Quaternary Research 75 (2011) 356–365

Contents lists available at ScienceDirect

Quaternary Research

journal homepage: www.elsevier.com/locate/yqres

Climate and material controls on periglacial soil processes: Toward improving periglacial climate indicators

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ARTICLE INFO

Article history:
Received 12 November 2008
Available online 18 February 2011

Keywords:
Periglacial
Permafrost
Paleoclimate
Solifluction
Thermal contraction cracking
Ice wedge

ABSTRACT

One of the distinguished efforts of A.L. Washburn was to reconstruct mean annual air temperature using periglacial features as climate indicators. This paper reviews existing periglacial indicators and proposes a strategy to improve their thermal resolution based on recent periglacial process studies, with a focus on solifluction and thermal contraction cracking and associated landforms/structures. Landforms resulting from solifluction reflect both the depth subjected to freeze–thaw and the thickness of frost-susceptible soils. The thickness of a solifluction structure can be used to infer the dominant freeze–thaw regime and minimum seasonal frost depth. Ice-wedge pseudomorphs have limited potential as a climate indicator because (1) they mainly reflect extreme winter temperatures, (2) their thermal thresholds depend on the host material, and (3) they need to be distinguished from frost wedges of other origin produced under different thermal and/or material conditions. Monitoring studies of currently active ice wedges suggest that ice-wedge cracking requires a combination of low temperature and large temperature gradients in the frozen active layer. Further field monitoring of periglacial processes and their controlling factors under various climate conditions and in various materials are needed, however, to improve the resolution of periglacial paleoclimate indicators.

Background

Conventional indicators of periglacial climate

Periglacial features, including landforms and subsurface structures produced by freeze–thaw or frozen ground processes, develop in response to suitable regional or local climatic conditions. Based on such a climatic dependence, Washburn (1980) defined the warm limits of diverse periglacial features in terms of the threshold values of the mean annual air temperature (MAAT) (Table 1) and reconstructed the global geography of the MAAT during the last glacial maximum (LGM). Features delimited by MAATs below −2° to −1 °C are generally associated with permafrost, but they generally contain little information limiting how low paleotemperatures were (i.e., lowest threshold values). Expressed by a simple parameter, these upper temperature threshold values (Table 1) have widely been applied to Quaternary paleoclimate reconstruction (e.g., Ballantyne and Harris, 1994; Huijzer and Isarin, 1997; Renssen and Vandenberghe, 2003). Moreover, the morphology (or dimension) of some periglacial features is also associated with climate (e.g., Romanovskij, 1973; Ballantyne and Harris, 1994; Grab, 2002). A combination of several periglacial features, often coupled with other paleoclimate (e.g., biological) indicators, provides more precise constraints on the climate while the features were active (e.g., Washburn, 1980; Ballantyne and Harris, 1994).

Subsequent to Washburn's work, the threshold thermal conditions indicated by each feature has been modified to reflect the extent as well as the core of the former distribution of the features, and they have been augmented by using annual precipitation (e.g., Harris, 1982; Haeberli, 1983; Karte, 1983; Harris, 1994). The freezing or thawing indices, computed from year-round air temperature records, or the mean air temperatures of the coldest month (MATCM) and the warmest month (MATWM) have also been used to define the distribution of periglacial features, because a number of features reflect extreme seasonal climatic conditions, e.g. the winter severity and/or the active layer thickness in summer, rather than the mean annual conditions (e.g., Harris, 1982; Karte, 1983; Burn, 1990; Huijzer and Isarin, 1997).

The threshold values of MAAT, however, are mostly empirical and lack precise justification (cf. Murton and Kolstrup, 2003). The present-day standard of Quaternary paleoclimate research, which has progressed with high-resolution cryostratigraphy, dating techniques, and climate models (e.g., Kasse et al., 2003; Murton et al., 2003; Van Huissteden et al., 2003; French et al., 2007; French and Shur, 2010), requires more precise, highly resolved indicators. Moreover, recent studies have highlighted the role of permafrost degradation in modifying periglacial features developed during past cold periods (e.g., Murton, 2007; French, 2008). Such a situation calls for

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doi:10.1016/j.yqres.2010.12.014

Table 1
Examples of periglacial climate indicators, defined by air temperatures.

	Warm limits by MAAT (°C)					Warm limits by MATCM (°C)	
	Péwé (1966)	Washburn (1980)	Karte (1983)	Ballantyne and Harris (1994)	Huijzer and Isarin (1997)	Karte (1983)	Van Huissteden et al. (2003)
Earth hummocks			+3				
Small periglacial involutions (amplitude<0.6 m)					−1		
Large periglacial involutions (amplitude≥0.6 m)					−8[a], −4[b]		
Small sorted patterned ground (diameter<1 m)			+4				
Large sorted patterned ground (diameter≥1 m)		0	−4	−2 to 0			
Rock glaciers		0	0 to +2	−2 to −1			
Organic palsas		0	−3 to 0				
Thermokarst depressions		−2 to 0	−1	−8 to −6			
Ice wedge polygons	−8 to −6	−5	−8 to −4	−6[a], −3[b]	−8[a], −4[b]	−20	−15
Soil wedge polygons			−4 to 0	−1[a], +1[b]	−1	−8	
Open-system pingos		−2	−1	−5 to −4	−4		
Closed-system pingos		−6	−5		−6		

MAAT: Mean annual air temperature. MATCM: Mean air temperature of the coldest month.
[a] Value for fine (clayey/silty) sediment.
[b] Value for coarse (sand/gravel) sediment.

'periglacial climate indicators' based on geological process and physics rather than those given only by simple empirical parameters.

Recent advances in geocryology

Periglacial process studies have advanced significantly in the last three decades in large part because of field monitoring, laboratory simulations, and numerical modeling. Automated techniques in field monitoring allow us to detect the timing and magnitude of ground movements and to relate these events to thermal and hydrological parameters (e.g., Hallet, 1998; Matsuoka, 2001a; Jaesche et al., 2003; Matsuoka, 2005; Kinnard and Lewkowicz, 2005; Harris et al., 2008a; Ikeda et al., 2008). Laboratory simulations can provide quantitative relationships between physical processes and parameters (e.g., Matsuoka, 1990; Harris et al., 1993, 2008b), as well as reproduce features difficult to monitor in the field, such as bedrock heave (Murton et al., 2001), frost weathering (Hallet et al., 1991), ice-wedge castings (Harris and Murton, 2005), and periglacial involutions (Ogino and Matsuoka, 2007). Numerical modeling provides insights into, and help in interpreting field data on, frost weathering (Walder and Hallet, 1985), rockfalls (Gruber et al., 2004), thermal contraction cracking (Lachenbruch, 1962), and frost sorting (Hallet and Waddington, 1992), as well as for predicting landform evolution (e.g., Werner and Hallet, 1993; Anderson, 2002; Kessler and Werner, 2003; Hales and Roering, 2007). These advances in periglacial process studies, and especially those addressing solifluction and thermal contraction cracking, permit us to revisit the corresponding periglacial climate indicators, using more realistic and precise criteria to gain greater resolution.

The distribution and morphology of periglacial features directly reflect the type and rate of periglacial processes, which depend on a number of factors including topography, geology, climate, hydrology, and vegetation (e.g., Åkerman, 1996; Hjort et al., 2007; Ridefelt et al., 2010). Thus, the simple link between air temperature and the occurrence of particular periglacial features is confounded by other parameters. Among these, the ground thermal regime and soil physical characteristics are most important when evaluating indicators of the regional climate. The ground materials not only control the type and rate of geomorphic processes but also affect the ground thermal regime. Other parameters tend to constrain the local distribution rather than the regional occurrence of features. For instance, topography (e.g., inclination of slopes), hydrology (e.g., soil moisture and groundwater level) and vegetation cover can vary substantially on a single mountain or within a drainage basin.

The purpose of this paper is (1) to demonstrate how the thermal regime and physical properties of the ground control periglacial processes and their products (landforms and sediments) and (2) to propose a strategy to improve the climate proxy potential of features produced by solifluction and thermal contraction cracking, which have often been used in paleoclimate reconstruction.

Solifluction

Review

Solifluction operates on slopes in a wide climatic range where moist fine-grained soil recurrently freezes and thaws regardless of the presence of permafrost (e.g., Washburn, 1967; Benedict, 1976; Ballantyne and Harris, 1994). Solifluction-derived landforms, including lobes, sheets, and stripes, are found in regions with MAAT from −20 °C to 7 °C (e.g., Washburn, 1979; Harris, 1981; Matsuoka, 2001b). In this respect, solifluction-derived forms or deposits, which are identified by morphology, grain size and/or macro- or microfabrics (e.g., Harris, 1981; Millar, 2006), indicate only the occurrence of freeze–thaw cycles. Using the data base from currently active sites, Matthews and Berrisford (1993) defined thresholds (warmer limits) for the initiation of solifluction in terms of the duration of snow cover and winter and/or summer mean temperature, but the thresholds differ significantly between alpine and polar regions. Thus, relict solifluction features may only imply effective winter freezing in the past, particularly where the ground rarely freezes today (e.g., Ballantyne and Harris, 1994). Careful analysis of the morphology or internal structure, however, may provide more detailed climatic information, because the magnitude and process of solifluction partly reflect the near-surface climate, affected by thermal, hydrological and biological characteristics of the slope (Matsuoka, 2001b).

Miniature slope forms, such as small sorted stripes with spacing of 10–30 cm and low stone-banked lobes (or sheets) with a front as high as 20 cm, prevail in high tropical mountains (e.g., Andes) where diurnal frost cycles occur throughout the year because of the lack of seasonality (Francou et al., 2001). Similar forms are widespread also in subpolar islands (Boelhouwers et al., 2003), mid-latitude high mountains (Matsuoka, 2005), and marginal periglacial mountains (Vieira et al., 2003), all experiencing frequent diurnal frost. These forms are considered to reflect rapid but shallow soil movements associated with diurnal frost (or needle-ice) creep (Matsuoka, 2001b). The advance of low stone-banked lobes often produces stratified slope deposits composed of alternating layers of clasts and fine debris, each of which is ~10 cm thick, which result from frontal clasts overridden by fine debris (Francou, 1990). Such a thin stratified structure (grèzes litées) in stratigraphy may represent a past environment dominated by diurnal frost and barren ground (Bertran et al., 1995), but detailed sedimentological analysis, for example, on the form of sedimentary units, grain-size grading and fabric properties, is required to distinguish from stratified structures induced by other scree processes unrelated to freezing (Van Steijn et al., 1995).

Millennial-scale solifluction activity reflects climate change during the Holocene. Long-term change in the rate of solifluction can be estimated using radiocarbon data from organic layers buried in advancing solifluction lobes (e.g., Smith, 1993; Matthews et al., 2005). In volcanic regions, advances of solifluction lobes concurrent with tephra falls also provide similar data (Hirakawa, 1989; Kirkbride and Dugmore, 2005). Most of these studies suggest several phases of acceleration or deceleration, though the temporal resolution is inherently low and the results are limited by the survival time of buried vegetation. Moreover, the radiocarbon dates can be compromised by contamination of modern vegetation (e.g., Matthews, 1993). Long-term solifluction activity reflects, in addition to atmospheric temperature, changes in precipitation, snow and vegetation conditions that may affect ground temperature and moisture (e.g., Oliva et al., 2009).

Prospects

Recent field monitoring and laboratory simulations have provided quantitative relationships between the rate of solifluction and controlling factors (e.g., Harris et al., 1993; Matsuoka, 1998; Jaesche et al., 2002; Matsuoka, 2010), as well as revealed detailed mechanisms of solifluction (e.g., Kinnard and Lewkowicz, 2005; Harris et al., 2008a, 2008b). Where diurnal frost action prevails, rapid but shallow frost (needle-ice) creep dominates soil displacements (Fig. 1) and produces miniature stripes and low stone-banked lobes (sheets). Where the slope is subject mainly to seasonal frost heave and thaw settlement, slower but deeper soil motion that is more uniform with depth promotes the advance of solifluction lobes with a frontal height of several decimeters (e.g., Smith, 1992; Matsuoka, 2001b; Jaesche et al., 2003; Kinnard and Lewkowicz, 2005; Matsuoka, 2010; Fig. 2). Thus, the predominance of diurnal or annual frost cycles can be evaluated by (1) the height of a solifluction lobe, for both stone-banked and turf-banked lobes, and (2) the thickness of a solifluction deposit underlain by a stone horizon or an organic layer which was overridden by the solifluction deposit. In general, lobes or sheets thinner than 20 cm, or stratified deposits in which each layer is ~10 cm thick or less, indicate the predominance of diurnal frost action (e.g., Bertran et al., 1995), whereas lobes or sheets 40–100 cm high or deposits ~50 cm thick originate mainly from annual frost action.

Caution is required, however, because surface conditions can vary spatially allowing both forms to coexist within a small area subject to the same regional climate. For example, the presence of a grass mat or seasonal snow on the surface minimizes diurnal frost action, highlighting the effect of annual frost action. The thickness of frost-susceptible (fine-grained) soil also controls the dimensions. Within a soil slope a few hundred meters long and subject to deep seasonal frost (e.g., ~1 m), diurnal frost action tends to prevail on the upper slope with thin soil (e.g., <0.2 m), because seasonal frost heave is limited by the well-drained soil and/or underlying coarse debris or bedrock that resists frost heave (Fig. 1). The thickness of fine debris may increase toward the lower slope, which favors seasonal frost heave and thaw settlement, possibly aided by increasing moisture availability (Figs. 2 and 3). As a result, miniature forms dominate on the top slope whereas larger forms become prominent near the base of the slopes (Matsuoka et al., 1997; cf. Ridefelt and Boelhouwers, 2006). In such a situation, larger forms are diagnostic of the regional climate. It is noteworthy, however, that the depth of solifluction rarely exceeds 50 cm of soil even where the freezing front propagates more deeply in thick (e.g., >1 m), fine debris (Fig. 2). This is because (1) seasonal frost heave is mostly confined within the top 50 cm (Fig. 3), due to desiccation of the subsoil as a result of ice segregation in the

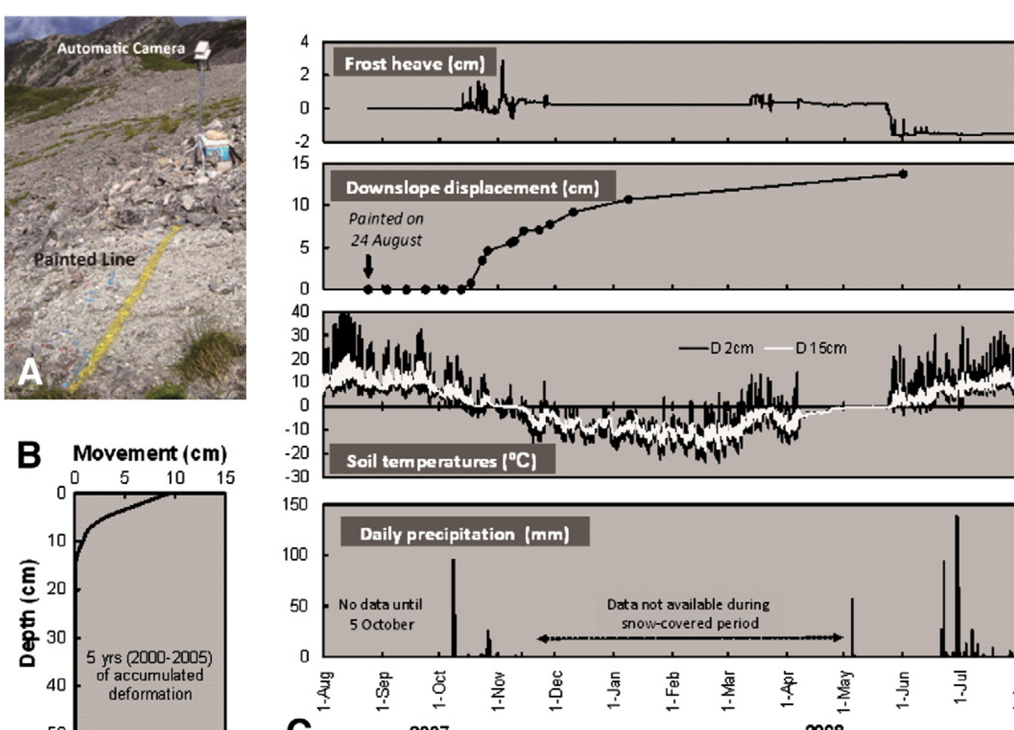

Figure 1. Soil displacement in a small solifluction lobe in the Southern Japanese Alps. (A) Visual monitoring of a line painted on the surface (drawn horizontally over 2.7 m in length) with an automatic camera: downslope to the left. (B) Vertical profile of total soil displacement after five years, showing shallow deformation within the uppermost 15 cm of soil. (C) From top to bottom: vertical surface soil displacement (frost heave and thaw settlement), downslope soil movement (based on data from the automatic camera), shallow soil temperatures (at 2 and 15 cm depths), and daily rainfall amount. Because the surface layer of fine debris is thin (20 cm) diurnal frost heave and creep dominate solifluction at this site despite the deep seasonal frost (>100 cm).

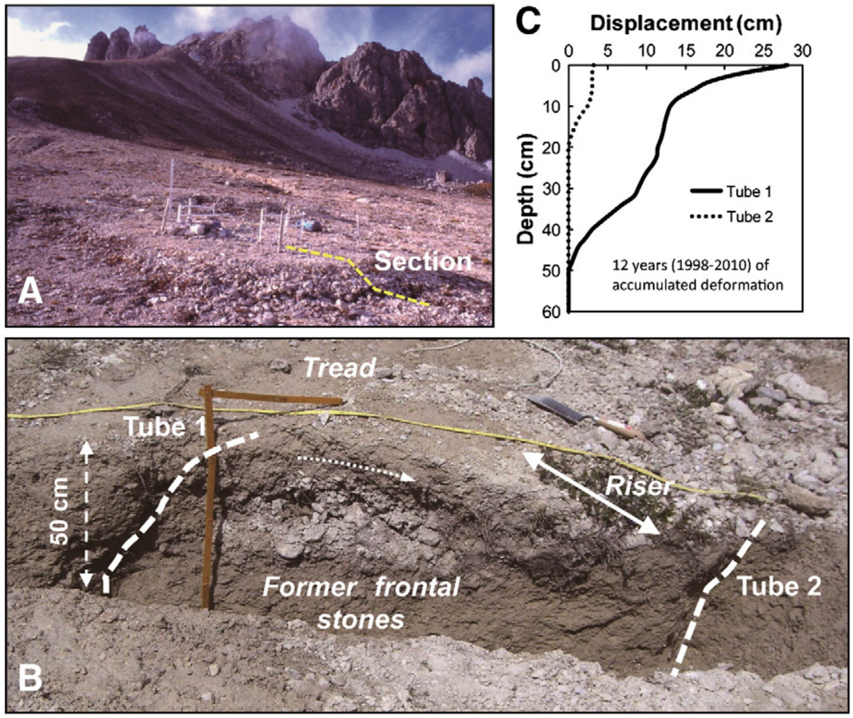

Figure 2. Soil structure and deformation in the frontal part of a stone-banked lobe (Lobe L in Matsuoka, 2010), Trais Fluors, the Swiss Alps. (A) Lobe L (tread width 5 m, length 5 m, riser height 0.4 m, and slope gradient 7°) in October 1997. (B) Excavated frontal section in July 2010, showing the locations of two tubes: Tube 1 on the tread and Tube 2 in front of the riser. (C) Subsurface soil deformation after 12 years. Tube 1 displays deeper and much faster movement than Tube 2, with the upper concave and lower convex downslope profiles: solifluction of the deeper soil appears to have been constrained by the former frontal stone aggregate, whereas the shallower soil seems to have moved beyond the former front.

upper layer (Matsuoka, 2001b), and/or (2) the void ratio decreases while shear strength increases in the lower layer (Harris and Davies, 2000; Matsumoto, 2010).

The growing evidence for two-sided freezing in permafrost regions marked by particularly cold winters (e.g., Mackay, 1981; Cheng, 1983; Shiklomanov and Nelson, 2007) may further allow identification of 'cold permafrost' using surface forms or subsurface structures. Where permafrost is absent or warm (~0 °C), seasonal frost penetrates only from the surface downward. The resulting one-sided freezing induces 'normal' solifluction in which the top ~50 cm of soil moves downslope and produces a lobe with a front also about 50 cm high or slightly higher. The lack of a shear plane at the base of movement promotes overturning of soil at the front (Fig. 4), although episodic lobe advance in places produces a front lacking overturned structure and having small shear planes (Kinnard and Lewkowicz, 2006). The overturned structure, typically observed in excavated turf-banked lobes (e.g., Hirakawa, 1989; Matthews et al., 2005), may reflect the predominance of one-sided freezing. Stone-banked lobes usually lack a distinct overturned structure fringed by an organic layer, but buried stone aggregates reflect the dimension of the former stone-banked front (Fig. 2B).

Where the active layer freezes both from the surface downward and from the permafrost table upward (i.e., two-sided freezing), the latter produces an ice-rich layer near the permafrost table. Thawing of the ice-rich layer often induces 'plug-like flow' in which the top 50–150 cm of soil move downslope en masse (e.g., Mackay, 1981; Lewkowicz and Clarke, 1998; Matsuoka and Hirakawa, 2000; Harris et al., 2008c). Such a deep movement with localized shear near the base may result in a thicker deposit, a higher front, and a subsurface structure free of overturning; field evidence proving this is still lacking, however. Moreover, depending on the spatial variation in the slope gradient or clay and water content of the soil, active-layer detachment slides are often triggered by the thawing of the ice-rich basal active layer. Their frontal structure is characteristically composed of deformed soil with a number of thrust planes and a well-defined shear plane at the base (e.g., Harris and Lewkowicz, 1993; Matsuoka and Hirakawa, 2000; Lewkowicz and Harris, 2005). These forms and/or structures reflecting two-sided freezing would be suggestive of the (former) presence of cold permafrost with mean annual temperature below about −5 °C (e.g., Lewkowicz, 1988; Shiklomanov and Nelson, 2007).

Thermal contraction cracking

Review

Ice wedges (Fig. 5A) and ice wedge casts (pseudomorphs) have long been used as the best indicator of continuous (or cold) permafrost and MAAT below −5 °C to −8 °C (e.g., Péwé, 1966; Washburn, 1980). The advantages of using these features for paleotemperature reconstructions include: (1) they are ubiquitous in permafrost terrain; and (2) they tend to form distinct surface patterns and structures in the vertical exposures that persist long even after the permafrost degrades.

The early MAAT criteria and related work need updating, however, in view of the results of more recent studies, including: (1) the disparity between air and ground temperatures; (2) the ground thermal condition at which thermal contraction cracking takes place; (3) the type of ground materials; and (4) the distinction from other frost wedges (active-layer soil wedges and sand wedges) that may develop under different climatic conditions. With respect to the first point, thermal insulators on the ground surface, such as snow and vegetation cover, enlarge the difference between the air and ground temperatures (Williams and Smith, 1989). For instance, a thick snow cover (>60 cm) can prevent rapid cooling of the ground and thermal contraction cracking even where MAAT is below −8 °C (Mackay, 1993). A field experiment at an Arctic drained lake site shows gradual inactivation of thermal contraction cracking with vegetation growth and associated snow entrapment (Mackay and Burn, 2002).

https://doi.org/10.1016/j.yqres.2010.12.014 Published online by Cambridge University Press

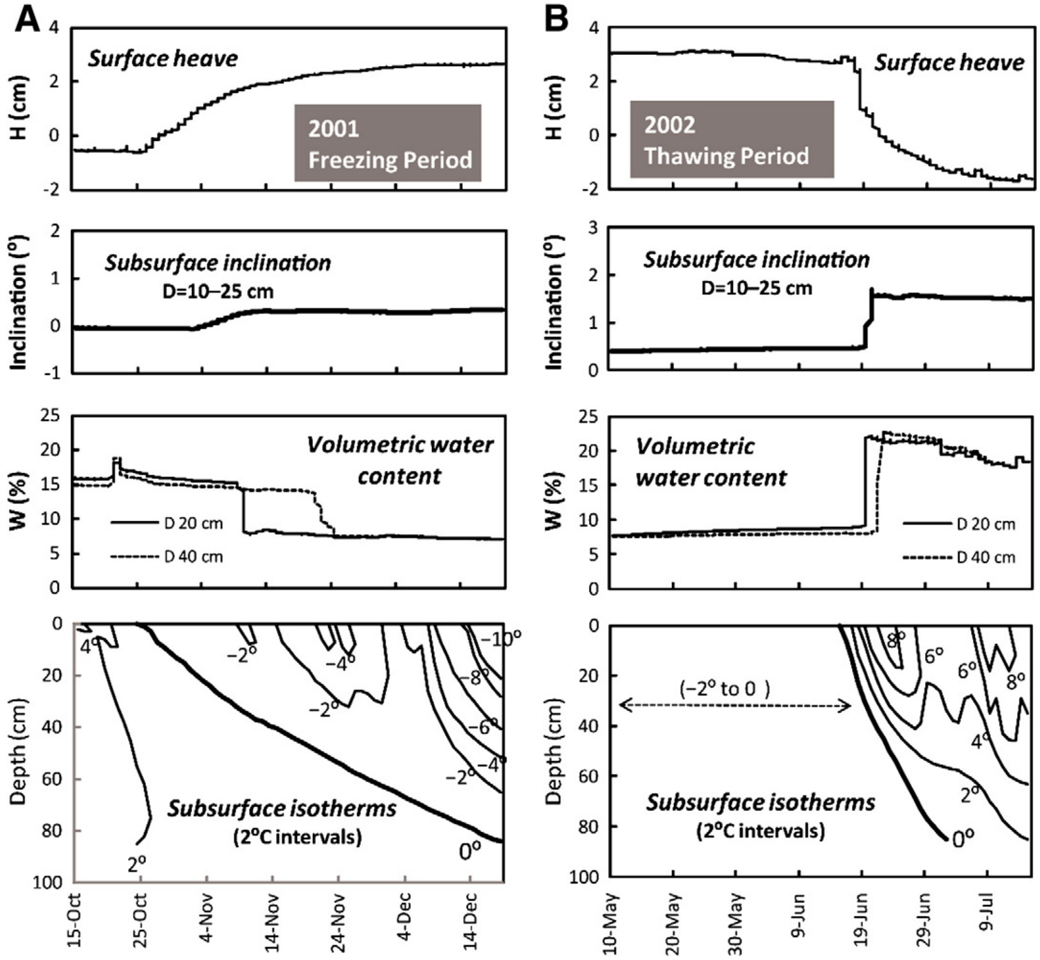

Figure 3. Soil movements of Lobe L (shown in Fig. 2) during the seasonal freezing period of 2001 (A) and the subsequent thawing period (B): five months of data are skipped between the two figures. From top to bottom: surface heave and settlement (H) on the order of 3–4 cm; tilting of a rigid inclinometer (installed between 10 – 25 cm depth), which only responds to large movement (positive values indicate downslope tilting relative to the position on August 8, 2001); soil moisture (W) at 20 and 40 cm depths, indicating relatively dry condition (~15% vol.) with minor wetting before freezing, sudden decrease in unfrozen water content due to freezing, and significant wetting (>20% vol.) by melting of surface snow and near-surface ice lenses during thawing; and subsurface isotherms at 2 °C intervals. Frost heave (ice lens growth) mostly takes place in the upper 50 cm, whereas thaw settlement rapidly progresses during thawing of the heaved soil but continues for an additional 2–3 weeks probably due to delayed seepage of meltwater. Note that an apparent tilting took place during frost heave (A) due to the slight difference in the angles between soil heave and the inclinometer, whereas rapid tilting during thawing (B) resulted mostly from gelifluction (Matsuoka, 2010).

Even where the insulating effect by snow cover is minimal, ice-wedge cracking essentially requires intensive cooling during a short period of the winter, rather than simply low mean annual or mean winter temperatures (e.g., Black, 1976; Mackay, 1986, 1992). Long-term, extensive observations by Mackay (2000) and recent automated observations (Allard and Kasper, 1998; Matsuoka, 1999; Fortier and Allard, 2005) have defined the ground thermal conditions at which cracking takes place rather precisely; these conditions, which control the magnitude of tensile stresses (e.g., Lachenbruch, 1962), can be expressed as the rate of cooling, thermal gradient, temperature of the top of permafrost (usually designated as TTOP), or their combination. Although available data are limited to a few climate and ground conditions, the results agree in suggesting a critical condition for cracking in polygon troughs: significant cooling must be sufficiently rapid to cause the mean temperature gradient in the frozen active layer, $G_{act} = (T_{gs} - T_{top})/D_{act}$, to reach about -10 °C m^{-1} or steeper in silty soils; here T_{gs} is the ground surface temperature, T_{top} is the temperature of the top of permafrost, and D_{act} is the active layer thickness (Fortier and Allard, 2005; Matsuoka and Christiansen, 2008; cf. Mackay, 1986). However, G_{act} is unlikely to be a universal and sufficient constraint. In fact, visco-elastic models suggest that rapid cooling intensifies thermal stresses in combination with lower near-surface temperatures, as represented by T_{top}, since viscous relaxation of stresses increases in warmer frozen soils (Lachenbruch, 1962; Mellon, 1997). In Svalbard, for example, cracking activity is insignificant in the early winter when the ground surface often experiences rapid cooling but T_{top} remains only a few degrees below 0 °C, but it becomes significant in the late winter when rapid cooling is associated with T_{top} below -10 °C (Christiansen et al., 2010). The threshold T_{top}

Figure 4. Frontal section of a 0.6-m-thick turf-banked solifluction lobe subject to one-sided (downward) freezing in Adventdalen, Svalbard, displaying the overturned vegetation (humus) and loam sandwiching a granular layer.

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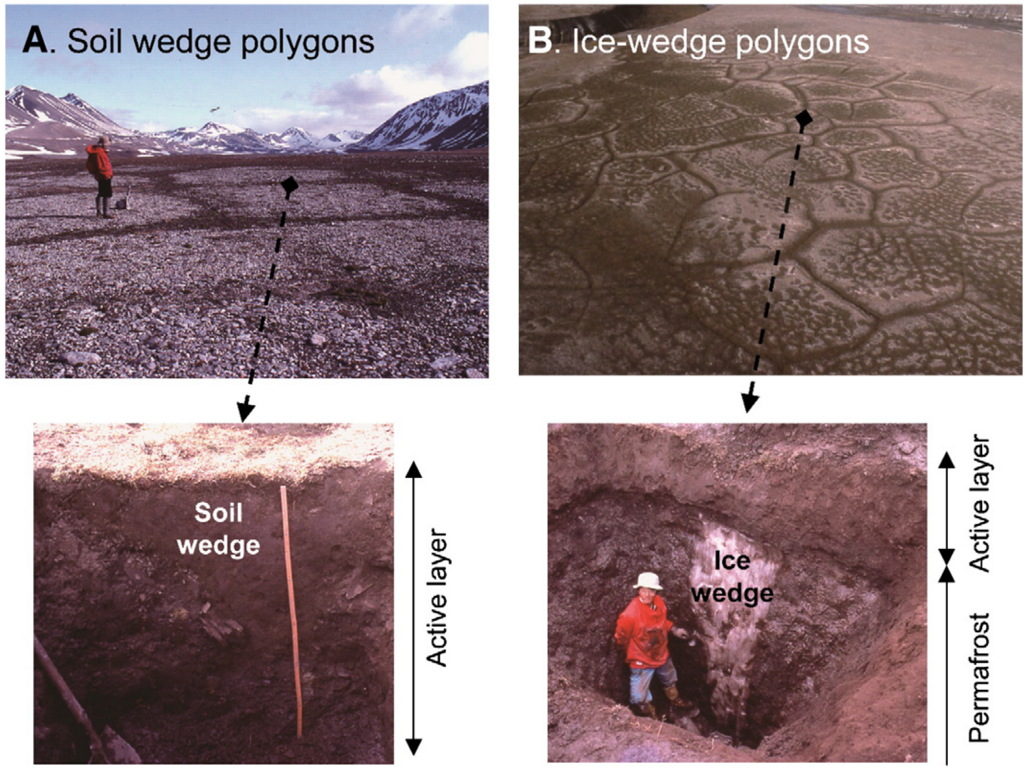

Figure 5. Two types of frost wedges. (A) Active-layer soil-wedge polygons in Kapp Linné (a coastal site), Svalbard: spacing of troughs 5–10 m, and depth of the soil wedge 1.1 m. (B) Ice-wedge polygons in Adventdalen (an inland site), Svalbard: spacing of troughs 10–30 m; ice-wedge height 4 m; and active layer depth 0.6 m. Relative to the coastal site, the latter site is considerably colder in the winter: the mean surface temperature was −7.0 °C at the coastal site and −10.7 °C at the inland site from November 1, 2007 to April 30, 2008. The arrows indicates the locations of excavated troughs.

is dependent on D_{act} and so far only roughly constrained, ranging from −15 °C (Fortier and Allard, 2005) to −10 °C (Matsuoka and Christiansen, 2008) in accordance with increasing D_{act} from 0.6 to 1.0 m.

The ground material is significant in estimating (paleo-)temperatures. In frozen ground, fine-grained (clayey or silty) soils favor deeper cracks that penetrate into the permafrost than coarse-grained (sandy or gravelly) soils. This is because the former have (1) larger coefficients of thermal contraction that contributes to larger horizontal tensile stress (e.g., Romanovskij, 1973; Mackay, 2000), and (2) smaller thermal conductivities favoring larger G_{act} (e.g., Williams and Smith, 1989) and, hence, larger tensile stresses. Reflecting such a contrast, fine soils can produce ice wedges in warmer permafrost areas (higher MAATs) than coarse soils (e.g., Romanovskij, 1973). Near the warmer margin of the frost wedge occurrence, however, coarse soils may be more favorable for thermal contraction cracking than fine soils, but producing only shallow soil wedges (see below). This is because, reflecting their unfrozen water content, wet fine soils start to contract with decreasing temperature only below −5 °C; above that temperature, cooling below 0 °C causes expansion due to ice growth. In contrast, coarse soils start to contract below −0.5 °C (Williams and Smith, 1989). In addition, the presence of peaty topsoils lead to thermal contraction cracking at higher MAATs (Hamilton et al., 1983), because the low thermal conductivity of dry unfrozen peat prevents warming of substrates in summer while a high thermal conductivity of frozen peat promotes ground cooling in winter (Williams and Smith, 1989).

Regarding the different types of wedges, ice-wedge casts indicative of past cold permafrost have to be distinguished carefully from other ice-free wedges. For instance, active-layer soil wedges, which develop from shallow cracks confined in the active layer (Fig. 5B), do not require the presence of permafrost; they can occur under marginal permafrost or even in seasonal frost environments in coarse-grained soils (e.g., Romanovskij, 1973). Sand wedges, which result from eolian infilling, typically develop in cold (and arid/windy) permafrost (Péwé, 1959; Bockheim et al., 2009) but can also occur in warmer environments (e.g., Romanovskij, 1973; Murton et al., 2000). When permafrost thaws, active-layer soil wedges and sand wedges mostly preserve the primary infilling structure (e.g., vertical foliation), whereas ice wedge casts reflect thaw transformation that often results in infilling material resembling the host sediment and showing deformation and slumping (Romanovskij, 1973; Harry and Gozdzik, 1988; Ballantyne and Harris, 1994). These characteristics permit us to identify the wedge types and define the conditions that prevailed, but more detailed structural criteria are required (e.g., Harry and Gozdzik, 1988; Murton and French, 1993; French and Shur, 2010).

Prospects

More detailed, comprehensive monitoring of thermal contraction cracking in various climate and ground conditions are needed to improve the ability to extract paleoenvironmental data from formerly active contraction polygons. Monitoring should include crack activity at high temporal and spatial resolutions, as well as snow depth and ground temperature profiles (Fig. 6). A combination of several sensors, including breaking cables (Mackay, 1992; Fortier and Allard, 2005), extensometers, and shock (acceleration) loggers, permit tracking the various aspects of ground motion in detail: the pre-failure ground contraction, timing of crack generation, and subsequent crack propagation (Matsuoka and Christiansen, 2008). The analysis of these will lead to more precise thresholds for ice-wedge cracking.

As noted previously, rapid cooling during mid to late winter favors ice-wedge cracking. Such cooling can occur several times per year where cracking is active (e.g., Allard and Kasper, 1998; Fortier and Allard, 2005), where there is little or no snow cover, and/or where the

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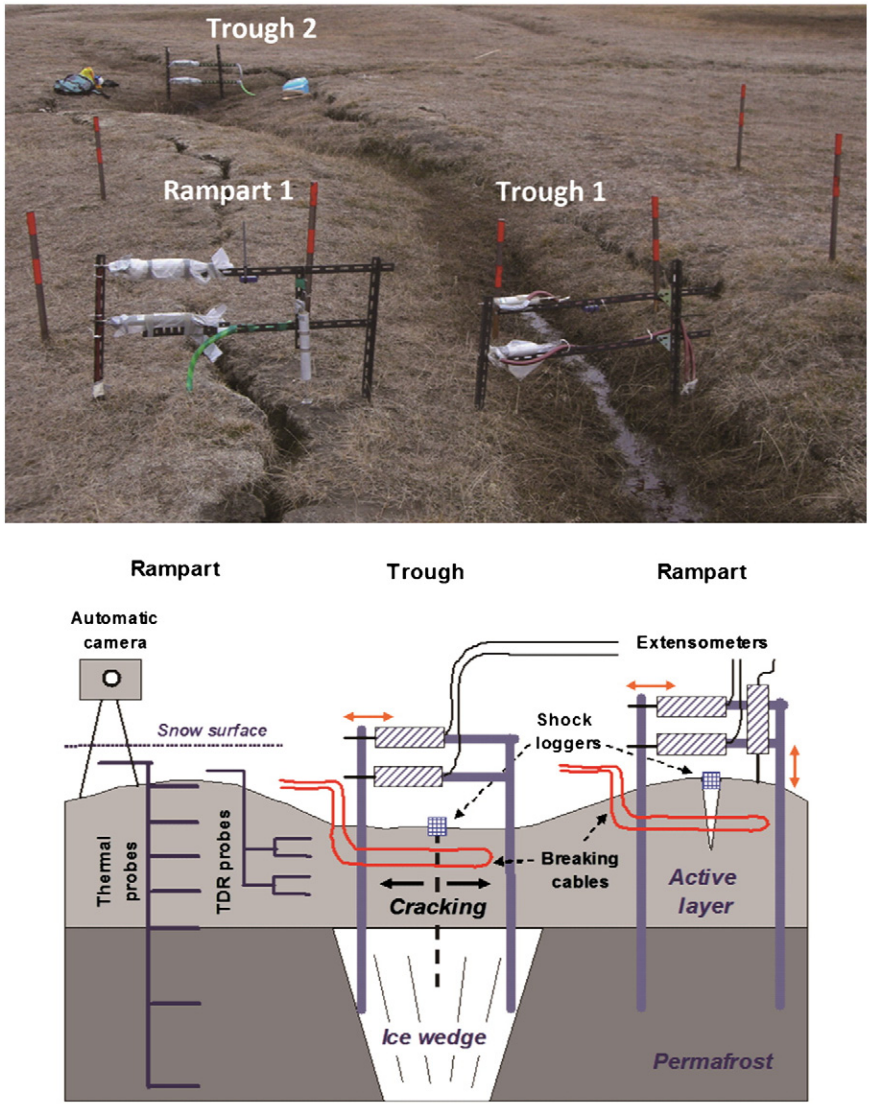

Figure 6. Automated field monitoring of ice-wedge cracking in Adventdalen, Svalbard (photo taken in June 2010). Three types of sensors, extensometers (horizontal and vertical soil displacements), shock loggers (acceleration events), and breaking cables (crack opening), permit us to study ground motion and cracking. Soil temperature and moisture (TDR) are concurrently monitored. Note the inward tilting of a pair of initially vertical benchmarks across the trough 1 and outward tilting across the rampart 1, which mainly reflected long-term deformation of the active layer.

atmospheric temperature is highly variable. For instance, extensometers and shock loggers indicate several events in a polygon trough corresponding to cold phases during a winter, most of which are likely to reflect minor cracking events toward a major failure (Matsuoka, 1999; Matsuoka and Christiansen, 2008). Continental climate with extremely low minimum temperatures are also favors cracking but is not a prerequisite. In fact, well-developed polygons are currently active in Svalbard, as indicated by frequent new cracks in the troughs, despite the minimum surface temperature exceeding −30 °C; the ground there, however, experiences large thermal fluctuations in the winter (Matsuoka, 1999; Christiansen, 2005). Intensive cracking activity takes place even in a relatively warm winter when there is little or no snow and the air temperature drops abruptly (Christiansen et al., 2010; Fig. 7B).

Paleoenvironmental studies have focused on the threshold conditions for ice-wedge cracking and growth, but there is also considerable merit in studying the paleo-environmental information reflected in the dimensions of polygonal patterned ground (e.g., diameter of polygons and height of wedges). The duration and magnitude of significant cooling events control the depth of crack, which in turn determines the horizontal extent of stress relief that controls the spacing of contraction cracks and polygon troughs (Lachenbruch, 1962). Quantitative relationships between the morphological and climatic parameters can be obtained by monitoring of ice-wedge cracking combined with morphological and structural analysis of ice wedge systems. Careful analysis is, however, required for the height of a past ice wedge, where the wedge grew syngenetically (possibly overestimating its height) or the wedge was truncated by active layer deepening or erosion (tending to underestimate its height) (cf. Romanovskij, 1973; French, 2007).

So far, the thermal thresholds are available only for ice-wedge polygons composed of fine-grained sediments. Future monitoring should also target ice-wedge cracking in coarse-grained sediments. It is empirically known that the depth of crack penetration varies with the type of sediment and ground temperature (e.g., Romanovskij, 1973), but this empirical rule is not unequivocally explained in terms of physical parameters. In addition to the thermal and related properties (e.g., the linear coefficient of expansion and the thermal conductivity), rheological properties, such as the effective viscosity that affects the magnitude of both horizontal tensile stresses and the tensile strength, may contribute to thermal contraction cracking (e.g., Lachenbruch, 1962; Romanovskij, 1973; Mellon, 1997; Matsuoka, 1999). Reliable values for these parameters have to be acquired through tests and measurements using field materials (Mackay, 2000).

https://doi.org/10.1016/j.yqres.2010.12.014 Published online by Cambridge University Press

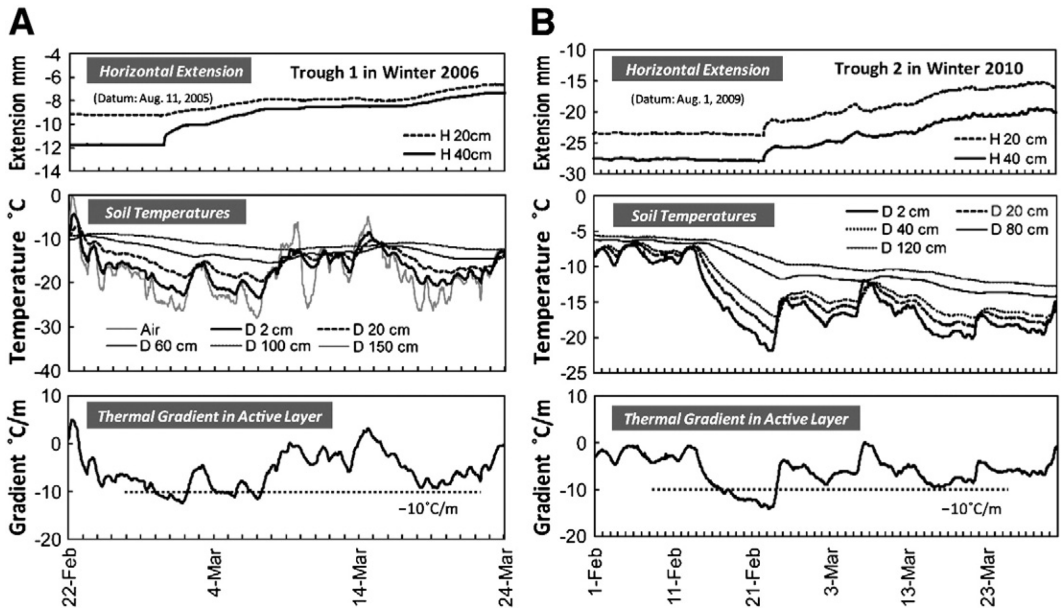

Figure 7. Two examples of extension (increasing signal) across troughs in winter at the Adventdalen site (see Fig. 6). (A) Data from Trough 1 (22 February–24 March, 2006). (B) Data from Trough 2 (1 February to 1 April). Both show that significant trough extension started when the frozen and pre-cooled ($T_{top} < -10$ °C) active layer experienced rapid cooling ($G_{act} < -10$ °C/m). The T_{top} value is represented by temperatures at 100 cm in Trough A and at 80 cm in Trough 2. Data from shock loggers, breaking cables, and crack mapping in April all reflect intense thermal contraction and cracking in the same periods (Christiansen et al., 2010). Details of the event in A are provided by Matsuoka and Christiansen (2008).

Field monitoring should also target the dynamics of sand wedges and active-layer soil wedges. Long-term observations of the growth of sand wedges have been undertaken in Antarctica (Black, 1973; Sletten et al., 2003), but extremely limited data have been published on the thermal conditions for cracking (Sletten and Hallet, 2003) though decade-long, continuous soil displacement data now exists for a number of sites in the Dry Valleys of Antarctica (e.g., Goehring et al., 2008). Essentially no measurement has been undertaken, however, to provide a basis for evaluating thermal thresholds for cracking of active-layer soil wedges. Since both wedge types are reported from diverse climatic zones, monitoring in a wide range of climates will promote quantification of the climatic controls of the dimensions of polygons and wedges.

Finally, thresholds for terrestrial thermal contraction cracking are also applicable, in principle, to the evaluation of permafrost conditions (e.g., temperature, internal structure, and the presence of ice) on the surface of Mars (e.g., Mellon, 1997). Recently, high-resolution orbiter images have revealed diverse types of (possibly active) small-scale polygons (typically 5–10 m in diameter) that are widespread in high latitudes (e.g., Levy et al., 2009) and robotic-arm excavations at the Phoenix landing site have exposed ice-rich permafrost below high-centered polygons (Mellon et al., 2009). The latitudal distribution may reflect overall lower temperature reducing viscous relaxation of tensile stress (Mellon, 1997), or the presence of ice-rich sediments (Levy et al., 2009). In practice, however, the large difference in mean surface temperature between Mars and the Earth would lead to important differences in thresholds for cracking and in polygon dimensions. For example, the lower temperatures on Mars would render the frozen ground more brittle and less deformable than on the Earth, which is responsible for lower stress relaxation and, hence, smaller polygons (Mellon, 1997).

Concluding remarks: towards improving periglacial climate indicators

Recent studies on the two periglacial processes considered herein suggest the following potential and limitations to using the resulting landforms and structures as climate indicators.

Solifluction forms and structures only roughly indicate the presence of freeze–thaw action, but their dimensions can indicate the predominance of either diurnal or annual freeze–thaw activity. Lobes 10–20 cm thick occur where diurnal frost activity is dominant or where fine debris at the surface is thin; on the other hand, features exceeding 50 cm in thickness develop where seasonal frost action dominates regardless of the presence of permafrost. Future studies targeting paleoenvironmental interpretations of the products of two-sided freezing indicative of the presence of 'cold' permafrost (permafrost temperature well below 0 °C or MAAT < -5 °C) have considerable merit.

Ice wedges and their pseudomorphs indicate ground thermal regimes that depend mainly on the combination of air temperatures and ground cover (snow and vegetation) conditions. Forms in fine-grained soils generally indicate cold permafrost, but those in soils overlain by peaty layers may occur in significantly warmer climates; by contrast, those in coarse sediments suggest colder conditions. Ice-wedge cracking requires mainly rapid cooling during mid to late winter, which leads to T_{top} lower than -10 °C and a thermal gradient in the active layer steeper than -10 °C m^{-1}; it depends weakly on the mean annual air/ground temperatures. The diameter of polygons and the height of ice wedges are expected to vary with ground thermal parameters, but further monitoring and analysis are required over a broader range of climate and the soil types.

Further improving periglacial climate indicators and acquiring higher resolution indicators require at least four research activities:

- detailed descriptions of key features (type, structure, morphology, dimension and activity) for various climates and materials;
- expanding monitoring network to wide climatic zones;
- comparing periglacial features in different materials but the same climate; and
- promoting laboratory simulations of processes difficult to monitor in the field.

Acknowledgments

Research was financially supported by the Grants-in-Aid for Scientific Research (No. 20300293) from the Japan Society for the

https://doi.org/10.1016/j.yqres.2010.12.014 Published online by Cambridge University Press

Promotion of Science. I thank helpful comments from Atsushi Ikeda, Bernard Hallet and anonymous reviewers to an early version of the paper.

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