Simulating the response of natural ecosystems and their fire regimes to climatic variability in Alaska[1]

D. Bachelet, J. Lenihan, R. Neilson, R. Drapek, and T. Kittel

Abstract: The dynamic global vegetation model MC1 was used to examine climate, fire, and ecosystems interactions in Alaska under historical (1922–1996) and future (1997–2100) climate conditions. Projections show that by the end of the 21st century, 75%–90% of the area simulated as tundra in 1922 is replaced by boreal and temperate forest. From 1922 to 1996, simulation results show a loss of about 9 g $C \cdot m^{-2} \cdot year^{-1}$ from fire emissions and 360 000 ha burned each year. During the same period 61% of the C gained (1.7 Pg C) is lost to fires (1 Pg C). Under future climate change scenarios, fire emissions increase to 11–12 g $C \cdot m^{-2} \cdot year^{-1}$ and the area burned increases to 411 000 – 481 000 $ha \cdot year^{-1}$. The carbon gain between 2025 and 2099 is projected at 0.5 Pg C under the warmer CGCM1 climate change scenario and 3.2 Pg C under HADCM2SUL. The loss to fires under CGCM1 is thus greater than the carbon gained in those 75 years, while under HADCM2SUL it represents only about 40% of the carbon gained. Despite increases in fire losses, the model simulates an increase in carbon gains during the 21st century until its last decade, when, under both climate change scenarios, Alaska becomes a net carbon source.

Résumé : Le modèle de végétation dynamique adapté à l'échelle globale MC1 a été utilisé pour étudier les interactions entre le climat, les feux et les écosystèmes sous des conditions climatiques passées (1922–1996) et futures (1997–2100) en Alaska. Les projections montrent que 75 % à 90 % de la superficie de la toundra simulée en 1922 sera remplacée par une forêt boréale ou tempérée vers la fin du 21e siècle. Selon les résultats de la simulation, de 1922 à 1996, les feux auraient détruit annuellement une superficie de 360 000 ha et les émissions associées aux feux auraient causé des pertes d'environ 9 g $C \cdot m^{-2} \cdot an^{-1}$. Pendant la même période, 61 % des gains en carbone (1,7 Pg C) ont été consumés par le feu (1 Pg C). Selon des scénarios de changements climatiques futurs, les émissions provoquées par le feu augmenteraient à 11–12 g $C \cdot m^{-2} \cdot an^{-1}$ et la superficie brûlée augmenterait à 411 000 – 481 000 $ha \cdot an^{-1}$. Le gain en carbone prévu entre 2025 et 2099 s'élèverait à 0,5 Pg C selon le scénario de changements climatiques le plus chaud (CGCM1) et à 3,2 Pg C selon le scénario HADCM2SUL. Les pertes attribuables aux feux avec le scénario CGCM1 sont donc plus grandes que les gains en C au cours de cette période de 75 ans alors qu'avec le scénario HADCM2SUL, elles ne représentent qu'environ 40 % des gains en C. Malgré l'augmentation des pertes attribuables aux feux, les simulations du modèle montrent une augmentation des gains en C au cours du 21e siècle jusqu'à sa dernière décade lorsque l'Alaska devient une source nette de carbone selon les deux scénarios de changements climatiques.

[Traduit par la Rédaction]

Introduction

Over the last 150 years, lakes and rivers in the Northern Hemisphere have shown a trend toward later freeze-up and earlier break-up dates (Magnuson et al. 2000). Permafrost degradation has been documented in Alaska (Lachenbruch et al. 1986) and in parts of Canada, with a northward shift of the southern limit of permafrost by about 200 km over the 20th century (Chen et al. 2003). Satellite observations have also suggested a lengthening of the growing season at high northern latitudes (Keeling et al. 1996; Myneni et al. 1997). As evidence for climate change becomes more tangible, concerns have arisen over the stability of the large carbon reserve in the Alaskan boreal and arctic regions. Moreover, the area burned by wildfires in North America's boreal forests was higher in the 1980s than in any previous decade on record (Murphy et al. 2000 cited by Harden et al. 2000).

The interactions between climate and fire are complex. Warmer temperatures alone would lead to increased fire activity by reducing the moisture content of fuels and increasing litterfall caused by drought stress, as long as ignition sources were not limiting. A concurrent increase in precipi-

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tation amount and storm frequency may more than compensate for the increase in temperature by reducing flammability and also fostering changes to a less fire prone or drought sensitive vegetation cover (Bergeron et al. 2004). On the other hand, increased rainfall will enhance plant growth and fuel build-up, and an increase in storm frequency will enhance the opportunity for lightning (Lynch et al. 2004). Different combinations of these factors can result in no change or even a decrease in predicted fire activity, and the exact effect of climate change on fire is likely to vary greatly regionally. As wildfires in the boreal forests appear to show tremendous interannual variation in both area burned and severity of burning (Harden et al. 2000), it is difficult to project their effect on the boreal carbon budget. Moreover, their indirect effects on permafrost may hasten thermokarst dynamics triggered by warmer temperatures and further complicate the final outcome.

An intercomparison of the response of dynamic global vegetation models (DGVMs) to transient climate (Cramer et al. 2001) shows that the models were most consistent at high latitudes, probably because of the overriding importance of temperature as a driving force in the response of those ecosystems (Kittel et al. 2000). All the models projected increases in net primary production (NPP) and vegetation biomass and northward shifts of the various vegetation zones under future climate change scenarios. However, none of the models included a dynamic fire module that could realistically simulate the interactions between wildfires, climate change, and vegetation. We are using our dynamic vegetation model MC1, which includes a state-of-the-art dynamic fire module. MC1 results have compared favorably with historical records and observations of fire impacts at the national and regional scales in the conterminous United States (Lenihan et al. 2003) and are currently used to forecast seasonal fire risk (http://www.fs.fed.us/pnw/corvallis/mdr/mapss/). Records of annual area burned from the 1920s to the 1940s, when fire suppression efforts were limited, are within 10%–15% of the simulation results. We used transient climate records for the state of Alaska and two climate change scenarios to project possible outcomes for the 21st century. Although conclusive model validation of the carbon budget of the state of Alaska is not possible at these large temporal and spatial scales and despite serious model limitations such as the absence of permafrost and inundated soils, we are nonetheless able to demonstrate the strength of our model to simulate historical vegetation distribution and the accuracy of our fire projections against the available fire records. The model results reported here are among the best available information to estimate the future role of fire in the carbon budget of high-latitude ecosystems in Alaska.

Materials and methods

Climate inputs

The Vegetation–Ecosystem Modeling and Analysis Project (VEMAP) was a collaborative program to simulate and understand ecosystem dynamics for the continental United States (Kittel et al. 2004). The project developed common data sets for model input, including a high-resolution, topographically adjusted climate history of the conterminous United States on a 0.5° grid with soils and vegetation cover. Data sets for Alaska were added to complete the coverage of the continental United States. The climate data set was developed at the National Center for Atmospheric Research (NCAR) (http://www.cgd.ucar.edu/vemap/datasets.html) in collaboration with Chris Daly (Oregon State University) and A. David McGuire (University of Alaska, Fairbanks).

Historical conditions: 1922–1996

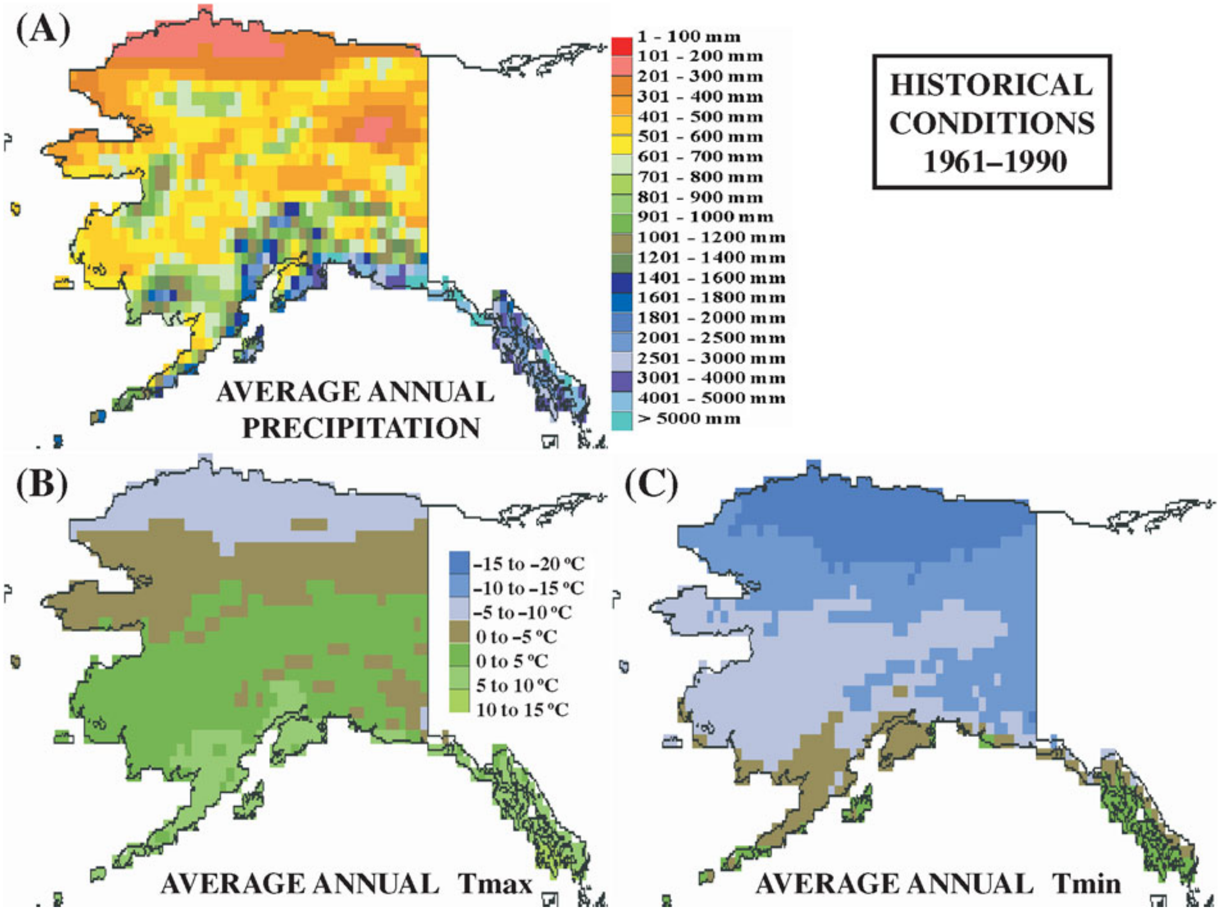
Monthly mean minimum and maximum temperature, monthly precipitation, and monthly mean humidity and solar radiation time series were derived from the National Climate Data Center (NCDC) Historical Climate Network (HCN), Global Historical Climate Network (GHCN), cooperative network monthly station data, Natural Resources Conversation Service SNOTEL data, and Canadian monthly station data (approx. 600 stations) for the period 1922–1996. Monthly precipitation and temperature anomaly series were created from station records and a 4-km 1961–1990 baseline climate generated by Chris Daly (Oregon State University) with his model, PRISM (http://www.ocs.orst.edu/prism/). These monthly anomalies (deltas for temperature, square-root ratios for precipitation) were then kriged to the 0.5° VEMAP grid to create temporally and spatially complete monthly anomaly fields for the period 1922–1996. The monthly gridded anomalies were converted back to temperature and precipitation series using the 4-km resolution baseline climate, regridded to 0.5°. Following the protocol of Kittel et al. (2004), daily temperature and precipitation records were stochastically generated from monthly values using a modified version of WGEN (Richardson 1981; Parlange and Katz 2000), parameterized with daily weather station records from the region (without separate wet vs. dry parameterizations used for the conterminous United States data set). Daily (and monthly) vapor pressure, daytime relative humidity, total incident solar radiation (SR), and irradiance (IRR) were simulated from daily temperature and precipitation using MTCLIM (version 3; Thornton and Running 1999; Thornton et al. 2000).

Historical climate is characterized by cold winters (Fig. 1C), a north–south gradient of temperatures with a strong arctic influence (Figs. 1B, 1C), and a gradient of moisture from the wet south coast to the dry interior (Fig. 1A) and the dry north coast.

Future conditions: 1997–2100

Transient climate change scenarios were based on coupled atmosphere–ocean general circulation model (AOGCM) transient climate experiments. The purpose of these scenarios was to reflect time-dependent changes in surface climate from AOGCMs in terms of both (1) long-term trends and (2) changes in multiyear (3–5 years) to decadal variability patterns, such as the El Niño Southern Oscillation. Scenarios were based on greenhouse gas experiments with sulfate aerosols from the Canadian Climate Center (CGCM1) and the Hadley Centre (HADCM2SUL; Mitchell et al. 1995; Johns et al. 1997), accessed via the Climate Impacts LINK Project Climatic Research Unit, University of East Anglia, and processed by the VEMAP data group (http://www.cgd.ucar.edu/vemap/datasets.html). To make the scenarios compatible with the Alaska historical data set, monthly GCM mean minimum and maximum temperature and precipitation time-series anomalies (relative to a 1961–1990 baseline) were spline fit to the

Fig. 1. Historical climatic conditions (1961–1990): (A) average annual precipitation, (B) average annual maximum temperature (Tmax), and (C) average annual minimum temperature (Tmin).



VEMAP grid and then recombined with the corresponding VEMAP baseline climate. These temperature and precipitation time series were then used to create monthly humidity and solar radiation series in the same manner as for the historical series (Kittel et al. 2004).

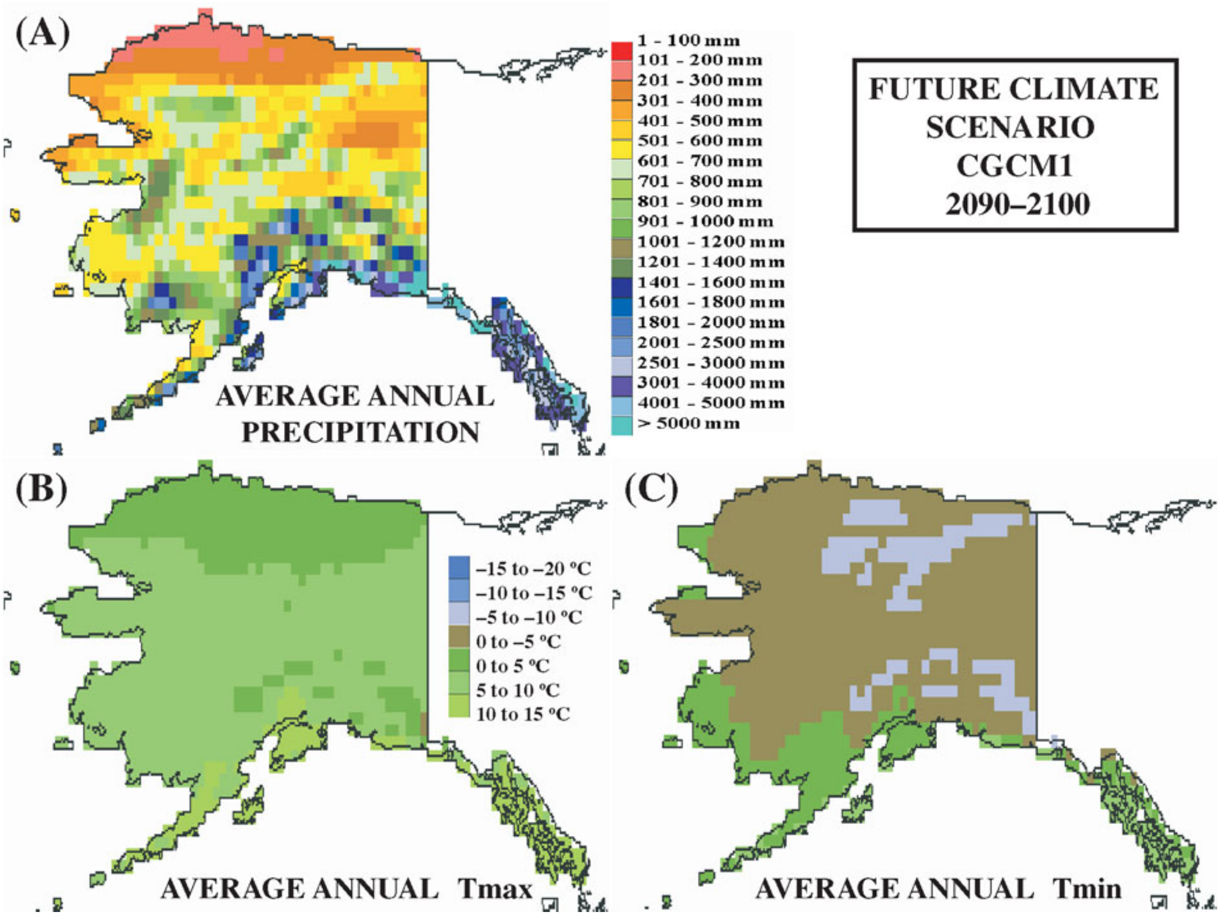
Under HADCM2SUL, the average annual temperature from 2090 to 2099 rose by almost 5 °C above the 20th century state-wide average (–4 °C from 1922 to 2000), while under CGCM1 it rose 8 °C above the 20th century state-wide average (Figs. 2 and 3). Annual average precipitation between 2090 and 2100 increased by 13% above the 20th century (1920–2000) state-wide average (925 mm) under CGCM1 and 5% under HADCM2SUL (Figs. 2 and 3).

Model

MC1 (Daly et al. 2000; Bachelet et al. 2000, 2001*a*, 2001*b*; Lenihan et al. 2003) is a DGVM derived from the MAPSS biogeography model (Neilson 1995) and the CENTURY biogeochemical cycling model (Parton et al. 1987). It simulates life-form mixtures and vegetation types (Table 1), fluxes of carbon, nitrogen, and water, and fire. The model is run separately for each grid cell, that is, there is no exchange of information across cells. The model reads climate data at a monthly time step but the fire module interpolates the data to create daily inputs.

A fire event is triggered when three thresholds are reached: (1) the 12-month self-calibrating Palmer Drought Severity Index (PDSI, Palmer 1965; Wells 2003; http://nadss.unl.edu/PDSIReport/pdsi/self-cal.html) is used as an indicator of moderate to severe drought to control the interannual timing of fire events; (2) the 1000-h fuel moisture content of dead fuels (Fosberg et al. 1981) is used as an indicator of extreme fire potential to control the seasonal timing of fire events; (3) the threshold of fine-fuel flammability (Cohen and Deeming 1985) is used as an indicator of the sustainability of fire starts to control the timing of fire events at the daily time step. The fire module determines the area burned as a fraction of each grid cell as a function of the current vegetation type, current drought conditions, and number of years since fire. Every simulated vegetation type is associated with a minimum and maximum fire return interval (Table 2; Yarie 1981; Kasischke et al. 1995*a*; Leenhouts 1998; Wong et al. 2003). A return interval between the minimum and maximum values for the current vegetation type is selected as a function of the current drought condition as indicated by PDSI. Potential fire behavior is based on current weather and estimates of the mass, vertical structure, and moisture content of several live and dead fuel size classes. Biomass is allocated to live and dead fuel classes as a function of current vegetation types (Albini 1976; Anderson 1982). Allometric functions are used to simulate crown height, length, and

Fig. 2. Future climatic conditions: (A) average annual precipitation, (B) average annual maximum temperature (Tmax), and (C) average annual minimum temperature (Tmin) for the decade 2090–2100 under CGCM1.



shape, and the depth of surface fuels (Botkin et al. 1972; Peterson 1985; Weller 1987; Keane et al. 1996). The direct effects of fire simulated by the MC1 fire module are the mortality and consumption of aboveground carbon and nitrogen stocks.

MC1 assumes that a variable fraction of the live and dead aboveground biomass, distributed in the different fuel size classes, is consumed as a function of its moisture content, except for 1-h fuels (fine litter), which are 90% consumed by each simulated fire event (Ottmar et al. 1993; Reinhardt et al. 1997). Root consumption and fire-induced root kill are not simulated. Live fuel moisture is simulated as a function of the soil moisture content (Howard 1978). The moisture content of each dead fuel size class is a function of antecedent weather conditions (Cohen and Deeming 1985). The model assumes that 30% of the nitrogen present in the biomass consumed by fire is returned to the soil. Gaseous and particulate fire emissions are simulated using emission factors (Ottmar et al. 1993) that are a function of the current vegetation types.

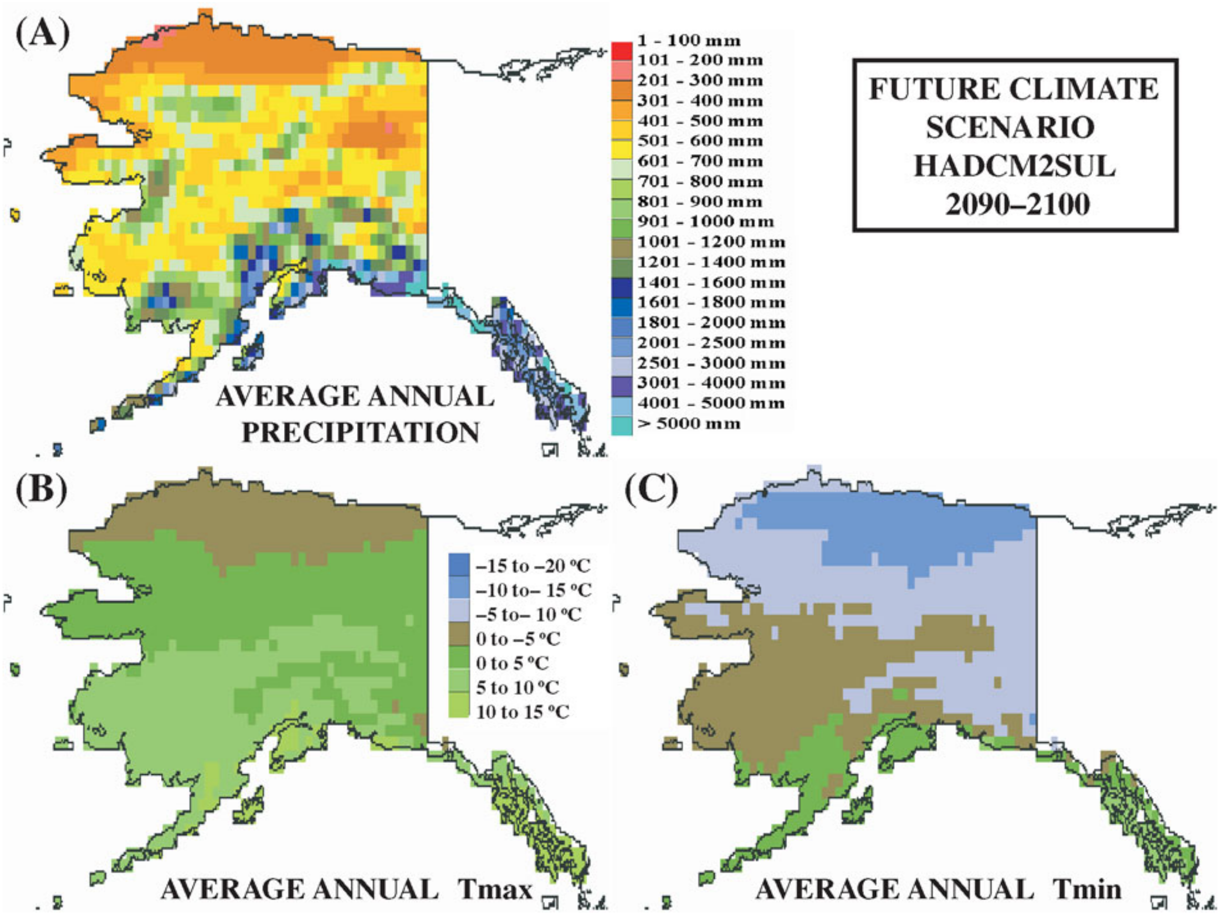
Model limitations

MC1 simulates potential vegetation dynamics without human-induced changes such as urbanization, agriculture, forest harvest, grazing, and air pollution, which can further complicate the system by possibly enhancing or impeding changes to the natural system (Chapin et al. 2004). We have assumed that given the low population density in Alaska, human impacts were limited and localized.

Fire occurrence is simulated as a discrete event, with no more than one event per year in each cell (0.5° latitude × 0.5° longitude, or 1382 km^2); thus, only large fires are represented. In the boreal region of North America, large (>400 ha) fires represent only 2%–3% of the total fires but account for about 98% of the area burned (Murphy et al. 2000 as cited by French et al. 2003). We thus feel confident our approach is well suited to simulating boreal fires. There is no constraint in the model on fire occurrence due to the availability of an ignition source, such as lightning- or human-caused ignition. Chapin et al. (2003) reported that human activities accounted for 62% of the fires in Alaska between 1956 and 2000 but only accounted for 10% of the area burned. Also, MC1 does not include fire suppression in this study. Historically, Native American tribes in Alaska, mostly fishermen and moose hunters, had little impact on the natural fire regime (Lynch et al. 2004). The state of Alaska covers 1.5×10^6 km^2, but even today only includes 600 000 inhabitants, so we have assumed that human impacts on the fire regime (such as ignition source or agents of suppression) remain limited.

The model does not include biotic disturbance agents such as pathogens and outbreak insects, which can add another layer of uncertainty associated with fire effects. Drought can stress trees and increase their susceptibility to insect attack,

Fig. 3. Future climatic conditions: (A) average annual precipitation, (B) average annual maximum temperature (Tmax), and (C) average annual minimum temperature (Tmin) for the decade 2090–2100 under HADCM2SUL.



while warm temperatures can concurrently hasten the life-cycle completion of insects. Climate change may also bring new species of insects and pathogens once previously limited by cold temperatures. Biotic disturbance regimes altered by these climate changes will further modify vegetation development trajectories and boreal carbon budgets both directly through mortality and reduction in the competitive abilities of species and indirectly through interactions with fire patterns (Malmström and Raffa 2000). Between 1989 and 1998, the mean annual area of Alaskan boreal forest experiencing active insect damage was about 9% greater than the mean annual area burned state wide (Malmström and Raffa 2000). We are currently developing new projects that will introduce insect damage in future simulations.

The model does not include wetlands, inundated landscapes, and saturated, anaerobic soils, which represent a significant fraction of the northern landscape. About 55% of Alaskan soils have been rated as moderately to well drained (Harden et al. 2001). We recognize that their inclusion in the model is essential to obtain a more accurate carbon budget, particularly as methane emissions account for a significant carbon loss, but the required input data sets are generally unavailable at a large scale (Kittel et al. 2000).

The model does not include permafrost. The location of permafrost across the state of Alaska can be estimated using published maps, but its evolution since the 1900s has not been documented accurately and thus cannot be considered as an available input to the model like a climatic variable would. Moreover, recent increases in temperature have enhanced permafrost degradation, which adds to the complexity of the issue. Permafrost thickness would be required to simulate its subsistence or disappearance across the landscape. Projections of permafrost extent and thickness in the next century that would allow simulating its interactions with changes in vegetation cover and fire regimes is a project in itself. We recognize that permafrost impacts on model results could be large. Permafrost restricts the effective depth of the soil profile. In the CENTURY-based biogeochemistry module of MC1, trees compete with grasses for soil water and nutrients, but grasses have a competitive advantage at shallow depths, while only woody life forms can reach the deeper soil layers. We could restrict the soil effective depth to mimic frozen soils, but then trees and grasses would compete for the same resources in the same shallow soil layers and the built-in competitive advantage to grasses may restrict tree growth too strongly. Permafrost affects water flow through the profile by either enhancing lateral flow and runoff, and thus limiting soil water availability in the profile, or by allowing anaerobic conditions and ponding, contributing to growth restriction or even root death. The CENTURY bucket model currently used in MC1 assumes percolation and drainage at the bottom of the soil profile and thus does not allow water accumulation in the upper soil layers above field saturation, which could mimic waterlogging. The pres-

Table 1. Comparison crosswalk table between the vegetation classes displayed in Fig. 4, the ecosystem types from the potential vegetation map of the major ecosystems of Alaska (USGS, approx. 1991), and the vegetation types simulated by MC1.

Vegetation classes (Fig. 4)	USGS Major Ecosystems of Alaska	MC1 vegetation classes	Biogeography criteria in MC1
Tundra	Alpine, Moist, Wet Tundra	Tundra	GDD ≤1
Boreal mixed forest	Lowland Spruce–Hardwood Forest, Bottomland Spruce–Poplar Forest, Upland Spruce–Hardwood Forest	Boreal mixed forest	Max. tree LAI ≥ 3.75; boreal zone; evergreen needle-leaf[a]
Maritime conifer forest	Coastal Western Hemlock–Sitka Spruce Forest	Maritime conifer forest	Max. tree LAI ≥ 3.75; temperate zone; evergreen needle-leaf[a]; CI < 16.5
Temperate conifer forest		Temperate conifer forest	max tree LAI ≥ 3.75; temperate zone; evergreen needle-leaf[a]; CI > 16.5
Woodland		Temperate conifer savanna	2 ≤ max. tree LAI < 3.75; evergreen needle-leaf[a]
Grassland		C3 grassland	Max. tree LAI ≤ 1 and max. grass LAI > 2
Shrubland	High Brush	Temperate arid shrubland	1 < max. tree LAI < 2 or max. grass LAI < 2; boreal or temperate zone; CI > 16.5
Wetland	Low Brush–Muskeg–Bog	Wetland	

Note: Biogeography thresholds used in MC1 to delineate the various vegetation types are described in the last column. CI, continentality index (maximum mean monthly temperature – minimum mean monthly temperature); GDD, annual growing degree-days; LAI, leaf area index.

[a]The tree life form is determined at each annual time step by locating the grid cell on a two-dimensional gradient of annual minimum temperature and growing season precipitation. Life-form dominance is arrayed along the minimum temperature gradient from more evergreen needle-leaf dominance at relatively low temperatures, to more deciduous broadleaf dominance at intermediate temperatures, to more broadleaf evergreen dominance at relatively high temperatures. The precipitation dimension is used to modulate the relative dominance of deciduous broadleaved trees, which is gradually reduced to zero towards low values of growing season precipitation. Life-form rules that define the tree and grass life forms are described in further detail in Daly et al. (2000).

Table 2. Fire return interval for the various vegetation types used by the model MC1 across the state of Alaska.

	Fire return interval (years)	
	Min.	Max.
Tundra	1000	7000
Boreal mixed forest	100	200
Maritime coniferous forest	350	500
Temperate coniferous forest	250	350

ence of permafrost also delays soil warming as air temperature increases during spring conditions, thus reducing the length of the growing season. In MC1, the shortest time step to delay soil warming, and thus decomposition and nutrient turnover, is 1 month, which may or not be realistic across the high-latitude gradient from coastal to continental arctic conditions.

The absence of a moss and lichen layer at the soil surface in the boreal forest is an additional weakness of the model. Mosses and lichens insulate the soil from the spring warming or the fall cooling, reinforcing the permafrost impacts on belowground processes at the beginning of the growing season. Bryophytes can also reduce evaporation from the soil surface, improving soil water availability for plant growth during the warm summers, but they can also capture rainfall and evaporate it directly, preventing it from entering the soil profile. They can be a nutrient source if they include nitrogen fixers, and they contribute significantly to fires as ground fuels. The moss layer can also delay the cessation of all belowground processes at the onset of winter.

Observations: validation data

The Alaska large-fire database used to compare model output with recorded area burned is available in ArcInfo coverage format in a draft form (http://agdc.usgs.gov/data/blm/fire/index.html; S. Rupp and S. Triplett, personal communication, 2004). Data are provided by the Bureau of Land Management, Alaska Fire Service. Between 1950 and 1987, records include only fires greater than 400 ha, but since 1987 fires greater than 40 ha have also been recorded. The map of large-fire location used in this study for validation purposes is available at http://agdc.usgs.gov/data/blm/fire/firehistory/akfirehist.jpg, and the published graph of area burned in the state of Alaska between 1950 and 1987 is available at http://fire.ak.blm.gov/weather/outlook/outlook.htm. Murphy et al. (2000) discussed in detail the limitations in the Alaska fire record.

The digital map of the Major Ecosystems of Alaska (approx. 1991) is available from the US Geological Survey at ftp://agdcftp1.wr.usgs.gov/pub/aklandchar/ecosys/ecosys.e00.gz. It is based on the map unit boundaries delineated by the Joint Federal–State Land Use Planning Commission for Alaska in 1973.

Results and interpretation

Biogeography

Historical conditions

Simulation results projecting the distribution of dominant vegetation under historical climate conditions from 1922 to 1996 were compared with the USGS potential vegetation map for Alaska (Fig. 4). The relation between the ecosystems of the USGS potential vegetation map and the vegetation types simulated by MC1, including their biogeography thresholds, is illustrated in Table 1. The model captures the broad patterns of vegetation distribution very well, with 66% of the land area dominated by tundra and 23% by the boreal mixed forest. Temperate evergreen forests dominate only 6% of the total area, mostly in southeastern Alaska, where maritime coniferous forests dominate. The discrepancies between

Fig. 4. (A) Potential vegetation map (resampled digital map of the Major Ecosystems of Alaska, US Geological Survey, approx. 1991) and simulated vegetation distribution under (B) historical climate conditions 1922–1996 and under two climate change scenarios, (C) HADCM2SUL and (D) CGCM1, for the decade 2090–2100 (data source: http://agdc.usgs.gov/data/blm/fire/firehistory/akfirehist.jpg). Note: Grey grid cells on the potential vegetation map correspond to wetlands.

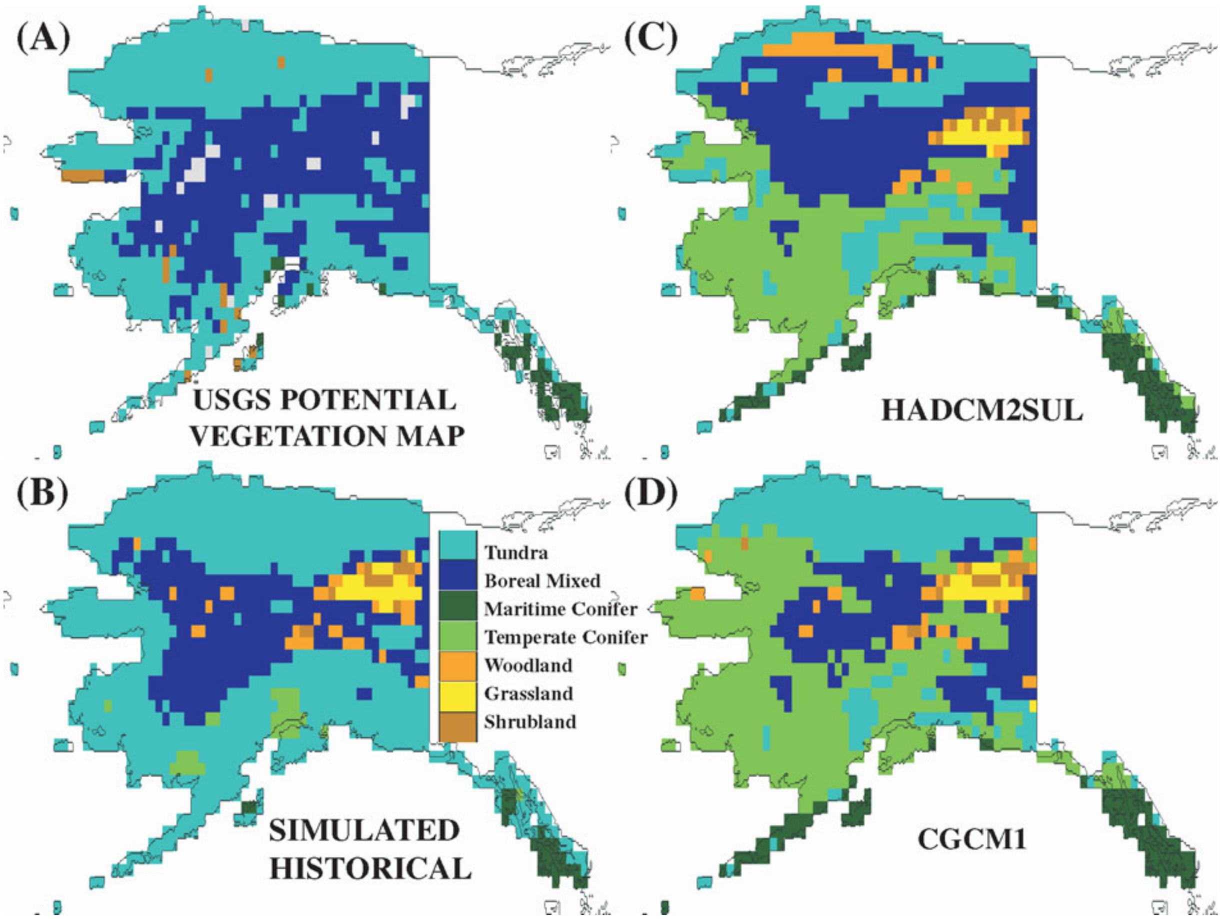
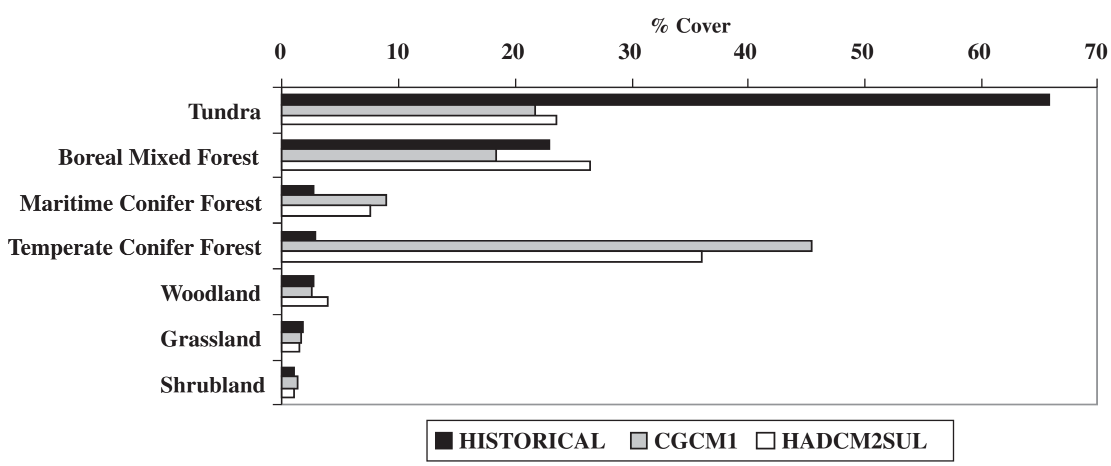
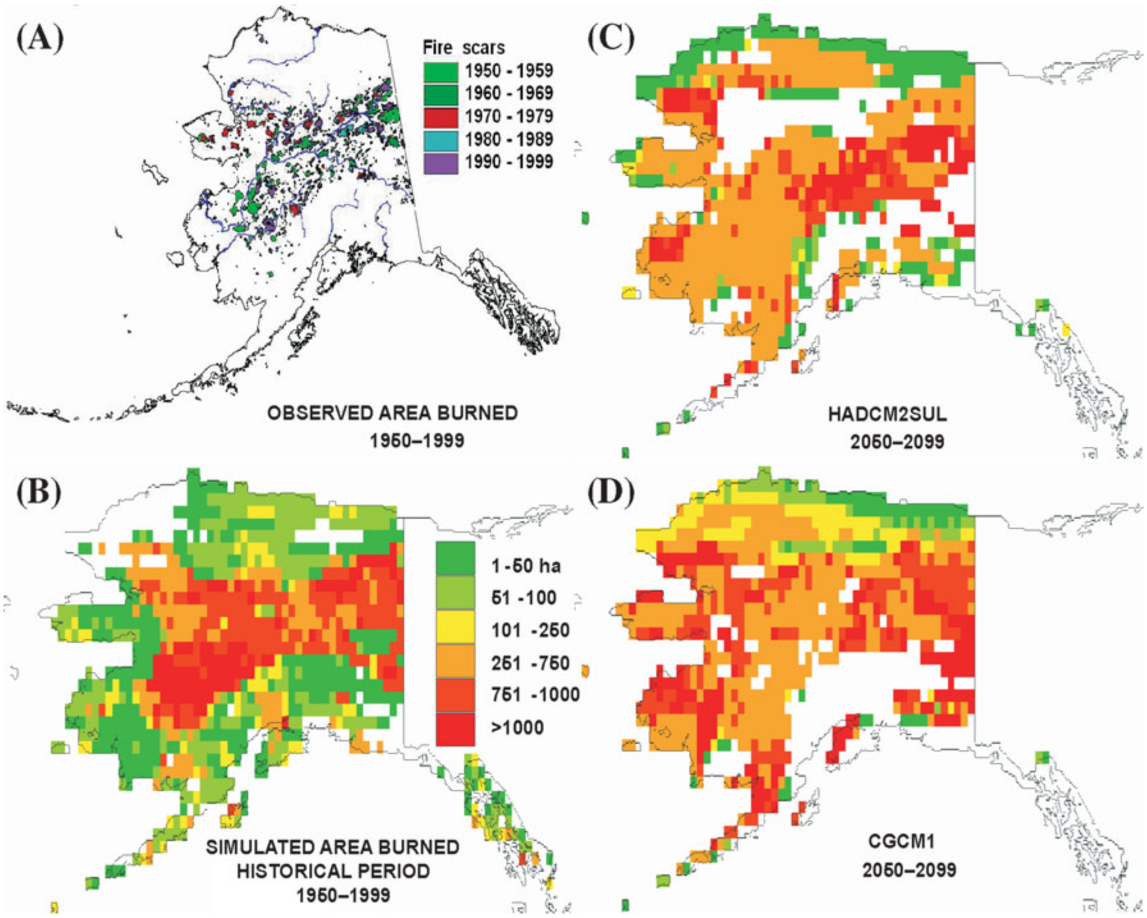


Fig. 5. Percent cover of the various vegetation types simulated across the state of Alaska under historical climate conditions (1922–1996) and under two climate change scenarios (HADCM2SUL and CGCM1).



the potential vegetation map and the simulated map depend on the biogeography thresholds defined in the model (Table 1) and the climate data that correspond to the grid cells where the discrepancy occurs (Fig. 1). In areas of southern Alaska where annual precipitation is higher but also where average annual maximum mean monthly temperature (MMT) and minimum MMT are higher than in interior Alaska, the model simulates areas of temperate coniferous instead of boreal mixed forests, which are defined in the model as colder areas where the minimum MMT drops below –16 °C (thresh-

Fig. 6. Observed (Bureau of Land Management, Alaska Fire Service) and simulated area burned in Alaska under historical climate conditions for 1950–1999 and under two climate change scenarios, HADCM2SUL and CGCM1, for 2050–2099 (data source: http://agdc.usgs.gov/data/blm/fire/firehistory/akfirehist.jpg).



old of intracellular freezing among temperate broadleaf trees). In other areas, the model simulates tundra, which is limited by a growing degree-day range between 50 and 735, instead of boreal forests, which are assumed to be less energy limited. Dry conditions in interior Alaska were interpreted by the model biogeography rules as sufficient to support grassland, while in reality the area supports a forest with low leaf area index. Simulated fires, which are frequent in this area, tend to tip the competitive balance towards grasses in the model, as most burned grasses are reestablished the year after a fire. Trees or shrubs regrow more gradually. The rapidly growing grasses gain an early advantage over woody life forms in the competition for water and nutrients, promoting even greater grass production, which in turn produces a more flammable fuel bed and more frequent fires (Lenihan et al. 2003). However, this discrepancy between simulation results and the published map only occurs on 3% of the total area.

Future conditions

Under climate change scenarios, 75% (HADCM2SUL) to 90% (CGCM1) of the tundra area simulated in 1922 disappears by the end of the 21st century. While some alpine tundra remains in the uplands of the Brooks and the Alaska ranges, arctic tundra only remains along the north coast, where arctic climate conditions maintain low minimum temperatures, especially under HADCM2SUL (Fig. 4). As both the minimum MMT and the continental index (maximum MMT – minimum MMT) increase, temperate coniferous forests greatly expand across the southern half of the state at the expense of the heat-limited tundra and the boreal forests, which are constrained by a minimum MMT of –16 °C, to cover 36% of the total area (Figs. 4 and 5). Maritime coniferous forests also expand in regions where the continental index remains below 15 °C. Interior Alaska remains dry under both climate change scenarios, and the model continues to simulate areas of grassland–shrubland that are maintained by reoccurring fires. An area of lower density vegetation also opens up in the north under CGCM1.

Fire

Historical conditions

Simulated area burned by wildfires between 1950 and 1999 compare well with records of large fires from the BLM Alaska Fire Service large-fire database (Fig. 6). Most of the fires occurred and were simulated in interior Alaska, where dry climate conditions prevail. They are associated in the model with the presence of mixed boreal forests and the open-canopy woodlands and grasslands the model projects

Fig. 7. Comparison between observed (Bureau of Land Management, Alaska Fire Service) and simulated area burned across the entire state of Alaska between 1950 and 1996 for historical climate conditions.

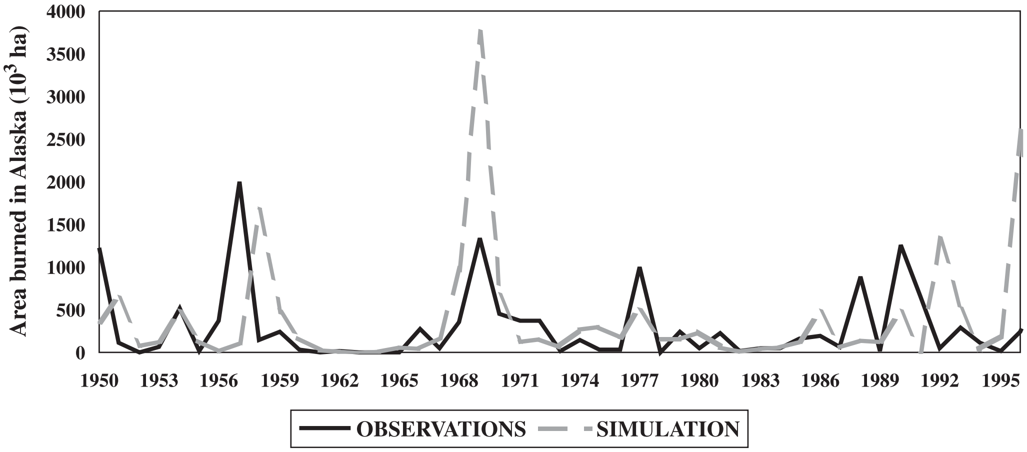
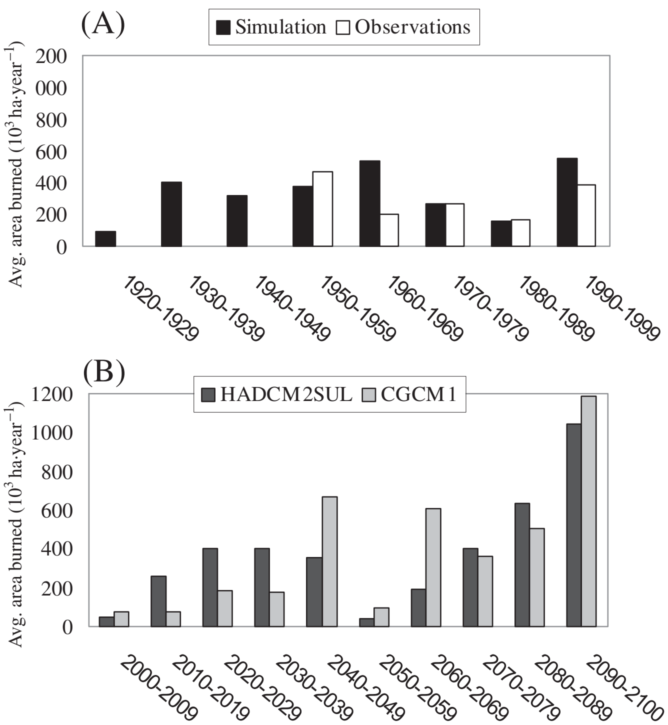


Fig. 8. Simulated annual average area burned by decade across the entire state of Alaska (A) under historical conditions (1922–1996) as compared with the decadal averages from the Alaska large-fire database (between 1950 and 1987, only fires >400 ha were reported; between 1988 and 2003, only fires >40 ha were reported) and (B) under HADCM2SUL and CGCM1 climate change scenarios.



for the dry interior region (Figs. 4 and 6). We compared simulated and observed timing of fire events during historical climate conditions for 1922–1996 (Fig. 7). The incidence of fires shows large interannual variability, which is typical of boreal regions (French et al. 2003). The average area burned between 1950 and 1996 as simulated by the model is around 360 000 ha·year^{-1}, and the average from the Fire Service records for the same period is 293 000 ha·year^{-1}. The overestimation of area burned by about 22% on average is partly due to actual constraints on ignition, which never occur in the model. Uncertainty in the records and the lack of accounting for smaller fires (<400 ha) may also account for some of the difference between the observed and simulated area burned (French et al. 2004). Even though wet soils would also be expected to limit the extent of fires, burned area was found to be highly correlated with wet soils in Alaska, where fuels and fire weather overcome soil wetness (Harden et al. 2001). Underlying permafrost would simply limit the fire impacts to shallow soil layers. The model lags a few years in simulating a few fires but generally compares well in time and extent with the observations. French et al. (2003) used empirical relationships to calculate that 1 year of burning could release as much as 36 Tg C or 62 g C·m^{-2} into the atmosphere in a big fire year, and they showed that emissions could vary widely from year to year. Our results show an annual average of 14 Tg C or 9 g C·m^{-2} released by wildfires, with large interannual variability.

Future conditions

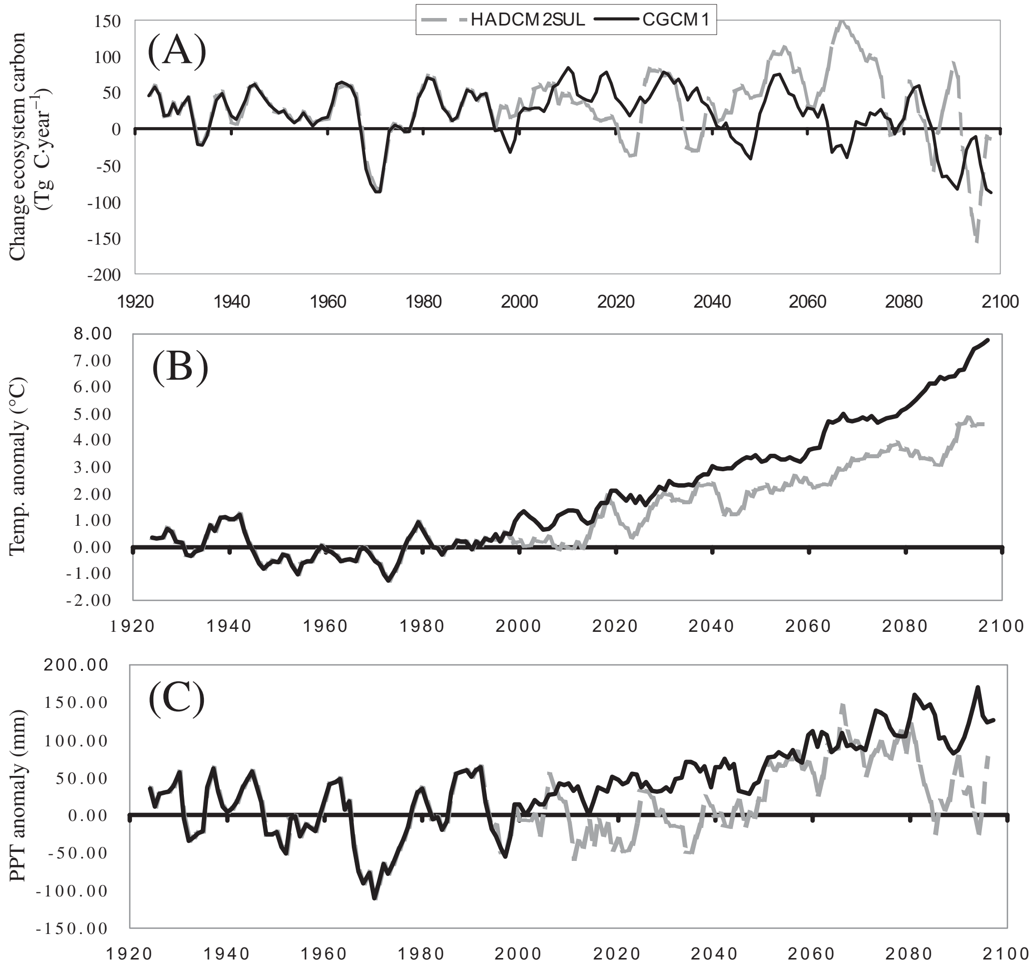
The model projects a northern expansion of temperate forests across the southern half of the state, primarily into tundra, which promotes more fires in the region because forests have a shorter fire return interval than tundra (Table 2). The simulated area burned state wide is 17%–39% greater between 2050 and 2100 than what was simulated between 1950 and 2000, with large interannual and interdecadal variability (Fig. 8). The projected area burned is larger under CGCM1 in the middle of the 21st century, when temperature differences between the two scenarios reach 2 °C. However, the greatest area burned occurs in the last two decades of the 21st century under both scenarios (16×10^6 ha burned), when large fluctuations in projected precipitation create drought conditions. The model simulates an average loss of 17–19 Tg C·year^{-1} due to fire emissions between 2025 and 2099, which corresponds to a 24%–33% increase above historical conditions.

Carbon budget

Historical conditions

The change in ecosystem carbon or net biological production (NBP) (Fig. 9A) shows periods of carbon storage with peaks in the mid 1940s, 1960s, and 1980s (70 Tg C) and pe-

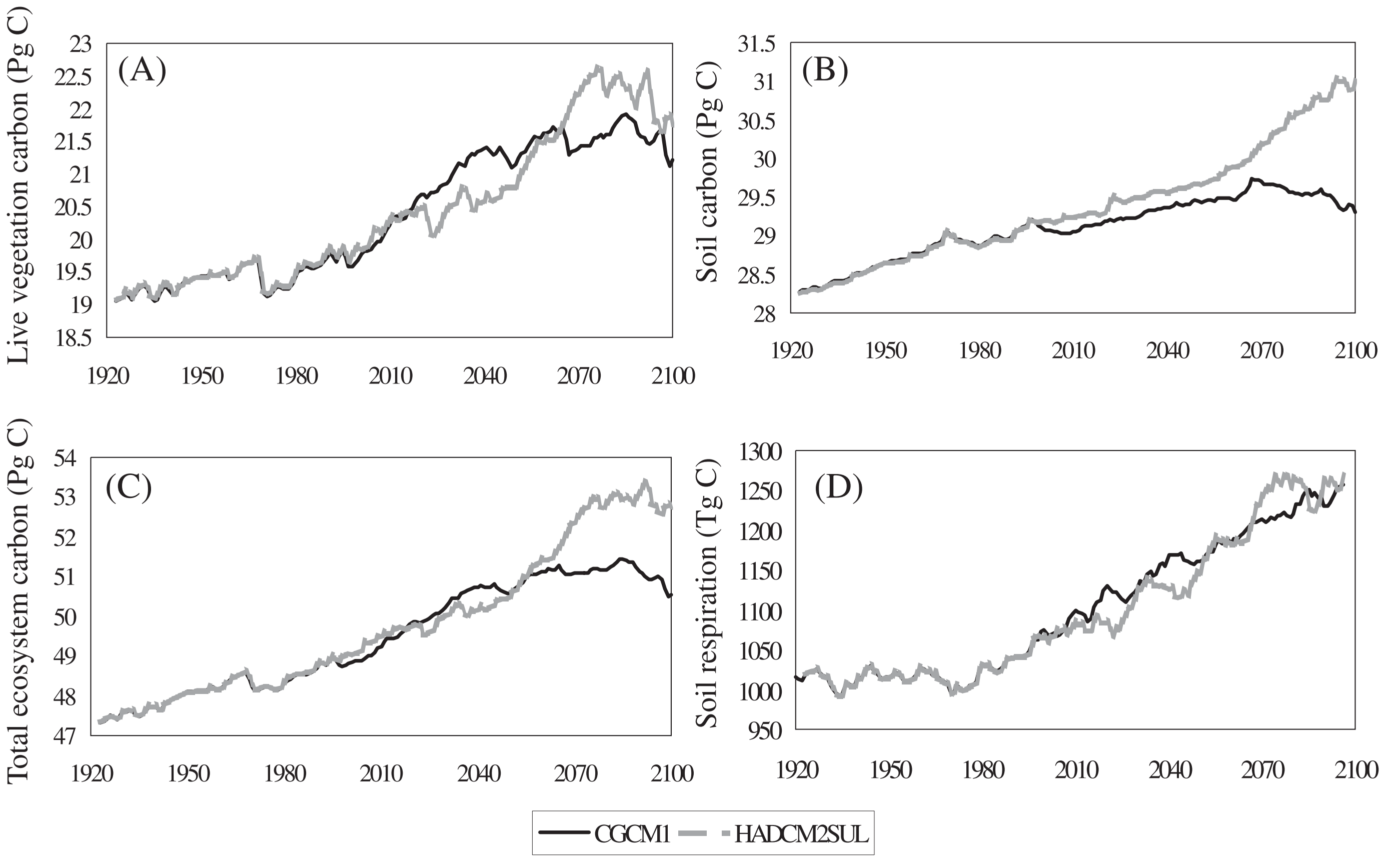
Fig. 9. (A) Five-year running-average change in ecosystem carbon (net biological production in Tg C·year^{-1}) under historical conditions 1922–1996 and under two climate change scenarios compared with variations in 5-year running-average (B) annual maximum temperature (Tmax) and (C) annual precipitation (PPT) normalized to the 1922–1996 average.



riods of carbon release in the mid 1930s and 1970s (86 Tg C). The 1976–1977 climate regime shift, which represents the most important long-term change during the period of record (Ebbesmeyer et al. 1990 as cited by Juday et al. 2003), is clearly visible in the climate record (Figs. 9B, 9C) and caused a strong response in the NBP trace. The simulated average NBP across the entire state is 16 g C·m^{-2}·year^{-1} (standard deviation (SD) = 42 g C·m^{-2}·year^{-1}), which compares well with the range of 18–53 g C·m^{-2}·year^{-1} estimated by Harden et al. (2000) with a mass balance model, 42 g C·m^{-2}·year^{-1} estimated from eddy correlation studies by Litvak et al. (2003) for a dry boreal forest stand, 41 g C·m^{-2}·year^{-1} for another dry boreal forest site, and 86 g C·m^{-2}·year^{-1} for a wet boreal forest stand, also in Canada, estimated by Bond-Lamberty et al. (2004). Another modeling study by Potter (2004) estimated the NBP of a relatively young taiga forest around Denali at 26 g C·m^{-2}·year^{-1} and of a tundra site at 1 g C·m^{-2}·year^{-1} using the DGVM NASA - Carnegie Ames Stanford Approach (CASA) bracketing the most different ecosystems in interior Alaska.

Estimates of net ecosystem production for the boreal region range from –8 g C·m^{-2}·year^{-1}, a small net source, to 130 g C·m^{-2}·year^{-1}, a net sink (French et al. 2003), without including losses to fires. MC1 simulates an average net ecosystem production value of 26 g C·m^{-2}·year^{-1} (SD = 42 g C·m^{-2}·year^{-1}) encompassing all ecosystems of Alaska between 1922 and 1996, with an average of 9 g C·m^{-2}·year^{-1} lost to wildfires. Harden et al. (2000) used estimates for NPP between 150 and 430 g C·m^{-2}·year^{-1} for the boreal region, and Bond-Lamberty et al. (2004) reported NPP values for a Canadian boreal wildfire chronosequence varying between a low of 50 g C·m^{-2}·year^{-1} following a fire and 521 g C·m^{-2}·year^{-1}. MC1 simulated an average of 667 g C·m^{-2}·year^{-1} (SD = 49 g C·m^{-2}·year^{-1}) for the state of Alaska. The model may have overestimated Alaskan boreal ecosystem NPP because, first, it does not limit nutrient availability sufficiently at these higher latitudes for which scarce data on nutrient inputs and soil content are available as reliable inputs to the model. Secondly, the model does not include a permafrost layer and thus probably overestimates water and nutrient availability through the entire soil profile. Thirdly, growth restriction due to frozen soils is also not accounted for in this version of the model. Field-based studies such as the Ameriflux micrometeorological flux towers network can provide crucial validation information at the local scale but are less useful as validation tools at the regional to continental scale. Moreover, there are four tower sites in Alaska, all in the tundra and none in the forests (Potter 2004). Conse-

Fig. 10. Simulated (A) live vegetation carbon, (B) soil carbon (including litter), (C) total ecosystem carbon, and (D) ecosystem respiration in Alaska under historical conditions and under two future climate change scenarios, HADCM2SUL and CGCM1.



quently, model validation of carbon fluxes in the different vegetation types in Alaska are hindered by the scarcity of observations.

Future conditions

For most of the 21st century and under both scenarios, variations in NBP show few periods of carbon losses comparable in magnitude to the 1970 loss but extensive periods of carbon gains, which are particularly large under HADCM2SUL. However, during the last decade of the 21st century, Alaska becomes a large carbon source (19–72 Tg C) under both scenarios (Fig. 9).

Both live vegetation and soil carbon (Fig. 10A) show increasing trends throughout much of the century under both future climate change scenarios. However, under CGCM1 the increase in soil carbon starts to level off early in 21st century, and the trend in live vegetation carbon follows suit by mid-century (Figs. 10A, 10B). Both carbon pools decline during the last two decades of the CGCM1 scenario. Soil respiration under this warmer scenario does not plateau at all, which clearly indicates that soil carbon decomposition is stimulated by the warming temperature and is affecting not only the most recently produced litter but also the longer term soil carbon pools (Fig. 10D). Under both scenarios, ecosystem carbon increases until about 2050, when it levels off under CGCM1, a couple of decades earlier than under HADCM2SUL.

Discussion

The paleoecological literature generally supports the hypothesis that rising temperatures at high latitudes will eventually lead to an elevational or latitudinal advance of tree line (Lloyd et al. 2003) so that tundra will be replaced by forests (Chapin et al. 2004). Climate changes observed during the 20th century (ex. Serreze et al. 2000) may in fact have already begun to affect vegetation distribution, as boreal forest expansion into tundra has been observed by Suarez et al. (1999) in northwestern Alaska and an increase in shrub biomass in northern Alaska tussock tundra has been reported by Chapin et al. (1995). Previous modeling efforts have projected large decreases in tundra area (–40% to –67%) under various climate change scenarios (Neilson et al. 1998). However, associated changes in disturbance regime (thermokarst activity and wildfires), variations in soil fertility, and slower tree growth in response to warmer and drier conditions (Barber et al. 2000) may modulate the response of the tree line to warming. Our model simulates a loss of 75%–90% of the tundra area as temperate forests expand across the southern portion of the state. Warm but also wet scenarios allow more temperate species to outcompete slow-growing cold-adapted vegetation. The inclusion of permafrost in the model might have delayed the abrupt changes from cold to temperate flora that are simulated to occur before the end of the 21st century.

Table 3. Annual average of carbon fluxes and pools across the state of Alaska for historical conditions and for two future climate change scenarios simulated by MC1 (standard deviations are in parentheses).

		Future (2025–2099)	
	Historical (1922–1996)	HADCM2SUL	CGCM1
Carbon flux (g C·m^{-2}·year^{-1})			
Net primary production	667 (49)	790 (44)	766 (74)
Soil respiration	641 (12)	747 (26)	744 (38)
Net ecosystem production	26 (42)	43 (40)	22 (57)
Fire emissions	9 (17)	12 (18)	12 (26)
Net biological production[a]	16	31	10
Carbon pool (kg C·m^{-2})			
Live vegetation carbon	12 (0.14)	13 (0.15)	13 (0.48)
Litter and soil carbon	18 (0.16)	19 (0.08)	19 (0.34)
Total ecosystem carbon[b]	30	33	32
Area burned (10^3 ha·year^{-1})	360	411	481

[a]Net biological production = net ecosystem production – fire emissions.
[b]Total ecosystem carbon = vegetation + soil carbon.

Fire records necessary to validate model results are difficult to obtain and sources can disagree. For example, the BLM Alaska Fire Service estimates that about 14.5 × 10^6 ha was burned in Alaska between 1950 and 1997, which is about 20% more than was reported by Birdsey and Lewis (2002). Our model simulates about 19 × 10^6 ha burned in Alaska during the same period. Our overestimation may be due in part to the lack of any fire suppression in our simulations, which, even if limited in Alaska, may affect actual fire occurrence since 1950. Most likely, however, it is due primarily to the lack of constraints on ignition source and, secondarily, the overestimation of fuel biomass due to the lack of permafrost and waterlogged conditions, which would otherwise limit vegetation growth and fuel loading. Under future climate scenarios (2025–2099), MC1 simulates a 14% (HADCM2SUL) to 34% (CGCM1) increase in area burned in comparison with the period 1922–1996. These results agree well with Flannigan et al. (2000), who found that an index of fire severity with a linear relationship to area burned (Flannigan and van Wagner 1991) increased by about 15% and 30% in Alaska under two future climate scenarios generated by the Hadley and Canadian GCMs, respectively. The increase in fire is due not only to warmer temperatures under both scenarios, but also to the resulting changes in vegetation type, since the temperate forests that replace the tundra in the southern portion of the state are assumed to have shorter fire return intervals (based on historical reconstruction of fire).

Fire plays a central role in carbon cycling by releasing carbon into the atmosphere, regulating vegetation cover, and thus controlling carbon accumulation patterns (French et al. 2003). Carbon released from fires in Alaska has been estimated for specific years (Kasischke et al. 1995*b*) or simulated over multiple years (French et al. 2003) but long-term estimates of carbon dynamics during the 20th century are only available as model results. Using historical climate records between 1922 and 1996, simulation results show a loss of about 1 Pg C (SD = 25 Tg) through fire emissions from almost 27 × 10^6 ha burned in Alaska. During the same period, the model simulates a carbon gain of 1.7 Pg C (SD = 60 Tg) over the entire state of Alaska (151 × 10^6 ha), which means the model simulates a loss to fires of 61% of the carbon gained in 75 years. Under the HADCM2SUL future climate change scenario, fire emissions account for 1.3 Pg C on 30 × 10^6 ha between 2025 and 2099. Under the CGCM1 scenario, fire emissions account for 1.4 Pg C on 36 × 10^6 ha. The carbon gain over the state is projected at 0.5 Pg C under CGCM1 and 3.2 Pg C under HADCM2SUL. The loss to fires under the warmer climate change scenario CGCM1 is thus greater than the carbon gained in those 75 years, while under HADCM2SUL it represents only about 40%.

Model results provide some sense of how the state of Alaska biosphere could change along two trajectories, one near the mild end of the temperature change gradient and one near the warmer end. The moderately warm HADCM2SUL scenario produces increased vegetation growth throughout the 21st century with a decline in the last decade. The warmer Canadian scenario produces more drought stress, resulting in less vegetation growth and more fire. However, both scenarios result in an overall net carbon gain over the 21st century (Table 3). Under CGCM1, the net gain is less than under historical climate, suggesting that there might be a warming threshold transition point below which plants can thrive and their carbon uptake can contribute to an “early green-up” phase. However, above this temperature threshold, regional drought-stress can occur and cause net carbon emissions (e.g., from drought and fires), which can trigger a “later brown-down”. Alaska may lose its tundra ecosystem if warming scenarios come to pass, but its carbon storage will continue to increase as expanding forests, under a favorable precipitation regime, will fix carbon and store it in the soil. However, a high temperature threshold might be reached that could reduce tree growth, and Barber et al. (2000) show it may already have happened for white spruce in the boreal forest of interior Alaska. Only when tree growth is reduced and decomposition stimulated by soil warming will the high-latitude large carbon store be at risk.

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