Climate Warming in Western North America

Evidence and Environmental Effects

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A Synthesis of Recent Climate Warming Effects on Terrestrial Ecosystems of Alaska

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Abstract
The instrument-based climate record in Alaska displays a strong trend of warming in the twentieth century. Climate in Alaska also displays a record of sudden regime shifts. Precipitation in central Alaska is highly variable and shows no strong trends. Effective moisture (P–PET) however, has decreased, resulting in widespread shrubland and drying of lakes and streams in regions of low or moderate precipitation. Overall, all climate models are uncertain; most show a slight increase in mass balance, permafrost warming across the state, and ground subsidence associated with thawing of ice-rich permafrost is commonly observed. Since buildings and infrastructure, as well as natural disturbances, can cause warming of the permafrost, it is difficult to distinguish from climate warming in some cases. The annual period of snow and ice cover is increasing, and growing season is increasing in length with greater naturalization of vegetation index (NDVI) greenness in the tundra region. North of the Brooks Range, tundra shrubs have advanced into the tundra, and warming experiments show that low shrub cover would significantly increase with additional warming. White spruce plantations at treeline include trees that grow more with warming as well as others that grow less with warming. Major species in the boreal forest region also include populations with similar responses, but growth on many of the most productive sites has declined. Recent high temperatures have caused widespread tree stress. Major outbreaks of tree-damaging insects have occurred due to both tree stress and direct temperature control on insects. Millions of acres of beetle-killed trees on the Kenai Peninsula are a potential fire hazard. The extent of forest fires in Alaska is positively associated with specific temperature factors. These changes are confronting people with a variety of challenges, ranging from obtaining subsistence food and potable water to maintaining health and safety. Scenarios of future Alaskan climate produced by general circulation models project significant future warming, which would exceed the apparent tolerance of some component species of current ecosystems.

Introduction
Climate warming caused by the increase of greenhouse gases comes about from both anthropogenic additions to the atmosphere (fossil fuel combustion, novel industrial and agricultural activities) and natural processes. In the Earth's photosynthetic capacity, either natural or through human land use changes that increase primary production and thus carbon uptake from the atmosphere. The Earth's high latitudes, especially the Arctic and Subarctic regions, have long been projected as the areas likely to experience the greatest magnitude of climate warming from these processes.

In situ formation of warm and cool air can occur radiatively in this region, and even though changes in the distribution and movement of air mass patterns either import heat or export cold across specific portions of the region, general circulation models (GCMs) indicate that, as a whole, higher latitudes will experience the greatest warming from global climate change (Houghton et al. 1995; IPCC 2001). As a result, a warming trend in the northern high latitudes should be detectable against the background noise of the Earth's natural climate system.

Climate change is naturally amplified in the Arctic and Subarctic through several processes. Changes in Arctic Ocean ice cover and boreal land cover affect albedo, or reflective power, with feedbacks that are significant at the planetary level (Bonan et al. 1993). Increased forest disturbance such as fire and insect-caused tree death can promote decomposition of fixed carbon in biomass and soils, and its release as greenhouse gases. A particular vulnerability that amplifies climate warming effects in the far north is the potential for abrupt change. Physical systems rapidly convert from the frozen state, or biological systems become less limited by freshwater conditions. Although the records are spatially sparse and limited in duration, the instrument-based record of temperatures in Alaska clearly demonstrates recent warming. Observations of temperature-dependent phenomena are particularly valuable, providing additional time perspective of climate variability and offering insight into some of the effects of warming. Change in Alaska is occurring rapidly, and consistent evidence of warming is seen in changing hydrology, permafrost, forests, disturbances, and other features.

In this chapter we present some of the evidence of climatic change in Alaska and mention some of its potential impacts on northern ecosystems and their human inhabitants. The general consensus of the Arctic Climate Impact Assessment (ACIA) is that warming is occurring at an alarming rate, and although natural variability plays a role, most warming is from anthropogenic sources: the increase of greenhouse gases in the atmosphere from the combustion of fossil fuels and changes in land use (Symon et al. 2005).

Synoptic Climate and Instrument Record
A recent synthesis of climate evidence across the Arctic region in general confirms widespread (although not quite universal) warming over the last 50 years (Serreze et al. 2000, Symon et al. 2004). Mean annual temperatures in Alaska, in particular, have increased significantly over that period, as represented by mean annual temperature data from widely separated weather stations (Fig. 9.1). The data reveal that the climate in interior and south-central, and southeastern Alaska has switched from a predominantly cool and moister period to hot and dry after a Pacific-wide regime shift in 1977 (Barber et al. 2004, Hibbs et al. 2004). Although the greatest magnitudes of warming occurred seasonally in winter, temperatures during the ecologically important spring and summer (May–August) increased as well (Fig. 9.1a, top). Autumn temperatures, however, have decreased slightly overall.

In addition to seasonal incoherence in the distribution of recent temperature increases, there was a considerable diurnal difference. In the highly continental climate region of central Alaska, the mean of daily high temperatures during the warm season increased only very slightly during the twentieth century, but the mean of daily low temperatures increased more than 0.3°C (Fig. 9.1a, bottom). Given this pattern of temperature change, the increase in diurnal range during the growing season has been disproportionately small. In recent years, the increase in growing-season length in central Alaska, while still highly variable, has been particularly impressive (Fig. 9.1). The concentration of warming in the daily temperature minima is consistent with a process operating to dampen heat loss (greenhouse gases) rather than amplify energy input (e.g., increased solar luminosity).

Records of direct precipitation over the last 50 to 100 years do not show the same strong trend as temperature. Some regions of Alaska (along portions of the southern coast) had increased precipitation, but the North Slope, interior Alaska, and the Kenai Peninsula did not experience a noticeable increase. Across much of the interior and central Alaska, annual total precipitation is quite low. Precipitation in the 100-year Fairbanks climate record, one of the longest in Alaska, is about 280 mm/year.

The pattern of variability over time in temperature and precipitation is distinct. Both the interior of Alaska and the central southeastern coast region have had a decrease in temperature and precipitation during the last half-century, whereas many coastal regions have had a significant increase in recent years (Fig. 9.1). The decadal-scale variability is superimposed on the longer-term century-scale warming. Instrument records reveal that the latest climate regime, begun in the mid-1970s, has been the most sustained period of hot and dry weather in more than a century (Barber et al. 2004) and much longer as determined by climatic reconstruction proxies (Miller et al. 1999, Overpeck et al. 1997) (Fig. 9.1, bottom).

A summer temperature reconstruction based on stable isotopes (carbon-13) and minimum leaf area density of low-elevation alpine white spruce in interior Alaska (Barber et al. 2004) shows continuity of similar temperature regimes back through 1800. Periods of warmer summer
Figure 9.3. Mean annual temperatures at three locations in Alaska. The 1956-1957 Pacific climate regime shift is indicated in the upper panel.

Figure 9.2. Warm season (May-August) temperatures at three localities in Alaska. Anchorage, 1966-2004 (south-central Fairbanks, 1906-2004 [central interior]; and McGrath, 1942-2004 [southwest interior]. Anchorage and Fairbanks have experienced urban population buildup and heat island effects (not severe in the season depicted); McGrath has remained a small village with minimal anthropogenic heat input.
temperatures versus cold can be reconstructed by tree-ring proxies as confirmed by recorded climate data, but precipitation is not sufficiently correlated to tree-ring properties to allow reconstruction.

The early part of the twentieth century was the coolest and wettest period of the century in Alaska, which marks the transition to the late-twentieth-century extremes of warmth particularly steep in the instrument-based records that span precisely that time period. Some of the climatic trends are synergistic in their effects. For example, as temperature increases, the timing of snowmelt occurs earlier in the spring, which truncates the period of snowpack accumulation and extends the period of liquid-moisture evaporation. Fall initiation of snowpack has been delayed, although not to the same degree. As a result, even with steady or even slightly increasing annual precipitation, plants experience greater evapotranspiration demand under the warming climate regime.

**Figure 9.3.** Growing season length (greatest number of consecutive days between latest spring and earliest fall dates with temperatures equal to or below freezing) in Fairbanks. Regression line added to allow visualization of overall rate of change on long (century) time scale. Rate of increase in overall season length is 4.6 days per century, with an average 16-day earlier date of last spring frost and a 25-day delay in first fall frost.

**Figure 9.4.** Normalized warm season (May-August) temperature versus normalized growth year (September-August) precipitation at Fairbanks. Top graph represents individual yearly values; bottom graph represents smoothed yearly values with five-year running mean.
Table 9.1: Area (ha) of water surface in closed-basin lakes and ponds by date and method of detection, with cumulative percentage change

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Effects on Hydrology

Effective moisture, or surface water balance, can be calculated as precipitation minus potential evapotranspiration (P-PET). Effective moisture has declined over central and northern Alaska as precipitation has failed to keep pace with warming, causing a water deficit in the summer months. For interior Alaska, the summer water deficit over the past 40 years has increased by almost 50 percent (~470 to ~770 mm), while the coastal plain summer water deficit has increased by about 200 percent (~40 to ~110 mm) (Hinzman et al. 2003). As a consequence of decreasing effective moisture, surface water levels have been declining in closed-basin lakes (no inlet or outlet surface flow) in most of Alaska.

Aerial photos taken in the 1960s were compared against Landsat ETM images from 2000–2002 in a study of different regions in Alaska with surface water in the form of lakes and ponds (Fig. 9.2). The study determined the regional trend in surface-water area loss (Russin 2005). The closed-basin water bodies were in nine regions across Alaska: (1) Copper River Basin, (2) Talkeetna, (3) Tektin National Wildlife Refuge, (4) Denali National Park, (5) Iono Lakes National Wildlife Refuge, (6) Martin Flats State Game Refuge, (7) Stevens Village, (8) Yukon Flats National Wildlife Refuge, and (9) Prudhoe Bay/Arc tic Coastal Plain. The study included more than 400,000 water bodies on approximately 48 million acres.

Analysis was based on GIS and remote sensing techniques. Water body change was detected over a 40-year time period, with a minimum of three time periods used for each area.

All study regions in subarctic Alaska lost surface-water area from 1960 to 2001 (Fig. 9.6). The Prudhoe Bay/Arc tic Coastal Plain region—which is underlain by deep, continuous permafrost (often in excess of 400 m)—had negligible change. Three areas experienced surface water losses prior to 1977; three after 1977, and two before and after 1977 (Table 9.1). Copper River Basin and Iono Lakes National Wildlife Refuge had the most substantial reduction in surface-water area, with most of the loss occurring prior to 1977.

These observed surface-water losses are likely the result of three mechanisms: increased evapotranspiration, increased active-layer water-holding capacity, and enlarged taliks (unfrozen layers sandwiched between burl permafrost and overlying seasonally frozen soil layers). Temperature is the driving factor for all three mechanisms, but a growing active layer and enlarged taliks are responses to increased temperature and associated changes in permafrost dynamics (e.g., an enlarged volume of unfrozen material). These mechanisms may also lead to increased subsurface drainage in areas where water outflow from the basin is no longer obstructed by an impervious frozen layer. The consequences of these changes include changing plant communities and drying salts.

Clearly, from 1950 to 2001, there has been a substantial change in closed-basin water bodies in many areas of subarctic Alaska. While the exact causes of surface-water area loss are unknown, the change occurred during a period of increasing temperatures when precipitation levels did not change significantly. Because the landscape-level change has occurred in regions of discontinuous
Climate Warming Effects on Terrestrial Ecosystems in Alaska

Figure 9.6. Percentage change in extent of open-water habitat in the nine study areas in Alaska.

has been particularly strong on the Tanana River at Nelana (central Alaska), where breakup has advanced earlier in the spring at the rate of 2.5 days per century since the record began in 1977 (Sagarin and Micheltorena 2003).

Other evidence of warming comes from changes in sea ice conditions in the Arctic. A 9 percent per decade decrease in perennial sea ice cover in the Arctic was documented between 1978 and 2000 using satellite data (Comiso 2001). In 2003, the floating sea ice cover of the Arctic Ocean shrank to record low levels (Serreze et al. 2003), but those were nearly matched the following two years, and a new record low was recorded in 2005.

Reduced sea ice cover represents a positive feedback to warming because open water has a lower albedo than ice, allowing more heat to be absorbed. Shrinking and disappearing sea ice has major ecological implications. Many fish, birds, and mammals live in, on, or under sea ice, or are otherwise constrained or aided by its presence. Changing sea ice conditions will impact these species as well as the people who rely on them for food and sustenance.

Sea ice affects coastal regions in other ways, and its presence protects against storm-driven shoreline erosion from fall and winter storms. Many coastal villages in northwestern Alaska are experiencing increased erosion from wave action, coastal erosion, and loss of integrated coastal ecosystems. Millions of dollars are needed to move these villages further inland to protect life and property. Enhanced buffering from these storms in ice-free seas could enhance exposure, further promoting accelerated erosion.

Sea Ice

The length of the cold season with snow and ice cover across the far-northern areas is decreasing, while at the same time the summer season is increasing (Manger 2000). In central and western North America and Eurasia during the period of reliable records (generally mid-1950s to present), there has been a strong trend toward shorter ice-covered seasons because of both later freeze-up and earlier breakup dates (Whitely et al. 1999). In the eastern portions of both continents, there has been little or no such trend. The trend toward earlier breakup of all water bodies studied showed some evidence of decrease in spatial extent (Klein et al. 2003). Some of these small lakes and ponds have disappeared.

Effects on Cryosphere

Glaciers

One of the most cited and striking examples of warming across Alaska in the twentieth century is the loss of glacier mass. The McClure Glacier in Alaska is an example of a glacier in negative mass balance. Its late-twentieth century elevation is about 45 m lower than it was in 1960 (Bahr et al. 1998), and the glacier is continuing to lose mass at an increasing rate (Nolan et al. 2005). The retreat of glaciers is being seen all across Alaska. Retreats from the Little Ice Age maximum extended positions of the terminus are measured in kilometers at many small land-terminating glaciers, and in excess of 10 km from southern coastal tidewater glaciers over the past several decades (Meier and Dryguy 2001).

The glacial geologic record (Wiles et al. 2004) and observation of recent changes in glacier length as well as mass-balance studies (Bahr et al. and Bahr et al. 2002) reveal the sensitivity of Alaska’s glaciers to multi-decadal to century-scale climate variability. Although some modest gains in glacier mass from increased snowfall have been documented (Bahr and Bahr 1999, Tribut and others 2003), and some tidewater glaciers have surged and advanced, the overall trend has been retreat, with accelerated losses since 1988 (Arclet et al. 2002, Dryguy and Meier 1999 and 2000) (fig. 9.7).

In the most comprehensive assessment, Arclet et al. (2002) used airphoto and satellite imagery to estimate thickness and volume change of 67 of Alaska’s glaciers. This work showed that recent losses, logged at a few glaciers where mass balance monitoring is underway (Tribut and others 2003), are characteristic of the losses of this large population of glaciers. Furthermore, their work has shown that previous estimates of world sea level significantly underestimated the contribution of melting Alaskan glaciers to the global rise of sea level (Arclet et al. 2002; Meier and Dryguy 2002).

Results from this study, in the acceleration of loss of this large glacier along the southern coast (Meier and Dryguy 2002), are strong reminders of the global effects of warming in Alaska and how it impacts the world by contributing to rising global sea levels. Extrapolating the thinning rate of the 67 measured glaciers to all of Alaska’s glaciers indicates a contribution to world sea level rise of 0.4 mm/yr (Wadhams et al. 2004). More local impacts include the loss of fresh water supply, increased river temperatures, and potential effects on the North Pacific climate via fresh water influx.

Permafrost

Permafrost (soil or rock that remains frozen for 10 years or more) extends across 35 percent of the Northern Hemisphere (Brown et al. 1997) and is particularly sensitive to climate change and human disturbance. Of particular interest, however, is the 18 percent of the circumpolar area and the 38 percent of boreal forest region in Alaska (Nowacki et al. 2003) defined as permafrost in lowland areas. These areas typically have high ice content associated with fine-grained material deposits and are at greater risk than mountainous areas for thermokarst (draining and settling of the ground surface following extraction of the frozen water). Thus, the potential consequences to human infrastructure and ecological processes are also much greater.

Ground temperatures obtained from deep bores in Alaska provide a centuries-long record of rising permafrost temperatures since the Little Ice Age (Leuschner...
and Marshall 1986). Permafrost temperatures in boreholes typically warmed 1-4°C during the early part of the nineteenth century. The data, which were obtained primarily along a north-south transect of Alaska from Prudhoe Bay to Gulkana (Osterkamp 2002), indicate only slight changes in permafrost temperatures into the mid-1980s. Since the late 1980s, however, permafrost temperatures along this transect and at other sites have generally warmed, initially in response to thicker snow covers.

Warming of the permafrost north of the Brooks Range up to 7°C (Fig. 9.8)—or even up to 13°C (Chow and Lefan 2003, Osterkamp 2003)—over the past 50 years is comparable in magnitude to the century-long warming seen there (1-4°C) (Lachenbruch et al. 1988), and about the same magnitude as the predicted warming of air temperatures for the next century. A calibrated permafrost temperature model (Osterkamp and Romanovsky 1999) calculates that soil temperatures in the active layer and in the permafrost have increased over the past four decades.

Although permafrost is generally considered stable in the continuous permafrost zone because mean annual ground temperatures (MAGTs) are usually -7 to -11°C, permafrost in the discontinuous zone is already undergoing widespread degradation because MAGTs are near 0°C. In Alaska, the discontinuous permafrost zone falls between the Brooks Range to the north and the Alaska Range to the south. Warming of the discontinuous permafrost is typically 0.5°C (Fig. 9.9) to as much as 1°C (Osterkamp 2001, Osterkamp and Romanovsky 1999) over the past 50 years. Thermal offset has allowed mean annual temperatures at the permafrost table to remain below 0°C with ground surface temperatures up to 4.5°C. Recent modeling indicates that 10-17 percent of the permafrost area of Alaska may thaw if climate warming continues as projected (Anisimov and Nelson 1996, Anisimov and Polashock 2003).

The lateral boundaries of permafrost bodies in discontinuous permafrost are constrained to the phase equilibrium temperature, typically slightly less than 0°C. Any warming of the permafrost will cause thawing at these boundaries and contraction of the permafrost bodies (Osterkamp 2000). Consequently, in the discontinuous permafrost that has warmed, the boundaries of permafrost bodies must be thawing. Thawing permafrost has been observed at several sites with rates of about 0.3 m yr⁻¹, indicating time scales of the order of a century to thaw the top 10 m of ice-rich permafrost. Thin, discontinuous permafrost is thawing at the base of the permafrost layer at a rate of 0.04 m per year at one site in interior Alaska (Osterkamp 2001) (Fig. 9.10).

Thawing of ice-rich permafrost (in which ice exceeds the pore space of the soil) or massive underground ice can cause the surface to settle or liquefy. The amount of
sentiment is directly related to the amount of excess ice or mosaic ice; however, the mode and rate of permafrost degradation and its ecological consequences depend on complex interactions of slope position, soil texture, ice content, and hydrology (Jorgenson and Osterkamp 2004, Smith 1993).

The dominant modes of permafrost degradation and resulting thermokarst topography include: (1) rapid lateral degradation of very ice-rich soils by thermal and mechanical erosion, resulting in thermokarst lakes; (2) degradation of ice-rich soils connected to groundwater movement, creating linear collapse-scars or (3) degradation of ice-rich soils isolated from groundwater, leading to round collapse-scars and pits; (4) differential settlement from thawing of relic ice wedges, creating high-centered polygons; and (5) minor settlement of ice-poor silty soils, resulting in hummocks or irregular mounds.

Thermokarst is developing in the boreal forests of Alaska where ice-rich discontinuous permafrost is thawing (Jorgenson et al. 2000, Osterkamp et al. 2000). Thawing destroys the physical foundation (ice-rich soil) on which boreal forest ecosystems rest, causing discernible changes in the ecosystem, including tree toppling or drowning at the ground surface subsides. Impacts on the forest depend primarily on the type and amount of ice present in the permafrost and on drainage conditions. As sites generally undrained by ice-rich permafrost, forest ecosystems can be completely destroyed.

In the Mentasta Pass area off the Tok Cutoff in eastern Alaska, wet wedge meadows, bogs, thermokarst ponds, and lakes are replacing forests. An upland thermokarst site on the University of Alaska campus consists of polygonal patterns of trenches, and pits caused by thawing ice-wedges polygons. Trees are destroyed in corresponding patterns.

In the Tanana Flats, the ice-rich permafrost supporting birch forests is thawing rapidly, and the forest are being converted to tundra vegetation. Boating and fishing. At this site, an estimated 80 percent of 2.6 x 10^6 ha was undrained by permafrost for a century or more ago. About 42 percent of this permafrost has been influenced by thermokarst development within the last one to two centuries. Subsidence at the above sites is typically 1 to 2 m, with some values up to 6 m. Much of the discontinuous permafrost in Alaska is extremely warm, usually within 1 or 2°C of thawing, and highly susceptible to thermal degradation.

Additional warming will result in the formation of new thermokarst, with its attendant impacts. In contrast to the lakes on the Seward Peninsula, the lateral expansion of thermokarst lakes near Mentasta has been noticeable, with some islands decreasing in size by 2.4 percent from 1948 to 1986 (Osterkamp et al. 2000). On the Tanana Flats in central Alaska, where degradation is primarily through lateral expansion of collapse-scars and lakes in ice-rich sites, the area of totally degraded permafrost increased from 19 percent to 49 percent over 47 years (Jorgenson et al. 2001). In Canada, tree-ring data and post-landslides such as collapse scar bogs (Halsey et al. 1995) have shown that the distribution of continuous permafrost has retreated northwards during the last 100–150 years.

Thermokarst terrain with high thaw settlement (> 1 m) destroys forests in some areas, converting them to aquatic, fen, or bog ecosystems with no similarity in

**Figure 9.9** Temperature trend in discontinuous permafrost.

**Figure 9.10** Change in depth to base of permafrost at Gulkana, Alaska. At a sufficient depth, the base or deepest front of permafrost yields to unfrozen soil or rock because of geothermal heat rising from the Earth. A rising permafrost base over time at a given location, as seen here, reflects cumulatively less cold infiltrating from the ground surface above.
species composition to the original forests. Moderate thermokarst (high-centered polygons, water tracks, piping) with intermediate thaw settlement (3-8 in) produces a mosaic of well- and poorly drained soils, creating conditions for more diverse tree and understory species. Areas with negligible or hummocky thermokarst in thawable soils become better drained and less aspericolic, and forest productivity can increase.

A quantitative determination of the impacts of thawing permafrost and thermokarst on infrastructure is difficult because the construction and habitation of a structure also have an impact on the permafrost, generally warming it. This makes it difficult to separate the effects of the structure itself from climatic warming on the underlying permafrost. Nevertheless, given the observed warming and thawing of undisturbed permafrost, it is clear that a portion of the observed damage to infrastructure must be associated with climatic warming. The economic and social consequences of permafrost degradation are likely to be enormous because much human use of the land in boreal and arctic regions is in areas that have high potential for thermokarst (Nelson et al. 2001, Osterkamp et al. 1997).

The impacts of thawing permafrost on human activities and the physical environment will depend on the permafrost ice content (Osterkamp et al. 1997). In ice-poor permafrost, warming and thawing will be limited to thermal effects and the effects of connecting ice to water. Impacts on the infrastructure in those areas will be minimal.

Thermokarst, however, is responsible for damage to homes, roads, airports, military installations, pipelines, and other facilities founded on ice-rich permafrost, creating severe maintenance and repair problems (Exch 1980, Osterkamp 1989, Osterkamp et al. 1997, Pérez 1984). Uneven topography from differential thaw settlement is a problem for agriculture and for recreational areas, such as golf fields and golf courses. In Russia, 40% of the forest zone has undergone permafrost degradation associated with human land use (Notthoff 1987).

Ecological Effects

Effects on Boreal Forest

The ecological effects of increasing temperatures, both observed and potential, on Alaskan forests are significant. In Alaska the boreal forest is a result of naturally occurring disturbances (especially fire and insects) that are triggered by warm temperatures, followed by development through successional stages influenced by characteristics of slope, aspect, elevation, drainage, and parent material. Boreal forest stands in Alaska are dominated by black spruce (Picea mariana, 55 percent of the forest cover), white spruce (Picea glauca, 36 percent of forest cover), and Alaska birch (Betula neoalaskana, 16 percent of forest cover). Only minor areas are dominated by cottonwood (Populus balsamifera), aspen (Populus tremuloides), or larch (Larix laricina) (Klud and Van Hoes 1990).

Most white spruce trees in the boreal forest grow on low-elevation upland sites and show a negative radial growth response to summer temperatures (Barber et al. 2000, Barber et al. 2000, Juday et al. 2001). With warmer summer temperatures, there is less radial growth (Fig. 9.21) due to temperature-induced drought stress (Barber et al. 2000). Floodplain white spruce sites tend to be more productive, and growth is typically positively related to summer temperature (Adams and Juday 1997) since moisture is less limiting.

Black spruce trees occur mostly on permafrost-dominated sites. Those on well-drained, north-facing ridge-top sites show a negative radial growth response to summer temperatures similar to white spruce, but with a different weighting of monthly temperature predictors of growth (Juday et al. 2001). Black spruce trees are also found on lower slopes and valley bottom sites, which are wet because of impeded drainage from the permafrost layer. These trees show a positive response to winter temperatures, with a strong negative response to April temperatures, when above-ground temperatures can stimulate photoinhibition while below-ground water remains frozen (Juday et al. 2001).

In boreal valley sites, black spruce trees also show a positive radial growth response to winter temperatures (Fig. 9.21). The majority of black spruce sites studied to date in Alaska exhibit a negative growth response to increasing temperatures, which would lead to predictions of considerable reductions in the amounts of this now-dominant species with a warming climate.

Aspen stands in Alaska have not been studied for their relationship to temperature; however, aspen cores a large portion of the western Canadian interior experienced several cycles of reduced growth between 1951 and 2000, especially in 1976-1981, when basal-area growth decreased by nearly 50% (Hogg et al. 2003). This variation in aspen growth was explained by moisture stress and insect defoliation, and another growth collapse likely occurred during a severe drought to the region during 2000-2001. There is no reason to believe that aspen sites similar to high, well-drained upland sites and white spruce growth on low-elevation upland sites in interior Alaska and projected growth based on climate-change models through 2099 are indicated (Fig. 9.21).

The scenarios indicate that by about 2070 or sooner, the climate will very likely no longer be able to sustain white spruce, based on past empirical relationships. Mortality would likely be due to temperature-induced drought stress or proximate agents, such as insects, that would be enhanced by the altered climate. White spruce will be able to persist in the landscape on floodplains and in regions where moisture is not limiting. The same relationships are indicated for black spruce trees, which display a negative sensitivity of radial growth to temperature (Fig. 9.21, bottom). The range of projected temperature increase to reach zero growth is 2-4°C.

Given these direct temperature controls of growth of the principal boreal tree species, a very different forest landscape may emerge in the next several decades in boreal Alaska.

Figure 9.21. Relationship of warm season (May-August) temperature and radial growth of white spruce on productive upland sites in interior Alaska. Tree growth is mean of 90 trees in eight stands. Temperatures for the earliest period have been reconstructed (Barber et al. 2000), those for the middle period are from instrumental records at Fairbanks, and twenty-first century temperatures are from general circulation model scenario output (Juday et al. 2005). Note the inverted temperature scale because growth is negatively correlated to temperature. In Alaska it would respond differently to local changes and deforestation episodes.

The integrated sum of photosynthesis over the period 1982-2003 in boreal North America confirms that the predominant growth anomaly during this time was reduced photosynthesis, although no systematic change in growing-season length occurred (Gower et al. 2003).

The Arctic Climate Impact Assessment (Hassel 2004) used the output of five different general circulation models (GHGFL—Geophysical Fluid Dynamics model, CSIRO—National Centre for Atmospheric Research, ECHAM—European Community Hamburg model, CCC—Canadian Climate Centre model, and the HAD—UK Hadley Centre model) to generate climate scenarios up to the year 2099. Future projections of radial growth of individual tree species have been modeled based on empirical relationships of growth to climate from previous studies and from projected temperature changes (Juday et al. 2005). Historical and reconstructed relationships between summer temperature and white spruce growth on low-elevation upland sites in interior Alaska and projected growth based on climate-change model data through 2099 are indicated (Fig. 9.21).
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**Effects on Treenline**

White spruce trees occurring at treeline in Alaska have mixed growth responses. Some trees have a positive radial growth response to spring temperature, some have a negative radial growth response to summer temperature, and others are not temperature sensitive (Wilkinson, 2003, Wilmking et al., 2004). The proportion of trees at each site belonging to the different responder types varies longitudinally across northern treeline in Alaska (Wilmking and Juday, 2005). In the western region most trees respond with increased growth in response to recent warming; in the central region, populations of responder types are mixed; and in the eastern region, most trees have decreased radial growth in response to warming temperatures.

This pattern is consistent with a decrease in precipitation from west to east in Alaska, and points to drought stress as the factor for reduced growth. Using correlation between tree growth and climate for the twentieth century, modeled carbon uptake of treeline white spruce populations for the twenty-first century predicts reduced carbon uptake in the eastern region in the twentieth century compared to twentieth-century uptake under GCM scenarios (Wilmking and Juday, 2005).

Spruce with negative radial growth responses to warming in the Brooks and Alaska Ranges still had greater total growth than the positive responders until the regime shift in 1977. After that shift, the positive responders grew more than the negative responders (Wilmking et al., 2004). These results were consistent across an east-west transect of both mountain ranges. The growth of negative responders was least correlated to prior July temperature, which explained most of the year-to-year variability in radial growth of this population of treeline white spruce.

More specifically, a threshold occurs for prior July temperature of 3°C. At Fairbanks (temperature at treeline in the Brooks or Alaska Ranges is about 4−5°C lower than at Fairbanks), this translates to an on-site threshold of about 11–12°C for July temperature in the Brooks and Alaska Ranges. In years with cooler temperatures, trees are not climatically sensitive, but in warmer years the trees became significantly sensitive and responded with decreasing radial growth (Wilking et al., 2004).

Under future, warmer conditions this July temperature threshold will be crossed more often, and as a result, the diminished growth in these treeline populations is likely to reduce the potential for continuous treeline movement under a warming climate. Treeline movement is more likely in the western regions of Alaska (Lloyd and Fastie, 2002), where wetter conditions reduce drought stress. Photographic evidence also indicates local shrub and treeline expansion during the last 50 years (Sturm et al., 2000) in western Alaska. A preliminary study (Wilmking et al., pers. comm.) of the northwestern-most white spruce in Alaska identified recent invasion of tundra by trees, whereas the central region of the Brooks Range seems to support stable treelines that have not changed during the last century.

A tree-ring width chronology (FTTH) of white spruce based on living and edict wood samples collected in the summer of 2003 from near latitudinal treeline in the Firth River area of northeastern Alaska spans the period from AD 1047 to AD 2002 and is based on 151 series (D’Arrigo et al., 2002). It shows very similar trends to those observed in an elevational treeline white spruce record (TTH, AD 1000–1200, 89 series) from the nearby Yukon Territory (D’Arrigo et al., 2004, Jacoby and Cook, 1998) (Fig. 5.14).

Figure 5.12. Relationship of winter temperature and radial growth of black spruce on a broad valley site (Site A), top, characterized by trees with a positive radial growth response to warming the best predictive temperature index for this population is the mean of winter monthly temperatures (December and January in the year of ring growth, and the same months plus February in the previous year). Tree growth data at Site A have been smoothed with five-year moving means. A negative radial growth response of black spruce to summer monthly temperatures on a subarctic river terrace site (Site B) is shown in the bottom graph. The best predictive temperature index for this population is the mean of May, June, and previous June and July mean monthly temperatures. Tree-growth data at Site B are individual-yearly values; the temperature scale is inverted because growth is negatively correlated. Twenty-first century values are a scenario output from GCM climate models used in the Arctic Climate Impact Assessment (Juday et al., 2000).
Figure 9.33. Mean decadal growth of white spruce at or near treeline (n = 1,135 trees; eight stands in the Brooks Range, seven stands in the Alaska Range). Note the crossover of growth relationship of the populations with a positive versus negative growth response to warmth at the time of the Pacific climate regime shift.

Figure 9.34. Tree-ring width (RW) chronologies for white spruce (Picea glauca) spanning the past millennium. The Firth record is from the Firth River area of northeastern Alaska (AD 1617–2000); the THH record is from the Yukon Territory, Canada (AD 1099–2000). Both series have been standardized to optimize retention of low-frequency trends using conservative methods (negative exponential function or regression line of negative or zero slope). Note growth changes in the twentieth century, with some growth decline at THH site (east).

Effects on Tundra

Shrub tundra is replacing tussock tundra in areas of warming climate (Bliss and Marteyn 1992), and tundra is showing increased shrubbiness in Alaska on the North Slope and the Seward Peninsula. A study comparing and contrasting photos between 1939 and 2000 shows an increase in shrub density on the North Slope in the Chandler River drainage (Seun et al. 2001). Most of the increased shrubbiness is a result of a "filling in" of shrunken areas where they were lacking earlier in 1949. A noticeable advance of shrubs was also documented on the Seward Peninsula along mountain valleys and riparian corridors between 1978 and 1992 using GIS and remote sensing (Siipasaari et al. 2001).

Increasing shrub cover suggests that directional changes are occurring on the Seward Peninsula consistent with experimental tundra warming (Chapin et al. 1999). Experimental warming of shrub tundra plots also suggests increased leaf area of existing shubs (Asbirt et al. 1999, Beet et al. 2001, Chapin and Shaver 1996, Hobbie and Chapin 1998). Modeling studies suggest that shrub tundra will continue to expand if warming continues (Kaplan et al. 2003). Such expansion has implications for the exchange of water, energy, and trace-gas exchange with the atmosphere (McGee et al. 2003, Thompson 2004, 2007).

Much of the landscape on the Seward Peninsula is in a transition zone between boreal forest and tundra, so it
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is expected that with warming, the treeline will advance onto the tundra, as it already has in some areas on the peninsula where glacial retreat has occurred (Lloyd et al. 2002). Recent climate warming may be an important factor in influencing vegetation change over the Seward Peninsula (Lloyd et al. 2002).

Changing land cover also affects the surface energy budget by changing the albedo, both directly because of vegetation changes and indirectly (and of greater magnitude) due to changes in snow distribution. Shrubs tend to trap the snow and thus provide a positive feedback from increased albedo (lower reflection of heat energy, and thus more absorption), which leads to warming. There has been a documented increase in the normalized difference vegetation index (NDVI) measured since 1990 (Jia et al. 2003, Myrnet al. 2004, Myren et al. 1997). This is evidenced by an increase in the growing season in interior Alaska (Barber et al. 2003) as well as north of the Brooks Range (Chapin et al. 1999).

The spring of 2004 saw the earliest leaf-out in Fairbanks (May 4) that has ever been recorded. With earlier and greater spring warming and later fall onset of freezing, the growing season has been extended (Barber et al. 2003). Consequently, the same amount of moisture is required over a longer growing season as was needed over a shorter growing season, presumably causing moisture stress. The recharge of water to the soil in spring, as contrasted with precipitation received during the growing season, is critical to vegetation growth through the summer months.

Modeling Results

A logistic regression model (Cale et al. 2005) predicts the potential equilibrium distribution of four major vegetation types in interior Alaska—tundra, deciduous forest, black spruce forest, and white spruce forest—based on elevation, aspect, slope, drainage type, fire interval, average growing-season temperature, and total growing-season precipitation. The hierarchical logistic regression model was used to evaluate how scenarios of changes in temperature, precipitation, and fire interval may influence the distribution of the four major vegetation types found in this region. The model was validated for interior Alaska (the model was used to develop the model), where it predicted vegetation distribution among the steps with an accuracy of 69-81 percent. When the model was independently validated for northwest Canada, it predicted vegetation distribution among the steps with an accuracy of 85-85 percent.

Effects of Forest Disturbance

Insects

On average, insect infestations annually affect a small amount of area as that affected by fire in the Alaskan forests (Werner et al. 2006). Records in Alaska indicate that insect infestation affects approximately 400,000 ha annually (Werner et al. 2006), but the effective area of insect-induced stand mortality is less than 100,000 ha because successive years of attack are required to cause stand mortality, and only a portion of the stands affected are host species for the attacking insects. In Alaska, the insects responsible for major infestations include the spruce beetle, the sparrow-marked black moth, the large aspen tortrix, the eastern spruce budworm, and the larch sawfly. Insect infestations in Alaska peaked during the late 1970s (impact more than 600,000 ha annually) with outbreaks of the sparrow-marked black moth, the large aspen tortrix, and the larch beetle, and during the 1990s (approximately 500,000 ha annually) with outbreaks of the spruce beetle, the eastern spruce budworm, and the larch sawfly (Werner et al. 2006). An outbreak of the larch sawfly during the late 1990s is estimated to have killed most of the mature larch in interior Alaska (>650,000 ha) (Werner et al. 2006).

Spruce bud beetles are found in background levels throughout Alaska, but a major outbreak has occurred in interior Alaska, and there is no evidence of major infestations. Controls over the population dynamics of insects responsible for tree mortality are not well understood. Some insects seem to cycle on a regular basis, such as the spruce budworm in eastern Canada (~30-year, Kern and Apps 1999, and the large aspen tortrix (~12 years) and the sparrow-marked black moth (~10 years) in Alaska (Werner et al. 2006). In contrast to infestations in eastern Canada, outbreaks of spruce budworm in Alaska have been observed only in the 1990s, when it infected around 600,000 ha (Werner et al. 2006). Werner et al. (2006) have hypothesized that this infestation may have been due to the temperature-induced drought stress experienced by white spruce in interior Alaska in association with warmer summers (Barber et al. 2000).

Factors that affect trees appear to be important in influencing insect infestations in western Canada as well, where decreasing effective moisture has caused multiple-year defoliation of aspens by the forest tent caterpillar as well as reduced tree growth (Hogg et al. 2005, Hogg et al. 2001). In Alaska (Werner et al. 2006), the engraver beetles tend to infest stressed and dying trees near the edges of stands in the interior (Werner and Posewitz 1997).

Effects of warming temperatures on the life cycles of insects also play an important role in insect outbreaks. For example, a single generation of spruce beetle can require either one or two years to mature, and this generally depends on temperature (Werner and Hofsten 1984). Warner summers in south-central Alaska during the 1950s and 1960s caused a shift from a two-year to a one-year life cycle for spruce beetle, which played a role in the large outbreak of the 1990s on the Kenai Peninsula (Werner et al. 2006). Werner et al. (2006) found that warmer summers also allowed a larger population of bark beetles to survive the winter, leading to a higher rate of subsequent population increase. Finally, unfavorable climatic conditions of warm temperatures and low precipitation have stressed regional populations of insects; for example, the eastern spruce budworm can "pitch out" wood-boring beetles (unless overwintered by numbers), but a slow-growing, weak tree with low growth reserves is less able to defend itself.

On the Kenai Peninsula, the buildup of the population of beetles to overwhelming size was assisted by an already stressed population of trees (Borg et al. 2006). This combination of events led to a massive outbreak of the spruce bark beetle (Dendroctonus rufipennis Kirby) that killed most of the mature white and Sitka spruce in the Kenai Peninsula forests, affecting two to three million acres over the last ten years (Borg et al. 2006, Wittmer 2004). Lack of host trees in this area is now the major limitation on future outbreaks. Under continued warming, we expect to see more insect infestations spreading to previously unaffected forest regions in Alaska.

Fire

Fire is another disturbance in the boreal forest associated with warm temperature anomalies. In Alaska, the area burned in a particular year is positively correlated with mean June temperature, and negatively correlated with the depth of snowpack near the end of winter (Duffy et al. 2005). Occasionally large fires account for most of the area burned in Alaska (Kasischke et al. 2006). Approximately 68 percent of the total area burned in Alaska from 1940 to 1998 occurred during 15 years characterized by a moderate-to-strong El Niño, which resulted in above-normal temperatures throughout the year, and normal precipitation from February to August in Alaska (Hess et al. 2003). Among the state's different ecoclimates, fire cycle decreases with increasing growing-season temperature, decreasing growing-season precipitation, and increasing lightning frequency (Kasischke et al. 2006).

Humans affect fire regimes in the boreal forest through the ignition and suppression of fires, and land use that alters the spatial distribution of fuels. In Alaska, lightning ignitions are largely distributed throughout the interior between the Brooks and Alaska Ranges, while human-caused ignitions are largely confined to the road network (Gabriel and Taake 1984, Kasischke et al. 2006). Although humans are an important source of fire ignition in boreal forests, they are not responsible for the majority of area burned. In Alaska, humans cause more than 60 percent of all ignitions, but these ignitions result in only 2 percent of the area burned (Kasischke et al. 2006).

The low amount of area burned from human-caused fires is the result of poor burning conditions at the time of ignition and responsive fire-suppression efforts; the fires started by humans are generally quickly detected and accepted by response teams. As fire conditions and suppression networks in Alaska, recent analyses indicate that fire cycles are longer near population centers (Chapin et al. 2005).

Fire suppression is important in prolonging longer fire cycles near population centers, but the alteration of fuels by human land use likely plays a role as well. For example, near Fairbanks, substantial harvesting of spruce forests during the settlement period in the early 1900s led to the replacement of many of these forests, through natural succession, by faster-growing aspen and birch forests, which are less flammable than spruce forests and inhibit the spread of fire. The combination of fire suppression and spatial alteration of fuels is presumably responsible for a fire cycle near Fairbanks that is estimated to be four times longer than the fire cycle of aspen in interior Alaska located away from population centers. The fire cycle in this area may also be influenced by insect infestations; affected stands may be more vulnerable to fire where flammability of forest stands increases with tree mortality.

Alaska has a reliable 56-year record of area burned (1950-2006). Total area burned has increased markedly and has been concentrated in specific fire years. Burning conditions, in general, are characterized by sustained periods with high daily-maximum temperatures (≥15°C).
climate warming, the net effect on radiative forcing of the atmosphere is not clear because drainage can either be enhanced or retarded by perennial degradation, and the response of drainage is likely to affect the release of CO₂ and CH₄, but in opposite directions (Rood et al. 2000, Rou- let et al. 1993, Roulet et al. 1998). For example, if perennial frost warming results in a drop in the water table, the release of CO₂ from aerobic decom- position is likely to be enhanced (Oechel et al. 1995, Christensen et al. 1998), but CH₄ emissions will likely decrease because methanogenesis is an anaerobic process (Rood et al. 2000, Roulet et al. 1992, Roulet et al. 1998). In contrast, if permafrost melting results in the expansion of lakes and wetlands as the result of water collection in areas of surface submergence, then CH₄ emissions are likely to be enhanced (Reeburg and Whalen 1992, Timo et al. 1997) while CO₂ emissions are reduced.

Observations of Changing Climate and Effects on Indigenous People

In recent decades, indigenous people in northern and western Alaska have observed notable changes in terrestrial ecosystems used by people (Huntington 1990, Hun- tington et al. 1998). Vegetation has shown accelerated rates of migration, with trees moving into formerly treed areas on the Chukchi and Baldwin Peninsula (Pungowiyi 1990) and willows becoming larger and more common in the Ktzwel region (Whiting 2001). Warmer summers have led to a profusion of many kinds of insects, including mosquitoes not seen before in certain areas (Pungowiyi 2002).

Geese and songbirds have been arriving earlier than in living memory, making them more susceptible to sudden cold snaps (Pungowiyi 2000). Some new birds, which have not been seen before, have been seen in the Ktzwel region (Whiting 2002). Winter has been relatively mild, with little snow, allowing ptarmigan populations to thrive (Pungowiyi 2000). Increases in lightning in north- western Alaska have led to more frequent forest and tundra fires (Whiting 2002). Moose have expanded their range into the northwestern Aniak (into the 300,000 acre Ktzwel region) (Whiting 2001). Beaver and muskrat have ex- tended their ranges in northwestern and western Alaska (Whiting 2002, Huntington et al. 1999).

These phenomena appear connected to changes in climate and are recognized by Alaska Natives as large-scale, directional changes of a kind and magnitude not seen before (Pungowiyi 2000). Many other changes have been reported—for example, in populations of
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