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Source: Arctic, Antarctic, and Alpine Research, 35(1) : 74-81

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1523-0430(2003)035[0074:FSOTAA]2.0.CO;2

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Arctic, Antarctic, and Alpine Research, Vol. 35, No. 1, 2003, pp. 74–81

Five Stages of the Alaskan Arctic Cold Season with Ecosystem Implications

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Abstract

We divide the Alaskan Arctic cold season into five stages based on transitions in climatological and thermophysical conditions in the atmosphere, snowpack, and soil active layer. Each of these stages has distinct characteristics which drive ecosystem processes. During the two autumnal stages (*Early Snow* and *Early Cold*) soils remain warm, unfrozen water is present, and the highest rates of cold-season soil respiration occur. The next two stages (*Deep Cold* and *Late Cold*) are characterized by a frozen active layer with decreasing temperature. *Thaw* is critical in determining the length of the growing season and the resumption of biological processes. *Deep Cold* and *Late Cold* result from a radiation deficit, show little interannual variation, and will be resistant to change under almost any reasonable climate change scenario. These are also the stages with the least amount of biological activity and have the least impact on the ecosystem. However, *Early Snow*, *Early Cold* and *Thaw* stages vary significantly from year to year, have more ecosystem implications, and are also the most likely to undergo significant change in timing and character as the arctic climate changes. This 5-fold subdivision is useful for framing discussions of biophysical interactions during the arctic winter and for focusing attention on critical cold-season periods.

Introduction

Arctic researchers increasingly view the cold season as the time of year when new biological discoveries are likely to be made. Older work (Pruitt, 1957; Formozov, 1946), detailing the complex interactions and adaptations of life to the cold season regime, has received new attention. Recent work (Jones et al., 1999; McFadden et al., 2001; Sturm et al., 2001a; Liston et al., 2002) has brought some of the important issues of climate change and arctic tundra ecosystems during the winter into focus. Although known for decades (Kelley et al., 1968), recent findings (Zimov et al., 1993, 1996; Oechel et al., 1997; Fahnestock et al., 1999) suggest that the winter CO_2 efflux due to heterotrophic soil respiration under the snow contributes significantly to annual CO_2 flux in tundra ecosystems and may have global significance (Zimov et al., 1996). In addition, litter-bag experiments at a tussock tundra site showed that nearly all litter mass and N loss occurred during winter (Hobbie and Chapin, 1996). In short, cold-season biological processes are important both in their own right and in the way they condition the ecosystem for the summer growing season.

To those without firsthand experience, the arctic cold season may seem long, dark, unchanging, and biologically ''dead,'' but in fact it has several distinct stages. Identifying these physically based stages is the main goal of this paper. The purpose of presenting this framework is to facilitate thinking about cold-season ecosystem activity and help to focus research on those stages of the cold season most likely to change in ways that impact the ecosystem. The five stages presented here are each defined by key transitions in surface atmospheric conditions, radiation, snowpack, soil thermal state, and soil unfrozen water content, and so represent as ''natural'' a subdivision as possible.

Background

Previous efforts to divide the arctic winter into stages have been based on subnivean temperature, small mammal activity, surface albedo, surface energy balance, and air temperature. Pruitt (1957) defined the ''snow period'' (essentially our cold season, defined below) as beginning when the snow cover reached a depth of 15 to 20 cm. He called this the ''hiemal threshold''—when the temperature at the ground surface became decoupled from the air temperatures, permitting small-mammal activity beneath the snow. More recent work (Taras et al., 2002) suggests this decoupling occurs at a greater depth (~80 cm) but confirms that decoupling is an important biophysical transition.

On the basis of soil temperatures observed near Barrow, Alaska, Kelley and Weaver (1969) subdivided the cold season into three stages: tundra active layer freezes, tundra soil continues to cool, and soil warms and snow melts. Similarly, Maykut and Church (1973) divided the arctic cold season near Barrow, Alaska, into albedo regimes. In the ''Autumnal transitional period,'' water and soil became frozen and snow covered. Next came a ''Winter stationary period'' continuing into late May, followed by a brief ''Spring transitional period'' when the albedo decreased as the snow-free tundra fraction changed. The three-stage systems defined by Kelley and Weaver (1969) and Maykut and Church (1973) can be mapped onto the five-stage system we define below, but while these authors found little to differentiate the bulk of the cold season when snow cover was effectively 100%, we find justification for dividing this period of the winter into several stages.

A somewhat more complex cold-season subdivision for the coastal Arctic, described by Weller and Holmgren (1974) and Dingman et al. (1980), was based on the surface energy budget. In this system, the ''Freeze-up period'' during the month of Sep-

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tember was defined as the period when the surface gained net radiation while losing heat through melting of snow. The bulk of the cold season, ''Winter period'' (October through May) was characterized by net radiation losses from the surface, largely offset by sensible heating from the atmosphere. The full melt period was subdivided into ''Pre-melt,'' when the soil and snow were heated to 0°C, followed by the ''Melt,'' during which the snowmelt actually occurred.

In a previous paper (Olsson et al., 2002) we used a 5-yr record of air temperatures from a transect from the Arctic Coast to the foothills of the Brooks Range to subdivide the cold season into three 3-mo periods: ''Early Cold'' (September–November), ''Deep Cold'' (December–Feburary), and ''Late Cold'' (March–May). The five-stage subdivision we present here builds on that work but uses naturally derived rather than calendar transition periods. These new subdivisions, *Early Snow, Early Cold, Deep Cold, Late Cold,* and *Thaw,* are based on newer, more comprehensive data that include time series of air, snow, and soil temperatures. In the new system the transitional dates between stages can vary from year to year and with location, more faithfully reflecting natural conditions, but the existence of substantially different stages during the winter is still highlighted.

Data

Observations and computation of various weather, snow, radiation, and soil conditions form the basis for our definition of the five stages. The data we present here (Figs. 1, 2) come from Imnavait Creek (68°37′N, 149°18′W) (Olsson et al., 2002) at the headwaters of the Kuparuk River basin, and from Ivotuk (68°28′N, 155°44′W) (Hinzman, 2002) in the central part of the Arctic Slope. We chose the Imnavait site because it provides our most comprehensive record; Ivotuk was chosen because it is about 250 km west of Imnavait Creek and 320 km due south of Barrow, Alaska. While the foothills location of Imnavait Creek tends to be warmer than the arctic plain during parts of the cold season, and experiences an earlier melt season (Kane et al., 1997; Olsson et al., 2002), it is still widely representative of much of the region. Time series data (temperature, solar radiation, etc.) have been combined with spot measurements (snow depth, snow type and stratigraphy, character, thaw depth) and computed quantities (radiation components, melt energy) to produce a typical evolution of the air, snow, and ground system. Together these define the physical and environmental conditions that characterize the arctic Alaskan cold season. Unfrozen water content was not measured at Imnavait, but we have adapted data obtained at Franklin Bluffs, Alaska, on the arctic coastal plain for the 1999–2000 season as an estimate of unfrozen water at Imnavait Creek in 1994–1995. We have used soil temperatures at both sites to adjust this water content time series.

The Start and End of the Cold Season

The term *winter,* as defined by the calendar (21 December through 21 March), is poorly suited to the Arctic, where winter-like conditions exist from September through May. For biophysical applications, it is more useful to speak of a cold season that begins when surface water begins to freeze (freeze-up) and precipitation arrives as snow and ends when most of the snow melts. The cold-season termination is known locally in Alaska as ''breakup.'' In arctic Alaska, the onset of freeze-up varies interannually (Romanovsky and Osterkamp, 1995; Osterkamp and Romanovsky, 1997) depending on regional weather patterns. Though the start of the cold season coincides closely with when average daily air temperature drops below freezing, the end is 2 to 3 wk *later* than the transition of average daily temperature from below to above freezing because of the extra time needed to melt the winter snowpack.

While freeze-up and breakup constitute the two most important thermal transitions of the year (with profound ramifications for plants, animals, humans, human transportation, and a myriad of other aspects of the arctic ecosystem), the timing and quantity of snowfall are also key drivers. The developing snowpack modulates surface and subsurface temperatures (Goodrich, 1982; Kane et al., 1997). As the timing of the end of the season is controlled to some extent by the total winter snowfall, which varies greatly, the length of the cold season differs substantially from year to year. The complex and nonlinear interactions among snow depth and timing, cloud cover and insolation can combine to produce extremes of ±20 d or more in cold-season duration. However, despite such various outcomes, the season still exhibits all five stages (defined below) each year. In Figure 1 we delineate the stages using the Imnavait data. In Table 1 we present the salient characteristics of the stages and their transitions.

Defining the Five Cold-Season Stages

(1) EARLY SNOW *AND (2)* EARLY COLD—*PHYSICAL CHARACTERISTICS*

Snowfall can occur in any month of the year in the Arctic but does not remain on the ground until the near-surface soil temperature approaches freezing, which is early September under current climate conditions. The first occurrence of freezing daily mean temperature after 1 September marks the onset of the first stage, *Early Snow.* This is often fairly concurrent with the first significant snowfall of the cold season. The timing and nature of *Early Snow* show considerable interannual variability depending on global and regional atmospheric circulation patterns. *Early Snow* is often characterized by storms of mixed rain and snow. Because the atmosphere is still relatively warm at this time, it can support greater precipitable water values that support more precipitation than later in the season. The latter half of *Early Snow* is also characterized by stable near-freezing soil and subsurface soil temperatures. This is due to the large store of latent heat and, to a lesser extent, sensible heat in the soil that prevents further cooling. This ''zero degree curtain'' (Sumgin et al., 1940; Outcalt et al., 1990) is enhanced if early August is rainy (a frequent occurrence) and produces saturated soils even before *Early Snow* begins.

Maximum daily temperatures are typically just above freezing during *Early Snow,* but diurnal temperature cycling, which is typically about 6°C in noncoastal locations (Olsson et al., 2002), combined with dwindling insolation, is usually sufficient to preserve the initial snow cover. The resulting *Early Snow* snowpack is practically an ephemeral one (Sturm et al., 1995), with icy layers and crusts produced by alternate melting and freezing.

The onset of the second stage, *Early Cold,* occurs when air temperatures remain consistently below freezing throughout the day and the net radiation budget becomes negative. Along with lower temperatures, snow continues to accumulate as the region experiences frequent periods of precipitation. Indeed, it is often during *Early Cold* that most of the winter snow accumulates (Fig. 1c). Periods of precipitation are interspersed with clear periods when the temperature can drop below −35°C.

Concurrently with the snow cover buildup, the zone of 70%

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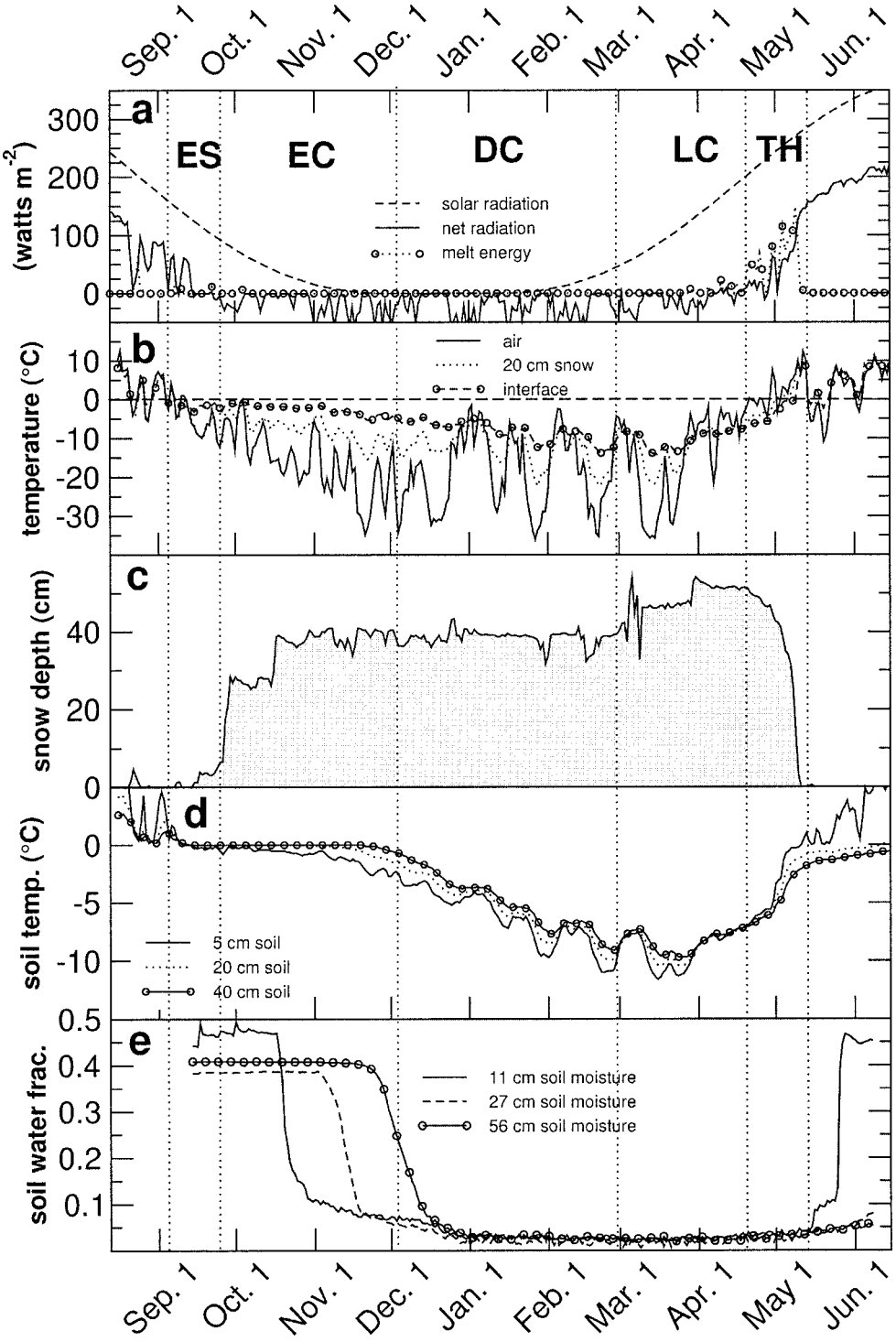


FIGURE 1. Data measured at Imnavait site for the 1994–1995 cold season except where noted. (a) Computed incoming solar radiation, net (longwave and shortwave) radiation, and melt energy; positive values represent an energy gain for the surface. (b) Temperatures of the air (1 m), ground/snow interface, and 20 cm snow temperature. (c) Snow depth measured with sonic sounder. (d) Measured soil temperatures at 5 cm, 20 cm, and 40 cm. (e) Soil moisture as total fraction of liquid water by volume. These moisture data were adapted from measurements at Franklin Bluffs, Alaska.

or greater sea-ice coverage moves incrementally southward, with the Beaufort and Chukchi Seas freezing by the first half of November and the northern Bering Sea freezing by the end of the period (Labelle, 1983). This eliminates proximate sources of water vapor as the sea ice expands and moisture to produce precipitation must be transported from increasingly greater distances and lower latitudes. Given this evolution, it is not surprising that the bulk of the cold-season precipitation (typically greater than 60% by our estimation) has occurred by the end of *Early Cold.*

Early Cold is characterized by a number of important changes taking place within the building snow cover and below surface in the soil. Unlike *Early Snow*, when the snowpack is nearly isothermal, low air temperatures occurring during cold outbreaks produce temperature gradients through the snowpack of sufficient magnitude to cause kinetic crystal growth (Colbeck, 1986) and the development of depth hoar (Trabant and Benson, 1972; Sturm, 1991; Benson and Sturm, 1993), a process that continues until breakup. This metamorphic process increases the snowpack's thermal insulation value, slowing the rate at which the soil cools and at the same time decreasing its conductivity (Sturm et al., 1997).

The termination of *Early Cold* is marked by the complete freezing of the moisture in the active layer (except thin films), in the same way that the start of the cold season is defined by

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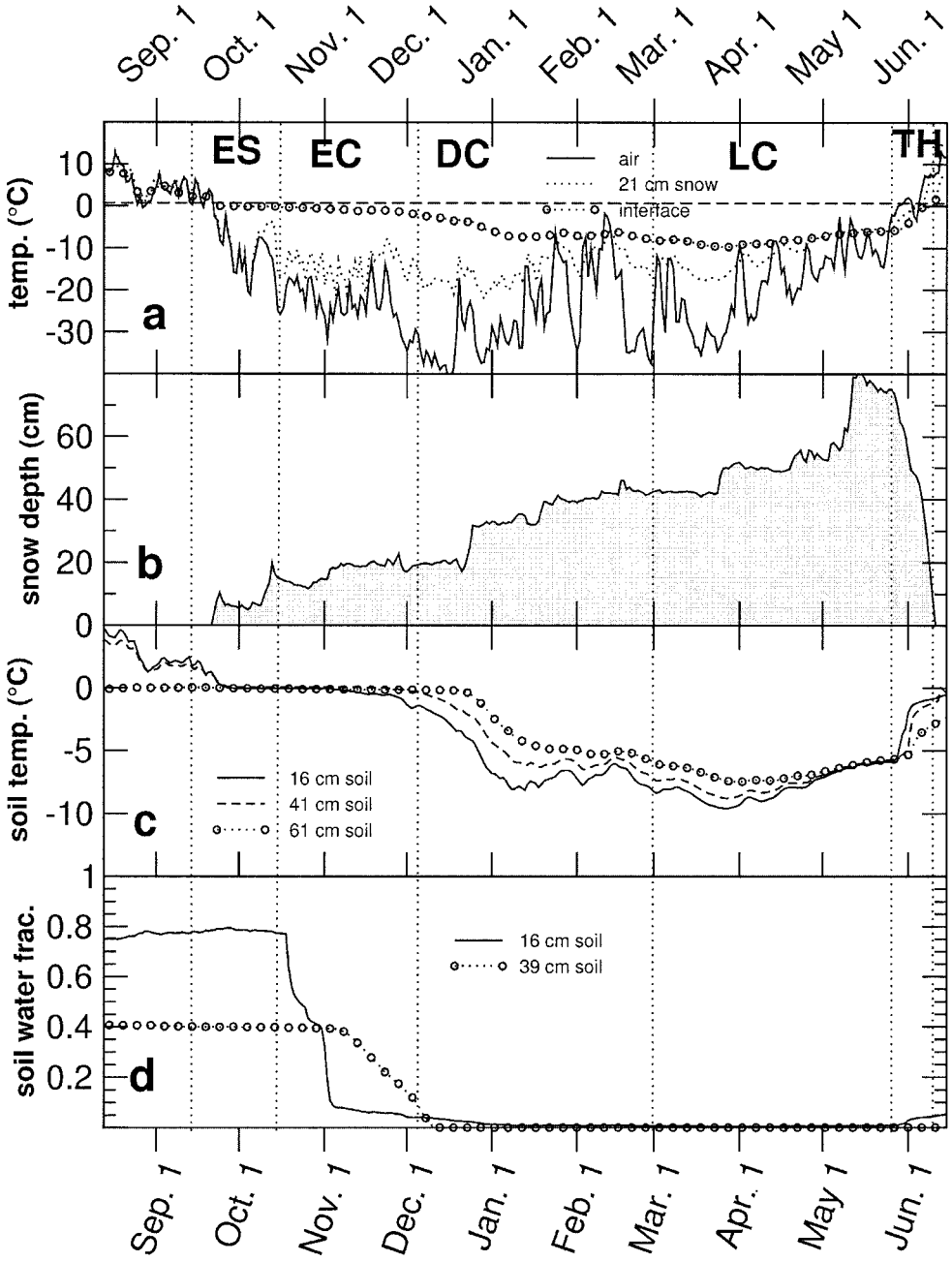


FIGURE 2. Data measured at Ivotuk for the 1999–2000 cold season. (a) Air temperatures (1 m), ground/snow interface temperature, and 20 cm snow temperature. (b) Snow depth measured with sonic sounder. (c) Measured soil temperatures at 16 cm, 41 cm, and 61 cm. (d) Soil moisture as total fraction of liquid water by volume.

the freezing of surface precipitation. The freezing takes place sequentially through the active layer of the soil, with deeper layers freezing only after an appreciable time delay as the sensible and latent heat content of the soil are expended (Osterkamp and Romanovsky, 1997; Romanovsky and Osterkamp, 2000). This freezing occurs both from above (about two-thirds of the depth) and from the bottom (about one-third of the depth) of the active layer. As the bulk of the soil moisture becomes frozen with the onset of *Early Cold*, soil temperatures begin to decline more rapidly. We fix the end of *Early Cold* when the deepest of the soil levels near the bottom of the shrub root zone (~40 cm), begins to show a drop in temperature indicating that all overlying layers of soil have now frozen. Also during *Early Cold* the calculated cloud-free net radiation at the snow surface begins to exhibit negative (loss) values (Fig. 1a), a tendency in agreement with Weller and Holmgren (1974) and Dingman et al. (1980).

For the parts of arctic Alaska for which we have good climate records, it is near the end of *Early Cold* that a brief but substantial atmospheric warming often occurs. This event, the result of strong warm-air incursions from much lower latitudes (Olsson, 2001), is often accompanied by strong winds and results in a dense wind slab capped by a melt crust. This capping event produces a durable stratum overlaying the softer, fluffier snow beneath, protecting it from wind transport during subsequent wind events.

(1) EARLY SNOW *AND (2)* EARLY COLD—*ECOSYSTEM INTERACTIONS*

The evolving early cold season has significant implications for the plants, animals, and soil microbes of the Arctic. By the end of *Early Snow* most arctic migrating bird species have left and ground squirrels have entered hibernation. Large arctic animals such as wolves, foxes, and caribou remain active, but their mobility becomes restricted by the developing snow cover that also protects subnivien microtines from predation (Pruitt, 1957).

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TABLE 1

Defining characteristics of the five stages

Name	Number	Sensitivity to Change	Salient Features
Early Snow (*Early Snow*)	1	High (timing)	above freezing max. daily temperatures, ephemeral snowpack, near 0°C soil temperatures
Early Snow–Early Cold transition			*air temperatures consistently below 0°C, snow accumulates, net radiation deficit, soil begins to freeze*
Early Cold (*Early Cold*)	2	High (timing and nature)	rapid snow buildup, onset of snow-pack metamorphosis, soil freezes
Early Cold–Deep Cold transition			*freeze-up of soil moisture in root zone, decreasing soil temperatures*
Deep Cold (*Deep Cold*)	3	Resistant	little snow accumulation, little to no solar radiation, snow metamorphosis continues, active layer temperatures fall
Deep Cold–Late Cold transition			*returning solar radiation, complete metamorphosis of subsurface snowpack, snow again accumulates*
Late Cold (*Late Cold*)	4	Resistant	active layer temperatures in phase with air temperature, air temperature begins increasing, active layer temperatures begin to rise
Late Cold–Thaw transition			*daily temperatures above freezing, snowpack starts decreasing*
Thaw (TW)	5	High (timing)	min. daily temperatures above freezing, snowpack melts, rapidly increasing soil temperatures

Arctic plants become dormant by the beginning of *Early Snow* and the snow cover development reduces exposure of most low-lying tundra plants to wind. Typically an average snow depth of 20 to 40 cm is emplaced by the end of *Early Cold* covering most tundra plants such that the stems of only taller birch and willow shrubs are exposed. Though photosynthesis of evergreen plants and mosses ceases, roots may remain active (Bilbrough et al., 2000) until the end of *Early Cold,* when temperatures throughout the active layer fall below freezing.

Given sufficient insulating snow cover (i.e., early and rapid onset of *Early Snow*), soil respiration continues during *Early Snow* and *Early Cold* (Grogan and Chapin, 1999; Jones et al., 1999) at rates that are the highest of the cold season (Oechel et al., 1997). The freeze-thaw cycles that occur in the soil active layer in *Early Snow* may also stimulate occasional pulses of soil respiration (Schimel and Clein, 1996). Jones et al. (1999) found that CO_2 efflux at a broad range of arctic Alaska sites was three to four times higher in November than April. This process is dependent on substantial snow accumulations that help keep the soils unfrozen. In those years with minimal snow accumulation in *Early Snow* and *Early Cold,* CO_2 efflux may be low during those periods because of the associated lower soil temperatures.

(3) DEEP COLD *AND (4)* LATE COLD—*PHYSICAL CHARACTERISTICS*

As the active layer freezes competely, *Early Cold* transitions into *Deep Cold* with *Late Cold* following. These two stages, the coldest of the annual cycle, experience frequent polar cold outbreaks, when frigid air from the polar region surges southward over northern Alaska, often producing temperatures as low as −40°C. During *Deep Cold,* insolation is minimal or nonexistent (Fig. 1a) and soil temperatures continue a general downward trend (Fig. 1d), although they respond in muted fashion to the periodic cycles of atmospheric warming and cooling.

The nature of *Deep Cold* and *Late Cold* is determined largely by the minimal insolation in the arctic night—the biggest single term in the annual radiation budget. The annual progression of insolation in the Arctic results from the tilted axis of the Earth in its plane of revolution rather than a result of climate factors. Regardless of a shift in climate, the same insolation curve would pertain. Given the tight coupling of *Deep Cold* and *Late Cold* to radiative forcing, their timing and nature is much more predictable and consistent from year to year. The other stages are to a much greater extent affected by short-term changes in weather and climate.

Intermittent cyclonic storm activity continues, but the precipitable water column, and therefore precipitation, is often minimal. During *Deep Cold,* there is typically little additional snowfall. As a consequence, the snowpack depth changes little, but the snow which was emplaced in *Early Snow* and *Early Cold* continues to metamorphose into depth hoar. By the onset of *Late Cold,* all but the surface layers of snow have been transformed to this type of snow, even layers that were initially dense slabs.

The transition to *Late Cold* is also marked by insolation values of similar magnitude to the net radiation loss. (However, due to the snow's high albedo, much of this solar input is reflected back to space.) The impact of returning insolation (Fig. 1a) is increasingly apparent in the latter half of the stage as the air temperature increases. However, even as the temperature trend is upward, the reservoir of very cold polar air in the Arctic Ocean Basin continues to plunge southward periodically during early *Late Cold,* producing some of the lowest temperatures of the cold season (Fig. 1b), despite the return of the sun. A secondary burst of snowfall often occurs near the end of *Late Cold.* It is at this time that the snowpack reaches its maximum depth (Fig. 1c). In the latter half of *Late Cold* the soil temperature profile becomes essentially isothermal and soil temperatures, tracking the snow and air temperatures, continue to rise until the onset of snowmelt (Fig. 1d).

(3) DEEP COLD *AND (4)* LATE COLD—*ECOSYSTEM INTERACTIONS*

Although biological activity is at a minimum during *Deep Cold* and *Late Cold,* the biosphere continues to interact with its geophysical environment. For example, though most tundra vegetation is snow covered by *Deep Cold,* taller shrub vegetation (>50 cm) continues to trap significant amounts of blowing snow, and drift traps along cutbanks continue to grow (Sturm et al., 2001a, 2001b). This process of the vegetation capturing and holding the blowing snow may be the most important vegetation interaction with the physical environment during this period. By *Late Cold* the only vegetation exposed above the snow surface are the tall 1- to 2-m stems of riparian willow shrubs along the rivers and creeks.

Little is known regarding root function and nutrient uptake during this period, but it is suspected to be low. While laboratory studies have shown that significant microbial activity can occur in soils at temperatures in the range of −6° to −10°C and even lower (Flanagan and Bunnell, 1980; Panikov and Dedysh, 2000), the decrease in temperature in the active layer during *Deep Cold* and early *Late Cold* reduces soil respiration and other biological activity to the lowest levels of the cold season (Zimov et al., 1993; Fahnestock et al., 1999). However, as temperatures begin to rise in *Late Cold,* Zimov et al. (1996) and Fahnestock et al. (1998) reported a rising level of CO_2 efflux.

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(5) THAW—*PHYSICAL CHARACTERISTICS*

Well before the thaw stage (*Thaw*) is reached, increasing insolation along with wind-driven sublimation of snow creates patches of bare ground in areas, such as ridge crests, that had thin seasonal snow cover during *Deep Cold* and *Late Cold*. It is not until minimum daily temperatures stay above freezing, however, that substantial thawing occurs (Kane et al., 1997) and the snow depth begins to diminish. This onset of *Thaw* tends to start in the foothills on the southern end of arctic Alaska, moving northward and to lower elevations as the season progresses (Olsson et al., 2002). By the end of the *Thaw* stage, temperatures remain at or above freezing throughout the diurnal cycle at all but the near-coastal locations. After the thaw has occurred in subarctic interior Alaska, warm solar-heated air from this region is available to be advected northward, hastening the thaw process brought about by the steadily increasing insolation. This warmer continental air also has a higher water vapor content, and as a consequence produces a larger longwave (infrared) radiative energy flux to the melting snowpack (Zhang et al., 1997).

Physically, water produced by surface melt percolates downward into the snowpack, warming it rapidly. While the center of the pack remains below freezing, water arrives at the pack base and refreezes (Marsh and Woo, 1984a, 1984b). Eventually, the pack becomes isothermal at 0°C, at which time excess water begins to pond in layers. Increasingly larger patches of bare ground emerge, and these further hasten the melt. The soil remains frozen throughout this stage, limiting infiltration and maximizing runoff and stream discharge. When 95% of the snow is gone, the stage is complete and the cold season is over for the year. The remaining snow is located in deep drift traps that can last well into the summer (Sturm et al., 2001a).

(5) THAW—*ECOSYSTEM INTERACTIONS*

During *Thaw*, vegetation plays an important role in the snowmelt process by decreasing the albedo of the surface as stems and leaves become exposed. Shrub stems of birch and willow that have been depressed by the snowpack often spring back and accelerate melt locally. Thaw signals the start of the plant growing season and bud break occurs within a couple of weeks of snowmelt. Snowmelt results in a flush of nutrients contained in melt water (Giblin et al., 1991). Microbes become active at snowmelt and plants may as well, though they do not appear to capture nitrogen and other nutrients at this time (Bilbrough et al., 2000; Brooks et al., 1997). Rates of arctic tundra CO_2 efflux increase dramatically in May during *Thaw* (Fahnestock et al., 1998) as would be suggested by the rapid rise in soil temperature shown in Figure 1.

Five Stages Elsewhere—an Example

To demonstrate the utility and applicability of our definition of the five stages of the Alaska cold season, we apply it to the data record of 1999–2000 (Fig. 2) from Ivotuk, Alaska. While the timing and some of the characteristics of the five stages were different than those at Imnavait Creek (Fig. 1), all five stages were present and key characteristics were repeated.

At both sites *Early Snow* started on the first occurrence of freezing daily mean temperature after 1 September. In both cases, air temperature transitioned from above to below freezing during *Early Snow* and the soil temperature profile became isothermal shortly after the first snow of the cold season. Snow accumulated and air temperatures decreased at each site during *Early Cold*. A significant temperature gradient developed in the snowpack (Figs. 1b, 2a), and the snow began its season-long process of metamorphosis. The beginning of active layer freeze-up that marked the beginning of *Early Cold* continued at both sites. By the onset of *Deep Cold*, temperatures in the top 40 cm (root zone) of the active layer soil had fallen below 0°C and soil moisture had dropped to its asymptotic winter levels.

Deep Cold at both sites was characterized by large excursions of air temperature resulting from synoptic scale weather patterns. During the cold spells the snowpack temperature gradient increased sharply, producing the highest values of the cold season and enhancing formation of depth hoar throughout the snowpack. Temperatures in the active layer remained stratified and continued to steadily decline. With the return of the sun in *Late Cold* the air temperature began to rise and snowpack temperature gradients at both sites weakened. Concurrently, soil temperatures rose, and by the end of *Late Cold* the active layer became isothermal. *Thaw* at both sites lasted about 10–15 d, characterized by above freezing daily temperatures and a rapid decrease in snow depth.

The most apparent difference depicted in the two figures is the seasonal pattern of snowfall (Figs. 1c, 2b). The significant accumulation of snow in *Deep Cold* seen at Ivotuk is atypical in our experience, but is again testament to the variability of the cold season. The other major difference is the late *Thaw* at Ivotuk, some 30 d later than at Imnaviat. This is significant because a late *Thaw* has a much bigger impact on growing-season length than a later onset of *Early Snow*. In the autumn, plants become senescent due to the decreasing insolation, a factor which does not depend on air temperature or the arrival of snow cover (Shaver and Kummerow, 1992). By contrast, a late *Thaw* occurs during a time of high insolation, resulting in a later soil thawing and warming that may shorten the growing season considerably.

Discussion

We have avoided giving specific dates for the five stage boundaries for the following reasons: (1) because the classification system is intended to be applicable to a wide variety of locations in the terrestrial arctic, and (2) even at a given location, the year-to-year variation in weather patterns will lead to a wide variation in the timing and duration of the stages, compressing some while expanding others. Clearly, given the wide variability in weather conditions, exceptions to our stage descriptions do occur. However, given the multiple criteria used to define the boundaries, all five stages can generally be observed in any given year and location.

Defining the five stages of the cold season is useful for several reasons. First, it focuses attention on the fact that the physical and environmental conditions vary through the cold season in ways that are predictable and that can be directly tied to ecosystem processes. As such, the stages approach provides a framework for the design of a research project. Processes that might be expected to occur in one stage are unlikely to occur in another. For example, measurements in *Deep Cold* and *Late Cold* are both difficult to make and unlikely to indicate much soil biological activity, but early in *Thaw* this might no longer be the case. Second, the subdivision helps to clarify what aspects of the cold season are sensitive to changes in climate and might therefore be the most important to investigate.

In terms of sensitivity to a change in climate, *Deep Cold* and *Late Cold* are the more robust stages because their nature largely derives from the cumulative effects of minimal winter insolation, a function of latitude rather than weather. Even in a

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warming climate, these stages are likely to be of similar nature and severity to today's climate: snow covered and cold, with a frozen active layer and minimal microbial activity, even after significant insolation returns in *Late Cold.* By contrast, changes in the timing and nature of *Early Snow, Early Cold,* and *Thaw*—the transition stages between the warm and cold season—are crucial to arctic ecosystem functioning. Processes such as snow cover accumulation and active layer freeze-up that occur in *Early Snow* and *Early Cold* play a critical role in the thermal and physical conditioning of the soil and near-surface environment as insolation diminishes. Timing of *Thaw* in large measure determines growing-season length. Significantly, *Early Snow, Early Cold,* and *Thaw* are also the most likely to be affected by Arctic climate changes.

The stages are a useful tool in understanding multiple-year time series data. Relating time series observations with the stages highlights biophysical relationships and transitions that may not be readily apparent using calendar dates or seasons as the temporal variable. An understanding of the stage transitions is also useful in developing hypotheses about ecosystem responses to the cold-season variability. Climate and land surface modelers can use the five stages as a basis to develop realistic climate change scenarios (for example a dry and cold *Early Cold*) and as a useful way of interpreting and framing discussion of the results.

Acknowledgments

The authors would like to thank L. Hinzman and his colleagues for providing some of the data used in this work. Jon Holmgren helped to collect the field data. This work was supported by the U.S. Army Cold Regions Laboratory and National Science Foundation in Grant Nos. 009615 and 9732077.

References Cited

Benson, C. S., and Sturm, M., 1993: Structure and wind transport of seasonal snow on the Arctic Slope of Alaska. *Annals of Glaciology,* 18: 261–267.

Bilbrough, C. J., Welker, J. M., and Bowman, W. D., 2000: Early spring nitrogen uptake by snow-covered plants: a comparison of arctic and alpine plant function under the snowpack. *Arctic, Antarctic, and Alpine Research,* 32: 404–411.

Brooks, P. D., Schmidt, S. K., and Williams, M. W., 1997: Winter production of CO_2 and N_2O from alpine tundra: environmental controls and relationship to inter-system C and N fluxes. *Oecologia,* 110: 403–413.

Colbeck, S. C., 1986: Classification of seasonal snow cover crystals. *Water Resources Research,* 22, 59–70.

Dingman, S. L., Berry, R. G., Weller, G., Benson, C., and LeDrew, E. F., 1980: Climate, Snow Cover, Microclimate and Hydrology. *In* Brown, J., et al. (eds.), *An Arctic Ecosystem: The Coastal Tundra of Barrow, Alaska.* Stroudsburg, PA: Dowden, Hutchinson and Ross, 30–65.

Fahnestock, J. T., Jones, M. H., Brooks, P. D., Walker, D. A., and Welker, J. M., 1998: Winter and early spring CO_2 efflux from tundra communities of northern Alaska. *Journal of Geophysical Research,* 103: 29023–29027.

Fahnestock, J. T., Jones, M. H., and Welker, J. M., 1999: Wintertime CO_2 efflux from arctic soils: implications for annual carbon budgets. *Global Biogeochemical Cycles,* 13: 775–779.

Flanagan, P. W., and Bunnell, F. L., 1980: Microfloral activities and decomposition. *In* Brown, J., et al. (eds.) *An Arctic Ecosystem: The Coastal Tundra of Barrow, Alaska.* Stroudsburg, PA: Dowden, Hutchinson and Ross, 291–334.

Formozov, A. N., 1946: The covering of snow as an integral factor of the environment and its importance in the ecology of mammals and birds. *Zoology,* 5: 1–152.

Goodrich, L. E., 1982: The influence of snow cover on the ground thermal regime. *Canadian Geotechnical Journal,* 19: 421–432.

Giblin, A. E., Nadelhoffer, K. J., Shaver, G. R., Laundre, J. A., and McKerrow, A. J., 1991: Biogeochemical diversity along a riverside toposequence in arctic Alaska. *Ecological Monographs,* 61: 415–435.

Grogan, P., and Chapin, F. S., III, 1999: Arctic soil respiration: effects of climate and vegetation depend on season. *Ecosystems,* 2: 451–459.

Hinzman, L. D., 2002: Climate data for the Arctic Transitions in the Land-Atmosphere System (ATLAS) project. Fairbanks, Alaska: http://www/uaf.edu/water/projects/atlas

Hobbie, S. E., and Chapin, F. S., III, 1996. Winter regulation of tundra litter carbon and nitrogen dynamics. *Biogeochemistry,* 35: 327–338.

Jones, M. H., Fahnestock, J. T., and Welker, J. M., 1999: Early and late winter CO_2 efflux from arctic tundra in the Kuparuk River watershed, Alaska, USA. *Arctic, Antarctic, and Alpine Research,* 31: 187–190.

Kane, D. L., Glieck, R. E., and Hinzman, L. D., 1997: Snowmelt modeling at a small Alaskan arctic watershed. *Journal of Hydrologic Engineering,* 2: 204–210.

Kelley, J. J., Jr., Weaver, D. F., and Smith, B. P., 1968: The variation of carbon dioxide under the snow in the Arctic. *Ecology,* 49: 358–361.

Kelley, J. J., and Weaver, D. F., 1969: Physical processes at the surface of the arctic tundra. *Arctic,* 22: 425–437.

Labelle, J. C., 1983: *Alaska Marine Ice Atlas.* Arctic Environment and Information Data Center, University of Alaska, Anchorage. 302 pp.

Liston, G. E., McFadden, J. P., Sturm, M., and Pielke, R. A., Sr., 2002: Modelled changes in arctic tundra snow, energy and moisture fluxes due to increased shrubs. *Global Change Biology,* 8: 17–32.

McFadden, J. P., Liston, G. E., Sturm, M., Pielke, R. A., Sr., and Chapin, F. S., III, 2001: Interactions of shrubs and snow in arctic tundra: measurements and models. Soil-vegetation-atmosphere transfer schemes and large-scale hydrological models, IASH Press, IASH Publ. No. 270, Wallingford, Oxfordshire, U.K.: 317–325.

Marsh, P., and Woo, M.-K., 1984a: Wetting front advance and freezing of meltwater within a snowcover 1: observations in the Canadian Arctic. *Water Resources Research,* 20: 1853–1864.

Marsh, P., and Woo, M.-K., 1984b: Wetting front advance and freezing of meltwater within a snowcover 2: a simulation model. *Water Resources Research,* 20: 1865–1874.

Maykut, G. A., and Church, P. E., 1973: Radiation climate of Barrow, Alaska, 1962–1966. *Journal of Applied Climatology,* 12: 620–628.

Oechel, W. C., Vourlitis G. L., and Hastings, S. J., 1997: Cold season CO_2 emission from arctic soils. *Global Biogeochemical Cycles,* 11: 163–172.

Olsson, P. Q., 2001: Fall warming events on the Arctic Slope of Alaska. Sixth Conference on Polar Meteorology and Oceanography, San Diego, CA, American Meteorological Society, 312–315.

Olsson, P. Q., Hinzman, L. D., Sturm, M., Liston, G. E., and Kane, D. L., 2002: Surface climate and snow-weather relationships of the Kuparuk Basin on Alaska's Arctic Slope., U.S. Army Cold Regions Research and Engineering Laboratory ERDC/CRELL TR-02-10, Hanover, NH, 42 pp.

Osterkamp, T. E., and Romanovsky, V. E., 1997: Freezing of the active layer on the Coastal Plain of the Alaskan Arctic, *Permafrost and Periglacial Processes,* 8: 23–44.

Outcalt, S. I., Nelson, F. E., and Hinkel, K. M., 1990: The zero-curtain effect: heat and mass transfer across an isothermal re-

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gion in freezing soil, *Water Resources Research,* 26: 1509–1516.
Panikov, N. S., and Dedysh, S. N., 2000: Cold season CH_4 and CO_2 emission from boreal peat bogs (West Siberia): winter fluxes and thaw activation dynamics. *Global Biogeochemical Cycles,* 14: 1071–1080.
Pruitt, W. O., 1957: Observations of the bioclimate of some taiga mammals. *Arctic,* 10: 131–138.
Romanovsky, V. E., and Osterkamp, T. E., 1995: Interannual variations of the thermal regime of the active layer and near-surface permafrost in Northern Alaska. *Permafrost and Periglacial Processes,* 6: 313–335.
Romanovsky, V. E., and Osterkamp, T. E., 2000: Effects of unfrozen water on heat and mass transport processes in the active layer and permafrost. *Permafrost and Periglacial Processes,* 11: 219–239.
Schimel, J. P., and Clein, J. S., 1996: Microbial response to freeze-thaw cycles in tundra and taiga soils. *Soil Biology and Biochemistry* 28: 1061–1066.
Shaver, G. R., and Kummerow, J., 1992: Phenology, resourse allocation and growth of arctic vascular plants. *In* Chapin, C. S., III, et al. (eds.) *Arctic Ecosystems in a Changing Climate: An Ecophysiological Perspective.* San Diego: Academic Press. 193–212.
Sturm, M., 1991: The role of thermal conduction in heat and mass transport in the subarctic snow cover. U.S. Army Cold Regions Research and Engineering Laboratory RR 91-19 NTIS ADA 243674, Hanover, NH. 84 pp.
Sturm, M., Holmgren, J., and Liston, G., 1995: A seasonal snow cover classification system for local to global applications, *Journal of Climate,* 8: 1261–1283.
Sturm, M., Holmgren, J., Konig, M., and Morris, K., 1997: The thermal conductivity of seasonal snow. *Journal of Glaciology,* 43 (143): 26–42.
Sturm, M., McFadden, J. P., Liston, G. E., Chapin, F. S., III, Racine, C. H., and Holmgren, J., 2001a: Snow shrub interactions in arctic tundra: a hypothesis with climatic implications. *Journal of Climate,* 14: 336–344.
Sturm, M., Liston, G. E., Benson, C. S., and Holmgren, J., 2001b: Characteristics and growth of a snowdrift in Arctic Alaska, U.S.A. *Arctic, Antarctic, and Alpine Research,* 33: 319–331.
Sumgin, M. I., Kachurin, S. P., Tolstikhin, N. I., and Tumel, V. F., 1940: *Obshchee merzlotovedeniye (General Permafrost Science).* Moscow: Akademiya Nauk SSSR. 240 pp.
Taras, B., Sturm, M., and Liston, G. E., 2002: Snow-ground interface temperatures in the Kuparuk River Basin, Arctic Alaska, U.S.A.: Measurements and model. *Journal of Hydrometeorology,* 3: 377–394.
Trabant, D., and Benson, C. S., 1972: Field experiments on the development of depth hoar. *Geological Society of America Memoir,* 135: 309–322.
Weller, G., and Holmgren, B., 1974: Microclimate of the arctic tundra. *Journal of Applied Meteorology* 13: 854–862.
Zimov, S. A., Zimova, G. M., Daviodov, S. P., Daviodova, A. I., Voropaev, Y. V., Voropaeva, Z. V., Prosiannikov, S. F., and Prosiannikova, O. V., 1993: Winter biotic activity and production of CO_2 in Siberian soils: a factor in the greenhouse effect. *Journal of Geophysical Research,* 98: 5017–5023.
Zimov, S. A., Davidov, S. P., Voropaev, Y. V., Prosiannikov, S. F., Semiletov, I. P., Chapin, M. C., and Chapin, F. S., III, 1996: Siberian CO_2 efflux in winter as a CO_2 source and cause of seasonality in atmospheric CO_2. *Climatic Change,* 33: 111–120.
Zhang, T., Bowling, S. A., and Stamnes, K., 1997: Impact of the atmosphere on surface radiative fluxes and snowmelt in the arctic and subarctic. *Journal of Geophysical Research,* 102: 4287–4302.

Ms submitted April 2002

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