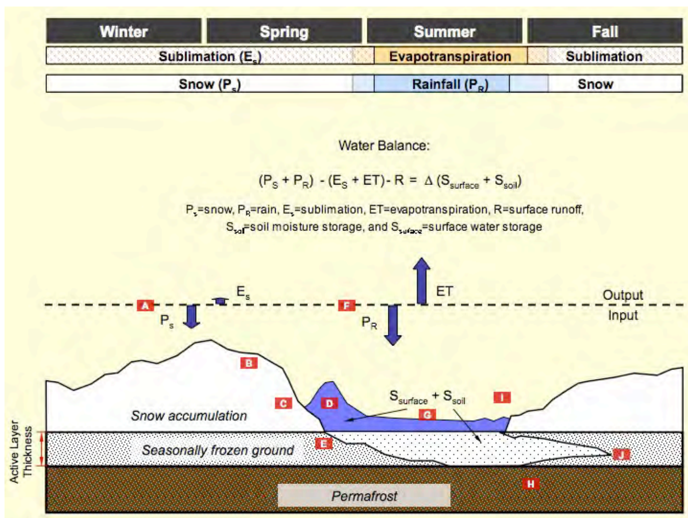
Section 3
Hydrology

Hydrologic processes will play a pivotal role in directing climate-influenced habitat change. This section illustrates the processes that determine overall water balance and seasonal fluxes and features typical of arctic environments. Figure 3.1 is a generalized representation of the current condition (baseline). Specific features of the illustration are labeled and keyed to the accompanying text. Figure 3.2 and accompanying text outlines the changes expected under a scenario of rising temperature. Figure 3.3 and accompanying text outlines changes expected under a scenario of increased temperature and precipitation. Predicted change can be interpreted in the figures by comparing baseline condition (grayed out) with predicted (white for snow, blue for surface storage).



Note: Figures 3.1–3.3 are reproduced in larger size on the following pages.

Figure 3.1. Baseline hydrological conditions.

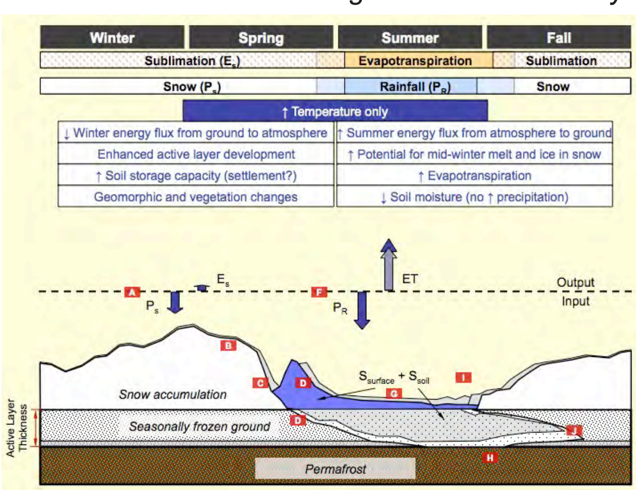


Figure 3.2. Hydrological changes expected under Scenario I (increased temperature).

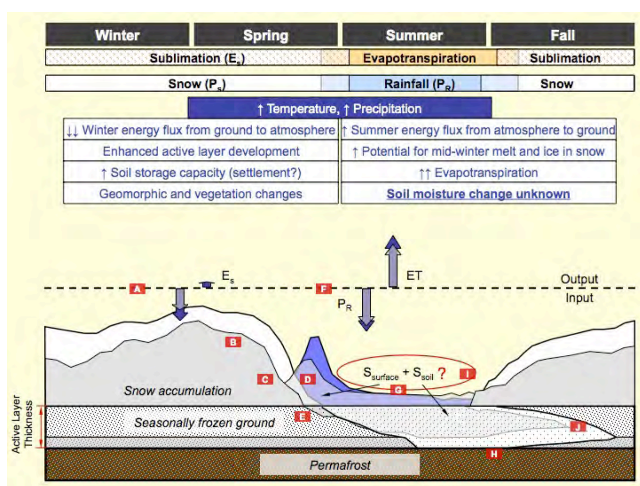


Figure 3.3. Hydrological changes expected under Scenario II (increased temperature and increased precipitation).

Baseline Case (Current Condition)

A) Winter precipitation less sublimation (PS-ES)

During the cold season, precipitation falls as snow and is temporarily stored in the snow pack at the ground surface. During the winter months, the surface energy balance yields a net loss of energy from the ground to the atmosphere, thus lowering soil temperatures. Snow cover, and snow depth in particular, helps to insulate the underlying ground surface from the cold winter temperatures. Therefore, with all else being equal, increasing snow accumulation would lead to higher surface soil temperatures, and decreasing snow accumulation will lead to lower surface soil temperatures. Snow accumulation is reduced to some extent by sublimation (water changing from the solid to gaseous state). Snowfall represents approximately 40% of annual North Slope precipitation.

Occasionally, air temperature will rise sufficiently to cause mid-winter snow melt. The resulting liquid water then moves downward in the snow pack before refreezing as an ice layer.

B) End-of-winter snow water equivalent (SWE)

The winter snow pack represents an effective seasonal surface water storage reservoir. Precipitation falling during this season, minus sublimation losses, remains in storage until the spring freshet when it is released. Consequently, the water stored in the snow pack at the end of winter is an important hydrologic quantity and is referred to as the end-of-winter snow water equivalent (SWE).

C) Spring snow ablation

During spring, the snow pack that developed over the course of the previous winter melts in a short period of time. Snow ablation (depletion via melting and evaporation) typically occurs within a one to two week period. During the melt period, incoming energy from the atmosphere leads to increasing snow temperatures. Once the snow pack reaches the freezing point, it becomes isothermal, and additional energy input is stored as latent heat associated with the water phase change from solid to liquid form. Once ablation is complete and the ground surface is no longer covered with snow, surface soil temperatures begin to increase. In addition, the loss of snow cover leads to a sudden and dramatic decrease of surface albedo that leads to more rapid warming and greater positive net energy flux at the surface.

D) Spring water storage and runoff

Water released from the snow pack during spring melt saturates surface soils, recharges local lakes, ponds, and emergent wetlands, and runs off the surface into streams. Shortly following ablation, streamflow rises sharply and then gradually recedes over the subsequent few weeks. Although much of the snowmelt drains into the Arctic Ocean, a considerable portion is at the surface in the form of lakes, wetlands, and soil moisture. Over the course of a few weeks following peak streamflow, the inundated land surface area, and thus surface water storage, rapidly decreases. This is largely due to lake and wetland drainage down slope until water surface elevations fall below their outlets. Surface water begins to infiltrate into the soil column as soil storage capacity increases with thaw and at the same time, surface and soil-stored water is lost to the atmosphere through evapotranspiration.

E) Soil thaw

Loss of snow cover and increasing energy flux from the atmosphere to the land surface causes thawing of surface soils. As soils thaw they become much more permeable. Consequently, water can more readily enter the soil column and drain vertically and laterally.

F) Summer precipitation less evapotranspiration (PR-ET)

During the warm season (average temperatures above 0 °C), precipitation occurs in the form of rain and condensation. The energy flux from the atmosphere to the ground during late spring and early summer is at its maximum and represents a net gain of energy for the ground surface. Unlike winter sublimation, the loss of surface moisture to the atmosphere is very close to precipitation inputs.

G) Summer water storage

During the summer months, rainfall and condensation tend to be equal or less than evapotranspiration. Therefore, there is a gradual drying of wetlands and soils that received snow melt recharge in the spring.

H) Active layer thickness (ALT)

The top portion of the soil column that thaws during the summer season is referred to as the active layer. The maximum depth to which the ground thaws is called the active layer thickness (ALT), and the lower extent of the active layer is defined by the permafrost table. The active layer plays a paramount role in hydrologic response as depth of thaw is directly related to the maximum soil water storage capacity.

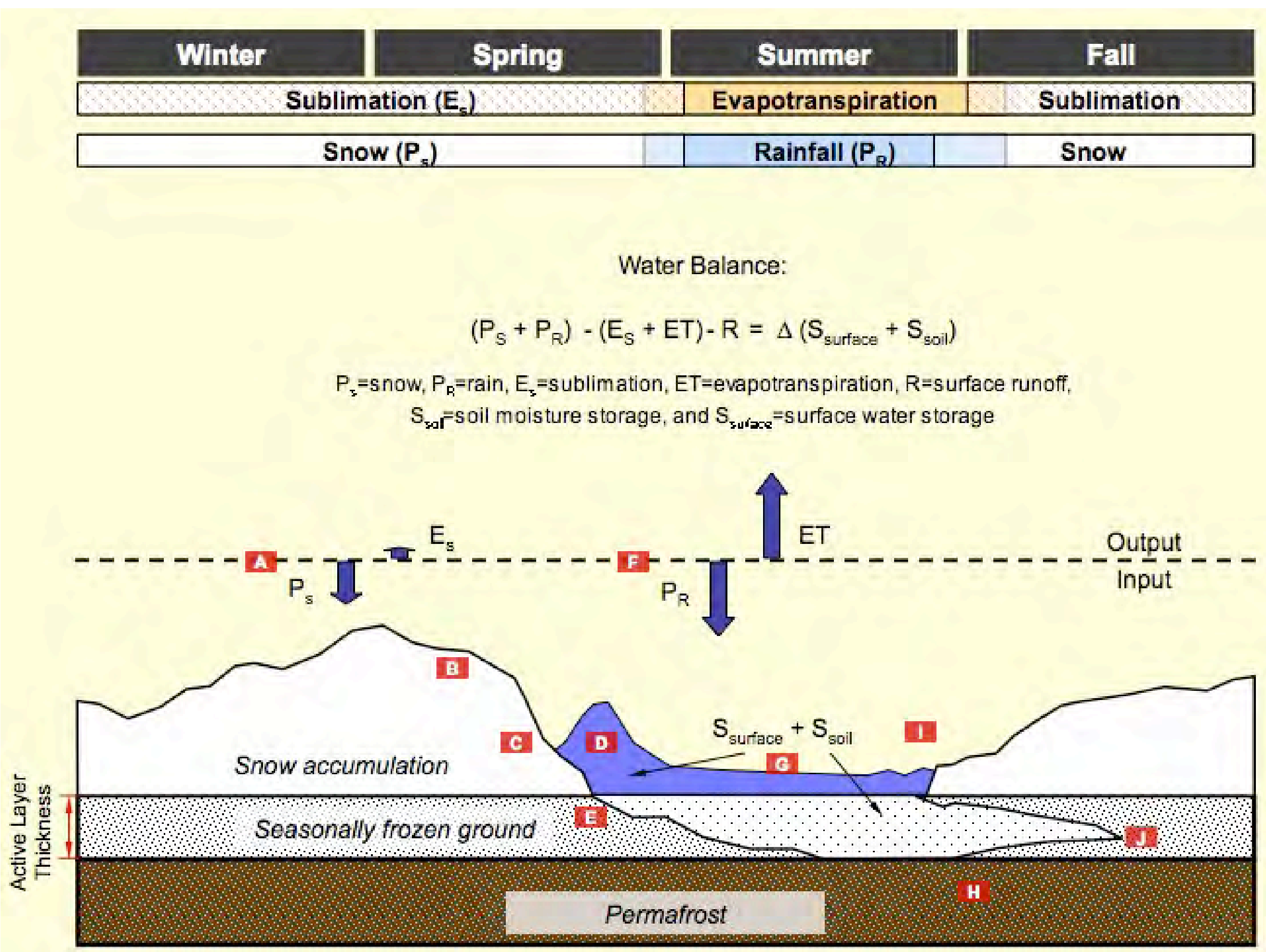
I) End-of-summer water storage

Late summer water storage is a function of total available basin storage (surface + soil), the sign and magnitude of PR-ET, and topographic gradients. The coastal plain is dominated by surface storage effects, while the foothills are more influenced by topographic gradients. As depicted in feature (G), surface water storage gradually decreases throughout the summer months. Although there is some summer recharge from storms, it tends to be minor, and end-of-summer surface storage is considerably below the maximum storage capacity. During the following spring, snowmelt will again recharge the surface storage.

J) Soil freeze-back

In early fall, surface soils begin to cool and eventually reach the freezing point. Between mid and late fall, the active layer undergoes freeze-back, primarily at the surface but also at the interface with the permafrost table.

Figure 3.1. Baseline hydrological conditions.



Future Scenario I: Increased Temperature Only

A)   Winter precipitation less sublimation (PS-ES)

Assuming no change in precipitation, there would be little change in snow minus sublimation. Changes in sublimation will likely be small.

Increasing temperatures, particularly in late fall and early winter may lead to increased frequency of mid-winter snow melt, rain on snow events, and ice formation.

B)   End-of-winter snow water equivalent (SWE)

As with feature (A), the change in sublimation will be small, and end-of-winter SWE will change little.

C)   Spring snow ablation

Increasing winter and spring temperatures will lead to earlier and more rapid snow ablation. However, changes in cloudiness may either enhance snow ablation (e.g., reduced cloudiness results in higher short wave radiation at the surface) or retard ablation (e.g., increased cloudiness reduces short wave radiation at the surface).

D)   Spring water storage and runoff

Earlier snow melt will lead to earlier surface storage recharge and peak streamflow. If it is assumed that snow melt rate changes are only minor, then peak flows and inundated areas can be expected to be similar during the initial runoff response.

E)   Soil thaw

Surface soils begin to thaw following the end of snow cover. Earlier loss of snow cover will, therefore, lead to earlier onset of active layer thawing.

F)   Summer precipitation less evapotranspiration (PR-ET)

Increasing air temperature will lead to increases in potential evapotranspiration (PET), which is a function of the energy available to drive evapotranspiration. PET is the rate of evapotranspiration that may be expected under saturated surface conditions. Actual evapotranspiration will increase as long as surface moisture is available. Because the baseline scenario (current condition) indicates that summer evapotranspiration already exceeds rainfall and condensation, a warming scenario that does not include increased rainfall suggests increasingly drier conditions and more rapid depletion of surface and soil water storage.

G)   Summer water storage

As stated above [feature (F)], the rate of moisture loss to the atmosphere will increase under a warming scenario. Consequently, the rate at which surface and soil water storage decreases throughout the summer is likely to increase. This is represented on the schematic as a steeper decline of the surface and soil water storage throughout the summer months.

H)   Active layer thickness (ALT)

Warmer winter temperatures will reduce the energy flux from soils to the atmosphere, and winter soil temperatures will not be as cold as in the baseline case. Increasing temperatures during the warm season will lead to greater energy flux to the soil from the atmosphere. Warmer soil temperatures and earlier onset of active layer thaw will enhance the potential depth of seasonal thaw. Other things being equal, increased depth of thaw will lead to greater soil water storage capacity and the conversion of surface ponding into subsurface storage. It is important to note, however, that this effect may be offset by ground settlement that tends to reduce ALT.

Increasing soil storage capacity and more rapid moisture export through evapotranspiration will lead to decreased hydrologic response to summer storms. Currently, the coastal plain exhibits very little or no stream response to summer rainfall events. This is due to the high surface storage capacity, low topographic gradient, and increasing storage deficits during summer. As rainfall arrives at the surface, it replenishes storage deficits and does not result in short-term streamflow increases. In contrast, significant rainfall events do result in streamflow response in the Arctic Foothills ecoregion. As the active layer and surface storage deficits increase, the foothills will have muted hydrologic response to storms but greater base flow due to suprapermafrost groundwater flow from soil moisture.

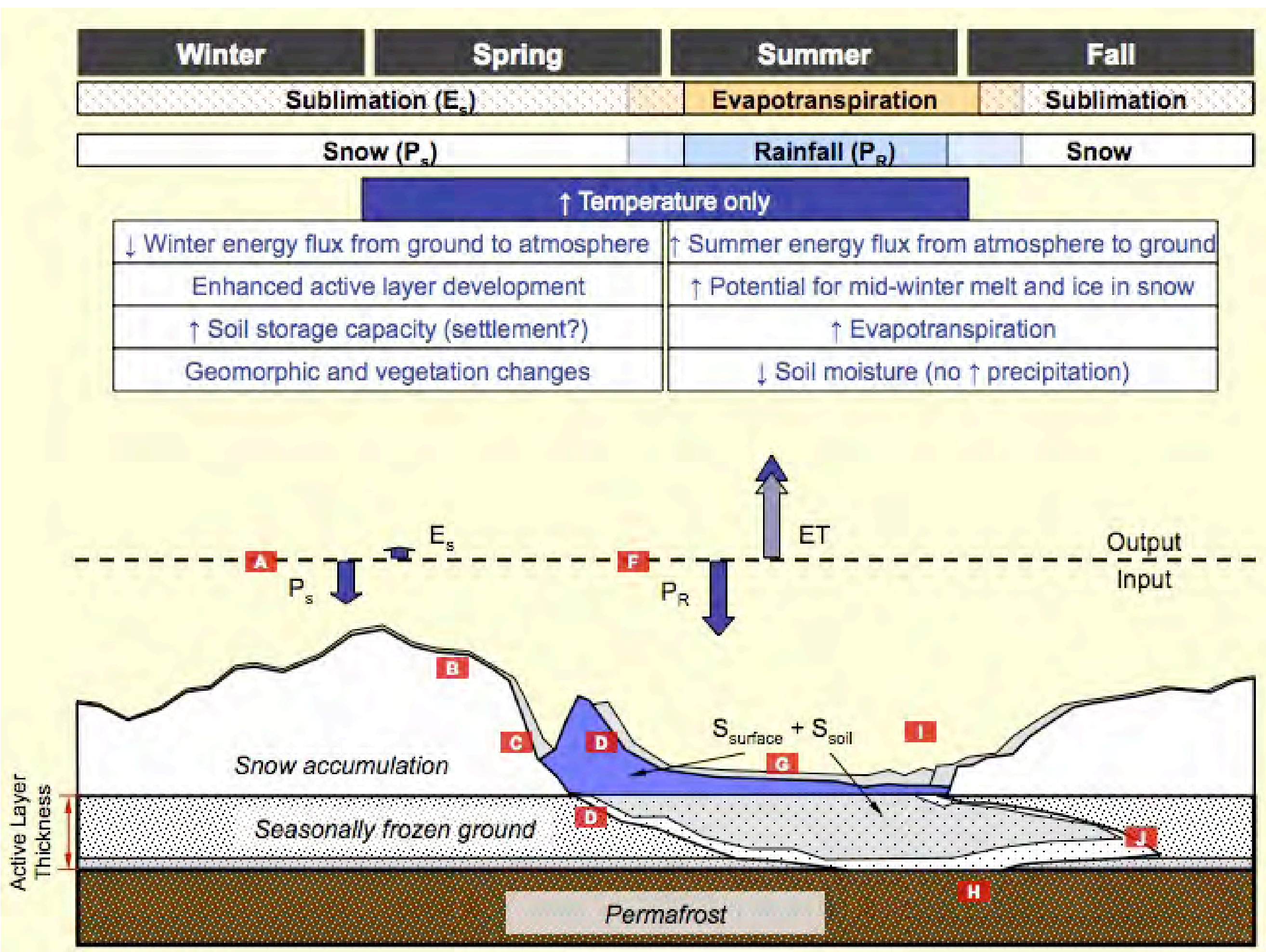
I) End-of-summer water storage

Although initial surface inundation and peak streamflows may be similar to current conditions, increasing evapotranspiration will lead to more rapid surface and soil water depletion. Furthermore, increasing ALT may convert more surface water storage into subsurface storage (soil moisture). Larger fall storage deficits will require a larger proportion of spring snow melt to recharge storage reservoirs. This will reduce peak streamflow the following spring to some extent.

J) Soil freeze-back

Warmer soil temperatures and a prolonged fall season will lead to later and more gradual soil freezing. In extreme scenarios this may even lead to talik formation if the active layer cannot fully freeze-back. In the foothills, prolonged thaw may lead to increased soil water drainage to streams. On the coastal plain, topographic gradients may be insufficient for soil water to drain.

Figure 3.2. Hydrological changes expected under Scenario I (increased temperature).



Future Scenario II: Increased Temperature and Increased Precipitation (Across All Seasons)

A) Winter precipitation less sublimation (PS-ES)

Increases in winter precipitation will dominate the snow storage, while changes to sublimation are likely to be minor. This scenario assumes a general increase in annual precipitation. However, most global climate models project that increasing arctic precipitation will be more pronounced in the winter.

B) End-of-winter snow water equivalent (SWE)

Increasing snowfall will result in higher end-of-winter SWE. As a result, there will be a greater volume of water released during the spring melt.

C) Spring snow ablation

Although higher temperatures favor earlier onset of snowmelt, the advance of melt is limited by the availability of radiation to drive the process. Furthermore, increased end-of-winter SWE will require additional energy to melt the snow pack. Consequently, the timing of melt and total snow ablation may not be significantly earlier than the current condition.

D) Spring water storage and runoff

The volume of water released from the snow pack will increase with increasing winter precipitation. If end-of-summer storage deficits are unchanged, the result will be increased surface storage and increased peak streamflow. If end-of-summer storage deficits are higher (i.e., drier conditions in fall), then a larger portion of SWE will be necessary to fill surface and soil moisture deficits.

Following the initial hydrologic response (e.g., peak flows and maximum inundated area), surface storage will still decrease rapidly according to the water surface elevation with respect to the elevation of drainage outlets. The rate of evaporative losses from water surfaces will be greater due to increased temperatures. However, total loss from evaporation and transpiration will depend on 1) total inundated area, 2) near-surface soil moisture, and 3) the timing of summer rainfall with respect to seasonally changing radiation and plant senescence.

E) Soil thaw

Soils will begin to thaw following snow cover loss. Although the start of thaw may not be different from the baseline case, the rate of thaw may be greater due to 1) warmer end-of-winter soil temperatures, 2) increased summer temperatures, and 3) increased heat transfer associated with summer rainfall.

F) Summer precipitation less evapotranspiration (PR-ET)

Increasing summer precipitation along with increasing summer temperatures will result in increased evapotranspiration. However, evapotranspiration is limited by soil and surface moisture. Soil moisture under changing climate is highly uncertain and the subject of many ongoing research efforts. It is not clear if increasing summer precipitation will be sufficient to outpace evapotranspiration.

G) Summer water storage

Increasing summer rainfall may reduce the rate of summer surface drying. However, it is unknown if rainfall will increase enough to outpace increasing evapotranspiration. Total summer storage is comprised of both surface water storage (e.g., lakes, ponds, and emergent wetlands) and soil moisture in the active layer.

H) Active layer thickness (ALT)

More rapid active layer thaw and a prolonged warm season will increase the potential soil thaw depth. As the active layer thaws more rapidly and to greater depth, the distribution of storage between surface water and soil moisture may shift downward. As a result, some surface water may become soil moisture and result in less inundated area, shallow soil moisture may percolate deeper in the soil column, and uplands may become drier despite overall higher soil water storage at the landscape scale.

The increased soil storage capacity associated with greater ALT serves to attenuate streamflow response to summer storms. As rainfall arrives at the surface, a greater volume can enter the soil column before saturating the soil and generating direct runoff. Soil water is then gradually released to the drainage network, resulting in lower peak flows and longer streamflow recession.

Although increased heat flux to the ground will cause a deepening of the active layer, a number of potentially opposing processes can also occur. The load-bearing capacity of frozen soils is greater than thawed soils. As a result, increasing active layer thickness may result in soil compaction that effectively reduces the maximum depth of thawed soil. Vegetation plays a strong role in controlling terrestrial-atmospheric energy and moisture fluxes. As temperature and active layer thickness increase, vegetation may respond in such a way as to insulate the ground and create a new stable state for the active layer. These feedbacks illustrate just how difficult it is to predict arctic hydrologic response to changing climate.

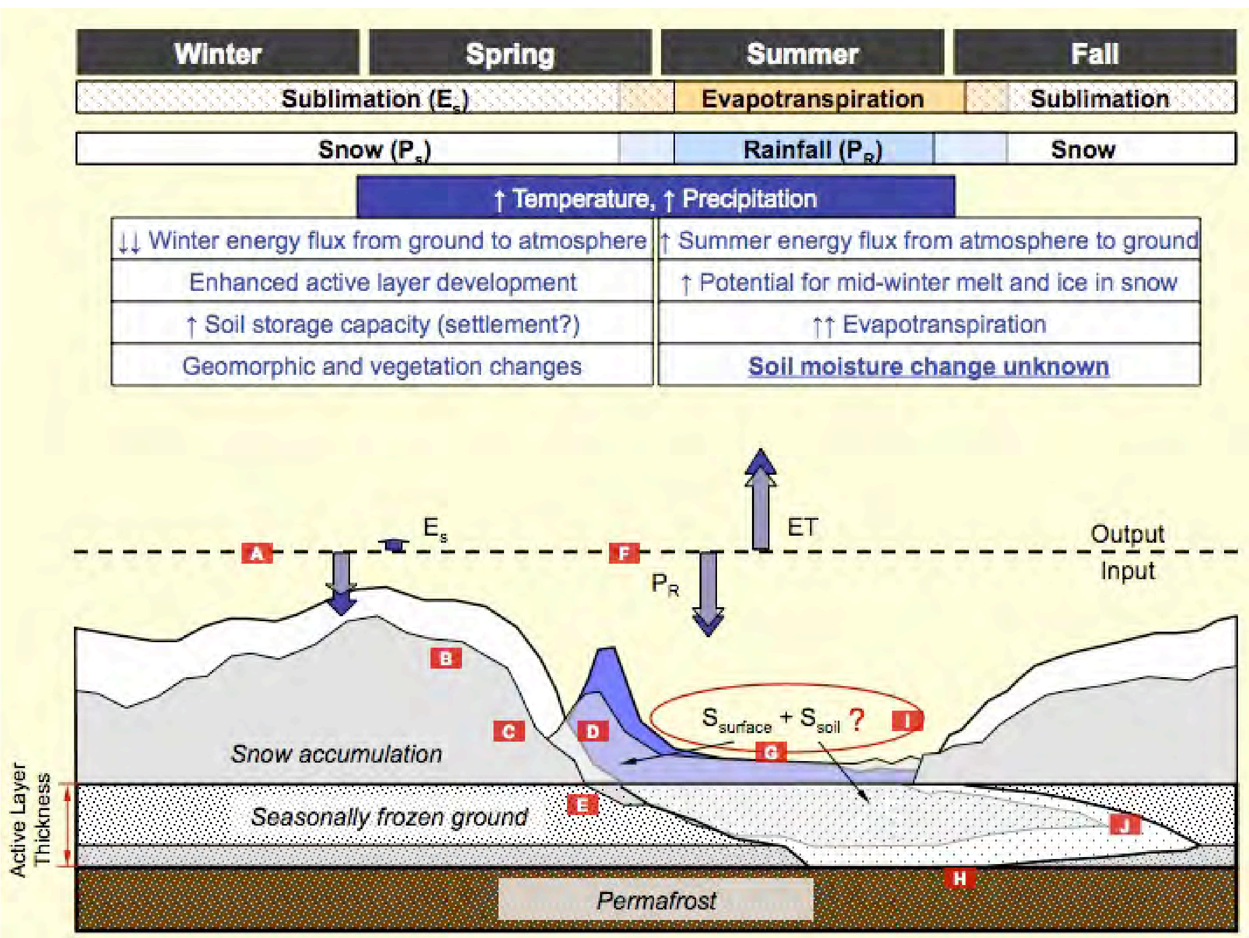
I) End-of-summer water storage

As the warm season is prolonged due to arctic warming, so too is the period for soil moisture to move down gradient and into drainage networks. This may lead to late summer and early fall soil drying, while stream baseflows increase due to additional shallow groundwater flow. If rainfall in early fall increases, much of the depleted storage may be recharged prior to winter freeze-back.

J) Soil freeze-back

As with Scenario I (temperature only), soil re-freeze will be delayed due to higher temperatures. In this scenario, however, the active layer thickness is greater, so the time to total freeze-back will take longer. In the extreme case, the active layer may become too thick to completely re-freeze in winter, and a talik will form. In the foot-hills, such an occurrence could mean continued soil drainage throughout winter and a larger spring soil moisture deficit. This winter drainage may lead to higher under-ice streamflow. The effect may be less important on the coastal plain where soil moisture may not have sufficient gradients along which to drain. Consequently, overall soil moisture distribution on the coastal plain will be largely determined by microtopography.

Figure 3.3. Hydrological changes expected under Scenario II (increased temperature and increased precipitation).



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