurs in northern New England (> 10.6%) under most scenarios (e.g. Fig. 2g,h,k,l).

Model tests

The overall vegetation patterns simulated by BIOME4 agree well with those in the NLCD 1992 data (Fig. 3a vs. b). The model's overall accuracy (μ) in predicting vegetation across New England was 0.77 with an overall Kappa statistic (κ) of 0.49 (Table 2), indicating that the BIOME4 is 'fair to good' at simulating vegetation for New England (Monserud & Leemans, 1992). For example, both the simulated vegetation and the

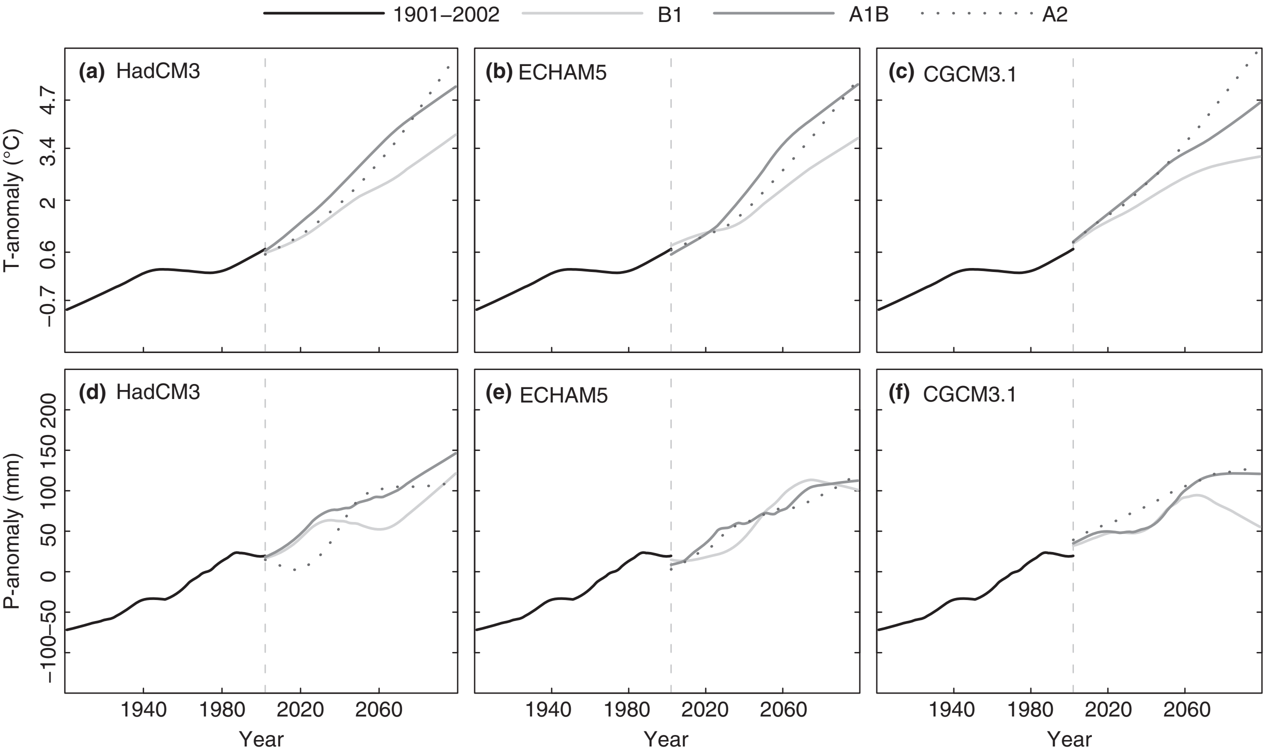

Figure 1 Projected annual mean temperature and precipitation over the years 2003–99 based on three GCMs: HadCM3 (a,d), ECHAM5 (b,e) and CGCM3.1 (c,f). Observed annual mean temperature and precipitation based on PRISM data are to the left of the dashed vertical line (the year 2002) in each panel, and to the right are the model projections under the B1, A1B and A2 storylines, respectively. The discontinuities between observed and simulated trajectories for some panels result because (1) the projected future changes are relative to 30-year mean climatologies for 1961–90 rather than that of the closest period (e.g. 1971–2000 or 1973–2002), and (2) a bias of the given GCM projections relative to actual observations.

Figure 2 The spatial patterns of projected change in annual mean temperature and precipitation for the periods 2041–70 and 2071–99 with reference to the baseline (1961–90) climatology under the SRES B1, A1B and A2 storylines. The data shown here for each storyline are means over all three GCMs forced by the given storyline.

NLCD 1992 data show that southeastern and northwestern Maine (Fig. 3) are dominated by boreal forest. The model's ability to simulate both the boreal conifer forest (as 'evergreen' for model test) and the northern deciduous hardwoods (as 'deciduous' for model test) in Maine is reflected by the model's high accuracy ($\mu_f > 0.74$) and the high overall accuracy ($\mu = 0.80$) and the 'fair to good' overall Kappa statistic ($\kappa = 0.60$) for this region (Table 2). In addition, the simulated mixed oak–hickory forest (as 'deciduous' for model test) in Connecticut and southern Massachusetts coincides with the deciduous forest type in the NLCD 1992 data, resulting in the high model's accuracy ($\mu_f = 0.85$) in simulating the deciduous forest for New England (Table 2). The BIOME4-simulated boreal conifer forest in Vermont and New Hampshire is mainly distributed in mountainous areas such as in Green Mountain National Forest and the White Mountain National Forest, agreeing well with the NLCD 1992 data (Fig. 3).

In addition to capturing general vegetation patterns across New England, the model was also able to accurately simulate vegetation with low human land use. The model's accuracy (μ_f) in simulating boreal conifer forest and northern deciduous hardwoods in White Mountain National Forest (area A in Fig. 3) was as high as 0.86 (Table 2), suggesting concordance of the modelled spatial patterns with that specified in the NLCD 1992 data (Fig. 3a vs. b) and consistent with the high overall accuracy ($\mu = 0.79$) and the 'fair to good' overall Kappa statistic ($\kappa = 0.58$) (Table 2) for this region. Similarly, the BIOME4-simulated vegetation distribution for the Acadia National Park (area B in Fig. 3) agrees well with that classified in the NLCD 1992 data as illustrated by the high overall accuracy ($\mu = 0.74$). Although the model's accuracy is relatively low ($\mu_f = 0.20$) in simulating the deciduous hardwoods in Acadia National Park, the number of grid cells dominated by this forest type accounts for only 15% of the total number of grid cells (1946).

We do note, however, that at finer scales the modelled vegetation may not reproduce the spatial patterns and texture apparent in the NLCD 1992 data. For example, the modelled boreal conifer forest in southeastern Maine and the northern New Hampshire is broader and more continuous than that

Table 1 Projected changes in annual mean temperature and total annual precipitation in New England for 2041–70 and 2071–99 compared to the baseline period 1961–90.

		HadCM3		ECHAM5		CGCM3.1	
Change in annual	Scenario	2055*	2085†	2055*	2085†	2055*	2085†
Temperature (°C)	B1	2.3	3.2	2.2	3.2	2.6	3.0
	A1B	3.2	4.6	3.2	4.6	3.2	4.0
	A2	2.7	4.8	2.6	4.3	3.3	5.2
Precipitation (%)	B1	4.7	8.0	7.9	9.9	8.0	6.4
	A1B	8.0	10.5	6.9	8.6	7.8	11.2
	A2	9.2	9.6	7.5	8.8	9.5	11.4

The 30-year (1961–90) mean annual temperature and precipitation in New England is 5.9 °C and 1109 mm.
*Refers to the period 2041–70.
†Refers to the period 2071–99.

Table 2 The accuracy assessment of our model simulations of forests in New England.

Region	Vegetation*	Model's accuracy (μ_f)	Producer's accuracy (φ_f)	Overall accuracy (μ)	Overall Kappa statistic (κ)
North-western Maine	Evergreen	0.87	0.74	0.80	0.60
	Deciduous	0.74	0.86		
White Mountain NF	Evergreen	0.86	0.70	0.79	0.58
	Deciduous	0.74	0.88		
Acadia NP	Evergreen	0.86	0.82	0.74	0.06
	Deciduous	0.20	0.25		
New England	Evergreen	0.62	0.72	0.77	0.49
	Deciduous	0.85	0.80		

NF, national forest; NP, national park.
*For comparison with NLCD 1992 data, we considered both the northern deciduous hardwoods and the mixed oak–hickory forest as deciduous forest cover type, and the boreal conifer forest as evergreen forest cover type.

classified in the NLCD 1992 data (Fig. 3a vs. b). This difference results, in part, from the fragmented nature of vegetation in New England due to human activities but which BIOME4 does not explicitly consider, so that the modelled vegetation tends to be more continuous than that classified in the NLCD 1992 data. In addition, the climate data used to run BIOME4 were derived from the 2.5 arc-minute PRISM and the 10 arc-minute CRU CL 2.0 data, and these data might not capture enough climatic variation at 30 arc-second spatial resolution across New England to simulate fine-scale vegetation patterns. In contrast, the NLCD 1992 data were derived from satellite images at 30-m spatial resolution, which should better capture the spatial variation of vegetation at a finer scale than our model simulations. Finally, because we combined northern

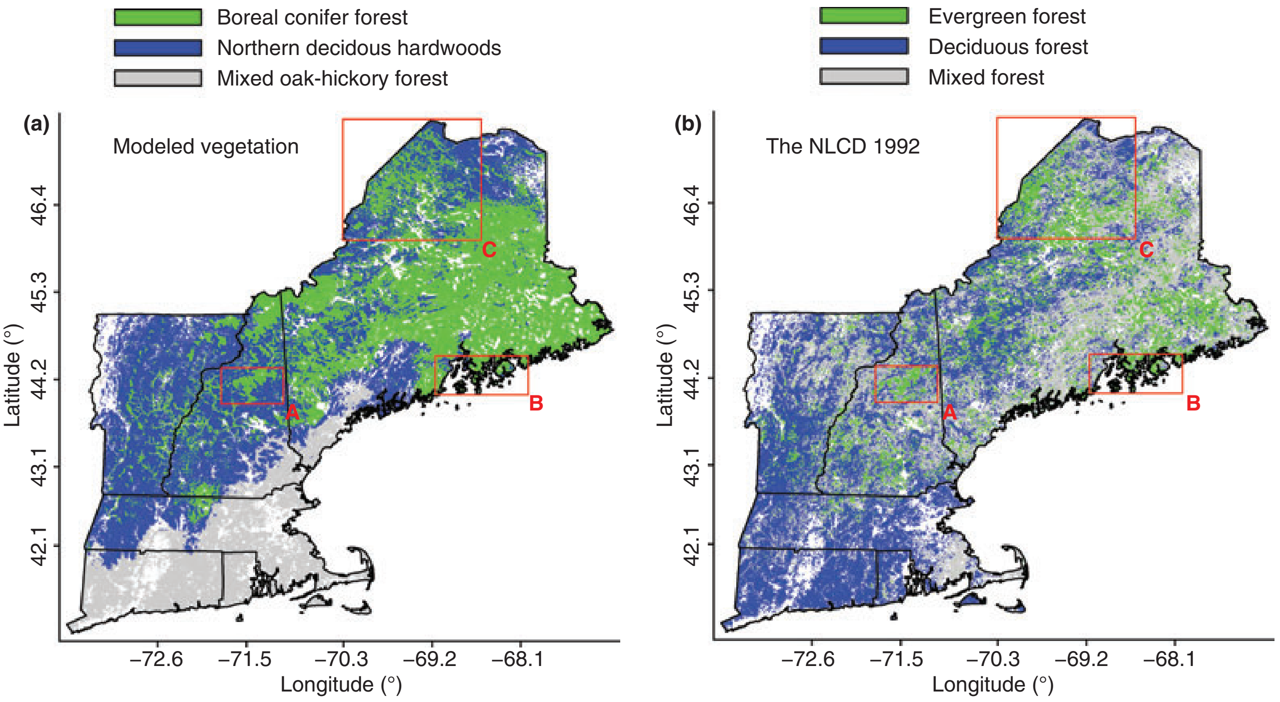

Figure 3 Comparison between (a) the modelled vegetation for the period 1961–90 and (b) the land-cover classification in the 1992 National Land Cover Data. Areas A, B and C are three subregions, i.e. the White Mountain National Forest, the Acadia National Park and north-western Maine, used to test the model simulation. White areas in (a) and (b) are lakes, land cover of limited extent and human use land covers that are excluded from model tests.

Table 3 Test of BIOME4-simulated LAI and NPP for forests in New England.

		Mixed oak–hickory forest		Northern deciduous hardwoods		Boreal conifer forest	
		Model	Observed	Model	Observed	Model	Observed
LAI (m^2 m^{-2})	Mean	3.34	3.84	4.04	3.38	3.09	2.99
	Min.	2.76	2.90	3.26	0.36	2.51	0.48
	Max.	4.07	4.50	4.51	7.30	3.48	7.40
NPP (g m^{-2} $year^{-1}$)	Mean	835	810	678	695	633	644
	Min.	696	660	542	199	322	440
	Max.	930	1010	819	999	816	914

Observed leaf area index (LAI) and net primary production (NPP) data are from Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, Tennessee, USA. (http://www.daac.ornl.gov). We used measured LAI in similar forests (see Table S2) to test modelled LAI for New England. Gridded NPP (Zheng et al., 2003) for forests dominated by boreal conifers was used to test modelled NPP in boreal conifer forest, and NPP for temperate deciduous broadleaf forest dominated by upland oaks was used to compare modelled NPP in mixed oak–hickory forests. Because of the lack of observed NPP data for northern deciduous hardwoods, we used NPP for forests dominated by aspen in Superior National Forest of Minnesota (USA) to test modelled NPP for northern deciduous hardwoods.

deciduous hardwoods and mixed oak–hickory forest together as deciduous forest in the model test, simulations of their specific distributions were not tested by the NLCD 1992.

The modelled optimum LAI for our three PFTs agree well with field observations in similar forests in other regions of the USA. Our modelled LAI averaged 3.34 in mixed oak–hickory forest, 4.04 in northern deciduous hardwoods, and 3.09 in boreal conifer forest, close (difference < 20%) to average observations of 3.84, 3.38 and 2.99 in similar forests respectively (Table 3). The magnitudes of modelled LAI in each grid cell (Fig. S1a) are within the ranges of LAI observations, as illustrated by the minimum and maximum LAI in each PFT from both simulation and observations (Table 3). Burrows *et al.* (2002), for example, reported a mean LAI of 3.45 in northern hardwoods in Park Falls, Wisconsin (USA) in July of 1999 based on eddy flux measurements.

BIOME4's ability to simulate vegetation for New England was also supported by the consistency of modelled optimum annual NPP with field measurements. The modelled mean annual NPP is 835 g m^{-2} $year^{-1}$ in oak–hickory forest, 678 g m^{-2} $year^{-1}$ in northern deciduous hardwoods, and 633 g m^{-2} $year^{-1}$ in boreal conifer forests, closely approximating (within 3%) average NPP of 810, 695 and 644 g m^{-2} $year^{-1}$ measured in similar forests, respectively (Table 3). The magnitudes of modelled NPP at a grid cell level (Fig. S1b) are also within the ranges of observed values (Table 3). Our simulated values are also supported by specific studies of forest NPP in the eastern USA: Brown & Schroeder (1999), for example, reported that annual NPP in eastern hardwoods ranged from 750 to 1150 g m^{-2} $year^{-1}$, with an area-weighted average of 970 g m^{-2} $year^{-1}$. In contrast, annual NPP in softwoods ranged from 580 to 980 g m^{-2} $year^{-1}$, with an area-weighted average of 870 g m^{-2} $year^{-1}$.

Model projections

Our model simulations of future forest distribution indicate a general shift from boreal conifers and northern deciduous

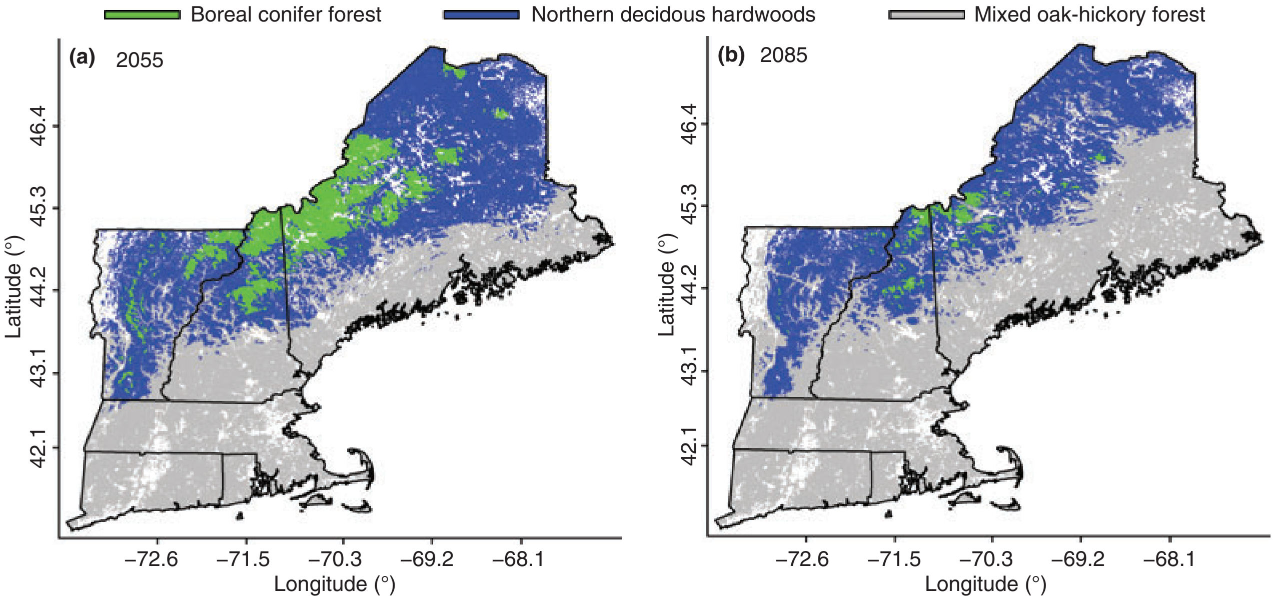

Figure 4 The distribution of mixed oak–hickory, northern deciduous hardwood and boreal conifer forests in two future periods 2041–70 (referred to as 2055) and 2071–99 (referred to as 2085) in New England. The vegetation type in each grid cell is based on the modal value of each grid cell across all nine climate changes scenarios.

hardwoods to mixed oak–hickory forest (Fig. 4). The mixed oak–hickory forest in southern New England is expected to move northward and increase in area through the mid and late 21st century under all CCS (Fig. 3a vs. Fig. 4; Table 4). We estimate, for example, that the northern boundary of the mixed oak–hickory forest will migrate northward by 0.7 latitudinal degrees (*c.* 75 km) by 2055 (Fig. 3a vs. Fig. 4a) and 1 latitudinal degrees (*c.* 101 km) by 2085 (Fig. 3a vs. Fig. 4b; HadCM3 B1 in Table 4). The corresponding southern boundary of the northern deciduous hardwoods is expected to shift northward, e.g. the latitudinal distribution of northern deciduous hardwoods is projected to increase by 0.6 latitudinal degrees (*c.* 67 km) by 2055 and by 0.8 latitudinal degrees (*c.* 90 km) by 2085 (Fig. 3a vs. Fig. 4h) under the ECHAM5 A2 scenario (Table 4). Under all scenarios, boreal conifer forest (e.g. Fig. 4d,j,p) is projected to contract to mountain ranges and to the region centred on the corner of northern New Hampshire and north-western Maine by 2085, because annual temperature in these areas tends to be the lowest across New England (Fig. S2a). The contraction of boreal conifer forest to higher elevations in mountain ranges can result in an apparent southerly shift in latitudinal range under most scenarios in 2085 (Table 4), i.e. when northern lowlands lose their boreal conifer forest.

Projected climate change is estimated to shift both the northern deciduous hardwoods and the mixed oak–hickory forest to higher elevations (Table 5). The average elevation of northern deciduous hardwoods is 279 m a.s.l. in the baseline simulation under current conditions, but increases by 52 m and 76 m by 2055 and 2085, respectively, averaged across all GCMs and scenarios. The average elevation is also projected to increase for the boreal conifer forest by 295 m and 556 m by 2055 and 2085, respectively (Table 5). The increase in apparent elevation of boreal forest, however, was primarily driven by losses of this community at lower elevations rather than a general shift to higher elevations, as boreal forests already occupy the highest elevations in our region (> 800 m a.s.l.; NLCD 1992).

Table 4 Projected latitudinal shifts of simulated forest types in New England.

		Mixed oak–hickory forest				Northern deciduous hardwoods				Boreal conifer forest			
		2055		2085		2055		2085		2055		2085	
GCM	Scenario	°	km	°	km	°	km	°	km	°	km	°	km
H3	B1	0.7	75	1.0	101	0.5	55	0.8	85	0.3	32	−0.1	−12
	A1B	0.9	95	1.4	147	0.8	86	0.9	98	−0.1	−13	−0.2	−16
	A2	0.7	77	1.6	165	0.7	74	1.0	104	0	3	−0.3	−30
	B1	0.7	79	1.4	145	0.2	26	0.9	99	0.5	53	0	0
E5	A1B	1.3	135	1.8	196	0.9	97	0.8	85	0	2	−0.1	−12
	A2	1.0	111	1.6	170	0.6	67	0.8	90	0.4	44	−0.1	−12
	B1	1.2	130	1.6	167	0.5	53	1.0	102	0.5	55	0	2
CG	A1B	1.5	160	1.8	192	1.1	118	0.8	86	0.1	10	−0.1	−15
	A2	1.4	148	2.1	219	1.0	105	0.6	60	0	2	0	1
Average shift		1.0	112	1.6	167	0.7	76	0.8	90	0.2	21	−0.1	−10

H3, HadCM3; E5, ECHAM5; CG, CGCM3.1.

Contraction of ranges upslope can cause some negative latitudinal shifts to occur (i.e. vegetation moves southward). This is the case for boreal conifers in high latitudes of New England, for example, where this forest type is expected to contract upslope into mountain ranges that can be at lower latitudes than northern New England. The projected latitudinal shifts are based on the average position of the PFTs in two future periods relative to the baseline period (1961–90).

Table 5 Projected altitudinal shifts of simulated forest types in New England.

		Mixed oak–hickory forest		Northern deciduous hardwoods		Boreal conifer forest	
GCM	Scenarios	2055 (m)	2085 (m)	2055 (m)	2085 (m)	2055 (m)	2085 (m)
HadCM3	B1	40	45	−75	−20	253	433
	A1B	47	56	−27	67	430	723
	A2	42	73	−51	93	364	874
ECHAM5	B1	47	54	−106	10	184	346
	A1B	53	98	−3	117	355	608
	A2	50	70	−58	67	219	539
CGCM3.1	B1	54	66	−34	31	194	321
	A1B	73	92	26	84	309	483
	A2	61	130	9	238	347	677
Average shift		52	76	−35	76	295	556

Changes in precipitation, temperature, and CO_2 can cause the expansion of northern deciduous hardwoods to lower elevations in some scenarios.

Projected climate change may cause a large portion of New England to be potentially dominated by mixed oak–hickory forest by the end of the 21st century (Fig. 4). The total number of grid cells dominated by mixed oak–hickory forest (37,261 of 215,509 of total simulated grid cells for 1976) increased under all scenarios; the magnitudes of the increases ranged from 99% under the HadCM3 B1 scenario to 276% under the CGCM3.1 A1B scenario by the 2055 period, and from 149% under the HadCM3 B1 scenario to 431% under the CGCM3.1 A2 scenario by the 2085 period (Table 6). New England is also expected to lose a large portion of the northern deciduous hardwoods and the boreal conifer forest (Fig. 3 vs. Fig. 4). Boreal conifer forests (89,634 grid cells under the baseline simulation) are expected to lose on average 61% of their areal extent in New England by 2055 and 91% by 2085 across all scenarios, while northern deciduous hardwoods (93,114 grid cells under the baseline simulations) are expected to lose 11% and 26% of their area by the 2055 and 2085 periods, respectively.

Our computational experiments, which examined the sensitivity of vegetation distribution to changes in CO_2 concentration, indicated that rising CO_2 concentration can reduce the losses of boreal conifer forests (Fig. 5). The spatial extent of the boreal conifer forest is reduced when CO_2 concentrations are held at baseline levels but with the same magnitude of climate change (Fig. 5a vs. b). When atmospheric CO_2 concentration

Table 6 Projected future changes in potential area of simulated forest types in New England.

		Mixed oak–hickory forest		Northern deciduous hardwoods		Boreal conifer forest	
GCM	Scenarios	2055 (%)	2085 (%)	2055 (%)	2085 (%)	2055 (%)	2085 (%)
HadCM3	B1	99	149	15	24	−57	−87
	A1B	143	237	26	0	−86	−99
	A2	110	282	32	−17	−79	−100
ECHAM5	B1	116	228	−23	−19	−25	−76
	A1B	210	345	−9	−45	−77	−97
	A2	173	272	−27	−18	−44	−95
CGCM3.1	B1	208	273	−51	−41	−34	−71
	A1B	276	330	−41	−44	−72	−92
	A2	240	431	−23	−77	−76	−99
Average change		175	282	−11	−26	−61	−91

The estimate percentage change is based on the number of grid cells for each forest type in the baseline (1961–90) and future (2041–70 or 2071–99) simulations.

was held constant at 333 p.p.m. in 2055 as in the baseline simulation, for example, climate change alone under the B1 storyline decreased the boreal conifer forest by 77%, 46% and 74% in the HadCM3, ECHAM4 and CGCM3.1 runs, respectively, which are greater losses than experienced under the same climate scenarios but with atmospheric CO_2 concentration increased to 487 p.p.m., e.g. 57%, 25% and 34% (Table 6).

Changes in summer precipitation are projected to either slow down (when summer precipitation increases) (Fig. 6b) or speed up (when summer precipitation decreases) the replacement of northern deciduous hardwoods by mixed oak–hickory forest (Fig. 6c). For example, the simulated extent of the northern deciduous hardwoods increased by 10% and 6% in 2055 under the HadCM3 and ECHAM5 B1 scenarios, where projected summer precipitation increased by 2–9% compared to the baseline precipitation (Fig. 6d,e). In contrast, the extent of the northern deciduous hardwoods decreased by 21% by 2055 under the CGCM3.1 B1 scenario, where projected summer precipitation decreased by 4–14% compared to the baseline precipitation (Fig. 6f). Changes in summer precipitation within a range of −10% to 10% relative to baseline precipitation, however, had minimal effect on boreal conifer forest.

DISCUSSION

We project that New England will lose the majority of its boreal conifer forest (91% averaged over scenarios) as well as some northern deciduous hardwoods (26% averaged over scenarios) in response to a projected 3.0–5.2 °C warming and 6.4–11.4% increase in annual precipitation by 2085. Mixed oak–hickory forest, in contrast, is projected to nearly triple in area (282% averaged over scenarios) in New England by the end of this century. We estimate that the northern deciduous hardwoods will shift northward by 0.8° latitude (*c.* 90 km) and by 76 m to higher elevations, while mixed oak–hickory forests will shift northward by 1.6° latitude (*c.* 167 km) and by 76 m to higher elevations (Tables 4 and 5). The corresponding

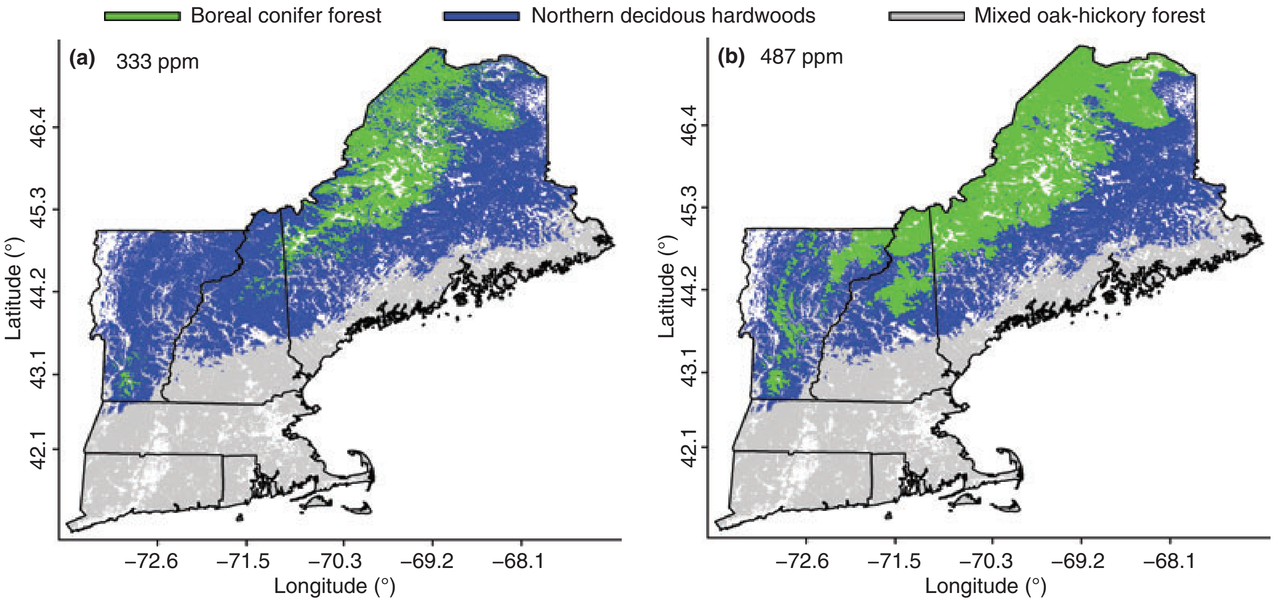

Figure 5 Increases in atmospheric CO_2 concentration promote the persistence of boreal conifer forest under projected climate warming scenarios. Panel (a) is modelled vegetation in 2055 under the B1 storyline with CO_2 concentration set at 333 p.p.m. Panel (b) is modelled vegetation in 2055 under the same storyline but with CO_2 concentration set at 487 p.p.m. The vegetation type in each grid cell is based on the modal value of each grid cell across all three GCMs.

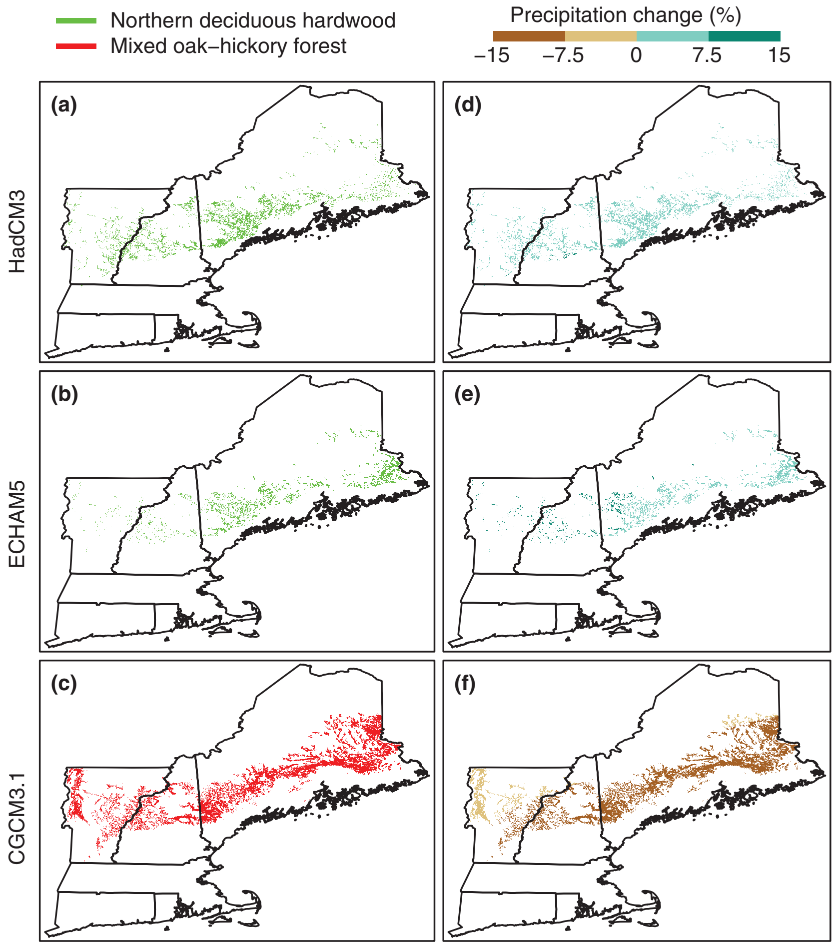

Figure 6 Sensitivity of simulated vegetation to changes in precipitation. An increase in summer precipitation of 5% and 7% results in northern deciduous forest rather than oak–hickory forest in areas of green for the HadCM3 (a) and ECHAM5 (b) GCMs. These changes in coverage represent a 6% and 10% increase in northern deciduous forest, respectively. A decrease in summer precipitation of 10%, in contrast, results in oak–hickory forest rather than the northern deciduous hardwoods in areas represented by red (c). This change represents a 40% increase in oak–hickory forest. Panels (d–f) show the changes in summer precipitation corresponding to the panels on the left. The data shown here are based on comparisons between simulations using projected precipitation for 2055 under the B1 storyline and simulations using the baseline precipitation. Temperature and CO_2 concentration in each pair of compared simulations are as same as those projected under the B1 storyline.

potential migration rates of 0.8–1.5 km per year are similar to rates calculated for biome shifts in other modelling studies (e.g. Malcolm *et al.*, 2002). Past migration rates of similar forests have been estimated to range from < 100 m year^{-1} (McLachlan *et al.*, 2005) to 250 m year^{-1} (Davis, 1989), suggesting that these forests may not be able to shift as rapidly as climate. Recent studies have already confirmed that climate warming in the 20th century has been associated with shifts of vegetation to both higher latitudes and elevations (e.g. Parmesan & Yohe, 2003; Beckage *et al.*, 2008), and increasing impacts are expected in the future (e.g. Thuiller *et al.*, 2005). These projections of vegetation shifts in New England are driven by projected regional climate change, and assume that climate is the major factor controlling the bioclimatic range limits of vegetation at regional scales (e.g. Dirnbock *et al.*, 2003) as the effects of land-use change, disturbance, etc. are not considered.

Although our model results project that the boreal conifer forest will move northward and contract to the northern New Hampshire and the northwestern Maine, climate change may not completely extirpate this forest type from New England this century. Previous studies based on regression tree analysis (e.g. Iverson & Prasad, 2001) projected the extirpation of spruce-fir forest types from New England under five CCS and doubled CO_2 concentrations. Our simulations based on BIOME4 and new GCM data driven by different storylines indicate that the boreal conifer forest may still persist in New England in the late 21st century under some scenarios but its distribution will contract to the ranges of mountains (see Fig. 4). The continued presence of boreal conifers in our simulations is likely because (1) the temperature in these scenarios does not increase enough (< 4.6 °C) to exceed the bioclimatic range limits of the boreal conifer species, and (2) the inclusion of the physiological effects of CO_2 on plant growth in BIOME4 offsets the negative effects of climate change on the boreal conifer forest (e.g. VEMAP Members., 1995; Lapola *et al.*, 2009). The greatest risks to the boreal conifer forest occur under the HadCM3 A1B (Fig. S3d) and A2 (Fig. S3f), the ECHAM5 A1B (Fig. S3j) and the CGCM3.1 A2 (Fig. S3r) scenarios, under which annual temperature is projected to increase by at least 4.6 °C.

Increasing atmospheric CO_2 concentration, which, of course, largely drives climate change, appears to reduce the negative effects of climate change on the distribution of the boreal conifer forest in New England (Fig. 5). Rising CO_2 concentrations can reduce plant transpiration by inducing the stomatal closure of plants that increases their water use efficiency (e.g. Claessens *et al.*, 2006), and thus causes higher rates of net canopy CO_2-fixation in relation to water loss

(Osborne *et al.*, 2000). Boreal conifer tree species are comparatively sensitive to drought (see USDA, NRCS 2009; The PLANTS Database, http://plants.usda.gov) and may gain more in NPP from elevated CO_2 than deciduous hardwoods (Tjoelker *et al.*, 1998). In our simulations, for example, annual mean NPP increased by 175 g m^{-2} $year^{-1}$ for boreal conifers and by 164 g m^{-2} $year^{-1}$ for deciduous hardwoods under high CO_2 (487 p.p.m.) simulations for 2055 compared to low CO_2 (333 p.p.m.) experiments. Differential responses to elevated atmospheric CO_2 could result in a relative competitive advantage of boreal conifers compared to deciduous hardwoods. This effect could result in increasing losses of boreal conifer forest in New England even if CO_2 levels were stabilized if global temperatures continued to rise in response to an accumulated thermal debt (e.g. Meehl *et al.*, 2005).

Previous studies have indicated the importance of changes in precipitation in addition to temperature in determining vegetation distribution (e.g. Dirnbock *et al.*, 2003; Rehfeldt *et al.*, 2006). Our study demonstrates that the effect of precipitation change, with a range of −10% to 10% relative to the baseline precipitation, depends on the target vegetation. For example, simulations that alternatively held precipitation at baseline or 2055 levels resulted in less than a 1% difference in the total number of grid cells projected to contain boreal conifer forest, while projected CO_2 concentrations and temperatures were allowed to change with the B1 storyline. In contrast, precipitation increases in summer reduce the replacement of northern deciduous hardwoods by mixed oak–hickory forest (Fig. 6a,b) while precipitation decreases in summer cause their further replacement (Fig. 6c). Northern deciduous hardwoods tend to be physiologically less drought-tolerant than the mixed oak–hickory forests (see USDA, NRCS 2009; The PLANTS Database, http://plants.usda.gov), so that changes in precipitation have the potential to shift the competitive balance between these two PFTs.

Although BIOME4 was developed to simulate equilibrial potential vegetation at global spatial scales, we were able to successfully use this model to simulate regional vegetation in a landscape that has a history of human activities and disturbance (Fuller *et al.*, 1998; Parshall *et al.*, 2003). The model's tests against both vegetation in relatively undisturbed regions and across the whole of New England demonstrate the potential of BIOME4 to simulate vegetation in New England (Table 2). Additional comparisons with two important ecological indicators (LAI and NPP) support the application of BIOME4 to forests in New England (Table 3). In contrast to niche-based, statistical models, we were able to account for the physiological effects of CO_2 on plant growth and vegetation distribution in BIOME4, showing that rising CO_2 can ameliorate increased water stress under elevated temperature thus affecting vegetation distribution (e.g. VEMAP Members, 1995).

We caution, however, that BIOME4 is an equilibrium vegetation model that assumes that vegetation is in equilibrium with climate and does not consider successional changes or transient states as the vegetation composition shifts. The rate at which vegetation responds to climate change depends on the time (or lag) required for vegetation to reach a new equilibrium in response to climate change. Our projections should therefore be viewed as the potential distribution of these forest types in New England under a given climate condition. In addition, BIOME4 assumes that climate is a major factor in determining vegetation distribution over a broad spatial scale. However, other factors, such as seed dispersal, local-scale disturbances and human activities, can be important factors controlling vegetation distribution in a given area, influencing the time for vegetation to reach an equilibrium with climate or even inhibiting the landscape from attaining its potential forest state. Finally, the aggregation of species into PFTs ignores the spectrum of species-specific migration rates and climatic tolerances, potentially hindering the accuracy of future projections and reducing the heterogeneity and complexity of spatial patterns of modelled vegetation distribution (e.g. Neilson *et al.*, 2005).

CONCLUSION

1. Annual mean temperature in New England is projected to increase by 2.2–3.3 °C in 2055 and by 3.0–5.2 °C in 2085 across emission scenarios for all three GCMs, compared to the 1961–90 annual mean temperature of 5.9 °C. Projected warming ranged from 2.2 °C under the ECHAM5 B1 scenario to 5.2 °C under the CGCM3.1 A2 scenario and was relatively uniform across New England. Annual total precipitation in New England is also expected to increase by 4.7–9.5% by 2055 and by 6.4–11.4% by 2085 under all scenarios, but increases in precipitation are more variable across years and scenarios than for temperature. Changes in annual precipitation are also spatially more variable across New England than for temperature. Compared to the baseline annual precipitation (1109 mm), the magnitude of increase is the lowest (< 6.2%) in southern New England under the B1 scenario and the highest (> 10.6%) in northwestern Maine under the A1B scenario.

2. The BIOME4-simulated vegetation pattern agrees well with the land cover in the NLCD 1992 data. When the model was tested against observed vegetation in the whole of New England, the overall Kappa statistic ($\kappa = 0.49$ indicating 'fair to good' fit) justifies BIOME4's application to New England even though BIOME4 was originally developed to simulate potential natural vegetation at the global scale. Comparisons with observed ecological indicators (LAI and NPP) in similar forests further justify the application of BIOME4 to New England.

3. Mixed oak–hickory forest in southern New England is projected to move north by 1.6 latitudinal degrees (*c.* 167 km) due to a regional warming of 3.0–5.2 °C by the end of this century. Projected future climate change is expected to shift both northern deciduous hardwoods and mixed oak–hickory forest upslope by 76 m by the end of the 21st century. The upslope movement of the northern deciduous hardwoods and oak–hickory forest coincides with an approximate 556 m upslope retreat, driven by losses of this forest type at lower elevations rather than colonization of higher elevations, of the boreal conifer forest by the end of the 21st century.

4. Projected climate change will result in reduced areas of the boreal conifer forest in New England this century. The magnitudes of losses of the boreal conifer forest range from 25% in 2055 under the ECHAM5 B1 scenario to 100% in 2085 under the HadCM3 A2 scenario. The extirpation of the boreal conifer forest from New England is most likely to occur in our simulations when annual mean temperature increases more than 4.6 °C. Projected climate change reduces the extent of northern deciduous hardwoods in most scenarios, but changes ranged from a 24% increase by 2055 under the HadCM3 B1 scenario to a 77% loss in 2085 under CGCM3.1 A2 scenario. Mixed oak–hickory forests, in contrast, are projected to increase by 149% in the HadCM3 B1 scenario to 431% in the CGCM3.1 A2 scenario by the end of the 21st century. Increases in atmospheric CO_2 concentration tend to reduce losses of boreal conifer forest in New England, while precipitation change influences the relative abundance of northern deciduous hardwoods and mixed oak–hickory forests.

ACKNOWLEDGEMENTS

This research was supported by the US Department of Energy's Office of Science (BER) through the Northeastern Regional Center of the National Institute for Climatic Change Research. We greatly thank Dr Patrick J. Bartlein for his help on downscaling climate data for this study. We greatly appreciate two anonymous referees for their valuable comments on this study. We thank Carrie Pucko for comments on previous drafts of this study, and Kimberley Kemper for discussions of New England vegetation.

REFERENCES

Bates, B.C., Kundzewicz, Z.W., Wu, S. & Palutikof, J.P. (eds) (2008) *Climate change and water*. Technical Paper of the Intergovernmental Panel on Climate Change. IPCC Secretariat, Geneva, 210 pp.

Beckage, B., Osborne, B., Gavin, D.G., Pucko, C., Siccama, T. & Perkins, T. (2008) A rapid upward shift of a forest ecotone during 40 years of warming in the Green Mountains of Vermont. *Proceedings of the National Academy of Sciences USA*, **105**, 4197–4202.

Brown, S.L. & Schroeder, P.E. (1999) Spatial patterns of aboveground production and mortality of woody biomass for eastern U.S. forests. *Ecological Application*, **9**, 968–980.

Burns, R.M. & Honkala, B.H. (1990) *Silvics of North America. Vol. 2. Conifers*. USDA, Washington, DC.

Burrows, S.N., Gower, S.T., Clayton, M.K., Mackay, D.S., Ahl, D.E., Norman, J.M. & Diak, G. (2002) Application of geostatistics to characterize LAI for flux towers to landscapes. *Ecosystems*, **5**, 667–679.

Claessens, L., Hopkinson, C., Rastetter, E. & Vallino, J. (2006) Effect of historical changes in land use and climate on the water budget of an urbanizing watershed. *Water Resources Research*, **42**, W03426. doi:10.1029/2005WR004131.

Daly, C., Taylor, G.H., Gibson, W.P., Parzybok, T.W., Johnson, G.L. & Pasteris, P. (2000) High-quality spatial climate data sets for the United States and beyond. *Transactions of the American Society of Agricultural Engineers*, **43**, 1957–1962.

Daly, C., Gibson, W.P., Taylor, G.H., Johnson, G.L. & Pasteris, P. (2002) A knowledge-based approach to the statistical mapping of climate. *Climate Research*, **22**, 99–113.

Davis, M.B. (1989) Lags in vegetation response to greenhouse warming. *Climatic Change*, **15**, 75–82.

Diffenbaugh, N.S., Sloan, L.C., Snyder, M.A., Bell, J.L., Kaplan, J., Shafer, S.L. & Bartlein, P.J. (2003) Vegetation sensitivity to global anthropogenic carbon dioxide emissions in a topographically complex region. *Global Biogeochemical Cycles*, **17**, 1067. doi:10.1029/2002GB001974.

Dirnbock, T., Dullinger, S. & Grabherr, G. (2003) A regional impact assessment of climate and land-use change on alpine vegetation. *Journal of Biogeography*, **30**, 401–417.

Farnsworth, E.J. & Ogurcak, D.E. (2006) Biogeography and decline of rare plants in New England: historical evidence and contemporary monitoring. *Ecological Applications*, **16**, 1327–1337.

Farr, T.G. & Kobrick, M. (2000) Shuttle Radar Topography Mission produces a wealth of data. *American Geophysical Union Eos*, **81**, 583–585.

Foster, D.R., Motzkin, G. & Slater, B. (1998) Land-use history as long-term broad-scale disturbance: regional forest dynamics in central New England. *Ecosystems*, **1**, 96–119.

Foster, D.R., Clayden, S., Orwig, D.A., Hall, B. & Barry, S. (2002) Oak, chestnut and fire: climatic and cultural controls of long-term forest dynamics in New England, USA. *Journal of Biogeography*, **29**, 1359–1379.

Foster, D.R., Oswald, W.W., Faison, E.K., Doughty, E.D. & Hansen, B.C.S. (2006) A climatic driver for abrupt mid-Holocene vegetation dynamics and the hemlock decline in New England. *Ecology*, **87**, 2959–2966.

Fuller, T.L., Foster, D.R., McLachlan, J.S. & Drake, N. (1998) Impact of human activity on regional forest composition and dynamics in central New England. *Ecosystems*, **1**, 76–95.

Gordon, C., Cooper, C., Senior, C.A., Banks, H.T., Gregory, J.M., Johns, T.C., Mitchell, J.F.B. & Wood, R.A. (2000) The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. *Climate Dynamics*, **16**, 147–168.

Hall, B., Motzkin, G., Foster, D.R., Syfert, M. & Burk, J. (2002) Three hundred years of forest and land-use change in Massachusetts, USA. *Journal of Biogeography*, **29**, 1319–1335.

Hansen, M., DeFries, R., Townshend, J.R.G. & Sohlberg, R. (2000) Global land cover classification at 1km resolution using a decision tree classifier. *International Journal of Remote Sensing*, **21**, 1331–1365.

IPCC (2007) Summary for policymakers. *Climate change 2007: impacts, adaptation and vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (ed. by M.L. Parry, O.F. Canziani, J.P. Palutikof, P.J. van der Linden and C.E. Hanson), pp. 7–22. Cambridge University Press, Cambridge, UK.

Iverson, L.R. & Prasad, A.M. (2001) Potential changes in tree species richness and forest community types following climate change. *Ecosystems*, **4**, 186–199.

Jungclaus, J.H., Botzet, M., Haak, H., Keenlyside, N., Luo, J.-J., Latif, M., Marotzke, J., Mikolajewicz, U. & Roeckner, E. (2005) Ocean circulation and tropical variability in the AOGCM ECHAM5/MPI-OM. *Journal of Climate*, **19**, 3952–3972.

Kaplan, J.O. & New, M. (2006) Arctic climate change with a 2 degrees C global warming: timing, climate patterns and vegetation change. *Climatic Change*, **79**, 213–241.

Kaplan, J.O., Bigelow, N.H., Prentice, I.C., Harrison, S.P., Bartlein, P.J., Christensen, T.R., Cramer, W., Matveyeva, N.V., McGuire, A.D., Murray, D.F., Razzhivin, V.Y., Smith, B., Walker, D.A., Anderson, P.M., Andrew, A.A., Brubaker, L.B., Edwards, M.E. & Lozhkin, A.V. (2003) Climate change and Arctic ecosystems II: modeling, paleodata-model comparisons, and future projections. *Journal of Geophysical Research*, **108**, 8171. doi:10.1029/2002JD002559.

Kelly, P.M. & White, J.M. (1993) Preprocessing remotely sensed data for efficient analysis and classification, Applications of Artificial Intelligence 1993: Knowledge-Based Systems in Aerospace and Industry. *Proceeding of SPIE*, **1993**, 24–30.

Kim, S.-J., Flato, G.M., Boer, G.J. & McFarlane, N.A. (2002) A coupled climate model simulation of the Last Glacial Maximum, Part 1: transient multi-decadal response. *Climate Dynamics*, **19**, 515–537.

Kim, S.-J., Flato, G.M. & Boer, G.J. (2003) A coupled climate model simulation of the Last Glacial Maximum, Part 2: approach to equilibrium. *Climate Dynamics*, **20**, 635–661.

Landis, J.R. & Koch, G.G. (1977) Application of hierarchical Kappa-type statistics in assessment of majority agreement among multiple observers. *Biometrics*, **33**, 363–374.

Lapola, D.M., Oyama, M.D. & Nobre, C.A. (2009) Exploring the range of climate biome projections for tropical South America: the role of CO2 fertilization and seasonality. *Global Biogeochemical Cycles*, **23**, GB3003. doi:10.1029/2008GB003357.

Loveland, T.R., Reed, B.C., Brown, J.F., Ohlend, D.O., Zhu, Z., Yang, L. & Merchant, J.W. (2000) Development of a global land cover characteristics database and IGBP DISCover from 1-km AVHRR data. *International Journal of Remote Sensing*, **21**, 1303–1313.

Malcolm, J.R., Adam, M., Ronald, P.N. & Michael, G. (2002) Estimated migration rates under scenarios of global climate change. *Journal of Biogeography*, **29**, 835–849.

McKenny, D.W., Pedlar, J.H., Lawrence, K., Campbell, K. & Hutchinson, M.F. (2007) Potential impacts of climate change on the distribution of North American trees. *BioScience*, **57**, 939–948.

McLachlan, J.S., Foster, D.R. & Menalled, F. (2000) Anthropogenic ties to late-successional structure and composition in four New England hemlock stands. *Ecology*, **81**, 717–733.

McLachlan, J.S., Clark, J.S. & Manos, P.S. (2005) Molecular indicators of tree migration capacity under rapid climate change. *Ecology*, **86**, 2088–2098.

Meehl, G.A., Washington, W.M., Collins, W.D., Arblaster, J.M., Hu, A.X., Buja, L.E., Strand, W.G. & Teng, H.Y. (2005) How much more global warming and sea level rise? *Science*, **307**, 1769–1772.

Monserud, R.A. & Leemans, R. (1992) Comparing global vegetation maps with the Kappa statistic. *Ecological Modelling*, **62**, 275–293.

Nakicenovic, N., Alcamo, J., Davis, G., *et al.* 2000. *Special Report on Emissions Scenarios: A special report of Working Group III of the Intergovernmental Panel on Climate Change: Other Information: PBD: 3 Oct 2000*. Cambridge University Press, New York; Pacific Northwest National Laboratory, Richland, WA, Environmental Molecular Sciences Laboratory (US).

Neilson, R.P., Pitelka, L.F., Louis, F., Solomon, A.M., Nathan, R., Midgley, G.F., Fragoso, J.M., Lischke, H. & Thompson, K. (2005) Forcasting regional to global plant migration in response to climate change. *BioScience*, **55**, 749–759.

New, M., Lister, D., Hulme, M. & Makin, I. (2002) A high-resolution data set of surface climate over global land areas. *Climate Research*, **21**, 1–25.

Osborne, C.P., Mitchell, P.L., Sheehy, J.E. & Woodward, F.I. (2000) Modelling the recent historical impacts of atmospheric CO_2 and climate change on Mediterranean vegetation. *Global Change Biology*, **6**, 445–458.

Parmesan, C. & Yohe, G. (2003) A globally coherent fingerprint of climate change impacts across natural systems. *Nature*, **421**, 37–42.

Parshall, T., Foster, D.R., Faison, E., Macdonald, D. & Hansen, B.C.S. (2003) Long-term history of vegetation and fire in pitch pine-oak forests on Cape Cod, Massachusetts. *Ecology*, **84**, 736–748.

Pope, V., Gallani, M.L., Rowntree, P.R. & Stratton, R.A. (2000) The impact of new physical parameterizations in the Hadley Centre climate model: HadAM3. *Climate Dynamics*, **16**, 123–146.

Ravindranath, N.H., Joshi, N.V., Sukumar, R. & Saxena, A. (2006) Impact of climate change on forests in India. *Current Science*, **90**, 354–361.

Rehfeldt, G.E., Crookston, N.L., Warwell, M.V. & Evans, J.S. (2006) Empirical analyses of plant-climate relationships for the western United States. *International Journal of Plant Sciences*, **167**, 1123–1150.

Rosen, P.A., Hensley, S., Joughin, I.R., Li, F.K., Madsen, S.N., Rodriguez, E. & Goldstein, R.M. (2000) Synthetic aperture radar interferometry. *Proceedings of the IEEE*, **88**, 333–382.

Schlesinger, M.E. & Malyshev, S. (2001) Changes in near-surface temperature and sea level for the Post-SRES CO2-stabilization scenarios. *Integrated Assessment*, **2**, 95–110.

Shuman, B., Newby, P., Huang, Y. & Thompson, W.E.B.B., III (2004) Evidence for the close climatic control of New England vegetation history. *Ecology*, **85**, 1297–1310.

Sitch, S., Smith, B., Prentice, I.C., Arneth, A., Bondeau, A., Cramer, W., Kaplan, J.O., Levis, S., Lucht, W., Sykes, M.T., Thonicke, K. & Venevsky, S. (2003) Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ Dynamic Global Vegetation Model. *Global Change Biology*, **9**, 161–185.

Song, M., Zhou, C. & Hua, O. (2005) Simulated distribution of vegetation types in response to climate change on the Tibetan Plateau. *Journal of Vegetation Science*, **16**, 341–350.

Tang, G. & Bartlein, P.J. (2008) Simulating the climatic effects on vegetation: approaches, issues and challenges. *Progress in Physical Geography*, **32**, 543–556.

Tang, G., Shafer, S.L., Bartlein, P.J. & Holman, J.L. (2009) Effects of experimental protocol on global vegetation model accuracy: a comparison of simulated and observed vegetation patterns for Asia. *Ecological Modelling*, **220**, 1481–1491. doi:10.1016/j.ecolmodel.2009.03.021.

Thuiller, W., Lavorel, S., Araujo, M.B., Sykes, M.T. & Prentice, I.C. (2005) Climate change threats to plant diversity in Europe. *Proceedings of the National Academy of Sciences USA*, **102**, 8245–8250.

Tjoelker, M.G., Oleksyn, J. & Reich, P.B. (1998) Seedlings of five boreal tree species differ in acclimation of net photosynthesis to elevated CO2 and temperature. *Tree Physiology*, **18**, 715–726.

VEMAP Members (1995) Vegetation/ecosystem modeling and analysis project: comparing biogeography and biogeochemistry models in a continental-scale study of terrestrial ecosystem responses to climate change and CO2 doubling. *Global Biogeochemical Cycles*, **9**, 407–437.

Vogelmann, J.E., Sohl, T. & Howard, S.M. (1998a) Regional characterization of land cover using multiple sources of data. *Photogrammetric Engineering & Remote Sensing*, **64**, 45–47.

Vogelmann, J.E., Sohl, T., Campbell, P.V. & Shaw, D.M. (1998b) Regional land cover characterization using Landsat Thematic Mapper data and ancillary data sources. *Environmental Monitoring and Assessment*, **51**, 415–428.

Willis, K.J. & Bhagwat, S.A. (2009) Biodiversity and climate change. *Science*, **326**, 806–807.

Woods, K.D. (2000) Dynamics in late-successional hemlock-hardwoods forests over three decades. *Ecology*, **81**, 11–126.

Zheng, D.L., Prince, S.D. & Wright, R. (2003) *NPP Multi-Biome: gridded estimates for selected regions worldwide, 1989–2001i, R1. Data set*. Oak Ridge National Laboratory Distributed Active Archive Center, Oak Ridge, TN, USA. Available at: http://www.daac.ornl.gov.

SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Figure S1 The simulated baseline (1961–90) (a) optimum leaf area index and (b) annual net primary productivity (NPP) in New England.

Figure S2 (a) The baseline (1961–90) annual mean temperature and (b) annual total precipitation in New England downscaled from the PRISM 2.5 arc-minute data.

Figure S3 The distribution of mixed oak–hickory forests, northern deciduous hardwood, and boreal conifer forests in two future periods 2041–70 (referred to as 2055) and 2071–99 (referred to as 2085) in New England under nine climate changes scenarios.

Table S1 Parameters used to define three plant functional types used in our simulations for New England.

Table S2 List of observed LAI data for comparisons with modelled LAI.

BIOSKETCHES

Guoping Tang is a postdoctoral research associate in the Department of Plant Biology at the University of Vermont. He is a geographer interested in vegetation modelling, climate change and climate impact assessment, and geo-spatial data analysis and visualization.

Brian Beckage is an associate professor in the Department of Plant Biology at the University of Vermont. His research investigates the structure and dynamics of ecological communities, including the effects of global climate change.

Author contributions: G.T. and B.B. conceived the ideas; G.T. collected the data and ran the models; G.T. and B.B. shared in the writing.

Editor: David Richardson

Ecological Monographs, 59(2), 1989, pp. 125–151

LATE-QUATERNARY HISTORY OF HIGH-ELEVATION VEGETATION IN THE WHITE MOUNTAINS OF NEW HAMPSHIRE[1]

RAY W. SPEAR
Fulton-Montgomery Community College, Route 67, Johnstown, New York 12095 USA

Abstract. The pollen and plant-macrofossil records from four small lakes in the subalpine and alpine zone of the White Mountains, New Hampshire, give a 13 000-yr paleoenvironmental history. The White Mountains were deglaciated before 13 000 yr BP. Downwasting of the continental ice sheet was rapid. The summits projected above the ice as nunataks for only a brief period of time. Residual ice may have existed in Franconia Notch until 11 000 yr BP. From 13 000 to 11 750 yr BP a barren periglacial desert covered the highest altitudes in the White Mountains. Tundra vegetation occupied the lower slopes and valleys. The mean annual temperature was roughly 5°–10°C colder than today.

Sparse tundra vegetation surrounded all four high-elevation sites from 11 750 to 10 300 yr ago and several taxa, particularly *Artemisia* and Caryophyllaceae, indicate disturbance. The summits were subjected to intense periglacial activity. The mean annual temperature was 4–6° lower than present. By 10 300 yr BP shrubs such as willow, juniper, and dwarf birch had invaded the tundra at Lake of the Clouds. Spruce woodland dominated the lower slopes and valleys.

At 10 300 yr BP spruce populations arrived at high-elevation sites. Macrofossils of fir, birch, and shrubs also occur in sediments of this age. The temperature increased to or exceeded modern levels. Tree species did not reach the Franconia Notch sites until 9750 yr BP. At these sites the establishment of subalpine forests spanned a much shorter time period. Forests with poplar, spruce, and birch replaced the spruce woodlands of low elevations.

Subalpine fir forests became well established by 9000 yr BP. Evidence from the alpine site shows that the fir trees were more abundant and treeline higher than today from 10 300 to 5000 yr BP. After 5000 yr BP, the pollen percentages of alpine indicators increased and the numbers of fir macrofossils dropped. Of the three sites in subalpine fir firest, only the lowest shows any evidence of a warmer interval in the early Holocene. Treeline is apparently a poor temperature indicator because wind and moisture are the major factors determining its position. Taxa of the Northern Hardwood Forest (e.g., white pine, hemlock, yellow birch, sugar maple, and beech) arrived at lower elevations by 6500 yr BP, but the zones of modern vegetation became established only after 2000 yr BP when spruce populations expanded at low elevations between 750 and 1200 m.

Key words: alpine; late-glacial; paleoecology; palynology; plant macrofossils; treeline; White Mountains.

INTRODUCTION

Over the last 12 000 yr the forests of eastern North America have undergone dramatic changes. The broad patterns of change have been demonstrated by mapping pollen data (Davis 1976, 1981*b*, 1983, Bernabo and Webb 1977, Webb et al. 1983, Gaudreau and Webb 1985, Webb 1986). The dynamics of regional and local vegetation change during this period are less well understood. Floristic diversity, the readjustment of range limits following glaciation, climatic variability and change, disturbance, disease, and biotic interactions all play a role in vegetation dynamics. Identifying how these factors interact to determine the structure and function of forest communities is difficult. Davis and Botkin (1985) used a forest growth model to simulate the response of forest in central New England to climatic change. Detailed paleoecological studies can also be designed to answer specific questions on the history of the forests and vegetation (e.g., Brubaker 1975, Jacobson 1979). I chose to study the history of vegetation at high elevations (above 1200 m) in the White Mountains of New Hampshire. Previous work (Likens and Davis 1975, Davis et al. 1980) established the regional forest history and demonstrated changes in the composition and distribution of plant communities along the steep climatic gradients of the moun-

---

[1] Manuscript received 9 January 1987; revised 11 March 1988; accepted 18 March 1988; final version received 3 June 1988.

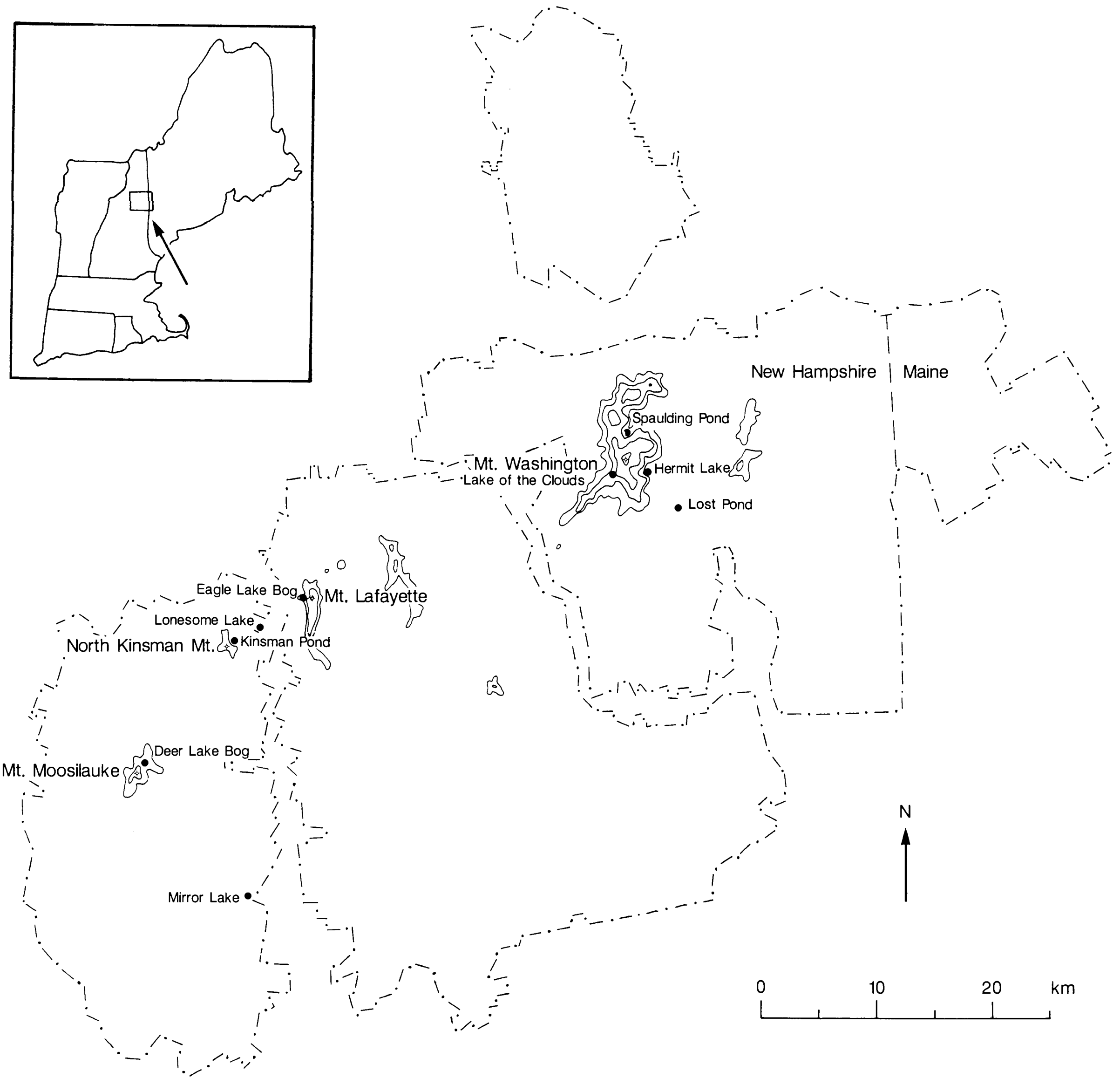

FIG. 1. Map of the White Mountain National Forest. The subalpine and alpine areas are denoted by the 1200-, 1400-, and 1600-m contour intervals. Bodies of water are shown as solid black dots.

tain slopes. I decided to do a detailed study on high-elevation vegetation because its fossil record might yield insights into vegetation dynamics unavailable from the study of lowland vegetation. High-elevation vegetation is a clearly defined unit of limited extent, has low floristic diversity (especially of tree species) and presumably has been in existence for 10 000 yr or more.

This study has three objectives: (1) to reconstruct the temporal pattern of deglaciation and plant colonization of the White Mountains of New Hampshire; (2) to determine when the modern subalpine fir forests and alpine plant communities developed from previous vegetation and from the lowland vegetation 14 000–9000 yr ago; and (3) to make inferences about climate and biotic interactions from the vegetation record. To meet these objectives I analyzed pollen and macrofossils preserved in lake (bog) sediments from four sites. The sites are located above 1140 m within an area of 2500 km^2 (Fig. 1). Three sites are in the subalpine fir forest and one site at 1542 m is in an alpine meadow. One problem with lakes at high elevations is that they accumulate most of their pollen from low-elevation forests (Davis et al. 1980, Spear 1981, M. B. Davis, *personal communication*). Plant macrofossils that come primarily from within 100 m of the basins (Dunwiddie 1987) provide a record of the local high-elevation vegetation. Comparison of the fossil records among my sites and with mid- and low-elevation sites (Davis et al. 1980, Davis 1983, and M. B. Davis, *personal communication*) provide much information on the vegetation dynamics in the subalpine and alpine zones and also provide a record of climate changes since glaciation.

STUDY AREA

Geology

The White Mountains are part of the mountainous terrain extending from northeastern Vermont through northern New Hampshire and southern Quebec to Mt. Katahdin in Maine (Fenneman 1938). Few peaks in the White Mountain National Forest are above 1600 m in elevation, but the relief is as much as 1150 m (Fig. 1). The bedrock of the White Mountain region consists largely of resistant gneiss, schist, and quartzite of the Littleton Formation (Devonian) (Billings et al. 1946). Both montane and continental glaciation shaped the White Mountains during the Pleistocene. Montane glaciers from 10 well-developed cirques in the Presidential Range preceded Late Wisconsin continental glaciation (Goldthwait 1970).

As the ice sheet thinned and retreated northward ≈14 000 yr ago, the high peaks were exposed as nunataks, with irregular tongues of ice persisting in the valleys (Goldthwait 1940, Goldthwait and Mickelson 1982). Deglaciation was followed by a period of intense frost activity. Continuous permafrost was present at an elevation of 1500 m, at least 300 m below its limit today (Goldthwait 1976). This period left periglacial features such as the frost-shattered bedrock and debris, stone nets, stripes and lobes, and rock glaciers in the alpine zone (Antevs 1932, Goldthwait 1940, Harries 1966). Some modern frost action can be found in the alpine zone, evidenced by solifluction features including miniature stone nets and stripes, and turf-banked and stone-banked terraces and garlands (Antevs 1932, Goldthwait 1940, and Harries 1966).

Climate

On the summit of Mt. Washington at 1917 m, the mean annual temperature is −2.8°C (NOAA 1975*a*). Frost can occur in any month of the year, but it is rare in July and August. Seven months have a mean temperature below 0°, and no month has a mean temperature above 10°. The lower portion of the alpine zone has a much milder climate than the summit; temperature and insolation are higher and precipitation and winds are lower (Bliss 1964, Harries 1966). Most of the climatic differences are due to altitude, but prevailing westerly winds and the dissipation of clouds passing over the summit also account for differences between the western and eastern slopes. Climatic data for the subalpine fir forest are summarized by Reiners and Lang (1979). For 7 mo of the year the mean temperature is below 0°, whereas for 4 mo it is >10°. The number of frost-free days is <90. At Pinkham Notch (613 m) the mean annual temperature is 4.4° (NOAA 1975*b*). Temperature data from treeline indicate that the lapse rate from Pinkham Notch to the summit is not a constant 0.53°/100 m, but steeper in the alpine zone than in the forested elevations (Harries 1966).

Precipitation in the White Mountains is distributed uniformly throughout the year. Total annual precipitation on the summit of Mt. Washington is 193.5 cm; at Pinkham Notch it is 146.7 cm. The elevations at which subalpine fir forests grow receive roughly 175 cm of precipitation a year (Reiners and Lang 1979). Fog drip and rime ice significantly increase this figure (Lovett et al. 1982). On the summit, 241 d of the year are cloudy from sunrise to sunset, whereas at Pinkham Notch only 167 d per year are cloudy. Heavy fog persists for some part of the day on 310 d per year on the summit but on only 24 d per year at Pinkham Notch. Intermediate elevations are in the clouds 30–50% of the time (Reiners and Lang 1979).

The alpine zone is subjected to extreme wind. The maximum wind speed recorded on the summit was 372 km/h in April 1934. Winds >161 km/h occur every month of the year. The mean annual speed on the summit is 56.6 km/h. The mean annual wind speed at treeline is 31.3 km/h, less than half that at the summit (Harries 1966). Once below treeline the wind speed drops dramatically. Average wind speed is 2.91 km/h at the 1158-m station in subalpine forest on Camels Hump, Vermont (Siccama 1974). At Pinkham Notch mean annual wind speed is 5.3 km/h, and there are fewer than 12 d/yr when it exceeds 40 km/h (Harries 1966).

Treeline appears to be correlated with an abrupt change in thermal lapse rate and wind speed. Both increase in the alpine zone. Macroclimate ultimately determines the position of treeline, but vegetation structure plays an important role in determining microclimate above and below treeline.

Vegetation

In the White Mountains today, the subalpine forests and alpine tundra above 1200 m are similar to boreal and arctic vegetation in eastern Canada. Montane coniferous forests of spruce and fir are found in the Appalachians from Tennessee and North Carolina northward into New England. Oosting and Billings (1951) denoted them as an Appalachian segregate of the spruce–fir boreal forest, comparing spruce–fir forests in the White Mountains of New Hampshire with those in the Great Smoky Mountains of Tennessee. Floristically, these Appalachian forests are nearly identical. Red spruce (*Picea rubens*) is important throughout the region, while Fraser fir (*Abies fraseri*) occurs in the south, and the closely related balsam fir (*Abies balsamea*) grows in the north (Myers and Bormann 1963). Except in the Catskill Mountains of New York (McIntosh and Hurley 1964), pure stands of fir occur at the upper elevations of the Appalachian coniferous forest.

Alpine tundra is found on the mountains of New York, New England, and the Gaspé Peninsula in southern Quebec. The presence of a diverse arctic–alpine flora on mountains in the Gulf of St. Lawrence led Fernald (1925) to theorize that arctic–alpine plants survived Wisconsinan glaciation on nunataks in this region. Fernald's theory has been disputed by glacial geologists who claim that the region was buried by ice (Ives 1974), and by botanists who have discovered new fossil localities of Fernald's "disjunct" species on glaciated terrain in eastern North America (Drury 1969, Miller and Thompson 1979). However, arctic–alpine species still could have persisted in ice-free or coastal sites. Farther south, the mountains in New England and New York were indisputably overridden by the Laurentide Ice Sheet (Goldthwait 1940, 1970). The members of the arctic–alpine floras found on these summits today must have migrated from the south as the ice retreated.

Zonation

Five vegetation zones have been recognized in the White Mountains: (1) northern hardwoods (hemlock phase); (2) northern hardwoods (spruce phase); (3) subalpine (spruce–fir phase); (4) subalpine (fir phase); and (5) alpine (Reiners and Lang 1979, M. B. Davis, *personal communication*). The first zone is called mixed hardwoods by Davis et al. (*unpublished manuscript*), and its extends up to ≈450 m in elevation. White pine (*Pinus strobus*), red oak (*Quercus rubra*), and hemlock (*Tsuga canadensis*), are limited to this zone. The northern hardwoods (spruce phase) occurs from 450 to 750 m and it includes the upper altitudinal limits of three species: yellow birch (*Betula lutea*), sugar maple (*Acer saccharum*), and beech (*Fagus grandifolia*). Spruce–fir forests are found between 750 and 1220 m. Above 1220 m red spruce becomes rare and balsam fir dominates the forest. Fir and black spruce (*Picea mariana*) krummholz are found in the lower part of the alpine zone, but upright trees seldom occur beyond 1500 m.

Subalpine fir forest 1200–1450 m

Many peaks in the White Mountains are high enough to support subalpine forest (Fig. 1). The subalpine forests are species poor. Fir makes up 99% of the tree layer, with a few trees of paper birch (*Betula papyrifera*), red and black spruce, and mountain ash (*Pyrus americana*). The diversity of shrubs and herbs is also low, but the forests are rich in bryophytes (Sprugel 1976, Reiners and Lang 1979, Sprugel and Bormann 1981). Bedrock and soils do not at present influence the distribution of this forest type (Flaccus 1959, Sprugel 1976, Reiners and Lang 1979, Spear 1981). Reiners and Lang (1979) and Sprugel (1976) show that the structure of the subalpine fir forest is related to wind. Reiners and Lang (1979) determined that subalpine fir forests display decreased canopy height, lower density, and more even-aged structure with increasing altitude because of increased wind exposure at higher elevations.

Treeline

In this paper, trees are defined as those individuals of a tree species >2 m tall (Hustich 1953). I use the following commonly accepted definitions: treeline is the limit of trees; forest limit is the boundary above which trees dominate <50% of the landscape (in New Hampshire treeline coincides with the forest limit); species limit is the elevation at which a species can no longer exist in any form (Hustich 1953, 1966, Wardle 1974, Black and Bliss 1978, Elliott 1979, Tranquillini 1979). Treeline in the Presidentials depends primarily on change in temperature with altitude (macroclimate). The species limit of black spruce krummholz at 1680 m (Antevs 1932) is correlated with a mean July temperature of 10° (Daubenmire 1954). Treeline also varies with steepness of slope, soil depth, snow cover, and exposure to the prevailing wind. On the west slopes of the Presidentials, treeline is between 1465 and 1525 m, but is as low as 1312 m on exposed northwest slopes. Treeline is higher on east slopes, usually occurring between 1525 and 1588 m, but does not span any mountain pass lying above 1400 m. Wind exposure is one of the most important factors in determining the exact position of treeline on the peaks (Monahan 1931, Griggs 1942, 1946, Bliss 1963, Harries 1966, Tiffney 1972, Reiners and Lang 1979). Wind damages trees in the winter by desiccation and reduces productivity by lowering the ambient temperature during the summer months (Marchand and Chabot 1978).

Alpine meadow above 1450 m

The Presidentials have the most extensive alpine zone, ≈20 km² (13.5 × 1–2 km) (Bliss 1963). The highest peak in the range is Mt. Washington at 1917 m, and six other peaks exceed 1500 m. The Franconia Range has a much smaller alpine area, 0.5 × 3.5 km, than the Presidential massif. The highest peak is Mt. Lafayette at 1603 m. Mt. Moosilauke is 1459 m high with an alpine zone of <10 ha.

Bliss (1963) identified nine plant communities in the alpine zone of the Presidentials: (1) sedge meadow, (2) sedge/dwarf-shrub heath, (3) sedge/rush/dwarf-shrub heath, (4) dwarf-shrub heath/rush, (5) dwarf-shrub heath, (6) *Diapensia,* (7) snowbank, (8) streamside, and (9) bog. Two environmental gradients are important in the distribution of alpine communities: (1) a gradient of increasing moisture and fog and decreasing temperature (from dwarf-shrub heath to sedge meadow), related directly to increasing altitude; and (2) a gradient of snow depth and snow melt (from snowmelt to *Diapensia*), inversely related to wind exposure (Bliss 1963, Harries 1966). Neither bedrock nor soil type is important in determining the distribution of alpine plant communities (Bliss 1963, Harries 1966). Rather, the development of soil is determined by the alpine vegetation cover.

METHODS

Sites

The four sites selected for pollen and macrofossil analyses (Fig. 1) occupy basins scoured in bedrock by glaciation. They have no inflowing streams and only small outlet streams.

Kinsman Pond (44°08′ N, 71°44′ W) is the largest (1.9 ha) and lowest of the sites at 1140 m. Its watershed is small, approximately eight times the size of the pond, and rises 170 m in 0.25 km to the summit of North Kinsman Mt. on the pond's west shore. The pond is surrounded by subalpine forest. The forest canopy is ≈10 m high. Balsam fir is the dominant tree in this forest. Heart-leaved paper birch is an important component, with low proportions of red spruce. Kinsman Pond and its watershed are underlain by Kinsman quartz monzonite (Williams and Billings 1938).

Two basins occur in a notch at 1275 m on the west flank of Mt. Lafayette (44°10′ N, 71°40′ W). The smaller basin (0.14 ha) ≈50 m northwest of Eagle Lake has a small pool of water surrounded by peat (Eagle Lake Bog). The watershed is small, <10 times the size of the site, and has gradual slopes rising only 10 m above the bog surface. Like Kinsman Pond, 6.5 km to the southeast across Franconia Notch, Eagle Lake Bog is underlain by Kinsman quartz monzonite (Williams and Billings 1938). Eagle Lake Bog is 110 m below treeline, surrounded by subalpine forest trees barely 2 m high, with stunted and flagged trees common on exposed ridges near the bog. Fir and black spruce are common in the boggy areas.

Deer Lake Bog (44°02′ N, 71°19′ W) is a *Sphagnum* bog at 1325 m, with a very small pool of water on its northeastern edge. The bog (0.09 ha) is in the middle of a large area of impeded drainage located in the col between Mt. Moosilauke and Mt. Jim. Subalpine fir forest with black spruce grows in the shallow *Sphagnum* peat of this poorly drained area, while fir forest ≈7.5 m tall also grows on uplands around the bog. Treeline is 115 m above the site. The bedrock underlying Deer Lake Bog and the surrounding slopes is composed of interbedded quartzite and mica schist of the Littleton Formation (Billings 1937).

Two lake basins (44°16′ N, 71°19′ W) lie 1.7 km southwest of the summit of Mt. Washington in the alpine zone 75 m above treeline. The core (see Fieldwork, below) is from Lake of the Clouds, the larger of the two basins (0.43 ha) at 1542 m. The 15-ha watershed is long and narrow, extending northeast from the lake to an elevation of 1676 m, and contains several springs (Buchanan 1975). The bedrock underlying Lake of the Clouds is the same as that at Deer Lake Bog: interbedded quartzite and mica schist of the Littleton Formation (Billings et al. 1946). Patches of fir and black spruce krummholz grow around their shores. Dwarf-shrub heath with *Vaccinium uliginosum, Vaccinium vitis-idaea, Ledum groenlandicum, Kalmia polifolia,* and dwarf birch (*Betula glandulosa*), and snowbank vegetation with *Houstonia caerulea* var. *faxonorum* and *Veratrum viride* are also common around the lake (Bliss 1963).

Low-elevation sites (<1100 m)

Davis and Shane (M. B. Davis, *personal communication*) have studied the fossil record in lake sediment from sites in the northern hardwood and spruce–fir forests on the lower slopes (Fig. 1). I have included summaries of the results from two of these sites, Lonesome Lake (900 m) and Lost Pond (650 m).

Fieldwork

The sediment cores were taken from a raft with a modified Livingstone sampler (5 cm in diameter). Surface cores with well-preserved mud–water interfaces were collected in plastic tubes (2.5 cm in diameter, 1 m in length) with pistons (Wright et al. 1965). Deer Lake Bog was sampled from the bog mat. Peat samples were collected from a cleaned face of a pit 80 cm deep. Below 80 cm the peat was humified enough to collect with a Livingstone sampler. The upper peat was resampled with a Russian peat sampler (Jowsey 1965).

Moss polsters were collected in east–west transects across the subalpine and alpine zones of three mountain peaks, Mt. Moosilauke, Mt. Lafayette, and Mt. Washington (Table 1). At designated elevations several clumps of moss were collected from a small area and the vegetation was sampled in a 2 × 5 m relevé (Müller-Dombois and Ellenberg 1974). The position of treeline and the presence of patches of krummholz were also recorded along each transect. The vegetation around the four coring sites was also sampled with 2 × 5 m relevés.

Laboratory techniques and pollen analysis

The Troels-Smith system of sediment characterization (Troels-Smith 1955) was used to describe the sediment. Samples from fresh cores were taken for pollen analysis and for loss-on-ignition determination with 1-mL volumetric spoons. The loss-on-ignition samples were dried for 24 h at 110°C and ashed for 2 h at 550° (Dean 1974). Seventeen samples were carefully removed from the cores, scraped to remove surface contamination, and sent to various radiocarbon laboratories for dating (Table 2).

Sediment samples were prepared for pollen analysis using the procedures of Faegri and Iversen (1975) and Cwynar et al. (1979). To determine the volume concentration of pollen and spores, a known amount of exotic *Eucalyptus* pollen was added to each sample (Stockmarr 1971). The conventions used in categoriz-

TABLE 1. Moss polsters collected in east–west transects across the subalpine asnd alpine zones of the indicated mountains.

No.	Mountain	Elevation (m)	Vegetation type	Elevation above or below treeline
1	Mt. Moosilauke	1402 west	fir krummholz	just below treeline
2	Mt. Moosilauke	1410 west	dwarf shrub heath-rush*	8 m above treeline
3	Mt. Moosilauke	1463 west	dwarf shrub heath-rush	61 m above treeline
4	Mt. Moosilauke	1463 east	snowbank community*	31 m above treeline
5	Mt. Moosilauke	1432 east	glade-fir krummholz†	at treeline
6	Mt. Lafayette	1219 west	subalpine fir forest	186 m below treeline
7	Mt. Lafayette	1372 west	subalpine fir forest	33 m below treeline
8	Mt. Lafayette	1524 west	dwarf shrub heath*	119 m above treeline
9	Mt. Lafayette	1584 west	sedge-rush dwarf shrub heath*	179 m above treeline
10	Mt. Lafayette	1584 east	snowbank community	84 m above treeline
11	Mt. Lafayette	1524 east	dwarf shrub heath-rush	24 m above treeline
12	Mt. Lafayette	1493 east	fir krummoholz	7 m below treeline
13	Mt. Washington	1372 west	subalpine fir forest	90 m below treeline
14	Mt. Washington	1494 west	dwarf shrub heath	32 m above treeline
15	Mt. Washington	1557 west	sedge meadow*	93 m above treeline
16	Mt. Washington	1905 west	sedge meadow	443 m above treeline
17	Mt. Washington	1905 east	sedge meadow	380 m above treeline
18	Mt. Washington	1676 east	dwarf shrub heath-rush	151 m above treeline
19	Mt. Washington	1372 east	alder thicket	153 m below treeline

* Plant communities recognized by Bliss (1963).

† Vegetation type recognized by Reiners and Lang (1979).

ing the pollen and spores follow those of Cushing (1963) and Birks (1976). Taxon names are after Fernald (1950).

At least nine different species of birch are currently found in New England and could be represented in the pollen assemblages. In the White Mountains, the three most important species of birch are dwarf, paper, and yellow birch. I measured the diameter of all birch grains from 14 samples at Lake of the Clouds and used a 20-μm cutoff to separate shrub (dwarf) from tree birch (Ives 1977). The two bottom histograms on Fig. 12 give measurements of 50 grains from fossil birch anthers. These two samples indicated the variation in grain size from a single individual.

Three species of alder, *Alnus crispa, Alnus rugosa,* and *Alnus serrulata* are found in New England today. Green alder (*Alnus crispa*) pollen can be separated from *Alnus rugosa* and *Alnus serrulata* pollen by pollen morphology (Bryan 1954, Cushing 1963).

The pollen sum, which includes all terrestrial taxa including fern and *Lycopodium* spores, was usually between 350 and 500. For the pollen data from surface samples below treeline the mean percentages are based

TABLE 2. Radiocarbon dates of samples removed from sediment cores.

Coring site	Depth of core sample (cm)	Sample code	Radiocarbon age (yr BP)
Kinsman Pond	50–60	GX-6052	220 ± 130
	280–293	GX-5604	8830 ± 165
	310–315, 320–337, 305–331.5*	I-10, 685	4490 ± 95†
Eagle Lake Bog	62–71.5	GX-6050	2760 ± 145
	297–307	GX-6051	7800 ± 190
	396–408	I-9907	9155 ± 145
Deer Lake Bog	120–128	I-9905	1485 ± 75
	180–191	GX-5602	2790 ± 130
	257–268	GX-5603	6465 ± 205
	384–395	I-9906	8620 ± 215
	425–435	QL-1134	10 960 ± 110
	463–488	QL-1133	13 000 ± 400
Lake of the Clouds	281–292 (site B1)	I-9908	8785 ± 135
	28–41 (site C1)	I-10 681	225 ± 85
	139–151 (site C2)	I-10 682	7010 ± 115
	222–237 (site C2)	I-10 683	10 300 ± 165
	280–295, 300–315, 285–315‡ (sites C2 and C3)	I-10 684	11 530 ± 420

* To obtain sufficient organic material for radiocarbon dating in these highly inorganic sediments, the bottoms of two cores were combined. The cores were taken < 1 m apart, and were correlated by pollen analysis before dating.

† This date was discarded. Not only is this sample significantly younger than sediment above it, but spruce pollen is a prominent type in the pollen assemblage. In New Hampshire, spruce is no longer an important component of the pollen sum by around 10 000 yr BP.

‡ As at Kinsman Pond, the bottoms of two cores were sent for radiocarbon dating.

on the 7 moss polsters at or just below the limit of continuous krummholz and the surface samples of four ponds in the subalpine fir forest (Table 3). The surface sample from Lake of the Clouds Bog plus 11 moss polsters are the basis of the mean percentages above treeline (Table 3). Pollen/spore taxa that are over-represented in particular moss polster samples have been removed from the pollen sum. These include Gramineae, Polypodiaceae, *Alnus,* and c.f. *Spiraea latifolia.* A taxon was considered over-represented when it covered a significant portion of the relevé (>10%) and its pollen/spores were present in unusually high proportions (20% or more) in the polster sample.

After the radiocarbon and pollen samples were taken, 2.5-cm lengths of core were removed for macrofossil analysis. After the potentially contaminated outer surface was removed, the sample was heated for 2 h in soapy water to disaggregate the sediment, then gently washed through a set of three United States Standard Series Sieves (595-, 250-, and 90-μm mesh, respectively). The coarse fraction was examined under 12× magnification and the medium fraction under 25× magnification. Plant and some animal fragments were picked from the sample and preserved in a drop of glycerin. The minimum needle number is the lowest number of needles represented by the needle fragments and twigs in a sample (Delcourt 1978). Seeds of yellow/black birch (*Betula lutea/B. lenta*) type, paper birch, and dwarf birch can all be distinguished in well-preserved material (Cunningham 1957).

RESULTS

Comparison of modern pollen and vegetation data

Vegetation.—Vegetation data from 10-m^2 relevés were collected at most of the sampling sites and modern pollen and macrofossil assemblages were compared to the local vegetation. The only trees in the subalpine zone relevés were fir, heart-leaved paper birch, red spruce, and black spruce. The spruce is of minor importance. Red spruce is occasionally found in the lower elevations of the subalpine fir forest and black spruce occurs near treeline in the Presidential Range. Mountain ash (*Pyrus americana*) occurs in the subalpine fir forest (Reiners and Lang 1979), but was not present in any of the relevés. Skunk currant, *Ribes glandulosum,* is one of the few shrubs in the subalpine fir forest. Ericaceous shrubs common to the alpine zone occur if there is an opening in the forest canopy.

The herbaceous vegetation in the subalpine fir forest is abundant but relatively monotonous. Approximately three-quarters of the relevés sampled had significant coverage of *Dryopteris spinulosa* and *Oxalis montana.* Other boreal forest herbs, such as *Coptis groenlandica, Clintonia borealis,* and *Maianthemum canadense,* occurred in a large number of relevés (40%) (Spear 1981). Krummholz and shrubs are common in the alpine zone. The krummholz species are fir, heart-leaved paper birch, and black spruce. Dwarf birch is abundant at Lake of the Clouds and green alder, *Alnus crispa,* is common on the east slope of Mt. Washington. Numerous ericaceous shrubs are found in the alpine zone. Three of the most abundant shrubs are *Vaccinium uliginosum, Vaccinium vitis-idaea,* and *Ledum groenlandicum.* The alpine zone has abundant herbaceous vegetation. Among the most common herbs are *Carex bigelowii, Juncus trifidus, Arenaria groenlandica,* and *Lycopodium selago,* which occurred in over one-third of the alpine relevés. Herbs found in the fir forests below treeline are also found in the alpine zone, especially around patches of krummholz.

Pollen.—Surface samples demonstrated that pollen assemblages from sites in the subalpine forest can be distinguished from those in the alpine meadow (Table 3). Five taxa—Ericaceae, Cyperaceae, Caryophyllaceae, *Lycopodium selago,* and *Houstonia caerulea* var. *faxonorum*—have pollen or spores that were found almost exclusively in the alpine moss polsters (Table 3). Plants from these taxa rarely occurred in relevés below treeline and almost never left a pollen/spore record in subalpine samples (Fig. 2). The only moss polster outside the alpine zone that contained >1% of these indicators was number 13, 90 m below treeline on the west slope of Mt. Washington, from the bank of a stream that forms the headwaters of the Ammonoosuc River. Many alpine plants have lower altitudinal limits along the stream's rocky banks.

The pollen assemblages from the subalpine fir forest and the alpine zone showed some subtle differences. There is more fir pollen below treeline than above (Fig. 2). This trend is particularly evident in the mean pollen percentages of the moss polsters (Table 3). Fir pollen percentages average 6.0% below treeline and 2.1% above treeline. Although the average percentages of spruce, birch, and *Alnus* are higher above treeline than below, the differences are not statistically significant. The average percentages of lowland trees in subalpine and alpine assemblages are nearly identical. *Lycopodium* is a prominent herbaceous genus in the vegetation near treeline (Fig. 2). Its percentages are highest at treeline and in the lower alpine zone. This distribution correlates fairly well with its distribution in the vegetation. Ragweed (*Ambrosia*), a lowland weed, was not found in any of the relevés and is not known to grow in either the subalpine or alpine zones. Its presence in the pollen assemblages must result from distant transport.

A comparison of the mean percentages of the seven polsters below treeline with the mean percentages for the four ponds below treeline shows distinct differences in spectra between the two types of sites (Table 3). Generally, the ponds show a more regional pollen rain with more spruce and lowland trees than do the subalpine polsters. These results are similar to the results of the surface sample studies of Ritchie (1974) and Birks (1977) at treeline. In contrast, pollen percentages

TABLE 3. Percentages of pollen of various taxa in surface samples from Mt. Washington, Mt. Lafayette, and Mt. Moosilauke.

Taxon		Moss polsters		Pond	
		Below treeline	Above treeline	Below treeline*	Lake of the Clouds
		Percentage of pollen sum			
Picea	Mean	1.6	2.4	5.0	5.2
	Range	0.5–2.1	0.7–4.0	2.4–7.9	
Abies	Mean	6.0	2.1	4.0	2.1
	Range	3.5–11.9	0.2–3.5	3.3–5.1	
Betula	Mean	46.0	49.8	44.6	48.2
	Range	37.4–56.7	41.1–56.8	40.5–47.9	
Alnus	Mean	2.0	2.8	0.8	2.8
	Range	0.7–3.3	1.0–6.7	0.3–1.4	
Pinus	Mean	8.2	8.2	6.9	6.7
	Range	5.0–13.4	4.9–13.1	4.2–8.0	
Tsuga	Mean	2.4	2.9	5.1	4.6
	Range	1.3–2.8	1.57–5.40	3.5–5.9	
Quercus	Mean	4.0	3.4	4.4	3.4
	Range	2.3–6.3	1.2–6.4	2.4–6.5	
Fagus	Mean	4.2	4.6	8.0	5.3
	Range	2.4–5.7	1.4—7.5	6.8–10.2	
Acer	Mean	2.0	1.6	2.0	0.7
	Range	0.8–3.5	0.3–3.0	0.7–4.1	
Ulmus	Mean	1.4	1.0	1.1	0.2
	Range	1.0–1.9	0–2.1	0.5–1.6	
Other trees	Mean	3.6	3.6	4.7	3.6
	Range	1.8–6.7	1.5–7.0	2.2–6.4	
Σ Lowland trees	Mean	25.8	25.3	32.2	24.5
	Range	14.4–37.4	19.5–32.2	27.2–39.2	
Salix	Mean	0.4	0.2	0.2	0.7
	Range	0–1.1	0–0.5	0–0.5	
Myrica	Mean	0.3	0.2	0.2	0
	Range	0–0.8	0–0.6	0–0.4	
Ericaceae	Mean	0.1	0.5	0.1	0.5
	Range	0–0.5	0–1.3	0–0.3	
Gramineae	Mean	3.1	3.3	1.8	4.4
	Range	1.1–8.7	1.4–7.6	0.7–2.7	
Cyperaceae	Mean	0.2	1.5	0.6	1.4
	Range	0–0.9	0–6.1	0–1.7	
Caryophyllaceae	Mean	0.1	1.3	0	0
	Range	0–0.4	0–5.0	0	0
Polypodiaceae	Mean	1.9	1.2	5.7	1.8
	Range	0.7–6.9	0–2.4	0.8–16.4	
Total *Lycopodium*	Mean	2.3	3.0	0.6	2.1
	Range	0.9–6.4	0.4–5.0	0.3–1.3	
Lycopodium selago	Mean	0	0.4	0	0.2
	Range		0–2.3	0	
Ambrosia	Mean	6.4	3.9	1.7	3.4
	Range	2.4–14.5	1.9–10.1	0.22–2.4	
Total herbs	Mean	18.1	16.5	12.5	15.8
	Range	9.8–24.1	10.1–22.7	8.6–21.0	
Long-distance transport (Σ lowland trees plus *Ambrosia*)	Mean	32.0	29.0	33.8	28.2
	Range	20.0–42.3	21.3–35.9	24.0–41.3	

* Ponds below treeline: Kinsman Pond 1140 m, Garfield Pond 1165 m, Eagle Lake Bog 1275 m, and Red Pond 1350 m.

from surface sediments at Lake of the Clouds, above treeline, are remarkably similar to the average percentages from the 11 alpine moss polsters (Table 3). These results are preliminary and should not be indiscriminately applied to other regions; however, the five taxa that distinguish the alpine samples (Ericaceae, Cyperaceae, Caryophyllaceae, *Lycopodium selago,* and *Houstonia caerulea* var. *faxonorum*) should be indicative of treeless open conditions in the fossil pollen record of the White Mountains.

Plant macrofossils.—Analyzing plant fragments in surface sediment samples reveals how well the local

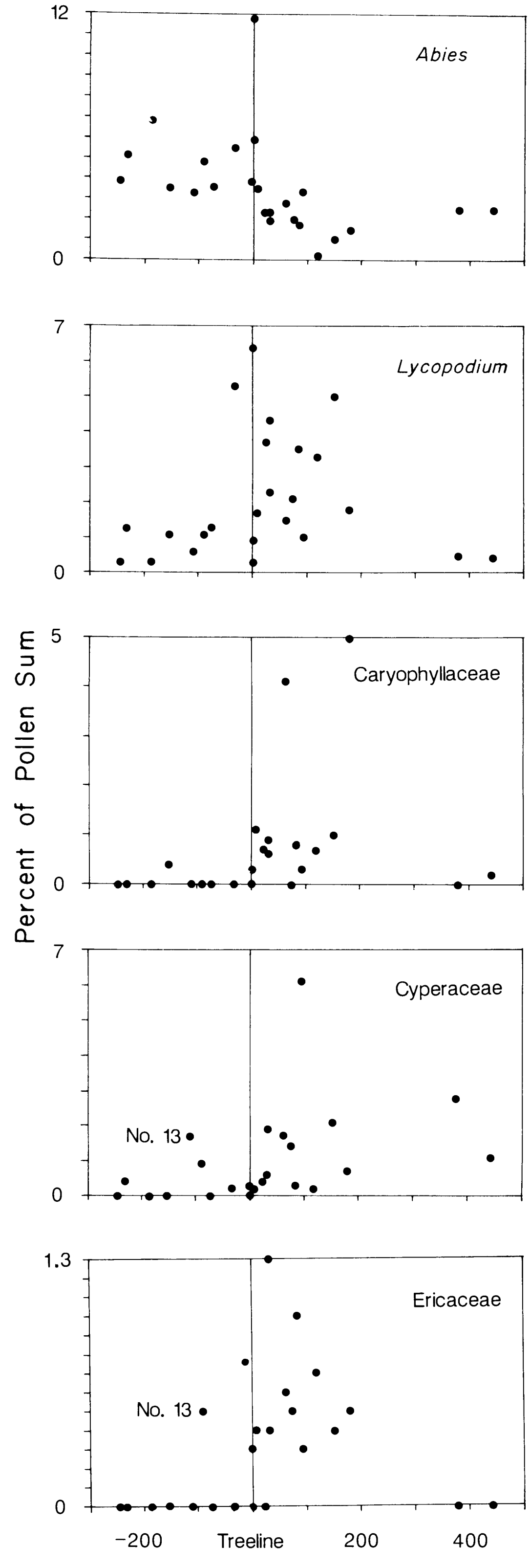

FIG. 2. Modern pollen distribution in the vicinity of treeline for Mt. Washington, Mt. Lafayette, and Mt. Moosilauke.

vegetation is represented in lake sediment (Dunwiddie 1987). Spruce needle fragments are over-represented in the sediment record in the White Mountains. Needle fragments from krummholz spruce and fir dominate the macrofossil record at "lower" Lake of the Clouds (1551 m). Although the relevé data collected at the site show that spruce is much less abundant than fir, spruce needle fragments are nearly as abundant as fir in the sediment. Hermit Lake (1200 m), a small pond just outside Tuckerman Ravine (a large cirque 1.5 km southeast of the summit of Mt. Washington) is in a red spruce–balsam fir forest with occasional black spruce (Teeri 1968). The importance value for spruce in the Hermit Lake forests is 5%; for fir it is 80% (M. B. Davis, *personal communication*). Spruce needle fragments are 10 times as abundant as those from fir in Hermit Lake sediment. Surface sediments from Spaulding Pond (1300 m), a small lake located on the floor of the Great Gulf cirque 1.6 km northeast of the summit of Mt. Washington, have green alder seeds. Green alder and black spruce are found at Spaulding Pond and are particularly common at treeline in the subalpine fir forest along the eastern flank of the Presidential Range today. Green alder should leave a clear record in macrofossil assemblages.

Study sites

With my major focus on local and high-elevation vegetation, I have presented summary diagrams for each site with only the taxa represented that have been important in the history of high-elevation vegetation. (All diagrams are plotted on a time scale constrained by available radiocarbon dates [Table 2 and Fig. 3].) I have presented complete pollen and macrofossil diagrams for only one representative site, Deer Lake Bog (Figs. 4 and 5) and used these diagrams to describe the regional vegetation history for central New Hampshire forests. On the summary diagrams, I included only three arboreal taxa: spruce, fir, and birch. Species belonging to these taxa, along with *Alnus,* also grow as shrubs and krummholz. These taxa are represented by one or more species in the subalpine and alpine floras of today. A number of herbaceous and shrubby taxa are included either separately or grouped together as Nonaboreal Taxa on the summary diagrams. The Nonaboreal Taxa consist of the sum of percentages from nine taxa. Four of the taxa—*Artemisia,* Gramineae, *Salix,* and Tubuliflorae—have species found in tundra today. Richard (1977) states that these taxa, among others, are characteristic of the late-glacial "tundra" of southern Quebec. An additional five taxa—Caryophyllaceae, Cyperaceae, Ericaceae, *Lycopodium selago,* and *Houstonia caerulea*—are those identified as being common in modern alpine/tundra pollen spectra.

Deer Lake Bog sedimentation record.—Sediments at Deer Lake Bog are dated by radiocarbon dates (Table 2) and at two levels by pollen stratigraphy (i.e., correlations). The abrupt decline in hemlock pollen at 205

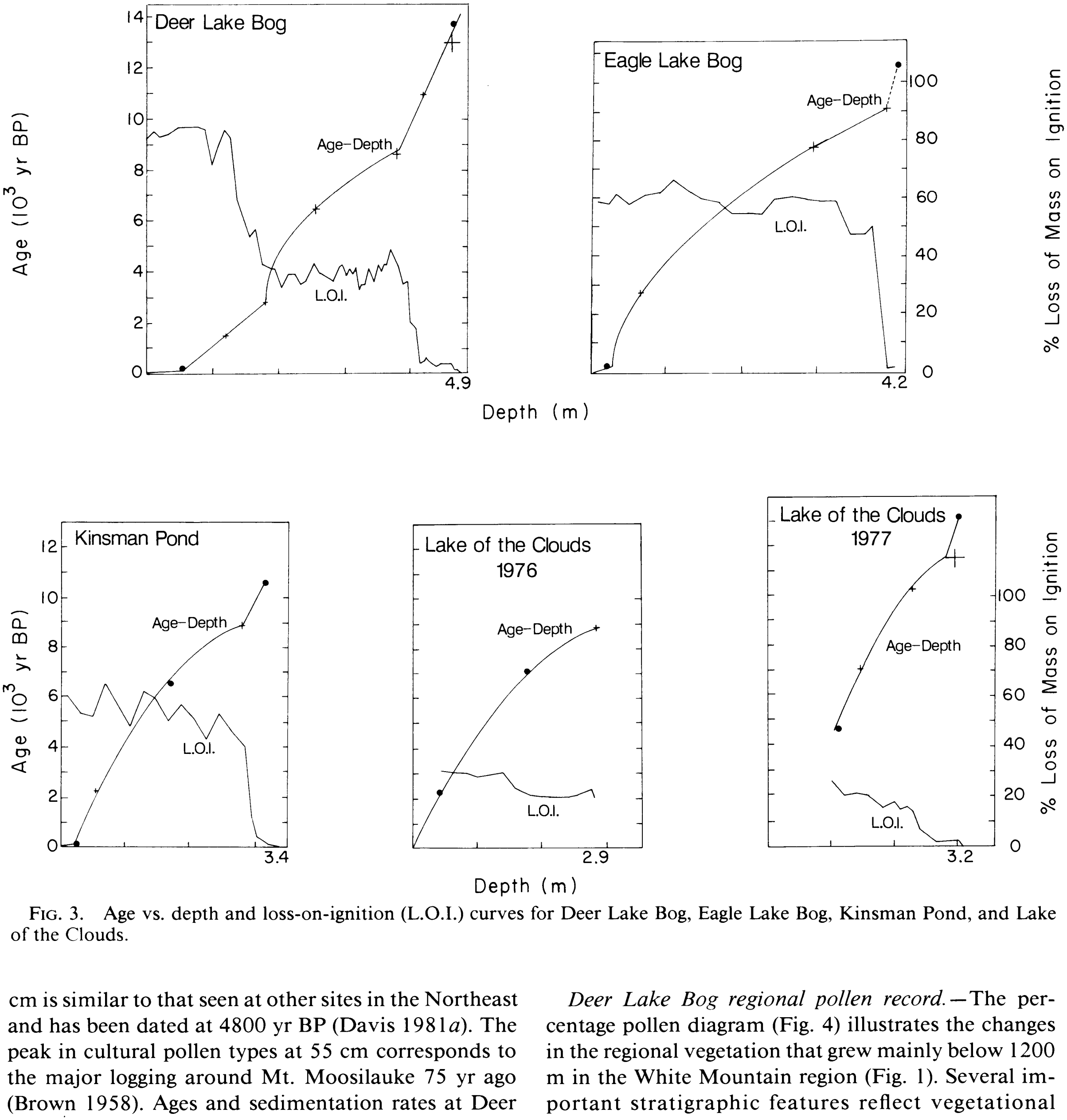

FIG. 3. Age vs. depth and loss-on-ignition (L.O.I.) curves for Deer Lake Bog, Eagle Lake Bog, Kinsman Pond, and Lake of the Clouds.

cm is similar to that seen at other sites in the Northeast and has been dated at 4800 yr BP (Davis 1981*a*). The peak in cultural pollen types at 55 cm corresponds to the major logging around Mt. Moosilauke 75 yr ago (Brown 1958). Ages and sedimentation rates at Deer Lake Bog were determined graphically because a simple curve did not fit through all the points and attempts to fit polynomial regressions to the dates were unsuccessful (Fig. 3).

Before 10 300 yr BP the pollen record is the sole source of information on vegetation at Deer Lake Bog (Figs. 5 and 6) because no macrofossils are found in sediments deposited before then. Sedimentation seems to have been relatively constant throughout this period, and the pollen influx data can be interpreted as indicative of changes in the vegetation. After 8500 yr BP irregular rates of sedimentation during basin infilling make pollen influx values highly variable and therefore less sensitive to vegetation changes (Davis and Ford 1982, Davis et al. 1984).

Deer Lake Bog regional pollen record.—The percentage pollen diagram (Fig. 4) illustrates the changes in the regional vegetation that grew mainly below 1200 m in the White Mountain region (Fig. 1). Several important stratigraphic features reflect vegetational changes (Fig. 4): (1) the rise in spruce pollen percentages at 11 500 yr BP and their decline around 10 000 yr ago and a second rise in the last 2000 yr; (2) the sharp peak (28%) in green alder pollen 10 500 yr ago; (3) the rise in birch pollen percentages immediately after the alder peak; (4) the rise of white pine (*Pinus* haploxylon type) percentages around 9500 yr ago; (5) the increase in percentages of hemlock pollen 8500 yr ago, their abrupt decline 4800 yr ago, and subsequent slow recovery; and (6) the increase in beech pollen percentages 6500 yr ago. These features are similar to those seen in other New England diagrams (Davis 1983, 1985, R. Davis and Jacobson 1985, Gaudreau and Webb 1985, Anderson et al. 1986).

From 14 000 to 12 000 yr BP the records at Deer

Lake Bog and other sites (Davis 1985, Davis and Jacobsen 1985) indicate that herb/shrub tundra covered the central New Hampshire landscape. Woodlands of poplar (balsam poplar according to Davis and Jacobson 1985), spruce, jack pine and later including fir became established after 12 000 yr BP. Spruce was undoubtedly a dominant tree in the mixed woodland forest. Trees of lesser importance in the forest were elm, larch, hophornbeam/ironwood, ash, maple, and oak. The dramatic decline of spruce followed by an increase in green alder for a brief period and then birch populations indicate the unstable, transitional nature of the forests 10 000 yr ago. The timing of this transition at Deer Lake Bog fits well within the boundaries of 11 000–9000 yr BP established for this pattern in the Northeast by Gaudreau and Webb (1985). The occurrence and abundance of alder is highly variable throughout the East. At Deer Lake Bog it occurs over a shorter time interval and is more abundant than at other sites. Alder is a nitrogen fixer and can greatly enrich the soil (Crocker and Major 1955). For this reason Davis (1983) and Gaudreau and Webb (1985) speculate on its significance to forest development in New England. Populations of white birch trees expand after those of alder decrease (Jackson 1983, M. B. Davis, *personal communication*).

By 9500 yr BP populations of white pine expanded in central New England (Likens and Davis 1975, Davis 1983) and pine forests became established. The composition of the lowland forests changed greatly over the next 3500 yr as pine populations decreased and populations of northern hardwood taxa increased. Hemlock and beech trees increased in abundance 8000 and 6500 yr ago, respectively. The mixed conifer/hardwood forest of New England became established by the end of this time. The decline in hemlock populations at 4800 yr BP (Davis 1981*a*, Webb 1982, Allison et al. 1986) altered the composition of the conifer/hardwood forest for a period of 2000 yr until hemlock populations gradually recovered. The expansion of spruce populations due to more favorable climatic conditions is apparent in the record at Deer Lake Bog and fits in with the general expansion of spruce populations during the last 2000 yr in the Northeast (Webb et al. 1983, Gaudreau and Webb 1985, Webb 1986).

Besides providing a record of local upland vegetation, the macrofossil diagram (Fig. 5) gives details on the process of infilling of the Deer Lake Bog basin. The maximum numbers of macrofossils from aquatic plants occur before 4000 yr BP. *Isoetes muricata* megaspores reach 651 at 5925 yr BP and *Potamogeton* seeds number 111 at 4650 yr BP. Seeds from *Potamogeton,* megaspores of *Isoetes,* and ephippia of Daphnia are all found until ≈1500 yr BP. After 4000 yr BP the numbers of macrofossils from taxa with paludal types such as Cyperaceae, *Juncus, Glyceria, Callitriche,* and *Sparganium,* increase. The pollen and seeds of sedges (Cyperaceae) are abundant from 2700 to 1000 yr BP as sedge peat filled the basin. Grass seeds from *Glyceria* are also abundant in this period, although little Gramineae pollen is seen in the record. Once the *Sphagnum* peat mat developed, the concentration of all macrofossils drops except for those of the Ericaceae family. In addition to the 203 *Chamaedaphne calyculata* seeds, 20 Ericaceae anthers were found at 625 yr BP. Ericaceae pollen is below 1% during the time of Sphagnum peat deposition.

Deer Lake Bog local pollen and macrofossil record.—The summary diagrams at Deer Lake Bog provide a clearer illustration of the changes occurring at higher elevations. The first 2200 yr of deposition (14 000–11 750 yr BP) were characterized by extremely low pollen influx (<1000 grains$\cdot$cm$^{-2}\cdot$yr^{-1}). Grass reached a maximum of 7%, but its pollen influx was only 22 grains$\cdot$cm$^{-2}\cdot$yr^{-1} (Fig. 6). Much of the pollen rain was undoubtedly from taxa growing a considerable distance away. Spruce pollen averaged 12.5%, while that of pine was higher, 17.5%. Birch values were especially high in the bottom sample, 15%. Pollen grains from thermophilous trees such as white pine, hemlock, sugar maple, red maple, elm, and oak were also found in these bottom samples. In fact, oak averaged 8.5% in the early assemblages.

At the start of the interval from 11 750–10 300 yr BP, total pollen influx had increased to ≈2000 grains$\cdot$cm$^{-2}\cdot$yr^{-1}. Spruce pollen reached a maximum of 22.6% and influx a maximum of 840 grains$\cdot$cm$^{-2}\cdot$yr^{-1} at 10 900 yr BP. Fir pollen remained <1% throughout this interval. Pollen influx from *Artemisia,* Caryophyllaceae, Gramineae, Cyperaceae, *Juniperus*/*Thuja,* and *Salix* pollen all reached maximum values at 10 900 yr BP.

Plant macrofossils including spruce needle fragments (Fig. 6) first appeared at the start of the interval 10 300–9000 yr BP. Spruce macrofossils (red/black spruce on the basis of resin ducts [Durrell 1916]) reached their highest concentration, 46 needles/50 cm^3 at 9950 yr BP as spruce pollen percentages fell below 5%. By 9650 yr BP the number of spruce needle fragments declined to the lower values of Holocene levels. Fir needles were found in the sample at 9650 yr but were absent in the sample at 9100 yr. Peak percentages of *Lycopodium* spores, 3.3%, are close to modern values found in the surface samples of ponds and moss polsters above treeline. One *Potentilla tridentata* seed was found in sediment dating from 9650 yr BP.

After 9000 yr BP the numbers of fir needles do not appear to increase until 8550 yr BP (Fig. 6), but this is perhaps an artifact. (Sediments from 9100–8550 yr BP were sent for radiocarbon dating and not sampled for macrofossils.) Relative proportions of fir pollen remained constant, 1–2%, until 1000 yr BP, when fir trees began growing on the bog mat. The macrofossil record of fir was also continuous for the last 9000 yr. Fewer macrofossils were deposited during the last thousand years of peat deposition. The regional rise in spruce

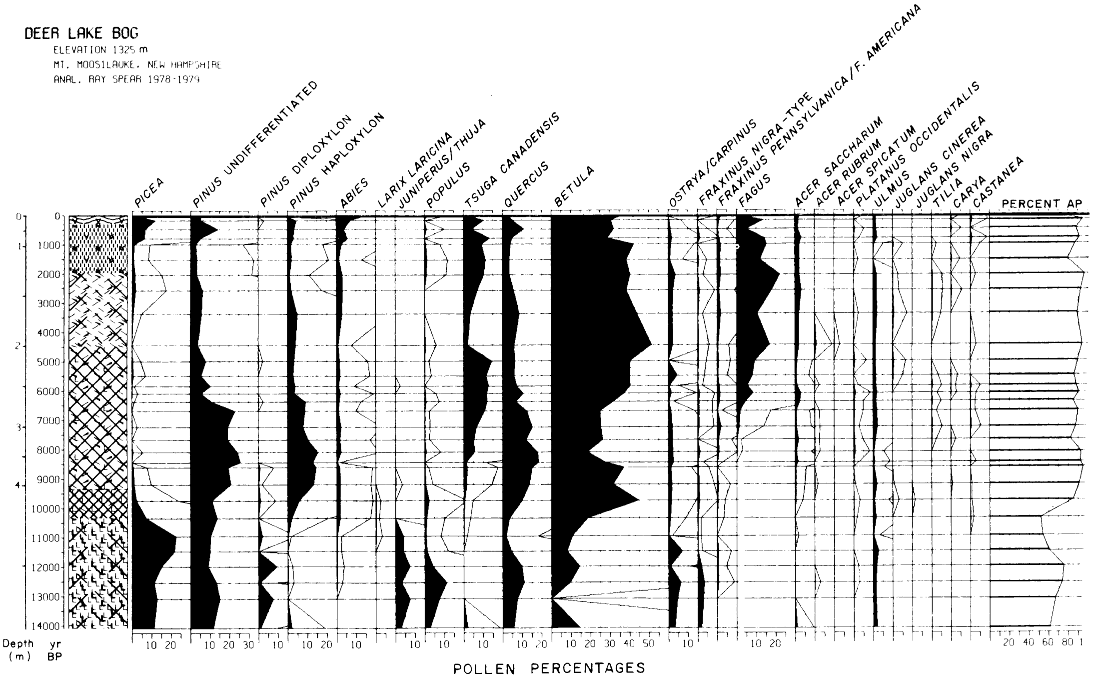

FIG. 4. Pollen percentage diagram for Deer Lake Bog.

pollen percentages during the last several thousand years is seen in the Deer Lake Bog record, and is particularly evident after 1000 yr BP. Spruce pollen reached 12% 175 yr ago, and 50 yr ago needle concentrations reached 6/50 cm^3.

Kinsman Pond and Eagle Lake Bog sedimentation. — Deposition time and estimated ages of sediment samples from the Kinsman Pond cores are derived from curves fitted by a polynomial regression (Dixon 1973) and graphical methods to five date levels (Fig. 3). The polynomial is fitted to four points, two radiocarbon-dated lake-sediment samples (Table 2) and two pollen-dated levels. The increase of beech pollen at 170 cm in the sediment represents beech's first appearance in central New Hampshire, dated at 6500 yr BP at Deer Lake Bog and Mirror Lake (Likens and Davis 1975). The other pollen-dated level is at 22.5 cm and has higher percentages of weed and herb pollen. This increase indicates cultural disturbance, mainly intensive logging, which began ≈100 yr ago (Chittenden

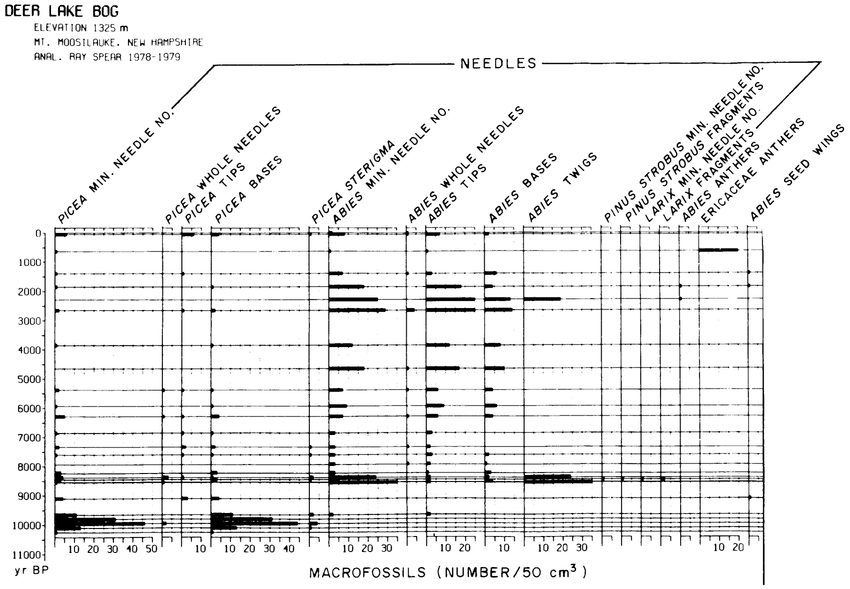

FIG. 5. Macrofossil diagram for Deer Lake Bog.

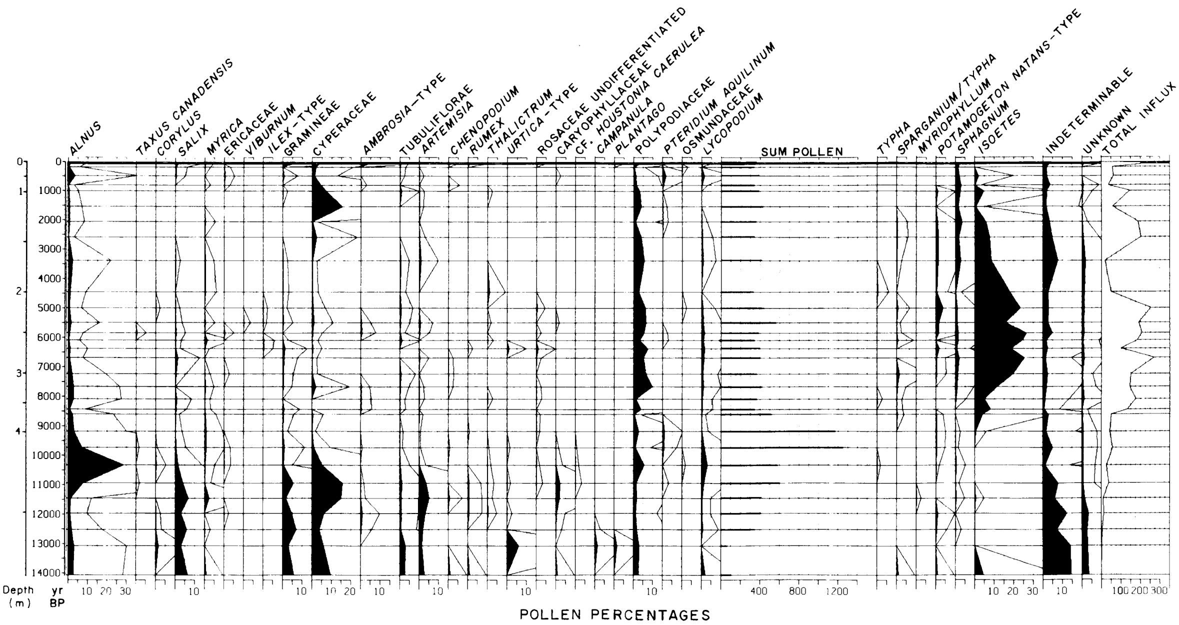

FIG. 4. Continued.

1905, Brown 1958). Sedimentation below the oldest radiocarbon date is estimated by correlation of the pollen sequence with radiocarbon-dated sites. The deepest pollen-bearing sediments in Kinsman Pond contain pollen assemblages characteristic of those deposited at other sites from 10 500 to 11 000 yr BP. The well-dispersed regional pollen types, white pine (*Pinus haploxylon*-type), hemlock, and birch are the best stratigraphic markers (See Fig. 4). White pine pollen percentages in the bottom three samples increase from 0.4% at 320 cm to 2% at 310 cm and 6% at 300 cm. No hemlock pollen is found at these three levels. Between 310 and 300 cm birch percentages increase from 11 to 46%. In addition, the two basal levels, 320 and 310 cm, have high pollen percentages of spruce (21.5%) and herbaceous taxa: Cyperaceae (18.2%), *Artemisia* (6%), and Caryophyllaceae (1%).

The age vs. depth plot for Eagle Lake Bog includes four points, three radiocarbon-dated levels (Table 2), and the origin. Since a polynomial regression does not

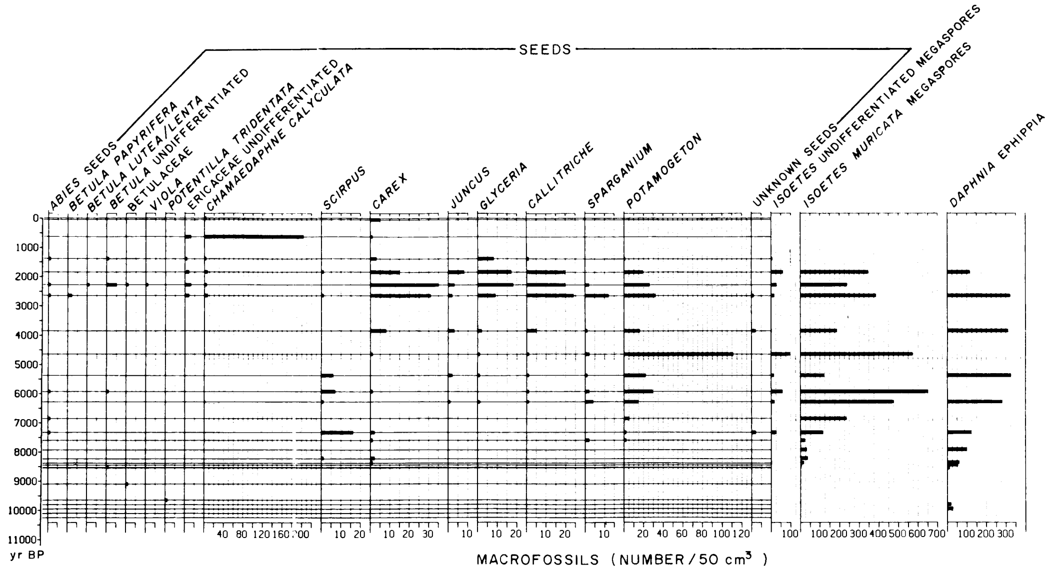

FIG. 5. Continued.

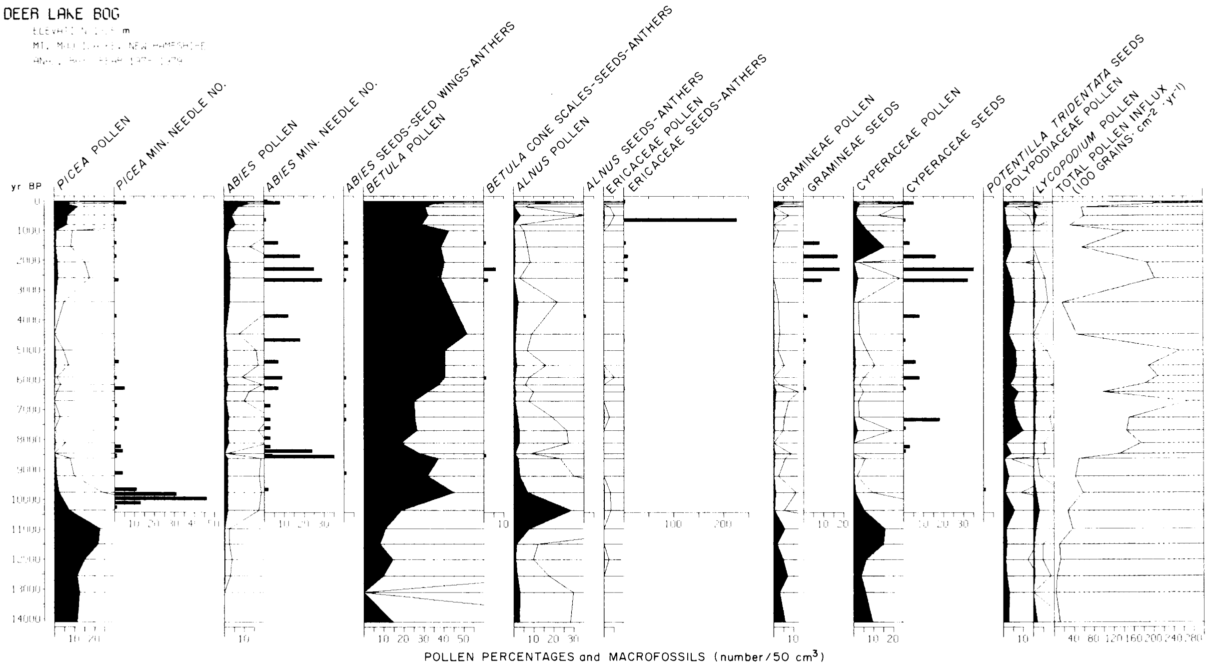

FIG. 6. Summary pollen and macrofossil diagram, Deer Lake Bog.

satisfactorily fit these points, the ages and sedimentation rates at Eagle Lake Bog are derived graphically (Fig. 3). As at Kinsman Pond, sedimentation for the lowest and post-European settlement levels are determined from straight lines. Sediment between 402 and 415 cm is gray silt, and dates from 10 500 to 10 000 yr BP. The sedimentary record at Eagle Lake Bog is a little shorter than at Kinsman Pond. White pine pollen percentages increase from 2% at 415 cm to 8% at 407 cm. At the same time birch percentages increase from 14 to 46%. Trace amounts of hemlock pollen, ≈0.5%, are found at both levels. There are also traces of herbaceous taxa: 2.3% Cyperaceae and 0.3% *Artemisia* pollen.

Sedimentation at the three sites in Franconia Notch was delayed after the regional ice retreat. Below 320 cm at Kinsman Pond, 17 cm of gray silt and fine sand were deposited. This sediment is barren of pollen and has very low percentages of organic matter (<2%). Lonesome Lake, at an elevation of 900 m on the same mountain slope 2.6 km east of Kinsman Pond (Fig. 1), has a similar pollen record (M. B. Davis, *personal communication*). The oldest radiocarbon date at Lonesome Lake is 10 610 ± 310 yr BP on a 20 cm long segment, 345–365 cm. The bottom pollen sample at 370 cm must be equivalent in age to that at Kinsman Pond. Sedimentation at Eagle Lake Bog began several hundred years later.

Kinsman Pond and Eagle Lake Bog pollen and macrofossil record.—I dated the 300-cm level at Kinsman Pond and the 407-cm level at Eagle Lake Bog by pollen stratigraphy at 9700 yr BP. The macrofossil record at Kinsman Pond began at 9750 yr BP (Fig. 7). A single charred spruce needle base was found at this level. The macrofossil assemblage in the next sample, from 9650 yr BP, contained abundant spruce and fir needles and a few birch seeds. The first appearance of macrofossils in Kinsman Pond sediment coincided with a change from inorganic silt and clay to gyttja. Loss-on-ignition values changed from 12.5 to 40% (Fig. 3). Sediment at Eagle Lake Bog also became markedly more organic at 9700 yr BP, when percent loss-on-ignition increased from 3 to 62% (Fig. 3). In contrast, at Deer Lake Bog and Lake of the Clouds, the first rise in organic matter and the first macrofossils occurred >500 yr earlier, indicating that the watersheds were stabilized by vegetation by 10 300 yr BP.

The largest number of spruce macrofossils was deposited between 9750 and 8500 yr BP at both sites. The highest spruce needles number is 29 needles/50 cm^3 at Kinsman Pond and 17 needles/50 cm^3 at Eagle Lake Bog (Figs. 7 and 8). While these numbers are much smaller than those at either Deer Lake Bog or Lake of the Clouds, they are far above the average Holocene values of 3 needles/50 cm^3 at Kinsman Pond and 4 needles/50 cm^3 at Eagle Lake Bog. In fact at Kinsman Pond, no spruce macrofossils were found in four samples spanning an interval of 1500 yr from 8500 to 6900 yr BP.

The macrofossil record of fir at Kinsman Pond suggests that fir trees were more prevalent in the forest after 5000 yr BP. Before 7000 yr BP, the average concentration of fir needle fragments was 9 needles/50 cm^3. After 7000 yr BP, the average was 27 needles/50 cm^3. The concentration of macrofossils of reproductive parts of fir, seed wings, seeds, and pollen sacs increased after

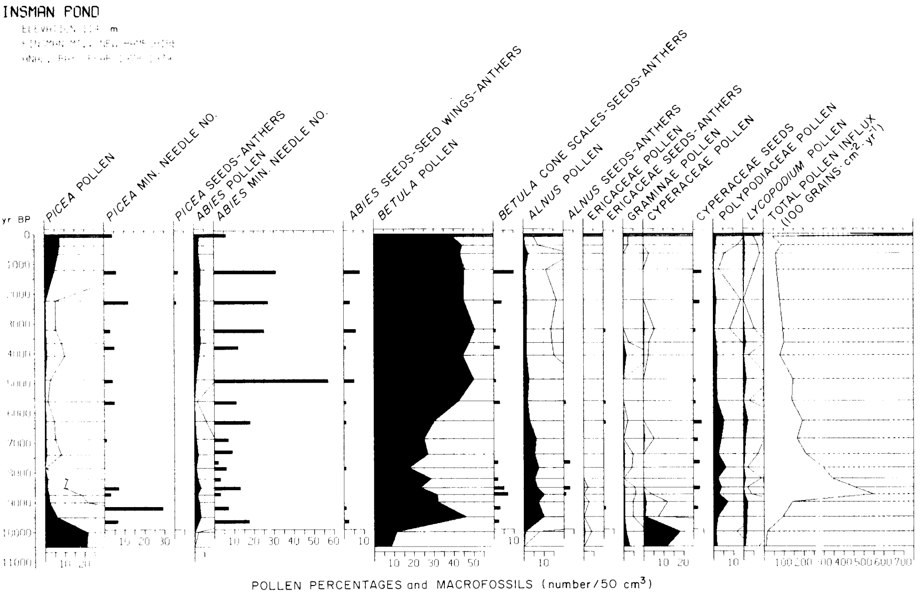

FIG. 7. Summary pollen and macrofossil diagram, Kinsman Pond.

5000 yr BP, as did fir pollen percentages. At Eagle Lake Bog (Fig. 8), the record of fir is more ambiguous. The sample at 8900 yr BP contained charred macrofossils; nearly half of the spruce and fir needle fragments and all those of white pine were burned.

Needle fragments of larch and a few pollen grains are found at Kinsman Pond, Eagle Lake Bog, and Deer Lake Bog at around the 9000 yr BP level. There were never more than 2 needles/50 cm^3 and no more than one or two pollen grains, < 1%, are found at any level.

Green alder pollen was more abundant from 10 000 to 5000 yr BP at both Kinsman Pond and Eagle Lake Bog than at either Deer Lake Bog or Lake of the Clouds. At Kinsman Pond, green alder pollen remained >5% and pollen influx exceeded 1000 grains·cm^{-2}·yr^{-1} from 9500 to 6200 yr BP. Several seeds and one anther were found in this interval. At Eagle Lake Bog, similar percentages and influx were found from 8800 to 5300 yr BP. There were no other indications in the fossil record of changes in the local forest composition. Seeds of *Potentilla tridentata* occurred in three levels at Eagle Lake Bog, 9050, 8900, and 3925 yr BP. They are probably from plants growing in open areas on the exposed ridges near the site.

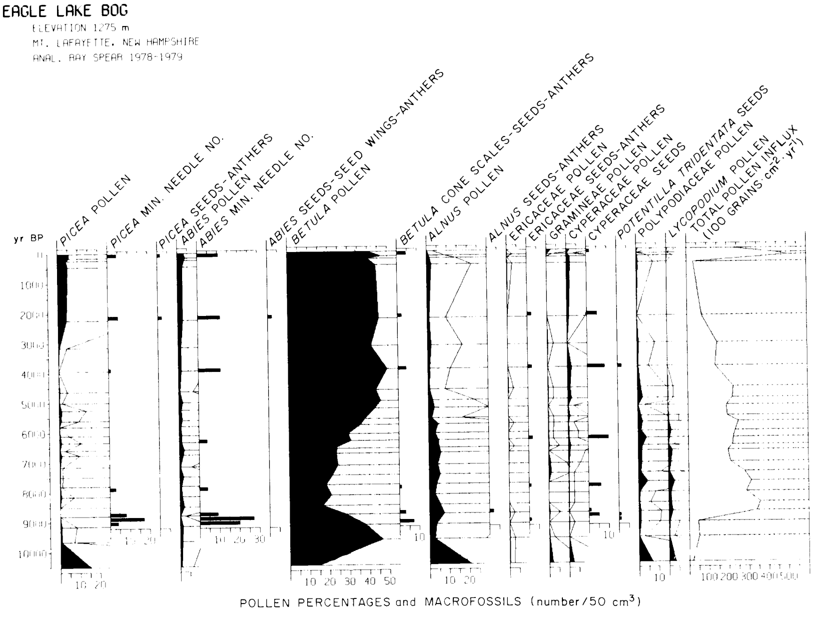

FIG. 8. Summary pollen and macrofossil diagram, Eagle Lake Bog.

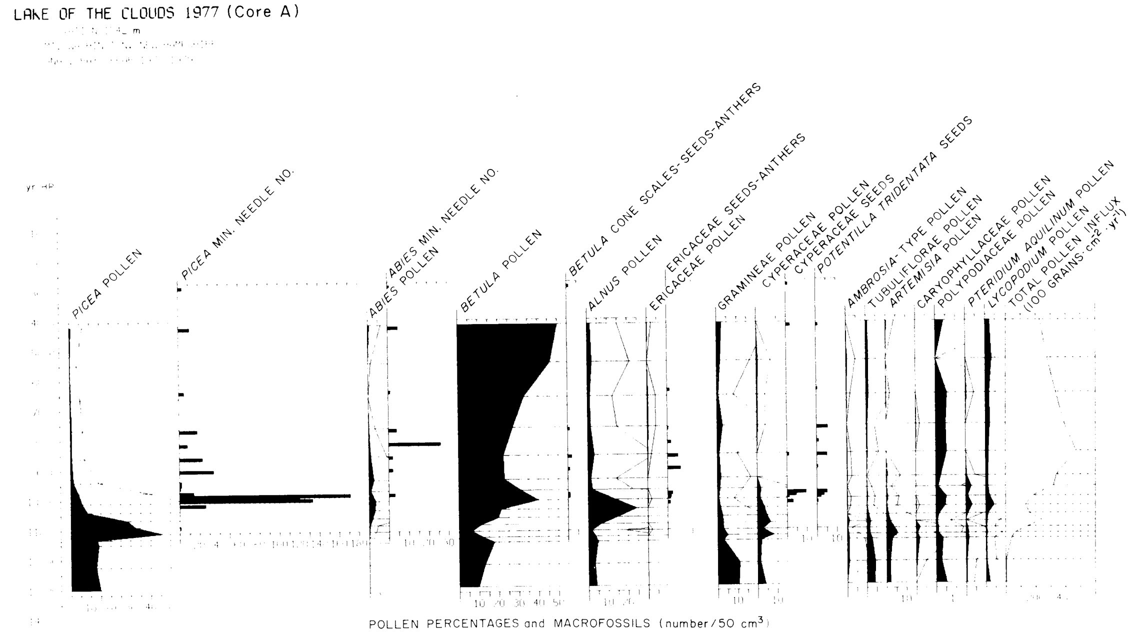

FIG. 9a. Summary pollen and macrofossil diagram, Lake of the Clouds 1977 (Core A).

Lake of the Clouds sedimentation.—Lake of the Clouds has a complex pattern of sedimentation. Several coring trips were made before a complete sedimentary record was obtained from the lake. Unfortunately, no one sequence contains the complete record since deglaciation. Pollen stratigraphy was used to correlate the 1976 and 1977 cores. For the 1976 (Fig. 3) core a polynomial regression is fitted to five points, the origin, the rise in spruce pollen at 40 cm (estimated at 2200 yr BP by correlation with the radiocarbon-dated 1977 core), the decline in hemlock pollen at 110 cm (dated at 4800 yr BP by regional correlation), the rise in *Isoetes* microspores at 180 cm (7000 yr BP, again by correlation with the radiocarbon-dated pollen stratigraphy of the 1977 core), and the basal radiocarbon date.

A second-order polynomial was fitted to four points on a plot of age vs. depth for the 1977 core (Fig. 3). One of the points is the hemlock decline of 4800 yr BP, which occurs at 110 cm. The other three points are the radiocarbon dates (Table 2). The radiocarbon date of 11 530 ± 420 yr BP is on sediment between 280 and 315 cm. Pollen assemblages in the lower portion of this section are similar to those at the base of Deer Lake Bog and probably date from approximately the same time, 13 000 yr BP.

Lake of the Clouds pollen and plant macrofossils.—For at least 1200 yr after the retreat of the continental ice sheet from 13 000–11 700 yr BP, deposition of pollen at Lake of the Clouds was extremely low (Fig. 9a) and no plant macrofossils were preserved. Vegetation was undoubtedly sparse at high elevations. The summary diagram, Fig. 9a, shows that the percentages of trees and shrubs that could have grown at high elevations were low (spruce 15%, birch 10–12%, and alder 5%). Grass pollen reached peak percentages of 10% between 13 000 and 11 700 yr BP, much higher than those reached at either Deer Lake Bog or Lost Pond (M. B. Davis, *personal communication*). Percentages of Tubuliflorae, nearly 5%, were also higher than at the other two sites. Wormwood (*Artemisia*) pollen percentages were also high in this interval, but lower than in the next period. Percentages of Caryophyllaceae pollen increased through this period. Pollen percentages from other arboreal taxa illustrate the importance of pollen transport in the early assemblages. Percentages of pine reached 33%, over half of which can definitely be assigned to the *Pinus* diploxylon type. The small size of pine grains in the *Pinus* diploxylon group indicates that most of the grains were from jack pine (*Pinus banksiana*). Oak percentages averaged 7%, and pollen from thermophilous taxa such as *Carya* (0.4%) were found in the bottom sample. Pollen from both poplar and *Juniperus/Thuja* were also relatively abundant (5 and 2%, respectively).

Although macrofossils were lacking from the record for 11 750–10 300 yr BP, the sedimentation rate remained constant and pollen influx data (Fig. 9a) can be interpreted easily. Alpine tundra must have become established at high elevations. Spruce, fir, and alder all showed high percentages and influx values at this time. Spruce reached a maximum of 46% just prior to 11 000 yr BP. By the end of the interval fir pollen reached 3%, while pollen of *Alnus crispa* peaked at 24%. From 11 750 to 11 000 yr BP, peak percentages and influx of several herbs occurred, most particularly *Artemisia* and Caryophyllaceae. Other herbs with high percentages from this interval were Gramineae (5%), Cyperaceae (8%),

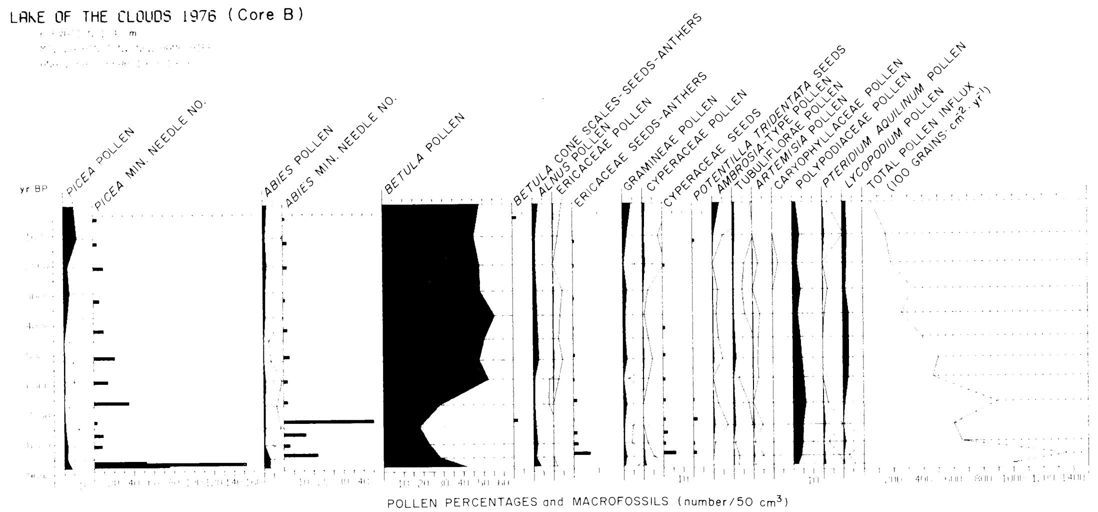

FIG. 9b. Summary pollen and macrofossil diagram, Lake of the Clouds 1976 (Core B).

Ambrosia (1%), and Tubuliflorae (2%). Their maximum influx values occurred later at 10 900 yr BP. This level had the highest total influx of any in the late glacial. Other herbaceous taxa seen in this early period were *Campanula, Armeria,* and *Polemonium.* Only one or two of these grains were found, so it is not clear if they were growing locally or blown in from a distant source.

From 11 000 to 10 300 yr BP, pollen percentages and influx of *Artemisia* and Caryophyllaceae decreased while the pollen influx of shrubs, *Juniperus/Thuja, Salix,* and dwarf birch increased. The accumulation rate of inorganic material dropped dramatically. *Juniperus/Thuja* reached a maximum influx of 770 grains·cm^{-2}·yr^{-1} at 10 900 yr BP, while maximum values of *Salix* (240 grains·cm^{-2}·yr^{-1}) and dwarf birch (3100 grains·cm^{-2}·yr^{-1}) occurred a little later, at 10 675 yr BP.

Pollen influx in sediments from 10 000 yr ago to the present at Lake of the Clouds is difficult to interpret. The average pollen influx values at Lake of the Clouds during the peak period of pollen influx is at least 1.4 times higher than lakes at low elevations or in the subalpine zone. Within-basin variation in pollen influx is illustrated by the two Lake of the Clouds cores (Figs. 9a and b). Maximum influx occurred 8700 yr ago in both, but influx rates of 135 000 grains·cm^{-2}·yr^{-1} in Core B are 2.9 times as high as influx in Core A (47 000 grains·cm^{-2}·yr^{-1}). The difference results from the pattern of sediment distribution within the lake basin. The deeper portions of lakes tend to receive the greatest amount of sediment, as a result of focusing (Davis and Ford 1982, Davis et al. 1984).

Red/black spruce arrived at Lake of the Clouds at ≈10 000 yr BP. The most striking feature in the Lake of the Clouds summary diagram (Fig. 9a and b) is the peak in the number of spruce needles (171 needles/50 cm^{3} in the 1977 core, Fig. 9a). This peak occurred after the maxima of spruce pollen percentages and influx at 11 100 yr BP. These first macrofossil assemblages also contained fir needle fragments and seeds of *Betula,* Ericaceae, Cyperaceae, and *Potentilla tridentata.* Fir needle fragments (three tips) first occurred at 9900 yr BP (Fig. 9a). However, the fir record did not become continuous until 9050 yr BP. In the 1976 core from Lake of the Clouds (Fig. 9b), fir did not appear until 8400 yr BP. Birch seeds were found in macrofossil assemblages dating from 10 000 yr BP. They could be from either dwarf birch or krummholz paper birch. Birch percentages and influx values in this interval are difficult to interpret. The peak in percentages occurred after the fall in *Alnus* and before the rise of pine percentages. Thus birch percentages were influenced by fluctuations in abundances of pollen from these two taxa. Birch influx increased at this time, but so did total influx, which may reflect sediment focusing.

The first macrofossil assemblages at 10 300 yr BP also contained seeds from taxa with alpine species. Up to 10 Ericaceae seeds were found per level, but this concentration was only one-third the concentration in sediments younger than 9000 yr BP (Fig. 9a). At the same time, 10 100 yr BP, Ericaceae pollen appeared in the pollen record. From 10 300 to 9850 yr BP, Cyperaceae pollen percentages were below peak levels but above modern levels (1–2%). Abundant seeds of Cyperaceae, 10 seeds/50 cm^{3}, occurred until 9900 yr BP. During the same interval, a maximum of 5 seeds of *Potentilla tridentata* were found per 50 cm^{3}. Pollen from the Caryophyllaceae disappeared at the beginning of this interval.

Percentages and influx of several spores (Fig. 9a) increased between 10 300 and 9000 yr BP. Clubmoss (*Lycopodium*) spores reached 3.5% of the pollen sum. In the surface sample data, *Lycopodium* spores attained the highest percentages just above treeline. Percentages and influx of Polypodiaceae and *Pteridium* spores reached high levels between 10 300 and 9700 yr BP, especially just after the maximum numbers of spruce macrofossils.

From 9700 to 9000 yr BP the macrofossil assemblages in the 1977 core contained a few pieces of degraded spruce needles. No fir needle fragments or Ericaceae, Cyperaceae, or *Potentilla tridentata* seeds were found. In the 1976 core (Fig. 9b), numerous spruce bases occurred in the first 300 yr of sedimentary record from 8800 to 8500 yr BP. These bases were all degraded, however, and no other macrofossils were found. The pollen record from 9700 to 9000 yr BP was difficult to interpret, because the arrival of white pine in lowland forests and the increase in birch distorted the pollen percentages, and because the total pollen influx was dramatically increasing. Increases in pollen influx of Gramineae and *Artemisia* at 9500 yr BP may reflect sediment focusing rather than a real increase in the abundance of these taxa in the vegetation.

Both the 1977 and 1976 cores have sediments dating from 9000 to 4000 yr BP. The macrofossil assemblages from this time interval are richer in species than from 4000 yr to the present. Although spruce pollen percentages were lower (1%) prior to 4000 yr BP, needle fragments were more abundant, averaging 40 fragments/50 cm^3 in both cores. Fir needles were most abundant between 9000 and 7000 yr BP in both cores. Too few birch macrofossils were found in either core to speculate about its abundance around Lake of the Clouds. However, a number of seeds were found in the 1977 core between 9000 and 7000 yr BP. Peak numbers of Ericaceae seeds were found in the 9000–7000 yr range: 26 seeds/50 cm^3 in Core A at 9100 yr BP, and 33 seeds/50 cm^3 in Core B at 8440 yr BP. Sedge (Cyperaceae) seeds show the same pattern in their distribution in the 1976 core. They were far more abundant from 10 000 to 5000 yr ago, indicating at least patches of open vegetation existed. Seeds of *Potentilla tridentata* were also more abundant during the first half of the Holocene. A maximum of 6 seeds/50 cm^3 was found at 7700 yr BP in Core A. Although pollen percentages of taxa with alpine species were higher in the second half after 5000 yr BP (Fig. 9b), pollen percentages of Ericaceae, Caryophyllaceae, and *Lycopodium* (2–3%) indicated that open conditions persisted throughout the interval, as did other pollen indicators of treeless vegetation: *Houstonia caerulea, Campanula,* and Liliaceae pollen. Bryozoan statoblasts (Phylactolacmata statoblasts) occurred from 8000 to ≈4000 yr BP. The quillwort (*Isoetes muricata*) first appeared at 6500 yr BP. Its numbers increased as the numbers of Bryozoan statoblasts decreased.

Vegetation development at high elevations

The summary diagrams suggest three stages in the establishment of the modern vegetation at high elevations (above 1150 m): (1) herb/shrub tundra; (2) treeline ecotone; and (3) modern alpine and subalpine zones.

During the first stage, herb tundra became established roughly 12 000 yr ago. The vegetation may have been similar to the sedge meadow (Bliss 1963) that covers the cone of Mt. Washington today. Alternatively the successional trend that Bliss (1963) describes for *Diapensia* communities, i.e., *Diapensia lapponica, Arenaria groenlandica, Agrostis borealis,* and *Juncus trifidus* colonizing areas subject to frost action, may better reflect the vegetation. However, comparison of fossil with modern pollen assemblages indicates that *Artemisia* was present and caryophylls were more prominent in the vegetation of this period than today. By 11 000 yr BP *Juniperus* or *Thuja* and probably *Salix* (willows) and ericads reached high elevations and grew as low, prostrate shrubs. The modern alpine shrub tundra communities could have become established at this time.

During the second stage, spruce arrived at roughly its modern altitudinal limit beginning ≈10 350 yr ago. It probably grew in patches of krummholz rather than as upright forest trees. The vegetation at treeline (1450 m) on the east slope of Mt. Washington (Harries 1966, Teeri 1968) may provide a modern analogue for the open vegetation of this time. Fir began to replace spruce between 10 000 and 9000 yr BP, but it did not become abundant in the subalpine forest and treeline ecotone until after 8500 yr BP. Although birch reached the high-elevation sites at about the same time as fir, it has never been abundant.

During the third stage, the modern alpine and subalpine vegetation became established around Kinsman Pond, Eagle Lake Bog, and Deer Lake Bog approximately 9000 yr ago. High-elevation vegetation was established, and subsequent changes in the mid- and low-elevation forest had little impact on the the subalpine and alpine zones.

DISCUSSION AND CONCLUSIONS

Deglaciation and plant colonization

Radiocarbon dates and pollen stratigraphy at the four high-elevation sites suggest rapid downwasting and retreat of the Laurentide Ice Sheet. If the summits stood above the ice sheet as nunataks they may have done so for only a short time (Bradley 1981, Gerath and Fowler 1982, Goldthwait and Mickelson 1982). The oldest date in the region, 13 870 ± 560 yr BP (sample GX 5429), is from the lowest elevation site, Mirror Lake at 200 m (Davis et al. 1980). Deer Lake Bog at an elevation of 1325 m on Mt. Moosilauke is 12 km northwest of Mirror Lake and dates from approximately the same time, 13 000 ± 400 yr BP. The two sites near Mt. Washington, Lost Pond (650 m; Davis et al. 1980) and Lake of the Clouds (1542 m), have sediments of comparable age. However, the mountain peaks could have been exposed for several thousand years prior to deposition of datable organic sediments in high-elevation basins. Residual ice in Franconia Notch could explain the later onset of sedimentation at Kinsman Pond, Lonesome Lake, and Eagle Lake Bog.

Extensive montane glaciation did not occur in the White Mountain cirques following the retreat of the continental ice sheet (Goldthwait 1970, Bradley 1981, Gerath and Fowler 1982). Two climatic parameters, temperature and snowfall, determine the elevation of montane glaciation. Goldthwait (1970) estimated that if, during the period of montane glaciation that preceded the last glacial maximum, precipitation were the same as today, the mean summer temperature would have had to be 9.5°C colder. Glaciers could also have been re-established in the cirques if snowfall were 1.5 times greater and mean summer temperature 5.5° cooler than the present (Goldthwait 1970). Consequently, the absence of montane glaciation from 14 000 to 11 000 yr BP meant that summer temperatures were either well above those of the full Wisconsinan or precipitation was drastically reduced.

Goldthwait (1976) estimates that the mean annual temperature was 4–6° cooler than today sometime after deglaciation 14 000 yr ago, because periglacial features are found down to an elevation of 1500 m in the Presidential Range. The paleoecological record supports Goldthwait's temperature estimates and helps to refine the timing and duration of the cold period. From before 13 000 to 11 750 yr BP, the pollen record is dominated by exotic arboreal pollen, indicating that little or no vegetation grew on the summits. Before 12 000 yr BP highly inorganic sediments were deposited at Lake of the Clouds, Deer Lake Bog, and Lost Pond (Fig. 10). Loss-on-ignition was $<2.5\%$. Deposition of inorganic particulate matter exceeded $150\ \mathrm{g \cdot m^{-2} \cdot yr^{-1}}$. The landscape must have been highly unstable. Wind eroded the fine material in the till and frost shattered the rocks forming the boulder fields of the summits. The temperatures were above full glacial minima but certainly well below those of today. By the end of this interval the Laurentide Ice Sheet had retreated to the north side of the St. Lawrence lowland and the Champlain Sea had formed. Sparse vegetation cover was established around 12 000 yr ago.

Herbaceous plants.—Herbaceous plant populations expanded in area and number between 13 000 and 11 000 yr BP. The herbs and shrubs listed as nonarboreal taxa (Fig. 11a), *Artemisia,* Gramineae, *Salix,* Tubuliflorae, Caryophyllaceae, Cyperaceae, Ericaceae, *Lycopodium selago,* and *Houstonia caerulea* var. *faxonorum* are often characteristic of open, sunny, and cold environments. The highest percentage (40%) of nonarboreal taxa occurred in the basal sample of Lost Pond 12 600 yr BP. Percentages at the higher elevation sites were smaller ($\approx 20\%$) indicating that the sparse vegetation grew on higher slopes and produced little pollen. A significant portion of the pollen rain at Lake of the Clouds and Deer Lake Bog was derived from arboreal taxa growing to the south. Between 11 500 and 11 000 yr BP, peak percentages in nonarboreal taxa occurred in the records of Lake of the Clouds (25%) and Deer Lake Bog (30%), but declined sharply at Lost Pond. The upper slopes became more densely vegetated during this period, while trees invaded the valleys and woodland was established at least to an altitude of 650 m.

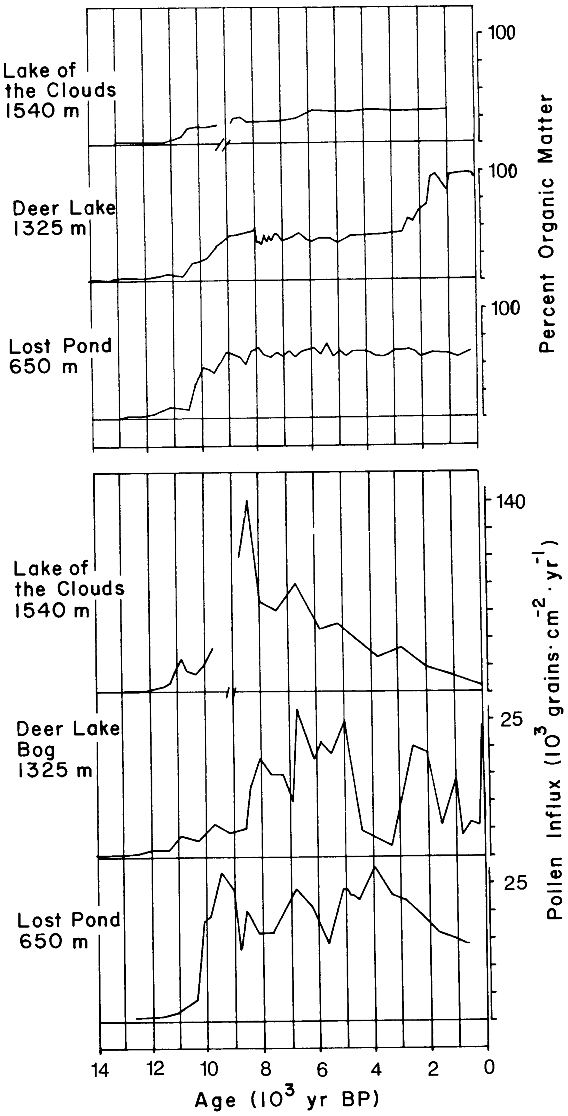

FIG. 10. Percent organic matter and pollen influx at Lake of the Clouds, Deer Lake Bog, and Lost Pond.

From 10 000 to 5000 yr BP nonarboreal taxa averaged 2.2% of the pollen rain at Lake of the Clouds, and rose to 3.9% after 5000 yr BP, exceeding the average of nonarboreal pollen at Deer Lake Bog (2.1%) and Lost Pond (1%). Macrofossil evidence also indicates the vegetation at Lake of the Clouds must have remained fairly open during the last 10 000 yr and especially during the last 5000 yr.

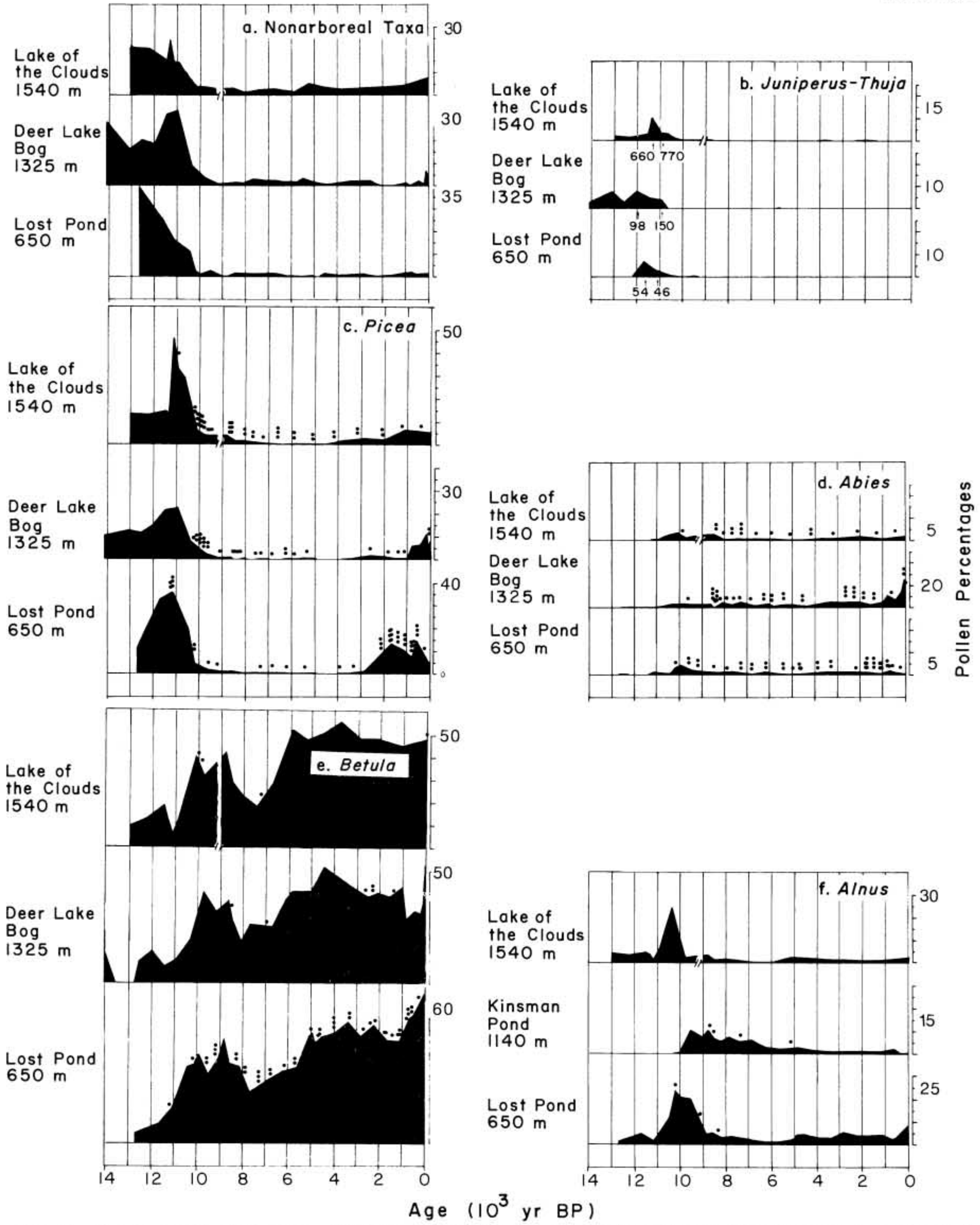

FIG. 11. Sumary pollen and macrofossil diagrams of specific taxa at Lake of the Clouds, Deer Lake Bog, and Lost Pond. Dots on the diagrams indicate the presence of macrofossils. · = macrofossils present; : = common; ⁝ = abundant.

Juniperus/Thuja.—The high-elevation sediments in the mountains contained no macrofossils of Cupressaceae, but pollen percentages of *Juniperus/Thuja* type (Fig. 11b) are comparable to those of Columbia Bridge, 65 km northwest of Mt. Washington. At Columbia Bridge, Miller and Thompson (1979) found a wide variety of *Juniperus communis* macrofossils, as well as ≈10% *Juniperus/Thuja* pollen in sediments dating from 11 500 yr BP. Webb et al. (1983) also suggest that *Juniperus* was the main contributor to the early peak in pollen at sites in southern Quebec. The peaks in *Juniperus/Thuja* pollen percentages and influx are relatively high and broadly synchronous at all three White Mountains sites, but differences exist among the sites. The ratio of *Juniperus/Thuja* pollen influx to total influx at Deer Lake Bog and Lake of the Clouds is 1.5 times higher than at Lost Pond. The influx values suggest that prostrate shrubs of *Juniperus* or *Thuja* were more abundant around the two high-elevation sites than in the lowland. They could have invaded the tundra-like vegetation on the high mountain slopes before the arrival of spruce. Pollen percentages indicate that *Juniperus/Thuja* plants disappeared from the region between 10 500 and 10 000 yr BP.

Spruce.—Spruce was one of the first forest trees to arrive in central New Hampshire (Davis and Jacobson 1985). Before 12 000 yr BP, spruce percentages were low, ≈10% (Fig. 11c). After this time, they began to rise, reaching maximum values between 11 500 and 10 900 yr BP at all three sites. Spruce macrofossils were found only at the lowest elevation site, Lost Pond, during this time. In the two upper-elevation sites, the high percentages of spruce pollen in the sediments resulted from upslope transport of pollen, for macrofossils did not occur at these levels. The charred spruce-needle base found at Lake of the Clouds in sediment dating from 10 900 yr BP may have been carried upslope by an updraft from a fire.

Red/black spruce (probably krummholz) reached the upper elevations at 10 300 yr BP. The spruce needles were found in sediments that recorded a precipitous decline in spruce pollen percentages. The decline in spruce pollen was not quite so abrupt and complete at Lake of the Clouds and Deer Lake Bog as at Lost Pond. Deer Lake Bog spruce percentages at 10 000 yr BP were nearly double those of Lost Pond and were equal to those of Lake of the Clouds. By 9000 yr BP spruce pollen percentages at Deer Lake Bog were 1%, equivalent to those at Lost Pond, while at Lake of the Clouds spruce percentages were still 4%. Populations of spruce probably persisted at high elevations.

Between 9000 and 3000 yr BP, spruce populations at all elevations were greatly reduced from their peak levels of 12 000–9000 yr BP. Lake of the Clouds had a continuous record of low pollen percentages and macrofossils from 9000 yr BP to the present. During the last 3000 yr, a dramatic regional increase in the amount of spruce pollen was recorded at all three sites (Fig. 11c). The largest increase was at the lowest site, Lost Pond, where spruce pollen percentages increased to 15% and macrofossil concentration tripled. The maximum percentage reached at Deer Lake Bog was 12%, whereas the maximum at Lake of the Clouds was 7%. Neither of the two higher elevation sites showed an increase in the number of spruce macrofossils. Spruce trees increased in number in the forests below 1100 m, the northern hardwood and spruce–fir forest at this time, but spruce populations did not increase in abundance above 1150 m in either the subalpine fir forest or the alpine meadow.

Fir.—Balsam fir arrived at all three sites in the White Mountains just before 10 000 yr BP (Fig. 11d). Its pollen, which is large and under-represented in the pollen record, slightly preceded its presence in the macrofossil record. Because fir pollen preceded fir macrofossils by a few hundred years at high-elevation sites, the pollen was probably blown upslope from low-elevation populations before trees began growing at high elevations.

Lost Pond had relatively abundant fir pollen and macrofossils from the time that fir first appeared (10 000 yr BP). Fir trees must always have grown in the lowland forests around the site. At Deer Lake Bog and Lake of the Clouds, fir macrofossils did not become abundant until 8500 yr ago. At Lake of the Clouds, fir needle fragments were more abundant between 10 000 and 7000 yr BP. Only scattered patches of fir krummholz occur near Lake of the Clouds today. Therefore, fir (krummholz) might have been more abundant there before 7000 yr BP. Fir might have been the first upright tree to grow near Deer Lake Bog. In the last 3000 yr, filling of the basin at Deer Lake Bog led to fir trees growing closer to and actually on the site of deposition.

Birch.—From ≈12 000–11 400 yr BP, birch pollen had peak values of 18.5% at Lake of the Clouds and 14.5% at Deer Lake Bog (Fig. 11e), the two high-elevation sites. There was no comparable peak at Lost Pond, where percentages of birch pollen remained at or below 10%. The size–frequency data from Lake of the Clouds show that most of the pollen in the 11 500-yr-old sample was between 20 and 22 μm in size (Fig. 12), therefore predominantly from tree birch. It must have blown over long distances from trees in the lowlands, the high percentages being the result of the low rates of local pollen production during this time. By 11 000 yr BP dwarf birch was most likely the major species in the White Mountain region, because the birch pollen from Lake of the Clouds was predominantly small (16 μm).

The pollen records were similar at all elevations after 10 000 yr BP. However, the macrofossil record of birch was quite different between the sites. Lost Pond was the only site with a continuous seed record from before 10 000 yr BP. Yellow and paper birch have both been prominent in the lowland forest for the last 10 000 yr. Although birch seeds occurred as early as 10 000 yr BP at Lake of the Clouds, the macrofossil record at the site and at Deer Lake Bog was sparse and discontinuous. All of the identifiable seeds in the macrofossil records from the four high-elevation sites were paper birch except for yellow/black birch type from the 2275 yr BP level at Deer Lake Bog.

Alder.—Percentages of green alder (*A. crispa*) at Lake of the Clouds and Lost Pond were highest between 10 500 and 10 000 yr BP (Fig. 11f), reaching levels of ≈24%. At Lake of the Clouds there are no alder macrofossils. Alder was probably not growing on the slopes

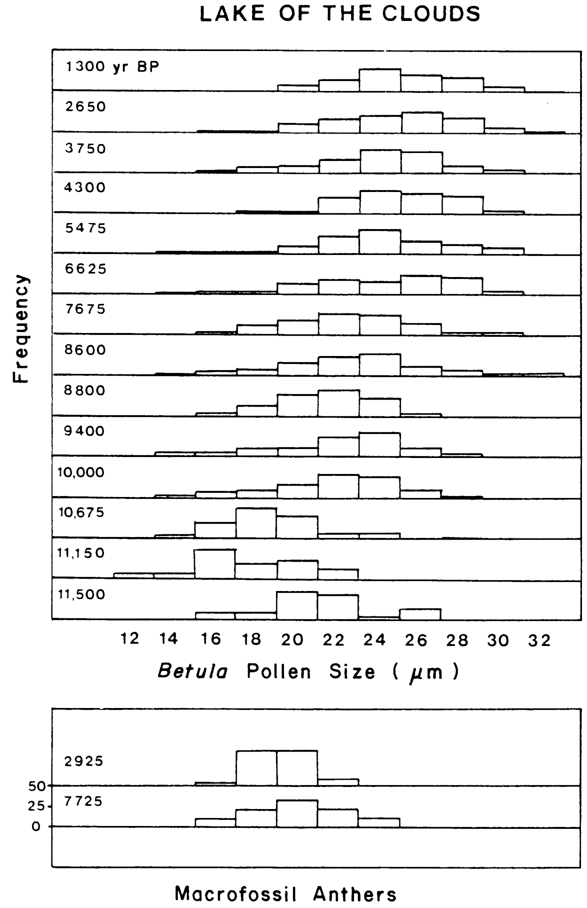

FIG. 12. Size frequency measurements of *Betula* pollen at Lake of the Clouds.

around Lake of the Clouds; peak alder percentages at this site probably represent pollen from the regional pollen rain. The pollen rain was predominantly from *Alnus crispa* populations before 8500 yr BP. After this time *Alnus rugosa* or *Alnus serrulata* pollen dominated the regional pollen rain. Kinsman Pond shows a more local pollen rain with green alder persisting until ≈6000 yr ago. Because alder is a prolific pollen producer it is over-represented in the pollen rain. In the northern boreal forests of the Mackenzie Delta, Northwest Territories, green alder percentages are comparable to the peak values reached in the White Mountains 10 000 yr ago. Here green alder does not form a significant portion of the vegetation cover; only scattered shrubs are present (Ritchie 1974). The ecological significance of the persistant populations of green alder in Franconia Notch is uncertain. In areas where large drifts of snow accumulate, birch krummholz and green alder replace fir and black spruce at treeline on the east slope of the Presidentials (Harries 1966). Similar conditions could have existed in Franconia Notch. The later expansion of populations in Franconia Notch could also explain why the sites contain a different vegetation history from Lake of the Clouds and Deer Lake Bog.

Development of modern subalpine and alpine plant communities

Between 11 750 and 11 000 yr BP the climate moderated slightly, allowing vegetation to become established on the mountaintops. Trees invaded the valleys, and extensive woodland had developed on the lower slopes. Sediments at all three sites became increasingly organic (Fig. 10) during this interval, indicating the establishment of more continuous vegetation. Pollen influx also increased 5–6 fold at Lake of the Clouds, 4-fold at Deer Lake Bog and 18-fold at Lost Pond (Fig. 10). Much of the herbaceous pollen—*Artemisia,* Caryophyllaceae, Gramineae, Cyperaceae, *Ambrosia,* Tubuliflorae—before 11 000 yr BP is from herbs that colonize unstable soils, indicating solifluction in the region. No *Artemisia* species are found in the alpine zone today. The southeastern outposts of the eastern arctic species of *Artemisia, A. borealis,* are the calcareous cliffs and the limestone, magnesium, and serpentine barrens of the Gaspé (Fernald 1950). Polunin (1948) reports finding *Artemisia borealis* growing on sunny south-facing slopes with unstable surfaces on the south coast of Baffin Island. In Greenland, *A. borealis* is found on xeric, sunny, unstable slopes on the west coast between 68° and 74° north (Bocher 1954). Fredskild (1973) found that *Artemisia* pollen percentages increased in lake sediments in the Godthabsfjord area of west Greenland 8000 yr ago. He suggests that this was due to a drier, more continental climate.

Caryophyllaceae pollen also indicates periglacial activity. Probably most of the Caryophyllaceae pollen belong to *Arenaria groenlandica* (mountain sandwort), a dominant species in several alpine communities in the White Mountains (Bliss 1963) and an important pioneer on disturbed soils (Harries 1966, Marchand and Roach 1980). Harries (1966) believes that in the alpine zone of the White Mountains *Arenaria groenlandica* grows solely on disturbed sites. *Agrostis borealis* and *Solidago cutleri* also invade such sites (Bliss 1963) and these species must be among those contributing to the high percentages of Gramineae and Tubuliflorae pollen before 10 300 yr BP. The late-glacial Cyperaceae pollen undoubtedly includes that of *Carex bigelowii* which inhabits windswept sites that are probably subject to frost action today (Bliss 1963, Harries 1966). A few shrubs, such as *Diapensia lapponicum, Vaccinium uliginosum,* and *Salix uva-ursi,* colonize soil polygons and windswept ridges (Antevs 1932, Bliss 1963, Harries 1966). Because pollen from *Artemisia* and Caryophyllaceae indicate that disturbed habitats were available for colonization, periglacial activity continued, and the mean annual temperature must have remained 4°–6°C below that of today.

Between 11 000 and 10 300 yr BP the pollen record indictates that shrubs such as *Juniperus, Salix,* and dwarf birch invaded the alpine zone. Moderation in the climate allowed the establishment of dwarf shrub heaths at this time. Other evidence of temperature moderation includes the opening of the basins in Franconia Notch which were open to sedimentation by 11 000 yr ago. If residual ice had filled Kinsman Pond and Lonesome Lake, it had melted or perhaps retreated into the valley bottom. During the 700 yr between 11 000 and 10 300 yr BP, pollen percentages and pollen influx of *Artemisia,* Caryophyllaceae, and other indicators of disturbance decline at all sites. The accumulation rate of inorganic matter also drops, indicating stabilization of the landscape. Periglacial activity was further reduced as temperatures rose closer to modern values; however, it probably did not cease until 10 300 yr BP. The modern alpine communities probably became established at this time.

10 300–9000 yr BP

Spruce and fir trees began growing at high elevations ≈10 300 yr BP, 1200 yr after they first grew in the valleys around Lost Pond and Mirror Lake (Fig. 11c). Before this vegetational change can be used as evidence of climatic change, other factors such as the migrational lag of species and soil development have to be eliminated.

One possible explanation for the 1200-yr lag in establishment of populations of spruce at high elevations could be a migrational delay in the arrival of the treeline species in central New Hampshire. The treeline species of spruce before 10 000 yr BP was probably black spruce. Evidence suggests, however, that it was present in the White Mountain region long before 10 300 yr BP. Watts (1979) believes that the three species of spruce migrated northward from Appalachia at different times. White spruce was reputedly the predominant species at the forest margin and the first to migrate north. Black spruce was present in the early white spruce forest, but red spruce migrated later, dominating the forests at the end of the spruce zone. Richard (1978) found pollen of both white and black spruce in sediments older than 11 400 yr BP at Mont Shefford, Quebec, 150 km north of the White Mountains. Needles with morphology similar to those found at high elevations have been found in sediments dating from 11 500 yr BP at Mirror Lake and 11 200 yr BP at Lost Pond (M. B. Davis and L. C. K. Shane, *personal communication*). Thus the lag in establishment of high-elevation spruce population is not explained by late migration of the high-elevation species into the central New Hampshire region.

Slow soil development at high elevations might also be used as an explanation for the delayed development of krummholz. Were several thousand years of plant succession needed before spruce and fir could colonize the summit boulder fields? This explanation is unlikely, for the White Mountains were completely overridden by the continental ice sheet, which left pockets of till at all elevations including the summit of Mt. Washington. Suitable soils for krummholz existed. Studies at Glacier Bay, Alaska (Cooper 1923, Lawrence et al. 1967), and Klutlan glacier, Yukon Territory (Rampton 1970, Birks 1980) indicate that under favorable climatic conditions spruce forests can develop on moraines <150 yr following the retreat of montane glaciers. In the White Mountains, vegetation studies by Flaccus (1959), Harries (1966), and Teeri (1968) show that treeline species can grown on nearly bare rock.

Climatic change is the most tenable explanation for the arrival of spruce and fir at the high-elevation sites around 10 300 yr BP. The underlying climatic factor determining the altitudinal species limit is most likely temperature (Daubenmire 1954). Wind, however, is the most important climatic factor determining the altitude of treeline and the distribution of alpine plant communities (Antevs 1932, Bliss 1963, Harries 1966). Above treeline, krummholz is usually found in sheltered sites (Tiffney 1972). Black spruce krummholz withstands the wind better than fir and can be found on exposed ridges swept free of snow in winter, as well as in more sheltered sites that have good snow packs in winter (Teeri 1968). The fact that Lake of the Clouds was surrounded by more extensive spruce than either Deer Lake Bog or Kinsman Pond may be explained by the higher, more wind-exposed location of Lake of the Clouds.

The climate 10 300 yr ago was marked by a dramatic increase in temperature of roughly 5°. Periglacial activity ceased. Temperatures rose from 4°–6°C below those of today to near modern values. The estimates of temperature at 10 300 yr BP range from 1° cooler to 0.5° warmer than today. If Lake of the Clouds had a climate similar to that found at 1700 m today, the altitudinal limit of krummholz, the mean annual temperature would have been 1° cooler than today. However, the black spruce that grew around the Lake of the Clouds 10 300 yr BP was more extensive than that growing around the lake today and undoubtedly more extensive than at the present-day species limit at 1700 m. Although spruce has the same altitudinal species limit as fir, it only forms krummholz patches at lower elevations near treeline (Harries 1966). The black spruce krummholz at Lake of the Clouds 10 300 yr ago was probably similar to that found today on the east slope of Mt. Washington at 1450 m, just above treeline. If so, the mean annual temperature was 0.5° warmer than today. No matter which estimate is correct, 10 300 yr BP is marked by a dramatic increase in temperature. The ecotone between the subalpine fir forest and alpine meadow was established close to its modern limit.

9000 yr BP to the present—record of climatic change and biotic interactions

Fluctuations in the altitudinal limits of white pine (*Pinus strobus*) and hemlock provide evidence of warmer temperatures at lower elevations in the White Mountains of New Hampshire from 9000 to 5000 yr

BP. Davis et al. (1980) estimate that the temperature was 2° warmer and that precipitation was reduced during this period. These climatic changes presumably affected high-elevation vegetation as well; however, the vegetation changes were less obvious at high elevations than at low. The subalpine vegetation itself drastically altered the microclimate of the slopes, increasing the magnitude of climate change necessary to affect the subalpine fir forest and the position of treeline. The ecotones between the subalpine spruce–fir and fir forest, and the fir forest and the alpine meadow, have not changed altitude much over the last 10 000 yr and do not appear to be sensitive to temperature change. The most important climatic factors in the distribution of vegetation in these zones are wind exposure (snow depth) and atmospheric moisture (Bliss 1963, Harries 1966). Alpine areas in the White Mountains with strong winds, cool temperatures, fog, and high atmospheric moisture are more similar to areas in the eastern Arctic than to the sunnier and drier alpine areas of the west (Bliss 1964). Upright trees of undiminished stature are found at treeline in the Rockies and Sierra Nevada, but are never found (Griggs 1946) and probably never have grown at the forest limit in the White Mountains.

Fir may have been more extensive at Lake of the Clouds from 10 000 to 5000 yr BP than it is now (Fig. 11b). Besides abundant fir and spruce needles, the maximum numbers of Ericaceae, Cyperaceae, and *Potentilla tridentata* seeds are found during this interval. The seeds indicate that open patches still existed, but extensive krummholz that traps snow (Tiffney 1972) would have created favorable microclimates for other species that probably grew around the site. Treeline may have been close to Lake of the Clouds. The climate there may have been similar to that of Eagle Lake Bog at 1280 m or the Horn at 1200 m. If so, the mean annual temperature was 2° higher.

The paleoecological record at Kinsman Pond can also be interpreted as indicating warmer temperatures from 9000 to 5000 yr BP. It contains fewer fir macrofossils and lower percentages of fir pollen during this interval than after it. If fir populations were actually less abundant in the subalpine forests, then spruce, paper birch, or some other species characteristic of lower elevations may have been more abundant. Spruce did not increase in abundance at high elevations from 9000 to 5000 yr BP. At Kinsman Pond, no spruce macrofossils were found during a 1500-yr period from 8500 to 6900 yr BP, and pollen percentages remained low throughout the first half ot the Holocene. Possibly there was an increase in paper birch in the forest, but the paleoecological record does not provide evidence for this. Green alder definitely grew around Kinsman Pond and Eagle Lake Bog until 5000 yr BP. The ecological significance of the persistant populations of *Alnus crispa* in Franconia Notch is uncertain. It grows on sheltered east slopes of the Presidential Range today. In areas where large drifts of snow accumulate, birch krummholz, birch trees, and *Alnus crispa* scrub replace fir and black spruce at treeline (Harries 1966). Similar conditions could have existed in Franconia Notch. The delayed establishment of species at Franconia Notch could also explain why the sites there contain a different vegetation history from Lake of the Clouds and Deer Lake Bog.

The spruce–fir phase of the subalpine forest from 750 to 1150 m is a relatively recent vegetation type in the White Mountains, but grew in mountains in southern Quebec 6000 yr BP (Webb et al. 1983). Maps from eastern North America (Gaudreau and Webb 1985, Webb 1986) show southward movement of the 5% spruce isopoll in the last few thousand years. Jackson (1983) theorizes that red spruce populations either migrated southward or expanded from small scattered populations at this time. I suspect that if red spruce populations did occur in the White Mountains before 10 000 yr ago, remanent populations must have persisted in favorable sites. Low but continuous percentages of spruce pollen support this view, although it is not known which species of spruce are represented in the pollen record. The failure of spruce to expand their high-elevation populations during the mid-Holocene warm interval is puzzling. Why did spruce not extend its altitudinal range upward as did white pine and hemlock? At the lower altitudinal range limit of spruce, temperature and moisture must play an important role in the competitive interactions of red spruce and northern hardwood species. The upper range limit of red spruce is determined by its ability to compete with fir in an environment subject to frequent wind disturbance. Above 1200 m, strong winds give fir a competitive advantage. Fir produces more seeds at more frequent intervals and has more vigorous seedlings which develop root systems faster than does spruce. Thus fir is better adapted to short-term disturbance (Sprugel 1976).

Perhaps warmer temperatures than today's gave northern hardwood species a competitive advantage over spruce in the lower elevations of the current range for spruce, while wind disturbance, which determines the upper limit of spruce range, remained unchanged. Thus the climatic conditions were unfavorable for the spread of red spruce populations from favorable sites until cooling after 3000 yr BP enabled them to expand at the expense of hardwoods at elevations between 750 and 1220 m.

Some evidence exists from high-elevation sites for climatic cooling after 5000 yr BP. At Lake of the Clouds, pollen percentages of alpine indicators increased (Fig. 11). This increase, plus a decrease in the number of needle fragments from spruce and fir, indicates less extensive krummholz and more open conditions around Lake of the Clouds. The development of peat deposits in the alpine zone coincides roughly with the expansion of spruce populations at lower elevations. Soil deterioration may account for some of the vegetation changes;

however, soils on mountain peaks have never been good. Much of the till left by continental glaciation on the summits was probably eroded before 10 300 yr BP, before the slopes were stabilized by vegetation. Modern vegetation studies (Bliss 1963, Harries 1966, Reiners and Lang 1979) show that although vegetation greatly affects the development of soils, the soils of the mountain peaks have little influence on alpine vegetation. The retreat of krummholz and the development of peat are likely the result of a cooler/moister climate that developed after 3000 yr BP.

The best physical evidence for climatic change is the development of peat at Star Lake Bog in the alpine zone of the northern Presidentials roughly 2000 yr ago (Spear 1981). As depth of peat increased, water was impounded, forming the shallow lake. However, there is no indication of renewed periglacial activity in the alpine zone during the last 2000–3000 yr BP. Goldthwait (1976) suggests that the temperature fell 4° and frost sorting occurred down to an elevation of 1700 m after 3000 yr BP. The pollen record shows no increase in the species indicative of periglacial disturbance. If periglacial activity has increased in the last 3000 yr, it must be at elevations too far above Lake of the Clouds to leave a paleoecological record. Further study of alpine peat deposits could yield important information on Neoglacial events in the White Mountains.

A final indication of cooling during the last 5000 yr occurs in the record of subalpine forests at Kinsman Pond. The number of fir macrofossils increases significantly after 5000 yr BP, suggesting that fir may have become more abundant in the forest at that time. Cooler and moister conditions might have enabled fir to expand its populations at its lower altitudinal limit when the subalpine forest as we know it today, with its spruce–fir phase, and fir phase developed.

In contrast to the continual changes in the vast lowland forests surrounding the White Mountain peaks, the high elevations have been remarkably stable. Changes in the lowland forest have had virtually no impact on the subalpine fir forest and alpine meadow. This knowledge gives insight into the dynamics of high-elevation vegetation and has allowed me to deduce the importance of climatic and biotic interactions in structuring the subalpine forest. Further detailed paleoecological study of White Mountain sites over a range of altitudes, especially at low and mid elevations, will allow us to reconstruct the history of vegetation in the mountains and to ask questions about the factors contributing to that history.

ACKNOWLEDGMENTS

I thank R. K. Olson, J. Ford, D. Mitten, M. B. Davis, and especially J. Harlan for their assistance with the fieldwork, I. W. Heathcote, C. K. Shane, and R. E. Moeller for help with laboratory analyses, R. C. Bright and W. A. Watts for help with plant macrofossil identification, E. J. Cushing for use of his computer program POLDATA version 5 to plot pollen and macrofossil diagrams, and K. Hadden and K. Gajewski for help in the preparation of the manuscript. I would like to thank M. B. Davis for introducing me to the White Mountains and for giving this project patient attention in each stage of its development. This research was funded by a National Science Foundation grant to M. B. Davis and by a Geological Society of America research grant to me.

LITERATURE CITED

Allison, T. D., R. E. Moeller, and M. B. Davis. 1986. Pollen in laminated sediments provides evidence for a mid-Holocene forest pathogen outbreak. Ecology **67**:1101–1105.

Anderson, R. S., R. B. Davis, N. G. Miller, and R. Stuckenrath. 1986. History of Late and Post-Glacial vegetation and disturbance around Upper South Branch Pond, Northern Maine. Canadian Journal of Botany **64**:1977–1986.

Antevs, E. 1932. Alpine Zone of Mt. Washington Range. Merril and Webber, Auburn, Maine, USA.

Bernabo, J. C., and T. Webb III. 1977. Changing patterns in the Holocene pollen record from northeastern North America: a mapped summary. Quaternary Research **8**:13–26.

Billings, M. P. 1937. Regional metamorphism of the Littleton-Moosilauke area, New Hampshire. Geological Society of American Bulletin **48**:463–566.

Billings, M. P., K. Fowler-Billings, C. A. Chapman, R. W. Chapman, and R. P. Goldthwait. 1946 (reprinted 1972). The geology of the Mt. Washington Quadrangle, New Hampshire. State of New Hampshire Department of Resources and Economic Development, Concord, New Hampshire, USA.

Birks, H. J. B. 1976. Late-Wisconsinan vegetational history at Wolf Creek, central Minnesota. Ecological Monographs **46**:395–429.

———. 1977. Modern pollen rain and vegetation of the moraines of the Klutlan Glacier, Yukon Territory. Canadian Journal of Botany **55**:2367–2382.

———. 1980. The present flora and vegetation of the moraines of the Klutlan Glacier, Yukon Territory, Canada: a study in plant succession. Quaternary Research **14**:60–86.

Black, R. A., and L. C. Bliss. 1978. Recovery of *Picea mariana-Vaccinium uliginosum* forests after burning near Inuvik, Northwest Territories, Canada. Canadian Journal of Botany **56**:2020–2030.

Bliss, L. C. 1963. Alpine plant communities of the Presidential Range, New Hampshire. Ecology **44**:678–697.

———. 1964. Plant productivity in alpine microenvironments on Mt. Washington, New Hampshire. Ecological Mongraphs **36**:125–154.

Bocher, T. W. 1954. Oceanic and continental vegetational complexes in southwest Greenland. Meddelelser om Grönland. **148**(1).

Bormann, F. H., and A. P. Nelson. 1963. Vegetation zonation in northcentral New England. American Institute of Biological Sciences-Ecological Society of America Field Trip.

Bradley, D. C. 1981. Late Wisconsinan mountain glaciation on the northern Presidential Range, New Hampshire. Arctic and Alpine Research **13**:319–327.

Brown, J. W. 1958. Forest history of Mount Moosilauke. Appalachia **24**:221–233.

Brubaker, L. B. 1975. Postglacial forest pattern associated with till and outwash in northcentral Upper Michigan. Quaternary Research **5**:499–527.

Bryan, M. S. 1954. Interglacial pollen spectra from Greenland. Danmarks Geologiske Undersögelse, Series 2, Number **80**:65–72.

Buchanan, C. 1975. A biological study of the Lakes of the Clouds (with emphasis on the zooplankton). Thesis. University of New Hampshire, Durham, New Hampshire, USA.

Chittenden, A. K. 1905. Forest conditions of northern New

Hampshire. Bureau of Forestry Bulletin Number 55. United States Department of Agriculture, Washington, D.C., USA.

Copper, W. S. 1923. The recent ecological history of Glacier Bay, Alaska. Ecology **4**:93–128, 223–246, 355–365.

Crocker, R. L., and J. Major. 1955. Soil development in relation to vegetation and surface age at Glacier Bay, Alaska. Journal of Ecology **43**:427–448.

Cunningham, F. E. 1957. A seed key for five Northeastern birches. Journal of Forestry **55**:844–845.

Cushing, E. J. 1963. Late-Wisconsinan pollen stratigraphy in east-central Minnesota. Dissertation. University of Minnesota, Minneapolis, Minnesota, USA.

Cwynar, L. C., E. Burden, and J. H. McAndrews. 1979. An inexpensive sieving method for concentrating pollen and spores from fine-grained sediments. Canadian Journal of Earth Science **16**:1115–1120.

Daubenmire, R. 1954. Alpine timberlines in the Americas and their interpretation. Butler University Botanical Studies **11**:119–136.

Davis, M. B. 1976. Pleistocene biogeography of temperate deciduous forests. Geoscience and Man **XIII**:13–26.

———. 1981*a*. Outbreaks of forest pathogens in forest history. Pages 216–227 *in* Proceedings of the IV International Palynological Conference (1976–1977). Volume 3. Birbal Sahni Institute of Paleobotany, Lucknow, India.

———. 1981*b*. Quaternary history and the stability of forest communities. Pages 132–153 *in* D. C. West, H. H. Shugart, and D. B. Botkin, editors. Forest succession—concepts and applications. Springer-Verlag, New York, New York, USA.

———. 1983. Quaternary history of deciduous forests of eastern North America and Europe. Annals of the Missouri Botanical Garden **70**:550–563.

———. 1985. History of the vegetation in the Mirror Lake Watershed. Pages 53–64 *in* G. E. Likens, editor. An ecosystem approach to aquatic ecology—Mirror Lake and its environment. Springer-Verlag, New York, New York, USA.

Davis, M. B., and D. B. Botkin. 1985. Sensitivity of cool-temperate forests and their fossil pollen record to rapid temperature change. Quaternary Research **23**:327–340.

Davis, M. B., and M. S. (J.) Ford. 1982. Sediment focusing in Mirror Lake, New Hampshire. Limnology and Oceanography **27**:137–150.

Davis, M. B., R. E. Moeller, and M. S. (J.) Ford. 1984. Sediment focusing and pattern influx. Pages 261–293 *in* E. Y. Haworth and J. W. G. Lund, editors. Lake sediments and environmental history: studies in paleolimnology and paleoecology in honour of Winifred Tutin. University of Leicester Press, Leicester, England.

Davis, M. B., R. W. Spear, and L. C. K. Shane. 1980. Holocene climate of New England. Quaternary Research **14**: 240–250.

Davis, M. B., R. W. Spear, and L. C. K. Shane. 1989. The history of low and mid-elevation vegetation of the White Mountains in New Hampshire. In preparation for submission to Ecological Monographs.

Davis, R. B., and G. L. Jacobson, Jr. 1985. Late-glacial and early Holocene landscapes in northern New England and adjacent areas of Canada. Quaternary Research **23**:341–368.

Dean, W. E., Jr. 1974. Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss-on-ignition: comparison with other methods. Journal of Sedimentary Petrology **44**:242–248.

Delcourt, P. A. 1978. Quaternary vegetational history of the Gulf Coastal Plain. Dissertation. University of Minnesota, Minneapolis, Minnesota, USA.

Dixon, W. J. 1973. BMD Biomedical computer programs. University of California Press, Berkeley, California, USA.

Drury, W. H., Jr. 1969. Plant persistence in the Gulf of St. Lawrence. Pages 105–148 *in* K. N. H. Greenidge, editor. Essays in plant geography and ecology. Nova Scotia Museum, Halifax, Nova Scotia, Canada.

Dunwiddie, P. W. 1987. Macrofossil and pollen representation of coniferous trees in modern sediments from Washington. Ecology **68**:1–11.

Durrell, L. W. 1916. Notes on some North American conifers based on leaf characters. Proceedings of the Iowa Academy of Science **23**:519–582.

Elliott, D. L. 1979. The stability of the northern Canadian tree limit: current regenerative capacity. Dissertation. University of Colorado, Boulder, Colorado, USA.

Eskenasy, M. D. 1979. The origin of the King Ravine rock glacier, Presidential Range, New Hampshire. Geological Society of America Abstract Programs **11**:11.

Faegri, K., and J. Iversen. 1975. Textbook of pollen analysis. Hafner, New York, New York, USA.

Fenneman, N. M. 1938. Physiography of eastern United States. McGraw-Hill, New York, New York, USA.

Fernald, M. L. 1925. Persistence of plants in unglaciated areas of boreal America. Memoirs of the American Academy of Arts and Sciences **15**:237–342.

———. 1950. Gray's manual of botany. American Book Company, New York, New York, USA.

Flaccus, E. 1959. Revegetation of landslides in the White Mountains of New Hampshire. Ecology **40**:692–703.

Fredskild, B. 1973. Studies in the vegetational history of Greenland: palaeobotanical investigations of some Holocene lake bog deposits. Meddelelser om Grönland **198**(4).

Gaudreau, D. C., and T. Webb III. 1985. Late-Quaternary pollen stratigraphy and isochrone maps for the northeastern United States. Pages 247–280 *in* W. M. Bryant and R. G. Holloway, editors. Pollen records of Late-Quaternary North America sediments. American Association of Stratigraphic Palynologists Foundation, Dallas, Texas, USA.

Gerath, R. F., and B. K. Fowler. 1982. Discussion of "Late Wisconsinan mountain glaciation in northern Presidential Range, New Hampshire" by Dwight C. Bradley. Arctic and Alpine Research **14**:369–370.

Goldthwait, R. P. 1940. Geology of the Presidential Range. New Hampshire Academy of Science Bulletin **1**:5–41.

———. 1970. Mountain glaciers of the Presidential Range in New Hampshire. Arctic and Alpine Research **2**:85–102.

———. 1976. Past climates on "the hill." Mt. Washington Observatory Bulletin. March and June issues.

Goldthwait, R. P., and D. M. Mickelson. 1982. Glacier Bay: a model for the deglaciation of the White Mountains in New Hampshire. Pages 167–171 *in* G. J. Larsen and B. D. Stone, editors. Late-Wisconsinan deglaciation of New England. Kendall-Hunt, Dubuque, Iowa, USA.

Griggs, R. F. 1942. Indications as to climatic changes from the timberline of Mount Washington. Science **95**:515–519.

———. 1946. The timberlines of northern America and their interpretation. Ecology **27**:275–289.

Harries, H. 1966. Soils and vegetation in the alpine and the subalpine belt of the Presidential Range. Dissertation. Rutgers University, New Brunswick, New Jersey, USA.

Hustich, I. 1953. The boreal limits of conifers. Arctic **6**: 149–162.

———. 1966. On the forest-tundra and the northern tree-lines. Annales Universitatis Turkuensis Series A2, **36**:7–47.

Ives, J. D. 1974. Biological refugia and the nunatak hypothesis. Pages 605–636 *in* J. D. Ives and R. G. Barry, editors. Arctic and alpine environments. Methuen, London, England.

———. 1977. Pollen separation of three North American birches. Arctic and Alpine Research **9**:73–80.

Jackson, S. T. 1983. Late-glacial and Holocene vegetational changes in the Adirondack Mountains (New York): a macrofossil study. Dissertation. Indiana University, Bloomington, Indiana, USA.

Jacobson, G. L. Jr. 1979. The paleoecology of white pine (*Pinus strobus*) in Minnesota. Journal of Ecology **67**:697–726.

Jowsey, P. C. 1965. An improved peat sampler. New Phytologist **65**:245–248.

Lawrence, D. B., R. E. Schoenike, A. Quispel, and G. Bond. 1967. The role of *Dryas drummondii* in vegetation development following ice recession at Glacier Bay, Alaska, with special reference to its nitrogen fixation by root nodules. Journal of Ecology **55**:793–813.

Likens, G. E., and M. B. Davis. 1975. Post-glacial history of Mirror Lake and its watershed in New Hampshire, U.S.A.: an initial report. Internationale Vereinigung für theoretische und angewandte Limnologie, Verhandlungen **19**:982–993.

Lovett, G. M., W. A. Reiners, and R. K. Olson. 1982. Cloud droplet deposition in subalpine balsam fir forests: hydrological and chemical inputs. Science **218**:1303–1304.

Marchand, P. J., and B. F. Chabot. 1978. Winter water relations of tree-line plant species on Mt. Washington, New Hampshire. Arctic and Alpine Research **10**:105–116.

Marchand, P. J., and D. A. Roach. 1980. Reproductive strategies of pioneering alpine species; seed production, dispersal, and germination. Arctic and Alpine Research **12**: 137–146.

McIntosh, R. P., and R. T. Hurley. 1964. The spruce–fir forests of the Catskill Mountains. Ecology **45**:314–326.

Miller, N. G., and G. G. Thompson. 1979. Boreal and western North American plants in the late Pleistocene of Vermont. Journal of the Arnold Arboretum **60**:167–218.

Monahan, R. S. 1931. Timberline study: a general review of influences affecting altitudinal limits of tree growth with special reference to their applications on the Presidential Range of the White Mountains, New Hampshire. Thesis. Yale University, New Haven, Connecticut, USA.

Müller-Dombois, D., and H. Ellenberg. 1974. Aims and methods of vegetation ecology. John Wiley and Sons, New York, New York, USA.

Myers, O., Jr., and F. H. Bormann. 1963. Phenotypic variation in *Abies balsamea* in response to altitudinal and geographic gradients. Ecology **44**:429–436.

National Oceanic and Atmospheric Administration. 1975*a*. Mt. Washington table of normals means and extremes–revised 1975. *In* J. A. Ruffner, editor. Climates of the States. Gale Research Company, Book Tower, Detroit, Michigan, USA.

———. 1975*b*. Climate of Pinkham Notch, New Hampshire. Climatography of the United States Number 20. National Climatic Center, National Oceanic and Atmospheric Administration, Asheville, North Carolina, USA.

Oosting, H. J., and W. D. Billings. 1951. A comparison of virgin spruce–fir forest in the northern and southern Appalachian system. Ecology **32**:84–103.

Polunin, N. 1948. Botany of the Canadian eastern arctic Part III, vegetation and ecology. Bulletin Number 204, Biological Series Number 32, National Museum of Canada, Ottawa, Ontario, Canada.

Rampton, V. 1970. Neoglacial fluctuations of the Natazhat and Klutlan Glaciers, Yukon Territory, Canada. Canadian Journal of Earth Sciences **7**:1276–1363.

Reiners, W. A., and G. E. Lang. 1979. Vegetational patterns and processes in the balsam fir zones, White Mountains, New Hampshire. Ecology **60**:403–417.

Richard, P. 1977. Vegetation tardiglaciaire du Quebec meridional et implications paleoclimatiques. Géographie Physique Quaternaire **31**:161–176.

———. 1978. Histoire tardiglaciaire et postglaciaire de la vegetation au Mount Shefford, Quebec. Géographie Physique Quaternaire **32**:81–93.

Ritchie, J. C. 1974. Modern pollen assemblages near the arctic treeline, Mackenzie Delta region, Northwest Territories. Canadian Journal of Botany **52**:381–396.

Siccama, T. G. 1974. Vegetation, soil and climate of the Green Mountains of Vermont. Ecological Monographs **44**: 325–349.

Spear, R. W. 1981. The history of high-elevation vegetation in the White Mountains of New Hampshire. Dissertation. University of Minnesota, Minneapolis, Minnesota, USA.

Sprugel, D. G. 1976. Dynamic structure of wave-regenerated *Abies balsamea* forests in the northeastern United States. Journal of Ecology **64**:889–905.

Sprugel, D. G., and F. H. Bormann. 1981. Natural disturbance and the steady state in high-altitude balsam fir forests. Science **211**:390–393.

Stockmarr, J. 1971. Tablets with spores used in absolute pollen analysis. Pollen et Spores **13**:615–621.

Teeri, J. A. 1968. The ecology of subalpine black spruce in New England. Thesis. University of New Hampshire, Durham, New Hampshire, USA.

Tiffney, W. N., Jr. 1972. Snow cover relationships of White Mountains alpine plants. Dissertation. University of New Hampshire, Durham, New Hampshire, USA.

Tranquillini, W. 1979. Physiological ecology of the alpine timberline. Springer-Verlag, New York, New York, USA.

Troels-Smith, J. 1955. Karakterising of lose jordater. Danmarks Geologiske Undersögelse, Series **4**.

Wardle, P. 1974. Alpine timberlines. Pages 371–402 *in* J. D. Ives and R. G. Barry, editors. Arctic and alpine environments. Methuen, London, England.

Watts, W. A. 1979. Late Quaternary vegetation of central Appalachia and the New Jersey coastal plain. Ecological Monographs **49**:427–469.

Webb, T., III. 1982. Temporal resolution in Holocene pollen data. Pages 569–572 *in* Proceedings of the Third North American Paleontological Convention, Volume 2. Busines and Economic Service, Toronto, Ontario, Canada.

———. 1986. Vegetational change in eastern North America from 18 000 to 500 yr B.P. Pages 63–69 *in* C. Rosenzweig and R. Dickinson, editors. Climate-vegetation interactions. Office for Interdisciplinary Earth Studies, University Corporation for Atmospheric Research, Boulder, Colorado, USA.

Webb, T., III, P. J. H. Richard, and R. J. Mott. 1983. A mapped history of Holocene vegetation in southern Quebec. Syllogeus **49**:273–336.

Williams, C. R., and M. P. Billings. 1938. Petrology and structure of the Franconia Quadrangle, New Hampshire. Geological Society of America Bulletin **49**:1011–1044.

Wright, H. E., E. J. Cushing, and D. A. Livingstone. 1965. Coring devices for lake sediments. Pages 494–520 *in* B. Kummel and D. M. Raup, editors. Handbook of paleontological techniques. Freeman, San Francisco, California, USA.

Arctic, Antarctic, and Alpine Research, Vol. 41, No. 3, 2009, pp. 362–372

Evidence of Climate Change Declines with Elevation Based on Temperature and Snow Records from 1930s to 2006 on Mount Washington, New Hampshire, U.S.A.

*Thomas M. Seidel**
Douglas M. Weihrauch†
Kenneth D. Kimball†§
Alexander A. P. Pszenny‡*
*Rita Soboleski**
Elena Crete‡ and
Georgia Murray†

*Mount Washington Observatory, P.O. Box 2310, North Conway, New Hampshire 03860, U.S.A.
†Research Department, Appalachian Mountain Club, P.O. Box 298, Gorham, New Hampshire 03581, U.S.A.
‡Institute for the Study of Earth, Ocean and Space, University of New Hampshire, Durham, New Hampshire 03824, U.S.A.
§Corresponding author: kkimball@outdoors.org

Abstract

Mount Washington, New Hampshire, has the longest northeastern U.S. mountain climatological record (1930s to present), both at the summit (1914 m) and at Pinkham Notch (612 m). Pinkham's homogenized daily temperature exhibits annual (mean = +0.07°C/decade, p = 0.07; min = +0.11°C/decade, p = 0.01), winter (min = +0.18°C/decade, p = 0.07), spring (max = +0.13°C/decade, p = 0.10), and summer (min = +0.11°C/decade, p = 0.01) warming trends. Though suggesting annual, winter, and spring warming (0.05 to 0.12°C/decade), mean summit temperature trends were not significant. Pinkham shows no significant change in date of first and last snow; however, the summit does but its period of record is shorter. Onset of continuous snow cover has not changed significantly at either site. Thawing degree days trended earlier at the summit (2.8 days/decade; p = 0.01) and Pinkham Notch (1.6 days/decade, $p < 0.01$), but end of continuous snow cover trended significantly earlier (1.6 days/decade; p = 0.02) only at Pinkham. Growing degree days showed no significant trends at either location. Pinkham exhibits more climatic change than the summit but less than regional lower elevations. Thermal inversions and high incidence of cloud fog commonly at or above the regional atmospheric boundary layer may explain the summit's resistance to climate warming. Caution is needed when extrapolating climate change trends from other mountains or proximate lower elevation climate data to upper elevations.

DOI: 10.1657/1938-4246-41.3.362

Introduction

Evidence of climate warming and reductions in seasonal snow cover on global (IPCC, 2007), hemispheric (Dye, 2002), and regional scales (Hayhoe et al., 2007) are numerous. Paralleling reported temperature increases are indices pointing to an earlier start to the growing season over the northern hemisphere (Schwartz et al., 2006). Northern hemisphere snow cover extent has decreased by 7.5 ± 3.5% during March and April (Lemke et al., 2007), and snow pack is decreasing in the mountains of the western United States with implications for water supply as well as entire ecosystems (Mote et al., 2005). However, there is a growing body of evidence that suggests considerable variability in climatic trends among mountains ranges. For example, mountain glaciers in Glacier National Park, Montana, U.S.A., are shrinking (Hall and Fagre, 2003), while glaciers on Mount Shasta, California, U.S.A., are expanding, despite a regional warming over the past half century (Howat et al., 2007).

Temperature fluctuations during the last century at high elevation sites around the world also exhibit horizontal and vertical variability (Diaz and Bradley, 1997). Europe (particularly western Europe), and parts of Asia displayed the strongest high-altitude warming during the period of record due primarily to increases in daily minimum temperature. Within central Europe's mountains temperature patterns trend warmer, but again with spatial and altitudinal variations (Weber et al., 1997).

Long-term, instrumented climatological measurements of the mountainous regions of the world are not numerous, and in eastern North America are relatively scarce. Two recent papers from New England show a warming at the summit of Mount Washington, New Hampshire, U.S.A., from 1935 to 2003 (Grant et al., 2005), an overall decrease in mean dew point temperature from 1935 to 2004, and an increase in annual fog frequency (Seidel et al., 2007). Other regional montane studies have focused on shorter time scale microclimatic studies (e.g. Friedland et al., 1992, 2003) or used longer temperature data sets for other purposes (e.g. lapse rate calculation; Richardson et al., 2004). Regional assessments have shown New England to be heterogeneously warming (Keim et al., 2003; Trombulak and Wolfson, 2004; Hayhoe et al., 2007).

New England precipitation studies show changes in the winter and spring hydrological records at lower elevations: an increase in the rain/snow ratio (Huntington et al., 2004), earlier dates for center-of-volume flow in rivers affected by snowmelt (Hodgkins et al., 2003), amount and timing of ice-affected river flow (Hodgkins et al., 2005), a decrease in the number of days with snow on the ground (Burakowski et al., 2007), and earlier springs (5 days) by seasonal evapotranspiration (Czikowsky and Fitzjarrald, 2004). Changes in snow cover amount and duration have many impacts in the Northeast including changes in nutrient cycling, mammalian species composition shifts (Carroll, 2007), and human recreation (Hamilton et al., 2007).

Snow cover patterns can affect alpine plant community composition (Walker et al., 1999; Bjork and Molau, 2007), physiology (Starr et al., 2000), phenology (Inouye and McGuire, 1991; Kudo, 1991; Huelber et al., 2006), and population genetics (Hirao and Kudo, 2004) and are an important factor in plant community distribution on Mount Washington (Bliss, 1963;

1523-0430/09 $7.00

Sardinero, 2000). Temperature also impacts arctic and alpine plant communities. Warming can lead to a shrubbier composition in alpine (Cannone et al., 2007) and tundra vegetation (Arft et al., 1999; Walker et al., 1999), and warming trends have been linked to range restrictions in alpine species at lower elevations (Pauli et al., 2007) and latitudes (Lesica and McCune, 2004). Seasonal increases in temperature reduced snow cover depth and duration in the Swiss alps, increasing the length of the growing season and causing earlier flowering (Keller et al., 2005).

Future predictions of climatic change and impacts on mountain ecosystems are frequently based on the most proximate low elevation data or on extrapolations from other mountain regions. But using surrogate climatic data to describe potential responses by mountain biota can result in compromised conclusions. For example, *Picea rubens* tree-ring growth on Mount Washington correlates more closely with the Mount Washington mid-elevation and summit temperatures than temperatures from low-elevation stations within a 40 km radius (Kimball and Keifer, 1988). Richardson et al. (2004) demonstrated that though general elevation–mean annual temperature relationships held across their northeastern U.S. mountain study sites, there was significant variation in air temperature lapse rates up the sides of the mountains and the pattern of variation was not consistent among mountains.

Mount Washington has the most complete northeastern U.S. instrumented temperature and snow records for both mid and high elevations, ranging from the 1930s to the present. In this study we compare seasonal and annual temperature trends, growing and thawing degree-day trends, and trends in two indices of snow season length for the summit and for Pinkham Notch, a mid-elevation site on the eastern side of the Mount Washington. Our study modifies previous conclusions from this region by Grant et al. (2005), discusses New England mountain climatic trends in relation to alpine ecosystems, and compares our observed trends with other high elevation sites in the world.

Site Location and Data Collection

BASIC DESCRIPTION OF REGION

Mount Washington (44°16′N, 71°18′W), the highest point (1914 m a.s.l.) in the northeastern United States, is part of the Presidential Range of the White Mountains of New Hampshire, a northern section of the Appalachian Mountains. The Presidential Range contains 2748 ha of contiguous alpine and subalpine vegetation surrounded by spruce and fir boreal forest with northern hardwood species at lower elevations. Treeline occurs at relatively low elevations and ranges from 1100 to 1700 m (Kimball and Weihrauch, 2000). The treeline-alpine ecotone is correlated with exposure to clouds and wind, slope, and aspect (Reiners and Lang, 1979; Kimball and Weihrauch, 2000).

The regional atmospheric mixing-layer typically is 1100–1500 m a.s.l. (Freedman et al., 2001), and on Mount Washington it exhibits diurnal and vertical migration that is influenced by daily solar heating, changing weather fronts, and the complex terrain of the surrounding mountain region. Above this mixing-layer is the "free atmosphere" where the winds approaching the Presidential Range are more geostrophic.

The climatological records are from the Mount Washington Observatory located on the rocky windswept summit of the mountain and from the Appalachian Mountain Club's (AMC) Pinkham Notch Visitor Center (612 m a.s.l.) in the upper extent of the northern hardwoods on the eastern side of the mountain (Fig. 1).

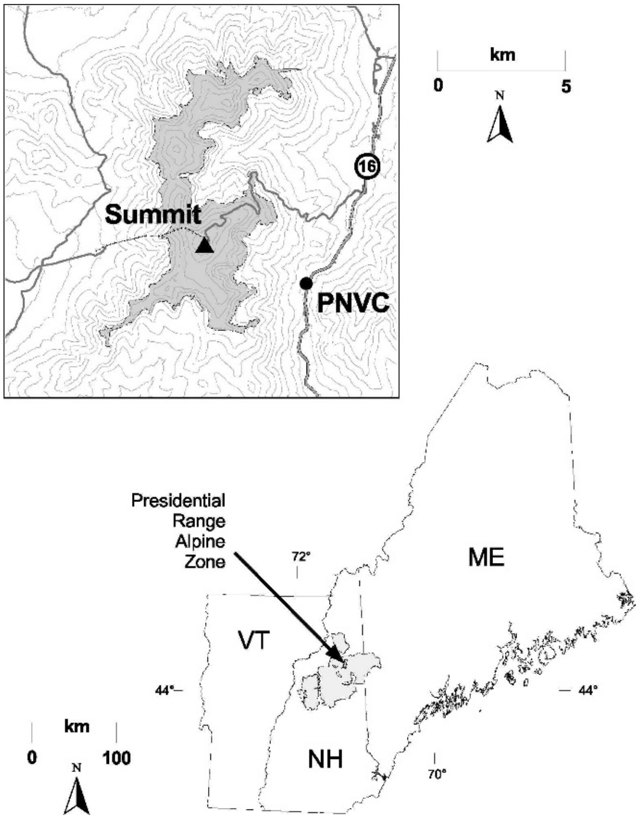

FIGURE 1. The White Mountain National Forest, outlined in black, in northern New Hampshire, U.S.A. The blowup shows the Presidential Range with contour intervals of 90 m, the alpine zone (gray shading), the summit of Mount Washington (triangle) and Pinkham Notch Visitor Center (circle) adjacent to NH route 16 (thick line). Additional landmarks are the Mount Washington Auto Road on the east side of the mountain and the Mount Washington Cog Railway on the west side of the mountain.

SUMMIT

The Mount Washington Observatory has maintained a staffed meteorological station atop the mountain since 1932. Consistent data records allow analysis of hourly temperature observations since 1935 and 6-hourly synoptic observations, including snow depth, since 1949. From 1932 through 1937 the observatory building occupied several locations on the summit, with additional moves in 1937 and 1980 (Grant et al., 2005). The area immediately surrounding the Observatory is felsenmeer.

Hourly temperature (T) observations are taken at the end of each hour using either a mercury thermometer mounted in a sling psychrometer or an alcohol-in-glass minimum thermometer. Daily minimum and maximum temperatures (min/max) are determined by reading the alcohol-in-glass minimum thermometer and the mercury maximum thermometer, respectively (Grant et al., 2005).

Snowfall and other hydrometeors are observed by trained and certified observers according to U.S. National Weather Service (NWS) standards. Snow depth is estimated by the observer based on a visual spatial average encompassing the summit at the time of the synoptic observation. Attempts are made to avoid overweighting due to drifts and wind-cleared zones. Snow depth is reported in 1.27 cm intervals, including trace.

Summit snow data from 1949 to 2004 were digitized as part of the same effort that produced the hourly and 6-hourly temperature and humidity data analyzed by Grant et al. (2005) and Seidel et al. (2007). For this study, Mount Washington Observatory staff updated the temperature and snow digital record to include 2005–2006. Due to the choice to define snow season criteria (see **Methods**) without using explicit snow depth, the snow data were not subjected to the time-intensive quality assurance checks for digitization errors.

PINKHAM

The Appalachian Mountain Club (AMC) has operated a NWS cooperative (COOP) station [COOP #276818, NCDC station ID 20018701, NWS Location ID HGMN3] at its Pinkham Notch Visitor Center (elevation 612 m a.s.l., 44°16′N, 71°15′W) in the White Mountain National Forest since 1930. The most recent National Climate Data Center (NCDC) station metadata (10 January 2008) describes the station thus: "Narrow north-south notch in mountainous and heavily wooded area."

Daily minimum and maximum temperatures were observed using standard liquid in glass (LIG) minimum and maximum thermometers in a cotton region shelter until switching to the maximum-minimum temperature system (MMTS), a thermistor in a plastic shelter (Quayle et al., 1991). The LIG instruments were moved due to construction in October 1966 and thermometer location changed again with the switch to the MMTS. Copies of NWS inspection reports give the installation date of the MMTS (number #4296) as 1 May 1986, while the NCDC Multi-network Metadata System gives the transition date as 1 October 1987. It is unknown when observers began to use the MMTS during this 17-month window.

The MMTS sensor is accurate to ±0.5°C, the temperature is displayed to the nearest 0.1 °F, recorded to the nearest whole degree Fahrenheit, and is calibrated annually against a specially maintained reference instrument (NCDC, 2006). The standard LIG-to-MMTS temperature adjustments (annual max −0.4°C, min +0.3°C, mean −0.1°C) were not applied to the LIG temperature data because the U.S. Historical Climatology Network (USHCN) data documentation (Quayle et al., 1991) and other studies do not recommend these adjustments for individual stations (Pielke et al., 2002; Hubbard et al., 2004). Pinkham inspection reports through 1990 (the latest stored on site) show positive scores for site location but occasional trouble with the LIG thermometers (e.g. 10/13/1976: "replaced min. therm.; was 2.54 different from max"). Due to the station changes described above the temperature data were subjected, as described below, to inhomogeneity tests for potential change points.

Temperature is nominally observed and recorded at 0700 Local Standard Time (LST); the only recorded change in schedule was from observations on the hour to 10 minutes past the hour. However, observations are often taken at 0600 LST (Michael Walsh, personal communication). Data have not been analyzed for a potential time of observation bias.

Both snowfall and snow depth are nominally measured at 0700 LST in a section of mature hardwoods in the center of the campus. New snowfall is measured once a day using a stake mounted to a 25.4 cm square piece of wood that is collected, measured, cleared and replaced. Snow depth is observed every morning using a wooden stake marked at 2.54 cm intervals and mounted in the ground prior to the first snowfall. In case of disturbance to this site, a backup stake is used.

Temperature and snow data for Pinkham from Data Set 3206 (DSI-3206): COOP Summary of the Day (January 1930–May 1948) and Data Set 3200 (DSI-3200): Surface Land Daily Cooperative Summary of the Day (June 1948–March 2007) were purchased from the NCDC (2003, 2006). The data were reformatted and selected for the following parameters: daily maximum temperature, daily minimum temperature, temperature at the time of observation, daily snowfall, and daily snow depth. NCDC quality control edits to the data were accepted as indicated by the Data Quality Flag (i.e. flag M = Switched TOBS with TMAX or TMIN). Data for only one flag, T (failed internal consistency check), were removed. As might be expected of a COOP station, there were more missing data from Pinkham than the summit's nearly complete record.

Methods

TEMPERATURE

Seasonal Averaging and Linear Trend Calculation

Daily maximum and minimum data were first used to calculate daily mean temperature [(max + min)/2]. The daily temperature data (max, min, mean) were then averaged into monthly values. Seasonal (Winter: December, January, February, etc.) and annual means were calculated from the monthly data. There had to be at least 20 days of data to compute monthly means. Seasonal and annual values were not computed if a month was missing from the respective interval. Pinkham temperature data for October and November 1979, December 1983, April 1990, June 2004, May 2005, and May 2006 did not have enough observations.

Linear regressions were fit to annual and seasonal mean data, and slopes are presented as decadal trends. The trend significance is indicated by the *p*-value from the linear fit. It should be noted that although summit annual and seasonal means are derived from the same data as that used by Grant et al. (2005), we do not calculate a value for winter 1935 (December 1934, January, February 1935) due to the missing-data criteria explained above. The significance values of the summit trends presented here are the *p*-values from the linear fit, not the Monte Carlo significance employed in the earlier study (Grant et al., 2005).

Homogenization of Data

Trends in climate data may exist due to actual climate change or to artifacts such as station relocations, instrument changes, or gradual alterations in the use of surrounding land; thus data should be subjected to a homogeneity test (Alexandersson and Moberg, 1997). Furthermore, to study climate trends derived from daily temperature data (e.g. number of days per year exceeding a maximum temperature, growing degree day accumulation) recent effort has been dedicated to homogenizing not only seasonal and annual averaged time series but daily data (e.g. Vincent et al., 2002; Brunet et al., 2006). This effort is made difficult due to the large daily variability of temperature.

The Standard Normal Homogeneity Test (SNHT; Alexandersson and Moberg, 1997) was applied to both Summit and Pinkham seasonal and annual mean temperature data. The summit temperature data were found to be homogeneous, similar to Grant et al. (2005), while Pinkham data were found to be inhomogeneous.

In order to homogenize Pinkham daily temperature data, the Spanish daily temperature homogenization method (Brunet et al., 2006), which is a hybrid of the SNHT (Alexandersson and Moberg, 1997) and a Canadian daily adjustment scheme (Vincent et al., 2002), was adopted and applied to the 1935–2003 Pinkham data. This time period was chosen because missing Pinkham data only allows annual mean values to be calculated through 2003. The regional comparison stations used for Pinkham consisted of the same 11 USHCN stations located within 1° latitude and 1° longitude of Mount Washington used previously to analyze the summit temperature (Grant et al., 2005). Monthly mean maximum and minimum data for these stations from 1930 to 2003 were obtained from the Department of Energy's Carbon Dioxide Information Analysis Center (available online http://cdiac.ornl.gov/epubs/ndp/ushcn/usa_monthly.html).

Only the point SNHT was used to identify inhomogeneities for Pinkham. Prior studies have shown the temperature trends on the summit to be less pronounced than regional trends (Grant et al., 2005); there was concern that Pinkham's potentially smaller trends would be overcorrected by the regional reference. Ideally a comparison test would be used with topographically similar stations; however Pinkham, similar to the summit, is unusual in New England due to its elevation and local topography. There was also concern that point inhomogeneities from Pinkham's station changes would be interpreted as trend inhomogeneities, given that the two known Pinkham station changes (1966 and 1986) occurred while the low-elevation stations in the region experienced localized cooling during the 1960s and warming trends since the late 1970s (Hayhoe et al., 2007).

In order to identify multiple change points in each series, the SNHT was applied to the entire series. If a change point was identified, the series was broken into two pieces, each of which was then subjected to the SNHT. This was repeated until either there were no significant change points or each section was smaller than 10 years. SNHT points were not identified within 5 years of the start or end of the series (Moberg and Alexandersson, 1997). Based on the Spanish daily adjustment method, 1966 and 1986 were picked as standard changepoints for Pinkham.

Adjustments for each monthly time series were calculated using the SNHT; starting from the end of each series data were adjusted to the most recent period to allow new data to be seamlessly added (Alexandersson and Moberg, 1997); i.e., 1967–1986 was adjusted to 1987–2003 and then 1935–1966 adjusted to 1967–2003. These monthly adjustments were then transformed into mid-month "target" values using a tri-diagonal 12 × 12 matrix (Sheng and Zwiers, 1998) and into daily adjustments by linear interpolation between the monthly targets (Vincent et al., 2002; Brunet et al., 2006).

The daily adjustments (Fig. 2) were applied to the Pinkham data, creating the Pinkham Daily Adjusted Temperature Series (PDATS). The calculated adjustments for the 1967–1986 period, which ended with the installation of the MMTS thermometer, matches the sign of the standard LIG to MMTS thermometer adjustments; namely, maximum temperatures were adjusted down and minimum temperatures were adjusted up.

DEVELOPMENT OF SNOW SEASON CRITERIA

The snow season was defined using the year associated with January (e.g. the season of 1 July 1968–30 June 1969 is winter 1969). Snow season criteria were mainly driven by the unique conditions atop Mount Washington. The high winds and complex terrain, including buildings, make it difficult to measure snowfall. Specific concerns are the ability to separate blowing snow from falling snow during times of high winds and to compare amounts measured during windy and calm conditions. The location and number of precipitation cans has also changed over the years. In short, it is difficult to place a confidence range on summit snowfall measurements, so we analyzed snow depth.

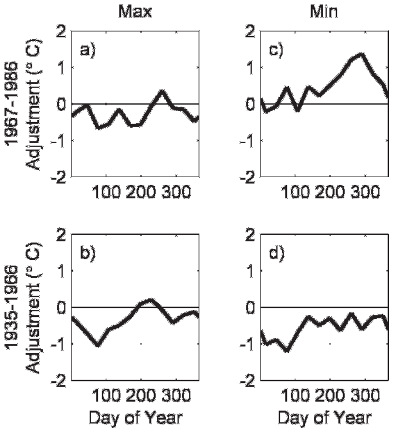

FIGURE 2. (a, b) Max and (c, d) Min daily adjustments applied to the 1967–1986 period and 1935–1966 period, respectively.

In order to avoid observer variability, snow depth observations were binned into two categories: (1) less than 2.54 cm, and (2) greater than or equal to 2.54 cm. This cutoff was chosen based on the protocol for determining snow cover disappearance in the International Tundra Experiment (ITEX) manual and other studies (Foster, 1989; Molau, 1996b). While methodology might vary over time and between individuals, we have high confidence that observers correctly identified what is essentially a snow/no snow cutoff. Choice of this criterion also avoids potential inhomogeneities in depth due to undocumented changes in method (Kunkel et al., 2007).

The binned data were used to calculate two separate length-of-season criteria. The first set of criteria, first and last snow (First/Last), are simply the first and last dates during a season that a snow depth, including trace amounts, was reported. The second criteria set, start and end of continuous snow cover (Start/End), are the dates after which and before which there was a continuous cover of at least 2.54 cm, allowing for "thaw" periods of no more than 4 days in length, with the end of cover most important to spring growth (Molau, 1996b). Continuous snow cover is generally used to calculate snow cover duration at higher elevations with persistent snow cover (e.g. Beniston et al., 2003). Allowing a short thaw period was necessitated by the summit data to extend the continuous snow cover season into spring. Otherwise a count-based calculation of snow cover duration (e.g. number of days with depth greater than a criterion per month) would be necessitated which would hinder the ability to identify the end of continuous snow cover, a biologically relevant event.

CALCULATION OF SNOW SEASON CRITERIA

Summit 6-hourly snow depth data were averaged to create a daily mean value. Pinkham daily snow depth is that reported at 0700 LST. Similar to temperature, the summit data were nearly complete while Pinkham was missing data. To have the most snow data possible, particularly at the tails of the season, missing Pinkham snow depth data were manually inspected and, if possible, missing depths were inferred.

After inspection, only 5 of 706 snow season criteria could not be used for Pinkham: first snow date for 1980 and snow cover start dates for 1980, 1983, 1990, and 1993. For the summit the end cover date in 1967 was not used because of suspected abnormal reporting practices: not once is total snow depth reported to be greater than 12.7 cm suggesting that observations were skewed low with reported conditions of less than 2.54 cm of snow artificially raised. In comparison, Pinkham does not show abnormally low snow depth during this winter.

TABLE 1

Annual and seasonal mean temperature with standard deviation and decadal trends with *p*-values for Pinkham (1935–2003) from PDATS data.

		Mean (°C)	Std. Dev. (°C)	Trend (°C dec.$^{-1}$)	Trend p
Annual	max	9.7	0.7	0.03	0.47
	min	−1.3	0.7	0.11	0.01
	mean	4.2	0.6	0.07	0.07
Winter	max	−2.8	1.6	0.11	0.24
	min	−13.8	1.6	0.18	0.07
	mean	−8.3	1.5	0.15	0.12
Spring	max	8.4	1.3	0.13	0.10
	min	−2.8	1.3	0.11	0.16
	mean	2.8	1.2	0.12	0.11
Summer	max	21.9	0.9	−0.02	0.76
	min	10.4	0.7	0.11	0.01
	mean	16.2	0.7	0.05	0.28
Autumn	max	11.6	1.1	−0.08	0.26
	min	1.1	1.0	0.08	0.19
	mean	6.3	0.9	0.00	0.99

For both First/Last and Start/End calculations the snow depth data were first divided into snow years (1 July–30 June) and indexed by day of winter (called DoW; 1–365 or 366 if a leap year). For each year First/Last data were calculated by finding all occurrences of snow depth greater than a cutoff criterion (C_{FL}), here 0 cm. The earliest occurrence (lowest DoW) of depth $> C_{FL}$ is the date of First snow depth and latest occurrence (largest DoW) of depth $> C_{FL}$ is the date of Last snow depth.

To determine Start/End dates for a snow year, all snow depth data greater than the Start/End depth cutoff criterion (C_{SE}), here 2.54 cm, were identified (snow subset) and the indices of these points stored (snow index). The first of the subset values was held as a potential Start date and the last as a potential End date. Next it was determined if a thaw period existed by taking the difference of adjacent snow index values. This difference index was transformed from days between snow to days of thaw by subtracting one. If these differences were less than the maximum number of continuous thaw days (d_{THAW}), here 4, then no thaw occurred and we used the above Start and End. If a thaw (or multiple thaws) occurred, the longest continuous period of the snow subset was selected and the first and last dates of this period were used as Start and End of snow cover.

CALCULATION OF THAWING DEGREE DAYS AND GROWING DEGREE DAYS

Thawing degree days (TDD, threshold T = 0°C) and growing degree days (GDD, threshold T = 5°C) were calculated for the PDATS data and summit data using ITEX formulae (Molau, 1996a). TDD and GDD are the sums of daily heat accumulation (H) greater than the respective threshold over a series of days, here starting 1 January. These formulae require either daily min/max temperature observations or 24 hourly observations per day.

For hourly T data, H is the sum of the day's hourly observations where T > threshold divided by 24. The summit's end of the hour observations were substituted for ITEX's recommended hourly averages. If hourly observations were missing they were replaced with an estimate, namely the mean value for that day and hour based on the full record.

For min/max data, H is defined as the daily mean T [(max + min)/2] if both min and max T are greater than the threshold or 0 if both min and max T are less than or equal to the threshold. For the third case, where the max is greater than the threshold but the min is lower, H is calculated by multiplying the daily T amplitude (max–min) by a scaling factor (Watanabe, 1978). Missing T values were estimated for periods up to 5 days using a mean based on the data preceding and following the missing data. If long stretches of data were missing, they were not estimated and that season's datum not used. The summit has no missing min/max data and Pinkham is only missing >5 continuous days a month three times during the critical winter and spring months: April 1990, May 2005, and May 2006. TDD for 1990, 2005, and 2006 were retained because the cumulative target value is reached prior to the period of missing data while GDD values were removed because the target value is reached after the missing data.

For each location the date at which TDD and GDD first reached an amount associated with the end of continuous snow cover and bloom of an early season plant, respectively, was identified. For the summit and Pinkham the approximate onset of bloom (15 May) for *Diapensia lapponica* and *Trillium undulatum*, respectively, was used (AMC, unpublished data). Specifics are described below. These values were plotted along with linear regressions with significance given by the associated *p*-values.

TABLE 2

Annual and seasonal mean temperature with standard deviation and decadal trends with *p*-values for the summit, 1935–2003.

		Mean (°C)	Std. Dev. (°C)	Trend (°C dec.$^{-1}$)	Trend p
Annual	max	0.9	0.6	0.03	0.37
	min	−6.5	0.7	0.06	0.19
	mean	−2.8	0.6	0.05	0.25
Winter	max	−9.4	1.6	0.09	0.35
	min	−18.6	1.8	0.08	0.45
	mean	−14.0	1.6	0.09	0.39
Spring	max	−1.0	1.2	0.11	0.14
	min	−8.6	1.4	0.12	0.16
	mean	−4.8	1.3	0.12	0.14
Summer	max	11.5	0.9	−0.05	0.32
	min	5.1	0.8	0.01	0.79
	mean	8.3	0.8	−0.02	0.69
Autumn	max	2.6	1.0	−0.02	0.72
	min	−4.2	1.1	−0.01	0.89
	mean	−0.8	1.0	−0.02	0.80

Results

TEMPERATURE TRENDS

The annual mean temperature at Pinkham from 1935 to 2003 using the PDATS data is 4.2°C, compared to the unadjusted value of 4.5°C. The most recent 30-year normal annual mean for Pinkham is 4.5°C (NCDC, 2004). PDATS seasonal means and standard deviations are presented in Table 1. The summit's annual mean temperature is −2.8°C; the seasonal means are shown in Table 2. At both stations the winter season is the most variable while the summer is the least variable.

Using the annual mean temperature from Pinkham and the summit, a lapse rate on the eastern slope of Mount Washington of 0.54°C/100 m was calculated. This value compares well with other annual mean lapse rates (range 0.5–0.70°C/100 m) on mountain ranges in the northeastern U.S.A., as tabulated by Richardson et al. (2004). The October 2001–September 2002 PDATS mean temperature, 5.9°C, is slightly cooler than extrapolated tempera-

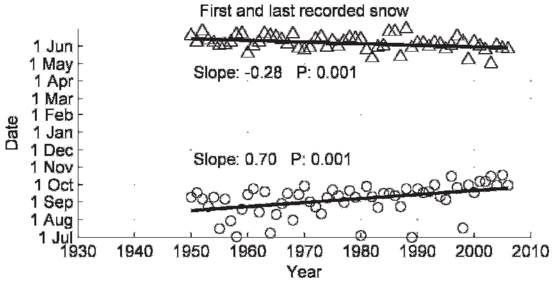

FIGURE 3. Summit day of first (circles) and last (triangles) snow with linear regressions for 1949–2006.

tures (~6.3°C) at similar elevations on Mount Moosilauke, New Hampshire, Whiteface Mountain, New York, and Mount Mansfield, Vermont, for the same time period (Richardson et al., 2004). Mean temperature at Pinkham from PDATS data (3.6°C) was colder than the estimated temperature at a similar elevation on the western slope of Camels Hump, Vermont (4.8°C) during a three-year period in the mid-1960s (Siccama, 1974).

Based on PDATS data, Pinkham shows significant ($p \leq 0.01$) warming trends for annual and summer minimum temperature (Table 1). Pinkham annual mean, winter minimum, and spring maximum temperature evidenced increases with a lower significance level ($p \leq 0.1$). Although not all are significant, warming trends in winter and spring at Pinkham are larger than summer and fall.

The summit shows annual, winter, and spring warming trends in maximum, minimum, and mean temperature although none are statistically significant (Table 2). Summer and autumn show cooling trends, which may correspond to the increased fog frequency (Seidel et al., 2007).

Similar to the summit, the Pinkham minimum trends (except for spring) are larger than the maximum trends. The Pinkham annual minimum trend is roughly twice as large as that observed on the summit. The Pinkham annual mean trend is slightly larger than that at the summit while the maximum trend observed at Pinkham matches that observed at the summit. Pinkham's annual mean trend (0.07°C/decade) is similar to that of the region (0.08°C/decade) over the last century (Hayhoe et al., 2007) and slightly smaller than the regional trend (0.10°C/decade) from 1931 to 2000 (Keim et al., 2003). Pinkham's annual mean trend is approximately halfway between the 1931–2000 trends for the southern (0.11°C/decade) and northern (0.0°C/decade) NOAA climate divisions of New Hampshire (Keim et al., 2003). Pinkham's annual mean warming is similar to that in NOAA climate divisions encompassing the mountains of Maine and to a lesser degree the mountains of Vermont (Keim et al., 2003), although these data are generally low-elevation stations.

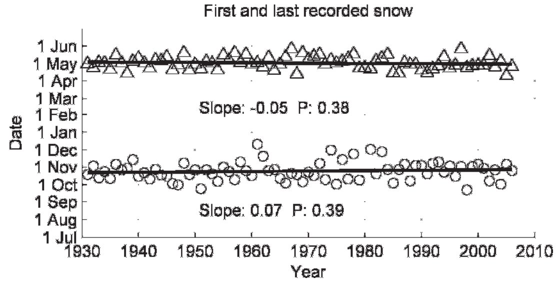

FIGURE 4. Pinkham day of first (circles) and last (triangles) snow with linear regressions for 1931–2006.

SNOW SEASON LENGTH

The trends in the dates of first and last observed snow depth on the summit are significant, with first snow occurring 7.0 days/decade later in autumn and last snow occurring 2.8 days/decade earlier in the spring (Fig. 3, Table 3). The trend in date of first snow (0.73 days/decade later) and last snow (0.50 days/decade earlier) at Pinkham are similar in sign but smaller in magnitude and not significant (Fig. 4, Table 3).

In contrast to the First/Last snow results, the summit shows no significant trends in the start (0.44 days/decade earlier) and end (0.18 days/decade later) of continuous snow cover (Fig. 5, Table 3). Pinkham (Fig. 6, Table 3) shows an earlier melt with a significant trend of 1.6 days/decade earlier. The trend in start of continuous snow cover at Pinkham is 0.50 days/decade later in autumn. Earlier end of continuous snow cover at Pinkham is consistent with the decreasing trend in snow depth based on independent Pinkham Notch samples taken in a 15-day window around 1 March for the Maine Cooperative Snow Survey (Hodgkins and Dudley, 2006). The much longer period of record for snow measurements at Pinkham (1931–2006) compared to the summit (1949–2006), with the 1960s being a regionally recognized cooler period (Zielinski and Keim, 2003) may explain the greater decadal rate of change and significance of the summit's first and last snow results.

DEGREE-DAY VALUES

Degree-day calculations are used to convert temperature data to a more biologically relevant metric; in montane environments cumulative degree days are strongly tied to phenological events and snowmelt. GDD with a 5°C threshold has often been used in relation to alpine plant development, while TDD with its threshold at freezing is related to snowmelt timing. In order to establish target TDD values for trend analysis, at each location the mean of the TDD values (based on min/max data) on the mean

TABLE 3

Mean, standard deviation, decadal trends, and *p*-values of the day of the year of the First/Last snowfall and Start/End of snow cover.

			Snow		Snow cover	
Station			First	Last	Start	End
Summit 1949–2006	mean	(day)	6 Aug (218)	6 June (157)	24 Nov (328)	29 Apr (119)
	std. dev.	(days)	27	13	28	22
	trend	(days dec^{-1})	7.0	−2.8	−0.44	0.16
	p-value		<0.01	<0.01	0.86	0.89
Pinkham 1931–2006	mean	(day)	26 Oct (299)	2 May (122)	2 Dec (336)	19 April (109)
	std. dev.	(days)	19	11	20	13
	trend	(days dec^{-1})	0.73	−0.5	0.5	−1.6
	p-value		0.39	0.38	0.64	0.02

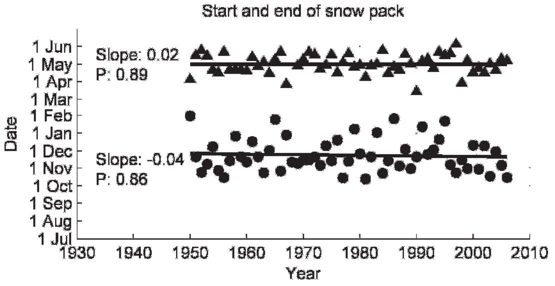

FIGURE 5. Summit day of start (circles) and end (triangles) of continuous snow cover with linear regressions for 1949–2006.

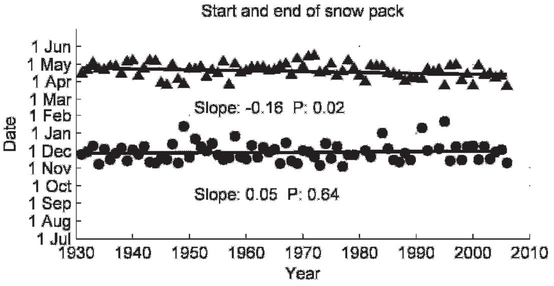

FIGURE 6. Pinkham Notch day of start (circles) and end (triangles) of continuous snow cover with linear regressions for 1931–2006.

date of the end of continuous snow cover, shown in Table 3, was calculated. This gave a target TDD value of 26 and 111 for the summit and Pinkham, respectively. Using a similar method, target GDD values of 15 and 137 for the summit and Pinkham, respectively, were derived from the approximate bloom date (15 May) of a species monitored by the AMC's phenology program at each location, namely *Diapensia lapponica* above treeline and *Trillium undulatum* at Pinkham.

The trend in summit GDD calculated from hourly data is advanced 9 days compared to GDD calculated from summit min/max data (Table 4). This suggests the min/max calculation of GDD estimates heat accumulation for the summit differently than hourly data. This can possibly be explained by an assumption in the calculation of min/max GDD, namely that a typical solar-driven diurnal heating pattern (radiative) with daily maximum in the afternoon and daily minimum shortly before sunrise is modeled using a sine-like relationship. The summit often does not follow this pattern; on 50% of days the summit experiences advective forcing and/or cloud immersion, which results in daily temperature extremes being recorded near midnight (Grant et al., 2005). During winter months only 23% of days can be classified as radiative increasing to 37% during summer (the remaining days are unclassified). This suggests that while both GDD calculations could be used to develop relationships between temperature and plant growth, the hourly calculation might more completely represent the actual accumulation of heat. Both min/max and hourly data sets give similar results when calculating TDD.

The date at which thawing degree days at the summit first reaches 26 advanced significantly ($p = 0.01$) using both min/max and hourly data sets (Fig. 7, Table 4). Pinkham showed a significant ($p < 0.01$) TDD advance (Fig. 8, Table 4) of about half the magnitude of that found at the summit. Growing degree-day trends are not significant for the summit or Pinkham (Figs. 7 and 8, Table 4) although all suggest earlier onset of spring warming. Although the observed negative trend in summit TDD would suggest earlier snowmelt, the end of the summit's snow cover shows no significant trend. However, at Pinkham the significant trend of an earlier end to the continuous snow cover of -1.6 days/decade matches the earlier accumulation of TDD (-1.6 days/decade).

INTERACTION OF SNOW AND TEMPERATURE

To explore the relationship between temperature and snow cover, a series of linear regressions was performed. Each of the many regressions used the end of snow cover date paired with an individual month's average max, min, or mean temperature (e.g. average April max vs. snow cover) or combinations of adjacent spring and winter months (e.g.. average April + March max vs. snow cover). The strongest relationship for the summit was that between end of snow cover and average April max temperature ($r^2 = 0.18$, $p = 0.001$). A similar analysis for Pinkham showed the strongest relationship between end of snow cover and average April + March max temperature ($r^2 = 0.51$, $p < 0.001$; PDATS data). There is both a difference in the timing of warmth important to snowmelt between the summit and Pinkham and the importance of that warmth; it appears just over half of the variability in timing of snowmelt below treeline can be attributed to temperature, while only 18% of the variability in melt timing on the rocky, windswept summit is due to temperature.

Snow cover also plays a role in protecting alpine plants from damaging frosts. Taschler and Neuner (2004) determined frost resistance for various species and concluded that protection by snow cover and other frost avoidance strategies influences the impact of low temperatures during the nascent growing season. In order to further explore the relationship between end of continuous snow cover and harmful late frosts on the summit, the number of days below minimum temperature thresholds within 60 days of the end of continuous snow cover was calculated. Two threshold temperatures were chosen, using Taschler and Neuner (2004) for guidance, to represent potential alpine plant frost damage: $-2°C$ for floral damage and $-5.5°C$ for vegetative damage. The summit experiences respective medians of 16 and 7 days below floral and vegetative thresholds post end of snow cover. There is large inter-annual variability (floral range of 2–55 days and vegetative 0–50 days) with early end of snow cover years having many days below the threshold temperatures. Using linear fits, neither measurement of frost risk shows evidence of significant trends (p-values of 0.63 and 0.87). As it might take several seasons of cold post-snow temperatures to exhaust a plant's energy reserves, decadal averages were analyzed; these also showed no trend. These results suggest that, for the alpine flora on Mount Washington that initiate growth soon after snowmelt, the balance between the risk of early growing season frost damage versus the gains of a longer growing season has not changed significantly during the period of record.

Discussion

Post-glacial climatic warming trends are frequently generalized as being horizontally and vertically synchronous across the landscape, along with the assumption that alpine ecosystems may be at great risk. Our results parallel those of Diaz and Bradley's (1997) that spatially, montane climate warming is complex. For different mid-latitude mountain regions in the world, the magnitude of climatic change is likely to vary considerably and be influenced by different factors. Mount Washington's summit temperatures over the last 70+ years, though trending towards warming, do not exhibit a statistically significant change ($p <$

TABLE 4

Mean, standard deviation, decadal trends, and *p*-values for day of the year when thawing degree days and growing degree days reach targets described in text.

	TDD				GDD			
	Mean (day)	Std. dev. (days)	Trend (days dec^{-1})	p	Mean (day)	Std. dev. (days)	Trend (days dec.$^{-1}$)	p
Summit								
min/max	30 April (120)	18	−2.8	0.01	17 May (137)	14	−0.4	0.64
hourly	30 April (120)	18	−2.7	0.01	8 May (128)	15	−1.1	0.20
Pinkham								
min/max	20 April (110)	9	−1.6	<0.01	16 May (136)	8	−0.5	0.27

0.05). Although our trends match those presented earlier for Mount Washington by Grant et al. (2005), the significance values differ. This earlier study did not account for temporal autocorrelation in their Monte Carlo simulations, has since been revised, and now concludes there were no significant temperature trends (Grant et al., 2008).

At our mid-elevation site there is a statistically significant warming in both annual and summer temperatures, with greater warming than that observed on the summit and less than that reported for lower elevations in the region. Pinkham mean temperatures show mixed results when compared to prior short-term studies in the White, Green (Vermont), and Adirondack (New York) Mountains. Nearby (48 km SW) and of similar elevation (222–1015 m a.s.l.), the U.S. Forest Service Hubbard Brook Experimental Forest experienced a significant warming trend from 1955 to present. Pinkham's end of snow cover, similar to the U.S. Forest Service's Hubbard Brook Experimental Forest's snow duration (Campbell et al., 2007), is happening earlier.

For biologically relevant temperature indices such as TDD and GGD, only TDD was statistically significant at both elevations in our study. However, only at our mid-elevation site did this manifest itself with a statistically earlier end of continuous snow cover. The summit's dates of first (autumn) and last (spring) measurable snow are significantly later and earlier, respectively, but the period of record is much shorter than for Pinkham; summit trends in onset and termination of continuous snow cover are not significant. Temporal trends of growing degree days were not significant at either site although they suggest earlier thawing and onset of the growing season.

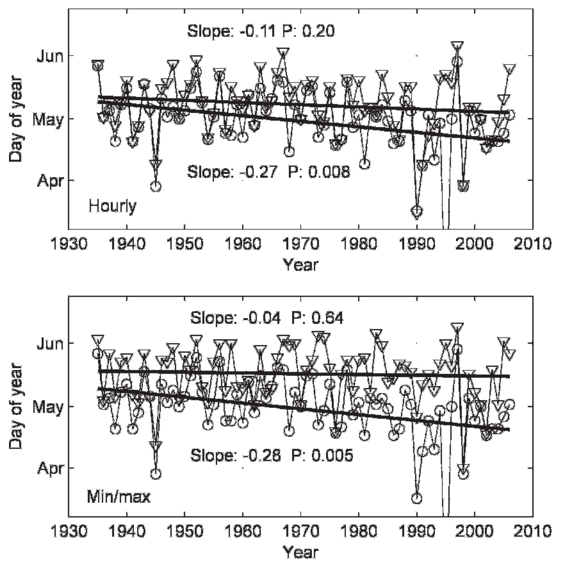

FIGURE 7. Day of year on which growing degree days (triangles) and thawing degree days (circles) first sum to target values (GDD = 15, TDD = 26), with linear regressions for the summit. The top panel show results based on hourly data and the bottom panel from min/max data.

Grant et al. (2005) estimated the summit of Mount Washington experiences free-atmosphere (troposphere) conditions on 50% of days in both summer and winter. This may explain why the summit exhibits a weak but not statistically significant warming trend, because during these conditions the summit would not necessarily be coupled with events observed from the surrounding regional lower elevation trends. Alpine areas in Europe, Asia, and other locations are experiencing warming trends, usually most significant in daily minima temperatures, and often greater in magnitude at higher elevations (Beniston et al., 1994; Diaz and Bradley, 1997). However, other alpine areas demonstrate less typical patterns, such as the Front Range, Colorado, U.S.A., where long-term trends indicate warming at mid-elevations, but a cooling trend within the alpine zone (Pepin, 2000).

There is evidence that resistance to climate warming at the higher elevations on Mount Washington has considerable tenure. Spear (1989), using pollen and plant macrofossil records from Mount Washington and surroundings, concluded that since 5000 yr BP, the subalpine forest and treeline-alpine ecotone boundary on Mount Washington has not exhibited demonstrable shifts. In contrast, mid and lower elevation tree species showed responses to climatic shifts in temperature. He concluded that alpine treeline is a poor temperature indicator for the region and hypothesized that wind and moisture determine the mountain's treeline position. Harding (2005) pointed out that alpine treeline in Scotland demonstrated a similar record of historical resistance. In his study of seed bank dynamics, seed dispersal, and colonization, he found no evidence for upward shifts in alpine vegetation under future warming scenarios, and suggested that models that do not

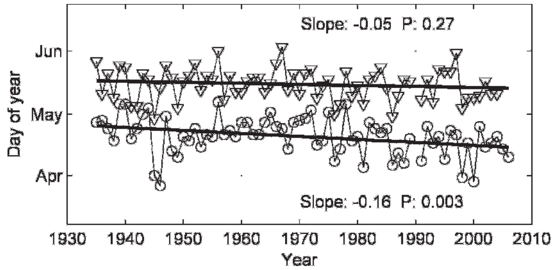

FIGURE 8. Day of year on which growing degree days (triangles) and thawing degree days (circles) first sum to target values (GDD = 137, TDD = 111), with linear regressions for Pinkham.

incorporate factors beyond temperature are likely to be poor predictors of future plant distributions.

Mount Washington's resistance to warming due to the frequency of being in the free atmosphere could also help explain why northeastern U.S. alpine ecosystems, which are remnant biogeographic islands from the last glacial period, are some of the lowest elevation alpine ecosystems at similar or more northern latitudes anywhere in the world. In addition, increasing frequency of fog events on Mount Washington (Seidel et al., 2007) could result in increased fog and rime ice deposition. Siccama (1974), Reiners and Lang (1979), Richardson et al. (2004), like Spear (1989), hypothesized that the transition from deciduous hardwood forest to coniferous spruce-fir forest to the alpine ecotone boundary on the mountains in this region is related in part to the cool, moist climate derived from frequent exposure to clouds. On northern New England mountains, Ryerson (1990) measured icing rates to increase exponentially above 800 m, with micro-topographic relief exposure a secondary control. He concluded that the dependence of icing rate upon elevation is largely a function of New England wind and cloud regimes and differs from other mountainous locations. However, evidence of an increasing cloud ceiling elevation at northeastern U.S. airports (Richardson et al., 2003) may have implications for the current cloud regime and mountain ecosystems, especially at mid-elevations. Research should be expanded to determine the applicability of increasing cloud ceiling elevation observations from airports to the region's mountains, where orographic effects are important.

The thermal structure of the lowest 2–3 km of the troposphere, the "planetary boundary layer," is complicated and includes inversions where temperature increases rather than decreases with height. Inversions are particularly common during winter over some middle and high latitude land regions. Inversions act to decouple surface temperatures from tropospheric temperatures on daily or even weekly time scales (Karl et al., 2006).

Our results support the conclusion that some mountains may only weakly follow regional low elevation surface climatic trends and may exhibit resistance to climatic warming with elevation. Factors may include temperature inversions, being in the free atmosphere at least a portion of the time, and sufficient atmospheric moisture availability to result in frequent cloud or fog exposure on the upper slopes. What regional climatic temperature increases would be sufficient to alter these dynamics is unknown.

Acknowledgments

Financial support was provided by the Office of Oceanic and Atmospheric Research at NOAA under grants NA06OAR4600209 and NA06OAR4600189. Larry Garland of the Appalachian Mountain Club provided the maps. Thanks are due to the dedicated observatory staff who have collected the summit data over the decades, especially current Director of Summit Operations Ken Rancourt; the volunteers who helped with data entry; those who helped establish, maintain and operate the Pinkham COOP Station; and those who helped reconstruct the Pinkham station history: J. Brooks Dodge, Jr., Michael Walsh, Peter Crane, and Dennis McIntosh. Two anonymous reviewers provided valuable comments on the original manuscript.

References Cited

Alexandersson, H., and Moberg, A., 1997: Homogenization of Swedish temperature data. Part I: homogeneity test for linear trends. *International Journal of Climatology*, 17: 25–34.

Arft, A. M., Walker, M. D., Gurevitch, J., Alatalo, J. M., Bret-Harte, M. S., Dale, M., Diemer, M., Gugerli, F., Henry, G. H. R., Jones, M. H., Hollister, R. D., Jonsdottir, I. S., Laine, K., Levesque, E., Marion, G. M., Molau, U., Molgaard, P., Nordenhall, U., Raszhivin, V., Robinson, C. H., Starr, G., Stenstrom, A., Stenstrom, M., Totland, O., Turner, P. L., Walker, L. J., Webber, P. J., Welker, J. M., and Wookey, P. A., 1999: Responses of tundra plants to experimental warming: meta-analysis of the international tundra experiment. *Ecological Monographs*, 69: 491–511.

Beniston, M., Rebetez, M., Giorgi, E., and Marinucci, M. R., 1994: An analysis of regional climate change in Switzerland. *Theoretical and Applied Climatology*, 53: 231–244.

Beniston, M., Keller, F., and Goyette, S., 2003: Snow pack in the Swiss Alps under changing climatic conditions: an empirical approach for climate impacts studies. *Theoretical and Applied Climatology*, 74: 19–31.

Bjork, R. G., and Molau, U., 2007: Ecology of alpine snowbeds and the impact of global change. *Arctic, Antarctic, and Alpine Research*, 39: 34–43.

Bliss, L. C., 1963: Alpine plant communities of the Presidential Range, New Hampshire. *Ecology*, 44: 678–697.

Brunet, M., Saladie, O., Jones, P., Sigro, J., Aguilar, E., Moberg, A., Lister, D., Walther, A., Lopez, D., and Almarza, C., 2006: The development of a new dataset of Spanish Daily Adjusted Temperature Series (SDATS) (1850–2003). *International Journal of Climatology*, 26: 1777–1802.

Burakowski, E. A., Wake, C. P., and Braswell, B., 2007: Trends in wintertime climate variability in the northeastern United States: 1970–2004. *EOS Transactions AGU*, 88: Fall Meeting Supplement, Abstract GC21A-0159.

Campbell, J. L., Driscoll, C. T., Eagar, C., Likens, G. E., Siccama, T. G., Johnson, C. E., Fahey, T. J., Hamburg, S. P., Holmes, R. T., Bailey, A. S., and Buso, D. C., 2007: *Long-term Trends from Ecosystem Research at the Hubbard Brook Experimental Forest*. Newtown Square, Pennsylvania: U.S. Department of Agriculture, Forest Service, Northern Research Station, General Technical Report NRS-17, 41 pp.

Cannone, N., Sergio, S., and Guglielmin, M., 2007: Unexpected impacts of climate change on alpine vegetation. *Frontiers in Ecology and the Environment*, 5: 360–364.

Carroll, C., 2007: Interacting effects of climate change, landscape conversion, and harvest on carnivore populations at the range margin: marten and lynx in the northern Appalachians. *Conservation Biology*, 21: 1092–1104.

Czikowsky, M. J., and Fitzjarrald, D. R., 2004: Evidence of seasonal changes in evapotranspiration in eastern US hydrological records. *Journal of Hydrometeorology*, 5: 974–988.

Diaz, H. F., and Bradley, R. S., 1997: Temperature variations during the last century at high elevation sites. *Climatic Change*, 36: 253–279.

Dye, D. G., 2002: Variability and trends in the annual snow-cover cycle in northern hemisphere land areas, 1972–2000. *Hydrological Processes*, 16: 3065–3077.

Foster, J., 1989: The significance of the date of snow disappearance on the arctic tundra as a possible indicator of climate change. *Arctic and Alpine Research*, 21: 60–70.

Freedman, J. M., Fitzjarrald, D. R., Moore, K. E., and Sakai, R. K., 2001: Boundary layer clouds and vegetation-atmosphere feedbacks. *Journal of Climate*, 14: 180–197.

Friedland, A. J., Boyce, R. L., and Webb, E. T., 1992: Winter and early spring microclimate of a subalpine spruce-fir forest canopy in central New Hampshire. *Atmospheric Environment; Part A, General Topics*, 26: 1361–1369.

Friedland, A. J., Boyce, R. L., Vostral, C. B., and Herrick, G. T., 2003: Winter and early spring microclimate within a mid-elevation conifer forest canopy. *Agricultural and Forest Meteorology*, 115: 195–200.

Grant, A. N., Pszenny, A. A. P., and Fischer, E. V., 2005: The 1935–2003 air temperature record from the summit of Mount

Washington, New Hampshire. *Journal of Climate*, 18: 4445–4453.

Grant, A. N., Pszenny, A. A. P., and Fischer, E. V., 2008: Corrigendum. *Journal of Climate*, 22: 1065–1066.

Hall, M. P., and Fagre, D. B., 2003: Modeled climate-induced glacier change in Glacier National Park, 1850–2100. *Bioscience*, 53: 131–140.

Hamilton, L. C., Brown, C., and Keim, B. D., 2007: Ski areas, weather and climate: time series models for New England case studies. *International Journal of Climatology*, 27: 2113–2124.

Harding, K. I. M., 2005: Potential movement of mountain vegetation zones in Scotland under predicted climatic change. M.Phil. thesis. University of Aberdeen, U.K., 127 pp.

Hayhoe, K., Wake, C. P., Huntington, T. G., Luo, L., Schwartz, M., Sheffield, J., Wood, E., Anderson, B., Bradbury, J., DeGaetano, A., Troy, T. J., and Wolfe, D., 2007: Past and future changes in climate and hydrological indicators in the US Northeast. *Climate Dynamics*, 28: 381–407.

Hirao, A. S., and Kudo, G., 2004: Landscape genetics of alpine-snowbed plants: comparisons along geographic and snowmelt gradients. *Heredity*, 93: 290–298.

Hodgkins, G. A., and Dudley, R. W., 2006: Changes in late-winter snowpack depth, water equivalent, and density in Maine, 1926–2004. *Hydrological Processes*, 20: 741–751.

Hodgkins, G. A., Dudley, R. W., and Huntington, T. G., 2003: Changes in the timing of high river flows in New England over the 20th century. *Journal of Hydrology*, 278: 244–252.

Hodgkins, G. A., Dudley, R. W., and Huntington, T. G., 2005: Changes in the number and timing of days of ice-affected flow on northern New England rivers, 1930–2000. *Climatic Change*, 71: 319–340.

Howat, I. M., Slawek, T., Rhodes, P., Israel, K., and Snyder, M., 2007: A precipitation-dominated, mid-latitude glacier system: Mount Shasta, California. *Climate Dynamics*, 28: 85–98.

Hubbard, K. G., Lin, X., Baker, C. B., and Sun, B., 2004: Air temperature comparison between the MMTS and the USCRN temperature systems. *Journal of Atmospheric and Oceanic Technology*, 21: 1590–1597.

Huelber, K., Gottfried, M., Pauli, H., Reiter, K., Winkler, M., and Grabherr, G., 2006: Phenological responses of snowbed species to snow removal dates in the central Alps: implications for climate warming. *Arctic, Antarctic, and Alpine Research*, 38: 99–103.

Huntington, T. G., Hodgkins, G. A., Keim, B. D., and Dudley, R. W., 2004: Changes in the proportion of precipitation occurring as snow in New England (1949–2000). *Journal of Climate*, 17: 2626–2636.

Inouye, D. W., and McGuire, A. D., 1991: Effects of snowpack on timing and abundance of flowering in *Delphinium nelsonii* (Ranunculaceae): implications for climate change. *American Journal of Botany*, 78: 997–1001.

IPCC, 2007: *Climate Change 2007: the Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge/New York: Cambridge University Press, 996 pp.

Karl, T. R., Hassol, S. J., Miller, C. D., and Murray, W. L. (eds.), 2006: *Temperature Trends in the Lower Atmosphere: Steps for Understanding and Reconciling Differences*. Washington, DC: Climate Change Science Program and the Subcommittee on Global Change Research, report, 165 pp.

Keim, B. D., Wilson, A. M., Wake, C. P., and Huntington, T. G., 2003: Are there spurious temperature trends in the United States Climate Division database? *Geophysical Research Letters*, 30: 1404.

Keller, F., Goyette, S., and Beniston, M., 2005: Sensitivity analysis of snow cover to climate change scenarios and their impact on plant habitats in alpine terrain. *Climatic Change*, 72: 299–319.

Kimball, K. D., and Keifer, M. B., 1988: Climatic comparisons with tree-ring data from montane forests: are the climatic data appropriate. *Canadian Journal of Forest Research*, 18: 385–390.

Kimball, K. D., and Weihrauch, D. M., 2000: Alpine vegetation communities and the alpine-treeline ecotone boundary in New England as biomonitors for climate change. *In* McCool, S. F., Cole, D. N., Borrie, W. T., and O'Loughlin, J. (eds.), *Wilderness Science in a Time of Change Conference—Volume 3: Wilderness as a Place for Scientific Inquiry*. Ogden, Utah: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station, 93–101.

Kudo, G., 1991: Effects of snow-free period on the phenology of alpine plants inhabiting snow patches. *Arctic and Alpine Research*, 23: 436–443.

Kunkel, K. E., Palecki, M. A., Hubbard, K. G., Robinson, D. A., Redmond, K. T., and Easterling, D. R., 2007: Trend identification in twentieth-century U.S. snowfall: the challenges. *Journal of Atmospheric & Oceanic Technology*, 24: 64–73.

Lemke, P., Ren, J., Alley, R. B., Allison, I., Carrasco, J., Flato, G., Fujii, Y., Kaser, G., Mote, P., Thomas, R. H., and Zhang, T., 2007: Observations: changes in snow, ice and frozen ground. *In* Solomon, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K. B., Tignor, M., and Miller, H. L. (eds.), *Climate Change 2007: the Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge/New York: Cambridge University Press, 337–383.

Lesica, P., and McCune, B., 2004: Decline of arctic-alpine plants at the southern margin of their range following a decade of climatic warming. *Journal of Vegetation Science*, 15: 679–690.

Moberg, A., and Alexandersson, H., 1997: Homogenization of Swedish temperature data. Part II: homogenized gridded air temperature compared with a subset of global gridded air temperature since 1861. *International Journal of Climatology*, 17: 35–54.

Molau, U., 1996a: ITEX climate stations. *In* Molau, U., and Molgaard, P. (eds.), *ITEX Manual*. Copenhagen, Denmark: Danish Polar Center, 6–10.

Molau, U., 1996b: Snow and ice. *In* Molau, U., and Molgaard, P. (eds.), *ITEX Manual*. Copenhagen, Denmark: Danish Polar Center, 11–13.

Mote, P. W., Hamlet, A. F., Clark, M. P., and Lettenmaier, D. P., 2005: Declining mountain snowpack in western North America. *Bulletin of the American Meteorological Society*, 86: 39–49.

NCDC, 2003: *Data documentation for Data Set 3206 (DSI-3206) COOP Summary of the Day—CDMP—Pre 1948*. Ashville, North Carolina: NCDC, 18 pp.

NCDC, 2004: *No. 20, Climatography of the United States 1971–2000*. Pinkham Notch, NH: National Climate Data Center, NOAA, U.S. Dept. of Commerce.

NCDC, 2006: *Data Documentation for Data Set 3200 (DSI-3200) Surface Land Daily Cooperative Summary of the Day*. Asheville, North Carolina: NCDC, 19 pp.

Pauli, H., Gottfried, M., Reier, K., Klettner, C., and Grabherr, G., 2007: Signals of range expansions and contractions of vascular plants in the high Alps: observations (1994–2004) at the GLORIA master site Schrankogel, Tyrol, Austria. *Global Change Biology*, 13: 147–156.

Pepin, N., 2000: Twentieth-century change in the climate record for the Front Range, Colorado, U.S.A. *Arctic, Antarctic, and Alpine Research*, 32: 135–146.

Pielke, R. A., Stohlgren, T., Schell, L., Parton, W., Doesken, N., Redmond, K., Moeny, J., McKee, T., and Kittel, T. G. F., 2002: Problems in evaluating regional and local trends in temperature: an example from eastern Colorado, USA. *International Journal of Climatology*, 22: 421–434.

Quayle, R. G., Easterling, D. R., Karl, T. R., and Hughes, P. Y., 1991: Effects of recent thermometer changes in the Cooperative Station Network. *Bulletin of the American Meteorological Society*, 72: 1718–1723.

Reiners, W. A., and Lang, G. E., 1979: Vegetational patterns and processes in the balsam fir zone, White Mountains, New Hampshire. *Ecology*, 60: 403–417.

Richardson, A. D., Denny, E. G., Siccama, T. G., and Lee, X., 2003: Evidence for a rising cloud ceiling in eastern North America. *Journal of Climate*, 16: 2093–2098.

Richardson, A. D., Lee, X., and Friedland, A. J., 2004: Microclimatology of treeline spruce-fir forests in mountains of the northeastern United States. *Agricultural and Forest Meteorology*, 125: 53–66.

Ryerson, C. C., 1990: Atmospheric icing rates with elevation on northern New England mountains, U.S.A. *Arctic and Alpine Research*, 22: 90–97.

Sardinero, S., 2000: Classification and ordination of plant communities along an altitudinal gradient on the Presidential Range, New Hampshire, USA. *Plant Ecology*, 148: 81–103.

Schwartz, M. D., Ahas, R., and Aasa, A., 2006: Onset of spring starting earlier across the northern hemisphere. *Global Change Biology*, 12: 343–351.

Seidel, T. M., Grant, A. N., Pszenny, A. A. P., and Allman, D. J., 2007: Dew point and humidity measurements and trends at the summit of Mount Washington, N.H. 1935–2004. *Journal of Climate*, 20: 5629–5641.

Sheng, J., and Zwiers, F., 1998: An improved scheme for time dependent boundary conditions in atmospheric general circulation models. *Climate Dynamics*, 14: 609–613.

Siccama, T. G., 1974: Vegetation, soil, and climate on the Green Mountains of Vermont. *Ecological Monographs*, 44: 325–349.

Spear, R. W., 1989: Late-Quaternary history of high-elevation vegetation in the White Mountains of New Hampshire. *Ecological Monographs*, 59: 125–151.

Starr, G., Oberbauer, S. F., and Pop, E. W., 2000: Effects of lengthened growing season and soil warming on the phenology and physiology of *Polygonum bistorta*. *Global Change Biology*, 6: 357–369.

Taschler, D., and Neuner, G., 2004: Summer frost resistance and freezing patterns measured *in situ* in leaves of major alpine plant growth forms in relation to their upper distribution boundary. *Plant, Cell and Environment*, 27: 737–746.

Trombulak, S. C., and Wolfson, R., 2004: Twentieth-century climate change in New England and New York, USA. *Geophysical Research Letters*, 31: L19202.

Vincent, L. A., Zhang, X., Bonsal, B. R., and Hogg, W. D., 2002: Homogenization of daily temperatures over Canada. *Journal of Climate*, 15: 1322–1334.

Walker, M. D., Walker, D. A., Welker, J. M., Arft, A. M., Bardsley, T., Brooks, P. D., Fahnestock, J. T., Jones, M. H., Losleben, M., Parsons, A. N., Seastedt, T. R., and Turner, P. L., 1999: Long-term experimental manipulation of winter snow regime and summer temperature in arctic and alpine tundra. *Hydrological Processes*, 13: 2315–2330.

Watanabe, N., 1978: An improved method for computing heat accumulation from daily maximum and minimum temperatures. *Applied Entomology and Zoology*, 13: 44–46.

Weber, R. O., Talkner, P., Auer, I., Bohm, R., GajicCapka, M., Zaninovic, K., Brazdil, R., and Fasko, P., 1997: 20th-century changes of temperature in the mountain regions of Central Europe. *Climatic Change*, 36: 327–344.

Zielinski, G. A., and Keim, B. D., 2003: *New England Weather, New England Climate*. Lebanon, New Hampshire: University Press of New England, 276 pp.

MS accepted January 2009