Yukon Water Availability Analysis

An Assessment of the Potential Impacts of Climate Change on the Balance between Precipitation and Potential Evapotranspiration in the Yukon, Canada

Prepared for
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by the
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July 2011
**Executive Summary**

The high-latitude ecosystems of the Yukon are vulnerable to climate change, including hydrologic changes. In order to predict potential changes in growing season water balance in the Yukon, the Scenarios Network for Alaska and Arctic Planning (SNAP) and Yukon College collaborated to develop a modeling tool for mapping future growing-season water availability. The model focused on estimating the growing-season balance between precipitation (P) and potential evapotranspiration (PET), a term used to describe the likely amount of water that could be returned to the atmosphere through the combination of evaporation and transpiration. Results showed that much of the Yukon is likely to remain water-limited during summer months, with the balance between P and PET remaining negative. Subtle changes were predicted in this balance, with some regional drying, particularly in the boreal regions. However, the greatest impacts to ecosystem hydrology may stem from associated climate-driven changes such as increases in growing season length and growing degree days and associated vegetation shifts; changing drainage from permafrost loss; and altered fire cycles.

**Introduction**

The global climate system is changing. The high-latitude ecosystems of the northern hemisphere, including the Yukon, have been identified among the most vulnerable, and the region is already showing evidence of climate change. Increases in temperature and changes in precipitation can have had profound effects on regional hydrology, including shrinking wetlands, ice recession, permafrost thaw, and an increase in fire frequency and intensity across the landscape as a result of increased drought and thunderstorms. Continuation of these trends will likely lead to further changes in the hydrologic cycle, with significant implications for the people and ecosystems that depend on the Yukon’s water resources. Because water availability is an important determinant of ecosystem structure and function, understanding changes during periods of peak biological demand is of critical importance. Recent efforts have sought to synthesize observational evidence with satellite-based analysis (e.g. Zhang et al. 2009) and projections from global and regional climate models to develop a better understanding of how biological communities may adapt.

In an effort to better understand where and when changes in hydrology are likely to occur, the Scenarios Network for Alaska and Arctic Planning (SNAP) and Yukon College collaborated to develop a tool for mapping future growing-season water availability in the Yukon, based on a similar tool developed by SNAP and The Wilderness Society in Alaska, USA.

This study is focused on water balance during the warm season. During the growing season, when biological demand is highest, evapotranspiration (ET) becomes the driving mechanism of landscape water loss. The term potential evapotranspiration (PET)
is used to describe the likely amount of water that could be returned to the atmosphere through the combination of evaporation and transpiration.

In much of the boreal and arctic, PET during growing season months typically exceeds incoming precipitation, resulting in an overall water deficit during this time (Woo et al. 1992). As the climate continues to warm and the growing season gets longer, scientists expect PET and precipitation (P) will both increase. If the increase in water lost from the landscape through PET is not offset by an equivalent increase in incoming P, the Yukon may experience more severe water-deficits during the growing season. This prediction is corroborated not only by anecdotal reports of landscape drying, but also by direct studies of the relationship between water availability and vegetation in the Yukon (e.g. Hogg and Wein 2005) and by observations of increased growing season ET based on satellite imagery (Zhang et al 2009).

PET is determined by the energy available to evaporate water, measured as temperature, and other environmental conditions including wind, cloudiness, plant growth, and humidity. The annual surface water balance can be simply estimated from the difference between precipitation and potential evapotranspiration (P-PET).

Estimates of PET reflect the amount of energy available to evaporate water from a saturated surface, and can be calculated from temperature data where direct measurements are not available. Over the latter half of the twentieth century, temperature-driven increases in summer evapotranspiration appear to have been partially responsible for net declines in summer water availability in arctic and boreal areas (Hinzman et al. 2005). These observations have led to hypotheses that continued increases in average temperatures may cause future evapotranspiration rates to exceed predicted increases in precipitation, thereby exerting increased drying across the landscape (Rouse et al. 1992; Rouse 2000).

Historically, precipitation in the Yukon tends to be heaviest during summer months, as shown by Climate Normals from Environment Canada. However, it should be noted that because the winter season is long, roughly 30-50% of the Yukon’s precipitation occurs as snow (Appendix A), accumulating in the snowpack and contributing to water storage across the landscape. The literature reflects some uncertainty as to the percentage of this wintertime storage lost through sublimation, with estimates ranging from approximately 10% to 45% (Pomeroy et al. 1999; Janowicz et al. 2004; Carey and Woo 2001; Pomeroy et al. 2006). The remainder is converted to runoff that serves to recharge rivers, lakes, and soils. Due to the uncertainty of sublimation estimates (Shutov et al 2006), the lack of data needed to drive a sublimation model, and the fact that the effects of sublimation on groundwater recharge are highly variable and localized, winter water storage is only indirectly accounted for in this study.
SNAP Climate Data and Modeling

Derivation of SNAP Climate Projections for Canada

The projections made in this project relied upon SNAP projections for future temperature and precipitation in the Yukon. These projections were derived from down-scaled outputs from General Circulation Models (GCMs), as described below.

GCMs are the most widely used tools for projections of global climate change over the timescale of a century. Periodic assessments by the Intergovernmental Panel on Climate Change (IPCC) have relied heavily on complex coupled atmospheric and oceanic GCMs driven by various emission scenarios. These models integrate multiple equations, typically including surface pressure; horizontal layered components of fluid velocity and temperature; solar short wave radiation and terrestrial infra-red and long wave radiation; convection; land surface processes; albedo; hydrology; cloud cover; and sea ice dynamics. GCMs include equations that are iterated over a series of discrete time steps as well as equations that are evaluated simultaneously. Anthropogenic inputs such as changes in atmospheric greenhouse gases can be incorporated into stepped equations. Thus, GCMs can be used to simulate the changes that may occur over long time frames due to the release of excess greenhouse gases into the atmosphere.

Different coupled GCMs have different strengths and weaknesses, and some can be expected to perform better than others for northern regions of the globe. SNAP climate researchers John Walsh and Bill Chapman evaluated the performance of a set of fourteen GCMs used in the Coupled Model Intercomparison Project. This analysis was performed for several regions, some of which overlapped: the majority of the northern hemisphere (20°-90°N); the far north (60°-90°N); Alaska; Greenland; and regions of western Canada, including all of the Yukon and BC and western NWT. Using the outputs for models’ 20th-century simulations (run with historical greenhouse gas and aerosol concentrations), they calculated the degree to which each model’s output concurred with actual climate data for the years 1958-2000 for each of three key climate variables (surface air temperature, air pressure at sea level, and precipitation).

The core statistic of the validation was a root-mean-square error (RMSE) evaluation of the differences between mean model output for each grid point and calendar month, and data from the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis, ERA-40. To facilitate GCM intercomparison and validation against the ERA-40 data, all monthly fields of GCM temperature, precipitation and sea level pressure were interpolated to the common 2.5° × 2.5° latitude–longitude ERA-40 grid. For each model, Walsh and Chapman calculated RMSEs for each calendar month and the three key climate variables. Because the errors were summed over all calendar months and all grid points in northwestern Canada, they represent the models’ ability to capture the spatial patterns within the domain over the full seasonal cycle.

In order to create climate projections for Alaska, SNAP has been relying on the five models that were the highest ranked when their performance in the two broader northern regions (20°-90°N and 60°-90°N) is taken into consideration in addition to their performance in Alaska alone. Using the same strategy, the same five models were used for climate
projections in Northern Canada. These five include MPI_ECHAM5 (Max Planck Institute for Meteorology, Hamburg Germany), GFDL_CM2_1 (Geophysical Fluid Dynamics Laboratory, NOAA, U.S.), MIROC3_2_MEDRES (the Center for Climate System Research at the University of Tokyo, the National Institute for Environmental Studies, and the Frontier Research Center for Global Change, Japan), UKMO_HADCM3 (Hadley Centre for Climate Prediction and Research, United Kingdom), and CCCMA_CGCM3_1 (Canadian Centre for Climate Modelling and Analysis). Detailed documentation for each model is available online at the CMIP3 website (http://wwwpcmdi.llnl.gov/ipcc/model_documentation/ipcc_model_documentation.php).

Downscaling model outputs

Because of the enormous mathematical complexity of GCMs, they generally provide only large-scale output, with grid cells typically 1°-5° latitude and longitude. For example, the standard resolution of HadOM3 is 1.25 degrees in latitude and longitude, with 20 vertical levels, leading to approximately 1,500,000 variables. Finer scale projections of future conditions are not directly available. However, local topography can have profound effects on climate at much finer scales, and almost all land management decisions are made at much finer scales. Thus, some form of downscaling is necessary in order to make GCMs useful tools for regional climate change planning.

Historical climate data estimates for the Yukon at 2km resolution are available from PRISM (Parameter-elevation Regressions on Independent Slopes Model), which was originally developed to address the lack of climate observations in mountainous regions or rural areas. PRISM uses point measurements of climate data and a digital elevation model to generate estimates of annual, monthly and event-based climatic elements. Each grid cell is estimated via multiple regression using data from many nearby climate stations. Stations are weighted based on distance, elevation, vertical layer, topographic facet, and coastal proximity. PRISM downscaling has been used in northwestern Canada by other researcher groups linking ecosystem processes to GCM climate change predictions (e.g. Wang et al. 2005), and has been shown to be particularly well-suited for modeling precipitation in mountainous regions of the Yukon and BC (Hamann and Wang 2005; Mbogga 2009).

PRISM offers data at a fine scale useful to land managers and communities, but it does not offer climate projections. Thus, SNAP needed to link PRISM to GCM outputs. This work was also done by John Walsh, Bill Chapman, et al. They first calculated mean monthly precipitation and mean monthly surface air temperature for PRISM grid cells for 1961-1990, creating PRISM baseline values. Next, they calculated GCM baseline values for each of the five selected models using mean monthly outputs for 1961-1990. They then calculated differences between projected GCM values and baseline GCM values for each year out to 2099 and created “anomaly grids” representing these differences. Finally, they added these anomaly grids to PRISM baseline values, thus creating fine-scale (2 km) grids for monthly mean temperature and precipitation for every year out to 2099. This method effectively removed model biases while scaling down the GCM projections.
Modeling Potential Evapotranspiration (PET) and Water Availability (P-PET)

A variety of methods can be used to estimate evaporation from land surfaces and resulting moisture balance, each with its own set of benefits and drawbacks (Shutov et al. 2006). Methods are selected based on available data and model reliability. For this project, the primary model used to estimate PET was the Preistley-Taylor model:

\[
PET = \left( \frac{1}{\lambda} \right) \cdot \alpha \cdot \left( \frac{s}{s + \gamma} \right) \cdot (R - G)
\]

where,
- \( \text{PET} \) potential evapotranspiration [mm day\(^{-1}\)]
- \( \lambda \) latent heat of vaporization of water at 20°C = 2.45 \([\text{MJ kg}^{-1}]\)
- \( \alpha \) adjusts PET for surface characteristics = 1.26 [unitless]
- \( s \) slope of the curve of the saturation vapor pressure curve [kPa °C\(^{-1}\)]
- \( \gamma \) psychrometric constant [kPa °C\(^{-1}\)]
- \( R \) net radiation [MJ m\(^{-2}\) day\(^{-1}\)]
- \( G \) heat flux from the ground surface [MJ m\(^{-2}\) day\(^{-1}\)]

Note that all calculations were performed with monthly averages; the daily rates derived here can then be multiplied by the number of days in a month to get accumulated monthly PET.

\[
\left( \frac{s}{s + \gamma} \right) - \text{slope of the vapor pressure curve and the psychrometric constant}
\]

\[
\frac{s}{s + \gamma} = 0.406 + 0.011T
\]

where,
- \( T \) is mean monthly air temperature [°C]

This equation was developed for and tested in boreal forest and tundra environments. It is unclear whether it is entirely appropriate for strongly marine influenced systems or high alpine environments.

1 Although \( \lambda \) varies with temperature, the variation is not large: \( \lambda_{0°C} = 2.50 \text{ MJ kg}^{-1} \), while \( \lambda_{100°C} = 2.25 \text{ MJ kg}^{-1} \). The conversion from kg water to mm m\(^{-2}\) is \( \frac{kg}{\text{kg}} \cdot \frac{(10^3 g)}{g} \cdot \frac{\text{cm}^2}{10^4 \text{cm}} \cdot \frac{10 \text{mm}}{\text{cm}} = 1 \).
R - net radiation

\[ R = (1 - \alpha) \cdot R_s - R_l \]

where,

\( \alpha \) is fractional albedo
\( R_s \) is incoming shortwave radiation [MJ m\(^{-2}\) day\(^{-1}\)]
\( R_l \) is net longwave radiation in the outgoing direction [MJ m\(^{-2}\) day\(^{-1}\)]

Note that if \( R_l \) is calculated by the standard sign conventions, this equation must be written as

\[ R = (1 - \alpha) \cdot R_s + R_l \]

a - albedo

<table>
<thead>
<tr>
<th>Surface</th>
<th>Albedo</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open Water</td>
<td>0.06</td>
</tr>
<tr>
<td>Wetland Tundra</td>
<td>0.15</td>
</tr>
<tr>
<td>Upland Tundra</td>
<td>0.16</td>
</tr>
<tr>
<td>Boreal Coniferous Forest</td>
<td>0.08</td>
</tr>
<tr>
<td>Boreal Deciduous Forest</td>
<td>0.16</td>
</tr>
<tr>
<td>Barren Land</td>
<td>0.20</td>
</tr>
<tr>
<td>Perennial Ice/Snow</td>
<td>0.40</td>
</tr>
</tbody>
</table>

This project used static albedo values. This is likely to introduce some error, as albedo would typically be higher when snow is present. Since albedo values in Eugster et al. (2000) are primarily mid-summer overcast-day minimum albedo values, they are at the low end of published values (see Betts and Ball 1997; Duchon and Hamm 2006). The albedo value used here for water is probably for low- or mid-latitude solar incidence angles and could likely be higher under polar light conditions (Barry and Chorley 2003). However, we are using the values selected by B. O'Brien.

Rs - incoming shortwave radiation at the surface

\[ R_s = k Ra(T_{max} - T_{min})^{0.5} \]

where,

\( k \) Hargreaves coefficient a constant [°C\(^{-0.5}\)]
\( Ra \) is solar radiation at the top of the atmosphere [MJ m\(^{-2}\) day\(^{-1}\)]
Tmax and Tmin are monthly average maximum and minimum temperature [°C]

The Hargeaves coefficient is a constant set to 0.16 in the interior and 0.19 in regions deemed to have a marine influence (Allen et al. 1998). It is likely that the spatial distribution and extent of areas that experience predominantly marine vs. interior airmasses may change in the future; thus our use of static k values may be somewhat inaccurate.

**Ra - incoming solar radiation at the top of the atmosphere**

\[
Ra = \frac{24 \cdot 60}{\pi} \cdot d \cdot S \cdot [\omega \sin(\phi) \sin(\delta) + \cos(\phi) \cos(\delta) \sin(\omega)]
\]

where,
- \( S \) is the solar constant [0.082 MJ m^{-2} min^{-1}]
- \( \omega \) is the sunset hour angle [radians]
- \( d \) is the inverse of the Earth-Sun distance
- \( \phi \) is latitude [radians]
- \( \delta \) is the declination [radians]

\( w \) - sunset hour angle

\[\omega = \cos^{-1}[-\tan(\delta) \tan(\phi)]\]

\( d \) - inverse earth-sun distance

\[\delta = 0.409 \cdot \sin\left(\frac{2\pi}{365} J - 1.39\right)\]

\( \delta \) - declination

\[d = 1 + 0.033 \cos\left(\frac{2\pi}{365} J\right)\]

where,
- \( J \) is the Julian Day of the year.

When the sun does not rise \( \omega \) is set equal to 0, and when the sun does not set \( \omega \) is set equal to \( \pi \). In order to calculate Ra at a monthly time step, we calculated average daily radiation for each day within the month and then average across the month.
RI - net longwave radiation in the outgoing direction

$$R_l = -f \varepsilon \sigma (T_{ave} + 273.15) I^4$$

where,
\begin{align*}
f &\text{ is a cloud factor, calculated below} \\
\varepsilon &\text{ is the emissivity, calculated below} \\
\sigma &\text{ is the Stefan-Boltzmann constant } [4.903 \times 10^{-9} \text{ MJ K}^{-4} \text{ m}^{-2} \text{ day}^{-1}] \\
T_{ave} &\text{ is average temperature } [°C]
\end{align*}

Cloud factor

$$f = \frac{R_s}{R_{cs}}$$

where,
\begin{align*}
R_s &\text{ is incoming shortwave radiation } [\text{MJ m}^{-2} \text{ day}^{-1}] \\
R_{cs} &\text{ is clear-sky shortwave radiation}
\end{align*}

Clear-sky shortwave radiation

$$R_{cs} = (0.75 + 2 \times 10^{-5} z) \times Ra$$

where,
\begin{align*}
z &\text{ is elevation } [\text{m}]^2 \\
Ra &\text{ is solar radiation at the top of the atmosphere } [\text{MJ m}^{-2} \text{ day}^{-1}] \\
\varepsilon &\text{ is used as a proportionality coefficient here, so the m units don't carry through and } R_{cs} \text{ remains in MJ m}^{-2} \text{ day}^{-1}
\end{align*}

Emissivity

$$\varepsilon = -0.02 + 0.261 e^{-0.00777 T_{ave}}$$

This equation produces net longwave radiation with the common sign convention. To use the net radiation equation used here, remove the leading minus sign.
Temperature

All temperatures needed are calculated from SNAP-downscaled CRU or GCM output, in combination with the 1961-90 PRISM climatology.

\[
PRISM.Tave = 0.5 \times (0.1 \times PRISM.Tmax + 0.1 \times PRISM.Tmin)
\]

\[
Tmax = 0.1 \times PRISM.Tmax + (Tave - PRISM.Tave)
\]

\[
Tmin = 0.1 \times PRISM.Tmin + (Tave - PRISM.Tave)
\]

Note that by calculating Tmax and Tmin in this way, we keep the diurnal temperature range constant over time. However, it is likely that the difference between maximum and minimum temperatures will vary with climate change. Average temperature and precipitation values for the Yukon are projected to increase over the course of the next century.

Results and Discussion

The existing negative water balance during the growing season across much of the Yukon is expected to persist in coming decades (Figure 1), with growing season precipitation significantly exceeding demands only in the Pacific Maritime ecoregion in the southwestern corner of the territory (Figure 2, Atlas of Canada), where the moisture balance is expected to remain highly positive. Given that this ecosystem is not moisture limited, it is unlikely that small changes in either direction would produce significant ecosystem shifts.

For the rest of the territory, including boreal and arctic regions, the balance between temperature-driven increases in PET and projected increases in P are likely to hang in the balance, with some regional drying, particularly in the central and southern portions of the territory, in both the A2 and B1 emission scenarios (as compared to the 1961-1990 baseline) and some moisture increases.

However, change is expected to be relatively subtle, as compared to the large existing variability in moisture balance across the territory. Model results for change in moisture balance, \(\Delta(P-PET)\) show some degree of continued boreal drying over the next two decades for both the A2 and B1 scenarios, with values for the remainder of the territory positive, but less than one centimeter (Figure 3). The implied increase in moisture in these areas is within the range of model uncertainty, and is likely to be offset by other factors, as described below. The drying effect is greater for the more extreme emissions scenario, A2.

Given the projected balance between the drying effects of increased heat, as expressed via PET, and the wetting effects of increased precipitation, it seems likely that the precise expression of these effects will be driven not only by the balance of P-PET, but also by changes in growing season length; permafrost, active layer depth, and
associated soil drainage; fire and post-fire succession; and ecosystem shifts. Each of these factors triggers complex feedback loops, as discussed below. Examining changes already underway over the last 50 years helps place these multiple effects into context and validate model results. All projections must be viewed as potential future scenarios, taking uncertainty into account, rather than absolute predictions.

**Lengthened growing season**

In much of the Yukon the growing season will get longer as average spring temperatures rise above freezing sooner, and fall temperatures drop below freezing later. Many areas, where PET has been historically limited by sub-zero temperatures (Where $T_{\text{mean}} \leq 0$, $\text{PET}=0$), will transition to become active evaporative environments in the future (Where $T_{\text{mean}}>0$, $\text{PET}>0$). Extension of the growing season is likely to cause a shift in the distribution of total growing season PET and contribute to landscape drying power. Historically, June and July have accounted for much of the total growing season PET, but this percentage is expected to decline as rising rates of PET in spring and autumn account for greater percentages of the growing season total PET. Without any adjustment in monthly P patterns, it is likely that months with the greatest increases in monthly PET will contribute the most to higher net growing season water deficits in the future.
Although high-latitude precipitation has increased slightly in the twentieth century, this has occurred primarily during winter and spring (Serreze et al. 2000; Hinzman 2005), when most animals and plants are within or just emerging from, a period of seasonal hibernation or dormancy. The GCM composite model predicts increases in precipitation across the Yukon, although the uncertainty of precipitation forecasts is generally greater than that of temperature predictions (Walsh et al. 2008), which show significant increases across the historical record (Hinzman et al. 2005), indicating our predictions of P-PET are more likely to be overestimates than underestimates. Uncertainty in accurate precipitation predictions is attributed not only to model shortfalls, but also to the difficulty of measuring precipitation accurately, as well as the temporal and spatial variability of rainfall and snowfall. Any improvement in model input variables is likely to improve future iterations of our analysis.

The effects of climate change on the contribution of winter precipitation to growing season water availability is unclear. Timing of spring thaw and changes in
drainage patterns may prove more important than changes in either total winter precipitation or changes in sublimation -- which are hard to predict, given uncertainty with regard to the effects of increased winter temperatures, humidity changes, and current impacts of sublimation. Janowicz et al (2004) results show that about 21% of winter snow pack is lost to ablation/sublimation process in the Wolf Creek Basin, located just outside of Whitehorse. He also reviewed a number of studies and concluded that “Sublimation seems to not be very significant in all reviewed studies ...” Carey and Woo (2001) also determined that sublimation was a small component of snowelt on all slopes – again in the Wolf Creek Basin. In this study, the authors looked at snowmelt processes from different aspects. There were large differences in snowmelt timing, etc., but they did not find that sublimation accounted for a major percentage of water loss.

It is likely that sublimation will increase slightly in areas that do not experience a vegetation shift, due to higher temperatures. Conversely, it may decrease slightly due to a shifting vegetation regime in some northern regions, where tundra converts to shrub tundra, and in some interior regions, if moving from spruce canopy to deciduous canopy will reduce sublimation of snow pack (Pomeray et al. 2006). However, the retention of winter precipitation for vegetative uptake in the summer is highly dependent on drainage, which in turn is dependent on slope, soil characteristics, presence and depth of permafrost, and active layer thaw rate. As such, the impacts of sublimation are highly site-specific, and are linked to several of the variables discussed below.

The ultimate consequences of these trends may be depletion of landscape water resources earlier each summer, exacerbating mid-summer drought and delaying the time it takes for autumn rains to replenish growing season draw-downs. Arctic and boreal ecosystems are historically conditioned to withstand net water deficits during the growing season (Woo et al. 1992), but increasing the severity of this deficit or changing its timing will likely disrupt the historic structure and function of these systems. Historically drought-prone areas will likely become drier, or increasingly dependent on inflows from other regions to replenish increased deficits accumulated throughout the growing season.

Vegetation change

It is likely that as climate warms and growing season moisture availability becomes more limited, this dynamic will contribute to the already complex interplay between climate and vegetation. While warmer temperatures will tend to drive a shift from tundra to shrub, from shrub to forest, and from coniferous to deciduous forest, moisture limitations may limit or reverse such changes, or trigger alternate trajectories, such as a shift to more grasslands. Exact trajectories are likely to vary by site, depending on slope, aspect, and soil drainage as well as on climate variables.

Scientists have already documented changes in land cover as a result of current warming trends. In some cases, these trends point toward increases in overall biomass, including northern advancement of treeline (Danby and Hik 2007; Lloyd and Fastie, 2003) and shrub expansion in Arctic tundra sites (Sturm et al. 2001). Further extension of the growing season might be expected to increase shrubbiness (Chapin et al. 2000).
Harper et al. (2011) predict an increase in the aggregation of trees in the Canadian forest-tundra ecotone. However, in conditions of drought stress, forest growth may decline (Lloyd and Fastie, 2003; Barber et. al. 2000). D’Arrigo et al. (2004) found a non-linear relationship between warming and forest growth near tree line in the Yukon, with increased growth up to an optimal value, and declines thereafter. They concluded that further warming in the absence of compensating increases in moisture would lead to a browning of areas that have been greening in recent decades.

Forest and shrub expansion would likely have positive feedbacks to climate change, while forest and shrub decline might provide negative feedback. Further advancement of boreal forests into historic tundra sites, or expansion of shrub tundra into areas previously occupied by wetland-sedge tundra would reduce growing season albedo and increase energy absorption, thereby inducing a positive feedback to warming (Chapin et al. 2000; Euskirchen et al. 2007, 2009). Loss of forest and shrub cover due to drought stress might be expected to have the opposite effect. However, either change is likely to be smaller than the changes in albedo due to increases in the length of the snow-free season (Euskirchen et al. 2009; Hinzman et al. 2005).

It should be noted that vegetation changes and shorter snow season have the potential to significantly impact estimates of growing season water availability, since loss of vegetation or change in vegetative cover reduces or alters transpiration. As such, failure to account for temporal changes in land cover variables may have lead to overestimation of the magnitude of drying power. In addition, actual values may be lower than predicted due to potentially higher surface albedo as result of some snowcover still present during the warm season. Failure to account for significant increases in albedo over the growing season (Blanken and Rouse 1994) as well as stomatal control over transpiration as soils dry out (Eugster et al. 2000) may have led to overestimates of PET rates.

Fire

Landscape drying may drive vegetation changes not only directly, but also indirectly, via changes in fire cycles. Fire is a primary driver of successional dynamics in the boreal region, and is predicted to increase in frequency with climate change, leading to an increase in early-succession vegetation on the landscape, as opposed to older coniferous forests (Rupp 2000). Increases in PET, and decreases in P-PET tend to increase fire frequency and severity during the growing season.

Not only does PET impact fire (since drier landscapes are more fire-prone) but the reverse is also true. The impacts of fire on PET change over the course of post-fire succession. Under drought stress, successional trajectories themselves may change. The immediate result of increased area burned is more widespread destruction of existing vegetation communities with further implications for altering albedo and PT alpha inputs that determine estimates of PET. In the short-term, removing vegetative ground cover will reduce plant transpiration and increase soil moisture (Yoshikawa, et al. 2003). However, as vegetation recovers in the years following fire, early-successional herbaceous ground cover is gradually replaced by a mixture of deciduous and coniferous
shrubs and trees, causing a steady increase in summer albedo (Euskerchen et al. 2009) and evapotranspiration rates. Both the albedo value and evapotranspiration rate from these deciduous-dominated stands are generally greater than in new burns or mature coniferous ecosystems that dominate the final successional stage of post-fire recovery (Liu & Randerson, 2008). Without accounting for changes in surface albedo or Prisetley-Taylor alpha values that would result from fire-induced vegetation shift, we may have underestimated future potential evapotranspiration rates. In the tundra, fire has historically played a small role in successional dynamics but Higuera et al. (2008) found that increased shrubbiness was well-correlated with increased fire frequency in tundra ecosystems across the paleorecord. Areas of mature mixedwood forests in the southwestern Yukon that burned in 1958 have shown poor regeneration of spruce (Hogg and Wein 2005) and have shifted to growth of aspen and grassland. Tree-ring analysis indicates that these forests are vulnerable if climate change causes drier conditions in the future.

**Permafrost**

Permafrost presence/absence, depth to permafrost, and the annual depth of the active layer play important roles in determining ecosystem structure and function, including ET rates, throughout much of the Yukon. Loss of permafrost stability will impact soil drainage and surface heat flux, while changes in annual thaw depths are likely to affect not only runoff and drainage but also the timing and depth to which plants can access soil moisture. Permafrost vulnerability to climate change is affected by not only mean annual temperature, but also topography, water, soil, vegetation, and snow. Jorgenson et al (2010) found that surface water, ground water, and snow depth had large effects on permafrost stability, and that vegetation succession provides strong negative feedbacks that make permafrost resilient to even large increases in air temperatures. Permafrost creates a strong heat sink in summer that reduces surface temperature and therefore heat flux to the atmosphere (Yoshikawa et al., 2003; Chambers and Chapin, 2002). Future losses of permafrost could amplify climatic warming and increase ET rates, indicating yet another positive feedback to the drought cycle. An increase in the depth to permafrost could result in greater runoff and precipitation infiltration, resulting in a decline in the surface water balance. Bonnaventure and Lewkowicz (2010) examined permafrost dynamics at three sites (Wolf Creek Basin, Ruby Range, and Haines Summit). They concluded that mountain permafrost in the discontinuous zone is likely to thaw entirely, given long-term temperature increases of 5°C. The tendency of wildfire to accelerate permafrost melting may also promote long-term soil drying (Swanson 1996) by removing the insulating organic layer and allowing a greater percentage of incoming solar radiation to be absorbed by the ground surface. While this may prove significant in areas where discontinuous permafrost is already showing signs of degradation and where fire is common, notably the boreal region, significant thawing of permafrost in the Arctic is not expected to occur this century. There is also suggestion that long-term melting of permafrost could increase water availability by converting water stored in frozen soils to freely available soil moisture. Rouss et al. (1992) found a
feedback loop between dry conditions and depth of active layer in wetland tundra. Dry years promoted deeper thaw depths in permafrost soils during the growing season due to larger ground heat fluxes and larger soil thermal diffusivities. This deeper thawing allowed tundra species to access deeper stores of water, thus at least partially offsetting the effects of moisture stress caused by climate warming. Although no permafrost projections were included as part of this analysis, maps showing projections for mean annual ground temperature and depth of active layer for the Great Bear Lake region and for Alaska give an idea of how conditions may be expected to change in the Yukon (Appendix B).

In conclusion, rising average temperatures are projected to drive increases in annual PET, which in some cases may exceed predicted increases in annual precipitation. This effect alone may make some portions of the Yukon drier during summer months over the course of the next century. However, other factors, including extension of the growing season into spring and fall, increase in growing degree days, vegetation shifts, changed drainage with permafrost thaw, and shorter fire cycles may ultimately play even larger roles in landscape drying and ecosystem change.

Considerable uncertainty is associated with projections for summer moisture balance, due to the limitations of the available model inputs as well as the complexity of feedback loops between temperature, precipitation, snow cover, vegetation, fire, albedo, soil temperature/permafrost, and other soil and groundwater dynamics. Consequences of reduced water availability are expected to be widespread, but vary in magnitude and scope. Declines in surface moisture are likely to initiate a range of secondary effects, and feedback to the surface energy balance and may ultimately cause long-term changes in the structure and function of impacted ecosystems, and in the human uses of these ecosystems.
Sources


Rouse and Steward (1972). a simple model for determining the evaporation from high latitude upland sites. J. Applied Meteorology, 11, 1063-1070.


Evidence of Recent Change in the Northern High-Latitude Environment. Climatic Change. 46(1-2), 159-207.


Appendix A: Historical Precipitation Normals


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<th>Community</th>
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<th>Mar</th>
<th>Apr</th>
<th>May</th>
<th>Jun</th>
<th>Jul</th>
<th>Aug</th>
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<th>Oct</th>
<th>Nov</th>
<th>Dec</th>
<th>Year</th>
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Appendix B: Permafrost Projections for Neighboring Regions

These projections were created by scientists from the Geophysical Institute Permafrost Lab (GIPL) at UAF, in partnership with SNAP.  [http://permafrost.gi.alaska.edu/](http://permafrost.gi.alaska.edu/) (See also Jorgensen et al. 2010, Grosse et al. 2011).

Figure B1 – Projections for active layer thickness (ALT), for the Great Bear Lake region for 2000s, 2030s, and 2090s. An increase of 20-50 cm is expected across the region, although changes are likely to be variable and site-specific.
Figure B2—Projections for mean annual ground temperature (MAGT) for the Great Bear Lake regions, 2000s, 2030s, and 2090s. In some areas of the watershed, mean annual ground temperatures are projected to be above freezing by the 2090s, leading to thawing of shallow permafrost.
Figure B2 – Projections for mean annual soil temperature at 2M depth for the 2000s, 2030s and 2090s for Alaska. Significant areas of thaw are expected in the boreal region.