Chapter 14

Forests, Land Management, and Agriculture

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Summary

The boreal region covers about 17% of global land area, and the arctic nations together contain about 31% of the global forest (non-boreal and boreal). The boreal forest is affected by and also contributes to climate change through its influence on the carbon cycle and albedo. Boreal forests influence global levels of atmospheric carbon dioxide and other greenhouse gases by taking up carbon dioxide in growth, storing carbon in live and dead plant matter, and releasing carbon through decomposition of dead organic matter, live plant and animal respiration, and combustion during fire. Human management influences on carbon uptake and storage include the rearrangement of forest age classes through timber harvest or wildfire suppression, selection of tree species, fertilization, and thinning regimes. The combined effect of all management actions can either enhance or reduce carbon uptake and storage.

Agriculture has existed in the Arctic as defined in this chapter for well over a millennium, and today consists of a mixture of commercial agriculture on several thousand farms and widespread subsistence agriculture. Potatoes and forage are characteristic crops of the cooler areas, and grains and oilseed crops are restricted to areas with the warmest growing seasons. The main livestock are dairy cattle and sheep, which have been declining, and diversified livestock such as bison or other native animals, which have generally been increasing in commercial operations. The five ACIA-designated models all project rising temperatures that are very likely to enable crop production to advance northward throughout the century, with some crops now suitable only for the warmer parts of the boreal region becoming suitable as far north as the Arctic Circle. The average annual yield of farms is likely to increase at the lower levels of warming due to climate suitability for higher-yielding crop varieties and lower probabilities of low temperatures limiting growth. However, in the warmest areas, increased heat units during the growing season are very likely to cause a slight decrease in yields since warmer temperatures can speed crop development and thereby reduce the amount of time organic matter accumulates. Under the ACIA-designated model projections, water deficits are very likely to increase or appear in most of the boreal region. By the end of the 21st century, unless irrigation is practiced, water stress is very likely to reduce crop yields. Water limitation is very likely to become more important than temperature limitations for many crops in much of the region. Overall, negative effects are unlikely to be stronger than positive effects. Lack of infrastructure is likely to remain a major limiting factor for commercial agricultural development in the boreal region in the near future. Even under model-projected levels of climate change, government policies regarding agriculture and trade will still have a very large, and perhaps decisive, influence on the occurrence and rate of agricultural development in the north.

Understanding the condition or character of the forest resource system that climate change affects is crucial in assessing forests and land management. Russia has made commitments to management of carbon stocks that are of global interest because of the amounts involved. Fire and insect disturbance at very large scales have generated resource management challenges in Canada. A large proportion of Alaska is managed as strict nature reserves and as resource lands for biodiversity and ecosystem services. Large forest disturbances associated with climate change have occurred in Alaska, disrupting ecosystems and imposing direct costs, but the large area of reserves improves the ultimate prospects of species surviving potential future climate change. In highly managed forests of Finland, Sweden, and Norway, forests are generally managed effectively and are increasing in volume, but the prospect of climate change puts at risk human expectations of specific future resource returns. In Iceland, temperature increases have improved tree growth at a time of a large afforestation program designed to increase forest land cover and sequester carbon.

About 6000 years BP (the end of the postglacial thermal maximum), radial growth of larch trees on the Taymir Peninsula of Russia surpassed the average of the last two millennia by 1.5 to 1.6 times. Tree growth and warm-season temperature have irregularly decreased in northernmost Eurasia and North America from the end of the postglacial thermal maximum through the end of the 20th century. Long-term tree-ring chronologies from Russia, Scandinavia, and North America record the widespread occurrence of a Medieval Warm Period about 1000 years BP, a colder Little Ice Age ending about 150 years ago, and more recent warming. Recent decades were the warmest in a millennium or more at some locations. Temperature and tree growth records generally change at the same time and in the same direction across much of the Arctic and subarctic. However, intensified air-mass circulation associated with a warmer climate has introduced a stronger flow of warm air into specific regions of the Arctic and enhanced the return flow of cold air out of the Arctic in other regions. Temperature and tree-growth trends are correlated but opposite in sign in these contrasting regions.

Between 9000 and 7000 years BP, trees occurred in at least small groups in what is now treeless tundra nearly to the arctic coastline throughout northern Russia. Around 6000 years BP, the northern treeline on the Taymir Peninsula (currently the farthest north in the world) was at least 150 km further north than at present. During the period of maximum forest advance, mean July temperature in northern Russia is estimated to have been 2.5 to 7.0 ºC higher than the modern mean. This record of past forest advance suggests that there is a solid basis for projecting similar treeline change under climate change producing similar temperature increases. It also suggests that the components of ecosystems present today have the capacity to respond and adjust to such climate fluctuations. The

greatest retreat of forest and expansion of tundra occurred between 4000 and 3000 years BP. In northeast Canada, the black spruce forest limit has remained stable for the past 2000 to 3000 years. In recent decades, milder winters have permitted stems that were restricted to snow height by cold and snow abrasion to emerge in upright form, and future climate projected by the ACIA-designated models would permit viable seed production, which is likely to result in infilling of the patchy forest–tundra border and possibly begin seed rain onto the tundra. In the Polar Ural Mountains, larch reproduction is associated with warm weather, and newly established trees have measurably expanded forest cover during the 20th century, although there is a time lag between climate warming and upslope treeline movement.

Across the boreal forest, warmer temperatures in the last several decades have either improved or decreased tree growth, depending on species, site type, and region. Some tree-growth declines are large in magnitude and have been detected at different points across a wide area, although the total extent has not been delineated. Temperature-induced drought stress has been identified as the cause of reduced growth in some areas, but other declines are not currently explained. Reduced growth in years with high temperatures is common in treeline white spruce in western North America, suggesting reduced potential for treeline movement under a warming climate. Tree growth is increasing in some locations, generally where moisture and nutrients are not limiting, such as in the boreal regions of Europe and eastern North America. The five ACIA-designated models project climates that empirical relationships suggest are very unlikely to allow the growth of commercially valuable white spruce types and widespread black spruce types in major parts of Alaska and probably western boreal Canada. The models project climates that are very likely to increase forest growth significantly on the Taymir Peninsula. The upper range of the model projections represents climates that may cross ecological thresholds, and it is possible that novel ecosystems could result, as during major periods of global climate change in the past.

Large-scale forest fires and outbreaks of tree-killing insects are characteristic of the boreal forest, are triggered by warm weather, and promote many important ecological processes. On a global basis, atmospheric carbon equal to 15 to 30% of annual emissions from fossil fuels and industrial activities is taken up annually and stored in the terrestrial carbon sink. Between 1981 and 1999, it is estimated that the three major factors affecting the terrestrial carbon sink were biomass carbon gains in the Eurasian boreal region and North American temperate forests, and losses in areas of the Canadian boreal forest. Particular characteristics of forest disturbance by fire and insects, such as rate, timing, and pattern of disturbance, are crucial factors in determining the net uptake or release of carbon by forests. The evidence necessary to establish a specific climate change effect on disturbance includes a greater frequency of fire or insect outbreaks, more extensive areas of tree mortality, and more intense disturbance resulting in higher average levels of tree death or severity of burning. Some elements of the record of recent boreal forest disturbance are consistent with this profile of climate change influence, especially forest fires in some parts of Russia, Canada, and Alaska and insect disturbances in North America.

Carbon uptake and release at the stand level in boreal forests is strongly influenced by the interaction of nitrogen, water, and temperature influences on forest litter quality and decomposition. Warmer forest-soil temperatures that occur following the death of a forest canopy due to disturbance increase the rate of organic litter breakdown, and thus the release of elements for new plant growth (carbon uptake). The most likely mechanism for significant short-term change in boreal carbon cycling as a result of climate change is the control of species composition caused by disturbance regimes. Successional outcomes from disturbance have different effects on carbon cycling especially because of the higher level and availability of nutrient elements (and thus decomposition) in organic litter from broadleaf trees compared to conifers. Net global land-use and land-cover change, especially aggregate increases or decreases in the area of forest land, may be the most important factor influencing the terrestrial sink of carbon. When water and nitrogen remain available at the higher growth rates typical of enhanced carbon dioxide environments, further carbon uptake occurs. Broadleaf litter produced under elevated carbon dioxide conditions is lower in quality (less easily decomposed) than regular litter because of lower nitrogen concentration, but quality of conifer litter in elevated carbon dioxide environments may not be as affected.

Different crop species and even varieties of the same species can exhibit substantial variability in sensitivity to ultraviolet-B (UV-B) radiation. In susceptible plants, UV-B radiation causes gross disruption of photosynthesis, and may inhibit plant cell division. Determining the magnitude of the effect of elevated UV-B radiation levels is difficult, because interactions with other environmental factors, such as temperature and water supply, affect crop reactions and overall growth. Damage by UV-B radiation is likely to accumulate over the years in trees. Evergreens receive a uniquely high UV radiation dose in the late winter, early spring, and at the beginning of the growing season because they retain vulnerable leaf structures during this period of maximum seasonal UV-B radiation exposure, which is amplified by reflectance from snow cover. Exposure to enhanced levels of UV-B radiation induces changes in the anatomy of needles on mature Scots pine similar to characteristics that enhance drought resistance. UV-B radiation plays an important role in the formation of secondary chemicals in birch trees at higher latitudes. Secondary plant chemicals released by birch exposed to increased UV-B radiation levels might stimulate its herbivore resistance.

14.1. Introduction

The Arctic has been defined in somewhat different ways in various studies, reports, and assessments, based primarily on the purpose of the project. While the most restrictive definitions limit the Arctic to treeless tundra, snow, and ice in the high latitudes, most definitions of the Arctic encompass some elements of the boreal forest. The definition used by the Arctic Monitoring and Assessment Programme (AMAP, 1997; see section 8.1.1, Fig. 8.2), for example, includes the productive boreal forests of northwest Canada and Alaska, but includes mostly marginal treeline forest and woodland in eastern Canada. Permafrost-free forests in the northern portion of the Nordic countries are within the AMAP-defined Arctic, but across central and eastern Siberia, the boundary follows the margin of sparse northern taiga and forest–tundra. This chapter focuses on the northernmost portion of the boreal forest region, but broadens consideration of the subject for two important reasons. First, many elements of the boreal forest are best understood as a whole (e.g., the gradients of changing tree responses to the environment from south to north), and this chapter includes an extensive and well studied Siberian transect that uses such an approach. Second, the five scenarios of climate change generated by the ACIA-designated models (section 4.4) project temperatures within the Arctic that today only occur in the boreal forest far to the south. If temperature increases of a magnitude similar to those projected by these models actually occur, the nearest analogues of climate (and eventually ecosystems) that would exist in the Arctic are those of more southerly boreal forest regions.

Sections 14.2 and 14.3 describe forest characteristics across the northern boreal forest to provide the context for understanding the importance of recent climate-related changes in the region and potential future change. Section 14.4 provides an overview of the climate scenarios generated by the ACIA-designated models and describes how the scenarios were used in different aspects of the assessment.

While many factors affect agriculture in the far north (e.g., changing markets, social trends, and national and international policies), section 14.5 focuses on the climate-sensitive aspects of crop production systems that would be affected under the scenarios of future climate, focusing on climate stations representative of areas with agricultural production or potential. Section 14.3 also considers the challenges that climate change poses for land management. Tree rings are one of the most important sources of information about past climates, especially in the sparsely populated far north, and section 14.6 reviews the record of climate and tree rings across the Arctic and northern boreal region.

Section 14.7 presents new information about the direct effects of climate on tree growth in the northern boreal forest, in both the distant and more recent past, and uses scenarios generated by the ACIA-designated models to project how climate change may affect the growth of selected tree species during the 21st century. Section 14.8 identifies key climate controls on large-scale population increases in insects that damage trees, and provides some recent evidence of these effects.

Forest fire is another major indirect effect of climate on the status of forests in the far north. Section 14.9 examines some of the climate-sensitive aspects of fire and possible future fire conditions and effects. The climate-related changes in growth, insect-caused tree death or reductions in tree growth, and fire are major factors that control the uptake and storage of atmospheric carbon (section 14.10). The implications of future climate change for forest distribution are briefly considered in section 14.11. Finally, section 14.12 summarizes some recently published information on the effects of increased UV-B radiation levels on boreal forest species, and section 14.13 reviews critical research needs.

14.2. The boreal forest: importance and relationship to climate

14.2.1. Global importance

The boreal region covers about 17% of the terrestrial area of the earth (Bonan et al., 1992), with a broad zone of forest in a continuous distribution across the Eurasian and North American landmasses. The boreal forest is defined as a belt of forest south of the tundra characterized by a small number of coniferous species including spruce, larch, pine, and fir and a limited number of broad-leaved species, primarily birch and poplar (see Appendix D for common and scientific names of tree and other woody species mentioned in this chapter). At the landscape scale, conifers dominate the boreal forest, although broad-leaved trees can be locally dominant. Forest and woodland in the arctic nations (excluding

Table 14.1. Total forest area in arctic nations and percentage of global forest area (Smith W. et al., 2001; US Forest Service, www.fs.fed.us/r10/spf/facts/spffact.htm).

	Total forest area (10^6 ha)	% of global forest area
Russia	851.4	22.4[a]
Canada	44.6	6.4[a]
United States (all)	302.4	8.0[a]
United States (Alaska boreal only)	35	0.9[b]
Finland	21.9	0.6[a]
Sweden	27.1	0.7[a]
Norway	8.8	0.2[a]
Iceland	0.034	<0.001[a]
Arctic nation total	1456.2	38.3
World total	3800	100

[a] “forest” category (FAO, 2002); [b] “boreal forest” category (Labau and van Hees, 1990)

Denmark), most of which is boreal forest, represent about 1.5 billion hectares (ha) of the total global forest area of 3.8 billion ha (in 2000), or about 38% of global forest area (FAO, 2001; Table 14.1). Russian forests (excluding woodlands), the vast majority of which are boreal, represent 22.4% of the global total, by far the largest proportion of any nation in the world (Table 14.1). Two of the remaining three countries with the largest percentage of global forest area (Canada and the United States) are also arctic nations.

This chapter focuses on the northern portion of the boreal forest, but many aspects of the topic must be considered from a broader perspective. The boreal forest contains trees growing at the highest latitude on earth, and along its northern margin it merges into the circumpolar tundra. The boreal region is the northernmost part of the world where agricultural crops are produced regularly on a significant scale and where a settled agricultural way of life has historical continuity.

The boreal forests of North America and Eurasia share some plant and animal species and display a number of other similarities. The boreal forests of Russia and especially Siberia are often referred to as the taiga, an indigenous term meaning "little sticks". The term taiga is equivalent to the true boreal forest.

The boreal forest is both affected by and contributes to climate change: both topics are examined in this chapter. Globally, the existence of large areas of boreal forest cover has a significant effect on the radiative balance of the planet (Bonan et al., 1992). The rough-textured, dark surface of land covered with boreal forest canopy intercepts and absorbs a high proportion of solar radiation, converting it to heat (Bonan et al., 1992). In contrast, the smooth, snow-covered surface of the tundra is highly reflective. In high-latitude regions where snow covers the ground for half of the year or more, the albedo effect of tundra versus boreal forest cover is magnified. Future expansion of the forest into present-day tundra regions resulting from a warming climate would thus amplify the warming further.

Another important influence of the global boreal forest on climate is its influence on levels of atmospheric carbon dioxide (CO_2) and other greenhouse gases (GHGs). Boreal forests take up CO_2 through photosynthesis, and store carbon in live and dead plant matter, including substantial long-term accumulations in large tree boles and in soil. Forests release CO_2 to the atmosphere through decomposition of dead organic matter, live plant and animal respiration, and combustion that takes place during fires. Both natural (e.g., fire) and anthropogenic (e.g., timber removal) disturbances are important influences on the boreal forest. An ecosystem disturbance is defined as a change in state or condition that disrupts the way in which the system has been functioning (photosynthesis, water regulation, etc.), causing it to reinitiate successional development. Disturbances vary by cause, rate, intensity, extent, timing, frequency, and duration.

Management-related factors influence carbon uptake and storage in the form of tree mass. These management practices include the rearrangement of forest age classes by timber harvest or suppression of wildfires, selection of tree species, fertilization, and thinning regimes. The combined effect of all management actions can either enhance or reduce carbon uptake and storage. For example, across the Russian boreal region, for many years after logging the forests that regrow take up less atmospheric CO_2 than nearby old-growth forests (Schulze et al., 1999). In the Boreal Cordilleran ecozone of Canada (see Wiken, 1986 for definition), it is estimated that total suppression of natural disturbances and their complete replacement by harvesting for maximum sustainable yield would increase carbon storage in soils and wood products over a period of a century or two (Price et al., 1997). Direct climate effects that increase or decrease tree growth in unmanaged natural forests also influence short-term uptake of atmospheric carbon.

The boreal forest and northern tundra together contain 40% of global reactive (readily decomposable to CO_2, methane, water, and mineral nutrients) soil carbon, an amount similar to the amount of carbon held in the atmosphere (McGuire et al., 1995b; Melillo et al., 1993). The extensive boreal forest plains of northeast Europe, western Siberia, and central and eastern North America that are within or immediately south of the discontinuous permafrost region occupy the zone of maximum carbon storage in soil organic matter on the earth. Climate change, interacting with human use and management of boreal forest, northern agricultural, and tundra ecosystems, would enhance the decomposition of carbon stored in soil organic matter and its subsequent release into the atmosphere, thus compounding climate change caused by anthropogenic GHG emissions.

The boreal forest is one of the most intact major vegetation regions of the earth, but in some areas boreal forest has been extensively converted to other land uses or severely damaged by air pollution (e.g., in Iceland and particular areas of Russia, respectively). Boreal forests in Finland, Sweden, Norway, and parts of Canada are generally intensively managed for timber production, and in such intensively managed stands, tree age structure, tree species, and spacing are controlled (section 14.3.4). However, huge areas of central and eastern Siberia and northwestern North America represent the most extensive remaining areas of natural forest on the planet (Bryant et al., 1997). Not all natural boreal forests consist of older trees: large areas are burned or subject to insect-caused tree mortality annually. Climatic factors, especially prolonged periods of warm weather, often create the conditions that result in fire and insect disturbances in boreal forests. The boreal forest is subject to rapid changes causing long-term consequences as a result of these climate-related effects.

The boreal forest is the breeding zone for a huge influx of migratory forest birds that perform many important roles (e.g., insect consumption, seed dispersal) in the boreal region and in other forests of the world during migration and winter residence in the south. Climate affects the population level of these migratory birds and their food resources. Climate-associated processes also determine the amount and quality of forest habitat available to migratory birds.

From 1981 to 1999, three regions of the world primarily affected the terrestrial carbon sink, which takes up and stores atmospheric carbon equivalent to 15 to 30% of annual global emissions from fossil fuel combustion and industrial activities (Myneni et al., 2001). The majority of the net terrestrial carbon change came from biomass carbon gains in the Eurasian boreal region and North American temperate forests, and carbon losses from some Canadian boreal forests. Some of the terrestrial biomass change was a response to direct and indirect climate effects. However, human use and management of the boreal forest was an important factor as well, and could be a significant future contributor to human management of the carbon cycle. Certain forest biomass carbon sinks can be used to meet national commitments to reduce GHG emissions under the Kyoto Protocol of the Framework Convention on Climate Change. Land and resource managers in the Arctic and boreal regions are interested in potential "carbon cropping", which might involve payments from organizations wishing to sustain or enhance carbon storage (Bader, 2004). Mechanisms to place values on the various carbon transfers are not fully in place. However, if effective, exchangeable systems of placing values on transfers of carbon are adopted at an international level, boreal forests could potentially generate a flow of wealth into arctic and subarctic regions from other parts of the world for boreal forest and land management treatments, offsets for emissions elsewhere, or policies designed to store or retain carbon.

Uncertainties remain about the influence of the boreal forest on each of the key processes that determine global carbon balance. For example, the uptake of atmospheric CO_2 by tree and other plant growth may either increase or decrease with increasing temperature, depending on the species, the geographic region where the growth occurs, the range of the temperature increase, and other climate factors such as precipitation that are likely to change in a changing climate. However, there has been substantial recent progress in understanding the response of elements of the boreal system to temperature. Across the boreal regions, a first generation of studies, models, databases, and measurements have provided a significantly better understanding of one of the most extensive and important vegetation types on earth. Continued and expanded data collection from research and management activities will provide a reasonable basis for determining the net contribution of the boreal forest to GHG balance and climate, the further changes a warming climate would induce in the boreal forest, and the agricultural and forest management opportunities available to the region in the future.

14.2.2. Arctic importance

Trees occur on only a small proportion of the land surface within the Arctic as defined in this chapter. Even so, forests and woodlands are important on a regional basis within the Arctic for several reasons. Where trees do occur, they serve as indicators of more productive terrestrial ecosystems with longer growing seasons than treeless tundra. Trees, even when present in small numbers on the arctic landscape, offer resources to arctic residents for a variety of uses. Finally, some areas of full-canopy forest within the Arctic are generally the most productive natural systems within the political jurisdiction where they occur (e.g., the galley forests along rivers that extend into the tundra in the northern Yukon and Northwest Territories in Canada). Specific reasons for the importance of boreal forest and agriculture in the Arctic include the following:

- Portions of the boreal forest devoted to forest products production are major contributors to the national economies of some arctic nations. Although the current zone of optimum climate for boreal forest growth is in the middle or southern boreal region, nearly all scenarios of climate change place the climatically optimum growth region within the present-day Arctic within a century or so.
- Residents of the boreal region depend on the products and resources of the forest for a variety of ways of life, including traditional ways of life.
- The major rivers of the boreal region transport large volumes of wood into the Arctic (Eggertsson, 1994; Ott et al., 2001), and this wood resource supports ecosystems that decompose the wood and feed organisms in rivers, oceans, and beaches. Climate change is very likely to affect all the processes in this system, including tree growth, erosion, river transport, and wood decay.
- Wood transported into the Arctic was an important resource for people in a naturally treeless environment during prehistoric times (Alix, 2001) and is still a useful and valued resource for many arctic residents today.
- The boreal forest collects, modifies, and distributes much of the freshwater that enters the Arctic Basin (see sections 6.8 and 8.2), and changes in boreal forests resulting from climate change would certainly affect many of these important functions.
- Portions of the boreal forest region have experienced some of the greatest temperature increases reported during the 20th century, and the responses of the forest system and the societal consequences in the region provide lessons that may be useful to other regions that could eventually experience similar change.
- Recent temperature increases in the boreal region have increased the frequency of occurrence of critical temperature thresholds for the production of agricultural crops currently grown in the region. Possible future temperature increases almost cer-

tainly would increase the land area on which crops could be produced successfully, and are very likely to increase the variety of agricultural crops that could be grown.

14.2.3. Climatic features

The boreal region is often assumed to be a zone of homogenous climate, but climate across the region is actually surprisingly diverse. During the long summer days, interior continental locations under persistent high-pressure systems experience hot weather that facilitates extensive forest fires frequently exceeding 100 000 ha. In maritime portions of the boreal region affected by air masses that originate over the North Atlantic, North Pacific, or Arctic Oceans, summer daily maximum temperatures are on average cooler than interior locations and seldom reach the high temperatures experienced at locations further inland.

Precipitation is abundant in the boreal zone of most of the Nordic countries, western Russia, and certain coastal and mountain regions of western North America. By contrast, in the topographically complex

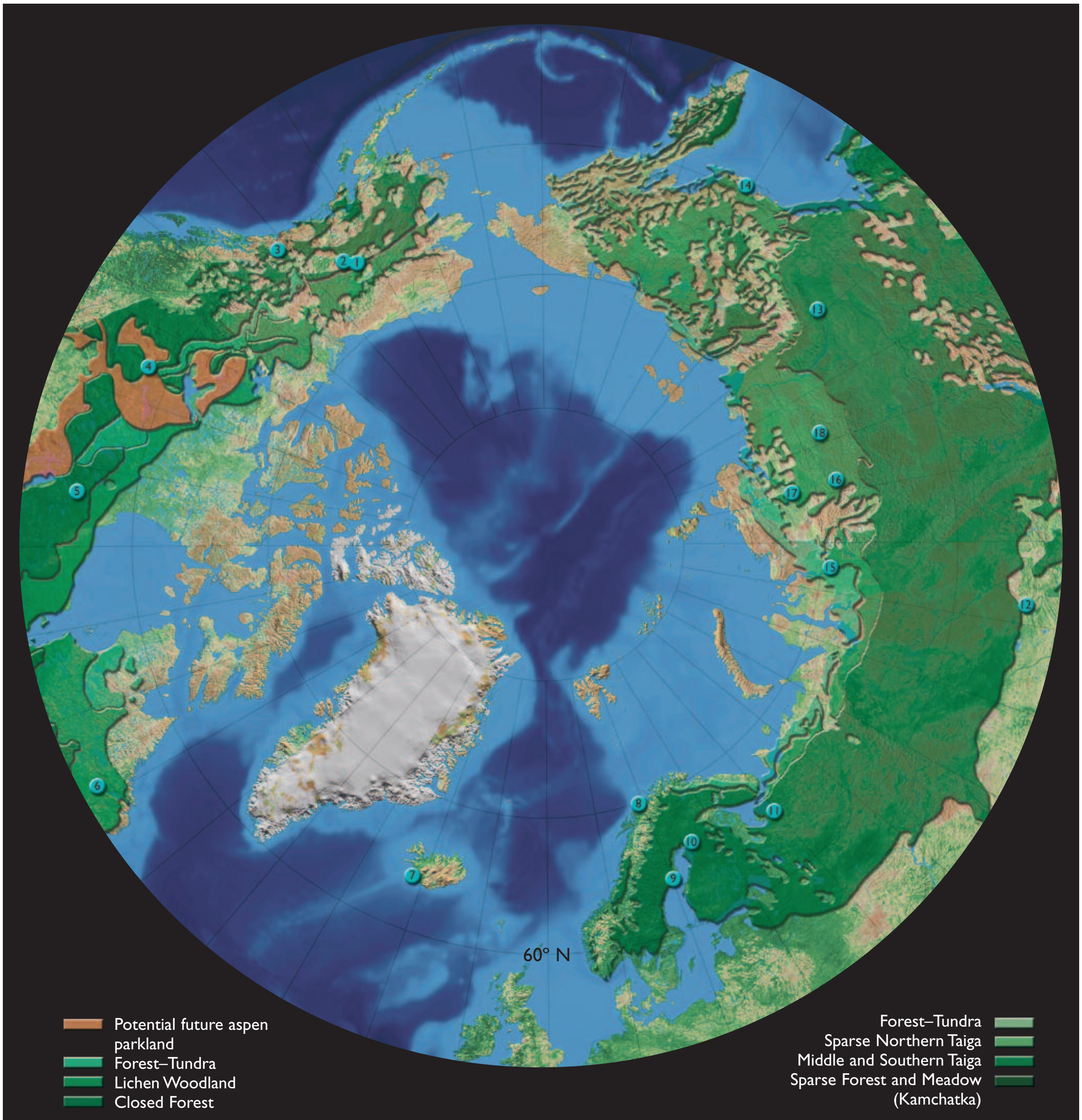

Fig. 14.1. Present-day boreal forest distribution, using a simplified formation system representing a gradient of decreasing productivity and species diversity from south to north (compiled using data from Anon., 1983; Anuchin and Pisarenko, 1989; Elliott-Fisk, 1988; Kuchler, 1970; Kurnayev, no date; Oswald and Senyk, 1977; Rowe, 1972; Viereck and Little, 1972). The area depicted in orange represents a broad zone where the ratio of precipitation to evapotranspiration is nearly one: under a scenario of doubled atmospheric CO_2, this area is projected to become too dry to support closed-canopy boreal forest, shifting instead to aspen "parkland", a woodland formation (Hogg and Hurdle, 1995). Numbered locations are sites analyzed in this chapter (see section 14.4.1, Table 14.2).

landscapes of Alaska, northwestern Canada, and central and northeastern Siberia, precipitation sometimes limits forest growth so that natural grasslands are part of the landscape.

Precipitation in the boreal region of western North America is influenced by storms in the southern Bering Sea and North Pacific Ocean and reaches a distinct maximum in late summer. In other parts of the boreal forest region, precipitation is more evenly distributed throughout the year or exhibits a winter maximum. East-central Siberia experiences low winter snow depths because the strong Siberian High suppresses precipitation. The boreal landscapes of far eastern Siberia and western North America are mountainous, whereas the topography of most of central and western Siberia and eastern Canada is characterized by low, smooth hills and level terrain. The mountainous boreal regions are characterized by sharply varying local climates (Pojar, 1996) and aspect-controlled differences in forest types (Viereck et al., 1986). As a result, the forest–tundra boundary is much more irregular there than on the plains of the central portions of the continents (Fig. 14.1). All of this regional climatic variation must be taken into account as a fundamental backdrop when considering climate change and ecological response in boreal forests.

In Eurasia and North America, both the northern and the southern boundaries of the boreal zone are not aligned at the same latitude east to west. The Icelandic and Aleutian Lows deflect storm tracks and advect relatively mild air masses northward as they approach the western margins of Eurasia and North America, respectively. As a result the boreal forest belt is located considerably farther north in both the Nordic countries and western North America than in the center of the continents (Fig. 14.1). In contrast, cold polar air flowing southward follows a persistent path along the eastern portion of both continents, and consequently the boreal forest belt reaches its southernmost limits there. In the center of the Eurasian landmass, the latitudinal position of the boreal forest depends mainly on the degree of continentality of the climate. The most northerly forests in the world occur along the Lower Khatanga River in central Siberia (72º 32' N), where the climate is more continental than either western or eastern Siberia. The warmer summer temperature in northernmost central Siberia is the critical factor that allows trees to survive there despite extreme winter cold.

14.2.4. Climate variability

Essentially all of the boreal forest in Alaska is north of 60º N, and practically all the boreal forest of eastern Canada is south of 60º N (Fig. 14.1). The boreal forest region is particularly prone to climatic variability because minor variations in key features of the atmospheric circulation can either intensify the advection of warm air into this naturally cold region, or enhance the distribution of cold air southward through the region. There is some evidence that the climate system in the far north operates in a way that positive (western continent) and negative (eastern continent) anomalies operate in synchrony with each other. It appears that the intensification of meridional air mass movement is especially effective in warming the western margin of the North American Arctic while cooling the eastern margin of the continent (Fig. 14.2), leading to east–west temperature anomalies in the arctic and boreal regions (see also section 2.6.2.1, e.g., Fig. 2.7, and section 6.7.2).

Periods of major climate change, including alternating glacial and interglacial conditions, have repeatedly and drastically affected the northern regions of the planet (section 2.7). During the late Pleistocene and several previous glaciations, the present-day boreal region was mostly covered with glacial ice and forest organisms were largely displaced south of the current limits of the region (Wright, 1983; Wright and Barnosky, 1984; section 2.7). The relatively small unglaciated portion of the present-day boreal region was almost entirely treeless and contained assemblages of species unlike any found today (Anderson and Brubaker, 1994).

Present-day boreal forest vegetation characteristically has a large ecological amplitude (i.e., the ability to survive across a wide range of environmental conditions). The paleoenvironmental record and modern instrumental measurements demonstrate major shifts in temperature regimes in this region even during the time that forest has been present (sections 2.7 and 14.6), and the presence of this large ecological amplitude in the trees indicates that this wide range of possible temperatures is a consistent enough feature of the environment that it has required an adaptive response. Rapid and large changes in weather over the short term send signals through boreal systems that initiate vital processes such as infrequent periodic tree reproduction (Juday et al., 2003). On the one hand, this high level of natural climate variability suggests that during periods of climate

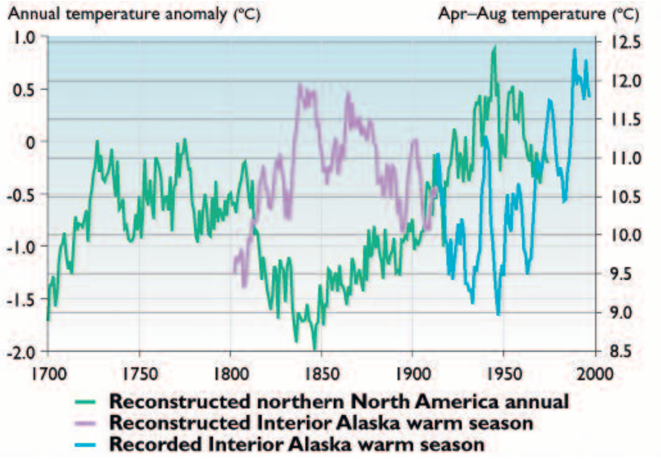

Fig. 14.2. Opposing high-latitude temperature trends from tree-ring reconstructions of mean annual temperature with a strong signal from eastern and central North America (data adapted from D'Arrigo and Jacoby, 1992, 1993; Jacoby and D'Arrigo, 1989) versus warm-season (Apr–Aug) temperature in western North America (observed and reconstructed from tree rings; Barber et al., 2004). Anomalies are calculated from the 1671–1973 mean.

change, the effects are more likely to be detectable at an earlier time in the boreal region than in many other parts of the earth. On the other hand, a long and persistent history of climate variability in the region suggests that organisms in the boreal forests of today may be among those better adapted to climate change because they have been filtered by many climate fluctuations in the past.

Because ecosystems quite unlike those of today have existed in the region in the past (Anderson and Brubaker, 1994; Wright, 1983; Wright and Barnosky, 1984), major climate change in the future is likely to produce ecosystems unknown today. The emergence of novel ecosystems (from the human perspective) is partly the result of individualistic species responses to changes in the environment. Each species has its own environmental requirements, tolerances, and thresholds, so that some species that co-occur today may not in the future, or existing sets of species may be joined by additional species (sections 7.3 and 7.6). Because of this property of individualistic responses to change in the environment, conservation efforts must be informed by monitoring the status of key species on a continuing basis. Conservation measures, such as modified harvest limits or fire management, must account for rapidly changing environmental conditions or changes in species populations not anticipated in management planning assumptions or outside historical experience in order to meet a goal of sustainability (Chapter 11).

The simultaneous pattern of temperature and precipitation anomalies can have important ecological impacts in forests of the north. Wildfire in the boreal forest is the product of short- to medium-term warm and dry conditions, usually associated with high-pressure dominance during the long days of summer (Johnson, 1992). Alternating periods of warm and dry versus cool and moist summer climate in central Alaska regulate the growth and reproduction of white spruce (*Picea glauca*), ultimately providing a mechanism to synchronize the production of seed crops to periods immediately following major forest fires (Barber et al., 2000; Juday et al., 2003). If future climate change alters not just the mean of climate parameters, but also the pattern of alternating warm/dry and cool/moist conditions, the resulting climate pattern could interfere with the reproductive success of one of the most widely distributed and dominant North American conifers. This example indicates the potential for subtle influences to be major factors in climate change effects.

14.2.5. Unique influences on climate

The Boreal Ecosystem–Atmosphere Study (BOREAS) was a large-scale, international interdisciplinary experiment in the northern boreal forests of Canada that began in 1993. Its goal was to understand how boreal forests interact with the atmosphere, how much CO_2 they were capable of storing, and how climate change will affect them.

Albedo measurements from BOREAS are among the lowest ever measured over vegetated regions, and indicate that the boreal forest (especially forest dominated by black spruce – *Picea mariana*) absorbs nearly 91% of incident solar radiation (Hall et al., 1996). In terms of water and energy balance, BOREAS found that the boreal ecosystem often behaves like an arid landscape, particularly early in the growing season. Even though the moss layer is moist for most of the summer, nutrient-poor soils and limiting climatic conditions result in low photosynthetic rates, leading to low evapotranspiration. As a result, relatively little of the available moisture is transferred to the atmosphere. Much of the precipitation penetrates through the moss layer into the soils, which are permeable, then encounters the underlying semi-impermeable layer and runs off. Most of the incoming solar radiation is intercepted by the vegetation canopy, which exerts strong control over transpiration water losses, rather than by the moist underlying moss/soil surface. As a result, much of the available surface energy is dissipated as sensible heat.

The BOREAS experiment also found that coniferous vegetation in particular follows a very conservative water-use strategy. Stomatal closure drastically reduces transpiration when the foliage is exposed to dry air, even if soil moisture is freely available. This feedback mechanism acts to keep the surface evapotranspiration rate at a steady and surprisingly low level (less than 2 mm/d over the season). The low evapotranspiration rates coupled with high available energy during the growing season can lead to high sensible heat fluxes and the development of deep (3000 m) planetary boundary layers, particularly during the spring and early summer. These planetary boundary layers are often characterized by intense mechanical and sensible heat-driven turbulence.

14.3. Land tenure and management in the boreal region

The influence of climate change on forest values and forest users depends on the amount and initial condition of the forest resource and the uses or intangible values of the forest for people, cultures, and economies. This section reviews forest extent, the overall allocation of forest land to different uses, the main patterns of forest use, the management systems, and the values generated by the boreal forest. Where these characteristics can be singled out by political jurisdiction or other means, the discussion is focused on the northern boreal forest. This discussion forms the basis for considering climate change impacts.

14.3.1. Russia

Russia contains the largest forested area of any nation, amounting to an estimated 763.5 million ha (FAO, 1999). The boreal forest of Russia can be thought of as three roughly parallel belts of southern and middle taiga, sparse northern taiga, and in the farthest north a forest–tundra region extending to the completely treeless tun-

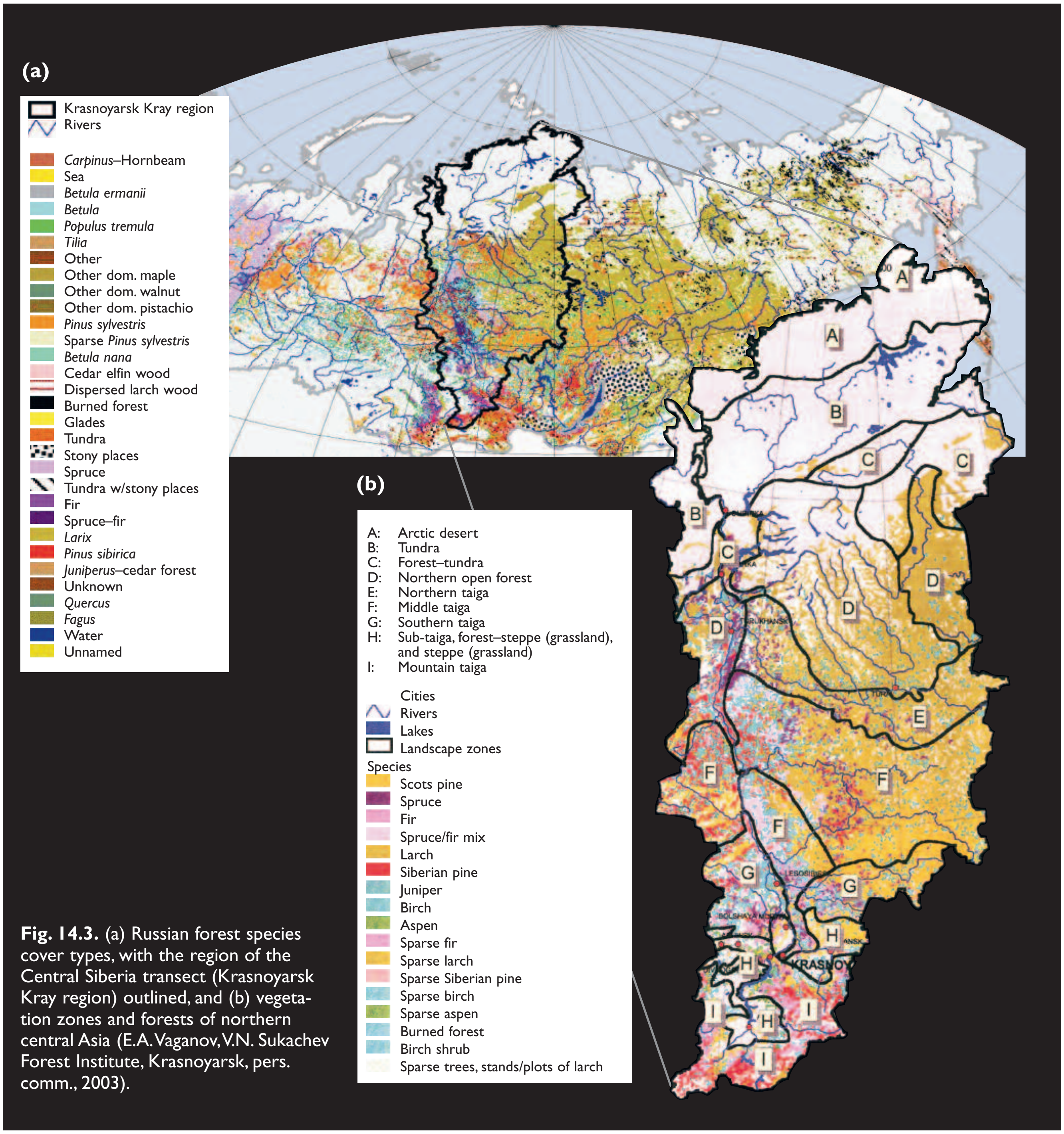

Fig. 14.3. (a) Russian forest species cover types, with the region of the Central Siberia transect (Krasnoyarsk Kray region) outlined, and (b) vegetation zones and forests of northern central Asia (E.A. Vaganov, V.N. Sukachev Forest Institute, Krasnoyarsk, pers. comm., 2003).

dra (Fig. 14.1). Dominant components of the Russian boreal forest change from spruce in the west (European Russia), to larch in the center and east, to pine at various locations (Fig. 14.3a). In Northern Eurasia, especially in central Siberia (Fig. 14.3b) and the Far East, as much as 70 to 75% of taiga forests appear to be close to a natural state. The remaining forests consist of fragments and other human-influenced forest, and areas dominated by marsh–bog complexes.

The central and especially the southern taiga zones have a long history of quite intensive land use. Although large untouched areas are rare or absent even in northern European Russia, a much greater proportion of these forests exists in a natural state than in similar vegetation zones in the Nordic countries. The Karelian Isthmus region in the St. Petersburg Oblast is a good example. The forests of the Green Belt along the Finnish–Russian (Karelian) border are one of the most important centers of boreal biodiversity in Europe (Silfverberg and Alhojärvi, 2004). Many species in the Red Data Book (endangered) lists for Sweden and Finland still exist in relatively healthy populations in northwest Russia. Disruption or decline in the ecological health of these remnant natural forests in Russia resulting from climate change is very likely to have severe implications for the survival of the listed species, at least within Europe.

Economically exploitable forests total about half of the forested areas under state forest management in Russia, and are generally located in the southern taiga. The northern boreal forest of Russia is generally too distant from transportation infrastructure and consumption centers to repay the costs of timber removal. The Russian forest sector is a major employer, with a work force estimated at 1.8 million people in the mid- and late 1990s. However, during the 1990s, public-sector forest management organizations often did not have enough money to retain employees, fight forest fires, enforce logging regulations, or make periodic inventories. A slow recovery of public-sector capacity has begun in more recent years.

A 1993 reform, the Basic Forest Law, started a movement toward market transactions in this sector. The Basic Forest Law allows forest leasing and auctions of standing timber, and forest leasing is the main market mechanism currently in use. In the early 1990s, rapid and unsustainable cutting of old-growth and mature forests began. After 1996, changes in forest management policy and better information resulted in a slowing of the pace of timber cutting in old-growth forests. The new Forest Code of the Russian Federation was issued in 1997. The Russian forest management system uses the term "Forest Fund" to refer to all forest and related lands under governmental jurisdiction, which in practice is nearly all the forest. The dominant part of the Forest Fund was (and still is) under the management of the Russian Federal Forest Service (Rosleskhoz), which manages about 94% of the total forest area in Russia, with another 4% belonging to agricultural organizations, 1% to the Committee of Environmental Protection, and 1% to other state bodies.

For management purposes, Russian forests have been divided into three categories based on economic and ecological characteristics. The first category comprises forests with a protective function, for example, watershed forests (20% of forested land), which are available for partial tree removal, sanitary tree felling, and small (maximum 10 ha) clear cuts. The second category consists of forests in inhabited areas and forests with low productivity (5.5% of forest area). The vast majority of the forest (74.5%) is included in the third category, industrially exploitable forests, where clear-cutting (up to 50 ha) is the main management practice. In addition to managing nearly all Russian forests, Rosleskhoz is also responsible for about 20% of the logging, in the form of partial and sanitary logging. Climate change impacts that disturb forest cover in category one and two forest are very likely to impose extra costs on managers, local governments, and forest users to stabilize or recover forests. In category three forest, actual or potential forest product values are at risk from climate change. However, because of the economic inaccessibility of the majority of this forest in the northern taiga, direct losses are likely to be relatively small overall.

As of 2000, the Russian Federation had 99 state zapovedniks, or strict scientific nature reserves, totaling 31 million ha or 1.82% of Russia's territory (Ostergren and Shvarts, 2000). Zapovedniks generally meet category I criteria of the World Conservation Union classification of protected areas (IUCN, 2000). During the 1990s, Russia established 35 national parks totaling 6.8 million ha (0.40% of Russia's territory). There are plans to establish additional zapovedniks and parks (Colwell et al., 1997). Practically all the national parks are located in Forest Fund areas and are managed by the state forestry authorities. These areas are managed for a range of scientific and biodiversity values, and the main concerns are climate changes that could reduce the chances for survival of the protected species or ecosystems.

The forests of Russia are an important component of the global carbon cycle because of the extensive area of forest land and the high storage of carbon in cold soils. In consideration of its extensive and significant forest resources, Russia has obtained substantial carbon emission credits as part of its participation in the Kyoto Protocol (Webster, 2002). Fulfilling the potential of Russian forests to offset carbon emissions will require sustaining, and to some degree rebuilding, a land management capability over a vast area with certain fundamental aspects: forest inventory and measurement, surveillance and detection of forest health problems, trained and deployable fire control and management forces, and various resource management specialists. Many of the benefits of increasing carbon sequestration in Russian forests can be obtained as a direct result of implementing policies that are widely agreed to be rational and beneficial (Shvidenko et al., 1997), including measures such as harvest levels in line with actual growth, effective fire control forces in regions of high-value timber, and adequate regeneration efforts.

14.3.2. Canada

The Canadian boreal forest represents nearly 6.4% of global forest area according to the United Nations Food and Agriculture Organization (FAO) definition of forest (Table 14.1). Forests play a large role in the Canadian environment, economy, culture, and history. Forest (tree-covered land with a full canopy) and woodland (tree-covered land with less than a complete forest canopy) cover nearly half (44%) of the Canadian landscape, totaling about 401.5 million ha, according to Canada's Forest Inventory 2001 (Natural Resources Canada, 2004). Canadian forest land totals about 309.8 million ha, according to the 2001 inventory, and about 294.7 million ha are not reserved and therefore potentially available for commercial forest activities, although much of the land has not been definitively allocated as to use. The large majority of Canadian forest is crown land held for the public, with 71% controlled by the provinces and 23% under federal control (Natural Resources Canada, 2003). A few percent of Canadian forest land is managed by territorial governments and the balance is in private hands. Of the 401.5 million ha of tree-covered land in Canada, 22.8 million (5.7%) are by law managed to remain in a natural state. On another 27.5 million ha (6.8%), timber harvesting is excluded by administrative policy (e.g., on unstable soils or as habitat buffers along important lakes or rivers). The most accessible forest land, and therefore the most likely to experience forest management activities, covers 144.6 million ha, or 36.0% of the total tree-covered area.

In the Canadian land tenure system, provincial governments are responsible for managing most of the land within their boundaries held for the benefit of the public (crown land). Until recently, the federal government held and managed land north of 60º N (an area not organized into provinces). However, in this area, by progressive steps, ownership and decision-making responsibility are passing to indigenous peoples and territorial governments. Simultaneously, lands of major conservation interest are being established as new national parks and wildlife refuges managed by the federal government. The indigenous peoples of Canada, who meet their cultural, spiritual, and material needs from their forest homeland, have a unique perspective and set of goals in forest management. Canadian land and forest management has changed significantly and is likely to change further as aboriginal title, treaty rights, and governmental responsibility to protect these rights are all more specifically defined.

The large majority of Canadian forest is boreal, with species such as white spruce, black spruce, aspen (*Populus tremuloides*), and paper birch (*Betula papyrifera*) having essentially transcontinental distributions. The forests of Canada are naturally dynamic, with large-scale disturbances quite typical. Across all Canadian forest types, insect defoliation affected 18.6 million ha in 2002 (Natural Resources Canada, 2003). Section 14.10.2 details the role of disturbance in the boreal forest with respect to carbon, and section 14.8.1 describes the role of forest insect disturbances in the boreal forest.

In the late 1990s, more than 300 communities in Canada depended largely on jobs in the forestry sector. During that period, the wood and paper industries and associated organizations employed more than 830 000 people, and paid more than Can$ 11.8 billion in wages annually. In 2002, 361 400 people were directly employed in the forest industry (Natural Resources Canada, 2003). Historically, Canada has been one of the largest suppliers of wood and paper products in the world, with 1995 shipments of manufactured forest products valued at Can$ 71.4 billion. Forest products exports from Canada contributed Can$ 39.6 billion to its net balance of trade in 2003 – almost as much as energy, fishing, mining, and agriculture combined. Canadian forests also contribute to uses and support industries providing billions of dollars in sales, including recreation, tourism, natural foods, furs, Christmas trees, and maple syrup. Much of the rapidly increasing recreation activity is forest-based. The number of visitor-days to forested national parks was 29.7 million in 1994.

Timber is harvested from about one million ha in Canada annually, or 0.7% of the total accessible, managed forest land. Allocations of timber resources are based on long-term goals for land use and forest management established in forest plans, and regional analyses and estimates of wood supply. On public (crown) lands, tenure arrangements with forest companies or communities to harvest timber are usually issued through contracts or licenses. Recent changes to legislation and tenure arrangements include provisions to license the harvesting of other forest resources such as blueberries or mushrooms. All harvest activities must also complement or integrate management objectives for wildlife, water, subsurface resources, hydroelectric energy, and transportation. The northwestern Canadian boreal forests of northern Alberta and British Columbia and the southern Yukon and Northwest Territories are the last regions of Canada to experience large-scale forest products harvest, beginning primarily in the 1980s. The installation of wood products processing facilities stimulated the expanded harvest in the northwestern Canadian boreal forest, and under the leasing system, large areas that are currently primary forest are now committed to eventual harvest.

The new emphasis in Canadian forest management typically includes the identification of objectives for the conservation of forests as a source of economic wealth, of habitat for wildlife and fish, of gene pools for biological diversity, and of water and carbon. Climate change calls into question the ability to adequately forecast future forest condition and growth and thus conduct meaningful planning. However, Canadian forest land managers are considering how to deal with climate change effects with specifically adapted silvicultural techniques for maintaining forest health, managing declining stands, regenerating disturbed areas with desired genotypes and species, and assisting in species migration (Parker et al., 2000).

Because of the large share of productive forest resources under Canadian provincial jurisdiction, an important source of leadership in developing coordinated forest policy has been the Canadian Council of Forest Ministers (CCFM), made up of the principal forestry officials of the provinces. Faced with public concerns about the extent of timber harvesting, and in response to the 1992 National Forest Strategy and the United Nations Conference on Environment and Development (UNCED), the CCFM developed a framework of criteria and indicators to define and measure progress toward sustainable forest management, in consultation with the entire Canadian forest community. The framework reflects the values of Canadians and identifies the forest features and uses they want to sustain or enhance, including indicators of environmental, social, and economic health.

Canada and 11 other countries have collaborated in the development of criteria and indicators for the conservation and sustainable management of boreal and temperate forests outside Europe (known as the "Montreal Process"). Climate change adds major uncertainty to basic assumptions about future forest condition, growth, and uses that are critical in making decisions in the present. For example, current forest harvest levels developed to meet the test of sustainability must be based on projections of future forest growth and mortality. A major climate shift would alter these factors in ways not fully understood but very likely to be disruptive. The challenge is to decide what forest activities should be allowed today based on an assumed future in which climate change outside the range experienced in previous planning horizons may be having an effect.

14.3.3. United States (Alaska)

Alaska is by far the largest state in the United States, occupying about 20% of the area covered by the remainder of the nation or an area greater than the Nordic countries combined. The two different types of forest found in Alaska are coastal rainforest in southeast and south-central Alaska and boreal forest in northern and Interior Alaska. The coastal forest in Alaska covers about five million ha, but the most productive areas of this forest type are south of the Arctic as defined in this chapter. However, much of south-central Alaska is either coastal forest or a boreal–coastal forest transition and is within the Arctic as defined here. The amount of land in the boreal region of Alaska that supports at least 10% forest cover is about 46 million ha, or 41% of the state. Statewide, about 6.4 million ha or 16.3% of total Alaska forest land is classified as "productive" forest, that is, land capable of an average growth rate of 1.4 m^3/ha/yr (Labau and van Hees, 1990). Even less of the Alaska boreal forest (12% or 5.5 million ha) is considered productive commercial timberland (Labau and van Hees, 1990).

Of the 114 million ha that make up Alaska, the federal government owned over 95% until Alaska became a state in 1959. The state government was granted the right to eventual ownership of 32 million ha (28% of Alaska) as a condition of statehood. To date the state of Alaska has received 27.4 million ha (85%) of its entitlement. Under terms of the Alaska Native Claims Settlement Act, Alaska Native corporations are entitled to receive 13.4 million ha (11.7% of Alaska), and most of that land has been conveyed. Individuals own only about 0.6 million ha (0.5% of Alaska). The private individual ownership category is expected to slowly increase as a result of government land sales and transfers. The federal government retains nearly 68 million ha in Alaska (60% of the state), including about 20.6 million ha in national parks and 31.1 million ha in national wildlife refuges.

National parks and wildlife refuges generally preclude resource development, but there are a few exceptions. In some circumstances, petroleum development can take place in wildlife refuges, and rural residents with a history of local use may obtain resources such as house logs, fuel wood, and poles for fish traps from national parks. Taking all federal land designations together, and including other protected land such as state parks, Alaska has probably the highest percentage (about 40%) of its area devoted to strict protection of natural habitats in the world. At least 25% of the productive boreal timberland in Alaska is reserved by law from forest harvest, and a similar amount is estimated to be reserved by administrative policy (Labau and van Hees, 1990). Climate change effects on this strictly preserved land base are likely to involve primarily the temporary reduction or increase in the populations of certain species resulting from land-cover change. The intactness and extent of these ecosystems enhance the prospect of species survival, even following large-scale climate change. The prescription offered by conservation biologists that best equips species to withstand major movement of optimum climate zones is to maintain large-scale, topographically diverse landscapes with naturally functioning ecosystems (Markham and Malcom, 1996). Such a strategy preserves complete gene pools and specially adapted ecotypes, and provides maximum opportunity for natural migration and disturbance recovery. The current boreal forest is largely the result of such adjustment by the biota to the many cycles of glacial and interglacial climate changes during the Pleistocene. The current land allocation situation matches the conservation biology prescription for climate change resilience better in Alaska than in almost any other major forest region of the world.

The Alaskan boreal forest is currently used for a variety of economic, subsistence, recreational, scientific, and other purposes. Local-scale logging has been a traditional use for much of the 20th century. The boreal region has only small-scale wood products facilities: mainly small sawmills and facilities to manufacture specialty products such as house logs and birchwood items. Employment in forest products manufacturing industries, mostly in the coastal region, peaked in 1990 at just under 4000 people, constituting 1.4% of total Alaska employment in that year (Goldsmith and Hull, 1994), but was only about 600 in 2002 (Gilbertson, 2002). In the 1990s, the two major wood products manufacturing facilities (pulp mills) in the state, which were supplied by long-term (50-year) contracts, permanently stopped operations. Much of the current economic activity associated with the Alaska boreal forest is generated by the basic activities of exercising the rights and responsibilities of ownership. These activities include forest inventory, monitoring conditions and trends, wildland fire management, access administration, and permits for use. Those administrative activities will occur under any scenario for the future, although they might need to be intensified under certain conditions that could be caused by climate change.

In Alaska, there are so few roads that timber removal generally must meet the costs of building or extending surface transportation routes. Much of the productive forest is distributed in scattered small stands across large landscapes. As a result, productive timberland is, with a few exceptions, not economically accessible. If low-cost forms of access (e.g., winter roads on frozen ground) can be used, the area of forest with positive stumpage value increases. Climate change will decrease the amount of time when winter access is safe on ice bridges and frozen winter roads (section 16.3.6, Fig. 16.21). Even a small amount of additional warming would initiate permafrost thawing across a sizeable portion of the interior boreal region (sections 6.6.1.3 and 16.2.2.3), severely disrupting the ground surface and causing widespread death of existing forest cover. Over the long term, however, sustained temperature increases are likely to expand the productive forest area significantly as long as available moisture does not become limiting (Juday et al., 1998).

The contributions of Alaska forests in providing subsistence food, fuel, building materials, and indirect ecosystem services, generally not measured by dollar flows, are very important in Alaska and probably exceed values from commercial timber operations in most of its boreal region. Alaska forests contribute ecosystem services especially important to the cash economy by providing commercial fisheries, sport hunting and fishing, and values of non-consumptive uses of the forest involving tourism, recreation, and enhancement of the quality of life. Scientific research is one of the most important current uses of the Alaska boreal forest, with climate change effects and carbon-cycle investigations major topics of continuing interest.

Over one million nonresident tourists visit Alaska annually, and the number of tourist visits has increased steadily over the last few decades (Northern Economics Inc., 2004). Forests create specific scenic resources for major segments of the tourism industry, including cruise ships and state ferry routes in Prince William Sound, and near the rights-of-way of the Alaska state highway system and the Alaska railroad. Beginning in the 1990s, large-scale insect-caused tree death and injury related to temperature increases was observed along several of the most popular and heavily traveled tourist routes. Recreation continues in those areas, but the quality of the experience for some visitors has been reduced.

If the assumption is made that demand and prices will rise and the public will allow expanded timber cutting under certain circumstances, the size and economic value of the Alaska forest products sector are likely to increase, and thus the risk of harm to this sector from climate change is likely to be greater. If the assumption is made that access will continue to be expensive (prohibitive), that manufacturing cost disadvantages in Alaska will persist, and that public attitudes about expanded timber cutting on the public lands are negative, the size and economic value of the Alaska forest products sector are likely to remain at current (historically depressed) levels or even decline further, limiting the magnitude of exposure to risk from climate change (Berman et al., 1999).

Much of recent Alaska forest management has been described as “opportunistic” (i.e., taking advantage of other events or projects to build programs and capabilities). Nevertheless, as inventory data accumulated, publicly built access systems expanded, and new scientific insights and data handling tools became available in Alaska, professional forest managers anticipated they would be able to set and accomplish goals more systematically. However, the increased uncertainty associated with climate change that has already been experienced in Alaska makes long-term forest planning and management considerably more difficult. It is already unclear whether there is a reasonable probability that the cost of planting or regenerating certain stands in certain regions can be recovered in future forest harvests because of higher risk from climate-triggered insect outbreaks (section 14.8.2) among other climate change effects (drying lakes, shrinking glaciers, and large burned areas) already obvious on the landscape.

14.3.4. Fennoscandia

Finland, Sweden, and Norway have certain features in common. These nations extend from at least 60º N (or further south) to north of the Arctic Circle. Across that distance, these nations encompass north-temperate deciduous or transition forest and an entire gradient of the boreal forest to treeline and tundra in the north. Despite its northerly location, the climate of the region is the warmest of equivalent latitudes in the circumpolar world, due to the strong influence of the Atlantic Gulf Stream. Annual precipitation varies from 1500 mm or greater in western Norway, which receives the strongest Gulf Stream influence, to 300 mm in the northeastern portions of Finland. The growing season lasts 240 days in the south and 100 days in the north. Mean temperatures range from 14 to 17 ºC in July to between -14 and 1 ºC in January and February. The combination of relatively mild temperatures and a deep, dependable snowpack that insulates the ground even in the coldest northern areas means that permafrost is practically absent in forest regions.

Most of Sweden and Finland is characterized by relatively even topography and is less than 300 m above sea level. In the west, mainly in Norway, the Scandinavian mountain range reaches elevations of 1000 to 2000 m above sea level. These mountain peaks are not forest-covered. The treeline varies from 700 m above sea level in the southern part of the mountains to 400 m in the northern part.

In northern areas of Fennoscandia (and the adjacent Kola region of Russia), the Saami (Lapp) people pursue reindeer husbandry in forest lands on the basis of ancient rights. The Saami are legally entitled to use lands that belong to others in order to feed and protect their reindeer herds. Saami earn their living from reindeer breeding, and reindeer move through different areas throughout the year. The forest owners affected by the herds are under a mandate to cooperate with Saami communities to ensure that reindeer can obtain their life-cycle requirements. However, because these rights are exercised across four national jurisdictions, legal systems, and boundary controls, traditional flexibility is hindered (section 12.2.5.2). This jurisdictional complexity works against adaptability of the herding system that would be desirable in response to climate change.

Due to the generally low buffering capacity of forest soils in the Fennoscandian region, a high level of air pollution, mainly originating abroad, has resulted in widespread soil acidification. Leaching of mineral nutrients has reduced soil-buffering levels by as much as half in recent decades. High levels of acid deposition (both sulfur oxides and nitrogen oxides) represent future risks to the growth and health of forest ecosystems. Overall, however, the measured rates of growth of managed and tended stands have increased considerably, which might be due at least partially to the nitrogen input.

14.3.4.1. Finland

Finland is the most heavily forested country in Europe, with 16 times more forest per capita than the European average. Forests as defined by the FAO cover 23 million ha or 74.2% of the land area. Finnish forests have been intensively harvested over the last few decades. Despite an active harvest program and the reduction of national territory after the Second World War (territory Finland ceded to the Soviet Union contained over 12% of its forest area and about 20% of the best saw-timber stands), Finnish reserves of wood volume are now greater than during the 20th century, and continue to increase. Today the annual aggregate wood growth in Finland is about 75 million cubic meters, while around 60 million cubic meters or less are harvested or die of natural causes. Of the total logged area, regeneration felling accounts for roughly one-third and thinning two-thirds.

In the boreal region, the most common trees are Scots pine (*Pinus sylvestris*), Norway spruce (*Picea abies*), silver birch (*Betula pendula*), and downy birch (*B. pubescens*). Usually two or three tree species dominate a given stand. Pure pine stands occur in rocky terrain, on top of arid eskers, and in pine swamps. Natural spruce stands occur on richer soil. Birch is commonly found as an admixture, but can occasionally form pure stands. About half of the forest area consists of mixed stands. Various kinds of peatlands are also an important part of the Finnish landscape. Originally, peatlands covered about one-third of Finland. Many peat areas have been drained for farming, forestry, and peat extraction purposes. Bog drainage reached a peak in the 1970s, when nearly 1% of the total land area of the country was being drained each year. About half of the original peatland area has been preserved in its virgin state.

Nearly all of the Finnish forest is intensively managed. In the 20th century, foresters favored conifers, especially pine, at the expense of other species. As the oldest generations of trees have been felled, the average tree age in the forest has become younger. Forest stand treatments and forest roads have fragmented large contiguous wilderness areas, and forest fires are largely prevented. Managed commercial forests of this kind now cover more than 90% of the productive forest land in Finland. About 40% of all endangered species in Finland are dependent on older age classes of natural forest (Parviainen, 1994). Any climate change effects (intensified physiological stress, fire) that disproportionately affect the remnant of older forest are virtually certain to have large negative impacts on the survival of the rare elements of biodiversity.

With the implementation of new forest management and protected area programs, the contribution of forest harvest and management to species endangerment will diminish. For the moment, however, forestry is still a greater threat to the preservation of species than any other human activity (Essen et al., 1992). One reason for this is that about half of the plant, animal, and fungus species found in the country live in forests. Of the various factors attributable to forestry that reduce biodiversity, the most important is the rarity of larger, old trees – both live and dead (Essen et al., 1992). Specialist organisms such as arboreal lichens, wood-rotting fungi, and mosses and invertebrates have declined with the decrease in large old trees, snags, and logs.

Finland has 35 national parks that cover 8150 km^2. Together with other nature reserves, the total protected area amounts to approximately 29 000 km^2 or about 9% of the total land area of Finland. The Finnish government adopted a special protection program for old-growth forests in 1996 and for broadleaf woodland in the late 1980s. There are ten programs designed to protect various types of natural features and areas. The goal is to extend protection to over 10% of the national territory. As of March 2003, about 3.57 million ha of land to be protected had been identified. About 2.7 million ha had been included in the programs by the beginning of 1999. Protecting habitats either totally or partially from human activities is the main strategy to recover endangered species and encourage biodiversity. In the case of forests, this mainly means protecting the remaining old-growth stands of both conifer and broadleaf forest growing on rich soil as public reserves, as these habitats have been declining most rapidly. There are elaborate arrangements for consultation with landowners during the planning stages of protection programs. This process of consultation was followed when the areas for inclusion in the Finnish Natura 2000 scheme were being designated. The largest category of protected areas is wilderness in Lapland. Wilderness areas, established by the Finnish Parliament, cover about 15 000 km^2 of forest, treeline, and tundra habitats. Although strictly limited, some forest cutting is permitted in some parts of these wilderness areas.

Most of the highly productive forests in Finland, especially those in the southern and central parts of the country, are privately owned. Nearly 60% of the Finnish forest is privately owned and one in five people belongs to a forest-owning family, including a broad spectrum of the population. The average size of a private forest holding is 30 ha. Most of the forest in

Lapland is publicly owned, and as a result, the northern forests have been managed more uniformly and on a larger scale than those elsewhere.

Until the mid- to late 1990s, one-third of Finnish export earnings came from forests, and the forest products industry as a whole was second only to metal products as an export sector. The economic activity (turnover) of the forest sector was roughly Eur$ 23.5 billion: management activities contributed Eur$ 1.7 billion, the forest industry Eur$ 17 billion, and machines and other equipment EurR$ 5 billion. The Finnish share of the total value of global export trade in forest products was about 7.6%. Forest management and the forest industry employed about 100 000 people while the rest of the sector employed about 50 000 people. About 100 000 timber sales deals are made every year between forest owners and forest industry companies.

An environmental program for forestry was adopted in 1994 based on forest-related principles approved at UNCED in 1992 and the 1993 general principles for sustainable forestry adopted by European forest ministers in Helsinki. The goal of Finnish forestry has become ensuring a sustainable economic return as well as preserving biodiversity and facilitating multiple use of the forests. Management practices that strongly alter the environment, such as bog drainage, deep plowing of forest soil, and use of herbicides to kill undergrowth, have been almost totally abandoned. Habitats important to biodiversity preservation have been excluded from timber operations, and both living and dead trees are now routinely left in cutting areas to support biodiversity. Natural regeneration of trees has increased in importance relative to planting nursery-grown seedlings and more attention is paid to preserving forest landscapes.

The Government of Finland adopted a National Forest Program (NFP) in March 1999. The goals of the NFP are to increase the industry's annual consumption of domestic wood by 5 to 10 million cubic meters by 2010, double the wood processing industry's export value, and increase the annual use of wood for energy to 5 million cubic meters. The Finnish government, in collaboration with the forest industry, will also ensure competitive conditions for the forest industry (e.g., supplying energy at a competitive price) and launch the technology and development programs needed for promoting the wood processing industry and wood-based energy production. In addition to wood product utilization, the NFP recognizes and promotes other forest uses, including hunting, reindeer husbandry, wild mushroom and berry picking, scenic and cultural values, recreation, and tourism.

The forest sector in Finland also makes a unique contribution toward the goal of zero net emissions of CO_2. At present, about 20% of the total energy production in Finland is based on wood, which is substantial compared to the global average. The forest industry produces about 80% of the wood energy by burning black liquor, a by-product of pulp mills, and sawdust and chips from the wood processing industry. Pulp mills are completely self-sustaining in terms of energy and are even able to supply other plants with energy. Households and small heating plants produce about 20% of the wood energy. They use primarily small dimension wood from thinning, chips made out of logging waste, and building waste.

14.3.4.2. Sweden

Sweden is a heavily forested nation. Closed forest and forest plantations cover slightly more than 60% of the nation or about 27.1 million ha. Forested shrubland, which is especially common in the transition between closed forest and tundra, represents another 3 million ha or 6.6% of Sweden. Inland waters make up almost 10% of the country, and the non-forest land base is about 23% (FAOSTAT, 2000). Sweden has 25 national parks that together cover more than 0.6 million ha, or 1.5% of the area of the country. Sweden also has 1563 nature reserves covering 2.6 million ha, or around 5.5% of the area of the country. There are also other nature reserves and protected areas that are increasingly provided by private forest owners, including individuals and large forest companies.

One striking feature in Sweden compared to other timber-producing countries is that the national government owns only 5% of the productive forest lands. In 1993, most government-owned timberland was transferred to a forest product corporation – AssiDomän – of which the national government owns 51% of the shares and the remaining 49% are traded on the stock exchange. Private individuals (families), owning approximately half of Swedish forests, are the largest single category of forest owners. Forest companies own approximately 3% and other owners account for 12%. Forest industry land holdings are concentrated in central Sweden and some portions of Norrland (the northern three-fifths of Sweden), where the industry also operates many large, modern production facilities. Swedish forest companies have globalized their operations and established themselves firmly in the European countries and on other continents. At present, some 8 million ha of forest land have been certified in Sweden. Certification is a process of formal evaluation and recognition by third-party evaluators that forest products have been produced using sustainable management practices, and can be advertised to consumers as such. Forest companies account for most of the certified forest land.

Sweden has the highest population (8.8 million) of the Nordic countries, but is still largely a sparsely populated country, especially in the north. Despite the relatively low population density, Swedish forests show the effects of many centuries of human use. Only in the northern interior are the forests less affected by humans. Today much of the northern Swedish forest is protected, either as nature reserves or by other means.

The average size of a private forest is about 50 ha. Until the Second World War, most private forest owners were farmers who lived on their property and were engaged in crop cultivation as well as harvesting wood products. Since then, the area of forest land devoted to this type of combined agriculture and forestry enterprise has dropped from more than 9 million to less than 4 million ha. Today many individual forest owners do not live on their forest land, but often in communities close to it or in more distant cities and towns. The bulk of forestry work on their properties is now performed by employees of forest owners associations or by other contractors.

Early in the 19th century, Sweden began the process of industrialization by rapidly expanding its sawmill industry. The major Norrland rivers were suitable for floating timber to the Baltic Sea, thereby opening up previously untouched inland forests to large-scale logging. The sawmill companies, which purchased very large tracts of land from farmers until 1905, soon gained a strong position in European timber export markets. Northern Sweden was very sparsely populated until the beginning of the 19th century. At that time, an accelerating agrarian colonization took place, and later in the century, large-scale exploitation of the virgin forest began. The first forest resources to be exploited were those close to the Bothnian coast, and subsequent exploitation moved further inland. This exploitation affected almost all forest land in northern Sweden between 1850 and 1950.

The boreal forest landscape in Sweden has changed dramatically during the last 150 years. Owing to the vigorous pursuit of forest management goals over the last 100 years, widespread stands of commercially tended Scots pine of generally similar age and structure make up most of the northern Swedish landscape (Essen et al., 1992). Less than 3% of the forest area of Sweden supports trees older than 160 years, and less than 2% of the conifer volume is in trees with diameters greater than 45 cm (Essen, 1992). The volume of standing dead trees in Sweden has decreased by more than 90% in the last century and the number of large-diameter conifers (>30 cm diameter at breast height) has decreased by more than 80% (Linder and Ostland, 1992). Many old-growth specialist species, including arboreal lichens, mosses, insects, wood-rotting fungi, and cavity-nesting insectivorous birds, have become rare in boreal Sweden (Essen, 1992). All the taxonomic groups that are represented in the 1487 threatened forest species in Sweden include elements that are largely dependent on old forests or the habitat elements of old forests (Berg A. et al., 1994). Many of these species belong to functional groups, such as insect predators, that perform vital functions for the health and productivity of the forest system as a whole. As a result, the northern Swedish forest is particularly vulnerable to any climate change effects that would accelerate tree mortality in older stands.

About 20% of the energy consumption in the country originates from forest biomass, and there is presently a significant annual accumulation of carbon in standing biomass, equal to approximately 30% of Swedish fossil fuel emissions. The high level of air pollution originating outside the country and the accelerated leaching of mineral nutrients in Sweden has resulted in widespread soil acidification. Forest-damaging amounts of both sulfur dioxide and nitrogen oxides are being deposited. The accelerated leaching of mineral nutrients caused by acid deposition has reduced nutrient levels by half on some sites in recent decades, although nitrogen deposition increases supplies of a critical nutrient. Air pollution poses a serious threat to the health and growth of Swedish forest ecosystems. In general, researchers believe that multiple stresses may act synergistically, resulting in acceleration of change due to other stresses (Oppenheimer, 1989). Climate change, therefore, must be understood to be acting in this context in Sweden, as one among many factors.

The present-day timber stock is 50% larger than it was when detailed measurements began in the 1920s. However, for much of the early and mid-20th century, forests in Sweden were exploited in ways that today, with increased knowledge and insights, are understood to be detrimental to the environment. Clear-cutting was extensive, and little effort was made to protect biodiversity. Since the 1992 UNCED meeting, the concept of sustainability has broadened. This is reflected in the revised Swedish forestry policy that went into effect in 1994. The 1994 Forestry Law states: "The forest is a national resource which should be managed so that it provides a good return on a sustainable basis and ensures the preservation of biodiversity". These two goals have equal priority (Lamas and Fries, 1995). Underlying this policy is the conviction that there will continue to be a demand for renewable products in the future and that Swedish forests can remain an important source of natural raw material produced using principles based on ecological cycles. Swedish forest legislation protects key woodland habitats, forests located near high mountains, and wetland forests. Special regulations govern four million ha of low-productivity woodlands that are not included in the productive forest land, allowing only careful low-intensity utilization and ensuring that their basic character remains unchanged. Hydrologic and other possible changes resulting from climate change represent potential risks for some of these protected features.

Swedish forest policy also states that forests should be able to sustain hunting and the gathering of wild mushrooms and berries as well as active silviculture. The "right of common access" is traditional in Sweden. People are entitled to hike through the natural landscape and to pick mushrooms and berries regardless of who owns the land. This tradition broadens the pool of resources and users with specific concerns about potential climate change impacts beyond those of timber and a comparatively small number of forest-land owners. Hunting and fishing remain widespread and significant activities in Sweden, and are considered in forest management since they are connected to key management practices. Sports such as orienteering, cross-country skiing, and other popular outdoor activities that take place in the forests involve a relatively large proportion of people in Sweden. Participants in these sports have a strong interest in the health of the forests and are an important forest and climate change constituency even though they do not remove products from the forest.

14.3.4.3. Norway

Forests and other wooded land cover about 37%, or 11.9 million ha, of the Norwegian mainland. Nearly 23%, or 7.2 million ha, is regarded as productive forest. The productive forest is distributed among 125 000 forest properties, and about 79% is owned by private individuals.

The total area of wilderness territory in Norway has been greatly reduced over the past 100 years. Wilderness territory is defined as areas more than 5 km from roads, railways, regulated watercourses, power lines, and vehicle tracks. Although the largest 20th-century reduction in wilderness occurred in the lowlands of southern Norway, where wilderness is now virtually eliminated, mountainous areas and northern Norway have experienced major wilderness reductions and fragmentation as well. Today, wilderness represents 12% of the total land area of Norway (not including Svalbard and Jan Mayen). Approximately 6.4% of mainland Norway has official protected area status. New protection plans, especially for additional national parks, are being developed, and about 15% of the land base will be within protected areas by the year 2010. Both public and private lands that lie within or adjacent to designated national parks are protected from construction, pollution, or any other encroachment.

Norwegian forests have been exploited intensively for the export of roundwood, sawn timber, and wood tar for hundreds of years. In addition, there is a long tradition of using the forests for domestic animal grazing and game hunting. For many years, the amount of wood removed from Norwegian forests has been less than the amount of growth. The biomass of standing trees has almost doubled since 1925, and the volume of standing forest increased by more than 95% from 1925 to 1994, to about 616 million cubic meters. In 1994, the net increase (increment or growth minus removal) in forest mass was 9.5 million cubic meters, or 1.5% of the total volume of standing forest. The 1994 volume of Norwegian forest consisted of 46% spruce, 33% pine, and 22% deciduous trees. The net volume increase was greatest for deciduous trees and pine. Because of their increase in total biomass, Norwegian forests have contributed to a net transfer of CO_2 from the atmosphere into storage during the 20th century. It is estimated that in 1994, the net amount of CO_2 assimilated by forest was 15 Tg, amounting to about 40% of Norway's annual anthropogenic emissions.

The increase in wood volume can be attributed to many factors. A sustained effort at intensive forest management has systematically removed older stands that have low rates of net growth, and increased the proportion of stands in the early and most rapid stages of growth. An extensive afforestation effort has established forests on sites such as formerly open wetlands that previously supported no (or only minor) forest cover. Natural forest vegetation has returned to uncultivated land that is no longer farmed, producing a gradual buildup of vegetation mass. Long-range transport of nitrogen in precipitation has had a fertilizing effect on Norwegian forests. Finally, the introduction of new species that grow faster than native species on some sites has also led to greater forest growth. It should also be noted that the particularly large increase reflected in the inventory data for the most recent years is probably due in part to new methods of calculation.

Today the forest is used first and foremost as a source of raw materials for sawmills and the pulp and paper industries. Government grants or subsidies for forest planting began as early as 1863 in Norway. In recent years, forest planting has been somewhat reduced to between 200 and 300 km^2 of forest planted or sown annually. Norwegian plantations are often spruce monocultures. The diversity and abundance of the fauna and flora in planted spruce forest is much lower than that in naturally regenerated forest.

Although the volume of forest has increased considerably since the beginning of the 20th century, the present-day forest has been transformed from its original condition. Clear-cutting, plantation establishment, introduction of alien species, ditching, and forest road construction are some of the intensive forest management measures that have been applied. Acidification and pollution have also affected forests. As in the other Nordic countries, management practices applied in order to produce increased tree biomass have caused a reduction in biological diversity. Much of the present forest growth is concentrated in large monocultures. Large stands of the same age class contribute to a reduction in the number of species of flora and fauna compared to what would occur in a more natural, mixed type of vegetation.

Forestry and the forest industry are important sectors of the Norwegian economy. Wood and forest products represent about 11% of Norwegian non-oil exports. This is slightly less than the export value of the Norwegian fishing industry, somewhat higher than both aluminum and natural gas export values, and twice the value of Norwegian high-technology exports. About 30 000 people receive their income from primary forestry and the forest industry. Most of the economic activity in the forestry sector takes place in rural districts and is particularly important in these areas that have fewer economic alternatives than urban zones. Norway began exporting large quantities of sawn lumber in the 16th century and continued this practice for several centuries. However, since the Second World War, most of the sawn lumber has been utilized nationally. Most homes in Norway are constructed of wood, with wooden interior fittings, and wood is an important part of the everyday life of Norwegians. In 1995, the wood processing industry used 5.1 million cubic meters of wood and employed about 16000 people.

Sawmills use about 50% of the Norwegian roundwood harvested. There are 225 sawmills in Norway operating on an industrial scale. Paper products have the highest export values of all the forest-based products. Paper and board products are currently produced by 36 different machines in Norway. Wood pulp and chemical pulp are produced by 17 production units. Every year, Norway exports about one million tonnes of newsprint. The pulp and paper industry employs 9000 people.

Until a century ago, wood was the dominant energy resource in Norway. Oil and hydropower are currently the major energy sources, and fuelwood use is only 7% of the volume used 100 years ago. However, the forest may play an increasing role as an energy supply. Shortages of electricity and CO_2 taxes on the use of oil have increased the interest in bio-energy. Today, the pulp and paper industry is by far the largest producer of bio-energy in Norway.

14.3.4.4. Iceland and Greenland

As a geologically young landform situated along the mid-Atlantic Ridge, isolated from both Europe and North America, Iceland has always supported ecosystems with a restricted set of species (only about 483 native and naturalized vascular species; Icelandic Institute of Natural History, 2001). Iceland is also the area of the Arctic with the most substantial human impact on ecosystems. At the time of Viking settlement in the year 874, vegetation was estimated to cover about 65% of Iceland, while vegetation covers only about 25% of the island today (Blöndal, 1993). This reduction in vegetative cover is most likely the result of intensive land and resource utilization by a farming and agrarian society over 11 centuries, although recent investigations also suggest that specific years of low temperature and low precipitation possibly initiated devegetated patches under a traditional winter grazing system that was abandoned in the early 20th century (Hellden and Olafsdottir, 1999). Estimates vary as to the percentage of the island originally covered with forest and woody vegetation at settlement, but a range of 25 to 30% is plausible (Blöndal, 1993). The native forest of Iceland was principally comprised of downy birch and a few tall willow species with a relatively large proportion of the woody stems in a dwarfed and somewhat contorted growth form. Rowan (*Sorbus aucuparia*) and European aspen (*Populus tremula*) occasionally occur among the birches. Tea-leaved willow (*Salix phylicifolia*), hairy willow (*Salix lanata*), and creeping juniper (*Juniperus communis*) are common shrubs in birch woodlands. Ancient native birch woodlands and scrub are key areas of terrestrial biodiversity in Iceland. The birch woodlands still retain most of their original biodiversity and are a key habitat for a number of threatened and endangered species, while the deforested land has suffered degradation and in some cases has been reduced to deserts.
When Norse farmers, traders, warriors (Vikings), and seafarers arrived in Iceland from western and/or northern Scandinavia, Ireland, and Britain in the 870s and 880s, the land was uninhabited and vegetated from the seashore to all but the glaciated mountains. At higher altitudes, the woodlands gave way to dwarf birch (*Betula nana*), willow, grasses, and moss. Birch woodlands separated by wetlands dominated the lowland vegetation. The lowlands were the only habitable land areas and were occupied in the first decade or two of settlement. Most of the first settlers were farmers that brought livestock to the island. Forests were cleared rapidly by burning and harvest for building materials and charcoal manufacture, generally within 50 years of the arrival of the settlers. Nearly all of the habitable woodlands were converted to pastures and only the mountainsides and highland margins retained forests, which survived to the early 19th century.

In southern and western Iceland, the land was cleared and barley was grown. Domestic livestock were sustained at numbers that caused landscape degradation, soil loss, and reduced long-term productivity of pastures. A pollen record from Iceland (Hallsdóttir, 1987) confirms the rapid decline of birch and the expansion of grasses between AD 870 and 900, a trend that continued to the present. As early as AD 1100, more than 90% of the original Icelandic forest was gone and by 1700, about 40% of the soils had been washed or blown away. Vast gravel-covered plains were created where there was once vegetated land.

The exploration of Greenland in 984 by Eric the Red gave the more adventurous Icelanders new farm sites and freedom from limits on establishing themselves in an occupied landscape. The small areas of woodlands in Greenland, dominated by birch and willow in the south and alder (*Alnus* spp.) in the valleys of the Western Settlement, were highly valued. The Greenland Norse were also farmers, although they supplemented their diets with seal, birds, and reindeer/caribou. For several centuries the Norse remained committed to their farming way of life but gradually experienced various stresses (land degradation, trade and genetic isolation, social and economic challenges), many of which were aggravated by a cooling climate. The Norse settlements in Greenland and Iceland were established shortly after the beginning of a period of relatively warm and stable climate that persisted from the 8th through the 12th centuries (Ogilvie and McGovern, 2000). With the gradual degradation of the landscape that reduced its capacity to support livestock and crops, and during a severe cooling period between 1343 and 1362, the Norse presence in the Western Settlement of Greenland ended and the Norse population disappeared. The mid-14th century cold interval in Greenland was probably the period of the lowest temperature in the last 700 years, and the effect of this climate change on the Norse community is highly likely to have been an important factor in its terminal decline (Ogilvie and McGovern, 2000). Simultaneously, arctic-adapted

Thule people arrived for the first time as the culmination of their historic expansion eastward and may have played a role in displacing the indigenous Norse people who first inhabited the area.

Until quite recently, efforts to conserve and increase the area covered by native woodlands in Iceland were hindered because the majority of woodlands were subject to grazing, mainly by sheep: the total area covered by birch woodlands does not appear to have changed significantly in recent decades. However, the establishment of the Icelandic Soil Conservation Service in 1907 was instrumental in fencing off and halting the most severe erosion processes in lowland areas of southern and northeastern Iceland. This is a unique context in which to consider climate change effects in the Arctic – humans have damaged soils and eliminated so much of the native forest ecosystems in Iceland that climate change represents an opportunity for forest recovery and expansion.

Starting in 1907 but especially since the mid-20th century, efforts were made to establish forest plantations in Iceland. Plantations now total at least 20000 ha (Gunnarsson K., 1999). Until the 1980s, the lands most suitable (from the standpoint of soils and climate) for re-establishing native birch forests remained committed to farming and winter fodder production for sheep and horses. However, since Iceland joined the European Economic Area and the World Trade Organization, Icelandic agriculture has been integrated into a trading and market zone in which many of its products fall under production quotas or price disadvantages. Farms are going out of traditional agricultural production, while other forms of land use, such as afforestation and horse farming, are increasing.

About 130 tree species from most of the cold regions of the world have been evaluated for their growth and survival characteristics (Blöndal, 1993). Different types of forest are being established for particular purposes. Native birch forest is often favored in recreation areas near urban centers and private land surrounding weekend recreation cottages in the country. Usually, land that is being converted to native birch forest must be fenced to exclude sheep. Protection forests are designed to prevent erosion of exposed soil, which is still quite common. Species commonly used in protection forests include downy birch, Sukachev larch (*Larix sukaczewii*), Siberian larch (*L. sibirica*), lodgepole pine (*Pinus contorta*), willows, and alder. A program to afforest denuded or severely eroded land has resulted in the planting of about one million trees in 70 areas subject to soil erosion. Production forests are designed to produce timber, fuelwood, and Christmas trees. Planted species include Sitka spruce (*Picea sitchensis*), Sukachev and Siberian larch, lodgepole pine, and black cottonwood (*Populus trichocarpa*). State-supported afforestation projects were initiated in recent years in all parts of Iceland, starting in eastern Iceland in 1991. The aims of the projects are to create jobs in the rural areas (e.g., in nurseries) and to support jobs on those farmsteads taking part in the project. Farmers who participate in the project can receive a state grant of 97% of the total cost of afforestation on their farmlands (Gunnarsson K., 1999).

Until the 1970s, most plantations were established within the existing birch woodland, but that is no longer the case. Approximately 2000 ha of mainly non-native tree species have been planted annually in Iceland since 2003. In the late 1980s, forest and woodland vegetation covered only about 1.33% of Iceland, amounting to about 117000 ha of native birch woodland and 15000 to 20000 ha of plantations. More than 80% of the woodland remaining in Iceland was less than 5 m in height in the 1980s, and 14.2% of the birch woods have trees taller than 5 m (Sigurdsson and Snorrason, 2000). In 2001, Iceland had 3 national parks totaling 189000 ha and 35 nature reserves totaling 225000 ha (Icelandic Institute of Natural History, 2001).

The main challenge to successful introduction of tree species to Iceland has been finding varieties that can survive and reproduce in an environment of strong, steady winds and growing seasons at the cool margin of tree tolerance. Warmer climate conditions generally increase the planting success and particularly the growth of trees in Iceland. The restructuring of the rural economy to include less production agriculture and more ecotourism, recreation, and amenity-based uses is supporting the trend toward the conversion of the Icelandic landscape from a treeless farming region into one with expanded forest area. The government program to sequester CO_2 has increased afforestation efforts. Most of the tree planting for carbon sequestration is contracted to farmers. A national goal is to expand forest cover on 1% of the national area per decade until forests cover about 5% of Iceland, which would represent the greatest forest extent in a millennium. In 1995, the estimated annual carbon sequestration by Icelandic plantation forests was 65 Gg CO_2, or about 2.9% of the annual national emissions at the time. By 1999, the annual sequestration totaled 127 Gg, representing 4.7% of annual emissions (Sigurdsson and Snorrason, 2000).

Approximately 70% of forests and woodlands in Iceland are privately owned, with the remainder owned by the national government or local municipalities. The Iceland Forest Service, an agency under the Ministry of Agriculture, is primarily responsible for implementing and monitoring government forest policy. The Environment and Food Agency of Iceland, part of the Ministry for the Environment, plays a role in the conservation of protected native birch forests. The main forestry laws in Iceland include the Forestry Law (1955), the Farm Afforestation Law (1991), and the Southland Afforestation Law (1997). In response to a request by the Icelandic parliament, the Iceland Forest Service prepared a comprehensive afforestation plan for all of Iceland as a part of its sustainable development strategy, resulting in the Regional Afforestation Law (1999). The Icelandic Soil Conservation Service, also under the Ministry of Agriculture, is responsible for implementing the Soil Conservation Law (1965), which also contains provisions for vegetation conservation and restoration.

14.4. Use and evaluation of the ACIA scenarios

14.4.1. Method of analysis

This chapter uses the projections from the five ACIA-designated models (section 4.2.7) primarily in the form of numerical output for key climate variables in specific grid cells across the area of analysis. Criteria for selecting the models and assumptions used in generating the scenarios are described in Chapter 4. For the purposes of this chapter, 14 sites were chosen that are broadly distributed across the northern part of the boreal region or are located somewhat south of the Arctic but currently experience climate conditions projected to occur in the Arctic and boreal region by the ACIA-designated models. Most of the locations were chosen because they have a medium- to long-term record of climate conditions. Many are major population, trade, or transportation centers within their respective regions, and the sites are representative of most regions in the far north where agriculture and forest management are currently (or potentially could be) practiced. The selected locations include areas in the eastern, central, and western portions of both Eurasia and North America (Table 14.2 and Fig. 14.1).

Table 14.2. Sites used for evaluation of the ACIA-designated model projections. Numbers in first column are those used in Fig. 14.1. Analyses of the Taymir Peninsula used the mean of the four stations on the peninsula.

	Location	Latitude	Longitude
1	Fairbanks, Alaska, United States	64.82° N	147.52° W
2	Big Delta, Alaska, United States[a]	63.92° N	145.33° W
3	Whitehorse, Yukon Territory, Canada	62.72° N	135.10° W
4	Fort Vermillion, Alberta, Canada	58.40° N	116.00° W
5	Thompson, Manitoba, Canada	55.80° N	97.90° W
6	Goose Bay, Newfoundland, Canada	53.30° N	60.4° W
7	Reykjavik, Iceland	64.13° N	21.56° W
8	Tromsø, Norway	69.83° N	18.55° E
9	Umeå, Sweden	63.8° N	20.30° E
10	Rovaniemi, Finland	66.55° N	25.00° E
11	Archangelsk, Russia	64.5° N	40.42° E
12	Novosibirsk, Russia	55.00° N	82.90° E
13	Yakutsk, Russia	62.00° N	130.00° E
14	Magadan, Russia	59.60° N	150.80° E
Stations on the Taymir Peninsula, Siberia			
15	Dudinka	69.67° N	86.28° E
16	Essej	68.78° N	102.62° E
17	Khatanga	71.97° N	86.78° E
18	Olenek	68.83° N	112.72° E

[a] Analysis of growing degree-days only

The primary variables obtained from the climate model output were mean monthly temperature, total monthly precipitation, and growing degree-days (GDDs). Mean monthly temperature is the mean of all mean daily temperatures for the month. Mean daily temperature is the mean of the high and low temperature on a given day. Mean monthly temperature integrates much of the short-term variability in weather and yet preserves a specific seasonal signal of warmth or coolness that has proven useful in many applications such as tree-ring studies. Total monthly precipitation reflects the amount and timing of moisture, and is widely used in forestry, agriculture, and hydrological applications. Growing degree-days are a measure of accumulated heat energy, calculated by taking the sum of daily mean temperatures above some defined threshold. For example, if the threshold value for accumulation is 5 °C, then a series of daily means of 2, 6, 7, and 5 °C would accumulate respectively 0, 1, 2, and 0 degrees for a total of 3 GDD.

Output climate variables from the ACIA-designated models were obtained generally for the period 1990 to 2099, which allowed an 8- to 10-year overlap with recorded data (typically 1990 or 1991 through 1998 or 1999). In some cases, recorded data from stations were truncated in 1989 or 1990, in which case model output from 1980 to 2099 was used in order to generate a 9- or 10-year overlap. For each station, the differences between monthly or annual values of the model output of temperature or precipitation representing the entire grid cell and the values measured at the specific recording station were calculated for the overlap period. From these differences, an overall mean difference was calculated and applied as a correction factor to the model output for all years beyond the overlap period. The variance was not adjusted, although some features of scenario versus recorded variance were examined (see section 14.4.4). The adjustment of general circulation model (GCM) output for use at a specific location is an active area of ongoing research (e.g., Hewitson, 2003).

14.4.2. Size and placement of grids

Grid cells in GCMs are large polygons that represent the surface of the earth. In the output from the ACIA-designated models, a specific climate variable is represented by a single value for an entire grid cell. Climate at a specific station location within the grid cell can be different from the value that expresses the entire grid cell for several reasons. It is important to understand what causes some of these differences in order to properly evaluate the scenario output and the ways it is used in this chapter.

The different GCMs use grid cells of different sizes and non-uniform placement. A grid cell that contains a given station location may be represented as a polygon largely north of the station in one model but largely south of the station location in another model, and the temperature values generated by the two models will have a different relationship to the location of the climate station. The topography of the earth is another factor that can cause a difference between a climate station and a grid cell value. In one model, values for high- and low-elevation terrain may be integrated to calculate a single number for the climate parameter, whereas in another model the grid cell may be placed in a way that the surface of the earth is more uniformly high or low.

Mountain terrain often produces sharp gradients in precipitation. As a result, grid cells that include mountains and smaller valleys can exhibit particularly large differences in precipitation compared to recording stations that are typically located in low-elevation valleys.

A small grid cell occupying an area along a mean frontal position in one model will show greater climate variability than a larger cell in the same model even if both contain the same recording station. In addition, different models produce different patterns of air-mass mixing, and as a result different patterns of mean frontal positions.

14.4.3. Range of scenarios

The ACIA-designated models project a range of annual and seasonal mean temperatures by the end of the model runs in 2099. The grid cell containing Ari Mas (on the Taymir Peninsula, Russia, and site of the mostly northerly treeline in the world; Jacoby et al., 2000) provides an example of these different temperature projections (Fig. 14.4). The Ari Mas grid-cell data was adjusted as described in section 14.4.1 using the mean annual temperature from four stations (Dudinka, Essej, Khatanga, and Olenek), which averages about -11.5 ºC for the available 20th-century record. The long-term mean (regression line fit) for the CGCM2 output of mean annual temperature adjusted to the stations reaches about -7.2 ºC at the end of the scenario period, while the CSM_1.4 output increases to about -2.7 ºC by the end of the period (Fig. 14.4). The CGCM2 projects a climate in this grid cell at the end of the century that is within the range of forest–tundra transition conditions, whereas the CSM_1.4 projects mean annual temperatures typical of regions where today some of the more productive examples of the northern boreal forest occur.

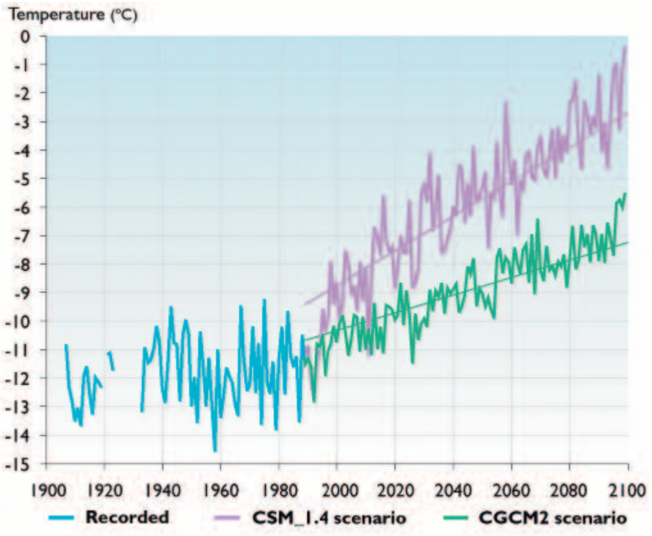

Fig. 14.4. Observations and projections of mean annual temperature for Ari Mas, Taymir Peninsula, Russia. Observations represent the average of four climate stations (Dudinka, Essej, Khatanga, and Olenek) on the Taymir Peninsula; projections from the CSM_1.4 and CGCM2 models have been adjusted using the average mean annual temperatures from these four stations (see section 14.4.1).

It was sometimes possible in the analyses to calibrate the response function of a key outcome, such as tree growth, to climate. Often the different scenarios produced similar results, with differences mainly in short-term (interannual) variability. The graphic display of such results overlaps to such an extent that clear trends are difficult to see. As a result, when the outputs of a climate variable of interest were broadly similar among the five scenarios, the high- and low-end members were used to examine the range of consequences produced by the scenarios. Using the Ari Mas region again as an example, warm-season (May–September) temperatures projected by the ECHAM4/OPYC3 model increase by about 5.7 ºC and those projected by CSM_1.4 by about 3.0 ºC (Fig. 14.5). The warm-season mean is closely related to the mean radial growth of trees at this location, so the scenarios provide some basis for understanding what the effects of temperature increases at this location might be on tree growth and vigor (sustained rate of growth over time). Note the generally close relationship in Fig. 14.5 between the regional tree-ring growth signal, and the mean May through

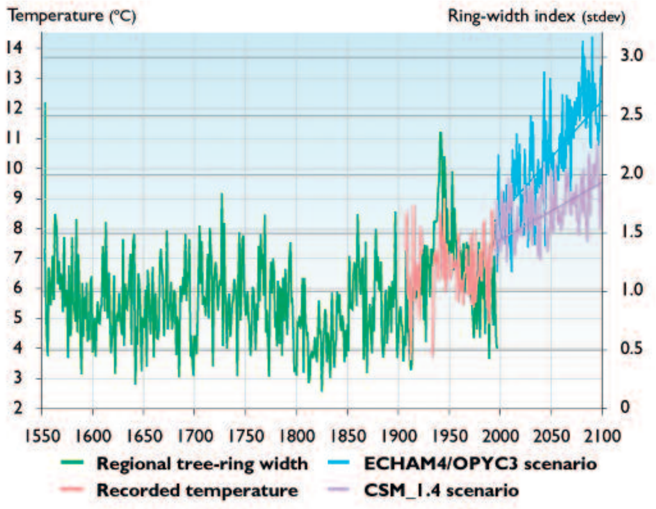

Fig. 14.5. Historic relationship between growth of Siberian larch and warm-season (May–Sep) temperature, and projections of future warm-season temperature increases in the Taymir Peninsula, Russia. Observed temperatures are the average of four stations (Dudinka, Essej, Khatanga, and Olenek); model-projected temperatures are for the entire grid cell and have been adjusted (see section 14.4.1). Tree-ring widths are shown as de-trended and normalized index values (standard deviation from the long-term mean of each measured series; data from Jacoby et al., 2000).

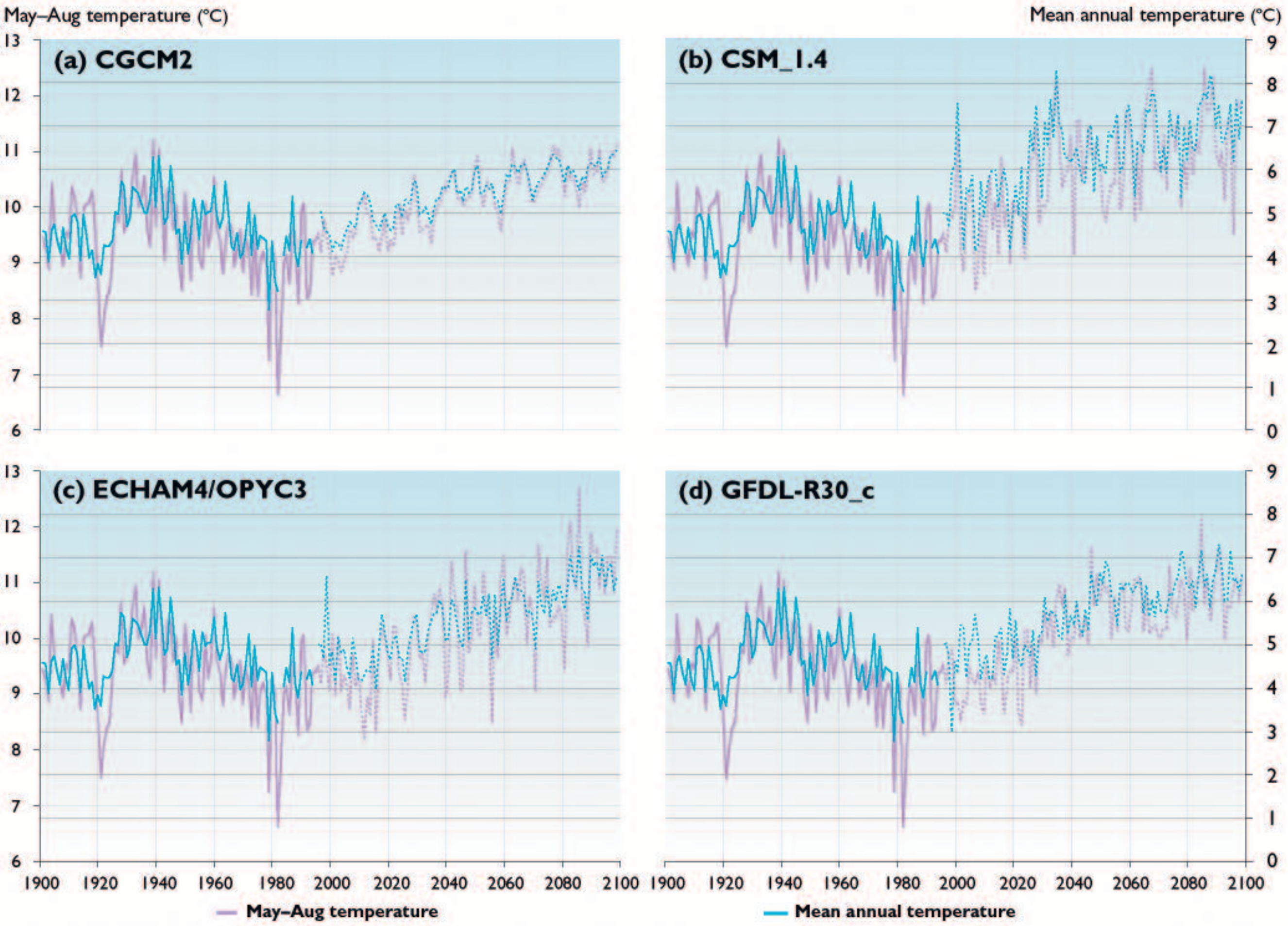

Fig. 14.6. Relationship between summer (May–Aug) and mean annual temperature at Reykjavik, Iceland, for the period of the instrumental record and the ACIA scenario period.

September temperature. Siberian larch is heat-limited and displays a positive growth relationship with temperature. The ECHAM4/OPYC3 scenario for the 21st century projects temperatures that would approximately double the rate of growth and make this marginal site a productive forest. The CSM_1.4 scenario does not project the same degree of warming, but would eliminate periods of severe growth limitation. It should be remembered that the relationship of tree growth to climate may change under an altered climate.

Very few of the examined model outputs showed cooling or no change. The exception in the analysis is Reykjavik, Iceland. The CGCM2 and GFDL-R30_C scenarios for the grid cell that contains Reykjavik project mean annual and summer (May–August) temperatures that generally are not warmer than 20th-century observations, and temperatures at the end of the 21st century that are near the long-term means of the recorded data (Fig. 14.6a,d). Other model scenarios for the Reykjavik grid cell generally show a warming trend (Figs. 14.6b,c), but the magnitude is notably less than most of the other grid cells and stations examined. The four scenarios shown in Fig. 14.6 produce higher values of both summer and mean annual temperature than were recorded during the mid-20th century temperature peak, but in this highly maritime area none of the scenarios greatly exceed the mid-20th century temperatures. Note the variability in the degree to which the different scenarios reproduce the historical relationship between the two variables depicted.

14.4.4. Variability and seasonality

In addition to long-term trends and magnitudes of temperature increase, the models produce different features of shorter-term climate variability. Often these short-term climate events or minor trends can be decisive ecological influences. For example, even one or two years of high temperatures can be enough to produce extreme risk of forest fire (Johnson, 1992) or optimal conditions for major outbreaks of insects that kill trees or reduce their growth (section 14.8). Conversely, cold spells during the growth season may harm or kill the trees. A short-term period of favorable weather may trigger production of especially large tree seed crops, with long-lasting ecological consequences (Juday et al., 2003). The stations used for scenario analysis exhibit the opposite temperature relationships mentioned in section 14.2.4 (Fig. 14.7). When summer (June–August) temperatures are high at Fairbanks, Alaska (western interior of North America) they tend to be low at Yakutsk, Russia (eastern interior of Eurasia) (Fig. 14.7a). When mean annual temperatures at Goose Bay, Labrador, Canada (eastern edge of North America) are high, they

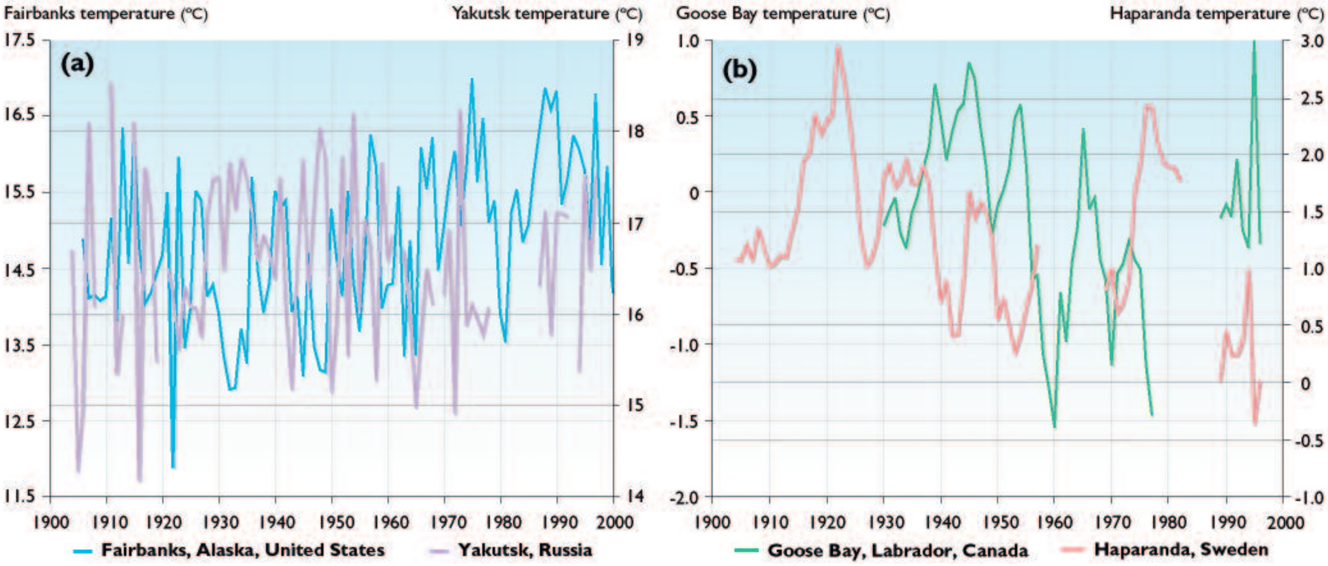

Fig. 14.7. Opposing temperature relationship between stations located in eastern and western portions of the continents in (a) summer (Jun–Aug) and (b) annual (5-yr running mean) (data from the Climatic Research Unit, University of East Anglia, United Kingdom, 2003).

tend to be low at Haparanda, Sweden (western edge of Eurasia) (Fig. 14.7b).

Because these features of climate variability are so important in producing ecological effects, it is important to understand the behavior of the models with respect to key features of short- and medium-term variability and seasonality.

Most of the ACIA-designated model outputs for the grid cells examined reproduce something close to the historical record of the difference between summer and mean annual temperature (Figs. 14.6 and 14.8a,c,d,e). However, sometimes scenario outputs show a different relationship between summer (May–August) and mean annual temperature than do the recorded data (Fig. 14.8b). All the model scenarios shown in Fig. 14.8 project temperature increases that are likely to substantially alter the moderately temperature-limited forest climate typical of the 20th century. The models project temperature ranges that would surpass thresholds for key factors such as outbreaks of insects that attack trees, new species, and altered ecosystem functions such as growth, fire, and decomposition.

The CGCM2 projections for the grid cell containing Fairbanks, Alaska, do not reproduce the same relationship between temperature trends in the coldest winter months and mean annual temperature as the recorded data (Fig. 14.9), and this disparity widens with time. It is possible that climate change would actually produce such an effect, but not all the scenarios agree on such features and realistic projections of effects on forests and agriculture require the resolution of such discrepancies. Different models have different strengths and weaknesses and all should be applied with an awareness of their limitations, circumstances in which they perform well, and novelties that could affect assessments of key climate-dependent processes. Differences in the relationship between seasonal and annual climate variables can have particular ecological importance. For example, a higher growing-season temperature may increase the growth of trees, leading to one set of effects, but an increasing mean annual temperature may promote thawing of permafrost, which could undermine or destroy the soil rooting zone for trees growing on such sites. Thus, different rates of temperature increase (annual versus seasonal) may produce quite different ecological effects.

14.4.5. “Surprises” in climate change effects

As with any use of models to project the future, a note of caution must be introduced. The projections of climate change and impacts on forests, land management, and agriculture are only as good as the climate models upon which they are based. It is also important to understand that the consequences of change introduced into ecosystems often include elements of chance and contingency that have very real and long-lasting consequences. Even though ecological events *could* unfold in a variety of ways under a given climate scenario, when events *do* take a certain pathway of cause and effect, this closes off another set of outcomes from that point onward. These elements of chance and contingency make it difficult to project ecosystem responses. However, other elements can assist in the projection of ecosystem responses. Including large areas in the analysis allows processes that have a variety of potential outcomes (e.g., the number of trees of a certain species reproducing successfully after a fire) to occur repeatedly. Any process that expands the pool of possible outcomes (either simulated outcomes or those that can be measured in ecosystems) of the climate change “experiment” will then cause the results or outcomes to begin to approach some distribution that may be described. Repeated outcomes can be produced not only as the result of multiple examples across space, but as the result of multiple outcomes across time in the same place. As a result, when larger areas and longer periods

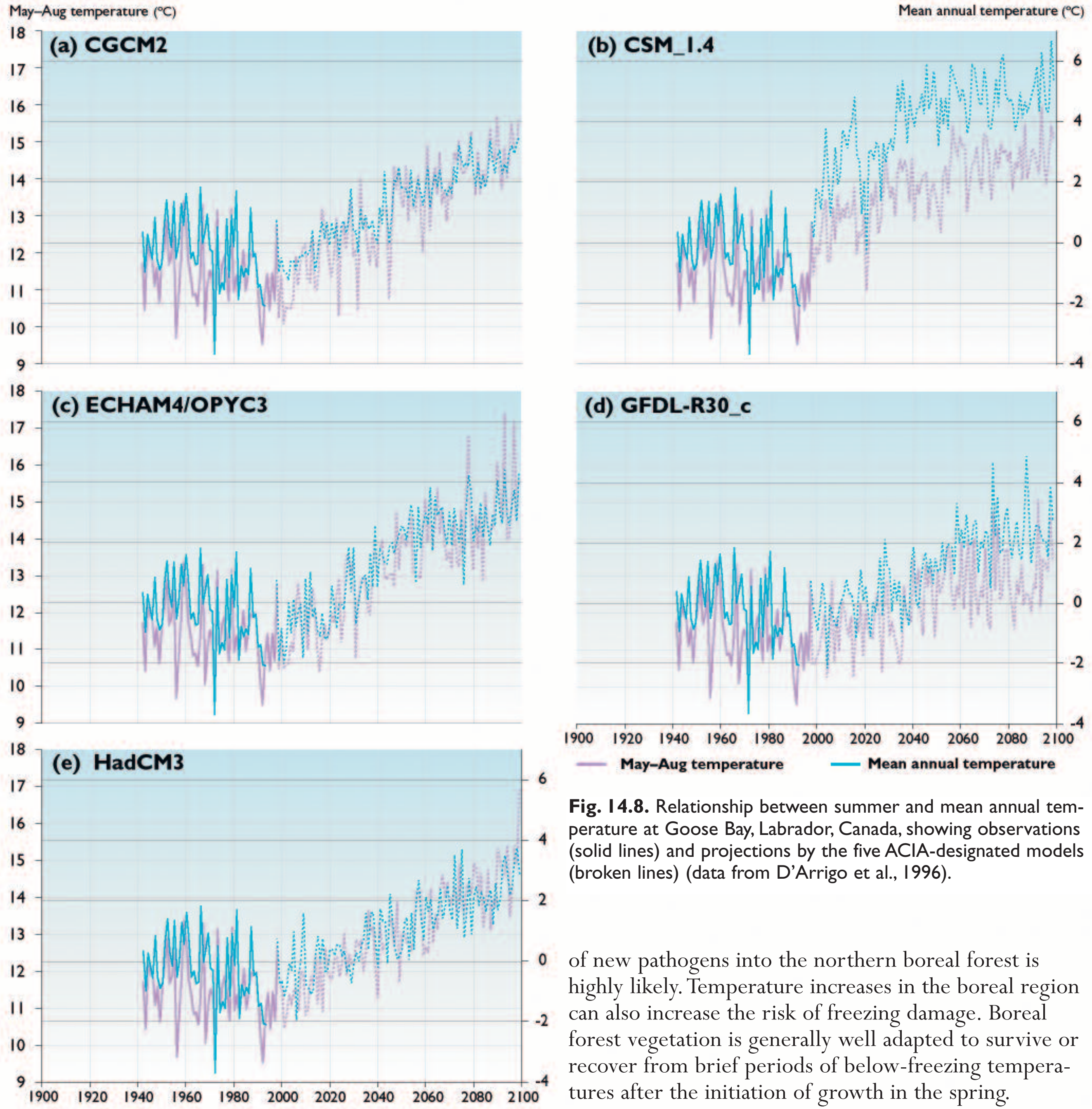

Fig. 14.8. Relationship between summer and mean annual temperature at Goose Bay, Labrador, Canada, showing observations (solid lines) and projections by the five ACIA-designated models (broken lines) (data from D'Arrigo et al., 1996).

of analysis are involved, the outcomes in ecosystems can be better described.

The climate changes that are the most difficult to project are low-probability, high-impact events. By definition, there are few examples of such events to learn from; in fact, analogous events may not have occurred during the period of the observational record. An example in forest ecosystems is the introduction and spread of tree pathogens such as insects or fungi. Newly introduced pathogens are capable of radically reducing the abundance of susceptible trees, with long-lasting consequences, yet there is no direct human experience of such events and no evidence of the influence of such outbreaks in the past. Temperatures maintain the current distribution limits of many of these insect and disease pathogens, so if temperatures increase, then movement of new pathogens into the northern boreal forest is highly likely. Temperature increases in the boreal region can also increase the risk of freezing damage. Boreal forest vegetation is generally well adapted to survive or recover from brief periods of below-freezing temperatures after the initiation of growth in the spring. However, some of these freeze-protection mechanisms provide protection from temperatures only slightly below freezing or are not fully effective at the earliest stage of growth. Earlier initiation of growth in the spring or even in the late winter in the boreal forest as the result of increasing temperatures is very likely to be followed by a return to seasonal cold temperatures well below freezing, resulting in damage that would not occur with later emergence from winter dormancy (Kellomaki et al., 1995, Prozherina et al., 2003).

The upper levels of temperature increases projected by the five ACIA-designated models are within the range of climate change that is likely to include "ecosystem surprises" for which no historic analogue exists and which ecological modeling is not likely to project.

However, humans are not passive spectators of change in forest ecosystems or forest resources that are important to them. Humans actively intervene, plan, and

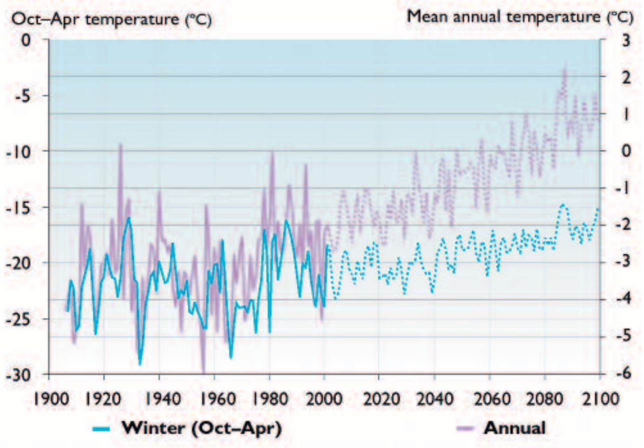

Fig. 14.9. Observed (solid line) and projected (CGCM2; broken line) winter and mean annual temperature at Fairbanks, Alaska. Observed data are a combination of University Experiment Station (1906–1948) and Fairbanks Airport (1948–2001) measurements (data from NOAA, 2002).

manage forest resources and agents of change in forests through their influence on factors such as fire that affect forests and forest values. Humans completely plan and create agricultural crops, livestock herds, and marketing systems. Therefore, it is possible for climate change to initially create effects that humans can respond to in a way that makes humans the most important agent of change. In some ways, these management responses are part of the standard set of land and resource management treatments that forestry and agriculture have developed over time. Reforestation, new species plantations, salvage of dead trees, sanitation treatments of a stand to reduce its vulnerability to pathogens, planting new crops, and altering soil or site conditions have a long history. Crops can be moved to suitable temperature zones, fields irrigated, and tillage systems adjusted to manipulate soil temperature within certain limits. What is new, and only now being developed, is the application of these techniques and the development of additional techniques to achieve specific goals associated with climate change in the far north, including both adaptation and mitigation approaches, and some appreciation of the associated costs.

14.4.6. Differences between the B2 and A2 emissions scenarios

For all of the model analysis presented in this chapter, projections based on the B2 emissions scenario (section 4.4.1) were used. This section evaluates the possible differences between model projections based on the A2 emissions scenario and those based on the B2 emissions scenario.

Broadly speaking, all the ACIA-designated models forced with either emissions scenario project that temperatures within the boreal region will reach levels higher than shown by reconstructions of climate for nearly the last 1000 years or essentially since the Medieval Warm Period. In every case, the models forced with the A2 emissions scenario project higher temperatures at the end of the 21st century than they project when forced with the B2 emissions scenario (section 4.4.2). The temperature increase over the 21st century projected by the model with the least warming when forced with the B2 emissions scenario (CSM_1.4) is about 3.5 °C, but the same model when forced with the A2 emissions scenario projects a temperature increase of about 4.5 °C over the same period. The greatest temperature increase over the 21st century – slightly more than 7 °C – is projected by the ECHAM4/OPYC3 model forced with the A2 emissions scenario.

Considered from the broadest perspective, the difference between projections based on the A2 emissions scenario versus those based on the B2 emissions scenario is largely a matter of timing of effects on forests rather than different kinds of effects. In practical terms, this means that the thresholds described in this chapter would be reached sooner under climate conditions projected using the A2 emissions scenario compared to conditions projected using the B2 emissions scenario. The difference in timing of these effects is projected to be as much as 40 years earlier in the case of the projections based on the A2 emissions scenario. The temperature projections of each model when forced with the A2 emissions scenario begin to diverge from projections using the B2 emissions scenario in amounts significant for forest processes after about 2040.

In general, the projections of the ECHAM4/OPYC3 model forced with the B2 emissions scenario best matched the features of annual temperature variability in the recorded data for the stations examined in this chapter; most of the other four scenarios produced less annual variability. The ECHAM4/OPYC3 model forced with the A2 scenario projects a pattern of steep temperature rise early in the 21st century, then a period of little change, and finally a steep rise again in the last decades of the century. This suggests that the effects of climate change under this scenario, with large variability possible from year to year, could be especially sudden. Under such a scenario the risk of extreme events, such as widespread forest fires and insect outbreaks, would probably be greatest.

14.5. Agriculture

14.5.1. Arctic agriculture in a global context

The population of the earth relies on two basic agricultural systems for its food supply. The first is the *large-scale commercial production* and trade of commodities. The commodities that dominate this system are cattle, sheep, and hogs for meat or wool, dairy animals for milk products, grains for animal feed and direct human consumption, and oilseeds. These products move through a complex, and in most cases efficient, network of trading, transportation, and processing before reaching the final consumer. Smaller volumes of products that are important for human dietary variety also move through this network, including potatoes, fruits and vegetables, and raw products for beverages and spices. As a result, a diverse array of foods is available to those consumers who have both physical and economic access to the network. Within the Arctic, this agricultural and trading system serves major population centers, areas served by regular surface transportation, and remote communities. The potential impacts of climate change on this system are global and complex: the system is vulnerable to direct climate influences on crops in multiple climate zones across the earth, to direct climate effects on shipping activities, and to subtle climate influences on prices and relative price differences (Rosenzweig and Parry, 1994).

The second type of agricultural system that supplies food for the population of the world is *subsistence* agricultural production. This system can be generally characterized as regional and largely self-reliant. It does not rely on a complex infrastructure to move either raw or manufactured products to the final consumer. Products are characteristic of the production region, and often obtained as part of a mixture of annual activities (section 12.2.2). These regions are categorized as those having low population density and/or a low economic profile. This subsistence agricultural system serves the majority of the area of the Arctic, but a minority of its population. Generally, regions in which subsistence agriculture predominates have either limited or no physical and/or economic access to the primary global food supply network. Potential climate change impacts on the subsistence system may be acute and local (e.g., a mid-season frost kills a crop) or subtle and longer term (e.g., climate effects on price advantages or barter opportunities).

In the Arctic, many examples can be found of a mixture of the two food systems or a cultural preference for the subsistence agricultural system despite economic and infrastructure advantages that would permit participation in the commercial production system. People in the Arctic can and often do move between participation in one system and the other. It is not clear that the commercial production system will completely displace the subsistence agricultural system in the foreseeable future.

14.5.2. Existing agriculture in the Arctic

Agriculture is a relatively small industry in high-latitude regions and consists mostly of cropping cool-season forage crops, cool-season vegetables, and small grains; raising traditional livestock (cattle, sheep, goats, pigs, poultry); and herding reindeer. This chapter focuses on crop production rather than livestock. Production of inexpensive reliable feeds is often a major constraint to animal agriculture and livestock production is often limited more by non-climate factors such as availability of processing facilities and appropriate waste handling rather than directly by climate. While agriculture is limited by climate in the Arctic, especially in the colder regions, it is also limited by lack of infrastructure, a small population base, remoteness from markets, and land ownership issues. Major climate limitations include short growing seasons (not enough time to mature or to produce high yields of harvestable crop), lack of heat energy (too few GDDs during the season), long and/or unfavorable winter weather that can limit survival of many perennial crops, and high moisture stress in some areas.

Most of the climate stations selected for examination have climates that are currently at least marginally suitable for growing barley, oats, green peas, and potatoes (although not always to full tuber maturity). Wheat has been grown in experimental plantings as far north as Rovaniemi, Finland, the southern Northwest Territories, Canada, and Fairbanks, Alaska, but is considered too marginal for commercial production in those areas and is a major commercial crop only in southern Siberia (represented by Novosibirsk). Canola (rapeseed) is also produced only in the southernmost part of the Arctic and is considered marginal even in many areas now considered suitable for barley (such as Interior Alaska). Sunflowers currently are produced in generally the same locations as wheat. The magnitude of temperature increases projected by the ACIA-designated models would remove these heat limitations in areas that are presently climatically marginal by the mid- or late 21st century.

All of the climate stations examined in this chapter have summers with enough heat to produce at least one successful harvest of forage crops, including legumes such as alfalfa or clover and grass such as timothy. Slight increases in growing-season temperature are likely to increase the probability of two or more successful harvests and thus increased yields in some areas. However, winter survival is a problem in most areas, thus alfalfa is produced only in the warmer parts of the Arctic, and clovers are successfully grown in areas with relatively mild winters. Perennial grasses, especially smooth bromegrass and timothy, are grown in much of the region.

Agricultural production statistics match the boundaries of the Arctic as defined in this chapter at the national level for Iceland. Relevant statistics are also available for the three northern counties of Norway (Troms, Finnmark, and Norland), Alaska (United States), and the Yukon and Northwest Territories (Canada). These areas are highlighted in this section. Data from the Norrbottens region of Sweden match this chapter's analysis area, and Finnish agricultural statistics are readily available at the national level and for jurisdictions with boundaries that are not as relevant for the purposes of this assessment. Complete and reliable statistics for agriculture in arctic Russia are not available.

Yukon Territory is one of the most mountainous parts of Canada, but agriculture occurs in valleys, especially in the southern interior, and in suitable landforms in the Northwest Territories. The data in this section on Canadian agriculture are from Statistics Canada (2001) and Hill et al. (2002). There are at least 170 farms in the Yukon and 30 in the Northwest Territories, representing a total capital value of over Can$ 67 million, primarily in land and buildings. Most of these farms are operated as part of a yearly cycle of other activities, and only 21 have gross receipts greater than Can$ 25000.

Farmers in the Canadian north take a diversified approach to agriculture. They produce traditional livestock and engage in sled-dog breeding and non-traditional livestock production primarily for local and specialty markets. Along with east-central Alaska, the western Yukon is the driest part of the North American subarctic. Because portions of these regions are already natural grasslands too dry for tree growth, irrigation would certainly be required to exploit improved temperature regimes for crop agriculture in such areas. In the central and eastern Canadian Arctic, temperatures currently fall well below minimums necessary for traditional agriculture, although warming trends have been detected. Across much of the Canadian far north, further temperature increases as projected by the ACIA-designated models would move the zone of temperature suitability northward into a region largely dominated by soils with relatively poor suitability for agriculture. Broad areas of Turbic and Organic Cryosols cover these northern landscapes (Lacelle et al., 1998), and in the glaciated north, soils are often thin, stony, and derived from nutrient-poor bedrock parent materials. Areas with pockets of suitable climates and soils that are close to population centers could possibly become locally important for food production for established populations with a goal of greater self-sufficiency.

Alaska has several climatically distinct agricultural regions, including interior valleys and lowlands with warm and dry summer climates, coastal regions with cooler summers and moderate to high precipitation, and a broad transition region in the major population zone of south-central Alaska. Most of the volume and value of agricultural production takes place in the transition and interior locations. In the dry interior valleys, heat sums necessary to mature grains such as barley are generally achieved, and recent warming trends have increased the probability of achieving this critical factor to a very high level (Sharratt, 1992, 1999). In interior valleys and the south-central transition region, there is a high correlation between years of excellent crop production and adequate and (especially) well-timed precipitation in the early and middle parts of the growing season. However, persistent and heavy precipitation in the late summer can delay, reduce, or even ruin crop harvest and recovery.

In 2001, Alaska had about 580 farms and ranches (sales greater than US$ 1000) covering 370000 ha, of which the great majority was unimproved pasture (Alaska Agricultural Statistics Service, 2002). Alaskan crop production in the same year included 10500 t of potatoes (including high-value, certified virus-free seed potatoes), 27000 t of hay, 800 t of carrots, and over 8800 m^3 of oats and barley. Alaska produced 11500 cattle, 1200 sheep, 1000 hogs and 15000 domestic reindeer in 2001. A few farm and ranch operations are producing species such as elk (wapiti), bison, yak, and musk oxen. Some Alaskan crops and pastures are already moisture-limited, especially in the warmer summer climate of recent years (Sharratt, 1994). If temperatures were to increase across the northern boreal region without a significant increase in precipitation in the early and middle parts of the growing season, the change would be unfavorable for agricultural production, except in the case of irrigated crop production that could take advantage of the greater growing-season heat units. Temperature increases projected for coastal areas generally are very likely to be favorable to agriculture, although much of the area consists of steeply mountainous terrain with little soil development. In the south-central transition region, Alaskan agriculture faces strong economic challenges from competing urban and suburban land uses in the region where historically it has been most well-established. Large new areas are available to be opened to agriculture, especially in central Alaska, but no recent public policy initiative to do so has emerged. National-level agricultural subsidy payments to Alaskan farmers have nearly doubled in the last decade to over US$ 2 million, or about 20% of net farm income.

In 2000, northern Finland contained about 10000 farms, of which about 50% were dairy farms. The average farm size is about 11.5 ha. While most northern Finnish farm properties remain in their traditional size and configuration (they have not been subdivided), the number that are producing significant amounts of dairy products has decreased rapidly in recent years (Häkkilä, 2002). Reasons for the decline include surpluses of dairy products, small quantities produced, and especially social change. These changes include migration of farm owners to towns and cities with professional opportunities that are more attractive and have higher income potential, and the conversion of inactive farms into country residences for town dwellers.

The three northern counties in Norway had 95000 ha of land in agricultural use in 1999, distributed among about 6500 owners (Statistics Norway, no date). The area receives nearly the maximum warming influence of the Gulf Stream in the Arctic, so there is a strong maritime influence especially in the south and west of the region. Climate factors limiting to agriculture are maximum temperatures and total warmth over the summer,

Fig. 14.10. Garden in northern Norway (photo: P. Grabhorn, 2003, Grabhorn Studios, Cerrillos, New Mexico).

which are strongly influenced by cool and wet maritime weather. Prolonged cool and wet maritime conditions can result in slow plant growth and failure of crops to complete development, wet fields and crops at harvest time, and, occasionally, outbreaks of plant diseases. Agricultural operations are generally small, with only 570 holdings larger than 30 ha and only 50 holdings larger than 50 ha. Employment in agriculture, horticulture, and forestry in the three-county region in 1999 totaled about 6500 full-time equivalents. Many of these farming operations fit into a diversified set of annual activities that include wage employment, fishing, cultural landscape maintenance, and personal consumption and non-cash trade of crops (Fig. 14.10). Nearly all farming operations include cultivated meadows for hay and pasture, and about one-third produce potatoes. Only 45 farms, all in the southernmost county of Nordland, produce grain or oilseed crops.

Temperature increases are likely to move the grain production boundary northward. Any climate changes that enhance the maritime effect, especially in the summer, would not be favorable to agriculture in this coastal region, although the effects are very likely to be lessened inland. Climate changes that increase growing-season length and daily maximum temperatures while maintaining or slightly decreasing persistence of growing-season clouds and rain are very likely to be favorable to agricultural production in this area. Agriculture in the far north of Norway will be strongly influenced by national and European policies to subsidize rural populations and landscapes, perhaps more than any other factor. The government of Norway is likely to have sufficient revenues available from petroleum to intervene strongly in the economics of northern community infrastructure and agriculture to achieve its social and environmental goals for some time to come.

Iceland is surrounded during the growing season by a cool ocean surface; locations farthest inland from the cool maritime influence are covered with glacial ice. As a result, lack of summer warmth has been the chronic limitation to Icelandic agriculture, especially in years when sea ice reaches the Icelandic shoreline. Icelandic agriculture traditionally has been limited to the crops most tolerant of cool season conditions. Even small temperature increases historically have removed this limitation and permitted the better harvests in the record. Iceland produces over 2 million m^3 of hay, more than 10000 t of potatoes, and over 4000 t of cereal grains (barley; Statistics Iceland, 2003). Glasshouse agriculture using abundant geothermal heat supports the production of about 1000 t of tomatoes and cucumbers. In 2001, the 3000 farms in Iceland produced 473000 sheep, 70000 cattle, 73000 horses, and 106000 liters of processed milk (Statistics Iceland, 2003). During the last decade of the 20th century, sheep and hay production declined at double-digit rates, largely for non-climatic market, policy, and social reasons. Numbers of cattle and horses (the famous Icelandic pony) have remained level. Icelandic agriculture employs about 4400 people, which is 3.1% of the total employment. Greenland supports about 20000 sheep and more than 3000 reindeer (Statistics Greenland, 2005).

The entrance of Iceland into the European Community Common Agricultural Programme (CAP) has been the major force in Icelandic agriculture in the last decade. In general, CAP subsidies have sustained smaller and higher-cost producers and rural economies, introduced limitations for certain products, reduced risk to producers, and transferred funds from consumers to producers. Temperature increases would almost certainly permit accelerated expansion of cereal grain production and possibly new oilseed crops, reduce the cost of winter forage for livestock, and allow increased per-unit crop yields where nutrients or water are not limiting. There are complex multiple goals in Icelandic agriculture, including food production, employment, rural stability, maintaining an attractive landscape, and environmental protection. Because the agricultural system exists in a highly interventionist public policy environment, a traditional economic analysis is likely to show that national and CAP policies will be more influential than temperature increases in the future of Icelandic agriculture (see section 14.3.4.4 for a discussion of the conversion of agricultural land to forests).

14.5.3. Approach to scenario analysis

The 14 sites that were chosen for examination of the ACIA-designated model projections were selected to represent areas in the northern part of the boreal region currently supporting commercial agriculture and areas north of current potential (section 14.4.1). The analysis did not consider soil effects, although soil can be a major limitation to agriculture, and changes in soil processes following climate change can have significant impacts on agriculture (Gitay et al., 2001). In much of the circumpolar north, large tracts of soils already exist that are suitable for agriculture if other constraints are removed, with the exception of much of northern Canada (composed of the Canadian Shield) and northern Fennoscandia, where the granitic soils are shallow and nutrient-poor.

This analysis used GDDs as a primary determinant of climate suitability for producing annual crops. Growing degree-days (a measure of accumulated heat energy) are considered a good predictor of plant phenology and are often used to project the timing of different plant growth stages. For example, farmers and extension agents often use GDDs to determine the optimum harvest time for forage crops. There are some major limitations to the use of GDDs. They may vary for different cultivars, locations (Sanderson et al., 1994), and growth stages (Bourgeois et al., 2000; Kleemola, 1991). Heat energy requirements for plants interact with other environmental factors, such as moisture stress and photoperiod (Bootsma, 1984; Nuttonson, 1955, 1957), and thus GDDs may not be a good predictor of plant development under stress. Heat requirements for plant development generally decrease with increasing photoperiod, so fewer GDDs are needed to reach a given growth stage at high latitudes than at low latitudes.

In areas with humid autumns, GDDs may equal or exceed the threshold for a given crop to reach maturity, but wet conditions may not allow the crop to dry enough to allow cost-effective mechanical harvesting. Thus, even though projections of future temperature may indicate that certain areas will be suitable for some crops, wet conditions may offset the temperature effects, especially in more humid zones within the northern boreal region.

Growing degree-day requirements for various crops from the highest-latitude areas possible were obtained from the literature, and similar GDD requirements were assumed for a given crop throughout the region. One problem with using GDD data from the literature for determining climate suitability for a given crop is that different reports use different base temperatures for calculating GDDs. For example, many agriculturists often use a base of 5 °C (GDD_5) for cool-season crops, while others use a base of 0 °C (GDD_0). Others use some experimentally determined growth threshold as the base temperature. If no information existed for GDD requirements using a base of 5 or 0 °C, the GDD requirements for the base temperatures of interest were estimated from published GDDs based on similar base temperatures. In some cases, insufficient information was available to do this for both 5 and 0 °C base temperatures. Despite its limitations, use of GDDs is a widely reported and accepted method to project approximate crop phenology. GDD analysis is generally accepted as a good way to estimate climate suitability for producing annual crops.

For projecting potential water stress, the analysis used model-projected potential evapotranspiration minus growing-season precipitation. Positive values indicate a potential water deficit, interpreted to mean potential water stress, which could limit crop yields or quality. However, many crops can produce acceptable yields even under water-deficit conditions, and water stress is also affected by soil water-holding capacity, rooting depth, precipitation timing (e.g., spring versus summer), and type of storms (e.g., infrequent large rainfall events may result in more water loss due to runoff than small, frequent storms).

This assessment ignored the effects of atmospheric CO_2 enrichment and changes in crop pests, other than to estimate the potential general effects. Few field studies have been done on CO_2 enrichment effects on crop plants at high latitudes and results from more temperate zones may not be transferable to northern areas, thus there is little data on which to judge the effects of CO_2 enrichment. Climate effects on pests are complex and difficult to forecast. Such an analysis is obviously needed to understand the complete range of climate change effects, but it is not yet feasible because of a lack of detailed knowledge of insect and other pathogens.

For this analysis, a few annual crops were selected that are currently well-adapted or marginally adapted to parts of the region and which have potential to become economically important crops for animal feed (cereal grains), human food, (beans, peas, potatoes, wheat), or oilseeds (canola, sunflowers).

14.5.4. Climate limitations and influences

If the ACIA-designated model projections of GDDs at a given site equaled or exceeded the minimum GDD requirements for a given crop, but were below the midpoint of the range given in Table 14.3, the crop was considered to be marginal. The analysis was based on 20-year averages, to smooth out much of the variability in the data and because it is assumed that farmers would require about 20 years under a changing climate regime to accept the change and adopt farming practices that included new crops.

The GDD requirements for some of the crops presented in Table 14.3 appear inconsistent between the two base temperatures (for example, with a base of 5 °C, seed peas appear to require fewer GDDs at the low end of the range than canola but require more GDDs at the upper end of the range). This may be due to errors in the literature, errors in the estimation of GDD requirements, or it may mean that either the 5 or 0 °C base temperature is not appropriate for some crops. This may cause errors in determining climate suitability for some crops. Table 14.3 also provides estimates of the number of accumulated GDDs required for forage crops to reach optimum time of harvest. These may be underestimates for high-latitude

Table 14.3. Growing degree-day requirements for various annual crops to reach maturity (Anon., 1996-2000; Ash et al., 1999; Dofing, 1992; Dofing and Knight, 1994; Juskiw et al., 2001; Miller et al., 2001; Nuttonson, 1955, 1957; Sharratt, 1999) and for forage crops to reach optimum harvest stage (Bootsma, 1984; Breazeale et al., 1999).

	Growing degree-days (5 °C base)	Growing degree-days (0 °C base)
Annual crops		
Peas (green for processing)	700–800	1000
Spring barley	700–900	1200–1500
Peas (for seed)	800–1150	1500–1700
Oats		1300–1700
Canola	950–1050	1350–1550
Potatoes	1000–1100	
Spring wheat	1000–1200	1400–1650
Dry beans	1100–1500	
Sunflowers (for seed)		1800–2000
Forage crops		
Alfalfa	350–450	
Red clover	450	
Timothy	350–450	

regions, as low soil temperatures may delay initiation of growth, especially in spring.

The models use large grid cells, which often contain large tracts of terrain not likely to become suitable for agriculture (such as high mountains), and some of the model projections are obviously not realistic even for present conditions. Therefore, this analysis uses proportional changes from the present, compared to observed weather data, rather than using actual temperature projections provided by the models. For each parameter analyzed (GDD_0, GDD_5, and growing-season water deficit), the highest and lowest projections for each location were selected (Table 14.4).

The success of perennial crops is governed more by winter survival success than by growing-season weather conditions. Factors such as the timing of killing frosts, warm spells during winter, length of the dormant season, snow depth, and winter temperatures all interact to affect winter survival. Warm periods during winter, especially prolonged temperatures above 0 °C, can be detrimental to perennial plants by reducing winter dormancy and depleting carbohydrate reserves (Crawford R., 1997). The complexity of these interactions makes projection of climate change effects on these crops difficult.

Table 14.4. Projected growing degree-days calculated from the highest and lowest model-projected temperatures for each site.

	2011–2030		2041–2060		2071–2090	
	Highest	Lowest	Highest	Lowest	Highest	Lowest
	Growing degree-days (0 °C base)					
Fairbanks	2175	1800	2525	1850	2625	2025
Big Delta	1925	1600	2225	1650	2350	1800
Whitehorse	1750	1500	2050	1525	2200	1700
Fort Vermillion	1875	1675	2000	1750	2050	1875
Thompson	2025	1825	2125	1900	2375	2050
Goose Bay	1550	1375	1600	1525	1775	1575
Reykjavik	1275	1125	1425	1050	1450	1150
Tromsø	1325	1100	1425	1125	1525	1300
Umeå	1700	1425	1875	1600	1900	1675
Rovaniemi	1650	1400	1825	1575	1850	1675
Archangelsk	1700	1450	1875	1600	2000	1625
Novosibirsk	2025	1900	2125	1850	2250	2000
Yakutsk	2025	1750	2150	1800	2325	1975
Magadan	1625	1400	1825	1425	2050	1775
	Growing degree-days (5 °C base)					
Fairbanks	1600	1125	2000	1150	2150	1325
Big Delta	1275	900	1600	875	1725	1050
Whitehorse	1275	900	1600	925	1725	1050
Fort Vermillion	1400	1200	1550	1275	1575	1400
Thompson	1350	1175	1450	1225	1650	1375
Goose Bay	1300	1100	1500	1275	1700	1350
Reykjavik	500	475	625	425	625	500
Tromsø	600	450	675	475	725	625
Umeå	1025	775	1200	975	1200	1075
Rovaniemi	900	700	1050	800	1050	875
Archangelsk	1000	825	1150	975	1250	1025
Novosibirsk	1325	1250	1425	1200	1500	1375
Yakutsk	1350	1075	1450	1100	1625	1325
Magadan	925	725	1075	625	1250	1050

14.5.5. Growing degree-day analysis

When GDD_0 was used for analysis, all of the ACIA-designated models projected that all the examined locations would be suitable for green pea production early in this century, and the high-extreme models (GFDL-R30_c and CGCM2) projected climates suitable for barley at all locations by 2030. When GDD_5 was used, the models projecting the most GDDs suggested that Reykjavik and Tromsø would be unsuitable for green peas by 2030 because of too many GDDs. The GFDL-R30_c projections indicate that potatoes could be grown in all locations except Reykjavik, Tromsø, Rovaniemi, and Magadan (marginal at Umeå and Archangelsk) by 2030, while the low-extreme models (HadCM3 and CSM_1.4) projected suitable climates only in Fairbanks, Fort Vermillion, Thompson, Goose Bay, Novosibirsk, and Yakutsk. Thus, projections for potatoes may be too low, as potatoes are already produced, at least marginally, in many areas not projected to be suitable by 2030. This may be because use of GDDs may not be a good way to project phenological development in potatoes (Shaykewich and Blatta, 2001). The models generally projected that all except the coolest sites in Scandinavia would have climates at least marginally suitable for potato production by the end of the century.

All of the models project a climate suitable for dry pea production at all locations except Reykjavik and Tromsø by 2030 when GDD_5 was used, although the low-extreme models (HadCM3 and CSM_1.4) projected that Goose Bay, all of the Scandinavian sites, Archangelsk, and Magadan would be unsuitable for dry peas using GDD_0. All the models project that all sites except Reykjavik and Tromsø would have climates suitable (marginally so at Goose Bay) for seed pea production by the end of the century. The high-extreme models (GFDL-R30_c and CGCM2) project that all of the North American sites (marginal at Big Delta, Whitehorse, and Goose Bay), Novosibirsk, and Yakutsk would have climates suitable for dry bean production by 2030, but the low-extreme models (HadCM3 and CSM_1.4) project that climate would only be marginal for such production even at the warmest sites (Fairbanks, Fort Vermillion, Thompson, Novosibirsk) by 2030. Models projecting the highest number of heat units at a given location indicated that all sites except Reykjavik, Tromsø, and Rovaniemi would be at least marginally suitable for dry beans by 2060, whereas the low extreme models (HadCM3 and CSM_1.4) indicated that only Fairbanks, Fort Vermillion, Thompson, Goose Bay, Novosibirsk, and Yakutsk would have climates suitable for dry bean production by 2090.

Most of the models projected that most locations would be at least marginally suitable for all of the cereal grains by the end of the century, although the low-extreme models, using GDD_5, projected that only Fairbanks, Fort Vermillion, Thompson, Goose Bay, Novosibirsk, and Yakutsk would be well-suited for wheat by 2090. The high-extreme model (GDFL-R30_c or CGCM2) for each location projected that climate would be suitable for canola at all locations except Tromsø by 2030 using GDD_0. All of the models projected that climate will be at least marginally suited for oilseed sunflowers by 2030 at Fairbanks, Thompson, and Novosibirsk, and the high-extreme models projected climates suitable for sunflowers at all sites except Goose Bay, Reykjavik, and Tromsø (marginal at Rovaniemi) by 2090. Conversely, the low-extreme models projected that only Fairbanks, Thompson, Novosibirsk, and Yakutsk would be well-suited, and Big Delta and Fort Vermillion would be marginally suited, for this crop by near the end of the century.

Assuming that the amount of heat is the main factor limiting yields and number of harvests of perennial forage crops, model projections indicate that warmer growing seasons are likely to increase the potential number of harvests and hence seasonal yields for perennial forage crops at all locations. Uncertainty about winter conditions make forecasts about survival potential for crops difficult. Warmer winters could actually decrease survival of some perennial crops if winter thaws followed by cold weather become more frequent. Crops adapted to certain types of winter stress are very likely to experience different types of stresses if winter temperatures rise and winter thaws become more frequent. However, lengthened growing seasons, especially in autumn, are very likely to result in northward extension of climate suitable for alfalfa production.

Table 14.5. Projected water deficits calculated from highest and lowest model projections at each site. Negative values indicate a water surplus.

	Water deficit (mm)					
	2011–2030		2041–2060		2071–2090	
	Highest	Lowest	Highest	Lowest	Highest	Lowest
Fairbanks	250	166	282	207	311	253
Whitehorse	150	116	287	165	321	265
Fort Vermillion	324	148	355	168	371	218
Thompson	335	132	394	157	413	169
Goose Bay	22	-122	62	-151	58	-125
Reykjavik	184	-20	216	-15	241	68
Tromsø	84	-31	117	14	134	44
Umeå	217	139	250	163	262	173
Rovaniemi	185	51	233	111	226	145
Archangelsk	191	-7	232	26	231	31
Novosibirsk	389	126	414	127	469	149
Yakutsk	289	152	315	161	328	165
Magadan	224	117	238	164	269	180

All of the models project high interannual variability in GDD accumulation. Thus, even though 20-year averages may indicate climates well suited for crops for a given area, the frequency of growing seasons cool enough to limit production may convince farmers that the risk for producing these crops is too high. Thus, the perception of risk by farmers could be a limiting factor for the northward advancement of many of the analyzed crops.

14.5.6. Precipitation and potential evapotranspiration analysis

Most of the models project increases in precipitation at most locations, with some forecasting up to 20% increases from values typical of the 1981–2000 baseline period by 2090. There was a great deal of variability among models, with no single model consistently projecting the greatest or smallest change in precipitation at all locations. A few models projected decreases in precipitation at a few sites. For example, CSM_1.4 projects a decrease of 5% at Whitehorse (compared to the present) during the 2011–2030 time slice, and HadCM3 projects a decrease of about 1% at Umeå and about 4% at Tromsø by 2090. All models projected a general increase in potential evapotranspiration over time at all locations, with much variability among models and locations.

Some locations in the subarctic already experience water deficits large enough to reduce crop yields. For example, Sharratt (1994) found that barley yields in Interior Alaska are depressed in about five of every nine years due to insufficient water supply. Water deficits were usually projected to increase over time, with GFDL-R30_c usually projecting the smallest deficits and HadCM3 and CSM_1.4 usually projecting the largest (Table 14.5). Most models project rather large water deficits, especially by the end of the analysis period. However, the HadCM3 model projects rather large water surpluses throughout the 21st century at Goose Bay, and slight surpluses at Reykjavik until about mid-century (Table 14.5). Most models, especially CSM_1.4, project fairly small deficits at Tromsø. These results indicate that water is likely to become a major limiting factor for production of most crops at all but the maritime sites in the boreal zone. Unless irrigation is supplied, which may not be economical for most of the crops analyzed, production is likely to be limited in many areas to drought resistant crops, such as cereal grains.

14.5.7. Indirect effects of climate change

This analysis has focused on the direct effects of changes in growing-season temperatures (using GDDs as indicators of crop growth potential) and moisture relationships on potential agricultural crop production in the next century in high-latitude regions. However, the indirect effects of climate change may have similar or greater impacts on agricultural development. Several of them are outlined here without in-depth analysis, to acknowledge their potential importance and alert those involved in future assessments that these factors need to be considered.

The development of crop disease requires suitable host, pathogen, and environmental conditions. Temperature and moisture are critical for the spread of many plant diseases (Gitay et al., 2001), especially in a region as severely heat-limited as the Arctic. Under conditions of increasing temperatures, the risk of crop damage increases in all regions of the northern half of Europe (Beniston et al., 1998). The severity and number of species capable of reaching infestation levels for North American agriculture are likely to increase with less severe winters (Shriner et al., 1998). These findings suggest that disease and weed pests are likely to increase throughout the Arctic under the ACIA-designated model scenarios. These problems are not likely to offset potential yield increases or eliminate the potential for new crops in most cases. However, it is possible that severe outbreaks could have that effect in specific cases. For example, temperature increases in Finland are very likely to increase the incidence of potato late blight to the point that it will significantly decrease potato yield in that country (Carter et al., 1996).

While the effects of atmospheric CO_2 enrichment were not considered in this analysis, higher CO_2 concentrations are very likely to result in yield increases for most crops (Warrick, 1988). Lack of experimental data on the effects of CO_2 fertilization on crops in high-latitude regions make in-depth analysis of this effect impossible at present.

Lack of infrastructure development, small population sizes (thus limiting local markets), and distance to large markets are likely to continue to be major factors limiting agricultural development in most of the northern part of the boreal region during this century. One possible indirect effect of climate change on agriculture in the Arctic and subarctic is the effect on transportation, including land transport and the prospect of regular ocean shipping across the Arctic Ocean with reduced or absent sea-ice cover (section 16.3). Thawing of permafrost that is susceptible to ground subsidence is disruptive to existing roads and railroads. Ultimately, once the permafrost thaws, in nearly all cases a more stable foundation will be available for permanent transportation routes and facilities. It is possible that shipping across the Arctic Ocean will greatly enhance economical trade in agricultural products, especially products unique to these regions, and thus enhance development of commercial agriculture in the far north.

If the net effect of climate change is greater global agricultural production, then food prices are likely to be lower for the world as a whole as long as the rather large subsidies that influence prices do not markedly increase or decrease. Lower world food prices would reduce the incentive to rely on local commercial or subsistence production in the Arctic. The opposite would be the case if climate change reduces global agricultural production.

A key component of the ability of global food production to cause a fundamental change in the supply/demand relationship for local agricultural production in the Arctic (other factors being equal) is shipping cost. Regular shipping across an ice-free Arctic Ocean (section 16.3.7) is very likely to lower transport costs for non-perishable bulk commodities such as grains or fertilizers, and thereby lower prices for the basic inputs to the agricultural system in the Arctic. In addition, it is possible that the availability of arctic shipping will stimulate the export of bulk finished products from the Arctic.
The volume of agricultural imports to or exports from the Arctic is unlikely to generate the economies of scale that would support regular shipping at a major cost advantage for agriculture within the Arctic, at least not within the 21st century. Mining and petroleum industries would be more likely to generate higher volumes of cargo, although there is only a limited ability to change the kind of cargo handled by specialized vessels. However, if arctic ports of call were integrated into a much higher volume of general trade between the northern continents, then price advantages for agriculturally related arctic cargoes would be much more likely.

The impacts of lower transport cost on arctic commercial agricultural production would be complex. Inexpensive, regular arctic shipping is very likely to provide a price advantage that could possibly stimulate production of unique arctic commodities such as reindeer and caribou meat or muskox milk or cheese, as long as other factors affecting demand were favorable. Conversely, lower shipping costs are likely to discourage arctic production of agricultural commodities widely produced elsewhere, unless overall global supply or demand structure shifted descisively. However, basic input data that would permit in-depth analysis, including trade levels and patterns in an ice-free Arctic, economics and practicalities of Arctic Ocean shipping operations, and the overall regional costs of arctic agricultural production compared to global supply and demand, are somewhat speculative or lacking.

In Europe, the United States, and Canada, public policy and interventions are major factors that agricultural producers confront in determining their activities. In many cases, national or trading-area agricultural transfer payments and rules of production correlate especially well with the annual profile of agriculture in the Arctic, as noted in section 14.5.2. There is no obvious trend to suggest that the significant role played by government policy will change during the 21st century. Russia is experiencing a period of population and economic consolidation in its northern regions, an adjustment required after investments made during Soviet times could not be sustained under new conditions. Given the strong dominance of state land ownership, government policy will be a decisive factor in the future of agriculture in most of the arctic nations. The current agricultural profiles also suggest that local markets are vital to the agricultural sector in the Arctic. If climate change results in increases in the human population of the

Arctic, then the entire agricultural sector is likely to expand given the historic positive relationship between regional population and agricultural production.

14.6. Tree rings and past climate

The most recent historical period of Northern Hemisphere warming of similar magnitude (but possibly different in its cause) to that of recent decades is the Medieval Warm Period (MWP) from about AD 900 to 1300 (see also section 2.7.5). However, interpretations of the climate during the MWP do not always agree. Some evidence suggests that the MWP was not general across the planet and did not exceed the current warming in its fluctuations (Hughes and Diaz, 1994; Mann M. et al., 1998). Other evidence suggests temperatures 1 to 1.5 °C higher than at present across the Northern Hemisphere during the MWP (Dahl-Jensen et al., 1998; Esper et al., 2002; Naurzbaev and Vaganov, 2000). The main tool to compare recent warming with temperature levels during the MWP and earlier periods in most terrestrial regions of the far north is long-term climate reconstructions based on tree-ring data, because few or no other historical records exist and marine proxies and ice cores cannot provide the required geographic coverage or detail.

Tree-ring chronologies serve as a useful basis for reconstructing natural temperature fluctuations in the high latitudes over millennial intervals, although the degree of reliability needs to be assessed carefully in each application. An important potential limitation is that the relationship between tree rings and climate may vary with time. However, compared to other indirect sources of climatic information, tree-ring chronologies have certain important advantages. First, tree rings record a complete annual sequence of climatic information. Second, in northern Eurasia, where trees reach a maximum age of 1100 years, there is a dense dendroclimatic network allowing spatially detailed quantitative temperature reconstruction for the last 500 to 600 years, and in some regions for more than two millennia (Briffa et al., 2001; Hughes et al., 1999; Vaganov et al., 1996, 2000).

The rate and magnitude of recent Northern Hemisphere temperature increases are unique within the last several centuries (Briffa et al., 1996, 2001; Mann M. et al., 1998). Some climate models that include anthropogenic effects calculate that the greatest temperature increase, in the range of 3 to 4 °C, should have occurred over the last several centuries in the high latitudes of the Northern Hemisphere (Budyko and Izrael, 1987; Kondrat'ev, 2002). However, temperature reconstructions (generally of the warm season or even a specific portion of the warm season) based on tree-ring chronologies from subarctic Eurasia, a region that makes up a large part of the projected zone of maximum warming, do not show temperature increases of the projected magnitude (Briffa et al., 1998; Naurzbaev and Vaganov, 2000). This may be partly due to the disproportionate influence of warm-season temperatures on tree growth, genuine local spatial climate variability, issues in calibrating the tree rings to estimate temperature with uniform reliability throughout the whole period of analysis, or errors or missing factors in climate model scenarios.

14.6.1. Past climate change in central Eurasia

This section examines several aspects of high-resolution proxy records based on tree-ring chronologies. Subarctic temperature reconstructions for Asia are compared with the main climatic forcing mechanisms during the last several centuries; reconstructed temperature fluctuations during the MWP are compared with recent temperature changes in the high latitudes of Eurasia using the millennial tree-ring chronologies; and recent temperature fluctuations are compared to temperature fluctuations during most of the Holocene in order to reveal warmest and coolest periods as well as rapid temperature changes in the central Asian sector of the subarctic.

The three main sources of information used for this analysis include local chronologies from the central Asian subarctic dendroclimatic network (Fig. 14.11) based on the analysis of living old trees (Briffa et al., 1998; Vaganov et al., 1996); super-long (at least two millennia) tree-ring chronologies constructed from cross-dating the abundant dead wood material of northern Siberia; and subfossil wood material excavated from alluvial deposits in terraces of small rivers in the Taymir and Lower Indigirka regions and even from sites north of the modern tree limit.

The radiocarbon dates of subfossil wood were used to define preliminary calendar time intervals and then the cross-dating method was used to identify absolute dates of tree-ring formation. Unfortunately, not enough material was available to build an absolute chronology for the entire Holocene in Taymir, so the "floating" chronologies (chronologies that cannot be tied with certainty to absolute calendar dates) with numerous radiocarbon dates were also used to analyze past temperature deviations (Naurzbaev et al., 2001).

A climatic signal was derived from raw tree-ring measurements using the regional (age) curve standardization

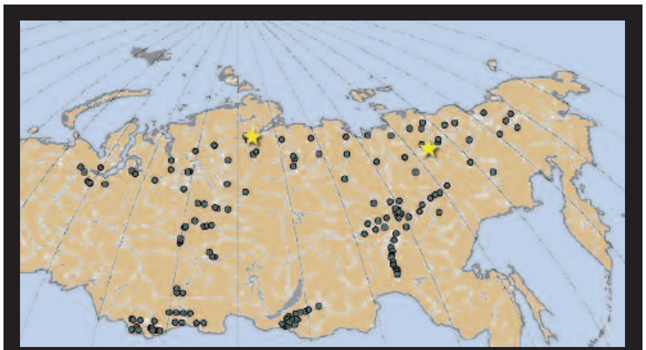

Fig. 14.11. Sites in the dendroclimatic network of the Asian subarctic (circles) and locations of millennial-length chronologies (stars) (Vaganov et al., 1996).

(RCS) approach (Briffa et al., 1996; Esper et al., 2002). This approach is applied to remove the age-dependent variations from single tree-ring series and to retain low-frequency climatic deviations (positive or negative trends) as well as high-frequency variations (year-to-year change) when averaging tree-ring index series. More details about this method of standardization can be found in Briffa et al. (2001), Esper et al. (2002), and Naurzbaev and Vaganov (2000). To verify results, tree-ring chronologies were compared to the instrumental climatic data averaged over a large sector of the subarctic. Finally, the longer-term tree-ring temperature reconstructions were compared to other proxy data including long-term variations in solar radiation (Overpeck et al., 1997), long-term variations in volcanic activity derived from ice-core measurements (Zielinski et al., 1994), and variations in CO_2 concentrations in air trapped in the GISP2 ice core (central Greenland; Wahlen et al., 1991).

14.6.1.1. Climate change in the central Asian subarctic during the last 400 years

Local tree-ring width series for each of 11 sites were obtained by averaging standardized series of individual trees. Between 60 and 70% of the variation in tree-ring width indices in the Eurasian subarctic is caused by changes in summer temperature (Briffa et al., 1998; Vaganov et al., 1996, 1998). A high correlation of local chronologies with temperature (r=0.69 to 0.84) allows the use of simple regression equations for reconstructing temperature based on tree rings. In effect, tree-ring width is used as a simple predictor of summer temperature. In order to avoid an additional procedure of statistical transformation of local tree-ring index series into normalized values of summer temperature variation, each of the local chronologies was normalized to the mean-square deviation in the generalized curve for the Asian subarctic region (Vaganov et al., 1996). To highlight trends and reduce short-term variability, the transformed local series was smoothed with a five-year running mean. Long-term temperature changes dominate the variability of the resulting generalized series. The generalized curve was compared with other temperature reconstructions for the circumpolar Northern Hemisphere (Overpeck et al., 1997) as well as with the main climatic forcing mechanisms.

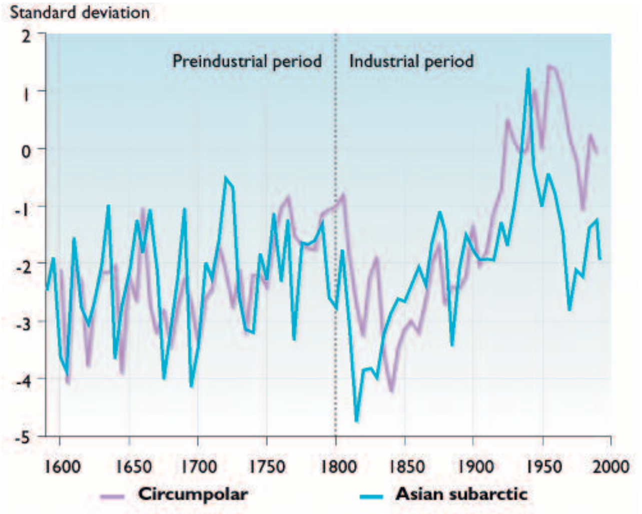

Fig. 14.12. Temperature variations from proxy records for the circumpolar Northern Hemisphere (data from Overpeck et al., 1997) and the Asian subarctic (see Fig. 14.11 for locations of sites).

Temperature variations in the Asian subarctic over the past 400 years correspond well to those observed across the circumpolar north (Fig. 14.12). Both curves clearly illustrate the temperature rise from the beginning of the 19th century to the middle of the 20th century. The main discrepancies between the two curves occur in the second decade of the 19th century and after the 1950s. The correlation of the two reconstructed temperature curves with each other for the preindustrial period is significant, but low (r=0.38, $p<0.05$), and markedly increases (r=0.65, $p<0.001$) for the industrial period (1800–1990) due to a distinct temperature increase that is shown in both curves.

More interesting is the correlation analysis of both curves (Fig. 14.12) with the main climatic forcing factors. The Asian subarctic generalized curve shows a significant correlation with all main climatic forcing factors: with solar radiation (r=0.32 for the entire period and 0.68 for the industrial period from 1800 to 1990); with volcanic activity (r=-0.41 for the entire period and -0.59 for the industrial period); and with atmospheric CO_2 concentration (r=0.65 for the period since 1850). The circumpolar reconstructed temperature curve (Fig. 14.12) is weakly correlated with solar radiation and atmospheric CO_2 concentration, and is not significantly correlated with volcanic activity. This is because more homogenous proxy data (only the tree-ring chronologies) were used for the Asian subarctic reconstruction, while different sources of proxy records (i.e. tree rings, lake sediments, isotopes in ice cores) were used in the circumpolar curve. The correlations with climatic forcing factors further indicate that natural factors (solar irradiance and volcanic activity) explain more of temperature variability that is common to the Asian subarctic and the circumpolar north than does the CO_2 concentration. Spatio-temporal analysis of reconstructed summer temperature variations in the Asian subarctic revealed that recent warming is characterized by an increased frequency of years with anomalously warm summers over the entire Siberian subarctic (the latitudinal dendroclimatic network seen in Fig. 14.11 that is replicated across a distance of 5000 km from east to west; Vaganov et al., 1996).

14.6.1.2. Medieval and current warming in northeastern Eurasia

Millennium length tree-ring chronologies were constructed from samples at two sites close to the northern treeline: east Taymir (71° 00' N, 102° 00' E) and northeast Yakutia (69° 24' N, 148° 25' E). These stands are made up of Gmelin larch (*Larix gmelinii*) and Cajander larch (*L. cajanderi*) in which the oldest living trees are up

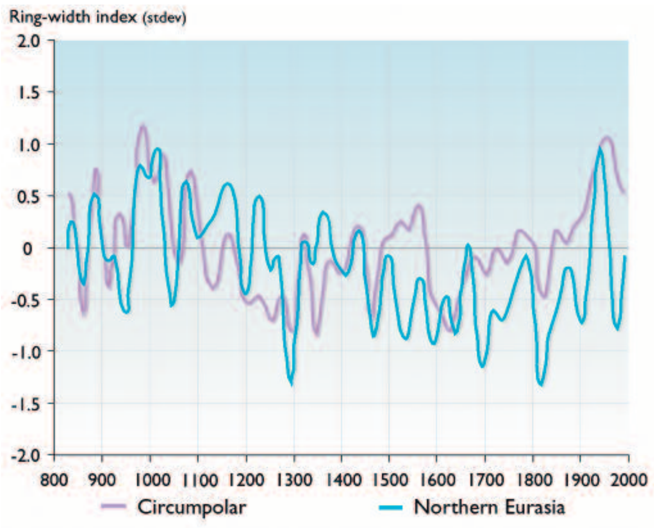

Fig. 14.13. Long-term changes in tree-ring growth in the circumpolar north (data from Esper et al., 2002) and northern Eurasia (combined chronology for east Taymir and northeast Yakutia).

to 1100 years old (Vaganov et al., 2000). Well-preserved dead tree trunks allowed extension of the record farther back than the maximum age of the living samples, allowing the construction of absolutely dated tree-ring chronologies for the periods from 431 BC to AD 1999 (Taymir) and from 359 BC to AD 1998 (Yakutia).

The RCS approach was applied to distinguish variation caused by climatic change. These representative tree samples were highly responsive to recorded temperature, allowing the reconstruction of temperatures over the last two millennia with annual resolution. With this long-term record, the instrumental record of 20th-century temperatures can be compared with temperatures during the MWP. Earlier results (Hughes et al., 1999; Vaganov et al., 1996) showed that tree-ring chronologies were highly correlated across distances up to 200 km (up to 500 km in northern regions), so these millennial chronologies represent temperature variations over a large sector of the Siberian subarctic.

The combined chronology for east Taymir and northeast Yakutia was compared with a generalized tree-ring chronology developed for the circumpolar region by Esper et al. (2002; Fig. 14.13). The two chronologies are significantly correlated during last 1200 years ($r=0.47$, $p<0.001$), suggesting that they represent long-term temperature trends in the entire Northern Hemisphere subarctic. Both curves show the warming characteristic of the MWP (10th to 13th centuries), decreasing temperature during the Little Ice Age (LIA, 14th to 19th centuries), and a 20th-century temperature increase.

The Siberian summer temperature reconstruction indicates that the warmest centuries were AD 1000 to 1200, with anomalies (from the mean over the last millennium) of 0.70 °C and 0.57 °C for those centuries, and the coolest centuries in were in the LIA period (AD 1600 to 1900), with anomalies for those centuries of -0.42 °C, -0.39 °C and -0.56 °C. The analysis of long-term trends of summer temperature leads to several conclusions. Present-day warming is estimated to represent an increase in summer temperature of approximately 0.6 °C above the coolest period of the LIA. The reconstruction clearly reveals the timing of the MWP, which occurs in the 10th to 13th centuries.

The reconstructed data indicate a greater warming (by about 1.3 °C) above the long-term mean during the MWP compared to the amount of cooling below the mean during the LIA. The results agree well with previous assessments of medieval warming in the Northern Hemisphere (Esper et al., 2002) but show less warming than that projected by climate models or historical analogues that use natural forcing factors such as solar variability and volcanic activity (Budyko and Izrael, 1987).

14.6.1.3. Climate change in the eastern Taymir Peninsula over the past 6000 years

The dendrochronological material discussed in this section was gathered from the Kheta-Khatanga plain region and the Moyero-Kotui plateau in the eastern Taymir Peninsula, near the northernmost present-day limit of tree growth in the world. The wood samples were collected from three areas: the modern treeline in the north forest of Ari-Mas; the modern altitudinal treeline (200–300 m above sea level) in the Kotui River valley; and alluvial deposits in terraces of large tributaries of the Khatanga River (one sample location is 170–180 km north of the modern treeline). The total number of wood samples exceeds 400. The RCS approach was used to standardize individual series (Naurzbaev and Vaganov, 2000). Approximate absolute dating (in contrast to the relative dating developed from the ring series) was established from radiocarbon dates of 45 samples of subfossil wood collected throughout the soil organic layer.

The result of this cross-dating is an absolute tree-ring chronology (see section 14.6.1.2) as well as a series of "floating" chronologies up to 1500 years in length, evenly spaced within the 6000-year interval of the mid- to late Holocene. The relative dating of the "floating" chronologies is based on the calibrated radiocarbon age of the samples (Stuiver and Reimer, 1993).

The resulting curve of the tree-ring index (relative growth) extends over 6000 years, and indicates favorable climatic conditions at about 6000 years BP.

This period of warmth represents the latter stages of the postglacial thermal maximum (Lamb, 1977a; section 2.7.4.2). The growth of larch trees at that time surpassed the average radial growth of trees during the last two millennia by 1.5 to 1.6 times. Tree growth (and temperature, accordingly) has generally decreased from the end of the postglacial thermal maximum through the end of the 20th century. Several samples of subfossil wood collected in the flood plain of the Balakhnya River were accurately dated in accordance with the "floating" chronology to the period from 4140 to 2700 BC. Due to the geography of the watershed, these samples could not have originated from the more southern regions, which means that during the postglacial thermal maximum the northern treeline was situated at least 150 km further north than at present. The postglacial thermal maximum can also be clearly identified by the relative levels of the stable isotope carbon-13 (^{13}C) in the annual tree rings (Naurzbaev et al., 2001). Increased ^{13}C concentration in annual layers of wood of this species (Cajander larch) is highly correlated with warm summer temperatures. High ^{13}C content is found in wood dated to the period that ring-width techniques reconstruct as warm, confirming high temperatures during this period.

Quantitative evaluation of mean deviations of average summer and annual temperatures demonstrates higher temperature variability during the postglacial thermal maximum compared to the 3.5 °C variability typical of the 20th-century instrumental temperature record. This corresponds well to the earlier published data (Naurzbaev and Vaganov, 2000) and to findings of subfossil wood in alluvial deposits of the Balakhnya River 150 km north of the present-day treeline. Reliable dates for the postglacial thermal maximum, in agreement with the "floating" chronologies, indicate that during this period sparse larch forest extended at least 1 to 1.5 degrees of latitude further north than the northernmost present-day forest limits in the Ari-Mas massif.

A comparison of the long-term temperature reconstruction for the Taymir Peninsula with other indicators of long-term temperature change in the high latitudes of the Northern Hemisphere during the Holocene (including summer melting on the Agassiz Ice Cap, northern Ellesmere Island, Canada; summer temperature anomalies estimated from the elevation of carbon-14 dated subfossil pine wood samples in the Scandes mountains, central Sweden; and temperature reconstruction from oxygen isotopes in calcite sampled along the growth axis of a stalagmite from a cave at Mo i Rana, northern Norway; Bradley, 2000) reveals several noteworthy features. During much of the Holocene, and especially starting between 9000 and 8000 years BP, the overall high-latitude Northern Hemisphere temperature steadily decreased, although there were shorter fluctuations with significant amplitude. This distinct trend of general temperature decrease agrees with the Taymir chronology. The concurrence of characteristic temperature fluctuations can be seen, for example, in significant decreases at about 6000 years BP and 4000 years BP, and increases at about 3000 years BP and 1000 years BP (the MWP). This coincidence suggests that some regions of the Arctic experienced long-term temperature changes in common with the high-latitude mean, which has been reconstructed using various proxy data. These interpretations of the data are supported, for instance, by the results of radiocarbon and dendrochronological dating of wood remains from the MWP collected north of the modern treeline in the Polar Urals (Shiyatov, 1993; section 14.11.1.3). However, some aspects of a reconstruction that infers higher summer temperatures during the postglacial thermal maximum than have been recorded in the late 20th century remain uncertain and subject to confirmation from additional research. The range in estimates obtained from different sources of the degree to which reconstructed temperature at its postglacial maximum exceeded the maximum warmth of the 20th or early 21st centuries is significant: from 0.6 °C (glacier and stalagmite layers and bottom deposits) to 3 to 3.5 °C (indicated by Taymir tree-ring chronology and the greatest extension of forest in the Scandinavian mountains; Kullman and Kjallgren, 2000). These deviations in reconstructed temperature may have been influenced by local conditions, different sensitivities of proxy sources to temperature change, or inadequate calibration models. Unfortunately, at present it is impossible to determine the cause of these deviations.

To detect the anthropogenic component of climate variations at high latitudes it is important to know whether temperature is *already* affected by increasing GHG concentrations and whether the rate of temperature increase is unprecedented in the period of instrument-based temperature records. A series of synthesizing studies, as well as simulations with GCMs, have established that anthropogenic emissions have had a significant influence on the rate of temperature rise in the Northern Hemisphere (Mann M. et al., 1998). However, in contrast to global trends, in the long-term northern Siberia tree-ring chronologies the present high-latitude summer temperature increase is less than that experienced during the postglacial thermal maximum. In this region, natural climatic forcing factors appear to have been more significant than the combination of human and natural factors producing the current summer warming thus far. In the Siberian study area, the amplitude of the current summer warming is not thus far greater than the warming during the MWP. This may be partly explained by the difference between regional and global trends, and partly by the difference between the trends in summer and annual temperature.

To determine with confidence whether the *rate* of temperature increase is unprecedented, it is necessary to obtain good quantitative data with high temporal resolution for the Holocene that will help to identify periods of drastic natural temperature increases in the past, and then to examine the amplitude of such drastic increases to see if there are natural limits during such periods of change. Therefore, one of the urgent tasks at present is the construction and analysis of super-long-term tree-ring chronologies for Eurasia with an adequate amount of subfossil wood. Such studies are being intensively conducted in Europe (Baillie, 2000; Leuschner and Delorme, 1988; Leuschner et al., 2000) and in north Asia (Hantemirov, 1999; Shiyatov, 1986; Vaganov et al., 1998). They should soon provide high-resolution tree-ring chronologies for Eurasia that can be used for quantitative reconstruction of temperature and for calibration of data obtained from other indirect sources of climatic information with lower temporal resolution.

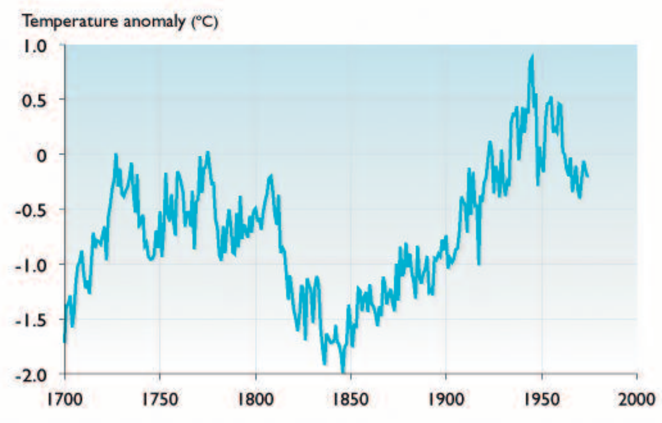

Fig. 14.14. Annual temperature anomaly (from 1671–1973 average) reconstructed from white spruce at the North American treeline (see D'Arrigo et al. (1996, 2001) and Jacoby and D'Arrigo (1989) for details of the reconstruction technique).

14.6.2. Past climate change in Alaska and Canada

One of the first large-scale climate reconstructions based on boreal tree-ring data was a study of treeline white spruce across northern North America (primarily Canada) covering about 90 degrees of longitude or about one-third of the circumpolar northern treeline extent (Jacoby and D'Arrigo, 1989). This chronology was based on the positive response of tree growth at treeline to temperature, and allows the reconstruction of mean annual temperature anomalies back to AD 1700 (Fig. 14.14). Key features of the reconstruction are intermediate temperatures during most of the 18th century, sharp cooling during the first half of the 19th century, gradual warming from the mid-19th century to a mid-20th century peak, and a slight cooling from about 1950 to the 1970s. If recent proxy data or climate records are available in a given locality to compare to the overall long-term record, unusual warming during the last decades of the 20th century is often noted (Jacoby and D'Arrigo, 1989; Jacoby et al., 1988). Some of the recent increase in annual temperatures in this reconstruction can be attributed to recovery from the last stages of the LIA.

Since the studies of the late 1980s, other analyses have added more spatial representation and longer temporal coverage (e.g., Mann M. et al., 1999) leading to an essentially complete coverage of the 20th century (Esper et al., 2002). The most recent reconstructed annual temperature curves confirm the major anomalies in annual temperature of the North American treeline curve of Jacoby and D'Arrigo (1989; see Fig. 14.14) that has served as the basis for standard Arctic temperature reconstructions (e.g., Overpeck et al. 1997).

The need to splice tree-ring records from overlapping generations of trees introduces some questions about whether the successively earlier generations of tree-ring records have been calibrated correctly and adequately preserve low-frequency variations (longer-term trends) in temperature. Methods of processing tree-ring data to preserve the low-frequency variations correctly have improved, allowing such trends and their possible causes to be identified (Esper et al., 2002). The Esper et al. (2002) reconstruction shows more prominent low-frequency trends, including the MWP and the LIA, than previous reconstructions or global or Northern Hemisphere-wide averages.

New technology, such as x-ray density (Jacoby et al., 1988) and stable isotope techniques, allow measurements of tree-ring properties in addition to ring width. Maximum latewood density of northern conifers increases when mid- to late growing season moisture stress is great (D'Arrigo et al., 1992). Maximum latewood density of boreal conifers also may represent an index of canopy growth where productivity is temperature-related, as indicated by satellite-sensed normalized difference of vegetation index (NDVI; representing "greenness" of the land surface) values (D'Arrigo et al., 2000). Carbon-13 isotope content is generally measured as "discrimination", which represents the difference in the amount of the isotope in sampled plant tissue compared to a reference standard. Less ^{13}C discrimination (greater ^{13}C content in sample) indicates production of the sampled plant tissue under a condition of restricted stomatal exchange, generally as a result of moisture stress (Livingston and Spittlehouse, 1996).

Maximum latewood density, ^{13}C isotope discrimination, and ring width of upland white spruce stands in central Alaska are well correlated with each other (Barber et al., 2000). Latewood density and ^{13}C isotope discrimination contain information specific to the climatic conditions of the year of ring formation, in contrast to ring width (which is influenced by two or more years of temperature), making them ideal for reconstructing past climates (Barber et al., 2000). No continuous instrument-based

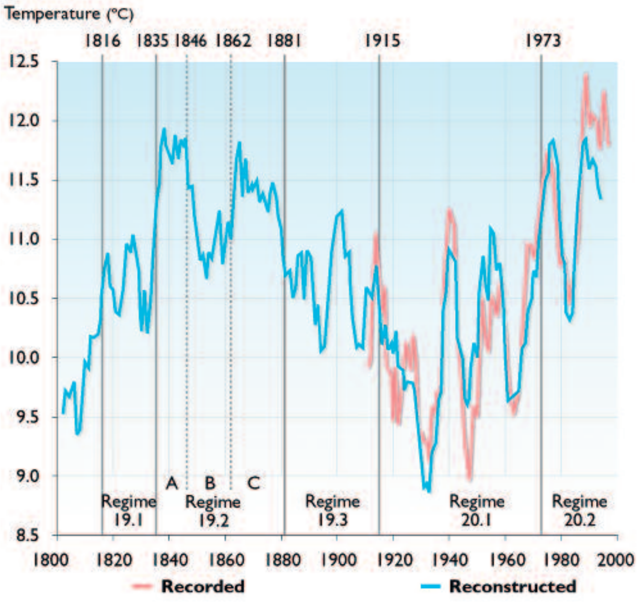

Fig. 14.15. Warm-season (Apr–Aug) temperature regimes and regime shifts in central Alaska from 1800 to 1996 from observations and reconstructed from tree-ring density and ^{13}C isotope discrimination (Barber et al., 2000; Juday et al., 2003).

temperature records exist for the western North American boreal region until the early years of the 20th century, and few records exist until the mid-20th century.

A 200-year reconstruction of warm-season (April to August) temperature at Fairbanks, Alaska, based on tree-ring density and ^{13}C isotope discrimination has been constructed (Barber et al., 2004; Juday et al., 2003; Fig. 14.15). The warm-season temperature reconstruction for Interior Alaska has been divided into multi-decade segments or warm-season temperature regimes. Regimes represent multi-decadal periods of characteristic temperatures that persist between periods of rapid climate change (Fig. 14.15). Note that the first half of the 20th century experienced extended periods of cool summers, which relieved moisture stress of low-elevation white spruce. The reconstruction of warm temperatures in the mid-19th century is out of phase with overall Northern Hemisphere means, but is strongly established by the proxies (^{13}C isotope discrimination and maximum latewood density) used in the reconstruction. A reconstruction of the annual Pacific Decadal Oscillation (section 2.2.2.2) index using western North American tree-ring records, accounting for up to 53% of the variance in instrumental records and extending back to 1700, also indicates that decadal-scale climatic shifts occurred in the northeast Pacific region prior to the period of instrumental record (D'Arrigo et al., 2001). These results suggest that rapid temperature shifts followed by semi-stable periods are a fundamental feature of climate change in that region.

From the perspective of two centuries, the recent very high rate of temperature increase in the second half of the 20th century in Interior Alaska is partially explained by a change from some of the lowest warm-season temperatures to some of the highest in the entire period (Fig. 14.15). Unlike the annual temperature reconstruction based on North America-wide treeline white spruce (Fig. 14.14), the Interior Alaska reconstruction (Fig. 14.15) indicates that the mid-19th century (Regimes 19.2A and 19.2C) was one of the warmest periods and the mid-20th century was a period of unusually cool summers. Because many of the climate records available in this part of the world begin only in the late 1940s or early 1950s (during the one of the coldest periods of the 20th century) and continue to the present (the warmest period of the last millennium), the instrument-based record indicates a higher rate of temperature increase than the longer-term reconstructions that incorporate several cycles of temperature increases and decreases. This suggests that the strong late 20th-century warming (during the warm season) in western North America may have a considerable component of natural climate variability in the signal.

There are still regions of the north that are not represented in large-scale temperature reconstructions, especially on the timescale of the past millennium. Recorded data and climate models strongly indicate that there are very likely to be important regional differences in temperature trends across the high-latitude north. In fact, a comparison of the large-scale reconstruction of northern North American annual temperature anomalies with the Interior Alaska reconstruction shows opposite trends (Fig. 14.2) that appear to be a consistent part of the climate system.

Across the Northern Hemisphere and beginning in different regions at different times, northern treeline trees display a reduced sensitivity to growing-season warmth (Briffa et al., 1998; Jacoby and D'Arrigo, 1995; Vaganov et al., 1999). The reduction in the positive response of some trees to warm-season temperature seems to have occurred around 1970, when warming resumed after a cool interval in the mid-20th century. When late 20th-century warming resumed, some trees continued to increase in radial growth in response to the warmer conditions. However the growth increase per unit of temperature increase was not nearly as great as previously, and in some trees temperature no longer had a reliable predictive relationship to tree growth at all (Briffa et al., 1998). In Alaska, much of the change can be attributed to greater moisture deficits associated with higher warm-season temperatures (Jacoby and D'Arrigo, 1995). In northern Siberia, the effect is attributed to shorter periods of thawed soil because of increased depth of snow cover, an indirect effect of warmer winters (Vaganov et al., 1999). For other areas, there are hypotheses about air pollution, UV radiation damage, and other factors (Briffa et al., 1998).

14.6.3. Past climate change in northwestern Europe

In the past few decades, Scandinavian tree-ring data have provided an increasing amount of information about past climate variability. Natural reserves preserve a number of old and mature forests virtually untouched by humans, especially in central and northern Scandinavia. Scots pine growing close to altitudinal or latitudinal distribution limits in the Scandinavian Mountains or in northernmost Sweden mainly respond to summer temperatures (with increased growth) and data from such sites have been used to interpret past climate variability (Briffa et al., 1990; Grudd et al., 2002; Gunnarson B. and Linderholm, 2002; Linderholm, 2002). Furthermore, due to the proximity to the Norwegian Sea, and hence the influence of maritime air masses brought in with westerly and south-westerly winds, precipitation may be a growth-limiting factor in moist areas, such as the western slopes of the Scandinavian Mountains or in peatlands (Linderholm et al., 2002, 2003; Solberg et al., 2002). High-frequency North Atlantic Oscillation signals (section 2.2.2.1) have been found in tree-ring data from east-central Scandinavia (Lindholm et al., 2001).

The relationship of climate to tree-ring variability in Scots pine in Scandinavia has been studied in a wide range of growth environments. A comprehensive study was made of pine growing on peatlands along a north–

south profile through Sweden to see if the trees contained high-resolution climate information (Linderholm et al., 2002). Peatland pines were also compared to pines growing on dry sites. Pines growing on peatlands are dependent on growing-season temperature and precipitation, as well as on local water-table variations, which are influenced by longer-term trends in both temperature and precipitation. There is a lag of up to several decades in the response of the pines to water levels, such that trees are integrating the immediate effects of growing-season climate as well as a delayed effect from the water table, making them unsuitable for high-frequency climate reconstruction. The sensitivity of pines growing on peatland also changes depending on climate. When the growing season is wet and cold, temperature is more important and trees respond positively to temperature, in particular to July temperature. Precipitation response increases to the south but is never as important as for pines growing on dry soils. Precipitation is important mainly in controlling water-table levels (Linderholm et al., 2002).

A 1091-year record of tree growth from AD 909 to 1998, developed from living and subfossil Scots pine in the central Scandinavian Mountains, provides evidence of low-frequency climate variation (Gunnarson B. and Linderholm, 2002). July temperatures had the largest effect on the growth of these trees, but growth was also positively and significantly correlated with October to December temperatures in the previous year. The response to precipitation during the vegetative period was negative although not significantly. The authors inferred that the chronology represents summer temperatures for the central Scandinavian Mountains, although it is suggested that care should be taken when interpreting the record. The chronology indicates prolonged excursions below the mean (cool conditions unfavorable for tree growth) in the mid-12th and the 13th centuries, and in the mid-16th and late 17th centuries (corresponding to the early and late LIA, respectively). Below-mean conditions in the late 18th century correlate with a "recent cold period" (Fisher et al., 1998; Grove, 1988; Jones and Bradley, 1992; Lamb 1977b). The chronology also provides evidence for the MWP in the 10th and early part of the 11th centuries as well as warmer periods during the mid-14th, mid-17th, and 20th centuries (Gunnarson B. and Linderholm, 2002).

14.7. Direct climate effects on tree growth

14.7.1. The Flakaliden direct warming experiment

14.7.1.1. Background

In 1994, a soil warming experiment began at Flakaliden near Vindeln, 65 km northwest of Umeå, Sweden (64º 07' N, 19º 27' E). The experiment is in a planted Norway spruce forest established in 1963. The environment of the area is representative of the northern portion of the European boreal forest. Mean annual temperature is 2.3 ºC and mean annual precipitation about 590 mm. The goals of the Flakaliden warming experiment are to:

- quantify the effect of soil warming on the seasonal course of plant respiration and phenology of trees at low (irrigated) and high (irrigated and fertilized) availability of soil nutrients;
- test and improve available mechanistic models used to project impacts of climate change on respiratory dynamics in plants and forest soils; and
- to estimate net carbon budgets for boreal Norway spruce at the tree, stand, and regional scale, in present and future climates.

The soil warming treatment was installed in late 1994 in the buffer zone of one irrigated and one irrigated and fertilized stand; air temperature was not directly modified. Each heated subplot has a corresponding unheated control plot. The reason for using treatments including irrigation was to reduce the risk of drying the soil as an effect of the soil warming. The experiment was not designed to produce the effects of climate warming *per se*, but to isolate the effect of one of the most distinctive features of the boreal forest that is thought to contribute to its great carbon storage – soil temperature. While aboveground production of plant material is relatively great in boreal forests, cold soils limit the rate of decomposition that releases the fixed carbon back into the atmosphere. Therefore, soil warming, if it were to increase decomposition of stored soil carbon more than it affected production, could have a disproportionate effect on carbon balance.

In the Flakaliden experiment, soil warming starts in April each year, about five weeks before the soil thaws in the unheated plots. The soil temperature is increased by 1 ºC per week, until a 5 ºC difference between heated and control plots is reached. In late autumn, when the soil temperature in the control plots approaches 0 ºC, the soil temperature of the heated plot is decreased by 1 ºC per week. If the control plots do not freeze before 1 November, the temperature reduction is still initiated (CarbonSweden, 2003).

14.7.1.2. Questions, hypotheses, and results

Based on the assumptions of doubled atmospheric CO_2 concentration and a 4 to 6 ºC increase in annual mean temperature, the following responses were hypothesized:

- Increased CO_2 and temperature will have a small positive effect on biomass production in boreal forest growing on nutrient-poor sites. The stimulating effect will mainly be due to a shorter period with frozen soils and increased nitrogen mineralization.
- Increased CO_2 and temperature will have a positive effect on photosynthesis in boreal forest growing on sites with good nutrient availability. The net effect on biomass production will, however, be reduced as an effect of increased plant respiration (mainly foliage).

- In boreal forests growing on poor sites, an increase in temperatures will stimulate soil respiration more than biomass production and therefore the net carbon balance will be negative until a new equilibrium is reached. The strength of the carbon source will depend on site index and the size of the soil carbon pool.
- The boreal forest ecosystems will be a major sink for atmospheric carbon, once new equilibriums between carbon fixation and decomposition of soil organic matter have been reached. The time to reach a new equilibrium will depend on site index and the size of the soil carbon pool.

Soil moisture is not normally limiting to growth at Flakaliden (Bergh et al., 1999). The earlier spring soil thawing and later autumn freezing in heated plots increased mineralization of soil organic matter, which increased the concentration of most nutrients in the needles (Bergh and Linder, 1999). The effect was most pronounced during the first years of warming, but was still apparent after the fourth season, by which time stemwood production had increased by approximately 50% compared to the control plots (Jarvis and Linder, 2000). Earlier access to water in spring results in an earlier start of photosynthesis (Bergh and Linder, 1999). After six seasons of warming at Flakaliden, stem volume production (m^3/ha/yr) was 115% higher on heated and irrigated plots than on unheated control plots; on heated, irrigated, and fertilized plots production was 57% higher than on unheated plots. The results indicate that in a future warmer climate, with increased nitrogen availability and a longer growing season, biomass production is very likely to increase substantially on both low- and high-fertility sites in the more humid parts of the boreal forest (Strömgren and Linder, 2002). However, the Flakaliden results are specific to soil warming without air-temperature change incorporated and it is too early to determine whether the observed responses are transitory or will be long-lasting.

In recent years, the large contribution of fine root turnover (growth and death within the growing season) as a factor in the annual production and storage of carbon in boreal forests has been recognized. Interactive effects of soil warming and fertilization on root production, mortality, and longevity at Flakaliden demonstrated that that nitrogen addition combined with warmer soil temperatures decreases the risk of root mortality, and annual fine root production is a function of the length of the growing season (Majdi and Öhrvik, 2004). Under scenarios of climate change that increase soil temperature, and maintain adequate soil moisture and sufficient nitrogen, root production (and carbon stored in roots at a given time) in boreal forests is very likely to increase, especially at low-fertility sites.

The influence of soil temperature on boreal forest growth and carbon storage in natural field situations does not appear to be as great as the potential demonstrated in the Flakaliden results. After four years of warming, a major temperature acclimation had occurred and there was only a small difference in soil CO_2 flux between heated and non-heated plots (Jarvis, 2000). The timing of soil thawing (date of near-surface soil temperature rapidly increasing above 0 °C) was not a good predictor for the start of spring photosynthesis in boreal coniferous forest at five field stations in northern and southern Finland, northern and southern Sweden, and central Siberia. The best predictor of the start of spring photosynthesis was air temperature (Tanja et al., 2003). In one case, photosynthesis commenced 1.5 months before soil thawing. At most sites a threshold value for air-temperature indices projected the beginning of photosynthesis in the spring, which varied among the sites by 30 to 60 days. The threshold values varied from site to site, probably reflecting genetic differences among the species and/or differences in the physiological state of trees in late winter and early spring induced by climate. A single physiological temperature threshold for the start of photosynthesis may not exist.

14.7.2. Climate effects on tree growth along the Central Siberia IGBP transect

The International Geosphere-Biosphere Programme (IGBP) is an international, interdisciplinary scientific research program built on networking and integration. It addresses scientific questions requiring an international approach, and undertakes analysis, synthesis, and integration activities on broad earth-system themes. The goals of the IGBP are to develop common frameworks for collaborative research, form research networks, promote standardized methodologies, facilitate construction of global databases, undertake model and data comparisons, and facilitate efficient patterns of resource allocation.

In the early 1990s, the IGBP developed the Global Change and Terrestrial Ecosystems project to establish terrestrial transects for global change research as one way to study ecosystem and climate change across large spatial scales (Canadell et al., 2002). The IGBP terrestrial transects run for more than 1000 km along specific environmental gradients such as temperature or precipitation, and along more conceptual gradients of land-use intensity. They often cross ecotones such as tundra–taiga that are believed to be highly sensitive regions with strong feedbacks to global change. The Central Siberia IGBP transect was one of five high-latitude IGBP transects.

14.7.2.1. Climate response functions of trees along a latitudinal gradient

The geography of central Eurasia offers an excellent opportunity to examine how climate influences tree growth along an uninterrupted transect from the cold tundra margin in the north to the semi-arid steppe grassland of central Asia. This transect approach can provide some ideas about how climate change might affect growth by comparing the climate factors that historically and currently control tree growth as one proceeds southward. However, several limitations to this approach should be considered. For example, the genetics of individual trees of even the same species change from north to south. However, the comprehensive view of how tree growth responds differently to climate from cool to warm regions is still quite useful.

To define the main climatic factors that influence tree-ring growth in various regions along the Central Siberia IGBP transect, correlation coefficients of tree-ring structure chronologies with monthly temperature and precipitation were calculated (Fritts, 1976; Schweingruber, 1988, 1996). Tree-ring data from Siberian larch, Gmelin larch, Siberian spruce (*Picea obovata*), and Scots pine from 46 sites located in regions from the forest–steppe zone in the south to the forest–tundra zone in the north were used (Fig. 14.16). The relationship of tree-ring width to climate was investigated at all sites.

To strengthen the climatic signal common to each region, tree-ring data from sites were averaged as regional chronologies when sufficiently high correlation of master chronologies from the same vegetation zone permitted. Averaged regional chronologies were obtained for the following vegetation zones along the transect: forest–tundra and the northern part of the northern taiga; northern taiga; middle taiga; and southern taiga. These chronologies were compared to regional climatic data averaged for several meteorological stations. In the south, in the forest–steppe zone, the similarity of master chronologies was lower and three regional chronologies were obtained. They were correlated with the data from the nearest meteorological stations.

Comparison of the climatic response functions obtained for trees growing at different regions along the transect show that there is a change in the climatic factor that defines tree-ring growth at sites located along the temperature gradient in central Eurasia (Fig. 14.17). Summer temperature is one of the most important external factors that define tree-ring growth at the northern treeline (Vaganov et al., 1996, 1999). It positively influences tracheid production and explains up to 70% of the variability in tree-ring width. In the middle taiga region, the effect of summer temperature on tree-ring growth decreases and the influence of precipitation increases (Kirdyanov, 2000). In this region, winter precipitation has a strong negative effect on tree-ring growth, while June temperature has a positive influence. In the forest–steppe zone, tree-ring climatic response functions are typical of regions with limited moisture (Fritts, 1976), where variability in tree-ring width is mainly explained by moisture variability (directly influenced by precipitation and indirectly influenced by temperature). The negative influence of spring and early summer temperature is explained by water loss from the soil at the beginning of growing season. High temperatures in the previous August and September affect soil water content, which is important for tree-ring growth activation during the next growing season (Magda and Zelenova, 2002; Schweingruber, 1996).

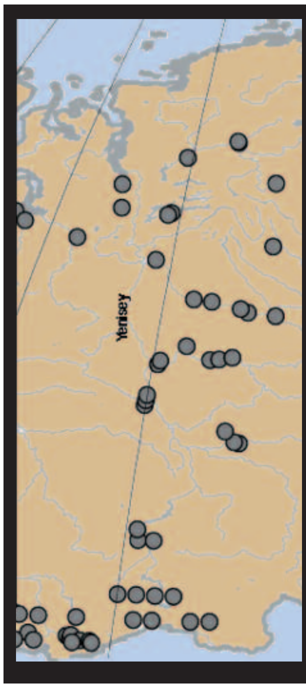

Fig. 14.16. Locations of tree samples along the Central Siberia IGBP transect (Yenisey meridian). See Fig. 14.3b for vegetation and forest types along the transect.

From north to south along the transect, the limiting positive effect of summer temperature in the forest–tundra zone is replaced by the limiting effect of spring precipitation (positive) and early summer temperature (negative) in the forest–steppe zone (Fig. 14.17). Climate conditions (soil moisture and temperature) at the beginning of and during the first part of the growing season play the key role in determining annual radial tree growth and wood production at various latitudes from the northern treeline to the forest–steppe zone. Moving from north to south, the start of the tree-ring growth season shifts to earlier dates. June and July conditions are important for trees growing at the northern treeline, whereas April through June precipitation and temperature influence tree-ring formation in the forest–steppe zone (Fig. 14.17).

Under climate scenarios projected by the ACIA-designated models, trees currently growing in a given central Eurasian forest zone begin to experience temperatures near the end of the 21st century that are typical today of the next zone to the south. The simplest response (linear) would be for the propagules of the vegetation of any zone under consideration to migrate northward and eventually reconstitute the zone further north – in effect a "migration" of the zone northward through regeneration over time, so that present-day zones are replaced in the same sequence by the zones found to the south (Fig. 14.17). However, novel features of the ACIA-designated climate projections and their effects through time could bring about a nonlinear forest response (Fig. 14.17). It is difficult to make specific projections of these outcomes, but it is possible that some processes resulting from the ACIA-designated climate projections will produce unique effects not seen within the range of temperature variability experienced during the last millennium.

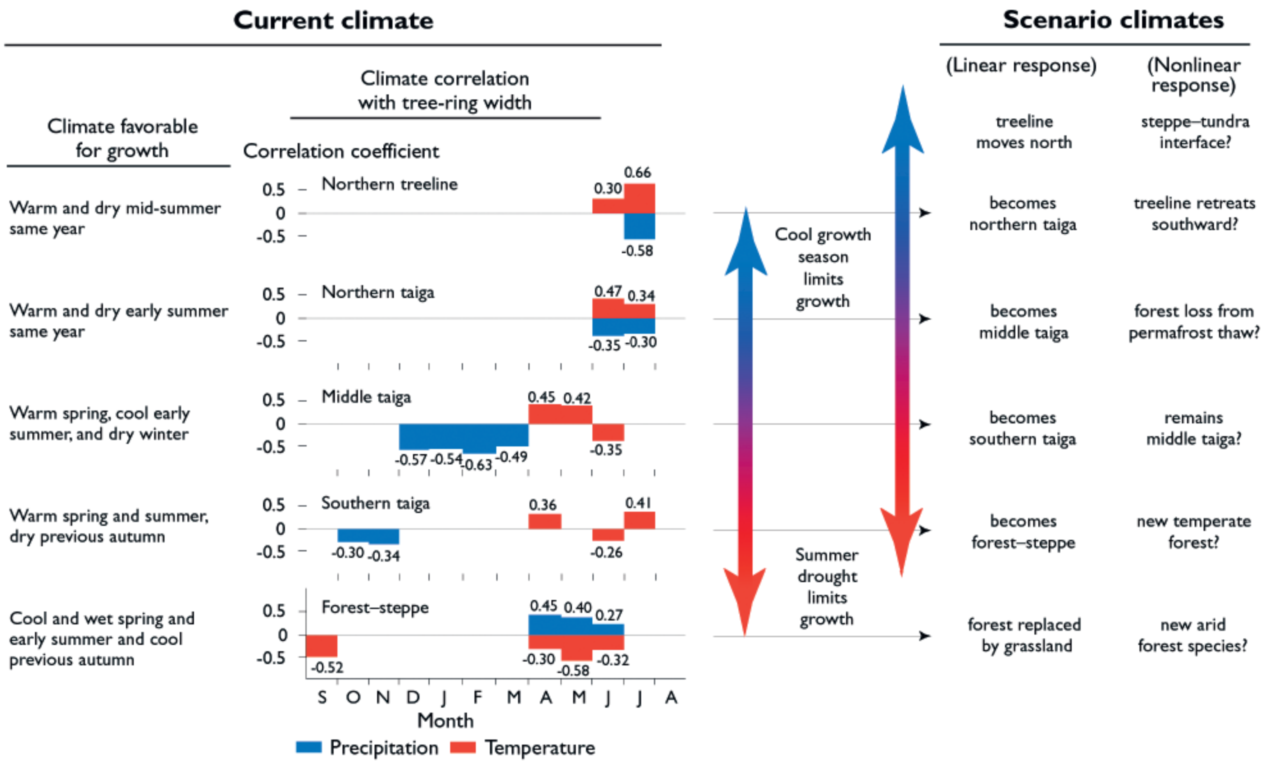

Fig. 14.17. Statistically significant correlations between tree-ring increments and climatic variables (typical climatic response functions) for different vegetation zones of central Eurasia along the Central Siberia IGBP transect. Potential alternative outcomes of climate change are also depicted. A warmer climate could result in the replacement of existing forest vegetation zones in sequence (linear response) or novel ecosystems could appear (nonlinear response) (data from Panyushkina et al., 1997; Vaganov, 1989; Vaganov et al., 1985, 1996).

For example, all the ACIA-designated models project an increase in temperature that is very likely to result in thawing of permafrost along the southern limit of its present-day distribution (section 6.6.1.3) and in low-elevation basins in the southern Yukon, central Alaska, and southwestern portions of the Northwest Territories. Permafrost thawing would transform forest soils and create site conditions that have few or no current analogues. The ACIA-designated models project annual temperatures in the southern boreal forest of central Eurasia at the end of the 21st century that are typical of present-day temperate forest, but the actual amount of moisture supply and species migration could either permit or hinder development of temperate forest in this region. The current territory of southern taiga is projected to be replaced by forest–steppe (Tchebakova et al., 1995) with a 15% phytomass decrease (Monserud et al., 1996) in a warmer climate, and forest degradation and decline during the transition period are very likely to result in a major increase in forest debris and forest flammability. The difference between the ACIA-designated model projecting the largest increase in central Eurasian growing season temperature, generally the CGCM2, and the model projecting the smallest increase, the CSM_1.4, was generally between 15 and 20%. At the extreme upper range of temperature increases in the CGCM2 and the ECHAM4/OPYC3 scenarios, it is possible that warming and drying effects will bring tundra into contact with semi-arid steppe (Fig. 14.17). Empirical relationships between evapotranspiration and vegetation in central North America project that aspen parkland will extend into the Arctic (Fig. 14.1, Hogg and Hurdle, 1995) under a scenario based on doubled atmospheric CO_2 concentrations.

14.7.2.2. Variability in the strength of climate influence on tree growth

Tree-ring variability in Gmelin larch, Siberian larch, and Scots pine was compared from sites located at different latitudes along the Central Siberian IGBP transect. Tree-ring width chronologies were calculated for trees from the forest–tundra zone (71° N); northern (64°–69° N), central (61° N), and southern (58° N) taiga regions; the forest–steppe zone (53° N); and high-elevation forest (51° N) (Table 14.6). To obtain comparable values of mean tree-ring width for different regions, only growth of mature trees (i.e., the period when the age trend in individual tree-ring width curves is not pronounced) was analyzed. The average age of larch trees growing in the south of the study area is less than that of northern larches. Hence, mean tree-ring width for larch from the north (61°–71° N) was calculated for the most recent 50 years for trees older than 200 years, and at the sites located at 57° to 51° N, for larch after 150 years of growth. Because pine trees are generally younger than larch growing at the same latitude, mean tree-ring width of pine was calculated for the most recent 50 years for trees older than 150 years.

Correlation coefficients of individual chronologies with the master time series and coefficients of sensitivity indicate the strength of environmental influences that synchronize tree-ring growth at the same site. The higher these two parameters, the greater the role of the environment in tree-ring growth. These statistics were averaged for the master chronologies obtained for sites from the same latitudinal belt.

Tree-ring width in larch trees increases from north to south up to the region of southern taiga (57º N; Table 14.6), then decreases in the high-elevation forest zone of Tuva (51º N). Correlation of individual chronologies with master time series and tree-ring sensitivity values indicate a decrease in environmental influence on tree-ring growth from north to south along the transect. The highest correlation coefficients were obtained for the northern treeline region and sites 200 km to the south, while the lowest were in the southern taiga region. Tree growth in the southern taiga region is less sensitive to environmental influences than growth of northern trees and trees at high elevations (Table 14.6). Similar changes in tree-ring variability along the transect were found for pine. Tree-ring width increases from the northern border of the pine area to the south taiga region. At the same time, the correlation of individual series with the master chronology and the sensitivity both decrease (Table 14.6). By contrast, in the forest–steppe zone (53º N) the year-to-year variability in pine growth is more closely synchronized to various local climate factors rather than an overall regional signal.

These trends in tree-ring width variability along the north–south Central Siberian IGBP transects repeated longitudinally are typical of tree growth in the entire Siberian boreal zone. Shashkin and Vaganov (2000) reported similar results (increasing larch tree-ring width with decreasing site latitude) from Yakutia, East Siberia, caused by the gradient of environmental factors that most influence tree-ring growth. At the northern treeline, lack of summer warmth is the main climatic factor that limits ring growth (Kirdyanov and Zarharjewski, 1996; Vaganov et al., 1996). As a result, at such high latitudes, large-scale patterns of summer temperature synchronize the growth of trees not only at the same site but also at sites located up to 800 km apart (Vaganov et al., 1999). Temperature increases projected by the ACIA-designated models are very likely to have positive effects on larch growth across this area.

The influence of summer temperature decreases moving southward from the northern end of the transect. This gradient of environmental control leads to the higher growth rate and lower sensitivity of tree growth to climate in the middle of the transect. However, toward the southern portion of the transect, there is once again an increase in the correlation between individual tree-ring growth and sensitivity of master chronologies. Summer warmth in these dry regions is a strongly unfavorable factor for pine ring formation at its southern limit of distribution and larch at its lower elevation limit. Therefore, at the southern end of the transect, the strong influence of environmental factors on larch growth in high-elevation forest and on pine in the forest–steppe zone synchronize ring formation in trees on similar types of sites. The broadly shared environmental controls caused better registration of environmental changes in tree-ring structure (higher sensitivity) among sites. Temperature increases projected by the ACIA-designated models are very likely to have negative effects on larch and pine growth across the southern part of the transect.

Table 14.6. Statistical characteristics of larch and pine ring-width chronologies from different locations along the Central Siberia IGBP transect.

Latitude (° N)	Number of sites	Number of trees	Mean tree-ring width (mm)	Correlation with master chronology	Coefficient of sensitivity
Larch					
71	4	48	0.19	0.8	0.42
69	3	55	0.21	0.82	0.4
64	7	71	0.24	0.67	0.34
61	5	78	0.37	0.66	0.28
57	3	28	0.53	0.58	0.21
51	4 and 3[a]	33	0.46	0.68	0.37
Pine					
66	1	19	0.32	0.65	0.3
61	3	24	0.36	0.6	0.22
58	2	24	0.52	0.55	0.21
53	4 and 2[a]	16	0.62	0.65	0.32

[a] Two samples at this latitude; the number of trees is the total for both sites.

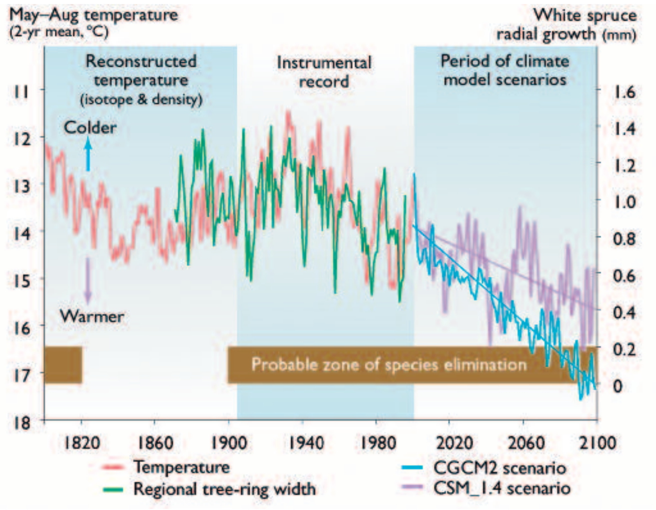

Fig. 14.18. Historic and reconstructed relationship between white spruce growth and summer temperature at Fairbanks, Alaska, and projections based on climate scenarios. The tree growth sample includes 10 stands across central Alaska. Summer temperature is an excellent predictor of white spruce growth. Because higher temperatures are associated with reduced growth and growth is the dependent variable, the temperature scale (left axis) has been inverted. Given the historical relationship between the variables and the scaling of these axes, the temperatures projected by the ACIA-designated models can be used to infer the approximate level of growth (right axis) possible in the future (data from Barber et al., 2004).

14.7.3. Response of high-latitude conifers to climate and climate change scenarios

14.7.3.1. White spruce in Alaska and Canada

The scientific literature on the relationship of tree-ring width to climate in northern North America is dominated by studies of white spruce carefully selected in order to allow the reconstruction of past climates (section 14.6.2) from trees that achieve greater growth with warmer temperatures and lower growth with cooler temperatures. However, radial growth of white spruce on upland sites across a broad area of central Alaska exhibits a strong negative response to summer temperature (Juday et al., 1998, 1999; Fig. 14.18). The negative relationship of radial growth to summer temperature is consistent throughout the 20th century, and occurs in a broad range of dominant and co-dominant trees in mature and old stands (Barber et al., 2000). The growth of white spruce on these sites is best projected by the mean of May through August temperature in the year of ring growth and the year prior (Barber et al., 2000). This relationship between climate and tree growth is sustained during the period of the instrumental record as well as the period for which summer temperatures were reconstructed from ^{13}C isotope content and latewood density.

Based on this strong relationship, the two-year mean of May through August temperature for the Fairbanks grid cell was calculated from the ACIA-designated model and used to project future growth of this species on similar sites (Fig. 14.18). Because the relationship between temperature and growth is negative (less growth in warmer conditions) and the response variable of interest is tree growth, the temperature axis is inverted (increasing temperature downward) in Fig. 14.18.

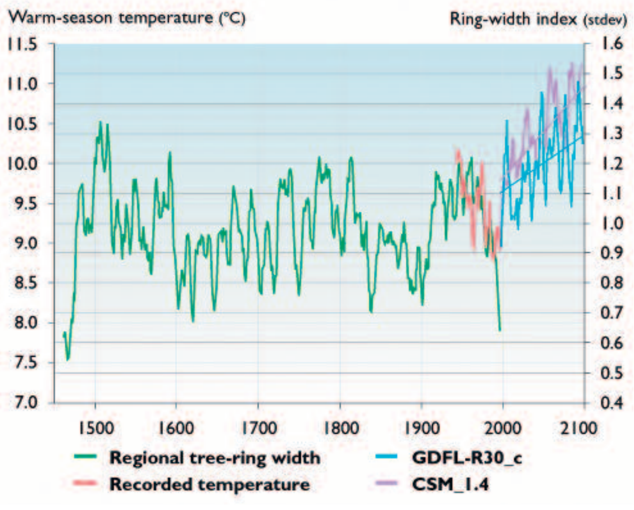

Fig. 14.19. Historic relationship between white spruce growth and mean warm-season temperature (June, July, and September of the growth year and April of the previous year, smoothed with a 5-year running mean), and climate scenarios for central and northern Labrador, Canada (data from D'Arrigo et al., 1996).

The CGCM2 scenario projects the highest temperatures for the Fairbanks grid cell, although with a reduced range of annual variability. The CSM_1.4 scenario projects the least warming in the grid cell of the five ACIA-designated models, with variability similar to the recorded data (Fig. 14.18). If white spruce growth maintains the same relationship to temperature in the scenario period as during the calibration period, under the CGCM2 scenario growth is very likely to cease (the empirical relationship reaches zero growth) by the end of the scenario period (Fig. 14.18). Under the CSM_1.4 scenario, white spruce growth is very likely to decline to about 20% of the long-term mean. The zone of temperature and tree growth in Fig. 14.18 that corresponds to 20% or less of long-term mean growth has been highlighted as a "zone of probable species elimination". The stressed condition of trees in such a climate is likely to predispose them to other agents of tree mortality such as insect outbreaks (section 14.8) and diseases. While the CSM_1.4 scenario does not produce warming by the end of the scenario period that is empirically associated with zero growth, this white spruce population would be growing in a climate that is very likely to greatly reduce its growth. This climate and reduced level of growth almost certainly would place the trees that occur there now at an elevated risk of mortality, primarily from fire (section 14.9.2.3) and insects (section 14.8.2). White spruce in this region that demonstrate this climatic response are among the largest, most rapidly growing, and commercially valuable in boreal Alaska.

At high latitudes and altitudes where moisture is not limiting, white spruce has a positive growth response to summer temperature. The growth of near-treeline white spruce north of Goose Bay, Labrador in eastern Canada is positively correlated with the mean of monthly temperature in June, July, and September of the growth year and April of the previous year (Fig. 14.19; D'Arrigo et al., 1996). Some of the trees in the sample may have been responding to non-climatic factors in the early 1950s and especially in the late 1990s when a significant short-term growth decline occurred (Fig. 14.19). The long-term tree-ring chronology available from the Labrador near-treeline sample reconstructs past climates in general agreement with those of several instrument-based and modeling studies of this sector of the North Atlantic (D'Arrigo et al., 1996).

If the relationship between temperature and tree growth in the Labrador sample is maintained in the future, the ACIA-designated model projections again provide a basis for evaluating possible tree growth responses. In the grid cell containing Goose Bay, the CSM_1.4 model projects the greatest amount of warming by the end of the scenario period and the GFDL-R30_c scenario projects the least warming. The long-term mean of ring-width growth in the sample is set to 1.0 and the variation is expressed in units of standard deviation (Fig. 14.19). As projected using the regression lines of the scenarios, the growth of trees that retained the historical relationship of temperature and growth would increase to about 0.3 standard deviations greater than the long-term mean under the GFDL-R30_c scenario and nearly 0.5 standard deviations greater under the CSM_1.4 scenario. Both scenarios produce warming and inferred levels of tree growth that this area has not experienced since the beginning of the proxy record in the late 1400s. These populations of white spruce are relatively small, slow growing, and generally not commercially valuable.

The response of treeline white spruce populations to climate change is of particular interest because this species delimits much of the North American treeline (Sirois, 1999) and the performance under warmer conditions of present-day populations is an indication of the possible future effects of climate change. In a large (>1500 trees) sample of treeline white spruce across the mountains of the Brooks and Alaska Ranges, over 40% displayed a statistically significant negative growth response to summer warmth while slightly fewer than 40% had a positive response to late spring warmth (Wilmking et al., 2004). In the negatively responding (warmer = less growth) white spruce population, July temperature explained most of the variability in growth. Growth was strongly reduced at temperatures above a July threshold of about 16 ºC at the Fairbanks International Airport climate station, which served as a representative common reference (temperatures were estimated to be 3 to 4 ºC cooler at the various treeline sites). Growth of positive responders is correlated to March or April temperature, and the relationship is generally not significant until the second half of the 20th century (Wilmking et al., 2004). D'Arrigo and Jacoby (Lamont-Doherty Earth Observatory, Columbia University, pers. comm., 2003) found a similar dual response in white spruce in the Wrangell St. Elias Mountains of Alaska. A study of eight sites at and near alpine and arctic treeline in three regions of Alaska found mixed populations of temperature response types as well, and growth decreased in response to increasing temperatures at all but the wettest sites after 1950 (Lloyd and Fastie, 2002). The negative growth response to temperature is more common in contiguous stands and tree islands (clusters of trees) than in isolated individual trees (Lloyd and Fastie, 2002; Wilmking et al., 2004). The intensity of the negative effect of July warmth increased after 1950, directly reflecting July temperatures above the threshold in a greater number of years in the second half of the 20th century compared to the first half (Wilmking et al., 2004).

The explanation for reduced growth in treeline white spruce with warming is that negative responders are experiencing temperature-induced drought stress, while the positively responding trees are not (Jacoby and D'Arrigo, 1995; Juday et al., 2003; Wilmking et al., 2004). At some treeline sites in northern Alaska, individual trees have shifted their response to climate during their lifetimes, following the regime shift to warmer conditions that took place in the last decades of the 20th century (Jacoby and D'Arrigo, 1995). Whereas previously the growth of the tree responded positively to summer temperature, it either became insensitive or began responding negatively to summer temperature after the shift to the warmer regime.

These results establish that from an ecological perspective, recent climate warming has been a major event that has strongly affected the growth performance of the majority of white spruce at and near treeline in Alaska and almost certainly in similar climate zones in Canada. The five ACIA-designated models project July temperatures by the mid- to late 21st century that the empirical relationship associates with very low or no growth in negatively responding white spruce. This suggests that under the ACIA-designated model projections, it is possible that the northern white spruce tree limit in Alaska and adjacent Canada would not readily advance. At the least, treeline is likely to become much more complex, with negative responders disappearing and positive responders expanding. It is possible that novel conditions would emerge, such as a portion of the southern tundra boundary in North America that is separated from the boreal forest by aspen parkland (see Fig. 14.1; Hogg and Hurdle, 1995). The actual response of treeline white spruce in this part of the world also adds support to the suggestion that aridification could be a major issue under climate scenarios projected by the ACIA-designated models.

14.7.3.2. Black spruce in Alaska and Canada

Stands dominated by black spruce represent about 55% of the boreal forest cover of Alaska and a large fraction of the northern boreal region of Canada. The response of black spruce to climate has not been studied as extensively or for as long as white spruce, so the literature on climate/growth relationships is more limited. This section draws on some recent dendrochronology work in Interior Alaska.

Different temperature factors are associated with black spruce growth on different permafrost-dominated sites (Fig. 14.20), a more variable response than the consistent temperature/growth response in upland white spruce (section 14.7.3.1). Of four sites examined, three show a negative growth response to increasing temperature, while one shows a positive response (Fig. 14.20). Both the Toghotthele site (Fig. 14.20a) and the Zasada Road 10 site in the Bonanza Creek Long Term Ecological Research site (Fig. 14.20c) show a negative growth response to summer temperatures (year of growth and 2 years prior). Radial growth of black spruce at both the northern and southern BOREAS sites in western Canada was also negatively related to summer temperature, or conversely, favored by cooler and wetter conditions (Brooks et al., 1998).

Black spruce at the Caribou-Poker Creeks Research Watershed (CPCRW) show both a negative correlation to April and May temperatures during the growth year and a positive correlation to February temperature two years prior to growth (Fig. 14.20b). A simple two-

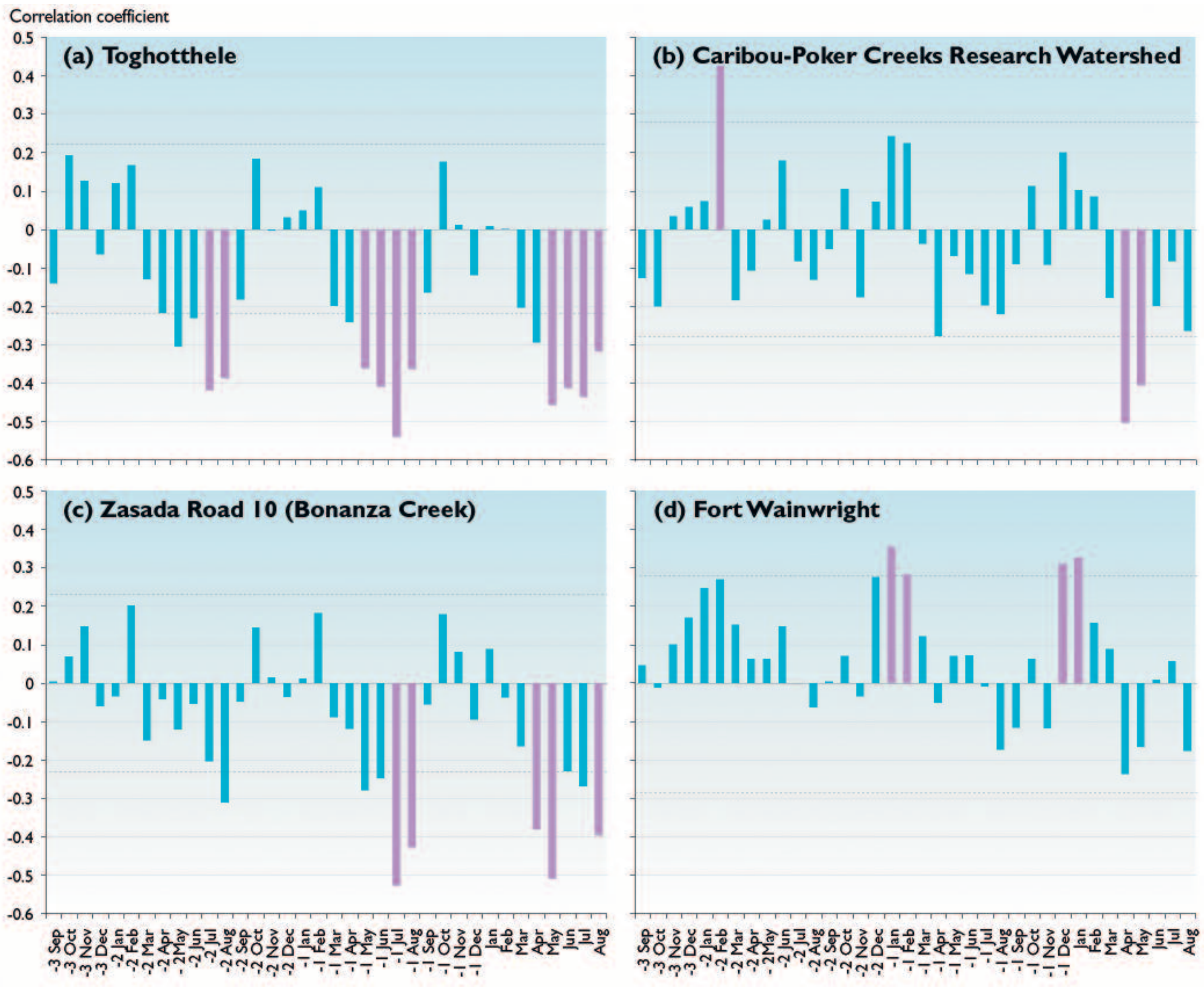

Fig. 14.20. Correlation of black spruce radial growth and Fairbanks mean monthly temperatures from four permafrost-dominated sites in central Interior Alaska. Lavender bars indicate a statistically significant correlation between the mean monthly temperature and tree growth, while blue bars indicate that the correlation is not statistically significant. Negative numbers preceding months on horizontal axis represent years before the year in which growth occurred (data from Juday and Barber, 2005).

month index of the mean of April and February temperature is highly correlated to the growth of the CPCRW trees (Fig. 14.21). The negative effect of warm early-spring temperatures on growth at CPCRW can be attributed to the onset of photosynthesis in spring when the ground is still frozen, causing desiccation and damage to the needles (Berg E. and Chapin, 1994) early in the growing season. The smoothed (5-year running mean) values are highly correlated during the 20th century (r=0.86), suggesting that tree growth of this species at sites similar to these could be projected using the ACIA-designated scenarios. While warm February temperatures favor growth, warmth in April depresses growth. With substantial temperature increases in the late 20th century, growth of this species has declined because the negative influence of April is stronger.

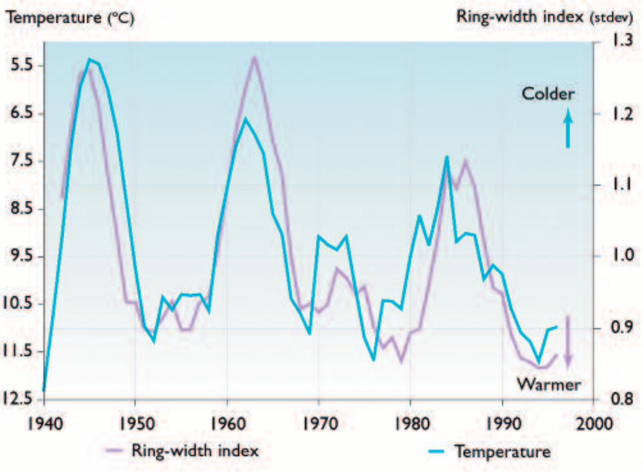

Fig. 14.21. Relationship of radial growth of black spruce at Caribou-Poker Creeks Research Watershed to the mean of April (growth year) and February (two years prior) temperature, smoothed with a 5-year running mean. Temperature axis is inverted as in Fig. 4.18 (data from Juday and Barber, 2005).

Growth of black spruce trees at Fort Wainwright is positively correlated with winter temperatures (Fig. 14.20d). When January in the year of growth and December, February, and January in the year prior to growth were all

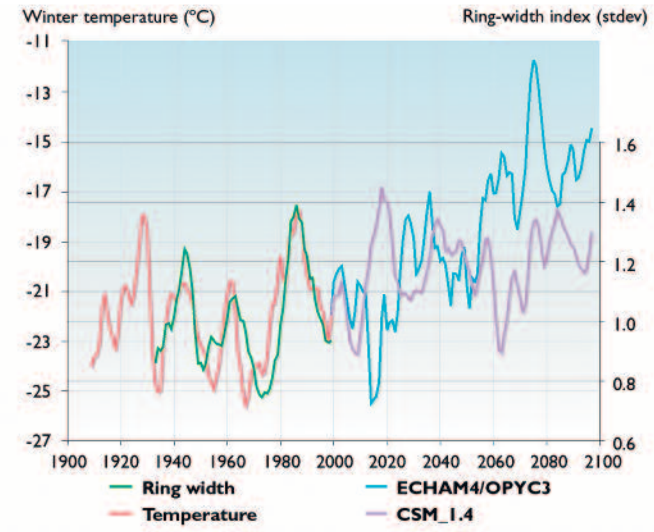

Fig. 14.22. Relationship of radial growth of black spruce at Fort Wainwright, Alaska (n=20 trees), to a 4-month climate index (mean of monthly temperature at Fairbanks, Alaska, in January of growth year and January, February, and December of previous year). Scenario lines show projections of the 4-month climate index (data from Juday and Barber, 2005).

warm, the trees grew better (Fig. 14.22). The smoothed (5-year running mean) values are highly correlated during the 20th century. This is one of the few species and site types in central Alaska for which the empirically calibrated growth rate can be inferred to improve under projected higher temperatures.

The positive relationship between monthly temperatures and black spruce growth at the Fort Wainwright site suggests that growth of tree on sites similar to it is very likely to increase (Fig. 14.22). The ECHAM4/OPYC3 model projects especially strong warming for the Fairbanks grid cell in the winter months that best predict black spruce growth at the site, so the empirical relationship, if it were maintained, suggests a major increase in growth rate (Fig. 14.22). Black spruce at the Fort Wainwright site occupy a low productivity system, so the overall significance of the projected growth increase is not clear. The CSM_1.4 model projects winter temperatures that do not increase as rapidly or to as great a degree, so the modeled relationship of these positively responding spruce suggests only a modest increase in growth under that scenario. Presumably, the positive effect of warm winter temperatures on growth is experienced though control of active-layer rooting depth and soil temperature in this permafrost-dominated ecosystem. However, well before the end of the scenario period, the CGCM2 (Fig. 14.9) and all the other models project mean annual temperatures above freezing in this grid cell. If temperatures increase to that extent, the temperature/growth relationship depicted in Fig. 14.22 is not likely to persist. However, this permafrost site is very near thawing, and warming of the magnitude projected by the ACIA-designated models would probably initiate thawing during the 21st century, leading to widespread ground subsidence and tree toppling, representing a new challenge for the survival of this species on such sites. Once the soil thawing process is complete, species with higher growth rates than black spruce, such as white spruce or paper birch, are likely to have a competitive advantage on the transformed site.

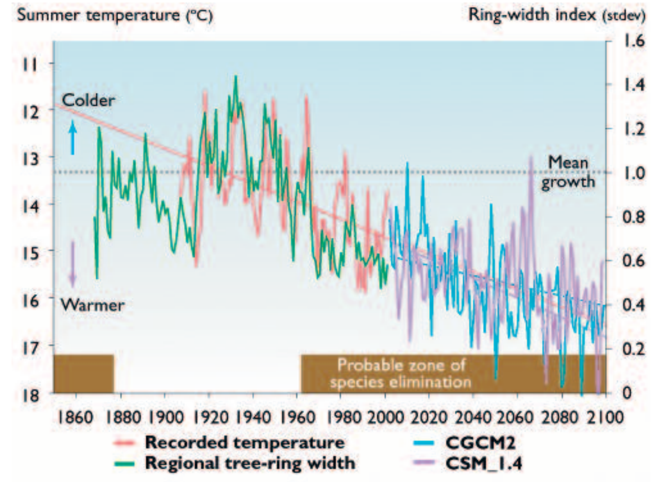

Fig. 14.23. Relationship between summer temperatures (mean of May and June in growth year, and June and July of the previous year) at Fairbanks and relative growth of black spruce at the Toghotthele site in central Alaska, and projections based on climate scenarios. Temperature axis is inverted as in Fig. 14.18 (data from Juday and Barber, 2005).

For the Toghotthele site (Alaska Native-owned land; Toghotthele Corporation), model projections of future temperature were compared to the empirical record of black spruce growth (Fig. 14.23) similar to the approach used for white spruce (section 14.7.3.1). The mean of four summer months is an excellent predictor of tree growth, with warm years resulting in strongly reduced growth. As for white spruce, the results suggest that if climates similar to those projected by the ACIA-designated models actually occur, by the end of the 21st century black spruce would experience climates that are very unlikely to permit the species to survive on similar types of sites. In this case, the CSM_1.4 model projects the highest levels of the particular set of monthly temperatures that drives the relationship while the CGCM2 model projects slightly lower levels. However, in both scenarios, warmth in individual years produces an empirical relationship very near zero growth before the end of the scenario period. Allowing for some differences in calibration and degree of climate control, the generally similar growth of black spruce in the BOREAS study areas suggests that elimination of black spruce is very likely to be widespread across the western North American boreal forest.

14.7.3.3. Scots pine in Scandinavia

Growth responses of Scots pine at the tree limit in the central Scandinavian Mountains varied throughout the 20th century (Linderholm, 2002). Long, hot, and dry summers have traditionally been thought to be optimal for pine growth in that environment. The greatest pine growth of the past three centuries occurred during the decade from 1945 to 1954. Although summer temperature was not particularly high during the mid-20th century, higher than average spring and autumn temperatures that extended the growing season were inferred to be the cause of this decade of high Scots pine growth (Linderholm, 2002). Despite similar apparently favorable temperature conditions in the latter part of the 20th century, growth of this species on similar sites declined during the most recent warmth. This suggests the occurrence of some additional unique component of the recent warming that negated the previously positive influence of extended growing seasons.

Fig. 14.24. Location of 14 Scots pine ring-width chronologies (see Table 14.7 for key to locations).

Table 14.7. Locations of 14 Scots pine ring-width chronologies. First column shows abbreviations used in Figs. 14.24 and 14.25.

	Location	Site type
N	Nikkaluokta	dry
AJ	Angerusjärvi	wet
LY	Lycksele	dry
ST	Stortjäderbergsmyran	wet
J	Jämtland	dry
ÅÖ	Årsön	wet
TS	Tannsjö	dry
SK	Skuleskogen	dry
NO	Norberg	dry
BM	Bredmossen	wet
SH	Stockholm	dry
HM	Hanvedsmossen	wet
GH	Gullhult	dry
AM	Anebymossen	wet

In addition to north–south transects of tree-ring chronologies, an east–west transect has been developed in Fennoscandia (Linderholm et al., 2003). Nine tree-ring width chronologies for Scots pine were compared for growth variability and response to climate along a gradient of maritime to continental conditions in central Fennoscandia. The study revealed higher growth variance and stronger response to climate in the oceanic area west of the Scandinavian Mountains, compared to the more continental areas further east. Pine growth responded positively to elevated summer temperatures in the western areas, and positively to high summer precipitation in the east. Generally, pine growth showed a weaker relationship with the North Atlantic Oscillation (section 2.2.2.1) than with temperature and precipitation. During the last half of the 20th century, pine growth in western Fennoscandia displayed reduced sensitivity to climate, while in the east, growth sensitivity increased. Indications of growth stress were found in one site east of the Scandinavian Mountains. Increasing temperatures have been accompanied by increasing precipitation in Fennoscandia throughout the 20th century, and it was suggested that a change in climate regime from subcontinental to sub-maritime caused those trees to experience climatic stress (Linderholm et al., 2003).

A selection of Scots pine tree-ring width chronologies was collected from across Sweden (Fig. 14.24). The standardized chronologies include wet (peatland) and dry (mineral soil) growth environments (Table 14.7). Despite large differences in climate responses among the sites, the collective growth trend for all the sampled Scots pine in the last few decades of the 20th century is negative (Fig. 14.25). Furthermore, a common feature is distinct changes in growth sensitivity to climate at all sites during that period, a feature that has also been observed at several other Fennoscandian sites. The pattern is one of widespread and simultaneous decreases in growth, and a decreased ability of previously established climate factors to project growth as accurately.

Because precipitation, in addition to temperature, has increased over the past decades in Scandinavia, it is unlikely that drought stress is the main reason for decreasing Scots pine growth throughout Sweden. The one exception is northernmost Sweden, where growth responses to summer temperatures are decreasing and responses to autumn precipitation increasing. However, in central Sweden, the evidence suggests a different cause of reduced growth and changing climate responses. On the eastern slopes of the Scandinavian Mountains, the climate could be regarded as subcontinental, with low annual precipitation and high temperature amplitude over the year. Tree-ring data from sites east of the mountains implies that over the past decades, the climate of this area has become more maritime, and consequently the trees may have suffered stress from surplus moisture rather than drought

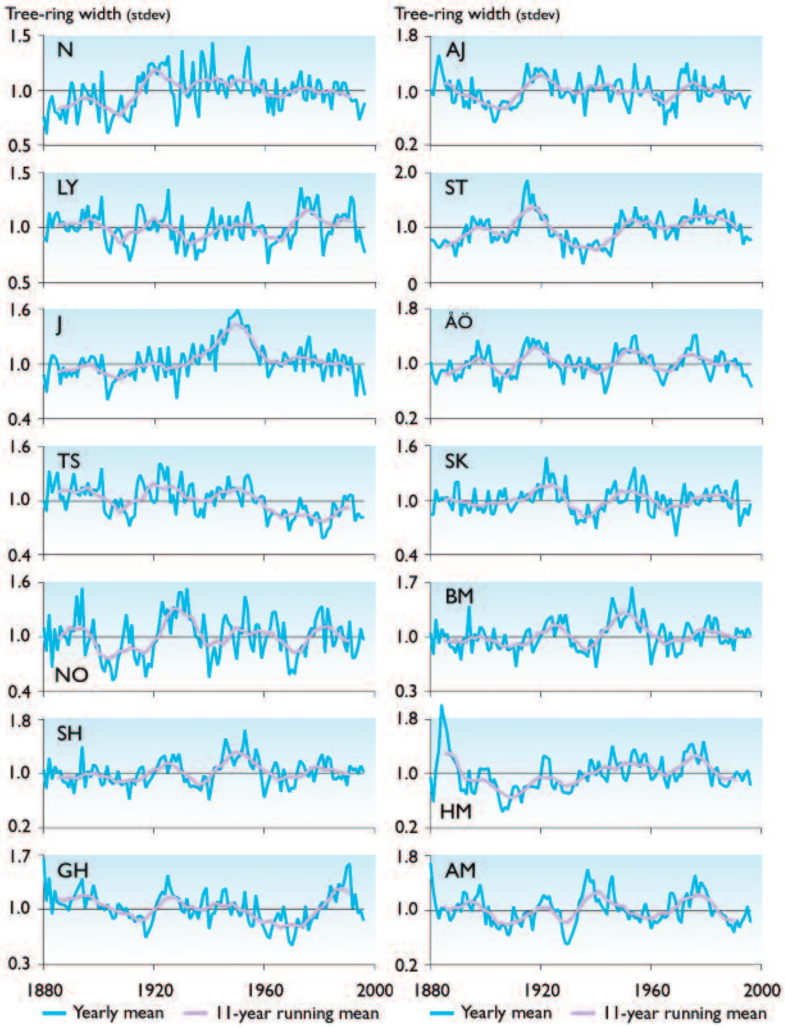

Fig. 14.25. Standard chronologies of Scots pine ring width (age de-trended and normalized with long-term mean set to 1.0 and standard deviation on 1.0) in Sweden. Curves express the degree of growth at any time relative to the long-term mean of the sample (see Table 14.7 for key to locations; data from Linderholm, 2002; Linderholm et al., 2002, 2003).

(Linderholm et al., 2003). In conclusion, the observed change in tree growth patterns and growth/climate relationships is most likely an effect of contemporary climate change. How trees respond to increased temperatures and/or precipitation is dependent on the local climate, and it seems that trees in oceanic areas are less affected than are those where the local climate regime has shifted from subcontinental to sub-maritime.

14.8. Climate change and insects as a forest disturbance

14.8.1. Role of insects in the boreal forest

The boreal forest is naturally subject to periodic large-scale tree mortality from insect outbreaks. The death of attacked trees has a considerable influence on the balance of carbon sequestered by forests. In boreal forests of Canada, insect-caused timber losses (tree death) may be up to 1.3 to 2.0 times greater than the mean annual losses due to fires (Volney and Fleming, 2000). The particular insects involved in large-scale boreal outbreaks include bark and wood-boring beetles, defoliating insects (often Lepidoptera – moths and butterflies), and insects that attack roots and cones. Major population increases of these insects (irruptions), along with wildland fires, are the major short-term agents of change in the boreal forest. Insect-caused tree mortality can appear suddenly and within 1 to 10 years affect hundreds of thousands to millions of hectares of boreal forest in an outbreak or related series of outbreaks. Moderate to severe defoliation of trees by insects has affected nearly 16 million ha/yr in the province of Ontario since 1975 (Parker et al., 2000). Nearly all boreal forest regions are subject to this phenomenon, and it may represent an adaptive response in which simultaneous mass tree death helps create conditions for the renewal of site productivity through processes that warm soils, mobilize nutrient elements, and promote simultaneous tree regeneration.

The peculiar vulnerability of boreal forests to large-scale insect disturbance may also be related to the low biodiversity of both the host tree populations (limited number of species) and the limited diversity within the complex of tree-attacking insects and the populations of predators and parasites that work to stabilize their numbers. Ecological theory predicts that systems with few species should be less stable or buffered against major population swings, and of the major forest regions of the world, the boreal forest is the least species-rich in most taxonomic groups. Finally, the occurrence of tree hybrids may be a non-climatic risk factor for large-scale insect outbreaks. Where the distributions of two wind-pollinated tree species of the same genus (e.g., spruce, pine, larches) approach each other or overlap, the trees can interbreed and form hybrid offspring. Because the hybrid trees contain a random assortment of the genes of both parents, they seldom inherit highly integrated and complete defense traits against insect attacks, such as specific chemical defenses or timing of events to avoid vulnerability. As a result, tree defenses against insect attack may be lowest in hybrid zones (Whitham, 1989; Whitham et al., 1994).

The level of tree mortality caused by insect outbreaks can vary from selective removal of a few percent of dominant trees to the death of nearly the entire forest canopy. Although vulnerable tree species and age classes must be present for insect outbreaks to occur, climatic events are often the trigger or proximate cause of the insect population increases and associated widespread tree mortality. In general, many organisms in the boreal region, and especially the northern boreal forest region, are heat-limited, that is, they could perform (survive, grow, and reproduce) better with more warmth if other factors did not become limiting. Many insect agents of tree mortality in the northern boreal region are heat-limited, so that sustained periods of abnormally warm weather are often associated with irruptions of insects that attack boreal tree species. Given that this is a natural climate-driven system, it is important to distinguish climate change effects from natural operation of this system.

Generally, a climate change effect imposed on the natural insect/tree-death system is very likely to result in a greater frequency of insect outbreaks; more extensive areas of tree mortality during outbreaks; and greater intensity of insect attack resulting in higher average levels of tree death within outbreak areas. The sustained operation of these factors as a result of temperature increases, if they occur, is very likely to begin to change regional tree species composition within decades. A necessary confirmation of the occurrence of a specific climate change effect on the insect/tree-death process would include the detection of differences between the composition of the new, altered forest and historical and recent paleo-ecological records of previous generations of forest. An alternative possibility that needs to be considered is that temperature increases (along with increased precipitation) will possibly produce a more favorable growth environment for certain tree species on certain site types. The increased vigor of the trees is very likely to increase their ability to produce defensive compounds and successfully resist the similarly likely increased levels of insect attacks.

It is possible that the influences of climate warming on forests and on tree-hosted insects (positive and negative for forest health) could occur simultaneously in different locations of the boreal forest, or could occur sequentially in one place at different magnitudes and rates of temperature increase. As a result, temperature increases could be associated with improved forest health in some locations and forest health problems elsewhere, and the climate change effect at any given time would be the sum of these two outcomes. In addition, the initial stage of temperature increases would produce one profile of the balance of forest health outcomes across the area of analysis, but later stages representing greater temperature increases would shift the previous balance of forest health outcomes. The empirical record strongly suggests overall negative effects on boreal forest health from sustained temperature increases.

14.8.2. Spruce bark beetle in Alaska

During the 1990s, the Kenai Peninsula in south-central Alaska experienced the largest outbreak of spruce bark beetles (*Dendroctonus rufipennis*) in the world (Werner, 1996). The forests of the Kenai lowlands are transitional between coastal Sitka spruce (*Picea sitchensis*) rainforests and the white spruce–paper birch boreal forests of semiarid central Alaska. From 1989 to 1999, more than 1.6 million ha of mature forest made up of white spruce and hybrids of Sitka and white spruce (called Lutz spruce) in south-central Alaska experienced at least 10 to 20% tree mortality, the threshold level for aerial mapping detection. In much of the insect outbreak area, virtually all mature canopy trees were killed. Extensive logging was initiated to salvage beetle-killed timber on private lands, and the US federal government has found it necessary to appropriate funds to reduce fuel hazards in stands with numerous dead trees after infestation.

The first documented spruce bark beetle infestation on the Kenai Peninsula occurred in 1950. Local infestations occurred throughout the 1950s and 1960s on the northern Peninsula (Holsten, 1990). A major outbreak in the northern Peninsula that started in 1970 prompted the US Forest Service to initiate annual surveys in 1971, which have continued with few interruptions. Annual survey maps show that spruce bark beetles are endemic to Kenai Peninsula forests, at least under the present climatic regime. In any given year, there is always a background level of infestation somewhere in south-central Alaska, with periodic major outbreaks in some years.

The relationship of spruce bark beetle to climate involves two direct temperature controls over populations of the insect, and an indirect control through host tree resistance. Two successive cold winters depress the survival rate of spruce bark beetle to such a low level that there is little outbreak potential for the following season (Holsten, 1990). The spruce bark beetle is a heat-limited organism on the Kenai Peninsula and normally requires two summers to complete its life cycle. During abnormally warm summers on the Kenai Peninsula, the spruce bark beetle can complete its life cycle in one year, dramatically increasing the potential for population buildup (Werner and Holsten, 1985). Tree health increases host resistance to beetle attack: healthy spruce trees can successfully resist moderate numbers of beetle attacks by opposing the wood-boring activity of females entering the tree to lay eggs with pitch under high turgor pressure (Holsten, 1990). The beetles are unable to overcome the flow of pitch and either are expelled or succumb in their pitch tubes. Host trees that are under stress, including either climate stress or stress from mechanical breakage, have reduced growth reserves, less pitch, and lower turgor pressure, and so are less able to resist spruce bark beetle attacks. When regional events stress entire populations of trees, spruce bark beetle reproductive success is greatly increased.

The US Forest Service conducts annual aerial surveys for “red needles” (i.e., conifer needles dead for a year or less) that allow areas of spruce killed by spruce bark beetle within the previous year to be mapped. These annual surveys describe the areal extent of active infestation; densities of recently killed trees can range from a few stems to hundreds of stems per hectare (USDA Forest Service, 1950-1999). There is a one- to three-year lag between beetles entering a stand and a stand-wide flush of red needles (tree death). Typically, a sequence of red-needle acreage maps on the Kenai Peninsula will show a rise and fall of active infestation over a 5 to 10 year period on a scale of 100 to 1000 ha. After that time, nearly all white/Sitka/Lutz spruce greater than 12 cm in diameter will generally be dead in the affected stand, and subsequent surveys will record no more red-needle acreage for that stand.

Following the 1976–1977 increase in North Pacific sea surface temperatures (Mantua et al., 1997), mean annual and summer temperatures in the Kenai Peninsula lowlands rose 0.5 to 1.5 °C. This temperature increase is the latest acceleration of post-LIA climate warming in south-central Alaska (Jacoby and D'Arrigo, 1995; Wiles et al., 1996). Until recently, the most visible manifestation of the late 20th-century warming was a drying landscape, with dried-up ponds, falling water levels (as much as 1 m) in closed-basin lakes, and black spruce invasion of peat wetlands across the landscape.

Annual red-needle acreages for the southern Kenai Peninsula can be compared with standardized climate parameters recorded at the Homer Airport climate station (Fig. 14.26). In Fig. 14.26c, the drought index is calculated by subtracting the normalized past October to present September total precipitation from the normalized May through August mean temperature of the named year. The drought index is designed to combine the moisture contribution of the previous winter with the growing season moisture and temperature of the identified year. The first major spruce bark beetle outbreak of the post-1950 period occurred in the early 1970s, following the extremely warm and dry period of 1968 to 1969 on the northern Peninsula (Fig. 14.26). Red-needle mortality dropped to nearly zero by 1975, following three cool summers (1973–1975). On both the northern and southern Peninsula, the sustained onset of warm summers beginning in 1987 was followed by substantial increase in red-needle mortality beginning in 1990 and reaching a maximum in 1996. Annually mapped red-needle area declined after 1996 on the southern Peninsula because beetles had killed most of the available mature spruce forest, not because climatic factors became less favorable for insect population increases.

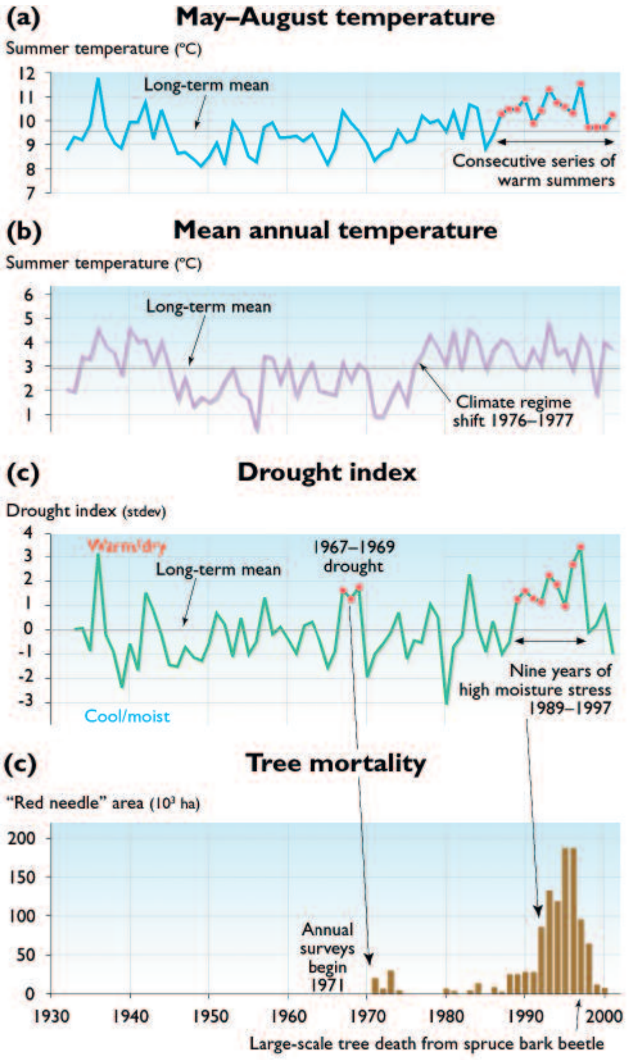

Fig. 14.26. Timing of climate events that release spruce bark beetles from population limits showing (a) mean summer (May–Aug) temperature; (b) mean annual temperature; and (c) drought index, compared to (d) actual outbreaks of bark beetles (represented by "red-needle" area) during the 20th-century period of record (data from E. Berg, Kenai National Wildlife Refuge, Soldotna, Alaska, pers. comm., 2002; NOAA, 2002 (Homer Airport station); Wittwer, 2004).

The association of red-needle mortality with drought is particularly striking in the southern Peninsula forests between 1989 and 1997 (Fig. 14.26c,d). Spruce bark beetles first attack the large, slowly growing trees in a stand (Hard, 1985, 1987). The survival of smaller trees (both young saplings and released understory pole-size trees) is attributed to their ability to produce enough pitch to swamp the beetle galleries and kill the larvae. An additional factor may be that large trees are more prone to drought stress. Due to cold soils, spruce trees in Alaska usually root within the upper 30 to 40 cm of the soil and extend their roots laterally for 1 to 2 m. This shallow, extended rooting can bring trees into competition with neighboring trees and surface vegetation, especially in times of drought. Ring widths in many large trees generally decline for 5 to 10 years prior to the tree succumbing to bark beetles, clearly signifying a stressed condition. Drought is the most likely source of this stress (Fig. 14.26, Hard, 1987).

While warm summers are the proximate cause of outbreaks, spruce stands must reach a certain level of maturity before serious spruce bark beetle infestation can develop, regardless of summer temperatures. For example, relatively light levels of tree mortality from bark beetles occurred in an area near the headquarters of the Kenai National Wildlife Refuge that burned in 1926, while a nearby stand of spruce 125 years old or older experienced heavy mortality. Many stands on the southern Kenai Peninsula were composed of trees that regenerated following a moderate spruce bark beetle outbreak in the 1870s and 1880s, and those trees only matured to a condition that would support substantial spruce bark beetle infestations in the mid-20th century. By the time that warm summers began in 1987, these trees had once again become prime bark beetle habitat.

Moisture appears to be another important factor in spruce bark beetle outbreaks on the Kenai Peninsula. Precipitation at the Kenai Airport station reached a record low from 1967 to 1969, and was at a near-record low (to that date) at the Homer Airport station during that period. This dry period was followed by an intense but short-lived spruce bark beetle outbreak from 1970 to 1974 (Fig. 14.26c,d). It appears that at least two warm summers are required to initiate a spruce bark beetle outbreak. This is consistent with the life cycle of the beetles, which normally takes two years to complete, except in the warmest summers when it can be completed in one year (Werner and Holsten, 1985). In 1987, summer temperatures shifted above the mean (Fig. 14.26a), beginning a sustained series of warm summers that is unique in both the instrumental and local reconstructed temperature records. After 1987, the spruce bark beetle outbreak accelerated, and no period of several cool summers, which could thermally arrest the outbreak, has occurred to the present. Red-needle acreage, however, has dropped steadily since 1996 because most of the mature spruce forest has been killed. Food supply, rather than climate, appears to have been the ultimate limiting factor in this outbreak, unlike outbreaks during the past 200 to 250 years studied by staff of the Kenai National Wildlife Refuge.

These results suggest that there is a continuing high risk from climate change to the management of forest land in south-central Alaska for spruce forest crops. Under recent climate conditions, and especially under scenarios of further temperature increases, spruce bark beetle irruption potential is very likely to remain high. As the small surviving (understory) spruce trees in the region mature to commercial forest dimensions, they will move into the prime size and age classes to serve as hosts for spruce bark beetle. Under these circumstances, the regional environment is very likely to remain effectively saturated with spruce bark beetles because climate limitations on beetles have been removed. Investments in regeneration and early tending of new commercial stands of spruce, should that be desired, would carry considerable risk because bark beetles would become effective agents of tree mortality at about the time that stands of spruce became large enough to generate commercial value.

14.8.3. Spruce budworm in North America

The eastern spruce budworm (*Choristoneura fumiferana*), is a defoliating insect affecting conifer trees, generally in the southern and central boreal region of the United States and Canada. The absolute distribution limits of spruce budworm generally follow that of white spruce throughout the boreal forest, and there are records of severe defoliation within 150 km of the Arctic Circle (Harvey, 1985). Outbreaks have at times extended over 72 million ha and lasted for up to 15 years (Fleming and Volney, 1995). The budworm is generally present in the forest at low background numbers. During irruptions, there can be more than 22 million budworm larvae per hectare in suitable forest stands (Crawford H. and Jennings, 1989).

Weather appears to be a critical factor in determining budworm distribution. Irruptions of budworm generally follow drought, and outbreaks also start after hot, dry summers (Fleming and Volney, 1995). Drought stresses the host tree population, reducing host resistance (Mattson and Haack, 1987). Elevated summer temperatures increase budworm reproductive output. Female budworms lay 50% more eggs at 25 °C than at 15 °C (Jardine, 1994). Finally, higher temperatures and drought can shift the timing of budworm reproduction so that natural parasitoid predators are no longer effective in limiting budworm numbers (Mattson and Haack, 1987). Conversely, cold weather can stop a budworm outbreak. Budworms starve if a late spring frost kills the new shoot growth of the host trees on which the larvae feed.

Given this weather/climate sensitivity of budworm, it was inferred that a warming climate would be associated with northward movement of spruce budworm outbreaks (Gitay et al., 2001). Such a northward movement appears to have happened. Before about 1990, spruce budworm had not appeared able to reproduce in the northern boreal forest of central Alaska. In that year, after a series of warm summers, a spruce budworm irruption occurred in the Bonanza Creek Long-Term Ecological Research site and visible canopy damage spread over tens of thousands of hectares of nearby white spruce forest. Populations of budworm have persisted in this area near the Arctic Circle, including a minor outbreak in 2002–2003.

14.8.4. Other forest-damaging insects in North America

Aspen is the most important deciduous tree species in the Canadian boreal forest, with more than 1000 Tg of carbon stored in the aboveground biomass. Aspen dieback has become conspicuous over parts of the southern boreal forest and aspen parkland in western Canada. In 18 aspen stands near Grande Prairie, Alberta, defoliation histories were reconstructed based on tree rings and records of past insect outbreaks. Several factors contributed to the observed dieback. Defoliation by the forest tent caterpillar (*Malacosoma disstria*) and drought in the 1960s and 1980s led to reduced growth and predisposed some stands to secondary damage by wood-boring insects and fungal pathogens. Thaw–freeze events during a period of unusually light late winter snow cover (1984–1993) also contributed to the observed dieback. Under climate warming, the severity of these stressors is very likely to increase, which would pose a serious concern for the future health, productivity, and carbon sequestration of aspen forests in the region (Hogg et al., 2002).

During the last two decades of the 20th century, warmer summers and winters in central Alaska were associated with a noticeable increase in the area of trees killed or damaged by insects. The mapped area of forest affected by spruce budworm, spruce coneworm (*Dioryctria reniculelloides*), and larch sawfly (*Pristiphora erichsonii*) defoliation mapped throughout Interior Alaska increased, totaling over 300 000 ha of combined infestations during the period 1991 to 1996 (Holsten and Burnside, 1997).

Various insects are known have the potential to become serious agents of tree mortality in a much warmer boreal and arctic region. Insect species not currently present are almost certain to be able to disperse readily into the region if higher temperatures allow, and some are likely to develop outbreak potential. In Ontario, Canada, forest land managers estimate that a range of forest-damaging insects that are currently prominent in the central and southern boreal area would move northward or, if already present, develop greater outbreak potential (Parker et al., 2000). These include spruce budworm, forest tent caterpillar, jack pine budworm (*Choristoneura pinus pinus*), and gypsy moth (*Lymantria dispar*). The bronze birch borer (*Agrilus anxius*) is a North American species that can cause severe damage to paper birch and may be effective in limiting the survival of birch along the southern (warm) margin of its distribution (Haak, 1996). The bronze birch borer is present today in small numbers as far north as Alaska. The ACIA-designated models project temperatures in northern boreal North America near the temperatures of areas where the bronze birch borer effectively limits birch today. In boreal Alaska, a larch sawfly outbreak killed most of the larger and older tamarack (*Larix laricina*) trees during a warm period in the decade of the 1990s, and aspen leaf miner (*Phyllocnistis populiella*) appeared at outbreak levels (142 000 ha) by 2003 (Wittwer, 2004).

14.8.5. Tree-damaging insects in northern Europe

Several insect species with tree hosts in northern European boreal forests regularly undergo population outbreaks, and most have some connection to direct or indirect effects of warm weather anomalies and temperature limitation. In Europe, temperature increases, especially in the form of warmer winters, are associated with increased outbreaks of various species of bark beetles and aphids (Beniston et al., 1998). Bark beetles (Coleoptera, Scolytidae) are present in the boreal region of northern Europe. A key feature of temperature sensitivity is the timing of the first flight of mature adults in the spring as they seek new host trees to attack. Many beetles do not begin spring flight until air temperature reaches a threshold level. The bark beetle *Ips typographus* (spruce engraver beetle) built up in numbers in trees damaged by heavy storm and snow damage in the late 1960s, and then broke out in large numbers during a series of warm years in the early 1970s (Heliövaara and Peltonen, 1999). A similar combination of widespread tree host susceptibility and warm summers in the future is very likely to result in similar outbreaks and tree death. There are no examples in Europe of bark beetles simultaneously expanding their distribution in one direction and retracting in another (Heliövaara and Peltonen, 1999).

Tree-defoliating insects that are responsive to temperature increases are capable of causing large-scale tree death and injury in the northernmost forests of Europe (Bylund, 1999; Danell et al., 1999). Geometrid moths of two species occur in the mountain birch (*Betula pubescens* ssp. *czerepanovii*) forest that makes up the ecotone between boreal forest and the tundra of Fennoscandia. The autumnal moth (*Epirrita autumnata*) and the winter moth (*Operophtera brumata*) regularly reach outbreak levels. Although both species lay eggs on a variety of trees and shrubs, the main host is mountain birch. Outbreaks of the autumnal moth mainly occur in mature forests in inland locations, whereas winter moth outbreaks are restricted to warm south-facing slopes or warmer locations along the Norwegian coast (Neuvonen et al., 1999).

Populations of both the winter and autumnal moth fluctuate from low to high levels with 7 to 12 years between peak densities (Altenkirch, 1990). Severe defoliation causes death of older stems, which usually triggers vigorous sprouting of new stems from the base of the tree. However, the recovery process in these forests is slow, and may take about a century (Bylund, 1999). Particularly severe defoliation may kill the underground portions of the tree as well. The natural cyclic fluctuation of geometrid moth numbers is driven by density-dependent factors that lag outbreaks in time, including buildup of natural enemies, disease, and reduced food quality, and by weather conditions at all life stages.
As do many of the macrolepidoptera, autumnal moths overwinter as eggs, making them vulnerable to extreme cold temperatures that reduce survival, particularly egg masses placed in the tree canopy above the snow limit. From 1960 to 1990, minimum winter temperatures below -36 °C, along with fine-scale patches of trees on dry and nutrient poor soils, efficiently predicted the distribution of autumnal moth outbreaks in birch forests of Finland (Neuvonen et al., 1999). Climate change involving an increase in minimum winter temperatures is likely to increase the frequency, and possibly the severity, of outbreaks of this species and others that overwinter as eggs (Neuvonen et al., 1999). Warmer summer temperatures with no change in winter temperatures are likely to reduce the geometrid moth outbreak potential, because higher densities of moth predators associated with warm summers would partially protect birch, while the winter limitation on the moths would still operate. Finally, if climate change consists of warmer summers and winters, as projected by the ACIA-designated models, current insect population/ecological models are not adequate to project overall effects on outbreaks (Neuvonen et al., 1999), although new insect distributions and novel climate limitations and outbreak patterns are very likely under such circumstances (see also section 7.4.1.4).

A diprionid sawfly (*Neodiprion sertifer*) is the most serious defoliator of pine forests in northern Europe. Outbreaks mostly occur on dry and infertile sites. Minimum winter temperatures below -36 ºC also limit this species. Outbreaks in Finland have been most frequent in southern and central inland areas, and temperature increases are likely to make outbreaks more common in eastern and northern areas (Neuvonen et al., 1999).

14.9. Climate change and fire

Fire is a major climate-related disturbance in the boreal forest, with pervasive ecological effects (Payette, 1992; Van Cleve et al., 1991; Zackrisson, 1977). Climate, disturbance, and vegetation interact and affect each other, and together they influence the rate and pattern of changes in vegetation (Neilson, 1993; Noble, 1993), the rate of future disturbance (Gardner et al., 1996; Rupp et al., 2002; Turner M. et al., 2003), and the pattern of new forest development (Turner D. et al., 1995; Rupp et al., 2000a,b). Understanding these interactions and feedbacks is critical in order to understand how scenarios of climate change will affect future fire regimes and the consequences these ecological changes will have for both boreal forests and forest management.

The total area burned in North America has been increasing concurrently with recent temperature increases and other climatic changes (Stocks et al., 2000). The annual area burned in western North America doubled in the last 20 years of the 20th century (Murphy et al., 2000). Based upon less precise statistics there appears to be a similar trend in the Russian Federation (Kasischke et al., in press).

Three factors are important when examining the impacts of climate change on the fire regime. First, boreal forests become increasingly flammable during succession as the forest floor and understory builds up. Conifer-dominated stands are usually more flammable than broad-leaved and other vegetation types. Early post-fire vegetation is generally less flammable than vegetation that is older and more structurally complex. Vegetation also influences fire probability indirectly through its effects on regional climate (Chapin et al., 2003), with early successional stands absorbing and transferring only half the incoming solar radiation compared to late successional spruce stands (Baldocchi and Vogel, 1995). A second factor is the need to understand the direct effects of climate on the fire regime. Finally, humans and their land-use changes affect the fire regime, and these changes must be considered to understand future fire effects in a warmer climate.

14.9.1. The role of fire in subarctic and boreal forest

Disturbance is the driving force behind vegetation dynamics in the boreal biome. Wildfires, alluvial processes (i.e., erosion and flood deposition), tree-fall events caused by wind, and tree-killing insect outbreaks all play major roles that can affect large portions of the landscape. However, fire is particularly significant because of its pervasive presence, its strong link to climate, and the direct feedbacks it has to permafrost dynamics, regional climate, and the storage and release of carbon. Fire often follows large-scale tree death caused by other factors.

In the mosaic of old burns that covers the boreal forest landscape, there are four tendencies related to ecosystem development, expressed over timescales of decades to millennia and over differing landscape areas. There is a tendency towards paludification, which is the water-logging and cooling of soil caused by the buildup of organic soil layers (Viereck, 1970). Paludification can be a significant factor in treeline retreat in maritime climates (Crawford R. et al., 2003). A second tendency is for the longer-lived conifer species to usurp the canopy from shorter-lived hardwood species (Pastor et al., 1999; Van Cleve et al., 1991) with an accompanying decline in living biomass (Paré and Bergeron, 1995). The third tendency is towards the formation of a shifting-mosaic steady state in which burned areas at different stages of secondary succession form the pieces of the mosaic (Wright, 1974). Finally, there is a tendency toward increased risk of burning with age. Fire opposes paludification by oxidizing the organic matter buildup on the ground surface (Mann D. and Plug, 1999). Fire also interrupts conifer take-over by restarting successional development with the earliest broadleaf-dominated stages. In addition, because the time since the last fire is an important determinant of vegetation composition and soil conditions, fire frequency has a large influence on carbon storage and release within the boreal forest (Kasischke et al., 1995; Kurz and Apps, 1999).

Fire is strongly controlled by both temporal and spatial patterns of weather and climate (Flannigan and Harrington, 1988; Flannigan et al., 2001; Johnson, 1992). Specific fire behavior responds to hourly, daily, and weekly weather conditions. Solar radiation, continentality, topography, and specific terrain features influence these fire-generating conditions. The general fire regime of an area responds to long-term, landscape-level climatic patterns. As climate varies, the controlling weather variables can vary in magnitude and direction (Flannigan et al., 2001).

Case studies from northern coniferous forests have documented weather variables prior to and during specific fire events (Flannigan and Harrington, 1988) and have identified good predictors of conditions that promote the rapid spread of fire. These conditions include warm temperatures, little or no precipitation, low relative humidity, and high winds. Synoptic weather conditions that produce cold frontal systems, drought, and low relative humidity have been found to be good predictors of area-burned activity (Flannigan et al., 2001). Upper-air circulation patterns have been related to area burned (Newark, 1975). Catastrophic burning events are related to the breakdown of the 500 mb long-wave ridge

(Nimchuk, 1983). Breakdown of these ridges generally occur at the same time as documented increases in lightning strikes and strong surface winds.

Regional and global links (teleconnections) between meteorological variables and weather anomalies have been identified. The most documented sources of teleconnections are the El Niño–Southern Oscillation and the Pacific Decadal Oscillation (section 2.2.2.2). Several North American studies have linked teleconnections to area-burned anomalies (Flannigan et al., 2000; Johnson and Wowchuk, 1993; Swetnam and Betancourt, 1990). Teleconnections may offer long-range forecasting techniques for temperature and precipitation anomalies that fire managers could use in estimating fuel moisture and fire potential (Flannigan et al., 2001). Winter sea surface temperatures in the Pacific are significantly correlated with warm-season temperature and seasonal area burned in Canada (Flannigan et al., 2000). Significant correlations between these factors vary by provincial region and phase of the North Pacific Oscillation (1953–1976 versus 1977–1995). Interestingly, the sign of the correlation changed from strongly negative (1953–1976) to strongly positive (1977–1995) for four regions (Alberta, western Ontario, eastern Ontario, and Quebec) at the same time they changed from strongly positive to strongly negative for two regions (British Columbia and Saskatchewan). During the same period, two regions did not change sign (Yukon/Northwest Territories and Manitoba). These shifts occurred at previously identified changes in climate regimes (e.g., section 14.6.2, Fig. 14.15), and emphasize the control the overall atmospheric circulation has on fire at a regional scale in understandable, but varying ways.

There are two different scenarios of the relay floristics (rate of spread of vegetation; *sensu* Egler, 1954) pathway for secondary succession after fire that operate in the boreal forest. The first is the process in which deciduous shrubs and trees initially colonize burned sites, but are replaced in the overstory in approximately 150 to 200 years by coniferous trees that may be limited in movement by the time necessary for successive generations to reach reproductive maturity or the presence of individual surviving trees dispersed throughout the burn. The second process is self-replacement, which occurs after fires by root sprouting (i.e., aspen, birch, balsam poplar (*Populus balsamifera*), and numerous shrub species) or by fire-stimulated seed release (i.e., black spruce) (Greene et al., 1999; Mann D. and Plug, 1999). Both of these successional processes occur in the boreal forest. However, the relative importance of each pathway in determining the structure of the boreal forest at a landscape scale is still unknown.

The fire regime describes the general characteristics of fire and its effects on ecosystems over time. It can be defined by specific components such as frequency, intensity, severity, size, and timing. These characteristics have been used to develop classification systems that aim to describe the principal types of fire regimes associated with different ecosystems. These fire/ecosystem regime categories are general and broad due to the large spatial and temporal variability exhibited by specific ecosystems (Whelan, 1995), but they do provide a conceptual model useful for understanding both fire behavior and fire effects in a particular system. Documentation of the components of fire regimes (i.e., ignition sources, frequency, extent, and severity) for specific ecosystems and geographic regions is also highly variable, and in many cases not well quantified.

Lightning strikes and humans are the two sources of ignitions in the boreal forest. Lightning is the most significant cause of fires (defined by total area burned), although this trend varies among regions. The number of lightning-caused fires generally declines as latitude increases because of decreased heating at the ground surface necessary to produce convective storms, and as climate becomes more maritime (cool layer at ground surface) (Johnson, 1992). Humans are responsible for high numbers of ignitions, but the fires started consume much less area because usually they are actively suppressed. Indigenous peoples ignited fires for specific purposes throughout history, but their impact on past overall fire regimes remains uncertain (Johnson, 1992; Swetnam et al., 1999). Specific regions of the boreal forest in North America experienced substantial anthropogenic fire impacts during the “gold rush” era of the late 19th and early 20th centuries.

14.9.2. Regional fire regimes

14.9.2.1. Russia

Fire statistics for the Russian Federation in general, and the Russian boreal region specifically, are incomplete at best (FAO, 2001; Stocks, 1991). Official fire statistics have been reported only for protected regions of Forest Fund land (section 14.3.1). Furthermore, only 60% of the Forest Fund land is identified as protected (FAO, 2001). Approximately 430 million ha of forested tundra and middle taiga in Siberia and the Far East receive no fire protection. The paucity of reliable statistics is a result of numerous issues including remoteness, lack of detection and mapping technology, lack of fire-management funding, and deliberate falsification of past records for political reasons. Humans were identified as the major source of ignitions at 65%, followed by lightning (16%), prescribed agricultural burning (7%), and other/unknown activities (12%) (Shetinsky, 1994). Intensive prevention and education programs have had little success in decreasing anthropogenic fires. Many of these fires grow to large sizes due to overstretched suppression resources.

Keeping these limitations in mind, the statistics for the protected Forest Fund area provide some perspective on fire across the Russian Federation. Between 18 000 and 37000 forest fires were detected annually from 1950 to 1999. The average annual area burned within the zone of detection for the decades 1950 to 1959, 1960 to 1969, 1970 to 1979, 1980 to 1989, and 1990 to 1999 was 1.54, 0.68, 0.48, 0.54, and 1.2 million ha, respectively (FAO, 2001). Indirect estimates of annual area burned in both protected and unprotected Forest Fund areas have been developed through modeling techniques. Shvidenko and Goldammer (FAO, 2001) used a modified expert system model (Shvidenko and Nilsson, 2000), available fire statistics, and forest inventory data on age and stand structure to calculate an estimated total of the annual average area burned in all of Russia over the past 30 years. The estimate applies to the Forest Fund and State Land Reserve area. An estimated 5.10 million ha burned annually, of which 3.94 million ha were in the boreal bioclimatic zones of forest tundra–northern taiga, middle taiga, and southern taiga (see Fig. 14.1).

Satellite remote-sensing techniques provide insight into the potential total extent of area burned across major regions of the Russian Federation. In 1987, an estimated 14.4 million ha burned in the Russian Far East and eastern Siberia (Cahoon et al., 1994), and in 1992, an estimated 1.5 million ha burned in all of the Russian territories (Cahoon et al., 1996). In 1998, following a very strong El Niño, an estimated 9.4 million ha burned in the Asian regions of Russia (FAO, 2001).

Only a small fraction of total forested area falls under any organized fire-suppression management. Fire suppression is headed by the Ministry of Natural Resources of Russia, which manages suppression efforts through regional offices. Fire-suppression resources include both ground and aerial operations. Aerial operations provide detection and monitoring services, and direct suppression resources (i.e., water and retardant drops, transport of ground personnel and smokejumpers). The State Forest Guard (ground suppression) and the Avialesookhrana (aerial detection and suppression) coordinate fire suppression with local and regional authorities. Under severe fire conditions, military detachments and local populations are recruited.

Operational policy and logistical allocation follow fire danger predictions, which are based on weather and climatic conditions. The Nesterov fire index, similar to the Canadian Forest Fire Danger Rating System or the US National Fire Danger Rating System, is used for prediction. Aircraft patrols and resource deployments are based on the fire index predictions. A severe limitation to successful fire management has been a lack of funding – in 1998, only US$ 0.06 per hectare was allocated to fire suppression (FAO, 2001). In addition, a lack of advanced technology equipment (i.e., satellite monitoring, radios, etc.), aircraft, and State Forest Guard personnel, as well as unfavorable land-use practices, have been identified as major weaknesses in the current system.

14.9.2.2. Canada

Canadian fire management agencies developed a large-fire database for all fires larger than 200 ha for the period 1959 to the present (Stocks et al., 2000). Fires larger than 200 ha represent only 3.5% of the total number of fires, but 97% of the total area burned over this period (Stocks et al., 2000). The fire perimeters were digitized and mapped in a geographic information system, and the database includes ancillary information such as ignition location and date, size, cause, and suppression action(s) taken. This database has been expanded into the past where data exist (as far back as 1918), and is continuously updated with each passing fire season. Total area burned in Canada has more than doubled since the 1970s, and the upward trend is well explained by warmer temperatures (Gillett et al., 2004).

Lightning-caused fires predominate throughout the fire record and account for almost all large fires in the northern portions of Canada. From 1959 to 1997, lightning ignited approximately 68% of all fires, and those fires accounted for approximately 79% of the total area burned (Stocks et al., 2000). Temporal trends show a steady increase in the number of lightning fires and their contribution to total area burned from the 1960s through the 1990s, which fire managers attribute to technology improvements in fire detection and monitoring. Anthropogenic fires are a significant contributor to the total number of fires almost exclusively in populated areas and along the road network, and suppression of these fires is effective in limiting the total area burned. Temporal trends show a steady decrease in total area burned due to anthropogenic ignitions, attributed to aggressive fire suppression tactics.

The average annual area burned between 1959 and 1997 was approximately 1.9 million ha, with interannual variability ranging from 270000 to 7.5 million ha burned (Stocks et al., 2000). Spatial trends identified a few ecozones (taiga plains, taiga shield, boreal shield, and boreal plains; see Wiken, 1986 for definitions) that accounted for the majority (88%) of total area burned between 1959 and 1997, primarily because these ecozones have a continental climate that is conducive to extreme fire danger conditions and have large uninhabited areas with low values-at-risk, so fires are allowed to burn unimpeded (Stocks et al., 2000).

Organized fire suppression and management has operated in Canada since the early 1900s, initiated as a response to large catastrophic fires, much like those experienced in the United States. Canada is recognized worldwide for its advanced technologies and operational efficiencies in both fire prevention and suppression. The Canadian Forest Fire Danger Rating System (CFFDRS) was first developed in 1968 and consists of two subcomponents: the Canadian Forest Fire Weather Index (FWI) system, which provides a quantitative rating of relative fire potential for standard fuel types based upon weather observations; and the Canadian

Forest Fire Behavior Prediction system, which accounts for fire behavior variability of fuel types based on topography and components of the FWI. The CFFDRS is used for training, prevention, operational planning, prescribed burning, and suppression tactics. In addition, researchers use CFFDRS for investigating fire growth modeling, fire regimes, and potential climate change impacts. The system is used extensively in Alaska in place of the US Fire Danger Rating System – a system developed primarily for the lower 48 states.

Aerial reconnaissance in conjunction with lightning detection systems has been employed heavily since the 1970s throughout the North American boreal forest (including Alaska). The lightning detection system identifies areas of high lightning strike density and allows for focused aerial detection operations. This has greatly increased detection efficiencies over remote regions.

14.9.2.3. United States (Alaska)

The fire record for the Alaskan boreal forest has evolved over time and consists of three datasets (Kasischke et al., in press). The first is a tabular summary of total annual area burned since 1940. The second is a tabular database that contains the location, ignition source, size, management option, and initiation and extinguishment date for all fires since 1956. The third is a geographic information system database of the boundary of fires since 1950. The spatial database includes all fires larger than 400 ha occurring from 1950 to 1987, and all fires larger than 25 ha occurring since 1988.

Early studies minimized the importance of lightning as an ignition source in the Alaskan boreal forest because anthropogenic ignitions were thought to be more important (Lutz, 1956). However, the implementation of digital electronic lightning detection systems along with a more thorough review of fire statistics led to the realization that lightning is not only widespread throughout the growing season in Interior Alaska, but is responsible for igniting the fires that burn most of the area (Barney and Stocks, 1983; Gabriel and Tande, 1983). Analysis of fire statistics from the Alaska Fire Service shows that while humans start more than 61% of all fires, these fires are responsible for only 10% of the total area burned (Kasischke et al., in press). The remaining fires are the result of lightning ignitions.

Convective storms and associated lightning can range in size from individual clouds to synoptic thunderstorms covering thousands of square kilometers. A single thunderstorm may produce most of the annual lightning strikes in an area (Nash and Johnson, 1996). In Interior Alaska, the lightning density gradient generally runs from high in the east to low in the west, parallel with the warmer summer climate in the interior continental areas and cooler maritime summer climate near the coast (Kasischke et al., in press). However, mapped fires that originate from lightning strikes are well distributed throughout the interior of the state. In contrast to lightning ignitions that are uniformly distributed between the Brooks and Alaska Ranges, anthropogenic fire ignitions are centered around major population centers (Fairbanks and Anchorage), as well as along the major road networks (Kasischke et al., in press).

The presence of the boreal forest landscape may promote convective thunderstorms. In central Alaska, the density of lightning strikes is consistently highest within boreal forest compared to tundra and shrub zones across a climatic gradient (Dissing and Verbyla, 2003). Within the tundra, the number of lightning strikes increases closer to the boreal forest edge. The paleo fire record indicates that wildfires increased once black spruce became established in Interior Alaska (despite a cooler, moister climate). This may have been due to two factors: 1) increased landscape heterogeneity and higher sensible heat fluxes (see section 14.2.5 BOREAS results) leading to increased convective thunderstorms; and 2) increased fuel flammability associated with the black spruce vegetation type (Chambers and Chapin, 2003; Dissing and Verblya, 2003; Kasischke et al., in press; Lynch et al., 2003). These results suggest that climate changes that promote the expansion or increase in density of conifer forest are likely to increase the incidence of fire, but that climate changes that decrease conifers (too-frequent burning, aridification) are likely to decrease subsequent burning.

The annual area burned in Alaska exhibits a bimodal distribution, with years of high fire activity (ignitions and area burned) punctuating a greater number of years of low activity. For the Alaskan boreal forest region, 55% of the total area burned between 1961 and 2000 occurred in just 6 years. The average annual area burned during these episodic fire years was seven times greater than the area burned in the low fire years.

With the observed increase in air temperature and lengthening of the growing season over the past several decades in the North American boreal forest region, a corresponding increase in fire activity between the 1960s and 1990s might be expected (Stocks, 2001). For the Alaskan boreal forest region (which was less affected by fire suppression efforts than southern boreal areas), such an increase in fire activity was not apparent until the summer of 2004, in which a record 2.71 million ha burned. That summer (May–August) also had the highest mean monthly temperatures recorded at Fairbanks since observations began in 1906. A record 1.84 million ha also burned in 2004 in the adjacent Yukon Territory, amounting to about 12.4% of all forest land in the Yukon. Between 1981 and 2000 there was only a slight (7%) rise in the annual area burned (297624 ha/yr) compared to the period between 1961 and 1980 (276624 ha/yr) (Kasischke et al., in press). These extensive 2004 fires have clearly established a significant upward trend in the area burned during the period of record. The frequency of large fire years has been greater since 1980 (five large fire years) than before 1980 (two large fire years), and the increase in average annual area burned between these periods is due to the increase in frequency of large fire years and the decrease in fire-fighting activities in remote areas. The fire data record for Alaska is consistent with an increase in large fires in response to recent climate warming, but not sufficient to determine definitively whether the increase is outside the range of natural variability. The official fire statistics for Alaska show that there were two large fire years per decade in the 1940s and 1950s, indicating that the frequency of large fire years has been relatively constant over the past 60 years, although this entire period is distinctly warmer than preceding centuries.

The number and size distribution of fires in the Alaskan boreal forest region is different during low and severe fire years. During low fire years between 1950 and 1999, an average of 17 fires greater than 400 ha occurred per year, with an average size of 7800 ha. In contrast, during high fire years an average of 66 fires greater than 400 ha occurred, with an average size of 20 300 ha (Kasischke et al., in press). In low fire years, 73% of the total area burned occurred in fires larger than 50 000 ha. In high fire years, 65% of the total area burned occurred in fires larger than 50 000 ha. In low fire years, 9% of the total area burned in fires larger than 100 000 ha, with no fires larger than 200 000 ha. In contrast, during high fire years, 33% of the total area burned in fires larger than 100 000 ha.

Fire managers in Alaska typically recognize two fire seasons (Kasischke et al., in press). The major fire season typically occurs after mid-June when the surface of the earth becomes warm enough to drive convective thunderstorm activity. However, there is an earlier fire season throughout the state associated with the extremely dry fuel conditions that occur immediately after snowmelt and before leaf-out in late spring. At this time, precipitation levels are low, and fuel moisture conditions are extremely low because of the curing of dead vegetation during winter and the lack of green vegetation. During this period, human activities result in the majority of fire ignitions.

Fire management and suppression is a relatively new concept in Alaska and did not become an organized effort until 1939. In the early 1980s, fire-management efforts in Alaska were coordinated under the direction of an interagency team. Beginning in 1984, fire-management plans were developed for thirteen planning areas. The fire plans provide for five separate suppression options:

- *Critical protection* is provided for areas of human habitation or development. These areas receive immediate initial attack and continuing suppression action until the fire is extinguished.
- *Full protection* is provided for areas of high resource values. Fires receive immediate initial attack and aggressive suppression to minimize area burned.
- *Modified action* provides for initial attack of new fires during the severe part of the fire season (May 1–July 10). Escaped fires are evaluated by the land manager and suppression agency for appropriate suppression strategies. On specified dates (generally after July 10), the modified action areas convert to limited action.
- *Limited action* is provided for areas of low resource values or where fire may actually serve to further land management objectives. Suppression responses include the monitoring of fire behavior and those actions necessary to ensure it does not move into areas of higher values.
- *Unplanned areas* exist where land managers declined participation in the planning process. These few unplanned areas receive the equivalent of full protection.

The formation of these plans radically changed wildland fire management in Alaska. The adoption of these cooperative plans created areas where fires were no longer aggressively attacked due to the low economic resource value of those lands. The primary goals of the plans are to restore the natural fire regime to the boreal forest ecosystem and to reduce suppression costs.

14.9.2.4. Fennoscandia

Finland, Sweden, and Norway account for a very small percentage of total area burned in the boreal forest and generally have not experienced large fires in modern times. This can be attributed to the ease of access throughout these relatively small countries and their highly managed forest systems. In addition, lightning fires are not a major factor, accounting for less than 10% of all fires (Stocks, 1991). Fire statistics for this region of the boreal forest are limited and discontinuous. The lack of natural fire is one of the principal causes of endangerment of a set of fire-dependent species in northern Fennoscandia (Essen et al., 1992).

Finland reported an annual average of 800 ha burned during the 1990s (FAO, 2001). Since record keeping began in 1952, there has been a steady decrease in average annual area burned from a high of 5760 ha in the 1950s to 1355 ha in the 1960s to less than 1000 ha from 1971 to 1997 (FAO, 2001). An average of 2224 fires per year occurred during the period 1995 to 1999. Anthropogenic fires accounted for 61% of all fires, followed by 29% of unknown cause, and 10% caused by lightning (FAO, 2001).

Boreal forest covers most of Sweden and includes a mix of flammable conifer trees, shrubs, and mosses. Sweden reported an annual average of 1600 ha burned during the 1990s (FAO, 2001). Humans are the major cause of fires accounting for approximately 65% (FAO, 2001) of the annual average of 3280 fires during the 1990s (FAO, 2001). Norway reported an average of 564 ha burned annually between 1986 and 1996 (FAO, 2001). The highest frequency of fires occurs in the boreal forest region of the eastern lowlands (Mysterud et al., 1998). On average, there were only 513 fires per year in Norway during the 1990s (FAO, 2001).

Forest fires are virtually absent from the sparsely forested regions of Iceland and Greenland. No fire statistics were reported to the FAO for the period 1990 to 1999.

14.9.3. Possible impacts of climate change on fire

Rupp et al. (2000a,b) used a spatially explicit model to simulate the transient response of subarctic vegetation to climatic warming in northwestern Alaska near treeline. In the model simulations, a warming climate led to more and larger fires. Vegetation and fire regime continued to change for centuries in direct response to a 2 to 4 °C increase in mean growing-season temperature. Flammability increased rapidly in direct response to temperature increases and more gradually in response to simulated climate-induced vegetation change. In the simulations, warming caused as much as a 228% increase in the total area burned per decade, which led to a landscape dominated by an increasingly early successional and more homogenous deciduous forest.

Turner M. et al. (2003) used the same model (Rupp et al., 2000b) to simulate fire-regime sensitivity to precipitation trends. Precipitation projections from global climate models have the largest associated errors and the highest variability between models. A simulated instantaneous 2 °C increase in average growing-season temperature was applied with two different precipitation regimes (a 20% increase and a 20% decrease from current precipitation levels) to explore the possible influence of climate change on long-term boreal forest ecosystem dynamics. Both scenarios projected an increase in the total number of fires compared to current climate. However, the distribution of fire sizes was surprising. As expected, the warmer and drier scenario resulted in fewer small fires and an associated shift toward larger fires. Also as expected, the warmer and wetter scenario resulted in fewer large fires and an associated shift toward smaller fires. However, the distribution of very large fires (burning >25% of the total landscape area) was unexpected. The warmer and wetter scenario produced more very large fires compared to the warmer and dryer scenario. The warmer and dryer climate scenario experienced frequent medium-sized fires, which prevented fuels from building up across the landscape and limited the number of large fires. In contrast, the warmer and wetter climate scenario led to frequent small fires, which allowed the development of well-connected, highly flammable late-successional stands across the landscape.

Additional model simulations suggest that vegetation effects are likely to cause significant changes in the fire regime in Interior Alaska (Rupp et al., 2002). Landscapes with a black spruce component were projected to have more fires and more area burned than landscapes with no black spruce component. Black spruce landscapes were also projected to experience numerous fires consuming extensive portions (more than 40%) of the landscape. These results agree with observations in the Canadian boreal forest where 2% of the fires account for 98% of the total area burned (Stocks, 1991). Large-scale fire events need to be realistically represented in ecosystem models because they strongly influence ecosystem processes at landscape and regional scales. These results have strong implications for global-scale models of terrestrial ecosystems. Currently, these models consider only plant functional types distinguished by their physiological (C_3 versus C_4 photosynthesis), phenological (deciduous versus evergreen), and physiognomic (grass versus tree) attributes. The fire-modeling results suggest the need for a finer resolution of vegetation structure and composition related to flammability in order to simulate accurately the dynamics of the fire regime, and to understand how different climate changes are likely to change ecosystems at several different scales.

Modeling results show that fire regime plays an important role in determining the overall relative abundance of ecosystem types at any given time, and that different fire regimes create qualitatively different patterns of ecosystem placement across the landscape (Rupp et al., 2000b, 2002). One process is "contagion" or the spread of effects into one type largely due to its juxtaposition with another. Simulation results in a typical Alaska black spruce landscape projected increases in the total amount of deciduous forest that burned in two landscapes that contained a black spruce component. The results indicated a much shorter fire return interval for deciduous forest in a landscape dominated by black spruce, and the fire return times were similar to actual fire intervals calculated by Yarie (1981) for deciduous forest in Interior Alaska (Rupp et al., 2000b, 2002). Landscape-level changes in the fire return interval of specific fuel types is an important effect of spatial contagion that currently cannot be addressed by statistical formulation within a global vegetation model. Although there is an excellent quantitative understanding of fire behavior as a function of climate/weather and vegetation at the scale of hours and meters (Johnson, 1992), the dynamic simulation of fire effects at landscape or regional scales remains rudimentary (Gardner et al., 1999). A long-term potential consequence of intensified fire regimes under increasing temperatures is that fire-induced changes in vegetation are likely to lead to a more homogenous landscape dominated by early-successional deciduous forest (Rupp et al., 2000b, 2002). This has significant implications for the regional carbon budget and feedbacks to climate (section 14.10.2). A shift from coniferous to deciduous forest dominance is very likely to have a negative feedback to temperature increases due to changes in albedo and energy partitioning (Chapin et al., 2000).

Stocks et al. (1998) used outputs from four GCMs to project forest-fire danger in Canada and Russia in a warmer climate. Temperature and precipitation anomalies between runs forced with current atmospheric CO_2 concentrations and those forced with doubled atmospheric CO_2 concentrations were combined with observed weather data for 1980 to 1989 in both Canada and Russia. All four models projected large increases in the areal extent of extreme fire danger in both countries under the doubled CO_2 scenarios. A monthly analysis identified an earlier start date to the fire seasons in both countries and significant increases in total area experiencing extreme fire danger throughout the warmest months of June and July. Scenarios are still of limited use, however, in projecting changes in ignitions.

Model projections of the spatial and temporal dynamics of a boreal forest under climate change were made for the Kas-Yenisey erosion plain in the southern taiga of western Siberia (Ter-Mikaelian et al., 1991). This study projected that the number of years in which there are severe fires would more than double under a summer temperature increase from 9.8 to 15.3 °C, the area of forests burned annually would increase by 146%, and average stored wood mass would decrease by 10%.

Flannigan and Van Wagner (1991) also investigated the impact of climate change on the severity of the forest fire season in Canada. They used projections from three GCMs forced with doubled atmospheric CO_2 concentrations to calculate the seasonal severity rating across Canada. Their results suggest a 46% increase in seasonal severity rating, with a possible similar increase in area burned, in a doubled CO_2 climate.

Flannigan et al. (1998) looked at future projections of wildfire in circumpolar boreal forests in response to scenarios of climate change. Simulations were based on GCM outputs for doubled atmospheric CO_2 concentrations. The simulation and fire history results suggested that the impact of climate change on northern forests as a result of forest fires may not be severe as previously suggested, and that there may be large regions of the Northern Hemisphere with a reduced fire frequency. These simulation results are attributed to a switch from using monthly to using daily GCM output. The scenarios still produced areas where the interval between fires is likely to decrease (i.e., more frequent fires), but they also produced regions of no change or with greater probability of an increasing interval (i.e., less frequent fires).

14.10. Climate change in relation to carbon uptake and carbon storage

14.10.1. The role of the boreal forest in the global carbon cycle

Within the terrestrial biosphere, forests cover 43% of the land area but are potentially responsible for 72% of the annual net primary productivity (McGuire et al., 1995b). The boreal forest covers roughly 1.37 billion ha and by itself (not including high-latitude tundra) contains approximately 20% of global reactive soil carbon, an amount similar to that held in the atmosphere (IPCC, 2001; Schlesinger, 1997). Climate change can affect high-latitude carbon cycling at multiple timescales. The most likely mechanism for significant short-term change in boreal carbon cycling resulting from climate change is a change in rates of organic matter decomposition in the forest floor and mineral soil resulting from major changes in species composition caused by alteration of disturbance regimes. Climate change can also strongly affect rates of carbon cycling through its control of the disturbance regime and the subsequent successional development of ecosystems (Barr et al., 2002; Gower et al., 2001; Jiang et al., 2002; Trumbore and Harden, 1997; Wang et al., 2001; Zimov et al., 1999). Climate-induced shifts in dominant tree species composition within the present boreal forest (Carcaillet et al., 2001; Hogg and Hurdle, 1995; Smith T. et al., 1992) are likely to have profound impacts on the global carbon budget (Gower et al., 2001; Kasischke et al., 2002).

Boreal and subarctic peatlands contain approximately 455 Pg of carbon accumulating at an average rate of 0.096 Pg/yr, and constitute a significant proportion of the total boreal carbon pool (Gorham, 1991). The majority of peat consists of molecules that are highly resistant to degradation (e.g., lignin and cellulose). Species composition in sphagnum bogs is highly resilient (likely to remain the same or recover after change) because the mosses modify the local environment to produce highly acidic conditions, and their resiliency increases with age (Kuhry, 1994).

14.10.2. The role of disturbance in the carbon cycle of the boreal forest

Four processes largely control the storage and release of carbon in boreal forests: the rate of plant growth; the rate of decomposition of dead organic matter; the rate of permafrost accretion and degradation; and the frequency and severity of fires (Kasischke et al., 1995). All four processes are affected by landscape-scale disturbance. Differences in carbon cycling between mature and recently disturbed forest ecosystems have been observed in both experimental studies and modeling experiments. Some studies (Arain et al., 2002; Valentini et al., 2000) suggest that the annual carbon budget of the mature northern forest is at equilibrium and in some cases losing carbon to the atmosphere. In addition, model results and field experiments show that when ecosystems are disturbed, significant losses of soil carbon and nutrients can occur (Schimel et al., 1994) for a number of years after the disturbance.

The effects of temperature and disturbance (i.e., fire and grazing) on carbon exchange over three years in five undisturbed sites and five disturbed sites in forests of northeast Siberia were measured by Zimov et al. (1999) and results show that disturbance increased the seasonal amplitude of net carbon exchange. Disturbance had a larger effect on seasonal amplitude than either interannual or geographic differences in growing season temperature.

Fire affects the storage of carbon in the boreal forest in at least five ways: it releases carbon to the atmosphere;

converts relatively decomposable plant material into stable charcoal; re-initiates succession and changes the ratio of forest-stand age classes and age distribution; alters the thermal and moisture regime of the mineral soil and remaining organic matter, which strongly affects rates of decomposition; and increases the availability of soil nutrients through conversion of plant biomass to ash (Kasischke et al., 1995). Each of these effects exerts an influence at different timescales. As a result, the effect of a given climate scenario on carbon storage in the boreal region will be greatly influenced by fire regime (section 14.9), and represents a complex calculation requiring a great deal of specific spatial and temporal information.

Relatively few studies have directly examined carbon emissions in the boreal forest resulting from fire. Nonetheless, valuable contributions have been made by studies that used remotely sensed data to estimate direct carbon emissions from combustion as fires are occurring (Amiro et al., 2001; Isaev et al., 2002). These studies show that wildland fires have the potential to release a significant amount of carbon directly into the atmosphere. The effects of wildfire on this initial carbon loss are highly variable and strongly influenced by forest type and fire severity (Wang et al., 2001). Soil drainage (defined by the water table, moss cover, and permafrost dynamics) is the dominant control of direct fire emissions (Harden et al., 2000).

Fires cause the release of carbon not only during but also for a short time after fires. Tree mortality after surface fires can be extensive, leading to a pulse of carbon released from heterotrophic respiration as fine roots die and aboveground fine fuels (i.e., needles) fall to the ground and decompose rapidly (Conard and Ivanova, 1997). The non-combustion post-fire release of CO_2 has the potential to affect global levels of atmospheric CO_2 over the short term, representing another mechanism by which the boreal forest can play a significant role in the global carbon cycle. Together, these direct and indirect fire-generated carbon emissions from boreal forests worldwide may exceed 20% of the estimated global emissions from all biomass burning (Conard and Ivanova, 1997).

Wildland fires change the distribution of soil organic carbon pools with respect to their turnover (release to CO_2) times. Soil carbon in the form of forest litter that has fallen recently has a relatively rapid turnover rate, but carbon in the form of charcoal is stable (very long turnover rates). If rates of burning increase in the boreal forest, an increasing proportion of soil organic carbon is likely to be converted to stable charcoal. This change in the soil organic carbon pool allocation is difficult to estimate, because forest type and fire severity strongly influence the effects of wildfire on carbon redistribution (Wang et al., 2001). Harden et al. (2000) show the importance of this shifting distribution of turnover times in soil organic carbon. They developed a system of ordinary differential equations to explore constraints on carbon losses to fire, using modern estimates of carbon production, decomposition, and storage; a model of fire dynamics developed for millennial timescales; and an assessment of the long-term carbon balance of a variety of boreal landscapes in North America. A sensitivity analysis found that their model results were responsive to the rate at which charred plant remains decomposed. Unfortunately, the specific characteristics of fires that result in maximum charcoal production are not well studied. However, it seems reasonable to infer that a moderate-intensity fire, in which combustion is enough to kill trees but not intense enough to consume them completely, would produce the greatest amount of charcoal. In complex mountainous or hilly terrain at high latitudes, north-facing slopes have higher fuel moisture content and as a result generally experience less complete combustion of fuels during fire (Van Cleve et al., 1991). This suggests that slope aspect might be an important factor in the conversion of plant biomass to charcoal.

Much of the difference in carbon cycling after disturbance can be linked to shifts in species composition and ecosystem age structure that enhance both peak summer CO_2 uptake and winter CO_2 efflux. The seasonal amplitude of net ecosystem carbon exchange in northern Siberian ecosystems is greater in disturbed than undisturbed sites, due to increased summer influx and winter efflux (Zimov et al., 1999). Disturbed sites differ from undisturbed sites during the summer, having 2.1 to 2.5 times the daytime CO_2 influx and 1.8 to 2.6 times the nighttime CO_2 efflux. Winter respiration in disturbed sites is 1.7 to 4.9 times that in undisturbed sites. Carbon cycling within disturbed ecosystems is more sensitive to interannual temperature variability than older forests, and disturbed sites also experience a greater difference in annual carbon exchange with the atmosphere in warm versus cold years than older forests (Zimov et al., 1999).

Two hypotheses have been advanced to explain why these differences in carbon cycling caused by enhanced CO_2 summer uptake and winter efflux occur in disturbed areas. One hypothesis is that the recent increase in March and April temperatures in high-latitude continental regions of North America and Siberia has advanced snowmelt and lengthened the growing season (section 6.4). The second hypothesis is that temperature-driven increases in summer carbon gain (greater CO_2 uptake from greater growth) balanced by increased winter respiration (greater CO_2 release from enhanced decomposition and live respiration) could enhance the seasonal amplitude of atmospheric CO_2 concentrations without a change in net annual carbon accumulation. Although the mechanism remains somewhat uncertain, experimental studies confirm that these responses do occur under appropriate conditions.

14.10.3. Climate and carbon allocation in the boreal forest

Changes in species composition modify the way carbon is allocated and stored. Deciduous forests experience greater carbon cycling (production and decomposition) than coniferous stands (Gower et al., 2001). Both aboveground net primary production and overall (including belowground) net primary production were roughly two times greater in a boreal deciduous forest than a coniferous forest. The fraction of net primary production allocated to coarse- and fine-root primary production is roughly two times greater for evergreen conifers than deciduous trees (Gower et al., 2001). Because of these differential allocation patterns between deciduous and evergreen stands, the amount of carbon in the soil of mature black spruce stands is approximately three times the amount of carbon in the biomass of the trees (Kasischke et al., 1995). Since the rate of decomposition is higher in deciduous than in coniferous forests (because of both litter quality and site conditions), nitrogen is probably more available in deciduous forests, further increasing production (Makipaa et al., 1999). These results make sense physiologically, since deciduous species have higher maximum rates of photosynthesis and productivity than evergreens and produce litter that quickly decomposes.

Aboveground carbon pools are directly related to stand age. Gower et al. (2001) found that the effect of stand age (young versus old stands) on net primary production is roughly equivalent to the effect of soils (fertile versus infertile) and annual variation in the environment (favorable versus unfavorable weather). Similarly, Wang et al. (2001) found that wildfire exerts a lingering effect on carbon exchange between the boreal forest and the atmosphere via its effect on the age structure of forests and leaf-area index (LAI) during succession. They also found a strong inverse linear relationship between aboveground net primary productivity and age that was largely explained by a decline in LAI. Modeling results from Kurz and Apps (1999) suggested that forest ecosystems in Canada were a carbon sink from 1920 to 1980 and a source from 1980 to 1989. They suggested that this was a result of a change in the disturbance regime, and this finding is consistent with recent fire statistics (see section 14.9.2). Sometime around 1977, a regime shift in the climate of the North Pacific Ocean occurred that has been suggested to be part of a low-frequency oscillation (Niebauer, 1998). One of the consequences of this shift was that the position of the Aleutian Low associated with El Niño moved even farther eastward than it did in previous El Niño years. This shift is consistent with a more easterly (less southerly) flow component across Interior Alaska, which could exert teleconnective influences on the fire-dominated disturbance regime of Canada (Bonsal et al., 1993). If the key control over fire occurrence was a one-time climate regime shift caused by a change in sea surface temperature in the northeast Pacific Ocean, it suggests that climate changes occurring at low temporal frequencies exert a strong influence on the rate of carbon cycling in regions where the disturbance regime is climatically driven. Kurz and Apps (1999) suggested that as stand age increases, the ability to sequester carbon decreases and the susceptibility to disturbance increases.

Modification of soil thermal regime and permafrost degradation as a result of fire have been documented. Warmer and drier (due to reduced cover of saturated mosses) conditions following a forest fire increase decomposition and decrease carbon storage (Kasischke et al., 1995). Simulation results suggest that a 5 ºC increase in average annual air temperature results in a 6 to 20% decrease in the total carbon stored in the soil over a 25-year period (Bonan and Van Cleve, 1992). In China, Wang et al. (2001) found that soil-surface CO_2 flux decreases immediately after wildfire because of a lack of root respiration, which accounts for about 50% of total soil-surface flux. In the time after a fire during which appreciable tree mortality occurred, the majority of respiration shifts from autotrophic to heterotrophic. As a consequence of the increased heterotrophic respiration and low net primary production in the early stages of succession, areas that have recently burned in the boreal forest tend to act as a carbon source for a brief period of time. Rapalee et al. (1998) found evidence that fire scars on the landscape are a net carbon source for about 30 years after burning, after which systems become a net carbon sink. Experiments in the Alaskan boreal forest showed that about 20% of the carbon in the soil surface layer is lost through decomposition during the first 20 to 30 years after a fire due to increased soil temperature (Van Cleve and Viereck, 1983). Rapalee et al. (1998) asserted that ecosystem changes in net ecosystem production are driven by changes in decomposition and species composition. They further suggested that changes in species composition are driven by fire-induced modification of the active layer. The relative importance of active-layer modification depends to a large extent on aspect and fire severity.

Fire frequencies in the sphagnum peatlands have decreased over the past 7000 years due to cooler and wetter conditions (Kuhry, 1994). Carbon emissions from peat combustion are still an order of magnitude greater than warming-induced oxidation (Gorham, 1991). Fire affects not only the amount of carbon in the forested peat systems but also the subsequent rate of accumulation. Kuhry (1994) found that peat accumulation in sphagnum-dominated peatlands of western boreal Canada decreased significantly with increasing fire frequencies. It is estimated that warming and drying would result in relatively rapid decomposition of peat soil organic carbon (Schimel et al., 1994). Valentini et al. (2000) investigated two contrasting land cover types, a regenerating forest and a bog, in the central Siberian region during July 1996, and found that net CO_2 uptake was limited by the decreasing soil water content in the regenerating forest. Their results showed substantial differences in both transpiration and carbon assimilation. The bog used the incoming solar energy principally for transpiration and, because of the constant availability of water, transpiration was not sensitive to seasonal changes in moisture conditions. The bog system also maintained high carbon assimilation potential compared to the regenerating forest. This trend was maintained and amplified in dry conditions, when the carbon uptake of regenerating forest decreased significantly. Bogs and peatlands represent a very different land cover type than forests and must be considered separately when assessing the role of disturbance in the boreal forest (see section 8.4.4.4).

14.10.4. Forest cover type, disturbance, and climate change

Vegetation response to climate change could feed back to cause large changes in regional and global climate through effects on terrestrial carbon storage (Smith T. and Shugart, 1993) and on water and energy exchange (Bonan et al., 1992; Chapin and Starfield, 1997). The rate and magnitude of this feedback is influenced by transient changes in the distribution of terrestrial ecosystems in response to changes in climate, disturbance regime, and recruitment rates. The long-term direction of ecosystem change is also sensitive to spatial patterns and processes operating at the landscape scale (Turner D. et al., 1995).

Current projections of vegetation response to climate change either assume that the disturbance regime does not change (Pastor and Post, 1986) or use globally averaged disturbance rates (Smith T. and Shugart, 1993). Conversely, projections of disturbance regimes in a warmer climate (Flannigan and Van Wagner, 1991; Flannigan et al., 1998; Kasischke et al., 1995) generally neglect rates and patterns of vegetation response to climate and disturbance. However, the climate–disturbance–vegetation interactions clearly influence the rate and pattern of changes in vegetation (Neilson, 1993; Noble, 1993) and disturbance (Gardner et al., 1996) through effects on fire probability and spread and pattern of colonization (Turner M. et al., 1997).

Landscape-scale interactions between vegetation and disturbance are particularly important in the forest–tundra ecotone (Chapin and Starfield, 1997; Noble, 1993; Starfield and Chapin, 1996) where vegetation change is very likely to have large feedbacks to climate (Pielke and Vidale, 1995). The potential colonization of tundra by forest is very likely to increase terrestrial carbon storage (Smith T. and Shugart, 1993), thereby reducing atmospheric carbon, but increase absorption of solar radiation (Bonan et al., 1992) thereby creating a positive feedback to regional temperature increases (Chapin and Starfield, 1997).

Modeling the transient dynamics of vegetation change allows investigation of both the short- and long-term responses of ecosystems to landscape-level disturbance and recruitment, and the subsequent feedbacks between climate and the biosphere. These spatial processes are responsible for long-term changes in vegetation distribution (Dale, 1997) in response to changing climate, and must eventually be incorporated into hemispheric- or global-scale spatio-temporal models of gradual climate change.

14.10.5. Land-use change

Atmospheric CO_2 and oxygen data confirm that the terrestrial biosphere was largely neutral with respect to net carbon exchange during the 1980s but became a net carbon sink in the 1990s (Schimel et al., 2001). However, the cause of this shift remains unclear. Several studies have indicated that land-use change is responsible for the majority of the terrestrial sink (Birdsey and Heath, 1995; Fang et al., 2001; Goodale et al., 2002; Houghton and Hackler, 2000; Houghton et al., 1983). Calculations of land-use changes can be used to calculate associated carbon fluxes (Houghton and Hackler, 2000). Since 1850, there has been a 20% decrease in global forest area, and during this period deforestation has been responsible for approximately 90% of the estimated emissions from land-use change (Houghton et al., 1999).

The majority of land-use change studies have focused on areas outside the boreal region, and a review of the critical issues provides perspective on the impact of land-use change on carbon cycling in the boreal forest. In the tropics, forests contain 20 to 50 times more carbon per unit area than agricultural land, and as a result, tropical deforestation during the early 1990s released 1.2 to 2.3 Pg of carbon annually (Melillo et al., 1996). Once tropical vegetation is cleared, soil mass is quickly lost through erosion and oxidation. When tropical forest soils are cleared of vegetation and cultivated, surface horizons experience exponential mass loss resulting in roughly a 25% decrease in carbon (Melillo et al., 1996). Despite the large atmospheric source of CO_2 from tropical deforestation, the terrestrial system is acting as a net sink for carbon (IPCC, 2001), and in a spatially explicit inversion analysis, Rayner et al. (1999) found no evidence of a large net source from the tropics. The spatial inversion analysis allows a more focused examination of carbon fluxes between discrete regions. For example, the lack of a large tropical net source suggests a tropical terrestrial sink of roughly the same magnitude. The exact nature of this tropical terrestrial sink remains a source of debate. Melillo et al. (1996) used gas flux studies to show that undisturbed tropical forests in the Brazilian Amazon are responsible for a net carbon uptake, but more work needs to be done to examine and quantify carbon flux from tropical forests experiencing and recovering from deforestation.

Like tropical forests, temperate and boreal regions in the Northern hemisphere have experienced substantial land-use changes in the past several hundred years. Siberian forests account for 20% of global forest area and net primary production; Valentini et al. (2000) estimated that approximately 800000 ha are harvested there annually. Fang et al. (2001) found that Chinese forests acted as a carbon source from 1948 to 1980, and as a sink from 1981 to 1998. Subsequent works focused on constraining spatial and temporal aspects of carbon fluxes and therefore, several atmospheric inversion analyses have indicated a large terrestrial carbon sink in the Northern Hemisphere (Ciais et al., 1995; Fan et al., 1998; Keeling et al., 1996; Rayner et al., 1999). Estimates of carbon flux in the United States derived from independent forest inventory methods (Birdsey and Heath, 1995; Caspersen et al., 2000; Turner D. et al., 1995) and ecosystem models (Hurtt et al., 2002) provide supporting evidence for the presence of a North American sink, although of a lower magnitude than that estimated by Fan et al. (1998). Goodale et al. (2002) found that growth rates in unmanaged forests of the eastern United States have changed little over the past several decades, suggesting that nearly all of the carbon accumulation in the region is due to forest regrowth from past disturbance rather than growth stimulated by increased atmospheric CO_2, nitrogen deposition, or climate change.

14.10.6. Nitrogen deposition and carbon dioxide fertilization

A process-based model that simulates the biomass production of Norway spruce in southeastern Norway under both current climate and climate change scenarios was used to project biomass production responses to three climate change scenarios (Zheng et al., 2002). Net primary production (dry mass) was projected to increase by 7% over the current 10.1 t/ha/yr under a mean annual air temperature elevated by 4 ºC over present-day levels. Doubling current ambient CO_2 concentration was projected to increase net primary production by 36%. The scenario of both elevated temperature and elevated CO_2 concentration led to an increase in net primary production of nearly 50%, which was higher than the sum of the two effects alone.

Nitrogen availability is often the limiting factor in net primary productivity. The majority of anthropogenic nitrogen inputs come from combustion (both biomass and fossil fuel) and agricultural fertilizer application (IPCC, 2001). Photosynthetic rate is correlated with the nitrogen content of leaves, since carbon assimilation is driven by the nitrogen-rich enzyme rubisco. Hence, reduced nitrogen availability decreases both leaf nitrogen content and photosynthesis. As a result, the carbon and nitrogen cycles are fundamentally coupled. Due to this coupling, increased anthropogenic nitrogen deposition must be considered in conjunction with elevated CO_2 levels. Elevated CO_2 levels have multiple direct effects on fundamental biochemical processes such as photosynthesis and respiration, which collectively determine net primary production.

Among studies that manipulated both CO_2 and nitrogen availability, the mean enhancement of photosynthesis by elevated CO_2 levels at the lowest level of nitrogen availability was 40%, while the mean enhancement at higher levels of nitrogen availability was 59% (McGuire et al., 1995a). These results indicate that for a fixed increase in CO_2 concentration, biomass increases proportionally to increased nitrogen availability. In a review of the literature, McGuire et al. (1995a) also found that compared to low nitrogen availability and baseline CO_2 levels, increased nitrogen availability and elevated CO_2 concentrations significantly increased biomass accumulation, sometimes by a multiple of more than two. Differential increases in biomass in response to elevated CO_2 concentrations are found in different species: in general, deciduous species exhibit twice the growth response of conifer species to elevated CO_2 levels (Ceulemans and Mousseau, 1994).

The carbon and nitrogen cycles are also strongly correlated with evapotranspiration. Arain et al. (2002) examined the response of net ecosystem productivity and evaporation to elevated atmospheric CO_2 concentrations and found that modeled and measured results showed a linear relationship between CO_2 uptake and evaporation. This coupling implies that as nitrogen deposition increases and plant tissue carbon to nitrogen ratios decrease, nitrogen cycling increases at a fixed level of evapotranspiration (Schimel et al., 1997). This is essentially an example of biological supply and demand. It becomes less efficient for plants to exert energy on translocation before senescence; hence, the quality of litter increases. The scaling of this response from tissue to plant level is seen in the results of modeling and field studies showing that nitrogen fertilization results in increased net primary productivity (Bergh et al., 1999; Chapin et al., 1988; McGuire et al., 1992; Vitousek and Howarth, 1991).

At the ecosystem level, studies indicate that conifers and deciduous species differ in their response to elevated CO_2 levels. Arain et al. (2002) examined the response of boreal net ecosystem productivity to elevated CO_2 levels in both a 70-year-old aspen stand and a 115-year-old black spruce stand. They found that the aspen stand was a weak to moderate carbon sink while the black spruce stand was a weak carbon sink in cool years and a weak carbon source in warm years (consistent with Fig. 14.23 and the BOREAS results described in section 14.7.3.2). These results emphasize the practical importance of the strong coupling between water flux and both the carbon and nitrogen cycles. When midsummer temperatures were high, the net ecosystem production of the black spruce stand decreased significantly due to increased respiration. At longer timescales, reduced litter quality resulting from elevated CO_2 levels has the potential to cause long-term negative feedbacks that constrain the response of net primary productivity. Litter nitrogen concentration is generally positively correlated with decomposition rates, and Cotrufo et al. (1994) found that in deciduous stands, cumulative respiration rates were lower for litter derived from elevated CO_2 conditions while rates for Sitka spruce remained relatively unaffected. These results have implications for the response of forests not only to elevated atmospheric CO_2 concentrations but also to the warmer temperatures that are very likely to ultimately accompany elevated atmospheric CO_2 levels.

Schimel et al. (1994) found that the amount of nitrogen lost from an ecosystem is an increasing function of the rate at which nitrogen cycles through the system. For example, nitrogen mineralization is a key index of soil inorganic nitrogen turnover, and is strongly correlated with evapotranspiration. Schimel et al. (1997) found evidence that losses of nitrogen trace gases are linked to the rate of mineralization of ammonium (NH_4) and nitrate (NO_3) from organic matter, a rate that increases as temperature and soil moisture increase. Similarly, the product of water flux and the ratio of NO_3 to dissolved organic nitrogen (DON) concentration directly control leaching losses of NO_3 and DON (Schimel et al., 1997). Hence, nitrogen losses are controlled by soil moisture and water flux. This linkage between hydrological and nutrient cycles is of critical importance in assessing the relevance of nitrogen mineralization enhanced by increasing temperature to net primary productivity. Schimel et al. (1997) noted that water, energy, and nutrient limitation of net primary productivity and carbon storage tend to equilibrate in near "steady-state" ecosystems. This implies that the greatest potential for discrepancies between carbon, nitrogen, and water cycling exist in recently disturbed ecosystems. In support of this, Schimel et al. (2001) noted that in general, carbon accumulation in recovering ecosystems is high and chronosequence studies show lower accumulation in undisturbed landscapes. The decreased net primary productivity observed in "mature" ecosystems is a consequence of the equilibration of the nitrogen and water fluxes. Hence, the impacts of temperature increases on enhanced nitrogen mineralization are likely to be greatest in recently disturbed ecosystems.

Finally, although it is not always the case that nitrogen limits growth, a review of studies exposing plants to both elevated CO_2 levels and increased soil nitrogen concentrations showed significant increases in net primary productivity (McGuire et al., 1995a). Under the assumption that "mature" ecosystems exhibit decreased net primary productivity because of equilibration of the nitrogen and water fluxes, elevated atmospheric CO_2 concentrations are likely to stimulate growth. The relatively rapid increase in atmospheric CO_2 concentrations makes this scenario even more plausible. Elevated atmospheric CO_2 concentrations have been shown to increase the amount of photosynthesis per unit of water transpired, also known as water-use efficiency. Schimel et al. (1994) suggested that if CO_2 or fertilizer and pollutant nitrogen increase global net primary production over the coming decades, it is possible for soil carbon increases to occur on a commensurate timescale. Despite this prospect of greater carbon storage, the estimated effect of increasing temperatures is sensitive to the feedback between primary production and decomposition via the nitrogen cycle. As soil organic matter is lost through enhanced nitrogen mineralization caused by temperature increases, more nitrogen becomes available for plant growth, which results in the formation of more soil organic matter, thus acting as a negative feedback (Schimel et al., 1994).

Another possible mechanism causing equilibration of the carbon, nitrogen, and water fluxes comes from examination of the tissue-level response to elevated atmospheric CO_2 concentrations. Acclimation to elevated CO_2 levels can occur through one or more of three processes of leaf-level carbon assimilation: carboxylation, light harvest, and carbohydrate synthesis (McGuire et al., 1995a). Under saturating light conditions at low levels of intercellular CO_2, assimilation is limited by the quantity and activity of rubisco, the enzyme that is primarily responsible for capturing atmospheric carbon in the production of sugars. At high levels of intercellular CO_2, the enzymatically controlled rate of carbohydrate synthesis, which affects the phosphate regeneration that is necessary for harvesting light energy, may regulate the fixation of carbon (McGuire et al., 1995a). Hence, mechanisms acting at scales from tissue-level biochemistry to ecosystem-level nutrient cycling exert influences to equilibrate carbon, nitrogen, and water fluxes in mature ecosystems. This equilibration has implications for recent patterns of carbon flux observed in terrestrial ecosystems. Schimel et al. (2001) noted that the terrestrial carbon sink must eventually become saturated because photosynthesis follows a saturating function with respect to CO_2. As the rate of photosynthesis slows, plant and microbial respiration must catch up eventually, reducing incremental carbon storage to zero. Simulations with the Terrestrial Ecosystem Model (Marine Biological Laboratory, Woods Hole) concur, suggesting that soon after atmospheric CO_2 concentration stabilizes, heterotrophic respiration comes into balance with net primary production and the CO_2-stimulated terrestrial carbon sink disappears (Melillo et al., 1996).

In addition to the effect that anthropogenic modification of carbon and nitrogen cycles has on atmospheric CO_2, temperature in boreal and tundra regions affects the intra-annual variability in atmospheric CO_2 concentrations. Surface temperature in the north is positively correlated with seasonal amplitude of atmospheric CO_2 the following year. Zimov et al. (1999) found that 75% of the annual increases in mean annual air temperature between 1974 and 1989 coincided with decreases in CO_2 amplitude at the Barrow monitoring station, consistent with observations of net summer CO_2 efflux from tundra and boreal forest during warm years.

14.11. Climate change and forest distribution

14.11.1. Historic examples of treeline movement

The contrast between the realm of life where trees dominate the environment and the very different realm where much shorter herbaceous and low woody vegetation is the exclusive plant cover is one of the most visually striking landscape characteristics of the circumpolar Arctic. A number of questions about arctic and associated alpine treelines have attracted scientific interest for well over a century. A simple catalog and summary of this body of research is too large for this chapter. Treeline questions have been studied from a number of different perspectives, but nearly all are at least implicitly connected with the question of how cold temperatures limit tree growth and survival. Because of this overriding interest in temperature limits, and because of the ubiquitous occurrence of treelines across the Arctic and the high state of preservation of dead trees in the cold environment of the Arctic, treeline studies have much to offer on the specific question of climate warming and cooling. This section focuses on some recent studies that shed light on the long-term history of dynamics and movement of tree limits as related to specific temperature-controlled processes. These studies provide a perspective of long-term continuity, and insight into the mechanisms that recent temperature increases have affected and that warming projected by the ACIA-designated models could affect.

14.11.1.1. Northern Eurasia

Between 9000 and 7000 years BP, forest occupied what is now treeless tundra nearly to the arctic coastline throughout northern Russia (MacDonald et al., 2000). These results are based on earlier subfossil wood collections carried out by Russian scientists as well as modern collections of subfossil wood from the Kola, Yamal, and Taymir Peninsulas, and at the mouth of the Lena River. Subsequently, the greatest retreat of forest and expansion of tundra (compared to the modern position of the treeline) took place between 4000 and 3000 years BP. During the period of maximum forest advance, the mean July temperature in northern Russia (at the coast) was 2.5 to 7.0 °C higher than the present-day mean, based on modern tree growth relationships to temperature. The northward advance of the treeline reflected a series of other environmental changes in the Arctic, including increased solar insolation, reduced sea-ice cover in the Arctic Ocean, increased climate continentality, and a significant intrusion of warm North Atlantic water into the Arctic (MacDonald et al., 2000). This documented record of past forest advance suggests that there is a solid basis for projecting similar treeline change from scenarios that project similar temperature increases. It also suggests that the components of ecosystems present today have the capacity to respond and adjust to such climate fluctuations.

A reconstruction of treeline movement in the Swedish Scandes Mountains using 173 dated remains of pine wood reveals a that there has been a gradual decrease in the elevation of the treeline (with small fluctuations), generally consistent with climate cooling. Elevational decline in treeline as the result of decreasing temperatures began sometime after 10700 calendar years BP. Although the rate of treeline recession was greater before about 8000 years BP than after, generally the treeline evidence here indicates a smooth long-term temperature decrease with only a few minor, brief warming and cooling episodes. The total elevational retreat of treeline was estimated to be 500 m, corresponding to a temperature decrease of 6 to 8 °C. Since the beginning of the 20th century, however, a local warming of about 1 °C has been associated with an upward movement of more than 100 m, which represents the largest adjustment of treeline position in the last 3000 years (Kullman, 2001).

14.11.1.2. Yamal Peninsula

Detailed analyses of the dynamics of polar treeline on the Yamal Peninsula are presented in Hantemirov (1999), Hantemirov and Shiyatov (1999, 2002), and Shiyatov et al. (1996). Holocene wood deposits in the southern Yamal Peninsula include a large quantity of subfossil tree remains, including stems, roots, and branches. This is the result of intensive accumulation and conservation of buried wood in permafrost. The existence of extensive frozen subfossil wood in the southern Yamal Peninsula in what is now the southern tundra zone was first noted and described by B.M. Zhitkov in 1912 during an ornithological expedition through the area. Tikhomirov (1941) showed that tree remains conserved in peat sediments were direct evidence that the northern treeline in the warmest period of the Holocene reached the central regions of the Yamal Peninsula (up to 70° N). Today the polar treeline is considerably further south on the Peninsula (67° 30' N).

Systematic collection of subfossil wood samples started in 1982 in the watersheds of the Khadyta, the Yadayakhodiyakha, and the Tanlova Rivers in the southern Yamal Peninsula, located between 67° 00' and 67° 50' N and 68° 30' and 71° 00' E. River flow in this area is from north to south, excluding the possibility that wood was transported to the collection site from more southerly locations; thus, there is very high confidence that the region experienced a considerably warmer summer climate in the relatively recent past.

Radiocarbon dating of the subfossil wood (53 dates) was cross dated with ring series from the samples, allowing the construction of a continuous tree-ring chronology 7314 years in length. Absolute dates can

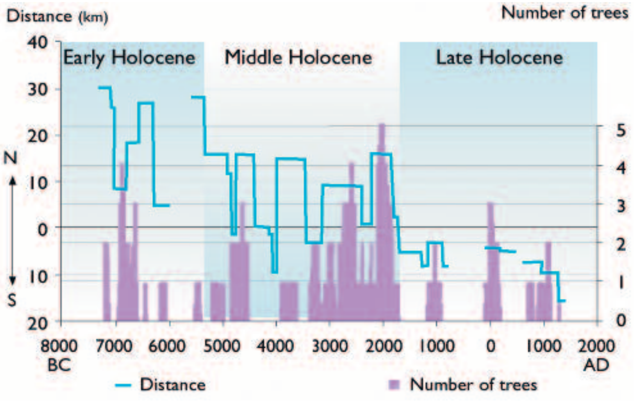

Fig. 14.27. Dynamics of the polar treeline on the Yamal Peninsula in Siberia, showing the relative distance of sampled tree remains from the present position of northernmost open stands of larch in river valleys and the number of tree samples for each radiocarbon date. A proposed division of the Holocene into three stages based on climatic shifts is indicated (data from Hantemirov and Shiyatov, 1999).

generally be assigned to the recovered wood remains. The result makes it possible to reconstruct the dynamics of tree limits on the Yamal Peninsula during the Holocene (Fig. 14.27). From at least 10 000 to 9000 years BP, trees grew across most of the Peninsula. The most favorable conditions for tree growth (almost certainly warm summers) occurred from 7200 to 6000 BC. From about 6000 to 5600 BC, climatic conditions became less favorable for tree growth (almost certainly cooler summers), but trees persisted and did not retreat to the south. Beginning about 5400 BC, trees were displaced southward. The forest stand density also greatly decreased during this period, and it can be considered as a transition to the next stage of the Holocene. From 5400 until 1700 BC, the polar treeline was still located at 69º N, well north of the present-day position. In unfavorable periods (4500–3900 BC and 3600–3400 BC), tree survival was mainly restricted to the river corridors, but in climatically more favorable periods (5200–4500 BC, 3900–3600 BC, and 3400–1800 BC), forests grew on hills and raised surfaces beyond the rivers.

One of the most important periods of displacement of polar treeline to the south and major reduction in the density and productivity of forest stands occurred about 1700 BC. This stage can be regarded as the end of the Middle Holocene and the beginning of the modern stage of treeline evolution on the Yamal Peninsula. During the last 1700 years, forest–tundra and forest associations have been primarily restricted to river valleys in the southern part of the Peninsula. Somewhat more favorable conditions occurred from 1200 to 900 BC, from 100 BC to AD 200 and during the MWP (AD 700–1400).

Treeline dynamics for the last 4000 years of the Holocene were reconstructed with even greater precision using more than 500 cross-dated tree stems with known coordinates of their burial places in valleys of different rivers on the southern Yamal Peninsula (Fig. 14.28). Treeline displacements northward and southward were relatively small and less important during the last 3600 years than those that occurred in the previous few millennia. Treeline generally moved at the most 5 km to the south of the present-day treeline, and subsequently northward only to the present boundary of open woodland in the river valleys. However, one particularly noteworthy major displacement of treeline to the south occurred during the second half of the 17th century BC. In this relatively short period (not exceeding 100 years), the boundary of larch open woodland moved southward nearly 15 to 20 km, and the treeline retreated a further 8 to 10 km during next 700 years. This major displacement of treeline in the 17th century BC appears to have been driven by strongly inclement climatic conditions (cold summers), representing the lowest reconstructed summer temperatures in the entire series. It was in the years immediately after 1657 BC that the temperature decreased sharply. Fourteen years during the interval 1630 to 1611 BC appear to have been extremely cold, reaching a nadir in two specific years, 1626 and 1625 BC. No other period during the reconstruction is even close. Moreover, it is clear that in 1625 BC, a severe freeze occurred in the middle of the summer (as indicated by characteristic anatomical structures of freeze injury in the tree rings). It is very probable that this short-term extreme climate event represented climate cooling following one of the largest volcanic eruptions of the last few millennia, which happened in about 1628 BC (possibly the eruption of the Santorini volcano in the eastern Mediterranean). The cooling appears to have reinforced another closely spaced cooling event that preceded it. The earlier of the two periods of extreme cold temperatures began sometime after 1657 BC, but in this case, it is difficult to determine the cause (Grudd et al., 2000). These events were the final circumstances that resulted in the most significant southward retreat of treeline during at least the last 4000 years. In spite of extremely favorable summer warmth that returned afterward and even persisted at various intervals, the treeline never returned to its previous boundary.

Another important result of dendrochronological dating of large samples of subfossil wood is the ability to calculate the relative abundance of Siberian spruce in forest stands of the area, which is an index or proxy for the degree of continentality of the climate. Figure 14.29 shows the change in the proportion of spruce in forest stands (the remaining part is all larch). In the first six centuries, from AD 900 to 1500, the proportion of spruce decreased from 22% to 3–5%. After that, the

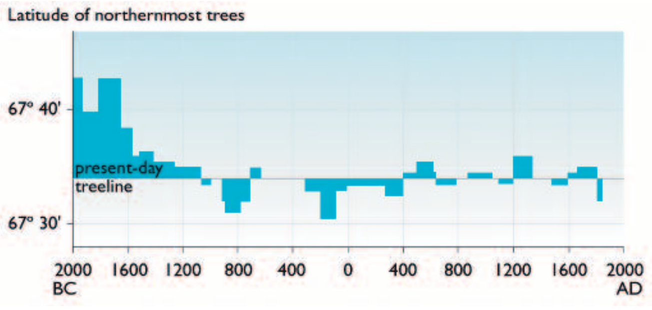

Fig. 14.28. Reconstruction of polar treeline dynamics on the Yamal Peninsula from 2000 BC, showing the latitude of recovered samples by year (reanalysis from the data of Hantemirov and Shiyatov, 2002).

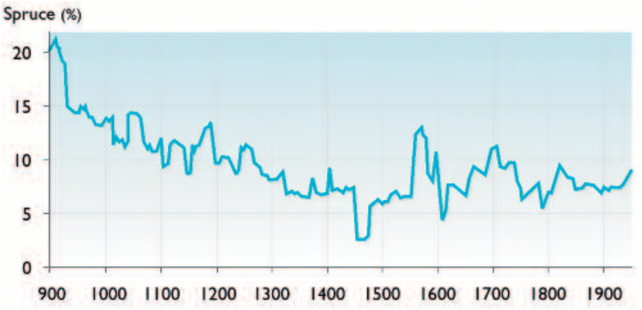

Fig. 14.29. Change in the proportion of subfossil remains made up of spruce compared to larch in Yamal Peninsula samples (data from Shiyatov, 2003).

Table 14.8. Altitudinal displacement of the upper treeline in the Polar Ural Mountains during the last 1150 years (Shiyatov, 2003).

Period (AD)	Number of years	Change in treeline	
		Direction	Rate (m/decade)
850–1280	430	rise	1–2
1280–1580	300	retreat	2–3
1580–1790	210	rise	0.5–1.0
1790–1910	120	retreat	2–4
1910–2000	90	rise	4–6

percentage of spruce stabilized in the range of 7 to 10%. The 20th century is characterized by an increasing percentage of spruce in forest stands in the valley of the River Khadytayakha, and a weak northward advance of the polar treeline.

14.11.1.3. Polar Ural Mountains

Significant spatial and temporal changes took place in the upper treeline ecotone in the Polar Ural Mountains (66º–67º N, 65º–66º E) during the last millennium (Shiyatov, 1993, 1995, 2003). Within the treeline ecotone, located between 100 and 350 m above sea level, open forests of Siberian larch dominate. Patches of closed forests of mixed larch and Siberian spruce grow at lower altitudes in the ecotone. Up to the present, these forests have largely developed under conditions of very little human influence. A large number of wood remnants persist on the ground at elevations up to 60 to 80 m above the present treeline, and within the ecotone between forest and tundra. These wood remains have been preserved for up to 1300 years because of the low rate of wood decomposition in the severe cold. The wood remnants provide material to extend a ring-width chronology back to AD 645, and to date the lifespan of a large quantity of dead trees.

In order to estimate the downward displacement of the altitudinal treeline (the highest altitudinal position of open forest over the last millennium), a transect 430 m long and 20 m wide was set up on the southeastern slope of Rai-lz Massif, from the highest location of larch wood remnants (340 m above sea level) down to the present upper treeline (280 m). The transect was divided into 10 x 10 m quadrats. All of the wood remnants were mapped and wood sections cut from the base of the trunk and roots were collected from each individual tree. The calendar year of establishment and death for each tree was determined by cross dating. Altogether, the life history of 270 dead trees was defined. In addition, 16 young living trees and seedlings were also mapped and their age determined.

Using these data, stages in the overall downward displacement of the upper treeline over the last 1150 years were reconstructed (Table 14.8). This time interval were divided into 5 periods, distinguished by differing directions of treeline shifts (rise or retreat in elevation) and differing rates of displacement. During the earliest period (430 years, AD 850–1280), the upper treeline rose from 305 to 340 m above sea level, an average of 1 to 2 m per decade. The highest altitudinal position reached by the treeline and the densest and most productive larch stands of the last millennium occurred in the 13th century. The second period (300 years, AD 1280–1580) was characterized by a substantial retreat of the upper treeline (from 340 m down to 295 m) at a mean rate of 2 to 3 m per decade. During the third period (210 years, AD 1580–1790), the treeline retreat stopped. The calculated rise (from 295 to 305 m, 0.5–1.0 m/decade) was not significant. The most extreme retreat was seen in the fourth period (120 years, AD 1790–1910), during which the upper treeline receded from 305 to down to 270 m, an average of 2 to 4 m per decade. During this fourth period, the upper treeline was at its lowest altitudinal position of the millennium. In the last period, from 1910 to the present, vigorous forest establishment took place on sites that were forested during the Middle Ages. The rate of change was the highest of the millennium (from 270 to 308 m, 4–6 m/decade).

The last period of expansion of forest vegetation is being studied using both direct and indirect evidence (old terrestrial, aerial, and satellite photographs; repeated stand descriptions of permanent plots and transects; morphology and age structure of stands; large-scale mapping within the ecotone; and meteorological and dendroclimatic data). Table 14.9 shows the change in area of different types of forest–tundra ecosystems within the ecotone during the 20th century. These data were obtained at the time of large-scale (1:10 000) mapping of the key study area (3085 ha) located at the bot-

Table 14.9. Area change in different types of forest–tundra ecosystems from 1910 to 2000 in the Polar Ural Mountains (Shiyatov, 2003).

	Tundra with individual trees		Sparse growth of trees		Open forest		Closed forest	
	Area (ha)	Change (ha/decade)	Area (ha)	Change (ha/decade)	Area (ha)	Change (ha/decade)	Area (ha)	Change (ha/decade)
1910	2403		349		328		5	
		-76		37		24		15
1960	2021		535		450		79	
		-26		-52		36		42
2000	1917		327		593		258	

tom of Tchernaya Mountain. Three maps were produced that show the spatial distribution of these types of forest–tundra ecosystems in the beginning, middle, and end of the 20th century. Change in area was estimated for two periods: the 50 years from 1910 to 1960, and the 40 years from 1960 to 2000.

Between 1910 and 2000, the area of tundra with only scattered individual trees decreased significantly from 2403 to 1917 ha, or from 78 to 62% of the main study area. The greatest rate of change (-76 ha/decade) was observed during the first 50 years, when isolated seedling establishment above the treeline was the most intensive. Data obtained on the change in area of the sparse trees mapping unit is very interesting since during the first period the change was positive (+37 ha/decade) but during the second period it was negative (-52 ha/decade). The decrease occurred when young established trees began producing seeds, resulting in an increase in stand density (many stands changed from tundra with isolated trees into open forest). This led to the great increase in open-forest area between 1960 and 2000 (from 450 to 593 ha, +36 ha/decade). However, the most impressive changes were seen in the case of closed forests. The area of closed forest increased from 5 to 258 ha over the 90 years with the transformation of open forests into closed forests. To date, more than 550 repeated terrestrial photographs have been taken which can be used to reconstruct stand parameters for the middle of the 20th century.

Temperature increases during the 20th century were the major cause of the expansion of trees in the treeline ecotone. The mean June temperature at the Salekhard weather station (50 km east of the study area) increased from 7.2 ºC (1883–1919) to 8.5 ºC (1920–1998). The July mean increased from 13.8 to 14.3 ºC. Mean temperatures in the winter months (November–March) increased from -20.8 to -19.6 ºC. The average June–July temperature, which is critical for tree growth, increased from 10.5 to 11.4 ºC (0.9 ºC). This means that the June–July isotherm rose 120 to 130 m in altitude (in this area the elevational temperature gradient is 0.7 ºC/100 m). However, the mean rise of the upper treeline was only 20 to 40 m owing to a deficiency of viable seeds on sites remote from fertile trees and stands.

In this area, larch cones open and seeds are disseminated only on days with elevated temperatures, typically from the end of June to the end of July (Shiyatov, 1966). Wind is the primary method of seed dissemination; other means of dispersal, such as animals and birds, are insignificant. Since seed dissemination takes place in summer when snow cover is absent, it is difficult for seeds to be transported upwards over long distances. Heavy larch seeds are carried no more then 40 to 60 m from the tree and most of these seeds become lodged in lichen–moss and shrub layers that prevent them from germinating successfully. That is why abundant seedling establishment took place only close to individual trees and within existing stands.

Impressive changes have occurred in the structure of existing stands during the last 90 years. Most sites with trees have become much denser and more productive (up to 4 to 5 times) and many tundra sites located within the treeline ecotone have been afforested. The degree of afforestation increased from 22 to 38% (based on data from Table 14.9). Thus, many factors affecting the forest–tundra ecotone position reacted to summer temperature changes. However, for the purpose of climatic reconstruction, the best proxies for reconstructing summer temperature are those obtained from existing stands (tree rings, biomass, stand and canopy density, degree of afforestation) rather than treeline movement, which experiences lag effects. Although the displacement rate of the upper treeline is a good reflection of long-term climatic fluctuations (Table 14.8), reconstruction of actual temperature changes is complicated because of the response lag caused by the slow growth of seedlings and the lack of seeds on remote sites. For example, the recent warming observed in summer months is of the same degree as inferred for the Middle Ages (about 1 ºC) but the upper treeline has not yet reached the altitudinal position where forests grew in the 13th century. To overcome such discrepancies, it is necessary to use corrective factors. These will be different for each period and study area. For example, in the Polar Ural Mountains over the 90 years of study, the upper treeline rose 20 to 40 m in altitude but the June–July temperature isotherm rose 120 to 130 m. Therefore, reconstructed temperatures based on recent treeline movement should be increased by a factor of four.

Recent treeline movement in the Polar Ural Mountains can be confirmed as a widespread phenomenon by comparison of high-resolution satellite images. The satellite and *in situ* data analysis shows an increasing proportion of the area with higher stand crown closure. Specifically, in 1968 about 23.5% of the area supported stands with crown closure ≥30%; in 1998 this area had increased to 50.0%. The border of stands with closure ≤10% moved between 100 and 400 m horizontally, depending on the site. Some areas that were tree-covered before the LIA still do not support any trees.

14.11.1.4. Northeast Canada

Treeline does not always fluctuate in a straightforward way in response to the control of direct climate warming and cooling on the establishment and death of trees from seed, as illustrated in northeastern Canada. Treeline populations in northern Quebec are primarily single-species stands of black spruce, surviving in the upright growth form in protected sites, or in a damaged, low growth form where spruce are exposed to wind and snow abrasion above the snowpack. Spruce macrofossils are generally absent in the northern Quebec tundra zone, and the forest limit and soil charcoal limit occur together, strongly suggesting that the forest limit has remained stable during the last 2000 to 3000 years. Apparently, climate change does not easily trigger treeline advance or retreat in this location (Lavoie and Payette, 1996).

However, the character of the forest near the treeline has changed over the last few millennia. The Hustich (1966) hypothesis suggests that the regenerative capacity of forests decreases from south to north. At the treeline of northeastern Canada, summer climate is too cold for frequent seed production, although black spruce produce cones that fail to complete only the last stages of development (Sirois, 1999). The black spruce present today were established thousands of years ago in northern Quebec and have persisted by layering since then. Sporadic and patchy fires kill black spruce in this environment and, in the absence of seed production from nearby survivors, trees are eliminated in the local fire patches. By this process, the treeline of today, which is out of equilibrium with the environment that created it, has developed a more irregular and patchy boundary with time (Payette and Gagnon, 1985).

Since the 1960s and 1970s, small-tree stands have emerged above the snowpack, triggered by less harsh conditions associated with snowier winters in the last few decades (Pereg and Payette, 1998). This process decreases the numbers of low, damaged growth-form individuals in the population, and increases the numbers of upright, tree-form stems. The temperature increases projected by the ACIA-designated models are likely to be sufficient to provide the final requirement of summer warmth that would result in viable seed production at treeline in northeastern Canada (see Fig. 14.8 for summer temperature scenarios at Goose Bay, Labrador). The initial effect of warming is very likely to induce viable seed production within the forest–tundra zone, resulting in infilling of the patchy forest–tundra border. It is possible that production of viable seeds at the absolute tree limit would begin seed rain onto the tundra, a process that probably has not occurred in appreciable amounts for thousands of years.

14.11.2. Scenarios of future treeline movement

White et al. (2000) developed a global vegetation model to project how increasing atmospheric CO_2 concentrations and other parameters, such as air temperature and nitrogen deposition, affect vegetation composition and distribution and carbon sequestration in biomass and soils north of 50° N. The model, Hybrid v4.1, is a non-equilibrium, dynamic global vegetation model, with a sub-daily time step, driven by increasing CO_2 and transient climate output from the Hadley Centre GCM (HadCM2) with simulated daily and interannual variability. Three emissions scenarios (IPCC, 1995) were used: IS92a, which results in an atmospheric CO_2 concentration of 790 ppm by 2100; CO_2 stabilization at 750 ppm by 2225; and CO_2 stabilization at 550 ppm by 2150. Land use and future nitrogen deposition were not included. The model projects an expansion of the area of coniferous boreal forest and mixed/temperate forests of the Northern Hemisphere by about 50%. This forest expansion mainly displaces tundra, driven by the direct effects of rising CO_2 levels and temperatures on tree photosynthesis and growth. The forest expansion and an associated increased carbon sink also depend on the indirect effects of increased nitrogen deposition and improved water-use efficiency. However, the model operates with only positive growth responses to increased temperatures, and new findings, such as negative growth responses in treeline trees to increased temperatures (section 14.7.3.1) and human activities forcing the treeline southward (Vlassova, 2002), suggest that the modeled increase in carbon uptake driven by forest expansion is unlikely to be fully realized.

The central Canadian Arctic is a region of low topographic relief and few barriers to tree migration, and is often depicted as an area of major northward tree movement in a straightforward response to increasing temperatures (e.g., Nuttall and Callaghan, 2000). The BIOME 1.1 model of potential natural vegetation under doubled atmospheric CO_2 conditions projects that about 60% of global tundra will be displaced, largely through the expansion of boreal forest (Skree et al., 2002). When the BIOME 1.1 model is forced with tripled CO_2 concentrations (see section 4.4.1, Fig. 4.12b), it projects that boreal forest will replace 70% of the tundra of northern Europe and virtually all the European Arctic coast will be occupied by what is today recognized as southern taiga or boreal forest (Cramer, 1997). These vegetation models of boreal forest response to climate change generally produce outputs based on the final adjustment of trees to occupy all climatically suitable areas. However, a variety of processes can reduce the potential for tree expansion or create environments that preclude tree establishment or survival, such as paludification (Skree et al., 2002). There is increasing evidence in the Russian Arctic of southward displacement of trees that are subject to strong maritime climate influences or intensified human land uses such as reindeer production systems (Vlassova, 2002). At the large scale, future northward movement of the treeline in response to climate change of the magnitude projected by the ACIA-designated models is very likely. However, a considerable lag period is possible, for example because of challenges to tree establishment (e.g., poor seedbed conditions in tundra) that must be overcome by repeated occurrences of low-probability events. Finally, environmental complexity and novelty in an altered, warmer world is likely to produce unexpected vegetation patterns at a local or regional scale.

14.12. Effects of ultraviolet-B on forest vegetation

During summer in the Arctic, levels of UV-B radiation are generally not much different than at mid-latitudes. However, the Arctic and Antarctic are uniquely susceptible to short-term, intense stratospheric ozone depletion, especially in their respective spring seasons. For general background on ozone depletion and UV-B radiation in the Arctic, see Chapter 5.

Exposure to increased UV-B (280–315 nm) radiation is known to inhibit plant growth, development, and physiological processes. Long-lived, slow-growing plants such as trees may show cumulative effects of increasing UV-B radiation levels (Sullivan and Teramura, 1992). Effects of elevated UV-B radiation levels on plant processes vary in severity and direction and among species, varieties, and clones, as well as among plant parts and developmental stages. To avoid UV-B radiation, plants have developed screening mechanisms including increased leaf thickness or cuticle (Johanson et al., 1995; Newsham et al., 1996), optical structures to scatter and reflect UV-B radiation (Kinnunen at al., 2001), and the accumulation of UV-B screening phenolics and flavonoids in the epidermal cells of the leaves (Alenius et al., 1995; Cen and Bornman, 1993; Day, 1993).

Until the 1980s, relatively few plant species had been screened to determine the effect of enhanced levels of UV-B radiation. Some species show sensitivity to present levels of UV-B radiation while others are apparently unaffected by very large enhancements in UV radiation levels. Dicotyledonous crop plants such as peas (Day and Vogelmann, 1995) and canola seem to be more susceptible to increased UV radiation levels than cereals such as wheat (Beyschlag et al., 1989) and barley, although many other factors play an important role in sensitivity. Even varieties of the same species can exhibit substantial variability in UV-B radiation sensitivity (Teramura and Sullivan, 1994). About two-thirds of the few hundred species and cultivars tested appear to be susceptible to damage from increased UV-B radiation levels. Crop damage caused by UV-B radiation under laboratory conditions generally has been attributed to impairment of the photosynthetic process (Day and Vogelmann, 1995). In addition to gross disruption of photosynthesis, UV-B radiation may inhibit plant cell division as a physiological change, causing reduced growth and yields (Teramura and Sullivan, 1994). Accurately determining the magnitude of the effect of elevated UV-B radiation levels in the field is difficult, because interactions with other environmental factors, such as temperature and water supply, affect the reaction and overall growth of the crop (Balakumar et al., 1993, Mepsted et al., 1996).

Sensitivity to UV-B radiation differs among forest species and populations and is influenced by environmental conditions (Laakso and Huttunen, 1998; Lavola, 1998). The composition of flavonoids varies according to altitude and latitude. High-latitude, low-altitude species and populations are more sensitive (fewer protective mechanisms) to enhanced levels of UV-B radiation than low-latitude, high-altitude species and populations (more developed defense mechanisms), reflecting natural levels of exposure to UV-B radiation (Sullivan and Teramura, 1992). Scots pine populations growing at high latitudes are rich in prodelphinidin (Laracine-Pittet and Lebreton, 1988) and significant differences in the characteristics of UV-absorbing compounds occur among species of pine (Kaundun et al., 1998) and birch (Lavola, 1998).

Ultraviolet-B radiation has many direct and indirect effects on plant growth and development (Caldwell M. et al., 1998). The direct effects are most damaging because the photons of UV-B radiation cause lesions in important UV-B absorbing biomolecules such as nucleic acids and proteins (Caldwell C., 1993; Greenberg et al., 1989; Taylor et al., 1997). Photoproducts of DNA formed by UV-B radiation are all toxic and mutagenic (Taylor et al., 1997) and altered DNA or RNA structures may interfere with transcription and replication causing slower protein synthesis as a result of UV-B radiation stress (Jordan et al., 1994; Taylor et al., 1997). The indirect effects of UV-B radiation on plants can also cause damage by the formation of free radicals and peroxides (Takahama and Oniki, 1997; Yamasaki et al., 1997). Excessive UV-B radiation levels can also alter patterns of gene activity (Caldwell M. et al., 1998).

Field studies in both crop plants and trees suggest that the primary effects of increased UV-B irradiance are subtle, light-induced morphological responses that alter carbon allocation (Bassman et al., 2003; Day, 2001). Recent UV radiation acclimation studies of subarctic and arctic plants have emphasized a multitude of responses ranging from avoidance and protective mechanisms to inhibition and accumulation of effects (Gwynn-Jones et al., 1999; Kinnunen et al., 2001; Laakso, 1999; Latola et al., 2001; Turunen et al., 2002). Enhanced UV-B radiation levels affect litter decomposition directly (photodegradation and mineral nutrient cycling) and indirectly (chemical changes in plant tissues) and also affect the biochemical cycling of carbon (Rozema et al., 1997). Both direct and indirect effects include physical, chemical, and biological components. The indirect effects of UV-B radiation are likely to be more significant than direct effects in subarctic and arctic forest ecosystems.

Subarctic conifers are long-lived with long generation times; damage from UV-B radiation is likely to accumulate over the years. Winter injuries to evergreens are caused by the interaction between freezing, desiccation, and photo-oxidation (Sutinen et al., 2001). In subarctic and arctic conditions, late-winter cold temperatures, enhanced UV radiation levels from intense solar radiation, and water deficiency are the major environmental risks. Xeromorphic leaves and small leaf size of subarctic and arctic trees reduce evapotranspiration. However, late spring in high-latitude subarctic and arctic ecosystems is characterized by high levels of solar radiation and fluctuations between freezing and thawing temperatures. The UV radiation dose received by evergreens in the late winter, early spring, and at the beginning of the short growing season is high due to reflectance from persistent snow cover. Some earlier studies considered evergreens to be tolerant, but other studies have revealed evergreens to be sensitive to enhanced UV-B radiation levels (Kinnunen et al., 2001; Laakso et al., 2000; McLeod and Newsham, 1997).

Experimental field evidence indicates that enhanced UV-B radiation levels mainly increase the amount of soluble UV-B-absorbing compounds in summergreen (deciduous) plants (Searles et al., 2001), but the protective functions of wall-bound phenolic compounds and epicuticular waxes in evergreens are more complicated. The responses are species-specific among pine species (Laakso et al., 2000).

In needles from mature subarctic Scots pine, enhanced UV-B radiation levels induced xeromorphic (change in plant anatomy to enhance drought resistance) characteristics, including smaller epidermal area and enhanced development of the cuticle layer (Latola et al., 2001). Ultraviolet responses increased the concentration of UV-B radiation absorbing compounds in the epidermal cells and induced high and accumulating proportions of oxidized glutathione (Kinnunen et al., 2001; Laakso et al., 2001; Latola et al., 2001; Turunen et al., 1998, 1999). The cumulative stress has been measured as gradually decreased total glutathione and an increased proportion of oxidized glutathione levels in one- to three-year-old needles (Laakso et al., 2001; Laakso and Huttunen, 1998).

The most consistent field response to enhanced UV-B radiation levels (Day, 2001; Searles et al., 2001) is an increase in concentrations of soluble or wall-bound UV-B radiation absorbing compounds in leaves, complicated by great seasonal, daily, and developmental variation, both in epicuticular and internal compounds. It is important to remember that not all soluble flavonoids are UV-inducible (Lavola, 1998) or acid-methanol extracted (e.g., cell-wall bound UV-B absorbing phenylpropanoids; Jungblut et al., 1995).

Increased amounts of epicuticular waxes and UV-absorbing compounds, such as flavonoids, and smaller leaf/needle surface area are plant defense mechanisms against UV-B radiation. In young Scots pine seedlings, UV-B radiation induces flavonoid biosynthesis (Schnitzler et al., 1997). However, an increase in UV-B absorbing compounds may result in a decrease in cell expansion and cell-wall growth (Day, 2001). Diacylated flavonol glucosides provide protection from UV-B radiation (Turunen et al., 1999) and biosynthesis of these compounds and the development of waxy cuticles appears to provide effective UV radiation protection in young needles (Kinnunen et al., 2001; Laakso et al., 2000). The protective effects come about through rapid development of epicuticular waxes, an increase in cutinization, and an initial increase followed by inhibition of UV-screening compounds (Kinnunen et al., 2001; Laakso et al., 2000).

Plant-surface wax morphology, chemistry, and quantity respond to environmental changes. Many diols (Turunen et al., 1997) are important as reflectors to avoid harmful UV radiation, but they also absorb in the UV spectrum. In many tree species, phenolics and other UV-protective substances are situated in the cuticle. The role of epicuticular waxes has been considered mainly to provide reflectance of UV radiation, but some wax components, for example, secondary alcohols (e.g., nonacosan-10-ol) and β-diketones, absorb UV radiation (Hamilton, 1995). Naturally established treeline Scots pines in a study site at Pallas-Ounastunturi National Park, Finland, were exposed to enhanced UV-B radiation levels during the full period of arctic summer daytime. The average needle dry weight increased and the wax content decreased in the UV-treated trees. The responses were observed both in the previous and current needle year (Huttunen and Manninen, in press).

A long-term study of silver birch conducted over three growing seasons showed changes in growth (i.e., shorter and thinner stems), biomass allocation, and chemical protection, while the effects on secondary metabolites in the bark were minor (Tegelberg and Julkunen-Tiitto, 2002). The changes did not occur until the third growing season, demonstrating the importance of long-term studies and the cumulative effects of UV-B radiation. A three-year study of Scots pine showed how varying defense mechanisms within the season, needle age, and developmental stage protected the Scots pine needles against increased UV-B radiation levels (Kinnunen et al., 2001). However, protective pigment decreased during the third year of exposure, suggesting that cumulative UV-B radiation exposure affects defense mechanisms and possibly makes these defenses insufficient for long-term exposure.

Quantitative changes were detected in secondary metabolites (plant chemicals) in leaves of dark-leaved willow (*Salix myrsinifolia*) exposed to enhanced UV-B radiation levels (Tegelberg and Julkunen-Tiitto, 2001). The changes in amount of secondary compounds are likely to have indirect effects at the ecosystem level on willow-eating insects and their predators, and on the decomposition process. Both dark-leaved and tea-leaved willow (*S. phylicifolia*) showed chemical adaptations to increasing UV radiation levels. However, the chemical adaptations were based more on clone-specific than on species-specific responses (Tegelberg and Julkunen-Tiitto, 2002). Several types of phenolic compounds in seedlings of nutrient-limited silver birch respond to UV-B irradiance, and seedlings are less susceptible to UV radiation when grown in nutrient-limiting conditions (Lavola et al., 1997). However, changes in secondary metabolites of birch exposed to increased UV-B radiation levels might increase its herbivore resistance (Lavola et al., 1997). A lower level of animal browsing on birch because of this chemical change induced by increased UV-B radiation levels could possibly improve the performance of birch over its woody plant competitors. Silver birch exposed to UV-B radiation with nutrient addition also displays an efficient defense mechanism through production of secondary metabolites, and demonstrates the additive effects of nutrient addition (de la Rosa et al., 2001). These results clearly establish that UV-B radiation plays an important role in the formation of secondary chemical characteristics in birch trees at higher latitudes (Lavola, 1998).

14.13. Critical research needs

14.13.1. Agriculture

This chapter focused mainly on annual crops and used GDD changes to estimate the likely effects of climate change on northern agricultural potential. Future assessments should include in-depth analyses of perennial crops and livestock, including scenarios of non-climate factors such as markets and public-sector policies. Better climate/crop models specific to agriculture in high-latitude regions (i.e., special crop varieties, long day-length effects) are also needed for future assessments. The effects of temperature increases and moisture changes on soil processes should be an integral part of the modeling systems. Future assessments should include many more climate stations in the boreal and arctic regions. The insect, disease, and weed issues that might accompany climate change and expanded agriculture should be addressed, particularly those with a unique high-latitude dimension unlikely to be part of ongoing studies in more southerly locations.

Given the critical role of public policy in agriculture in the region, an important issue is to identify national and international policies that might accommodate expansion of specific crops in the Arctic and subarctic. The infrastructure change needed to accommodate climate change should be identified along with the economic impacts of constructing new infrastructure and the economic impacts once such projects are constructed.

14.13.2. Boreal forests and climate change

Given the large contribution of the boreal forest to the terrestrial carbon sink, and the large stocks of carbon stored in it, the health and vigor of the forests of this region are a particularly significant input to climate change policy, science, and management. Climatic models, recorded data, and proxies all indicate that there are significant spatial differences in temperature anomalies and associated tree growth changes around the globe. Not all species and relatively few forest site types have been examined to determine the strength and components of climate controls on tree growth. In order to determine the current and probable future carbon uptake of the boreal forest, these spatial differences in climate change and tree growth need to be systematically identified, both in the contemporary environment and in the recent historical past. Particular attention should be given to ways of effectively combining ground-level studies, which clarify mechanisms of control and provide a historical perspective, with remote-sensing approaches, which provide comprehensive spatial coverage that can be easily repeated and updated.

Trees and forest types that change their level of sensitivity to climate or even switch their growth responses to temperature are of particular interest. If climate controlled growth more at some times and less at other times in the past, tree-ring data used to reconstruct past climates will contain biases from the recorded climate data they are calibrated to, especially if the calibrating climate data cover a short time period. It is important to establish whether this change or reduction in tree growth–climate relationships is unique to the 20th century, whether the climate change of recent decades is largely responsible for growth declines, and finally, whether the mechanism of climate control is direct or indirect.

In order to better understand, measure, and project the rapid transformations that are possible in the boreal forest, better information is needed about the role of climate and forest condition in triggering major fire and insect disturbances. The specific characteristics of fires that result in maximum production of charcoal, which represents an important form of long-term carbon sequestration, are not well documented. Further insight is needed into the interaction of climate, forest disturbance, forest succession (and its influences on carbon dynamics), water and nitrogen dynamics, and changes in carbon stocks.

Forest advance into tundra has the potential to generate a large positive temperature feedback. Unfortunately, the understanding of change at this crucial ecological boundary comes from a small number of widely separated studies undertaken to achieve many different objectives. A coordinated, circumpolar treeline study and monitoring initiative will be necessary to address definitively the question of how and why this boundary is changing at the scale required to address its potential global importance.

Improvements in scaling output from GCM grid data to applications at a given location will improve confidence in scenario projections.

14.13.3. Boreal forests and ultraviolet-B radiation

The UV-radiation survival strategies of evergreens appear to include simultaneous inhibition and avoidance. Late-winter and early-spring conditions are critical, and their importance should be studied further and effects quantified. More studies, longer-term studies, and well-designed ecosystem studies that incorporate the previously established effects from individual plot studies and consider the multiple influences of UV-B radiation on plant chemistry in settings of ecological communities will be needed to understand the cumulative effect of increased UV-B radiation levels on forests at high latitudes.

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