PERMAFROST AND PERIGLACIAL PROCESSES
Permafrost and Periglac. Process. **16**: 5–17 (2005)
Published online in Wiley InterScience (www.interscience.wiley.com). DOI: 10.1002/ppp.518

The Transient Layer: Implications for Geocryology and Climate-Change Science

Yuri Shur,[1]* *Kenneth M. Hinkel*[2] *and Frederick E. Nelson*[3]

[1] Department of Civil and Environmental Engineering, University of Alaska-Fairbanks, Fairbanks, AK, USA
[2] Department of Geography, University of Cincinnati, Cincinnati, OH, USA
[3] Department of Geography, University of Delaware, Newark, DE, USA

ABSTRACT

Research treating permafrost-climate interactions is traditionally based on a two-layer conceptual model involving a seasonally frozen active layer and underlying perennially frozen materials. This conceptualization is inadequate to explain the behaviour of the active-layer/permafrost system over long periods, particularly in ice-rich terrain. Recent research in North America supports earlier Russian conclusions about the existence of a *transition zone* that alternates in status between seasonally frozen ground and permafrost over sub-decadal to centennial time scales. The transition zone is ice-enriched, and functions as a buffer between the active layer and long-term permafrost by increasing the latent heat required for thaw. The existence of the transition zone has an impact on the formation of a cryogenic soil structure, and imparts stability to permafrost under low-amplitude or random climatic fluctuations. Despite its importance, the transition zone has been the focus of relatively little research. The impacts of possible global warming in permafrost regions cannot be understood fully without consideration of a more realistic three-layer model. The extensive data set under development within the Circumpolar Active Layer Monitoring (CALM) program will provide a significant source of information about the development, characteristics, behaviour, and extent of the transition zone. This paper is focused on the uppermost part of the transition zone, which joins the active layer at sub-decadal to multi-centennial time scales. This upper part of the transition zone is known as the *transient layer*. Copyright © 2005 John Wiley & Sons, Ltd.

KEY WORDS: transient layer; transition zone; active layer; permafrost; climate change

INTRODUCTION

Classification is a necessary step in the development of any area of scientific endeavour. Classifications are, without exception, imperfect devices developed to draw attention to and facilitate understanding about selected aspects of 'reality.' Stated alternatively, all classifications are (or should be) developed for specific purposes. It follows that a classification developed for one purpose may perform poorly for others. Problems arise when terminology and definitions based on rigid classificatory criteria become so entrenched that they acquire filter-like qualities that limit the ability of individuals to refine or reconceptualize scientific problems (Cline, 1949).

The concept of permafrost, traditionally defined as *earth materials that remain continuously at or below 0°C for at least two consecutive years* (e.g. van Everdingen, 1998), is a product of classification. Stripped to its essentials, this widely accepted definition qualifies a binary concept (above or below 0°C) by attaching an arbitrary, open-ended temporal

* Correspondence to: Y. Shur, Department of Civil and Environmental Engineering, University of Alaska-Fairbanks, Fairbanks, AK, USA 99775, USA. E-mail: ffys@uaf.edu

Received 28 May 2004
Revised 19 November 2004
Accepted 27 November 2004

criterion. Not surprisingly, as geocryological knowledge has accumulated, the traditional definition has been subject to qualification, although its basic tenets appear to remain widely accepted. The notion of 'permanency' is now widely discounted (e.g. Washburn, 1980, p. 21) although the term *permafrost* is so entrenched that efforts to dispense with it (e.g. Bryan, 1946) are considered futile. Introduction of such qualifying terms as *cryopeg*, *cryotic ground*, and *pereletok* attest to the limitations of the traditional definition.

Implicit in the traditional definition is the idea that two types of frozen ground exist: seasonally frozen and permafrost (e.g. van Everdingen, 1998). Discounting seasonally frozen ground outside permafrost regions, the traditional definition gives rise to a simple conceptual model: a layer of ground subject to freezing and thawing on an annual basis, underlain by a frozen zone that rarely or never undergoes phase change. Stated alternatively, the substrate in permafrost regions is typically divided into two layers on the basis of the thermal state. The upper part is known as the active layer, which thaws in summer and refreezes in winter. Below this is the layer that remains frozen throughout the annual cycle. Further qualification is necessary where the positions of the permafrost table and bottom of the active layer do not coincide.

To be consistent with the traditional definition of permafrost, the active layer must also include that portion of the soil profile subject to thaw once in two years. The term *pereletok* (restricted by some authors to locations outside the permafrost regions) has been used in Russian literature to describe such situations, in which a layer of ground remains frozen through one or two summers and then thaws (van Everdingen, 1998). This paper is focused on a similar concept: a part of the substrate that undergoes freeze-thaw transitions at much lower frequencies, ranging from sub-decadal to multi-centennial. This zone lies between the active layer and underlying permafrost, and has characteristics that differ from both because it occasionally joins the active layer and experiences the influence of both permafrost and the active layer (Figure 1). Termed the *transition zone*, it has a tremendous impact on the processes that gradually transform the upper permafrost, on the formation of soil and cryogenic structures, and on the thermal stability of permafrost in the face of climatic variations. Despite the importance of this layer, it has been the focus of only a few works. The objectives of this paper are: (1) to briefly review the literature related to the transition zone; (2) to develop the theoretical basis of variations in active-layer thickness over time; (3) to compare theoretical and field results; and (4) to discuss the

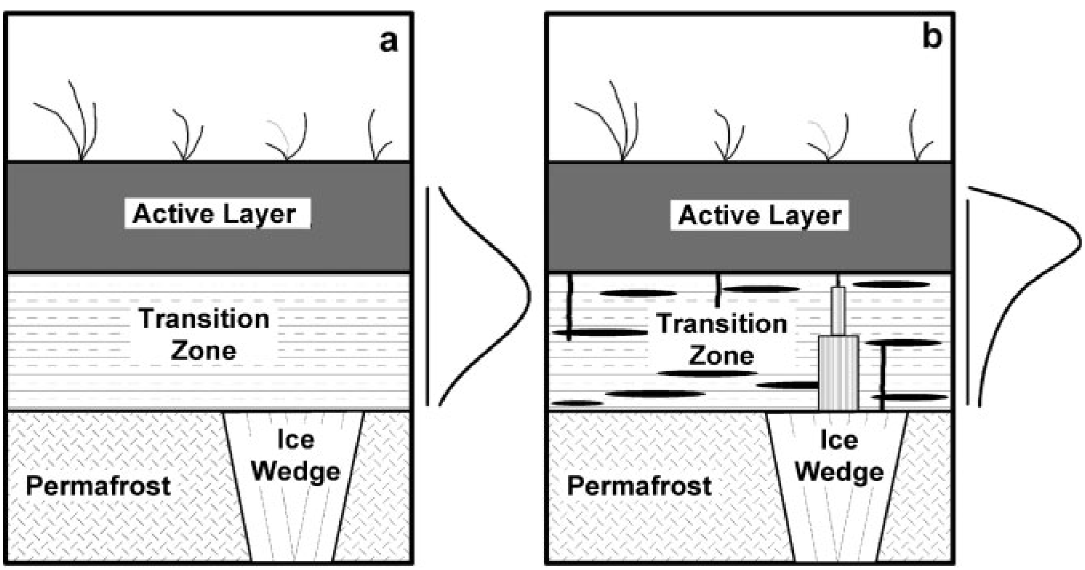

Figure 1 Schematic diagram of a three-layer conceptual model, with curve centred on mean annual active layer and showing relative probability of annual thaw depth (a) immediately following very deep thaw and (b) with ice enrichment of the transition zone after several centuries. Note formation of ice lenses (filled ellipses), upward growth of secondary and tertiary ice wedges into transition zone, and development of ice veins (from Figure 4 in Lewkowicz, 1994).

impact of the transition zone on selected periglacial processes.

THE TRANSITION ZONE

Numerous studies have demonstrated that the genesis and properties of the uppermost permafrost differ from those of the underlying permafrost. The uppermost layer is subject to episodic thaw at sub-decadal to multi-centennial scales, and is referred to in this paper as the transition zone. Its lower bound is the long-term permafrost table and is coincident with the top of primary ice wedges. The transition zone is important for permafrost stability and serves as a buffer between the active layer and ice-rich permafrost, especially permafrost containing massive ice. This zone can aggrade and change properties over time through ice segregation (Mackay, 1983) and by infiltration of meteoric water or melting ground ice from the overlying active layer, with subsequent refreezing (Hinkel *et al*., 2001).

Yanovsky (1933) studied pedogenic processes in permafrost regions and noted they were not limited to the active layer during the study period. He concluded that there is a layer that is typically a part of the uppermost permafrost but which, under certain conditions, occasionally joins the active layer. He coined the term *transient layer* to describe this part of the soil profile. Based on this finding, Sumgin *et al*. (1940) recommended considering the year in which the active-layer thickness measurement was made, and adding to the measured depth an additional increment that varies under different conditions. The additional increment could range from 1 m in dry, highly permeable soils to 0.1 m in wet organic soils.

Shur (1975, 1988a, 1988b) reconceptualized the active-layer/permafrost system by analysing the stability of the system. The existence of a transition zone, of which the transient layer forms the upper part, explains the fact that wedge ice and other massive ice formations are frequently located well below the active layer, as observed at Barrow (Brown, 1969; Estabrook and Outcalt, 1984) and elsewhere (Shur, 1975). Ice veins or rejuvenated ice wedges (secondary or tertiary growth stages) have been observed extending from primary ice wedges upward into the transition zone, indicating reactivation of ice-wedge growth (Lewkowicz, 1994).

Numerous studies demonstrated that the transition zone is ice-rich, with the ice content often much greater than the permafrost immediately beneath it (Tsitovich, 1932; Bykov and Kapterev, 1940; Kachurin, 1946; Mackay, 1971; Sellmann *et al*., 1975; Pollard and French, 1980; Shur, 1982, 1988a, 1988b; Zhestkova and Shur, 1982; Cheng, 1983; Burn, 1986; Kokelj and Burn, 2003; Kanevskiy, 2003). Ice occurs as segregation lenses, ice veins, and contemporary ice wedges, as shown in Figure 1b (Shur, 1988a; Lewkowicz, 1994). In the continuous permafrost zone, the cryogenic structure of this layer differs significantly from that of the underlying permafrost. In Russian permafrost literature, the characteristic net-like cryogenic structure found in the transition zone is known as *ataxitic*; Murton and French (1994) refer to it as *suspended*.

Owing to latent-heat effects, the ice-rich transition zone resists thaw and tends to promote thermal stability at greater depths (Shur, 1977; Smith, 1988). Although the transition zone is enriched with ice relative to the active layer, the upper region contains less ice than underlying regions because it undergoes thaw more frequently. Following Yanovsky (1933), we refer to the uppermost part of the transition zone as the *transient layer*. Deep thaw is largely driven by changes in the surface energy equation due to interannual variations and climatic fluctuations, and occurs at the sub-decadal to multi-centennial scale.

During most warm years, the transient layer protects the underlying ice-rich materials from thaw. However, using a combination of field observation and modelling, Lewkowicz and Clarke (1998) linked deep thaw to shear displacement occurring within the thawed transition zone during particularly warm summers, resulting in increased rates of solifluction on slopes.

Deeper in the transition zone, thaw is less frequent and occurs in response to a climate shift, disturbance of the organic layer, or change in surface vegetation (Shur and Jorgenson, 1998; Shur *et al*., 2003). This deeper zone is more stable since thaw is forced by a different set of processes. Furthermore, deep thaw is a response to local disruption or regional climate shift, and the effects may not be observed everywhere in regions of continuous permafrost. From this perspective, the upper transient layer is more amenable to probabilistic modelling, and is the focus of this paper.

A Probabilistic Approach—Hydrological Analogue

To demonstrate the probabilistic basis of the transient layer, we draw an analogy with a well-known frequency-magnitude example from hydrology: the notion of a recurrence interval. Given a time series of maximum annual flood events as measured by discharge (Q). Each maximum flood event in the series of N events is magnitude ranked from 1 (highest) to N (lowest). The recurrence interval is calculated as the time between two annual discharge events of equivalent or larger magnitude, given by $R = (N+1)/M$, where R is the recurrence interval in years, N is the duration of the time series, and M is the rank of the given annual discharge. The probability of a flood event of a particular magnitude is the reciprocal of the recurrence event ($P = R^{-1}$). Thus, a 25 year flood (Q_{25}) has a 4% chance of occurring in any given year. The mean annual flood has a recurrence interval of approximately 2.33 years, or a 43% chance of occurring in any given year.

Stream discharge is impacted strongly by precipitation in the watershed, but also reflects the effects of snowmelt, infiltration, evaporation, and transpiration. Discharge can, therefore, be considered a spatially integrated response to several factors operating in the drainage basin upstream from the gauging station. Because the forcing and modulating functions have strong temporal components, so too does the occurrence of floods. Similarly, annual thaw depth is affected strongly by summer heat input but is modulated by the effects of rainfall, soil moisture, wind speed, cloudiness, and other components of the surface energy equation. Maximum thaw depth is therefore a temporally integrated response to factors that operate throughout the thaw period. In many situations, it appears that the condition of the previous autumn also has an impact on the soil thermal properties. In this case, the active layer and transient layer may display a hysteretic effect, retaining the memory and influence of previous events and conditions (Miller *et al*., 1998).

As a first approximation, maximum annual thaw depth should exhibit a similar probabilistic nature if system conditions remain stable over the period, and this approach has been used by Lewkowicz (1994) to estimate the probability of observing secondary ice wedges following ice-wedge growth rejuvenation. This conceptualization is illustrated in Figure 1a, which shows a normal distribution curve with 43% probability located at the base of the average *annual* active layer. Permafrost, or ground frozen for two or more years, would be encountered at the curve peak (probability of 50%). Below this, the probability of thaw decreases. Toward the surface, there is a greater chance of thaw and, at some point immediately beneath the ground surface, the probability is 100%. In a sense, this represents the 'truly active' active layer in that, despite interannual variations in forcing factors and resulting thaw depth, it consistently experiences a freezing-thawing cycle.

The transient layer has spatially variable thickness but, over the sub-decadal to centennial time scale, the base is coincident with the tops of primary ice wedges. However, between periods of maximum thaw events, the properties of the transient layer evolve as interstitial ice, ice lenses, ice veins, and veins of wedge ice accumulate. Ice enrichment requires greater heat input to thaw compared to the initial condition; the amount is equal to the product of the enriched ice mass and the latent heat of fusion. This means that the probability of thawing to a greater-than-average depth will decrease with time as ice accumulates and the thermal inertia increases. The normal distribution curve shown in Figure 1a is therefore only valid for the period immediately following extremely deep thaw. As ice enrichment of the transient layer occurs over time, the curve loses its symmetric Gaussian form and becomes increasingly skewed, as shown in Figure 1b. The part of the curve representing higher probability will be progressively displaced toward the surface.

THEORY OF THAW VARIATION

The annual active-layer thickness within this basic unit is a function of the defining factors, and can be approximated from the Stefan equation.

$$z = \sqrt{\frac{2kI}{\rho W l}} \tag{1}$$

where: z = the active-layer depth (m), k = soil thermal conductivity ($\mathrm{W\,m^{-1}\,K^{-1}}$), I = thawing index (K·s), ρ = density ($\mathrm{kg\,m^{-3}}$), W = water content fraction and l = latent heat of water fusion (0.334 $\mathrm{MJ\,kg^{-1}}$ at 0°C).

If we are considering the average condition within the soil column, z, k, I, and W can all assume the subscript$_{av}$. A discrete change in the average thaw depth (Δz) varies around the mean value in response to changing conditions in (Δk, ΔI, ΔW). The thaw depth variation is described as:

$$\begin{aligned} z_{av} + \Delta z &= \sqrt{\frac{2(k_{av}+\Delta k)(I_{av}+\Delta I)}{\rho l (W_{av}+\Delta W)}} \\ &= \sqrt{\frac{2k_{av}I_{av} + 2I_{av}\Delta k + 2k_{av}\Delta I + 2\Delta k \Delta I}{\rho l (W_{av}+\Delta W)}}, \end{aligned} \tag{2}$$

which can also be expressed as a fractional (or percentage) change:

$$\frac{z_{av}+\Delta z}{z_{av}} = 1+\frac{\Delta z}{z_{av}} = \sqrt{\frac{1+\frac{\Delta k}{k_{av}}+\frac{\Delta I}{I_{av}}+\frac{\Delta k \Delta I}{k_{av} I_{av}}}{1+\frac{\Delta W}{W_{av}}}} \quad (3)$$

By rearranging,

$$\Delta z = z_{av}\left(\sqrt{\frac{1+\frac{\Delta k}{k_{av}}+\frac{\Delta I}{I_{av}}+\frac{\Delta k \Delta I}{k_{av} I_{av}}}{1+\frac{\Delta W}{W_{av}}}}-1\right) \quad (4)$$

and

$$\frac{\Delta z}{z_{av}} = \sqrt{\frac{1+\frac{\Delta k}{k_{av}}+\frac{\Delta I}{I_{av}}+\frac{\Delta k \Delta I}{k_{av} I_{av}}}{1+\frac{\Delta W}{W_{av}}}}-1 \quad (5)$$

The expression $\frac{\Delta k \Delta I}{k_{av} I_{av}}$ can be much smaller than $\frac{\Delta k}{k_{av}}$ and $\frac{\Delta I}{I_{av}}$. If $\frac{\Delta k}{k_{av}}$ and $\frac{\Delta I}{I_{av}}$ are on the order of 0.1, the cross product can be safely neglected. However, it can become important if there is a significant increase in $\frac{\Delta k}{k_{av}}$ and $\frac{\Delta I}{I_{av}}$.

The approximate change in the response function with variations in the defining factors can be also found using the well-known theoretical equation of error:

$$\Delta f = \sum_{i=1}^{k} \Delta a_i \left| f_i'(a_1, \ldots, a_k) \right| \quad (6)$$

In relation to variations in active-layer thickness, this equation can be rewritten as:

$$\Delta z = \sum_{i=1}^{k} \Delta a_i \left| z_i'(a_1, \ldots, a_k) \right| \quad (7)$$

where: $\Delta a_i =$ variation of a factor a_i, and $z_i' =$ partial derivation of z in respect to a_i.

Error theory can be applied to each of the factors governing thaw in the Stefan equation [equation (1)]. Note that this applies to the average condition within the soil column.

$$\frac{dz}{dk} = \frac{1}{2}\frac{2I}{\rho W l\sqrt{\frac{2kI}{\rho W l}}} = \frac{z^2}{kz} = \frac{1}{2}\frac{z}{k} \quad (8a)$$

$$\frac{dz}{dI} = \frac{1}{2}\frac{2I}{\rho W l\sqrt{\frac{2kI}{\rho W l}}} = \frac{1}{2}\frac{z}{I} \quad (8b)$$

$$\frac{dz}{dW} = -\frac{1}{2}\frac{2Ik}{W^2 \rho l\sqrt{\frac{2kI}{\rho W l}}} = -\frac{1}{2}\frac{z}{W} \quad (8c)$$

Because the above expressions represent the average condition within the soil column, the impact of changes from variation of the governing factors can be expressed as:

$$\Delta z = \frac{1}{2} z_{av}\left(\left|\frac{\Delta k}{k_{av}}\right| + \left|\frac{\Delta I}{I_{av}}\right| + \left|\frac{\Delta W}{W_{av}}\right|\right) \quad (9)$$

Equation (9) overestimates maximum variations of active-layer thickness, primarily because the error equation predicts small variations but also because it does not account for interdependence between factors. Such dependence definitely exists between thermal conductivity and water content, for example, because higher water contents increase the thermal conductivity. The water content and its distribution in the frozen active layer prior to thawing depend on the soil water content the previous autumn, and subsequent migration of water to the freezing front. They completely define the latent heat and the rate of thawing, but only partly impact the thermal conductivity of the thawing soil because the latter also depends on precipitation during the thaw season.

Equation (10) describes the relation of the defining factors in the Stefan equation succinctly.

$$\Delta z = \frac{1}{2} z_{av}\left|\frac{\Delta k}{k_{av}} + \frac{\Delta I}{I_{av}} - \frac{\Delta W}{W_{av}}\right| \quad (10)$$

Equations (4) and (10) show that variations in active-layer thickness result from changes in the thawing index, the thermal conductivity of the thawing portion of the active layer, and from the latent heat potential of its ice-rich frozen portion. Not only do Δk and ΔW have an impact equivalent to changes in the thawing index (ΔI), they can be even more important because their relative changes can exceed that of the thawing index. A wet summer following one or several dry summers can produce the greatest increase in active-layer thickness because the dry summer results in low latent heat in the frozen active layer and the thermal conductivity of the thawing soil increases in the wet summer. It should be emphasized that a warm summer does not necessarily imply warm soil temperatures in the active layer because soil temperature and active-layer thickness depend on both air temperature and soil properties.

THEORY AND OBSERVATION

Active-layer Data Records

Vasil'ev (1981) evaluated active-layer thickness at meteorological stations in western Yakutia. Maximum annual thaw depth was interpolated from soil temperature, measured at standard depths. These data are

Table 1 Average active-layer thickness and maximum variation for sites in West Siberia (based on Vasil'ev, 1981).

Site	Monitoring period, years	z_{av}, m	$\frac{z_{\max} - z_{av}}{z_{av}}$, %
Kusur	8	1.15	17
Dzhangky	5	0.9	22
Druzhina	17	1.2	12.5
Batagay-Alyta	8	0.8	18
Verkhoiansk	17	1.2	6
Srednekolymsk	17	1.55	10
Zhigansk	6	1.85	8
Ust'-Moma	16	1.0	20
Zyrianka	17	2.45	12
Bestiakh fur farm	13	1.55	3
Predporozhnaia	17	1.0	17
Sangary	14	1.7	11.2
Deliankir	15	1.05	14
Oimiakon	15	1.6	10
Krest-Khal'dzhy	17	1.45	11.7
Okhotsky Pervoz	14	1.65	6
Allakh-Un'	14	3.0	6

presented in Table 1, and show the relation between the average active-layer thickness and its maximum variation (%) during the periods of record. Pavlov (1998) studied the active layer at the Marre-Sale experimental site on the Yamal Peninsula; his data are presented as Table 2. Interannual variations of active-layer thickness in the Yakutsk region were studied by Skriabin *et al*. (1998), and are shown in Table 3.

Figure 2 presents data from Tables 1, 2, and 3, with the variation (%) as a function of average active-layer thickness. All data show that relative variations decrease with increasing active-layer thickness. In the same region a thinner active layer is associated with organic soils. Shur (1988b) demonstrated that variations in the active layer are greater in organic soils than in mineral soils, and Pavlov's (1998) data support this contention. Data in Tables 1, 2, and 3 cannot be directly interpreted as the relative thickness of the transient layer because periods of monitoring do not exceed 20 years and many are less than 10 years. The data demonstrate, however, that the relative thickness of the transient layer varies widely depending on climate and soil, and in some cases can exceed 30% of the active-layer thickness (site 8a in Table 3).

Applications

Both equations (4) and (10) can be used to analyse the impact of varying factors on the change in active-layer thickness and, consequently, on the thickness of the transient layer. It is common to look for a change in the thawing index (ΔI) as the primary determinant. Existing climatic data make it easy to verify this supposition. Using climate information for Barrow, Kotzebue, and Fairbanks in Alaska, we evaluated possible changes in active-layer thickness due to variations in the thawing index.

At Barrow, the average thawing index for the period 1950–2003 is 8.9°C-months. In 1989, the thawing index was more than twice this magnitude, when it reached 18.3°C-months. This yields an increase in the active-layer depth by 50% if we use equation (10) and by 43% if we calculate the increase directly from equation (4). Similarly, the maximum expected variation of active-layer thickness due to changes in the thawing index is equal to 13% in Kotzebue, 7% in Fairbanks, and 4.5% in Yakutsk using thawing index data for the period 1954–1965 (Are and Demchenko, 1972). These variations are shown as series 1 in Figure 3.

Series 2 in Figure 3 is based on the active-layer data presented by Vasil'ev (1981) for 17 meteorological stations in western Yakutia (Table 1) and climate data for the same stations (*USSR Climate Data Handbook*, 1966). An impact of the thawing index change (ΔI) on

Table 2 Variations in active-layer thickness from 1978 to 1994 at the Marre-Sale experimental site on the Yamal Peninsula (based on Pavlov, 1998).

Site	Terrain unit	z_{av}, m	$\frac{z_{\max} - z_{av}}{z_{av}}$, %
1	Poligonal tundra with sedge-dwarf birch-moss cover	1.13	15
2	Polygonal tundra with cloudberry–polytrichum–sphagnum cover	0.83	24
3	Peatland with cotton grass-sedge cover	0.75	28
4	Sedge swamp	0.65	20
5	Polygonal tundra with thick moss-lichen cover	0.43	17
6	Polygonal tundra with frost boils	1.23	11
7	Bare sandy south facing slop	1.48	13
8	Hummocky tundra	1.26	11
9	Grass tundra	1.13	11

Table 3 Variations in active-layer thickness near Yakutsk (based on Skriabin *et al*., 1998).

Site	Monitoring period, years	Terrain units	z_{av}, m	$\frac{z_{max} - z_{av}}{z_{av}}$, %
		Sites on a gentle slope with sparse pine forest and thin surface vegetation		
4	8		2.9	8
5	16		3.58	8
6	9		3.24	13
6b	12		3.96	4
7	9		2.48	9
7b	11		1.88	6
9	13		1.57	10
10	13		1.87	13
11	13		1.94	17
		Sites in shallow linear depression with dense birch forest and thick moss. Peat layer from 0.3 to 1.3 m		
4a	12		1.05	19
3a	12		0.43	16
8	15		1.14	19
8a	9		1.03	34
4/100	9		1.24	19

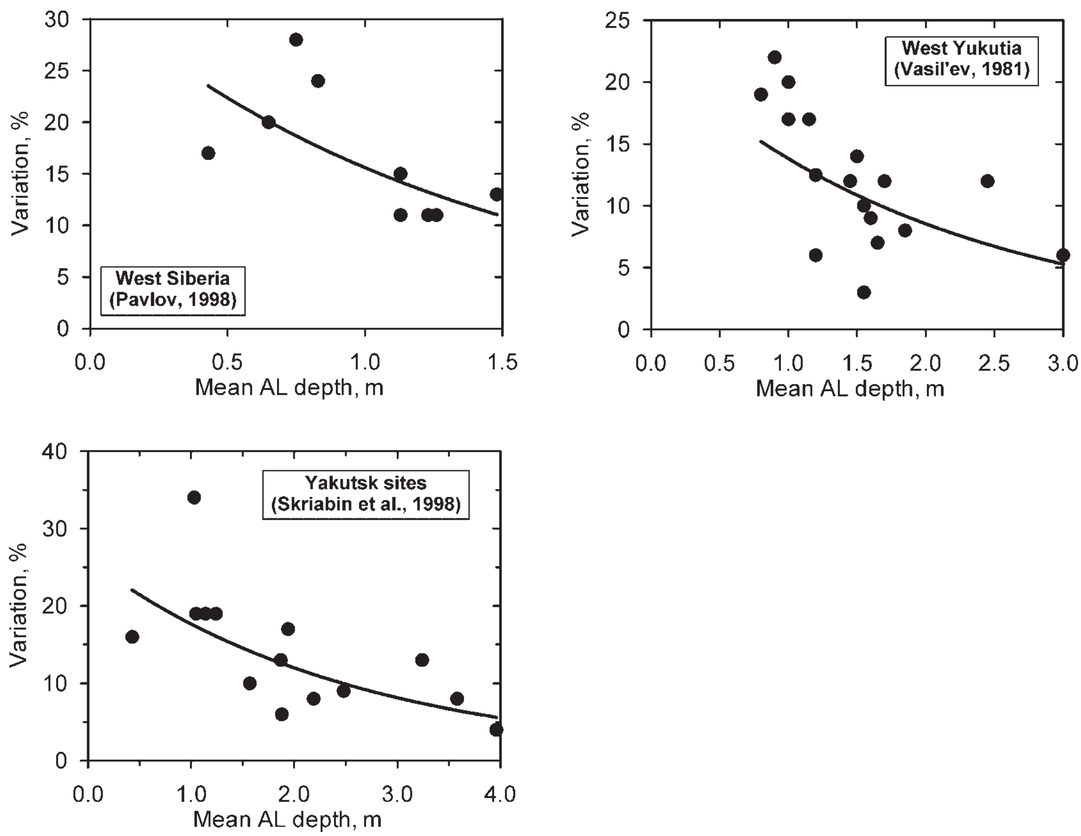

Figure 2 Variations of active-layer thickness vs. average active-layer thickness. Curves are based on an exponential best-fit function. AL = active layer.

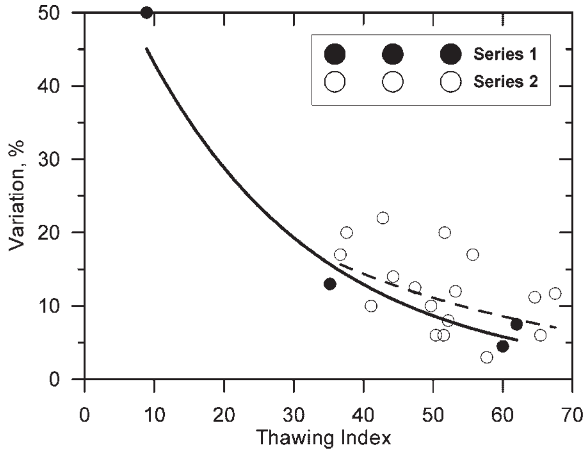

Figure 3 Estimates of variations in active-layer thickness with thawing index (°C-months) for series 1 (Barrow, Kotzebue, and Fairbanks for the period 1950–2003 and Yakutsk for the period 1954–1965), and series 2 (data presented in Table 1 for 17 sites in western Yakutia, based on Vasil'ev (1981) and on climate data for these sites from *USSR Climate Data Handbook* (1996)). Curves based on an exponential best-fit function.

variations in active-layer thickness is more apparent in soils where water content is less variable. These include sites with constantly saturated soil. Our data from the Circumpolar Active Layer Monitoring (CALM) site at Barrow (Miller *et al*., 1998; Hinkel and Nelson, 2003) show that variation in active-layer thickness within central marsh (saturated soil) is smaller than that observed on the dry beach ridge or polygonized upland. Most field measurements demonstrate a stronger dependence of active-layer variations on the thawing index than theoretically evaluated because field data also reflect the impacts of other factors. Both theoretical estimates and field measurements show that relative variations in active-layer thickness increase with a decrease in the thawing index. This confirms the predictions of Shvetsov (1963) that variations in the thawing index have greater impact on the variations in the active-layer depth in the High Arctic than in continental areas at lower latitudes.

Active-layer thickness can be defined as an average for arbitrarily defined periods of time. The probability of this depth being achieved in any given year is about 43%. The probability of thawing to smaller depths increases from 43% to 100%, while the probability of thawing to greater depths decreases from 43% to 0%. Vasil'ev (1987) evaluated the probability of the active-layer thickness for four sites in Central Yakutia. The data are based on interpolation of soil temperature for a period of 20 years. Average active-layer thickness and its maximum relative variations are presented in

Table 4 Variations in the active-layer thickness in central Yakutia (based on Vasili'ev, 1987).

Site	Monitoring period, years	z_{av}, m	$\frac{z_{\max} - z_{av}}{z_{av}}$, %
Yakutsk	20	2.17	12.2
Pokrovsk	20	2.09	5.7
Amga	20	1.59	20.1
Isit'	20	1.12	9.0

Table 4, and the exceedance probability distribution is shown in Figure 4.

Brown (1969) and Nelson *et al*. (1998b) reported results from a programme initiated at Barrow in the 1960s by the US Army's Cold Regions Research and Engineering Laboratory (CRREL). A robust sample ($n = 320$) of end-of-season thaw depths were collected from 19 plots over the period 1962–1968, and again from 1991 to the present. Thaw depths in the 1960s were substantially greater given the same air temperature forcing, as measured by the thawing index (Nelson *et al*., 1998b; Hinkel and Nelson, 2003). This reflects a fundamentally different response between the two periods, and implies a significantly higher soil thermal inertia in the 1990s. Nelson *et al*. (1998b) interpreted this 'Markovian' behaviour as suggestive of ice enrichment near the base of the active layer. This apparently occurred following a year of shallow thaw during the intervening period (1969–1990), which allowed ice segregation to occur at the new level and 'reset' the system to a shallower value of active-layer thickness. Ice-content data from the Barrow CRREL plots, presented by Hinkel *et al*. (1996, Figure 3) for 1963 and 1993, are consistent with this interpretation.

The average annual thaw depth and recurrence interval for the CRREL sites at Barrow is presented in Figure 5. Most of the measurements made in the 1960s cluster at longer recurrence intervals (low-probability events), indicating, as expected, that the data are temporally autocorrelated. Conversely, most of the more recent measurements are found at shorter recurrence intervals. The single exception is 1998 (see Figure 5), which was the warmest summer on record in northwestern North America. The high temperatures that year triggered relatively deep thaw. Although thaw extended into the uppermost part of the transient layer, penetration was insufficient to 'reset' its characteristics, and subsequent years have experienced relatively shallow active layers. Field investigations at Barrow under CALM II will continue to address the long-term evolution of the active layer

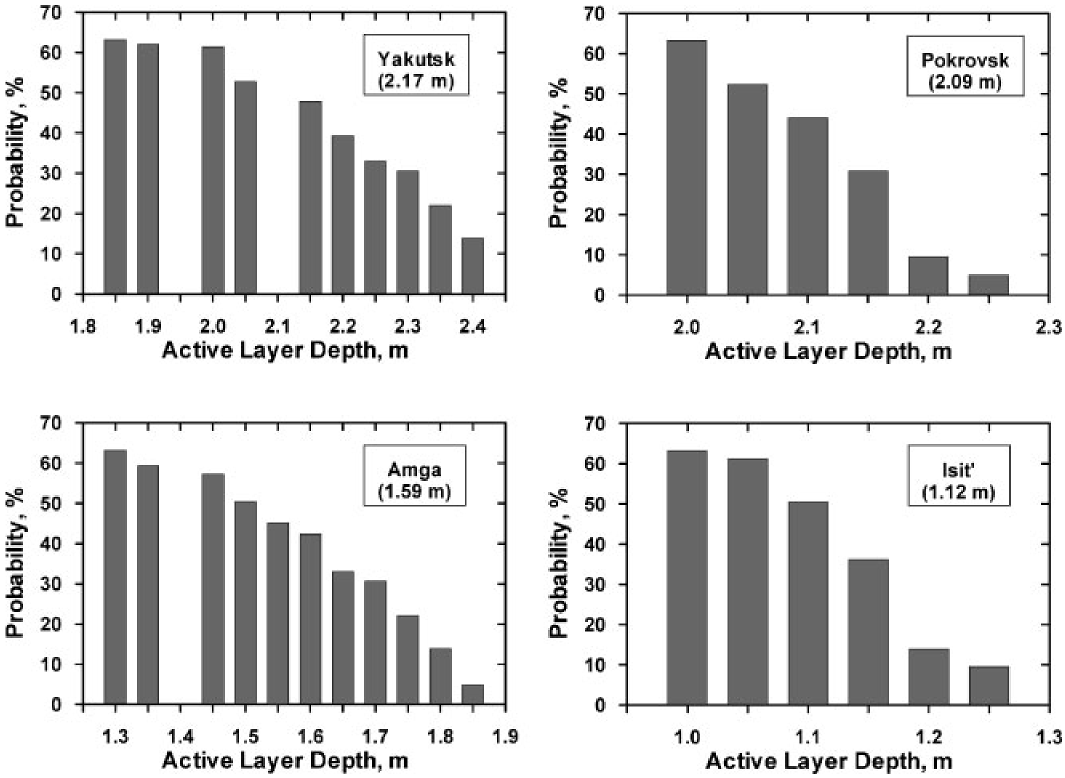

Figure 4 Exceedance probability of active-layer thickness for sites in central Yakutia.

at the CRREL sites, including collection and stratigraphic analysis of soil cores.

The relative stability of active-layer thickness in many Arctic landscapes indicates the existence of self-regulating mechanisms that inhibit change at depth, and contribute a robustness to the upper permafrost with respect to external climatic forcing. Although the magnitude, frequency, and variability of these forcing processes are not well documented, the data record from Barrow indicates that they may be responsible for abrupt, long-lasting 'Markovian' changes in active-layer thickness.

DISCUSSION

Given a sufficiently long time series of maximum annual events, it is possible to determine the probability of thaw depth of specific magnitude relative to a long-term average. This relation will only hold true if the system remains relatively stable during the time period. Returning to the hydrology analogy, a major change in the drainage basin, such as that which occurs following urbanization, alters the discharge response to the forcing factors. In this case, the system has shifted to a new state and the statistical parameters used to previously describe the discharge response will no longer be valid.

We hypothesize that analogous controls apply to the maximum annual thaw. Over some unknown time scale, the transient layer experiences a regime shift due to natural processes, as the Barrow CRREL example indicates. It is the Markovian behaviour of the active layer with state transitions at decadal or lower frequency that might limit the utility of the frequency-magnitude approaches discussed above.

Given the interannual variability of maximum thaw depth, we must consider the impact of the length of the record or time series. We must also account for the additional fact that it can vary over short lateral distances, resulting in highly variable thaw depths across space.

Because most active-layer records are short (Brown *et al.*, 2000), they have limited utility for revealing frequency-magnitude aspects of the transient layer. End-of-season measurements can, however, provide useful information on spatial variability if an appropriate spatial sampling scheme is implemented (Brown *et al.*, 2000). This is particularly true if various major vegetation-landscape

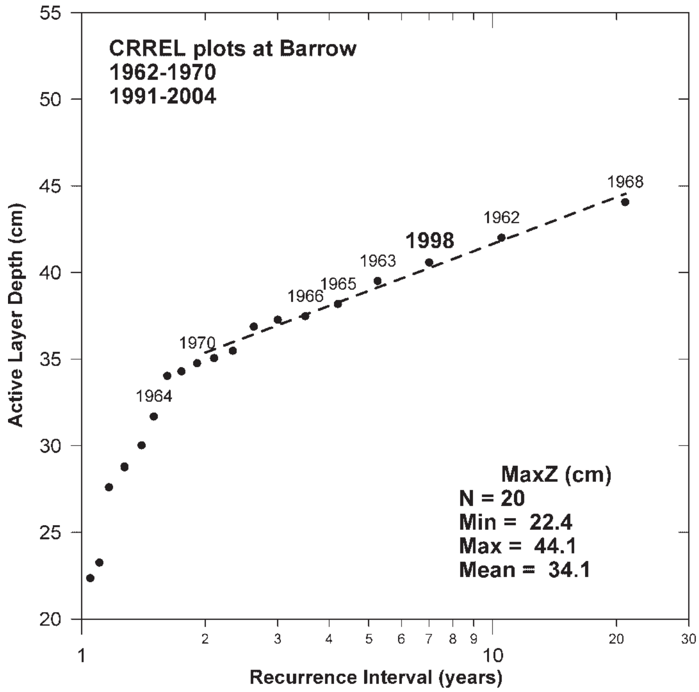

Figure 5 Maximum annual thaw depth (cm) at CRREL sites at Barrow ($n = 320$) and recurrence interval (years) over the discontinuous period of record. z = active-layer depth.

units are sampled regionally. Because the summer air temperature field is reasonably homogeneous at such scales, the impact of local modulators such as vegetation and soil type, soil moisture, snow thickness patterns, and topography can be observed (Nelson *et al.*, 1997, 1998a; Klene *et al.*, 2001; Shiklomanov and Nelson, 2002).

Longer time series of maximum annual thaw depth have the potential to provide insights under certain conditions. First, given the high degree of spatial variability in thaw depth, a spatially representative sample is required. Second, replicate measurements must be made at the same time period each year, and near the same location. Finally, data must be collected using a method applied consistently throughout the time period, so as to avoid introduction of systematic error. Data meeting these requirements have been collected since the early 1990s under the auspices of the CALM program (Brown *et al.*, 2000). In Alaska, the record at most sites currently incorporates nine thaw seasons; at the Alaskan CALM grids, maximum annual thaw depth is the average of 121 measurements made at surveyed nodes extending over a 1 km^2 area. The Alaskan sites demonstrate a high degree of interannual synchronicity in their response despite separation distances of up to 200 km. At these tundra sites, maximum thaw depth is strongly correlated with the regional summer air temperature pattern over the period of record, indicating that there was no regime change in the active layer over this period (Nelson *et al.*, 1998b; Hinkel and Nelson, 2003).

The following two examples show the importance of the transient layer for permafrost development.

Formation of ice wedges. It is widely presumed that the upper surface of ice wedges merges with the base of the active layer. According to Shumskii (1959), ‘The depth of occurrence of the tops of ice wedges in most cases is equal to the contemporary or ancient active layer. If the top of an ice wedge does not reach the bottom of the active layer, it means that an ice wedge does not grow at the present time.’ When thin ice veins are found between the top of an ice wedge and the bottom of the active layer (see Figure 1), it is often interpreted as an indicator of climate cooling. Thin ice veins between the main body of an ice wedge and the base of the active layer, however, can be explained simply by the existence of a transient layer in which an ice wedge attempts to develop and in which ice veins are episodically destroyed (see Lewkowicz, 1994). This process is extremely important for the self-protection of the ice-wedge system against thermokarst. Shur (1974) studied the causes of thermokarst and concluded that the transient layer can be a regulator that protects the system against thermokarst developed in response to short-term climate change. The Russian geographer V. Sochava (1974) concluded that geosystems in which different parts are intimately adjusted to each other are extremely unstable. The transient layer therefore serves as a buffer between the active-layer and ice-rich permafrost in that it protects ice wedges during the occasional increase in the active-layer depth by providing room for fluctuation of active-layer thickness above ice wedges. Episodic thaw penetration into the transient layer is evidenced by truncation of secondary and tertiary ice-wedge veins (Figure 1b, and see also Figure 4 in Lewkowicz, 1994).

Formation of syngenetic permafrost. Syngenetic permafrost forms simultaneously with the accumulation of sediment on the soil surface. This process, which currently occurs on floodplains and slopes in the Arctic, was extremely active in periglacial areas during the Late Pleistocene. During syngenetic permafrost formation, permafrost aggradation and sediment deposition occur simultaneously and in the same direction. Given a constant active-layer thickness, the rate of permafrost formation could be equal to the rate of sedimentation and would be constant if the rate of sedimentation is constant. The existence of the transient layer, however, makes syngenetic permafrost growth intermittent.

The transient layer, not the thickness of the additional sediment, controls the thickness of the additional increment of permafrost. The cryogenic structure found in syngenetic permafrost includes both relatively thick ice layers, and soil with microlenticular or thin reticular cryogenic structures between the ice layers. According to Popov (1967), such a structure is formed by the addition to permafrost of a thickness equal to the distance between ice layers which is greater than the thickness of annual sediment accumulated. This is possible because of fluctuations in active-layer thickness and continuous accumulation of sediment. Ice layers reflect maximum thaw depth, and the layer of soil with lenticular cryogenic structure is formed between extremes in active-layer thickness or, in our interpretation, in the transient layer. With accumulation of sediment on the soil surface, the transient layer also moves upward.

CONCLUSIONS

The two-component (active layer/permafrost) conceptual model is inadequate to explain the behaviour of the active-layer/permafrost system, particularly in ice-rich terrain. Furthermore, it does not properly describe the cryogenic structure of the upper permafrost and temporal changes in active-layer thickness. The existing definitions of permafrost and the active layer do not identify the layer of soil that changes its state from perennially frozen to seasonally thawed over a duration exceeding two years. Perhaps there is an alternate way to describe the changing thermal character of the uppermost permafrost but, accepting existing definitions of the active layer and permafrost, we must accept the existence of a layer that episodically dithers between the two end members. The traditional definition obscures effective understanding of the formation and properties of the upper permafrost and syngenetic permafrost, and makes a realistic determination of the stability of arctic geosystems under climatic fluctuations virtually impossible. Numerous studies of the active layer do not typically target the transient layer, and our current knowledge of this layer is very limited. To date, records of sufficient length to determine site-specific frequency-magnitude patterns of thaw depth are largely unavailable. Although the duration necessary to capture the regime characteristics are unknown, we suggest that a minimum period of three decades is probably required. In this case, it would entail analysis of 30-year records, much like climate normals.

The thickness of the transient layer plays a crucial role in evaluating the potential response of the active-layer/permafrost system to climatic change, and surface disturbances and for development of thermokarst. For well-developed thermokarst terrain to evolve, the long-term maximum thaw depth must be achieved consistently from year to year during the thawing season. Although the ice-rich transition zone acts to retard the rate of degradation, progressive thaw under monotonic climate warming would lead to its destruction, with attendant thaw consolidation. Elimination of the ice-rich transition zone may be accompanied by long-lasting increase in the thickness of the active layer. Thaw penetration into spatially heterogeneous ground ice in the underlying permafrost triggers differential thaw settlement at the surface. Currently, such process can be observed at sites where the vegetation has been disturbed or destroyed. However, to an observer measuring active-layer thickness using traditional methodology (e.g. mechanical probing), thaw penetration into an ice-rich transient layer may not be apparent owing to thaw consolidation and net subsidence of the surface. Conversely, following colder summers, ice may be added to the upper permafrost and possibly result in a subsequent decrease of active-layer thickness.

The transient layer plays a prominent theoretical role in the second phase of the CALM program, and several of CALM II's observation and experimental subprograms are devoted to it. The data record produced from the CALM network's 125+ sites (Brown *et al*., 2000) will help to allay the data paucity problem.

ACKNOWLEDGEMENTS

The Alaska EPSCoR program, 0092040-EPS grant supported this study. This work is supported by the U. S. National Science Foundation (NSF) under grants EPS-0092040 to YS, OPP-9732051 and OPP-0094769 to KMH, and OPP-0095088 and OPP-0352958 to FEN. Any opinions, findings, conclusions, or recommendations expressed in this material are those of the author(s) and do not necessarily reflect the views of the NSF. We thank Jerry Brown, Torre Jorgenson, Kolia Shiklomanov, and Alan Heginbottom for useful discussions about various aspects of the problem addressed in this paper. Steve Kokelj and an anonymous reviewer provided helpful comments.

REFERENCES

Are AL, Demchenko R. 1972. Some results of long-term monitoring of the active layer in the Yakutsk vicinity. In *Experimental Studies of Heat Exchange in Frozen Soils*. Akademiia Nauk SSSR: Moscow; 91–96 (in Russian).

Brown J. 1969. Soil properties developed on the complex tundra relief of northern Alaska. *Biuletyn Peryglacjalny* **18**: 153–167.

Brown J, Hinkel KM, Nelson FE. 2000. The Circumpolar Active Layer Monitoring (CALM) program: research designs and initial results. *Polar Geography* **24**: 165–258.

Bryan K. 1946. Cryopedology—the study of frozen ground and intensive frost-action with suggestions on nomenclature. *American Journal of Science* **244**: 622–642.

Burn CR. 1986. *On the Origin of Aggradational Ice in Permafrost.* Ph.D. thesis, Carleton University, Ottawa, Ontario.

Bykov NI, Kapterev PN. 1940. *Permafrost and Construction on it*. Transzheldorizdat: Moscow; 371 pp. (in Russian).

Cline MG. 1949. Basic principles of soil classification. *Soil Science* **67**: 81–91.

Cheng G. 1983. The mechanism of repeated-segregation for the formation of thick layered ground ice. *Cold Region Science and Technology* **8**(1): 57–66.

Estabrook GF, Outcalt S. 1984. An algorithm for clustering profile data and its application to near-surface ice content data from wet coastal tundra soils near Barrow, Alaska. *Mathematical Geology* **16**: 193–205.

Hinkel KM, Nelson FE. 2003. Spatial and temporal patterns of active layer thickness at circumpolar active layer monitoring (CALM) sites in northern Alaska, 1995–2000. *Journal of Geophysical Research-Atmospheres* **108**: 81688168, DOI:10.1029/2001JD 00927,2003.

Hinkel KM, Nelson FE, Shur Y, Brown J, Everett KR. 1996. Temporal changes in moisture content of the active layer and near-surface permafrost at Barrow, Alaska: 1962–1994. *Arctic and Alpine Research* **28**: 300–310.

Hinkel KM, Paetzold R, Nelson FE, Bockheim JG. 2001. Patterns of soil temperature and moisture in the active layer and upper permafrost at Barrow, Alaska: 1993–1999. *Global and Planetary Change* **29**: 293–309.

Kachurin SP. 1946. *On Genesis of the Most Widely Occurring Ground Ice*. USSR Academy of Sciences: Moscow-Leningrad.

Kanevskiy MZ. 2003. Cryogenic structure of mountain slope deposits, northeast Russia. In *Proceedings of the Eighth International Conference on Permafrost*, Vol. 1. AA. Balkema Publishers: Lisse, The Netherlands; 513–518.

Klene AE, Nelson FE, Shiklomanov NI, Hinkel KM. 2001. The n-factor in natural landscapes: variability of air and soil-surface temperatures, Kuparuk River basin, Alaska. *Arctic, Antarctic, and Alpine Research* **33**: 140–148.

Kokelj SV, Burn CR. 2003. ''Drunken forest'' and near surface ground ice in Mackenzie Delta, Northwest Territories, Canada. In *Proceedings of Eighth International Conference on Permafrost*, Vol. 1. AA. Balkema Publishers: Lisse, The Netherlands; 567–572.

Lewkowicz AG. 1994. Ice-wedge rejuvenation, Fosheim Peninsula, Ellesmere Island, Canada. *Permafrost and Periglacial Processes* **5**: 251–268.

Lewkowicz AG, Clarke S. 1998. Late-summer solifluction and active layer depths, Fosheim Peninsula, Ellesmere Island, Canada. In *Proceedings of the Seventh International Conference on Permafrost*. Cente d'études nordiques, Université Laval: Québec; 641–646.

Mackay JR. 1971. Ground ice in the active layer and the top portion of permafrost. In *Proceedings, Seminar on the Permafrost Active Layer*. National Research Council of Canada Memorandum **103**: 26–30.

Mackay JR. 1983. Downward water movement into frozen ground, western arctic coast, Canada. *Canadian Journal of Earth Sciences* **20**(1): 120–134.

Miller LL, Hinkel KM, Nelson FE, Paetzold RF, Outcalt SI. 1998. Spatial and temporal patterns of soil moisture and thaw depth at Barrow, Alaska U.S.A. In *Proceedings of the Seventh International Conference on Permafrost*. Centre d'Etudes Nordiques, Université Laval: Québec; 731–737.

Murton JB, French HM. 1994. Cryostructures in permafrost, Tuktoyaktuk coastlands, western arctic Canada. *Canadian Journal of Earth Sciences* **31**: 737–747.

Nelson FE, Shiklomanov NI, Mueller G, Hinkel KM, Walker DA, Bockheim JG. 1997. Estimating active-layer thickness over a large region: Kuparuk River basin, Alaska, U.S.A. *Arctic and Alpine Research* **29**: 367–378.

Nelson FE, Hinkel KM, Shiklomanov NI, Mueller GR, Miller LL, Walker DA. 1998a. Active-layer thickness in north-central Alaska: systematic sampling, scale, and spatial autocorrelation. *Journal of Geophysical Research* **103**: 28963–28973.

Nelson FE, Outcalt SI, Brown J, Shiklomanov NI, Hinkel KM. 1998b. Spatial and temporal attributes of the active-layer thickness record, Barrow, Alaska U.S.A. In *Proceedings of the Seventh International Conference on Permafrost*. Centre d'Etudes Nordiques, Université Laval: Québec; 797–802.

Pavlov AV. 1998. Active layer monitoring in northern West Siberia. In *Proceedings of the Seventh International Conference on Permafrost*. Centre d'Etudes Nordique, Université Laval: Québec; 875–881.

Pollard W, French HM. 1980. A first approximation of the volume of ground ice, Richards Island, Pleistocene Mackenzie Delta, Northwest Territories, Canada. *Canadian Geotechnical Journal* **17**: 509–516.

Popov AI. 1967. *Frost Features in the Earth's Crust (Cryolithology)*. Moscow State University Press: Moscow (in Russian).

Sellmann PV, Brown J, Lewellen RI, McKim H, Merry C. 1975. The classification and geomorphic implications of thaw lakes on the arctic coastal plain, Alaska. U.S. Army CRREL Research Report; 21 pp.

Shiklomanov NI, Nelson FE. 2002. Active-layer mapping at regional scales: a 13-year spatial time series for the Kuparuk region, north-central Alaska. *Permafrost and Periglacial Processes* **13**: 219–230. DOI:10.1002/ppp.425

Shumskii PA. 1959. Ground ice. In *Fundamentals of Geocryology. Part I: General Geocryology*, Shvetsov PF, Dostovalov BN (eds). Akademiia Nauk SSSR: Moscow; 274–327 (available in English as National Research Council of Canada Technical Translation **1130**: 1964, 118 pp).

Shur YL. 1974. Causes of thermokarst. In *Physical-Geological Processes in Freezing and Thawing Soils*, Chistotinov LV (ed.). USSR Institute of Hydrogeology and Engineering Geology: Moscow; 31–46 (in Russian).

Shur YL. 1975. On the transient layer. In *Methods of Geocryological Studies*, Shvetsov PF, Chistotinov LV (eds). USSR Institute of Hydrogeology and Engineering Geology: Moscow; 82–85 (in Russian).

Shur YL. 1977. *Thermokarst*. Nedra Publishing House: Moscow (in Russian).

Shur YL. 1982. Intermediate layer of syngenetic permafrost. In *Formation of Soil Cryogenic Structure*, Zhestkova TN (ed.). Nauka, Moscow; 170–172 (in Russian).

Shur YL. 1988a. *Upper Horizon of Permafrost and Thermokarst*. Nauka, Moscow (in Russian).

Shur YL. 1988b. The upper horizon of permafrost soils. In *Proceedings of the Fifth International Conference on Permafrost*, Vol. 1. Tapir Publishers, Trondheim: Norway; 867–871.

Shur YL, Jorgenson MT. 1998. Cryostructure development on the floodplain of the Colville River Delta, northern Alaska. In *Proceedings of Seventh International Conference on Permafrost*. Centre d'Etudes Nordiques, Université Laval: Québec; 993–1000.

Shur Y, Jorgenson MT, Pullman ER. 2003. Widespread degradation of ice wedges on the Arctic Coastal Plain in northern Alaska in response to the recent warmer climate. *EOS, Transactions of the American Geophysical Union, Fall Meeting Supplement,* Abstract C11A-05, **84**(46): F347.

Shvetsov PF. 1963. Short-term changes in soil temperature in the Subarctic and their engineering significance. In *Presentations to the International Permafrost Conference*. Akademiia Nauk SSSR: Moscow; 245–252 (in Russian).

Skriabin PN, Varlamov SP, Skachkov YB. 1998. *Interannual Variations in the Thermal Regime of Soils in the Yakutsk Region*. Russian Academy of Sciences: Novisibirsk (in Russian).

Smith MW. 1988. The significance of climatic change for the permafrost environment. In *Proceedings of the Fifty International Conference on Permafrost*. Tapir Publishers, Trondheim: Norway; 18–23.

Sochava VV. 1974. Geotopology as a part of geosystem study. In *Topological Aspects of Geosystem Studies*, Sochava VV (ed.). Nauka, Novosibirsk; 3–36 (in Russian).

Sumgin MI, Kachurin SP, Tolstikhin NI, Tumel' VF. 1940. *Obshchee merzlotovedenie (General Permafrostology)*. Akademiia Nauk SSSR: Leningrad-Moscow 240 pp.

Tsitovich NA. 1932. Permafrost studies in the lower Enisey River. In *Reports of the Commission for Permafrost Investigations*, Vol. 1. Akademiia Nauk SSSR: Leningrad; 89–100 (in Russian).

USSR Climate Data Handbook. 1966. Issue 24, part 2. Hydrometeoizdat, Leningrad; 398 pp.

van Evergingen RO. (ed.). 1998. *Multi-Language Glossary of Permafrost and Related Ground-Ice Terms*. International Permafrost Association.

Vasil'ev IS. 1981. *Patterns of Seasonal Thawing of Soils in West Yakutia*. Nauka, Novosibirsk (in Russian).

Vasil'ev IS. 1987. Variability of the thickness of seasonally thawed and seasonally frozen soil on the Lena-Amga Watershed. In *Environmental Conditions of Developing Regions in Siberia*. Permafrost Institute of the Akademiia Nauk SSSR: Yakutsk; 66–76 (in Russian).

Washburn AL. 1980. *Geocryology: A Survey of Periglacial Processes and Environments*. Halsted Press: New York; 406 pp.

Yanovsky VK. 1933. Expedition to Pechora River to determine the position of the southern permafrost boundary. In *Reports of the Committee for Permafrost Investigations*, Vol. 5(2). Akademiia Nauk SSSR: Moscow-Leningrad; 65–149 (in Russian).

Zhestkova TN, Shur YL. 1982. Cryogenic structures of syngenetic permafrost. In *Formation of Soil Cryogenic Structures*. Nauka, Moscow; 148–168 (in Russian).

