

ASSESSMENT OF LATE PLEISTOCENE TO RECENT CLIMATE-INDUCED
VEGETATION CHANGES IN AND NEAR SHENANDOAH NATIONAL PARK
(BLUE RIDGE PROVINCE, VA)

R.J. LITWIN¹, B. MORGAN¹, L.S. EATON², and G. WIECZOREK¹

¹U.S. Geological Survey, MS 926A, Reston, VA 20192

²Geology and Environmental Science, MSC 7703, James Madison University, Harrisonburg, VA 22807

Abstract

Pollen evidence from a shallow core in the Blue Ridge Mountains (Big Meadows, Virginia) and from other outcrops in and adjacent to Shenandoah National Park, indicates that from the Late Pleistocene through the Holocene (45-0 ka) regional vegetation in north-central Virginia fluctuated from warm temperate forests to fully developed taiga (boreal forest). Present day analogues to these vegetation zones can be found in the forests and forest transitions extending from central Georgia (approximately 32° N latitude) to central Ontario, Labrador, and northern Newfoundland (approximately 52-55° N latitude). It is still undetermined (on the basis of our pollen evidence) whether it was ever cold enough in the study area during this 45,000-year time interval to develop alpine tundra extensively along these ridge tops.

Current evidence from the study area suggests that the forests in and around Shenandoah National Park changed frequently in composition through the studied time interval. Most of the forests previously established in the study area were of types that favored notably cooler mean annual temperatures than the forest type that is established currently in the proximity of the Park. Although this fossil record is a partial one, the radiometric carbon evidence verifies that we now have discovered pollen assemblages of full-glacial age (last glacial maximum) in the Blue Ridge, of multiple forest types.

Background

An especially intense and prolonged summer storm in the area of Graves Mill (eastern border of Shenandoah National Park, June, 1995) created numerous debris flows and intensive scouring of streams and hillsides on the eastern front of the Blue Ridge (Morgan et al., 1999a; Wieczorek et al., 2000; study area, Fig.1). This scouring exposed numerous depositional remnants along stream banks that contained evidence of prehistoric debris flow and solifluction activity. ^{14}C analyses (Eaton and McGeehin, 1997) confirmed that these remnants were of Late Pleistocene and Holocene age.

The vegetational, climatic, and geomorphic history of the Blue Ridge during the Pleistocene and Holocene are poorly understood. We used this debris flow cluster event to begin intensive investigations of these deposits to increase our understanding of the role of debris flows in regional denudation (Eaton, 1999; Morgan et al., 1999b; Eaton et al., 2003a), the character of sediment processes in the Blue Ridge during the Late Pleistocene (Smoot, 2004a, 2004b), and a reconstruction of climate-driven vegetation change in the study area during the past 40,000 years (this report). This paper presents a first-order assessment of Late Pleistocene to Recent vegetative changes in the Blue Ridge in the area of Shenandoah National Park.

The outcrop remnants exposed by the 1995 storm are packages of sediment varying from 0.5 to 10 meters thick (Litwin et al., 2001); each deposit differs in age, but all accumulated during a relatively brief time interval. Palynological investigations coupled with ^{14}C analyses are used here to synthesize and reconstruct a chronological vegetation (proxy climate) sequence, linking prehistoric forest types with ages determined through accelerated mass spectrometric (AMS) ^{14}C analysis. Additionally,

we attempted to obtain a single long reference section by vibracoring at Big Meadows, Shenandoah National Park (Fig. 1). This attempt was partly compromised by vibration-induced fluidization of sediments in the deeper drilling runs, but we still were able to discern important aspects of the prehistory of Big Meadows from the shallower runs of the core.

This report contains a description of investigations at Big Meadows and more regional considerations regarding climate and vegetation change. Questions that are addressed in this report include the following. What is the character of the shallow subsurface at Big Meadows? What are the ages of the sediment in the subsurface? What are the implications for the geologic history of Big Meadows and for archeological studies that are ongoing in the area? What are the characterization and timing of the vegetation changes within the Blue Ridge of Shenandoah National Park over the last 45,000 years? When did the forest that now occupies the park first develop? Does core evidence document previous (prehistoric) intervals of meadow vegetation at the core site? Lastly, what is an initial estimate for the minimum range of climate change established in the study area over the past 45,000 years, as documented by climate-driven changes in vegetation?

This study was undertaken in cooperation with the National Park Service, during a recent revitalization (trimming and controlled burning) of the Big Meadows site for the purpose of preventing forest encroachment on the meadow. Permission to core was granted during this revitalization, as part of a joint scientific assessment on the archaeology and climate history of Big Meadows.

Samples and results

The suite of outcrop and core samples we analyzed and interpreted for this report are presented in Figure 1, and Tables 1 and 2. Their geographic locations are shown graphically in figure 1; the locations and elevations are listed in Table 1. Where possible, they have been cross-referenced to localities originally described by Eaton (1999). The ages assigned to each sample are noted in Table 2. Most have been dated directly by AMS ^{14}C radiometric analyses of macrofossil fragments (wood, charcoal, or seeds) after handpicking directly from the pollen-bearing sediment (using a binocular microscope). The uncorrected radiocarbon ages (14-C yr) and calibrated radiometric ages (ky cal BP) are presented, along with their attendant 1-sigma and 2-sigma analytical errors. Core samples are denoted by a “©” after the sample ID name; all other samples were taken in the field by the authors from unweathered outcrop.

The raw results from the pollen analyses of the core and outcrop datasets are presented in Table 3. Pollen identifications were based on Bassett et al. (1978), Lewis et al. (1983), Lieux (1980a, 1980b, 1982, 1983), Lieux and Godfrey (1982), Richard (1970), and Traverse (1988), as well as on pollen slides prepared directly from vouchered specimens that were sampled by one of us (RJL) from the George Mason University Herbarium.

Each sediment sample was spiked with a *Lycopodium* tracer to insure our chemical processing did not destroy or degrade the fossil pollen in each sample. In addition to the counts, we have noted the forest zone type (abbreviated) that we think best matches the each sample, based on the modern forest zone criteria (below). Badly damaged, crumpled, or unknown pollen morphotypes were assigned to the category “other”.

Minor occurrences of other taxa (e.g., *Salix*, *Comptonia*, etc.) also were included in this category.

The total pollen per gram of sample varied markedly across the sample set. Several samples had to be standardized in their count because the abundances of pollen were low in the sample; analyses were ended after two microscope slides were counted for each sediment sample, or a tabulation of 300 specimens was reached (whichever came first). We think that some taxa (e.g., *Quercus*) are over-represented in these samples, while other taxa (e.g., *Acer* and *Fraxinus*) are under-represented. Accordingly, we have used multiple presence/absence/abundance criteria for assigning each fossil sample to a probable prehistoric forest type. We used both the core and outcrop datasets to derive these pollen-based forest zone assignments.

Once we established prehistoric forest zones, we interpreted the data two ways. We examined the Big Meadows core dataset independently, to answer site-specific questions. We then combined the outcrop and core datasets to produce a compiled vegetation history. We think the compiled vegetation history is justified, as it based on a suite of well-dated point samples recovered from small depositional remnants of Quaternary age that are distributed across a small geographic area (Fig. 1) along the eastern face of the Blue Ridge.

INTERPRETATION OF PREHISTORIC VEGETATIONAL CHANGE

Use of modern analogue forest types

We base our interpretation of the pollen data on a comparison to the modern analogue suite of forests that are now established along the U.S. Atlantic Coast. At

present, we have used a generalized version of the forest characterization by Kuchler (1975) for the eastern United States, combined with a map of the ecoregions of Canada (Ecoregions working group, 1989; Fig. 2). Multiple characterizations exist for modern eastern U.S. forests, including those of Eyre (1980) and Braun (1950). We also refer readers to the U.S. Forest Service website, for further details and comparisons of some of these forest characterizations (<http://www.fs.fed.us>). For the purposes of this initial study, we found the zonation of Kuchler (1975) to be the most useful for comparison to our pollen assemblages. Given that our study only includes samples younger than 50 ka, we propose that these relatively young fossil samples can be assigned accurately to existing modern forest analogues. We do acknowledge that a small number of our prehistoric pollen assemblages may have been derived from forest types with no modern counterpart.

The main forest zones that we think previously may have been established in the Blue Ridge study area are those that now occupy the U.S. Atlantic seaboard: 1) Southern Mixed Forest, 2) Oak-Hickory-Pine Forest, 3) Appalachian Oak Forest (present study area), 4) Northern Hardwoods Forest, 5) Northern Hardwoods-Spruce Forest, 6) Northeastern Spruce-Fir Forest, and 7) Boreal Forest (Fig. 2; Kuchler, 1964, 1975; Ecoregions Working Group, 1989). Our fossil data set currently includes samples that we think represent each of these forest types, with the exception of Southern Mixed Forest. Although we do not yet have direct evidence, we predict that the very warmest of the forests established in the study area intermittently during the Holocene thermal maximum (~4-6 ka) possibly may have been of that type; it has not yet been investigated.

Currently we are in the process of cross-testing this initial model for transposing pollen assemblages into forest zones, by analyzing modern pollen assemblages in the modern soils associated with each of these forest types, from southern mixed forest northward through boreal forest and into subarctic tundra. The sample set we have accumulated (for this cross testing) thus far is indicated by plain white circles in Figure 2. It is hoped that these modern pollen assemblages will enable us to perform dissimilarity coefficient cross-tests on our fossil dataset to refine the pollen-to-forest zone assignments we propose in this report. The results of this cross testing and the refinement of our temperature index will be presented elsewhere; they are ongoing studies, and are not the primary focus of this report.

Modern taxa ranges

The first-order forest zone reconstruction for this study primarily was based on comparison of the composition of each of our pollen assemblages to our compilation of modern forest genera distributions within the area bounded by 20^o-60^o N latitude and 50^o-90^o W longitude. The generic ranges were compiled from the species distribution maps of Little (1971, 1976; Fig. 3; see also “<http://www.na.fs.fed.us/spfo/pubs/silvicsmanual>”). The number of species used to derive these composite ranges varied per genus, and is noted for each in Figure 3. The 31 genera we tabulated represent the cumulative mapped ranges of 147 plant species that exist between the tropics and the subarctic tundra. This compilation provided a characterization of the general plant components for each of the forest zones mapped by Küchler (1975), that we have used as our baseline modern forest analogues. We note two related points here. The first is that the zonation presented here

is preliminary, and will be updated and refined as necessary. The second point is that nearly half of these genera range through all the forest types we observed, from boreal forest to oak-hickory-pine forest. Because of this, we emphasize that our assignment of fossil pollen assemblages to modern forest analogues could not be based solely on the presence/absence of particular taxa, but also had to include the relative proportions of taxa within those assemblages. Consequences of this condition are addressed elsewhere in this report (see *Discussion*, p. 38). Grasses, composites, and lower vascular plant taxa were not included in the modern genus compilation of Figure 3. The genera that potentially range through all forests from boreal forest to oak-hickory-pine forest include: *Abies*, *Acer*, *Alnus*, *Betula*, *Cornus*, *Corylus*, *Fraxinus*, *Juniperus*, *Pinus*, *Quercus*, *Sanguisorba*, *Shepherdia*, and *Ulmus*. Although these genera all occur within the forests above, each may be a rare to uncommon element in any of them, and may not be represented consistently in the pollen record of that forest.

We used an overlay of mean annual temperature (MAT) isotherms for North America to help derive approximate temperature ranges for each genus we compiled, to construct Figure 3 (source: <www.kgs.ukans.edu/PRS/info/pdf/doveton.PDF>). We marked the midpoint of each genus's temperature range in Figure 3 with a diamond, to indicate its probable favored temperature regime and zone of probable abundance. The genera were then arranged on the graph by decreasing midrange values ("decreasing thermal optimum", Figure 3). The wider portion of each taxon's range bar represents one half of its total documented range, as centered on its total range midpoint. If one accepts a normal population distribution between the endmembers of a taxon's range, the wide bar approximates the subrange in which its populations- and therefore its total pollen-

should be relatively more common. Again, this is a first-order assessment only, and ranges and midpoints may shift as the compiled genera are emended. For example, the range noted for the genus *Kalmia* here is based on one species distribution (*K. latifolia*) which does not extend into spruce-fir forest or boreal forest. Other *Kalmia* species (e.g., *K. angustifolia* and *K. polifolia*) do occur in boreal forest (Larsen, 1980, p. 474), but their range distribution maps were not found for inclusion here. After constructing the range chart we used the MAT isotherms to help us interpolate warm and cold boundaries for each of the forest zones mapped by Kuchler (1975). These are shown as a series of horizontal bands in Figure 3 (converted to °C).

Criteria for 'pollen / forest zone' assignments

Those pollen assemblages we interpret as being derived from full boreal forest (here 'Zone 7' of Figure 2) are characterized by the following taxa: *Abies* (fir), *Acer* (maple), *Alnus* (alder), *Betula* (birch), *Cornus* (dogwood), *Corylus*, (hazelnut) *Juniperus* (juniper), *Larix* (larch), *Picea* (spruce), *Pinus* (pine), *Sanguisorba* (burnet), and *Shepherdia* (buffaloberry). *Populus* (poplar) and *Salix* (willow) also occur in this zone, but were not included in Figure 3. Larsen (1980, p. 31) additionally notes the presence of *Quercus* (oak), *Fraxinus* (ash) and *Ulmus* (elm) in the warmer, southern boreal forest. These putative occurrences are corroborated by their respective plant distribution maps in Little (1971) and distribution maps in Lewis et al. (1983), when compared to boreal forest boundaries noted by the Ecoregions Working Group (1989). *Fraxinus niger*, *Quercus macrocarpa*, *Q. rubra*, and *Ulmus americana* are all present in areas currently characterized as boreal forest (Little, 1971; Ecoregions Working Group, 1989). In general

the taxa above are exceptionally cold tolerant and characterize modern boreal forest vegetation (Larsen, 1980; Porsild and Cody, 1980; Ritchie, 1987). The following pollen index taxa appear to be absent from boreal forest: *Carya* (hickory), *Castanea* (chestnut), *Fagus* (beech), *Ilex* (holly), *Juglans* (walnut), *Liquidambar* (sweetgum), *Liriodendron* (tulip poplar), *Magnolia* (magnolia), *Morus* (mulberry), *Nyssa* (black tupelo), *Ostrya-Carpinus* (ironwood and hornbeam), *Platanus* (sycamore), *Sabal* (palmetto), and *Tsuga* (hemlock)(Larsen, 1980). The relative temperature index values (i.e., a ratio of oak and ash pollen to spruce and fir pollen, (see p.16, and Figs. 4 and 5)) of fossil assemblages that we have assigned to this forest zone are less than or equal to 1.2. The most commonly represented tree taxa in this zone are *Picea*, *Larix*, *Pinus*, *Abies*, and *Betula* (Ritchie, 1987). Modern examples of this forest zone can be found in the Gaspé Peninsula, elsewhere across much of southern Quebec, eastward across most of Newfoundland and southwestern Labrador, westward across southwestern Ontario and southern Manitoba, and northwestward to Alaska.

The next warmer forest zone-- the northeastern spruce-fir forest—has all of the component taxa of the boreal forest (above). It appears to differ from boreal forest by the first appearance of two taxa that are somewhat less cold tolerant: hemlock and beech (*Tsuga* and *Fagus*). This zone also includes a minor amount of oak (*Quercus*), probably derived mostly from northern red oak (*Q. rubra*). Northern red oak is broadly distributed in almost all of the forests in this study, with the exception of restricted occurrences in boreal forest and southern mixed forest (Little, 1971). Küchler (1975) has mapped this zone in northwest and southeast Maine, and in the highest mountains in Vermont, New Hampshire, and New York (Fig. 2). Those fossil samples that we assign to this zone

produce relative temperature index values between 1.2 and 1.85 (Fig. 5). K uchler (1975) described this zone as a predominantly coniferous forest (e.g., *Picea* and *Abies*), with a minor component of deciduous angiosperm taxa, including *Acer*, *Betula*, and *Populus*.

The next zone, the northern hardwoods-spruce forest ('Zone 5' of Figure 2), hosts several new taxa in addition to those noted above. *Juglans* (*J. cinerea*, butternut), *Ilex* (*I. verticillata*, common winterberry) and *Tilia* (basswood) have documented first appearances in this forest zone (Little, 1971; K uchler, 1964, 1975). A few other taxa- *Carya* (hickory), *Castanea* (chestnut), and *Liriodendron* (tulip poplar) *Ostrya-Carpinus* (ironwood and hornbeam)- were not noted in this zone by K uchler (1964), but have documented ranges that marginally overlap it (Little, 1971) and have first appearances within this zone in our fossil pollen samples. [*Ostrya* and *Carpinus* are presented in this study as a combined "form taxon" because they bear morphologically similar pollen (Ritchie, 1987, p.41).] The northern hardwoods-spruce forest lacks *Liquidambar*, *Magnolia*, *Morus*, *Nyssa*, *Platanus*, and *Sabal*. The relative temperature index values for pollen samples within this zone range from 1.85-7.0 (Figure 5). This forest type is found across north central Maine, the mountains of New Hampshire and Vermont, and across the Adirondacks of New York (excepting for its highest elevations, which are inhabited by spruce-fir forest; see Fig. 2). The Ecoregions Working Group (1989) does not differentiate the northeastern spruce-fir forest from the northern hardwoods-spruce forest, but combined the two into a High Cool Temperate Ecoclimatic Province. We have attempted to differentiate them here, on the basis of the genus-level first occurrence datums noted above.

As before, the next warmer forest zone—the northern hardwood forest (‘Zone 4’ of Figure 2) —contains all of the taxa of the previous zone, plus the addition of *Nyssa* (black tupelo) and *Platanus* (sycamore). *Picea* (spruce) abundance drops in this zone. This zone marginally contains the first occurrences of *Morus* (mulberry) and possibly *Magnolia* (“cucumbertree” or magnolia), based on their distribution maps (Little, 1971). The northern hardwood zone conspicuously lacks *Liquidambar* and *Sabal*. *Carya*, *Castanea*, and *Liriodendron* more regularly occur in the warmer part of this zone (southern half). It has common maple, birch, hazelnut, pine, and hemlock (*Acer*, *Betula*, *Corylus*, *Pinus*, and *Tsuga*) (Küchler, 1964). It also contains abundant oak (*Quercus*), linden (*Tilia*), and elm (*Ulmus*) (Little, 1971, 1976). This zone occurs at lower elevations across Maine, New Hampshire, Vermont, western Massachusetts, New York, and across the northern border of Pennsylvania. It extends southward across higher elevations of the Allegheny Plateau through the mountains of eastern West Virginia (Küchler, 1975; Fig. 2). Samples from our study that we assign to this zone bear relative temperature index values of 7-70 (Fig. 5). It is characterized in part by a strong admixture of deciduous hardwoods and conifers.

The Appalachian oak forest (“Zone 3” of Fig. 2) is characterized by a maximum abundance of oaks (*Quercus*), the common presence of pine (*Pinus*), hickory (*Carya*), walnut (*Juglans*), and scattered local presence of birch (*Betula*), beech (*Fagus*), and hemlock (*Tsuga*) (Küchler, 1964; Little, 1971, 1976). It contains the first appearance of *Liquidambar* (sweetgum). Additionally, the coldest parts of this forest zone host the last occurrences of spruce (*Picea*) and larch (*Larix*). *Abies* (fir), *Sanguisorba* (burnet) and *Shepherdia* (buffaloberry) appear to have their last occurrence in this forest zone; *Abies*

and *Sanguisorba* occurs as a refugial cold-loving plants at the highest elevations (~3500') in the Blue Ridge. Appalachian oak forest is the zone presently established around Big Meadows, and along the crests of the Blue Ridge, the Appalachians, and much of the Allegheny Plateau (Pennsylvania), usually above elevations of approximately 1000' (~305 m; Fig. 2). Up until the middle of the last century, chestnut (*Castanea dentata*) was an important co-dominant overstory component in this forest locally as well, although it probably is underrepresented in the pollen record, even at its historic maximum abundance. Since the infestation of chestnut by the blight *Cryphonectria parasitica* (*Endothia parasitica* per Anderson, 1974), it no longer exists as a significant overstory component in forests. It persists only as a shrub-to-small tree-sized understory component, with only minor contribution to the pollen record (Anderson, 1974). The relative temperature index values in the Appalachian oak forest zone range from 70-140 (Fig. 5). The dominance of oaks in this forest zone is illustrated by the average total percentage of oak pollen in each of our Appalachian oak forest samples- 47%; in the northern hardwoods forest, northern hardwoods-spruce forest, and spruce-fir forest the respective percentages are 38%, 26%, and 16%. Kuchler (1964) characterized it as a "tall, broadleaf deciduous forest" dominated by white oak (*Quercus alba*) and northern red oak (*Q. rubra*).

The next warmer forest zone, the oak-hickory-pine forest ("Zone 2" of Fig. 2), is the warmest zone for which we presently have fossil pollen evidence. This zone is characterized by the greatest relative abundance of hickory (*Carya*), dominant oak, pine, dogwood (*Cornus*), sweet gum (*Liquidambar*), tulip poplar (*Liriodendron*), black gum (or tupelo, *Nyssa*), and a marked decrease-to-absence of hazelnut (*Corylus*), hemlock

(*Tsuga*), and birch (*Betula*) in the pollen record (Küchler, 1964; Little, 1971, 1976). *Corylus* and *Tsuga* have their last appearances within this forest zone. Spruce (*Picea*), larch (*Larix*), fir (*Abies*), buffaloberry (*Shepherdia*) and burnet (*Sanguisorba*) are absent from this forest zone. The warmest part of this zone hosts the first appearance of palmetto (*Sabal*); this genus is present in this forest zone along the U.S. Atlantic coast from approximately Savannah, Georgia, to Cape Fear, North Carolina. This forest zone currently is established at lower elevations (below ~305 m elevation) in the study area, on warmer valley floors and across the Piedmont (Fig. 2). The relative temperature index values from our samples range upward from 140 (Fig. 5) and their upper limit has not yet been established.

It is from this combination of presence/absence, co-occurrence, and relative maximum abundance of the diagnostic pollen taxa in our samples that we have reconstructed probable forest zones and forest zone boundaries for the pollen assemblages in this study. As noted earlier, we are in the process of directly comparing the pollen assemblages from our Blue Ridge fossil samples to modern pollen assemblages taken from modern soils within each of the forest zones Küchler (1975) has mapped (Fig. 2). This cross test includes all the major forest types established along the eastern seaboard (including boreal forest in Newfoundland and the Gaspé Peninsula). Additionally, we are trying to characterize the in-situ pollen signal beyond the boreal forest, across the forest-tundra ecotone (zonal boundary).

Analytical methods

In order to interpret the Blue Ridge's climate history, we examined fossil pollen preserved in the Big Meadows core and pollen from outcrops uncovered by recent debris flow activity in the study area (Morgan and Wiczorek, 1996; Wiczorek et al, 2000; Litwin et al., 2001; Eaton, 1999; Eaton et al., 2003b; Litwin et al., 2001; Litwin et al., 2004a, 2004b; Smoot, 2004a, 2004b). The presence and relative abundance of particular pollen taxa within these samples permitted us to derive-- stepwise-- the relative temperatures under which each sample was deposited, the probable forest succession through time, and the probable climate history of the study area.

We isolated fossil pollen assemblages from core and outcrop samples using the laboratory procedures outlined in Litwin et al. (1993). Briefly, this involves decalcification of the sample in HCl, demineralization in HF, filtration of clays through 8 μm filters, specific gravity fractionation of organic and inorganic fractions in an aqueous ZnCl_2 suspension by centrifugation, oxidation, slide preparation, and light microscope identification of specimens. Each sample was characterized on the basis of a standard count of 300 pollen grains per sample unless otherwise indicated. More than 14,000 specimens were counted by the senior author for this study.

The samples were categorized both by relative age and by absolute age. AMS ^{14}C dating was used to determine an absolute age (or age range) where possible for each sample. Most of our outcrop samples bore enough macrofossil carbon fragments to permit independent dating of each sample. However, the Big Meadows sample set did not contain sufficient organic material for every sample to be dated independently. In lieu of this, we constructed an age model for the core samples based on those AMS ^{14}C

dates we were able to recover downhole. The core chronology we used is addressed later in this report (p. 20).

We reconstructed the relative thermal signal of each sample by developing a relative temperature index (noted earlier), based on the character of the fossil pollen assemblage within each Quaternary sediment sample. We experimented with a large number of ratios to find and construct a plausible relative measure of paleotemperature. One fairly promising model we constructed was based on the sum of the standard deviations of cold-loving taxa versus the sum of the standard deviations of warm-loving taxa in each pollen assemblage. That model was useful, but did not provide greatly different results than a simpler model that we present here; the standard deviations model will be presented elsewhere. The model we used here was based on a standardized count of 300 specimens. We tabulated the ratio of pollen specimens from selected warm arboreal taxa (*Quercus* and *Fraxinus*- oak and ash) versus the number of specimens from selected cold arboreal taxa (*Picea* and *Abies*- spruce and fir). This model, although fairly simple, provided a dependable and direct differentiation of our samples with respect to relative paleotemperature. Because some of the warmest pollen assemblages in this study had essentially no cold tree pollen taxa represented, we added a minimum denominator “correction” value of 1. Because oak pollen was strongly dominant in the samples of our study, this first-order “thermal index” initially was more sensitive in discriminating relative temperatures among pollen assemblages from warmer climate conditions. We corrected this preliminarily by plotting the index on a log scale, which increased the sensitivity for differentiating among cooler forest types, but somewhat decreased sensitivity for differentiating warmer forest types (see Fig. 4). Overall, this adjustment

enabled us to differentiate relative temperature (thermal signal) among samples more effectively (y-axis, Fig. 4). This index was established as a first relative means of differentiating our sample set. Note that the dataset presented in Fig. 4 is the compilation of both subsets- core and outcrop; the relative confidence levels in their chronology are detailed below.

Our samples cluster into three categories of dating accuracy. All samples that yielded radiocarbon dates of less than 20,265 yr BP were calibrated by the computer program CALIB, according to the methods proposed by Stuiver and Reimer (1993). This program corrected our radiocarbon ages for known fluctuations in atmospheric CO₂ through time, and also enabled calculation of 2-sigma error confidence limits on the calibrated ages. These samples are preceded by the letter “A” in their sample code (ex.: “ALRA7”, Fig. 4 and Table 1), and denote those ages which we think have the highest fidelity.

Radiocarbon samples that yielded ages in excess of 20,265 ¹⁴C yr BP exceeded the limits of this calibration window, although they also bear 2-sigma error confidence limits. These samples, although dated as accurately as possible, are undoubtedly somewhat less accurate than those samples for which calibration was possible. The code designation for this second type of samples is preceded by the letter “B” (ex.: “BNFMR6”).

Core samples for which no direct AMS ¹⁴C dating was possible –interpolated dates (between datable control points) for which no confidence limits were calculated-- are preceded by the letter “C” (ex.: “CAS”). These samples denote the lowest level of

chronological fidelity, but are still highly useful in determining the paleovegetation history of the Blue Ridge.

The study samples can be divided grossly into two time intervals: Holocene and Recent (0-10 ka), and the Pleistocene (10-45 ka). The Pleistocene-Holocene boundary is fixed at 10 ka by international convention (Harland et al., 1990). We have informally divided the Pleistocene interval in this report into two parts: the Late Glacial (15-45 ka), and the “transition” between the Late Glacial and the Holocene boundary (10-15 ka). This “transition” includes the Interstadial 1 warm event and the Younger Dryas cold event, as reported elsewhere (Harland et al., 1990; Dansgaard et al., 1993; Alley et al., 1993). For convenience of discussion we have further subdivided our samples into 5,000 yr increments, for a total of nine time windows (Fig. 4); each subinterval was examined for the range (extent) and frequency of forest change (see pp. 33-36).

Note that some of the carbon-bearing samples associated with the prehistoric debris flows did not yield pollen, and, therefore, yielded no proxy climate information. However, those carbon samples successfully established the age of the debris flows from which they were obtained, and indicate local periods of Quaternary deposition along the eastern flank of the Blue Ridge. They are plotted separately, adjacent to the x-axis (Fig.4, bottom).

Next we applied the presence, absence, and relative abundance criteria noted in the modern forest zonation (above) to our compiled pollen dataset (the relative temperature index graph), to derive Figure 5. The forest boundaries in Figure 5 were established internally using only our pollen dataset (Table 3). This zone assignment formed our interpretive model for characterizing the forest succession over the past 45 ky

BP. The positions of the forest boundaries (with respect to our relative temperature index, along the y-axis, Fig. 5) are preliminary; each is now in the process of being cross-tested and refined by the modern soil/ forest dataset. The model will be emended in the future as necessary. The ^{14}C calibration limit is noted on the graph at 20,265 yr BP (for reference). Because so many separate questions exist regarding the geologic history and vegetation prehistory of Big Meadows, we also examined the core data subset separately to answer these site-specific questions (below).

LOCAL INVESTIGATIONS AT BIG MEADOWS

Location

The Big Meadows-Upper Rapidan River study area is located in north-central Virginia along the eastern front of the Blue Ridge Mountains (Fig. 1). The core site is situated within the Big Meadows 7.5' topographical map (1:24,000 series). The main core was taken from near the geographical center of the meadow, at approximately 3460' elevation (Fig. 6). A second, shorter, polycarbonate-lined core was taken adjacent to the main core, in order to obtain an undisturbed sample of the very shallowest part of the section. The meadow core site is surrounded by Appalachian oak forest. Because our vibracoring locally disrupted sedimentary fabric within the core (fluidization) in the lower core runs, a 10-cm sampling interval was chosen as the closest sample spacing in the upper part of the core that was deemed likely to provide reliable (unmixed) results.

Character of the shallow subsurface

The shallow subsurface in Big Meadows is interpreted as a thin (<0.5 m) soil developed on the depositional interval below it, which is also partly incorporated into the soil. The remaining 4 meters consists of a matrix-supported mixture of oxidized sand, silt, and clay, with varying concentrations of poorly-sorted, poorly-stratified, angular-to-subrounded pebbles. The top of this lower layer is reduced below the surface soil layer. The bulk of the sediment penetrated by our coring probably was deposited by cryogenic processes (solifluction, J.Smoot, USGS, pers. comm.). Smoot interprets low-relief steps and lobes on the surface of Big Meadows to be the geomorphic expression of shallowly buried solifluction features-- relict cold-climate landforms that are still visible under younger sediment. Relict Pleistocene deposits generated by cryogenic processes are more abundant in the study area than previously was recognized (Eaton et al., 2003; Smoot, 2004a; Smoot, 2004b). Although these textures and features presently are found in areas subjected to persistent ground ice (permafrost), they also can develop in somewhat warmer settings.

It appears that the character of these shallow subsurface deposits also creates heterogeneity in the soil moisture in the meadow (with depth). The top of a relatively-impermeable layer of clay-rich sediment occurs at approximately 35 cm below ground surface, and acts as a shallow aquitard which permits the lowest elevations in the center of the meadow to become a functional marsh periodically throughout the year. Effectively this turns the soil just below the meadow ground surface into a shallow perched aquifer. The increased moisture and cooler microclimate (because of the 3460' elevation) permit the growth of *Sanguisorba canadensis* (Canadian burnet, a cool-climate

bog plant), *Betula populifolia* (gray birch), *Abies balsamea* (balsam fir), *Picea rubens* (red spruce) and *Cornus stolonifera* (Red-osier dogwood, which favors moist soils) to grow as refugial taxa on this ridgetop site (W. Cass, NPS, 2002, pers. comm.).

Thickness of Holocene/Pleistocene deposits

The vibracoring done at Big Meadows penetrated approximately 6 m of Holocene and Pleistocene-aged poorly consolidated sediment before refusal. Approximately 4.8 m of this interval was recovered as cored sediment. There was no indication that our coring reached bedrock (Catoctin Formation), so the maximum depth of Pleistocene and Holocene deposition on the Catoctin Fm. at Big Meadows is as yet undetermined. The Catoctin crops out around the edge of Big Meadows, especially to the south, and even sporadically within the meadow itself (Allen, 1963; Gathright, 1976; Rader and Evans, 1993). Quaternary deposits in the meadow therefore are non-uniform in thickness and distribution. Our coring suggests their thickness is a minimum of 6 m in the center of the meadow. Only the topmost sediments (upper 2.5 m) were sufficiently undisturbed by our coring to permit serial pollen analysis.

Age assignment of core samples and age-depth models

Like many of the outcrop samples we used in this study (Fig. 1), several samples from the Big Meadows core provided individual AMS ¹⁴C dates; as such, their chronology is the most strongly established. However, other samples from the two cores did not provide individual dates; these were time-calibrated by interpolation between the dateable samples (Tables 1, 2). In addition, carbon from several shallow archaeological

survey pits (from a study by C. Nash, James Madison University) also were analyzed by AMS to derive ^{14}C ages, in an effort to establish time control for the shallowest part of the sample set from the core (Table 1). Radiocarbon analyses were run on samples 3, 29, 36, 39, 53, 59, and 60 (see Table 1).

The results did not yield uniformly older dates with increasing depth. It is likely that some of the results were affected by contamination of reworked (older) carbon or by vertical mixing—collapse of sediment bearing younger carbon from a higher stratigraphic level down hole, during repenetration of the core hole. The first circumstance would give older than anticipated ages, and the latter would give younger than anticipated carbon ages at depth. The corrected ages and initial age results for radiocarbon samples we used in the Big Meadows core are shown in Table 2, along with all other radiocarbon samples used in this study. The samples are listed in Table 2 in order of increasing age; those that are preceded by an asterisk (also in boldface type) were used to develop age control in the Big Meadows core. The ^{14}C age results suggest that four different age models (Fig. 7) immediately may be proposed for those sediments recovered at Big Meadows.

The first model is built on the greatest possible number of ^{14}C AMS radiocarbon samples (samples 3, 5, 6, 29, 36, and 39), but favors the youngest possible age at depth. Using this age model, three of the four deepest samples (samples 53, 59 and 60) are “off-trend”. They would have to have been contaminated with young carbon (from upsection, perhaps during repenetration of the hole during coring-- samples 59 and 60), or with redeposited old carbon (‘dead’ carbon-- sample 53). Under this first age model, we would interpret the core section at Big Meadows to be almost entirely Pleistocene in age, and overlain only by a thin veneer of late-Holocene-to-Recent organic-rich soil. It requires

that three of the eight ^{14}C AMS radiocarbon analyses from the meadow be rejected as inaccurate. We note that two of the carbon dates used for age model 1 (samples 5 and 6) were averaged into a single value. The charcoal isolated from these two soil samples within Big Meadows were dated by ^{14}C AMS radiocarbon analysis in collaboration with the James Madison University (C. Nash) archaeological assessment: gridpoints 800N/475E (sample 5), and 800S/50E (sample 6, Fig. 6). The calibrated ages (per Stuiver and Reimer, 1993) for these samples were 2.61 ± 0.14 ky cal BP and 2.84 ± 0.08 ky cal BP, respectively. Averaged together, they set the age of the 30 cm depth interval at Big Meadows as 2.73 ± 0.14 ky cal BP. The very next core sample below this depth interval (sample 29) provided a calibrated age of 23.87 ± 0.77 ky cal BP, from a depth of 47 cm in Big Meadows. In this age model we therefore place a depositional hiatus in the core between 15 and 47 cm depth, and tentatively between 30 and 47 cm depth.

The second model is a variant of the first; it accepts sample 53 as one of the accurate age datums and suggests that sample 39, like samples 59 and 60, was affected by young carbon from re-penetration of the hole during coring. This model suggests that all “off-trend” ages are clustered at depth, and are similarly contaminated by ‘young’ carbon. It also requires that three of the eight total ^{14}C AMS radiocarbon analyses from the meadow to be rejected as inaccurate. However, it proposes a significantly older age for the 320 cm depth interval (~67 ka) than is proposed for the first age model (~44 ka). This increase in age requires a tenfold drop in the net deposition rate of Pleistocene sediment below ~135 cm core depth, suggesting a major change in depositional style occurred. We do not recognize a textural fabric change or grain size change within the core in that interval that might support this.

In the third age model, samples 3 and 59 are proposed to be the most accurate ages. This model does not mandate a hiatus (paraconformity) in the core. Age model three would require that more ^{14}C AMS radiocarbon analyses from the meadow be rejected (five of eight). Using this chronology, the Holocene/Pleistocene boundary would be placed between 106 and 116 cm depth in the core (on the basis of a standardized 10 ka age). This age model requires that some of the coldest vegetation signals in the core (samples 29 and 45 through 49) -- representing some of the coldest forest types -- had to have been established in Big Meadows during the Holocene, when mean annual temperatures were much too high to sustain these forest types in the study area.

Age model four proposes that the shallowest (sample 3) and deepest (sample 60) samples provide the most accurate ages, because they define the least possible total age span. This model suggests that “old” carbon affected the results of the intermediate-depth “off-trend” samples (five samples), in a non-linear way. This solution requires that five of the eight ^{14}C AMS radiocarbon analyses from the meadow be rejected. The Holocene/Pleistocene boundary also would be placed between 163 and 173 cm depth. As before, it would require that exceptionally cold-tolerant forests be established at Big Meadows during the warm climate of the Holocene.

We have adopted model 1 for use in this study for the following reasons. It incorporates the greatest number of radiocarbon ages, and the millennial-scale chronology (absolute time context) that it defines for the warm and cold pollen assemblages in the core is consistent with known larger climate trends in the latest Pleistocene and Holocene. Age model 2 could not be substantiated by textural changes in

the core, as required below ~135 cm core depth. Age models 3 and 4 require more carbon dates to be rejected, and also require the establishment of cold forests in the study area during the warmest interval of the Holocene. Because much of the Holocene interval was as warm or warmer than present climate conditions we find that scenario and those respective age models to be untenable. We think the lowest of the accumulation rates in this age model—0.8 cm/ky-- indicates the approximate position of a temporal hiatus (paraconformity), between 15 and 47 cm depth. The depth of this temporal hiatus varies. Much of the meadow has approximately 30-50 cm of Holocene sediment, expressed as the 'B2', 'B1', 'O' and 'A' soil horizons (below). In the core itself the hiatus is somewhat shallower, between 15-21 cm depth.

If we accept model 1, the lack of latest Pleistocene and early Holocene sediment in the core perhaps may be attributed to an early Holocene dry interval that is recognized elsewhere in eastern North America (Muller et al, 2003). This dry period may have affected the deposition or contributed to the erosion or deflation of the core site at Big Meadows through the 10-5 ky cal BP interval. It also may have contributed to the erosion of any sediment that had been deposited at the Big Meadows core site between 20-10 ky cal BP.

Pleistocene-Holocene vegetation history of Big Meadows

On the basis of the radiometric and fossil pollen evidence we recovered directly from the Big Meadows site, we suggest the following trends occurred in vegetation from approximately 37 ky BP to the present. The oldest pollen evidence we have from the

core so far (sample CAW, number 59¹) suggests that Big Meadows was vegetated by boreal forest at approximately 37 ka (Fig. 8; Tables 1 and 2). Big Meadows would have been vegetated similarly to what the highest peaks of the Gaspé Peninsula or Newfoundland are today. The meadow site warmed rapidly over the next 2000 years, transitioning through spruce-fir forest and northern hardwoods-spruce forest into northern hardwoods forest (sample CAV, number 58). This transitional interval would have been similar to the forests (and climates) currently found in Nova Scotia, central Maine, and upper New York state, respectively. From 36 ka to 28 ka the meadow experienced a net cooling all the way to boreal forest conditions. Three large cooling pulses occurring between 36 -33 ka (samples CAU and CAT, numbers 57 and 56), between 33-31 ka (samples CAS and CAR, numbers 55 and 54), and between 31-29 ka (samples BAQ and CAP, numbers 39 and 53). From 29.5-28 ka the meadow site was vegetated again by boreal forest (samples CAP and CAO, numbers 53 and 52). However, the evidence does not yet enable us to determine conclusively whether the Big Meadows site was unvegetated barrens, fully open alpine tundra, sparsely forested forest-tundra, or densely forested ridgetop during those past intervals when it was surrounded by boreal forest. It may have been cold enough that this ridgetop site was mostly open alpine tundra, with snowpack that melted late in the spring or summer, and was sparsely vegetated with small copses of shrubs or stunted trees (if so, then probably spruce, fir, larch, birch, buffaloberry and pine). Pollen evidence suggests it varied (see “*First occurrence of “modern” ...*”, p. 29). By approximately 26 ka the site once again experienced a substantial but very brief warming, back to northern hardwoods forest (sample BAM,

¹ Although sample 59 is noted as a control point in age model 3 of Fig. 7 (at 22.78 ka), its projected age in age model 1 (used here) is 37.15 ka.

number 36). Precipitous cooling transformed the site back to boreal forest by 25 ka, a change of three forest zones in a span of only 1000 years (samples CAL and CAK, numbers 51 and 50). Evidence suggests the forests in and near Big Meadows fluctuated between 25 and 22 ka (Figs. 8, 9), changing at least twice from boreal forest through spruce-fir forest to northern hardwoods-spruce forest, then back through spruce fir forest to boreal forest again (samples CAJ to AAD, numbers 49-45 and 29). This time interval (24.5-21.5 ka) contains the coldest pollen assemblages we have been able to document so far in the Blue Ridge study (samples BMAD1 and HCR1, numbers 31 and 26). The timing of these samples corresponds to a Northern Hemisphere insolation minimum (Berger, 1978; Fig. 9). These samples probably represent the vegetation in place at or very near to the time of the last glacial maximum (LGM) in the Northern Hemisphere, when the glacial terminus was only about 200 miles north of the study area.

The largest time gap we observed in the core spans from approximately 23.8 ka to ~2.73 ky cal BP. We cannot determine from core evidence alone what vegetation changes occurred in this interval. We also cannot determine if additional sedimentation had occurred at the site through part of this interval (such as during the latest Pleistocene between 23 and 10 ky cal BP), only to be stripped away later, during the middle Holocene. Core evidence by itself (sample ASHEN2, number 3) indicates that Appalachian oak forest, the forest type now surrounding Big Meadows, was in place at the site by ~2.73 ky cal BP. Outcrop evidence from elsewhere in and near Shenandoah National Park (samples AWRA, AWRD, and ALRA7, (numbers 11, 10, and 8)) provide evidence constraints for when Appalachian oak forest may have been first established in the study area (below).

AMS ¹⁴C dates: implications for archaeological studies

The radiometric dates we were able to obtain from the shallow subsurface at Big Meadows have important implications for documenting native human occupation in and around Big Meadows. Four usable dates were obtained from the top meter of sediment in the meadow (samples ASHEN2, '800N', '800S', and AAD (numbers 3, 5, 6, and 29; Tables 1 and 2)). Archeological excavations from James Madison University field crews (under the supervision of Carole Nash) have recovered approximately 15 diagnostic artifacts from Big Meadows so far, objects that represent human occupation prior to European settlement of this area. Of these, Nash reported only 2-3 objects that were buried in the A/B, B-1 or B-2 soil horizons; all others were found on the ground surface. Artifacts similar or identical to all three of the most deeply-buried of the "diagnostics" had been found elsewhere in the region in association with charcoal, and were determined to be younger than 6000 yr BP (C. Nash, 2003, JMU, pers. comm.).

The radiometric ages we obtained from the top meter of sediment were from 14 cm (number 3), 30 cm (numbers 5 and 6), and ~47 cm depth (number 29; see Table 1). These four samples suggest a depositional hiatus or unconformity exists between the deepest of these samples (number 29) and the next shallower ones (numbers 5 and 6). This hiatus is noted in the age/depth plot in Fig. 7. Comparison of the depth of the oldest diagnostic objects found in the meadow with the depth and ages of the radiometric dates we obtained indicate that all diagnostics were found above our radiometric sample at ~47 cm depth, and therefore all were found above this unconformity. The oldest dates we were able to document above the unconformity are 2.61 ± 0.14 ky cal BP and $2.84 \pm$

0.08 ky cal BP. Both of these samples are from ~30 cm depth, but from opposite sides of the meadow (Fig. 6). They provide the averaged age of 2.73 ± 0.14 ky cal BP for the 30 cm depth horizon in our age/depth model, just above the hiatus (above, Fig. 7).

The position and ages of the radiometric evidence leads us to suggest that no true depositional context can be expected for any artifacts older than $\sim 2.73 \pm 0.14$ ky cal BP. Any artifacts found in the meadow that bear relative ages older than this very likely will have been deposited directly on the non-depositional unconformity surface in the meadow, in a time interval between 23.87 ± 0.77 ky cal BP and $\sim 2.73 \pm 0.14$ ky cal BP. To state this differently, we predict that all vertical succession (increased relative age with depth) for human artifacts found in Big Meadows will terminate regularly between 30 -50 cm depth below ground surface, or about 2.73 ± 0.14 ky cal BP.

First occurrence of “modern” (presently established) forest

One of the questions we hoped to answer in this study was the timing of the first occurrence of Appalachian oak forest in the Blue Ridge, within the area of our study. Our dataset is still sparse through the Pleistocene- Holocene boundary (~ 10 ky cal BP), and we cannot yet document that forest transition directly. The best present evidence we have that can address that question is two outcrop samples from nearby Wilson Run (Table 1, Fig. 1). Samples AWRD and AWRA (numbers 10 and 11) suggest that northern hardwoods-spruce forest was still present in the study area between approximately 12 ky cal BP and 13 ky cal BP (Fig. 5). These samples fall within the Younger Dryas climate interval, the last major cold pulse before the late Pleistocene final deglaciation. This suggests that Appalachian oak forest likely became established in the

study area only more recently than 12 ky cal BP. An earlier pollen study done by Craig (1969) suggested that the southern part of the Shenandoah Valley changed from forests heavily populated with spruce and pine to forests dominated by oak approximately 9520 \pm 200 BP. When we correct Craig's radiometric date for atmospheric CO₂ flux (per Stuiver and Reimer, 1993), it suggests that the time of that transition may be closer to 10.748 \pm 510 ky cal BP (2 sigma error). The older end of this calibrated age (11.258 cal ky BP) is fairly close to some ages suggested for the young boundary of the Younger Dryas (Alley et al., 1993; Freidrich et al., 2001; Hajdas et al., 2003). The very earliest date that Appalachian oak forest could have existed in the study area is bounded by a third Wilson Run sample, sample AWRB2 (number 13). That sample suggests that a cold tolerant forest--northeastern spruce-fir forest-- was established in the study area at approximately 17 ky cal BP (Fig. 9). It is feasible, but as yet undocumented, that Appalachian oak forest first could have existed in the study area briefly, during the more sparsely documented interval between 17 ky cal BP and 13 ky cal BP.

Evidence of prehistoric meadows

Although a hiatus in sedimentation exists at Big Meadows, we can make an indirect assessment of when in the past this site may have been a meadow. There appear to be several times when this may have been likely- each may have been intermittent, and lasted less than a millennium. The first opportunity may have been times of the most intense cold climate in the study area. We are still trying to determine how cold the ridge tops became during periods of maximum glaciation in the Northern Hemisphere. It may have been cold enough that they could have become alpine meadow, alpine tundra or

forest-tundra type open parkland. This is still equivocal; solifluction deposits in the meadow may not have required permafrost to form (French, 1996, p.151), suggesting that the climate could have been warmer than the -1° C mean annual temperature isotherm (the southern limit of permafrost, Figs. 2, 3; Brown et al., 1997). If it had been alpine meadow, alpine tundra, or open parkland during the coldest periods, the most likely intervals would be 37.5-36.5 ka, 30-28 ka, and intermittently between 25.5-22 ka.

Fossil pollen provides some limited evidence to address this directly. Five samples from the Big Meadows core probably represent boreal forest conditions. We tested these five core samples by comparing two pollen indicators within each of them to the values of those same two indicators in the present-day soil at Big Meadows. The two measures for comparison were the percentage of grasses (Poaceae) and the percentage of all herbaceous taxa (here, Poaceae, composites, and amaranths). The surface soil in the present-day Big Meadows contains approximately 9% grass pollen, and a combined herbaceous pollen total of 18%. Only one of the five 'boreal forest' core samples had percentages that approached these modern analogue values. Sample CAK (number 50) contained approximately 13% grass pollen and 13% total herbaceous pollen. We think this sample most probably represents open (alpine) meadow surrounded by boreal forest, at approximately 25.17 ka. At the moment, the present evidence collected for this study does not enable us to answer conclusively whether meadow or alpine tundra existed at this ridgetop during the peak of the last glaciation (last glacial maximum).

Another set of conditions that might favor meadow development could be those brief periods of intense climate change. These are identified by large shifts in forest type over a relatively short period (1-2 ky duration). Open meadow may have developed at

the Big Meadows site because the climate might have been shifting faster than forests could adapt to the environmental stress of rapidly changing temperature and precipitation. During such times, shorter-generation plants-- herbaceous plants, grasses, and low shrubs—might tolerate such stresses relatively better than closed-story forest, which would die off and create an open-story environment advantageous for opportunistic herbs and grasses. A series of strong or abrupt shifts are found at 37.5-35.5 ka, 33-29 ka, 26-23 ky ka, and 21.5-20 ky BP (Fig. 8). We have identified eight possible times in our core sample set when the Big Meadows site may have been a meadow, based on the same two pollen indicators used above. Three of these fossil samples exceeded modern analogue values in both criteria. Because of this, open meadow almost certainly occurred at the site at those times- 2.3 ky cal BP (sample ASHEN2, number 4), 23.87 ka (sample AAD, number 29), and 31.7 ka (sample CAR, number 54; Fig 9, Table 1). The latter two samples are from intervals of strong climate shifts, and suggest the site was open meadow surrounded by northern hardwoods-spruce type forest. The first of these three samples is of Holocene age, and probably represents temperate open meadow surrounded by Appalachian oak forest (as today). Three totally different core samples did exceed the modern percentage of grasses with respect to their pollen totals, but did not meet or exceed the total herbaceous percentage of our modern analogue soil sample. Accordingly, these probably (but less definitely) represent meadow; they occurred at 24.21 ka (sample CAF, number 45), 24.81 ka (sample CAI, number 48), and 25.17 ka (sample CAK, number 50, noted earlier). The first two were surrounded by northern hardwoods-spruce forest and the third was surrounded by boreal forest. The final two core samples have values that approach- but are somewhat less than- the modern meadow

values for both pollen criteria. These both represent northern hardwoods forest and were found at 25.53 ka (sample BAM, number 36) and 35.85 ka (sample CAV, number 58). These two samples may represent intervals of meadow, or of open-story woodland at the site. All of the likely prehistoric meadow horizons are noted in a schematic of the Big Meadows core (Figure 8).

VEGETATION RESPONSIVENESS TO CLIMATE

Compiled vegetation history

We have attempted to address questions of scale and context regarding the Blue Ridge's vegetation responsiveness to 'geologic-scale' climate change by compiling our core and outcrop datasets (Figs. 4, 5, and 9). We have done this by integrating our samples on the basis of (calibrated) radiometric age. By necessity, those samples whose age exceeded present calibration limits, or which were derived through interpolation (core samples), are plotted with different symbols. Because the sample set is generally well-constrained with independently-derived ages, and is recovered from a localized geographic area, we have used it as a compiled dataset to synthesize the vegetation history (proxy climate history) of the Blue Ridge in the study area (Figs. 9, 10).

Range of forests per time increment

We were able to document and to compare the probable number of different forest zones that existed in the study area during each successive 5 ky increment from 45 ka to the present (Fig. 9). We counted all forest zone changes linearly—i.e., where we possessed pollen evidence of two geographically non-contiguous forest types within a specific time

increment, we acknowledged the implicit presence of all intermediate forest types, occupying the temporal position between those two data points. For example, prehistoric boreal forests in the study area that experienced climatic warming did not directly become Appalachian oak forest, but probably transitioned through three other forest phases first (Figs. 5, 9). This permitted us to estimate the total vegetation variability within each time increment, as an approximate measure of the high and low temperature extremes (amplitude of change) occurring within it.

Some 'range-through' occurred across time interval boundaries. In several instances, the youngest documented sample in a given time increment represented a different forest zone than the oldest documented sample in the next younger time increment. In order for us to derive an estimate that did not overcount or undercount the probable number of forest zones that occurred within each time increment, we systematically assigned all such boundary 'range-through' to the next younger zone. The only exception to this was the time interval 10-5 ky cal BP; no dated pollen evidence fell within this interval at all, so we deferred assigning a forest zone range value to it. The range of forest types observed within these 5 ky 'windows' varied, partly as a function of data saturation.

The increment 45-40 ka bore only one forest type, whereas the next younger interval (40-35 ka) bore four—all of them colder than the forests presently established in the area (Fig. 9). 35-30 ka provided evidence of two forest types: the next younger interval (30-25 ka) produced four forest types within its 5 ky 'window'. The absolute coldest forest types noted in this study occurred between 25-20 ka; this time interval provided evidence of four different forest types. The interval 20-15 ky cal BP provided

evidence of three forest types, but importantly, they generally were warmer forest types than would be predicted if the Last Glacial Maximum (LGM) occurred at 18 ky cal BP, a commonly accepted date (CLIMAP, 1976; Dreimanis, 1977; Delcourt and Delcourt, 1981). We think the LGM is now more appropriately placed approximately 25.5-20.5 ky cal BP, perhaps even 24.5-21.5 ky cal BP, where the strongest evidence of boreal forest occurs in our data set, and where Northern Hemisphere solar insolation was at or near minimum values (Berger, 1978; Fig. 9). The interval 15-10 ky cal BP shows evidence of two forest zones (one of them a range-through), but the next younger interval (10-5 ky cal BP) provided no direct evidence of the range of its forests. Interpolation of those samples immediately older and younger than this interval suggests that a minimum of two forests may be inferred in this data gap, but in the absence of any fossil evidence in this interval at all, we assigned no forest range value. Finally, the most recent time increment (5-0 ky cal BP) provided evidence of two forest types. We qualify this by noting that oak-hickory-pine forest presently is established in the lowest elevations of the study area (below ~1000' elevation) and sample 4 may be assigned marginally to that forest type.

These fossil assemblages (Fig. 9) indicate that many time intervals within the Late Pleistocene portion of our study experienced a variety of forest conditions, and not just uniformly cold boreal forest. This suggests climatic fluctuation was common during the Late Pleistocene. In addition to determining forest range within each increment, we also tried to determine how frequently forests shifted in the study area over the past 45,000 years.

Frequency of forest change

Whereas the range of forest change was an approximate measure of the amplitude of climate change, frequency of forest change is an approximate measure of how often climate changed enough to force a turnover in forest type. On the basis of the data accumulated at present, the number of forest zone changes observed appears to be in part a function of the data saturation within any given time window. Although our data are still accumulating, we have seen evidence of a minimum of 37 forest shifts in the study area over the past 40,000 years, suggesting that forests experienced climate-related turnover on average less than every 1100 years (Fig. 9). Up to 14 forest zone changes (13 plus 1 range-through) can be documented within one 4-ky time increment (26-22 ka), suggesting that forest composition may have responded to climatic changes in the study area as frequently as every ~285 years. This high frequency of change is similar to the rates found in two shorter time intervals in the core. Three different forest zones have occupied the study area within the 500-year period from 26-25.5 ka, and up to four different forest zones have occupied the study area within the 500-year period from 25.5-25ka. These indicate vegetational response rates (forest changes) of 0.17 ky and 0.125 ky, respectively. From this we conclude that the vegetation shifts probably were in phase or nearly in phase with climate changes, and that no appreciable vegetational 'lag' was present. This recently also has been documented in Switzerland, in central and southern Europe, in the Netherlands, and in the Canadian Maritime region (Ammann et al, 2000; Freidrich et al., 2001; Hoek, 2001; Williams et al., 2002). This 'in phase' trend of vegetation response to climate does not appear to conform to the vegetation

disequilibrium model suggested for southern Ontario, which simulated a vegetation lag response to climatic warming since the Little Ice Age (Campbell and McAndrews, 1993).

We also made an initial estimate of the frequency of change within each of the 5 ky subintervals (Figs. 5, 9). Evidence of forest changes varied considerably among the time increments. The oldest increment, 45-40 ka, had no evidence of forest change. The next younger increment, 40-35 ka, had six shifts. 35-30 ka had one forest zone shift. The next increment, 30-25 ka, had eight. The cold interval between 25-20 ky cal BP recorded sixteen forest zone shifts (for an average forest shift frequency of 312 years, Fig. 9). The period of 20-15 ky cal BP recorded three full transitions. The period of 15-10 ky cal BP recorded 1 full transition. No forest transitions were directly recorded from 10-5 ky cal BP (due to a data gap), but inference suggests the interval had to have included at least 2 transitions. The final increment recorded 2 full transitions, albeit marginally. The maximum frequency of forest change within any of these defined 5 ky increments (312 years) is a similar sub-millennial scale to the vegetation response rates we noted spanning shorter time intervals in the dataset (125 and 170 years). The absolute magnitude of these response times is about the lifespan of a single tree; it suggests that climate change stresses certainly acted on the individual plants in these forests at century-scale or perhaps decade-scale. It also suggests that climate probably was a steady and persistent agent of change in these forests over the studied time interval (0-45 ka). Accordingly, we suspect that Blue Ridge forests may have changed significantly in composition -- with sub-millennial frequency-- throughout the 45-ky interval of our study. We predict that such shorter rates of change likely will be documented within additional time increments as we are able to increase the robustness of our data set through those intervals.

Discussion

It would be highly implausible for forests to change at sub-millennial frequency if forest change occurred only at the ecotone (fringe) between two forest zones, as an advancing or retreating “wavefront”. Such a mechanism could not shift forests along a changing climatic gradient with the frequency we observed in our data. Instead, one of us (RJL) thinks the mechanism that enables forest zones to change in-phase or nearly in-phase with climate shifts must happen internally and locally or extra-locally within each forest zone, within embedded microenvironments. The key to vegetative responsiveness to climate very likely can be found in the distribution of refugial plants such as those at Big Meadows. Plants that occupy environments that are at the extremes of their climatic tolerance can expand their ranges quickly and exponentially when climate does shift toward conditions more favorable to their survival and population growth. This “embedded” refugial growing stock, along with ungerminated (dormant) seeds from rare or recently extirpated plants in the soil, probably enables fairly rapid restructuring of plant communities and whole forests in a “mosaic-style” or kaleidoscopic transformation. For example, sustained strong climatic cooling at the Big Meadows site would favor the growth and expansion of populations of balsam fir (*A. balsamea*), red spruce (*Picea rubens*), Canadian burnet (*Sanguisorba canadensis*), red-osier dogwood (*Cornus stolonifera*), eastern hemlock (*Tsuga canadensis*), eastern white pine (*Pinus strobus*), bigtooth aspen (*Populus grandidentata*), and gray birch (*Betula populifolia*) at the expense of populations of blackjack oak (*Quercus marilandica*), scarlet oak (*Quercus coccinea*), chestnut oak (*Quercus prinus*), post oak (*Quercus stellata*), pin oak (*Quercus*

palustris), American holly (*Ilex opaca*), mulberry (*Morus rubra*), pignut hickory (*Carya glabra*), mockernut hickory (*Carya tomentosa*), black walnut (*Juglans nigra*), shortleaf pine (*Pinus echinata*), and Virginia pine (*Pinus virginiana*). All occur within the Appalachian oak forest, but taxa in the first group are much more characteristic elements of cooler forest zones, such as northern hardwoods forest, northern hardwoods-spruce forest, etc. Such a mosaic pattern may be seen near the present southern boundary of the boreal forest zone (Larsen, 1980).

Another mechanism that apparently may enhance and hasten forest transformation is a “leveraging” effect of increased sensitivity when tree taxa are shifted to the very limits of their climatic tolerance. They not only suffer increased mortality due to physical climate stress (changes in temperature and precipitation), but more rapidly succumb to biologic stressors such as insect and fungal attack. Examples of a biological leveraging effect on the mortality rates of already climatically-stressed taxa probably include the current insect infestations of hemlock wooly adelgid (*Adelges tsugae*) on the eastern hemlock, balsam wooly adelgid (*Adelges picea*) on the fraser fir population of the Great Smoky Mountains, and red oak borer (*Enaphalodes rufulus*) on northern red oak and black oak, as well as fungal infestations of dogwood anthracnose (*Discula destructiva*) on flowering dogwood, beech bark disease (*Nectria coccinea* var. *faginata*) on American beech, and butternut canker (*Sinococcus clavigigenti-juglandacearum*) on butternut. Examination of the most abundant pollen (plant) taxa in our samples show that each has fluctuated in abundance in response to climate throughout the entire geologic interval we studied, frequently over rather short geologic intervals. Figures 8-10 illustrate this biological responsiveness; Figure 10 is a local synthesis of the climate-

driven vegetation response in the Blue Ridge study area over the past 45 ky. To our best knowledge it is the longest such synthesis ever reconstructed for the Blue Ridge.

Minimum range of climate change

Current fossil evidence suggests that a total of six different forest types can be documented to have existed in the study area over the past 45 ky (Fig. 9). The data suggest that the area experienced climatic conditions ranging from those that supported a warm-temperate oak-hickory-pine forest to those that supported boreal (taiga) conditions, during the past 45 ky. As an initial estimate, this suggests that the mean annual temperatures (MATs) in the study area varied *at least* 11.5^o C (20^o F) over this 45-ky interval (the approximate temperature range spanning the coldest limit of oak-hickory-pine forest (~13^o C or 55^o F) to the warmest limit of the low boreal forest (~1.5^o C or 35^o F); Fig. 2). The evidence also indicates that this serial vegetational change (proxy climate change) was not gradual or unidirectional through this time period. Forests of nearly every type were developed repeatedly in the study area, especially forest types that were markedly colder than those presently established there. Several short temporal intervals (1-2 ky) show evidence of abrupt warming or cooling (spanning three or four different forest zones). The evidence demonstrates that the late Pleistocene climate in the study area was *not* uniformly cold, nor was Holocene climate uniformly warm in the study area.

A modern geographic analogue that spans the minimum climatic range that we can document in our fossil evidence (to date) would extend from the base of the Blue Ridge Mountains in Virginia to central Ontario or Newfoundland. From the current

evidence we cannot demonstrate conclusively that the climate in the immediate study area got cold enough for permafrost to develop during this 45-ky time interval. We currently are in the process of testing and refining our model, through finer-scale pollen analyses of another local core of full-glacial age (from Kinsey Run). We are comparing the pollen assemblages within it to pollen in modern soil samples from low, middle, and high boreal forest. It is within the zone of high boreal forest that climate conditions support the establishment of permafrost; this corresponds approximately to the -1°C mean annual temperature isotherm (Brown et al., 1997). This core also should provide an additional test of the shorter-term responsiveness of vegetation to climate change.

CONCLUSIONS

We conclude the following from the current results of our climate history study component for the landslide hazards study of the Blue Ridge Province:

A.) Forests currently developed on the Blue Ridge are not characteristic of forests that have been present in this area over the past 45,000 years. Evidence preserved in and around Shenandoah National Park suggests that demonstrably cooler forests were established in the study area from approximately 45,000 years ago to as recently as 12,000 years ago.

B.) Appalachian oak forest probably existed along the Blue Ridge intermittently over the past 15,000 years, and commonly over the past 10,000 years. We found no evidence that this forest type existed in the study area at all prior to 15 ky cal BP. During the warmest part of the Holocene, from approximately 6 ka to 4 ka, the ridgetop forests may have shifted periodically to oak-hickory–pine forest or even to southern mixed

forest. No direct evidence has yet been found to corroborate this prediction, however; studies are ongoing.

C.) Forest zones have changed rapidly and frequently in composition over the time interval studied. Our composite climate history evidence, although still fragmentary and quite incomplete, already documents approximately 37 forest zone changes in the study area over the past 45,000 years. Several intervals indicate that forests in the study area may have responded to climate change as frequently as every 125-300 years. Late Pleistocene forests in the Blue Ridge were not uniformly cold, but fluctuated strongly in response to regional-scale climate shifts, over short geologic time intervals.

D.) We have demonstrated that it is possible to create a moderately long serial record of Pleistocene to Holocene vegetation change for higher elevation regions in the eastern U.S., despite the general lack of individual sites favorable for recovering long Quaternary core records.

E.) The *minimum* range of climate change established in the study area over the past 45,000 years-- on the basis of the coldest and warmest modern analogue forest types we recognize in our prehistoric samples-- probably exceeds a range of 11.5⁰ C (20⁰ F) in mean annual temperature. The current MAT of the study area (at the base of the Blue Ridge) is approximately 13⁰ C (55⁰ F). We note these figures are preliminary- we are still working to document and to quantify the most probable absolute temperature shifts for the study area.

F.) Our evidence currently supports an older age for the last glacial maximum (LGM) in the Northern Hemisphere, based on radiometrically-dated pollen assemblages ~ 200 miles south of the LGM terminal moraine. The coldest vegetation assemblage we

recognize in this study definitely occurs from 20.5-25.5 ky cal BP, and possibly from 21.5-24.5 ky cal BP. Our evidence does not appear to support placement of the last glacial maximum at 18 ka.

G.) Archaeological evidence appears to compare somewhat equivocally with radiometric evidence we have recovered at shallow depth in the Big Meadows core. At present we think the oldest relatively dated archaeological tool evidence was deposited on a non-depositional surface (unconformity) that may have developed at Big Meadows during the warmest interval of the Holocene. This unconformity may represent as much as 20 ky of geologic time.

H.) Clay-rich intervals at shallow depth in the meadow, deposited during the Late Pleistocene by solifluction and/ or gelifluction, created an aquitard that now increases soil moisture at the ground surface of the meadow throughout much of the year, in the lowest parts of the depression. This perched water table, combined with the cooler microclimate of the high elevation ridgetop (~3500') helps to enable the growth of refugial taxa in the meadow, specifically *Sanguisorba canadensis* (Canadian burnet, a cool-climate bog plant), *Betula populifolia* (gray birch) and *Cornus stolonifera* (Red-osier dogwood, which favors moist soils).

ACKNOWLEDGEMENTS

We most gratefully acknowledge Gary Somers (Chief, Natural and Cultural Resources, NPS, Shenandoah National Park) for permission to conduct this climate history study in Big Meadows and surrounding areas of the Park. We additionally thank Wendy Cass and Reed Engle (botanist and historian, NPS, Shenandoah National Park), Tom Blount (NPS,

Big South Fork, TN) and Carole Nash (archaeologist, James Madison University), for their assistance in recovering the Big Meadows core. We thank Nancy Durika (USGS) for field assistance and for processing pollen samples for study by the authors (Litwin). We gratefully acknowledge Joe Smoot and Milan Pavich (USGS), and Wendy Cass (NPS), for reviewing the manuscript and for suggesting improvements.

REFERENCES CITED

- Allen, R.M., 1963. Geology and mineral resources of Greene and Madison Counties. *Virginia Division of Mineral Resources, Bulletin 78*, 102 p.
- Alley, R., Meese, D.A., Shuman, C.A., Gow, A.J., Taylor, K.C., Grootes, P.M., White, J.W.C., Ram, M., Waddington, E.D., Majewski, P.A, and Zielinski, G.A., 1993. Abrupt increase in Greenland snow accumulation at the end of the younger Dryas event. *Nature*, 362:527-529.
- Amman, B., Birks, H.J.B., Brooks, S.J., Eicher, U., von Grafenstein, U., Hofmann, W., Lemdahl, G., Schwander, J., Tobolski, K., and Wick, L., 2000. Quantification of biotic responses to rapid climate changes around the Younger Dryas- a synthesis. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 159:313-347.
- Anderson, R.C., 1974. The chestnut pollen decline as a time horizon in lake sediments of eastern North America. *Canadian Journal of Earth Science*, 11:678-685.
- Bassett, I. J., Crompton, C.C., and Parmalee, J.A., 1978. An atlas of airborne pollen grains and common fungus spores of Canada. *Canada Department of Agriculture, Monograph No. 18*, 321 p.
- Berger, A.L., 1978. Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quaternary Research*, 9:139-167.
- Braun, E.L., 1950. Deciduous forests of eastern North America. Hafner Press, Macmillan, New York, 596 p.
- Brown, J., Ferrians, O.J., Heginbottom, J.A., and Melnikov, E.S., 1997. Circum-Arctic map of permafrost and ground-ice conditions. *U.S. Geological Survey Circum-Pacific Map Series*, Map CP-45.
- Campbell, I.D., and McAndrews, J.H., 1993. Forest disequilibrium caused by rapid Little Ice Age cooling. *Nature*, 366:336-338.
- CLIMAP Project members, 1976. The surface of the ice-age Earth. *Science*, 191:1131-1144.
- Craig, A.J., 1969. Vegetational history of the Shenandoah Valley, Virginia. *United States Contributions to Quaternary Research. Geological Society of America Special Paper 123*:283-296.
- Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjörnsdottir, A.E., Jouzel, J., and Bond, G., 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature*, 364: 218-220.

- Delcourt, P.A., and Delcourt H.R., 1981. Vegetation maps for eastern North America: 40,000 YR B.P. to the present. In: Romans, R. (ed.), *Geobotany II*, Plenum Press, New York, p. 123-166.
- Dreimanis, A., 1977. Late Wisconsin glacial retreat in the Great Lakes Region, North America. *Annals of the New York Academy of Science*, 288:70-89.
- Eaton, L.S., 1999. Debris flows and landscape evolution in the upper Rapidan Basin, Blue Ridge Mountains, Virginia. PhD. Thesis, Univ. VA, 154 p.
- Eaton, L.S., and McGeehin, J.P., 1997. Frequency of debris flows and their role in long term landscape evolution in the central Blue Ridge, Virginia. *Geological Society of America, Abstracts with Programs*, 29(6):410.
- Eaton, L.S., Morgan, B.A., Kochel, R.C., and Howard, A.D., 2003. Role of debris flows in long-term landscape denudation in the central Appalachians of Virginia. *Geology*, 31(4):339-342.
- Eaton, L.S., Morgan, B.A., Kochel, R.C., and Howard, A.D., (2003). Quaternary deposits and landscape evolution of the Blue Ridge of Virginia. *Geomorphology*, 135:1-16.
- Ecoregions Working Group, 1989. Ecoclimatic regions of Canada, first approximation. Environment Canada, *Ecological Land Classification Series*, 23, 118 p.
- Eyre, E.F., (ed.), 1980. Forest cover types of the United States and Canada. *Society of American Foresters*. Washington, DC. 148 p.
- French, H.M., 1996. The periglacial environment. Addison Wesley Longman Limited, Essex. 341 p.
- Friedrich, M., Kromer, B., Kaiser, K.F., Spurk, M., Hughen, K.A., and Johnsen, S.J., 2001. High-resolution climate signals in the Bølling-Allerød Interstadial (Greenland Interstadial 1) as reflected in European tree-ring chronologies compared to marine varves and ice-core records. *Quaternary Science Reviews* 20:1223-1232.
- Gathright, T.M., II, 1976. Geology of the Shenandoah National park, Virginia. *Virginia Division of Mineral Resources, Bulletin* 86, 93 p.
- Hajdas, I., Bonani, G., Moreno, P.I., and Ariztegui, D., 2003. Precise radiocarbon dating of Late-Glacial cooling in mid-latitude South America. *Quaternary Research* 59:70-78.
- Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., and Smith, D.G., 1990. A geologic time scale 1989. Cambridge, Cambridge University Press, 263 p.
- Hoek, W.Z., 2001. Vegetation response to the ~14.7 and ~11.5 ka cal. BP climate transitions: is vegetation lagging climate? *Global and Planetary Change* 30:103-115.
- Küchler, A.W., 1964. Potential natural vegetation of the conterminous United States. *American Geographical Society, Special Publication* 36, 116 p.
- Küchler, A.W., 1975. Potential natural vegetation of the conterminous United States (map). *American Geographical Society*.
- Larsen, J.A., 1980. The boreal ecosystem. Academic Press, London. 500 p.
- Lewis, W.H., Vinay, P., and Zenger, V.E., 1983. Airborne and allergenic pollen of North America. The Johns Hopkins University Press, Baltimore. 254 p.

- Lieux, M. H., 1980a. An atlas of pollen of trees, shrubs, and woody vines of Louisiana and other southeastern states, part I. Ginkgoaceae to Lauraceae. *Pollen et Spores*, 22(1):2-57.
- Lieux, M. H., 1980b. An atlas of pollen of trees, shrubs, and woody vines of Louisiana and other southeastern states, part II. Platanaceae to Betulaceae. *Pollen et Spores*, 22(2):191-243.
- Lieux, M. H., and Godfrey, 1982. An atlas of pollen of trees, shrubs, and woody vines of Louisiana and other southeastern states, part III. Polygonaceae to Ericaceae. *Pollen et Spores*, 24(1):21-64.
- Lieux, M. H., 1982. An atlas of pollen of trees, shrubs, and woody vines of Louisiana and other southeastern states, part IV. Sapotaceae to Fabaceae. *Pollen et Spores*, 24(3-4):331-368
- Lieux, M. H., 1983. An atlas of pollen of trees, shrubs, and woody vines of Louisiana and other southeastern states, part V. Lythraceae to Euphorbiaceae. *Pollen et Spores*, 25(3-4):321-350.
- Little, E.L., Jr., 1971. Atlas of United States trees. Volume 1. Conifers and important hardwoods. *U.S. Department of Agriculture, Misc. Publ.* 1146.
- Little, E.L., Jr., 1976. Atlas of United States trees. Volume 4. Minor eastern hardwoods. *U.S. Department of Agriculture, Misc. Publ.* 1342.
- Litwin, R.J., Frederiksen, N.O., Adam, D.P., Andrie, V.A.S., and Sheehan, T.P., 1993, Continental-marine correlation of Late Pleistocene climate change: Census of palynomorphs in core OL-92, Owens Lake, California. In: Smith, G.I., and Bischoff J.L., eds., Core OL-92 from Owens Lake, southeast California: *U.S. Geological Survey Open-File Report* 93-683, p. 333-391.
- Litwin, R.J., Morgan, B.A., Eaton, L.S., Wiczorek, G.R., and Smoot, J.P., 2001. Proxy climate evidence from late Pleistocene deposits in the Blue Ridge of Virginia, In: Adams, D.B. and others, U.S. Geological Survey Appalachian regional integrated science workshop proceedings. *U.S. Geological Survey Open File Report* 01-0406, p.78.
- Litwin, R.J., Morgan, B.A., and Eaton, L.S., 2004a, Vegetation trends in the Blue Ridge of Virginia, 0-45 ka, and the role of fossil pollen as proxy indicators of Pleistocene and Holocene climate. *Upland archaeology in the East: Symposium IX, James Madison University, Harrisonburg, VA. Program with Abstracts*, p. 13.
- Litwin, R.J., Morgan, B.A., Eaton, L.S., and Wiczorek, G.R., 2004b, Proxy climate trends in the Blue Ridge of Virginia, 0-45 ka, and comparison to the Greenland ice-core record. *Geological Society of America, 39th Annual Meeting, Northeastern Section, Washington, D.C., Final Program* (online). ([www.http://GSA.confex.com/gsa/2004NE/final_program/abstract_69931.htm](http://GSA.confex.com/gsa/2004NE/final_program/abstract_69931.htm)).
- Morgan, B.A., and Wiczorek, G.R., 1996. Debris flows and landslides resulting from the June 27, 1995, storm on the North Fork of the Moormans River, Shenandoah National Park, Virginia: *U.S. Geological Survey Open-File Report* 96-503.
- Morgan, B.A., Wiczorek, G.R., and Campbell, R.H., 1999a. Map of rainfall, debris flows and flood effects of the June 27, 1995, storm in Madison County, Virginia: *U.S. Geological Survey Investigations Series* I-2623-A.
- Morgan, B.A., Wiczorek, G.R., and Campbell, R.H., 1999b. Historical and potential debris flows and flood hazard map of the area affected by the June 27, 1995,

- storm in Madison County, Virginia: *U.S. Geological Survey Investigations Series I-2623-B*.
- Muller, S.D., Richard, P.J.H., Guiot, J., de Beaulieu, J.-L., and Fortin, D., 2003. Postglacial climate in the St. Lawrence lowlands, southern Québec: pollen and lake level evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 193:51-72.
- Porsild, A.E., and Cody, W.J., 1980. Vascular plants of Continental Northwest Territories, Canada. National Museums of Canada, Ottawa, 667 p.
- Rader, E.K., and Evans, N.H., (eds.), 1993. Geologic map of Virginia- expanded explanation. *Virginia Division of Mineral Resources*, 80 p.
- Ritchie, J.C., 1987. Postglacial vegetation of Canada. Cambridge University Press, New York. 178 p.
- Smoot, J., 2004a. Sedimentary characteristics of Late Pleistocene periglacial stratified-slope deposits in the Blue Ridge of central Virginia. *Geological Society of America, 39th Annual Meeting, Northeastern Section, Washington, D.C., Final Program* (online). ([www.http://GSA.confex.com/gsa/2004NE/finalprogram/abstract_70146.htm](http://GSA.confex.com/gsa/2004NE/finalprogram/abstract_70146.htm)).
- Smoot, J., 2004b. Sedimentary fabrics of stratified slope deposits at a site near Hoover's camp, Shenandoah National Park, Virginia. *U.S. Geological Survey Open-File Report 2004-1059*. 52 p.
- Stuiver, M., and Reimer, P.J., 1993. Extended ¹⁴C database and revised CALIB radiocarbon calibration program. *Radiocarbon*, 35:215-230.
- Traverse, A., 1988. Paleopalynology. Allen and Unwin, Inc., Boston. 600 p.
- Wieczorek, G.R., Morgan, B.A., and Campbell, R.H., 2000. Debris-flow hazards in the Blue Ridge of central Virginia. *Environmental and Engineering Geoscience*, 6(1):3-23.
- Williams, J.W., Post, D.M., Cwynar, L.C., Lotter, A.F., and Levesque, A.J., 2002. Rapid and widespread vegetation responses to past climate change in the North Atlantic region. *Geology*, 30(11):971-974.

FIGURE 1. Site map of study area, Shenandoah National Park and environs, VA. Outcrop localities noted by small white circles. Sample numbers correspond to Table 1. The boundary between the warmer oak-hickory pine forest and the cooler Appalachian oak forest approximates the 1000' elevation contour (lower right corner).

CS

Tony -

Hope this helps. I am
revisiting these samples
because we have developed
a more inclusive temperature
model.

Ron

Received via mail
09/04/22