Measurement of dynamic parameters of frozen soil by ultrasonic techniques

D.Y. Wang, Y.L. Zhu, W.W. Peng, S.P. Zhao & H.P. Li
State Key Laboratory of Frozen Soil Engineering, CAREERI, CAS, Lanzhou, Gansu, China

ABSTRACT: Velocities of ultrasonic waves in frozen fine sand, silt (loess) and clay with different water contents were measured at different temperatures. According to the value of wave velocities, dynamic elastic parameters of the frozen soils were calculated in terms of elastic theory. The calculation shows that both the dynamic elastic modulus and shear modulus of the three types of frozen soils increase with decreasing temperature. The Poisson's ratios of the frozen soils tested increase as the temperature increases. Both the dynamic elastic modulus and shear modulus of frozen silt and clay increase with water content increasing in the range of lower water contents; whereas they decrease with increasing water content in the range of higher water contents. The Poisson's ratios of the frozen silt and clay increase with water content increasing over the wide range of water contents tested.

1 INTRODUCTION

The dynamic elastic modulus $E$, dynamic shear modulus $G$ and Poisson's ratio $\mu$ are not only the essential mechanical parameters, which reflect the dynamic properties of frozen soils, but also indispensable mechanical indices for frozen soils foundation designation and other engineering as well. Owing to difficulty with the experimental technique, however, only a few experiments have been carried out to establish the dynamic mechanical properties of frozen soils all over the world.

He et al. (1993) conducted uniaxial compression vibration tests on saturated frozen silt at different temperatures, frequencies and loads, and analyzed the influence of the maximum stress, strain and temperature on the dynamic elastic modulus. Xu et al. (1996) studied the relationship between dynamic stress and dynamic strain using the Material Test (Vibration) System (MTS), and obtained the frozen soils dynamic elastic modulus expression: $E = 1/(a + bs)$ where $a$ and $b$ are the test parameters. He et al. (1999) measured the Poisson's ratio and variation of the volume of the specimen with axial strain of frozen clay, frozen loess and frozen sand at different negative temperatures using a diametrical strain gauge in the MTS Vibration System.

As the ultrasonic technique is one of the most effective non-destructive methods for testing materials, it is widely used in testing the properties of concrete and unfrozen earth materials. However, it is seldom used in testing the properties of frozen soil. Sheng et al. (2000) measured the ultrasonic velocities of both dilatational and shear waves in Tomakomai silt mixed with fragments of rubber tires over the temperature range of $-0.2^\circ C \sim -10^\circ C$ by the Immersion sing-around method, and discussed the relationships between ultrasonic velocities and unfrozen water content. Nakano et al. (1972) measured the ultrasonic velocities of dilatational and shear waves in water-saturated frozen soils as a function of temperature by using both pulse first-arrival and critical angle methods. They found that a close correlation exists between dilatational wave velocities and unfrozen water content.

The observed hysteresis in the velocities of silt and clay during a frozen-thaw cycle was thought to have been caused by hysteresis in the unfrozen water content. The authors also found that in the frozen state, there is a general tendency for shear wave velocity to decrease with ascending temperature, but the effect of temperature on shear wave velocity is not so pronounced as that on dilatational velocity. Nakano & Arnold (1973) measured ultrasonic velocities of dilatational and shear waves as well as damping of dilatational in frozen Ottawa Sand as a function of water content by using the critical angle method at a frequency of 1 MHz. Yang & Li (1997) studied the compressive strength, tensile strength and elastic modulus of frozen sand and frozen clay by testing the ultrasonic velocities of dilatational and shear waves as well as the damping of dilatational waves.

In order to study the dynamic mechanical characteristics of frozen soil further, ultrasonic velocities in the frozen clay, frozen loess and frozen sands were measured by a UVM-2 Ultrasonic Measuring System. Then, the dynamic elastic mechanical parameters of frozen soils were calculated in terms of elastic theory, based on the measured ultrasonic wave velocities, and the effect of temperatures and water content on the dynamic parameters was discussed.

2 TESTING PRINCIPLE

The tests were conducted by using the improved UVM-2 Ultrasonic Measuring System made in Japan. The pulse sing-around method is used to measure the time of sound propagating a fixed distance $l$. There are two transducers. One serves as a transmitter, and another
serves as a receiver. A pulse signal will be transmitted from the transmitter perpendicularly to the receiver, and then it will be reflected. The reflected pulse will reach the transmitter again, which will then emit the second pulse. The pulse signal will be cycled (or recycled) in this way. Measuring the sing-around period $T$ is equivalent to measuring the time of sound propagating in the medium. If the distance $l$ is known, the sound velocity can be calculated by the following formula:

$$C = \frac{2l}{T} \quad (1)$$

However, as the electronic circuit has a certain delayed time $T_0$, the measured period $T$ includes the time of sound going through the frozen sample and the electronic circuit. Therefore, the sound velocity should be calculated by the following equation:

$$C = \frac{2l}{T - T_0} \quad (2)$$

Ultrasonic velocities of the dilatational and shear waves through a frozen soil sample can be obtained from Equation 2. According to the elastic theory, the dynamic elastic modulus $E$, dynamic shear modulus $G$ and Poisson’s ratio $\mu$ can be calculated as follows:

$$E = \frac{\rho V_p^2 (3V_p^2 - 4V_s^2)}{V_p^2 - V_s^2} \quad (3)$$

$$G = \rho V_s^2 \quad (4)$$

$$\mu = \frac{V_p^2 - 2V_s^2}{2(V_p^2 - V_s^2)} \quad (5)$$

where $V_p =$ velocity of the dilatational wave, $V_s =$ velocity of the shear wave, $\rho =$ density of the specimen.

3 SPECIMEN PREPARATION AND TESTING METHODS

The materials tested in this research programme are Harbin clay, Lanzhou loess and Beijing fine sand. The liquid limit and plastic limit of Harbin clay are 35.5% and 22.0% respectively, and of Lanzhou loess are 24.6% and 17.2%, respectively. The procedure of preparing samples is as follows. The dry soil is put into a vessel, and a specific amount of water is poured in and mixed to achieve the required uniform water content. After the moist soil was stored overnight in an airtight container to allow the moisture to equilibrate, then it was carefully packed into a mould, which was encased in a supporting copper jacket and tamped a certain number of times until the desired density was attained to form a 62 mm-diameter by 12.5 cm-deep specimen. The specimen-filled mould was then placed into a freezing cabinet at $-20$°C and was frozen quickly for 24 hours. This was thought to represent typical conditions in terms of deposition in normal unfrozen ground, which after freezing for about 24 hours would also represent the structure in frozen soil.

The copper jacket was then removed, and the specimen was cut on a lathe in a cold room. The nominal size of specimen after cutting was around 30 ~ 40 mm long. Thermometers were inserted into the center of the thermostat tank to monitor the test temperature. The sample was then placed into the thermostat tank, through which the cooling fluid was circulated by a pump. Transducers were pressed against both top surface of the sample with a thin petroleum jelly sheet for better coupling.

4 TEST RESULT DISCUSSION

4.1 Relationships between dynamic elastic mechanical parameters and temperature

The relationship between dynamic elastic modulus and temperature is shown in Figure 1. It can be seen from this figure that the elastic modulus increases with decreasing temperature. According to regression analysis on the test data, the relationship between the dynamic elastic modulus and temperature can be described as follows:

$$E = A \ln \theta + B \quad (6)$$

where $E =$ the dynamic elastic modulus of frozen soils, $\theta =$ the absolute value of temperature in °C, and
\( A \) and \( B \) = the parameters associated with the type of soil and water content. The values of \( A \) and \( B \) of the three types of frozen soils are listed in Table 1 for different water contents. Figure 1 also shows that the coarser the soil grain, the higher the dynamic elastic modulus is, namely, \( E_{\text{sand}} > E_{\text{silt}} > E_{\text{clay}} \).

The variation of dynamic shear modulus of frozen soils with temperature is plotted in Figure 2. The dynamic shear modulus increases with decreasing temperature. It can also be seen from this figure that the coarser the soil grain, the greater the dynamic shear modulus is, namely, \( G_{\text{sand}} > G_{\text{loess}} > G_{\text{clay}} \). The relationship between \( G \) and \( \theta \) can be described by the following equation based on the regression analysis of the test data:

\[
G = C \ln \theta + D \tag{7}
\]

where \( G \) = the dynamic shear modulus in GPa, \( \theta \) = the absolute value of temperature in °C, and \( C \) and \( D \) = the parameters related to the type of soil and water content. The values of \( C \) and \( D \) are shown in Table 2.

Figure 3 shows the relationship between the dynamic Poisson’s ratio and the temperature of three types of soil. It can be seen from this figure that the dynamic Poisson’s ratio decreases with decreasing temperature. It is in good agreement with the results of investigations by He & Zhu (1999) obtained with the MTS. Nakano et al. (1972) reported that the dynamic Poisson’s ratio of frozen sand is almost not related with temperature, however, it decreases with ascending temperature for frozen silt and clay. The law of dynamic Poisson’s ratio varying with the type of soil is not similar to the static Poisson’s ratio obtained by He & Zhu (1999). In this investigation, the result is that the dynamic Poisson’s ratio is becoming greater as the soil grain becomes finer, namely, \( \mu_{\text{clay}} > \mu_{\text{loess}} > \mu_{\text{sand}} \).

### 4.2 The relationship between the dynamic elastic mechanical parameters and water content

The curves of dynamic elastic mechanical parameters of frozen clay and frozen loess versus water contents at various temperatures are shown in Figure 4, 5 and 6, respectively. It is obvious that the dynamic elastic modulus and the dynamic shear modulus of frozen loess increases as water content increases for the range of lower water contents, whereas they decrease with increasing water content for frozen clay at the range of higher water contents. The increase in modulus of soil frozen at lower water contents is due to the increasing cementing area of the ice matrix with increasing water content. However, after reaching a certain water content, the soil grains were increasingly separated from each other, and the cementing effect becomes reduced, so the frozen soil modulus is lower.

Figure 6 shows that the dynamic Poisson’s ratio of frozen clay and loess increase with increasing water content for the range of water contents investigated.

### Table 1. Values of the parameters \( A \) and \( B \) in Equation (6).

<table>
<thead>
<tr>
<th>Frozen soil type</th>
<th>Water content (%)</th>
<th>( A )</th>
<th>( B )</th>
<th>( R^2 )*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Frozen clay</td>
<td>31.0</td>
<td>2.61</td>
<td>2.53</td>
<td>0.9948</td>
</tr>
<tr>
<td>Frozen loess</td>
<td>18.9</td>
<td>4.16</td>
<td>10.76</td>
<td>0.9990</td>
</tr>
<tr>
<td>Frozen sand</td>
<td>17.6</td>
<td>2.88</td>
<td>21.76</td>
<td>0.9994</td>
</tr>
</tbody>
</table>

\( * \) correlation coefficient.

### Table 2. Values of the parameters \( C \) and \( D \) in Equation (7).

<table>
<thead>
<tr>
<th>Frozen soil type</th>
<th>Water content (%)</th>
<th>( C )</th>
<th>( D )</th>
<th>( R^2 )*</th>
</tr>
</thead>
<tbody>
<tr>
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<td>31.0</td>
<td>0.98</td>
<td>0.88</td>
<td>0.9963</td>
</tr>
<tr>
<td>Frozen loess</td>
<td>18.9</td>
<td>1.74</td>
<td>4.07</td>
<td>0.9982</td>
</tr>
<tr>
<td>Frozen sand</td>
<td>17.6</td>
<td>2.02</td>
<td>8.40</td>
<td>0.9940</td>
</tr>
</tbody>
</table>

\( * \) correlation coefficient.

Figure 2. Relationship between dynamic shear moduli and temperature.

Figure 3. The dynamic Poisson’s ratio versus temperature for frozen soils.
5 CONCLUSION

The results presented herein show that the ultrasonic technique can be used to measure the dynamic elastic mechanical parameters of frozen soil in the laboratory, furthermore, it may be applicable by transferring to equivalent field campaigns. Although there are several measuring methods to survey the dynamic elastic mechanical parameters of frozen soil, the ultrasonic technique should be preferred.

Both the dynamic elastic modulus and shear modulus of the three types of frozen soils tested increase as temperature decreases and their relationships can be described by a unique formula, with different values of parameters.

Both the dynamic elastic modulus and shear modulus of frozen silt increase with water content increasing over the range of lower water contents, whereas they decrease with increasing water content in the range of higher water contents for frozen clay.

The dynamic Poisson’s ratios of the frozen soils tested increase with temperature increasing and water contents increasing.

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Spatial and temporal variation in thaw depth in Siberian tundra near Tiksi

K. Watanabe & H. Kiyosawa
Faculty of Bioresources, Mie University, Tsu, Japan

K. Fukumura
Department of Environmental Engineering, Utsunomiya University, Utsunomiya, Japan

T. Ezaki
Fujita Corporation, Tokyo, Japan

M. Mizoguchi
Department of Biological and Environmental Engineering, The University of Tokyo, Tokyo, Japan

ABSTRACT: Siberian tundra is characterised by permafrost overlaid with an active layer. The spatial and temporal variation in the thaw depth in the active layer may play an important role in biological and hydrological processes. In this study, we carried out field investigations of flat tundra and a hillslope in a watershed near Tiksi, Siberia, at various scales, from 1997 to 2000. The ground started to thaw in early June and refroze in mid September. The progress in the ground thaw was proportional to the square root of the thawing index. The maximum thaw, which was observed at the end of August, ranged from 1.2 to 0.2 m and averaged 0.4 m. The spatial variation was analysed using a variogram method. The ground thaw had a fractal dimension (= 1.8) at each scale. While there was no anisotropy in the thaw depth in the flat area, the thaw depth on the hillslope periodically fluctuated perpendicular to direction of maximum slope.
The thermal conductivity increases with depth from the live-plant layer through the organic soil to the silt (Watanabe et al. 2000). The hilltops are covered by shale, with some lichens (Umbilicaria sp., Rhizocarpon sp., Dryas sp., Festuca sp., and Luzula sp.). The permafrost is estimated to be over 500 m thick (Fartyshev 1993). A network of ice-wedge polygons with diameters of about 7 m occur in the tundra. A field of these polygons was studded with non-sorted circles (frost boils) about 0.5 m in diameter (Washburn 1980).

This survey was carried out between 1997 and 2000. The average air temperature was $-13.5\degree C$, with a maximum recorded temperature of $32.6\degree C$ and a minimum of $-53.6\degree C$. The air temperature remained below 0\degree C from mid September to mid May. The average annual precipitation was 345 mm. Compared to the ground thaw season in 1997, there was less precipitation in 1998 and relatively warmer temperatures in 1999 and 2000. The prevailing winds were northeasterly in summer and southwesterly in winter (Kodama 1998). Due to relatively high wind speeds (mean of 5 m/s), less than 0.1 m of snow accumulated in the wetlands and it piled into drifts (2–20 m thick) in the valleys during winter. The drifted snow remained until August, and supplied water for the wetlands.

3 METHODS

An area of $1000 \times 1000$ m, consisting of a network of 121 precisely located permanent stakes, driven into the permafrost at 100 m horizontal intervals (F10G), was established in a tundra flat (Hinzman 1997). Nested (hierarchical) square grids with intervals of 10, 5, and 1 m were fitted inside the southeast end of the area (F10G, F5G, F1G) (Fig. 1). A meteorological station was located 100 m south of the area. Parts of the area were covered by a mat of moss forming a 0.03–0.2 m layer of live plants. Likewise, an area of $100 \times 100$ m was chosen on a hillslope (average gradient 2°), and nested grids with intervals of 10, 5, and 1 m were set up in the area (S10G, S5G, S1G).

The thaw depth of the ground was measured monthly at each grid intersection from June to September using a steel rod, calibrated in centimetres and pushed vertically into the ground. Since it was difficult to recognise the ground surface where it was covered by a thick mat of moss, we placed a wooden panel (0.2 m $\times$ 0.2 m, weight 500 g) on the plants and set this as the ground surface. Mackay (1995) discusses some potential sources of error associated with this methodology. Soil temperatures were measured at depths of 0.1 and 0.2 m using a thermistor, and the thaw depth was checked every August. In August, the ground surface level, ground water level, gradient, soil type, and soil moisture were also investigated at each intersection.

Additionally, the vegetation cover in each 10 m square grid was classified by the dominant species observed (Watanabe et al. 2001).

4 RESULTS AND DISCUSSION

4.1 Temporal variation in thaw depth

4.1.1 Difference by year

Table 1 lists the thaw depth in each grid during the maximum thaw season between 1997–2000 (i.e. active layer thickness). The average thaw was $0.40 \pm 0.15$ m and the maximum and minimum values were 1.05 m and 0.15 m, respectively. Compared with 1997, the ground thaw was shallower in 1998, as a result of a dry summer, and deeper in 1999 and 2000, which were warmer summers. A similar trend was not seen in the maximum and minimum values of thaw depth however (Table 1).

Ground thaw is strongly influenced by the properties of vegetation and soil (Nelson et al. 1997, 1999, Hinkel et al. 2000). In this watershed, thaw depth varies with vegetation type, the vegetation cover ratio and micro-undulations, which induce water runoff (Watanabe et al. 2001). Figures 2a–c map the thaw depth in F10G in...
1997, 1998, and 1999, respectively. The ground thawed to a deeper level in the northwestern part of the grid than in the southeastern part. At some points near the northwestern midpoint of the grid, the thaw depth was close to 1 m. The thaw depth was locally deeper at or near water-filled depressions and near non-sorted circles. It was shallower in moss-covered areas. The locations of points at which the ground thaw was deeper were similar in all three years, as were the depth of thaw, despite the year-to-year spatial variation in the thaw and mean thaw depth.

4.1.2 Seasonal change
The soil in the observation area started to thaw in mid June, reaching a maximum depth in late August to early September, before refreezing after mid September. Several previous studies have addressed the correlation between thaw depth and air temperature records (e.g. Gray et al. 1988, Jorgenson & Kreig 1988). Figure 2 shows the seasonal change in the average thaw depth and the standard deviation at F10G and S10G in 1998. The thawing index is a time-temperature integral, calculated by summing the mean daily temperatures above 0°C (Harlan & Nixon, 1978). In both observation grids, the ground thaw progressed as the thawing index increased, and reached its maximum depth in late August. The average thickness of the active layer was

<table>
<thead>
<tr>
<th>year</th>
<th>day</th>
<th>N</th>
<th>Avg (cm)</th>
<th>max</th>
<th>min</th>
<th>Std (cm)</th>
<th>Cov (%)</th>
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</thead>
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<tr>
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<td></td>
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<td></td>
</tr>
<tr>
<td>1997</td>
<td>8/15</td>
<td>625</td>
<td>35.1</td>
<td>48</td>
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<td>4.79</td>
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<tr>
<td>1997</td>
<td>8/16</td>
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<td>105</td>
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<td>11.07</td>
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<tr>
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<td>75</td>
<td>26</td>
<td>10.74</td>
<td>23.78</td>
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</table>

N indicates number of samples; Avg, Std and Cov are average, standard deviation and coefficient of variance.

Figure 2. Maps of thaw depth in the 100 m × 100 m observation area on flat tundra, F10G (a–c), and hillslope, S10G (d). (a) August 16, 1997; (b) August 15, 1998; (c) August 19, 1999; (d) August 20, 1998. The solid lines are the thaw depth contour with a 5 cm interval.
53.5 cm in F10G (range 33 to 83 cm) and 32.6 cm in S10G (range 16 to 60 cm). The standard deviation increased with increasing thaw depth, indicating that variation increased as the ground thawed.

The relationship between thaw depth, $Z$, and the thawing index, $DD$, was $Z_{F10G} = 1.9 \text{ DD}^{0.54}$ ($R^2 = 0.99$) for F10G and $Z_{S10G} = 2.0 \text{ DD}^{0.53} - 8.5$ ($R^2 = 0.99$) for S10G (shown as solid lines in Fig. 3). In both areas, the ground thaw was roughly proportional to the square root of the thawing index, indicating that the trend in the seasonal change in thaw depth in this watershed is primarily homogeneous. These relationships agree with the form of the Stefan solution, which has been used in studies of active layer thickness (e.g. Hinkel & Nicholas 1995, Nelson et al. 1997). The active layer on the hillslope was shallower than that on the tundra flat, despite similar air temperatures and vegetation. This was presumably due to differences in slope aspect and the existence of a snow patch just above the hill-slope, which continually supplied flowing cold water to S10G during the ground-thaw season.

4.2 Spatial variation in thaw depth

4.2.1 Topography

The depth of thaw differed between the tundra flat and hillslope, although the trend in ground thaw progression was similar at the two sites (Fig. 3). We analysed the spatial variability in both observation grids (F10G & S10G). Figures 2b and 2d map the thaw depth in F10G and S10G respectively, in the maximum thaw season of 1998. S10G was oriented in the direction of the slope. The average thaw depth was deeper in F10G than in S10G. In F10G, locally deep thaws were observed near non-sorted circles, which consisted of silt with a high thermal conductivity. This factor was compounded by the lack of vegetation on their surface and a reduced albedo. Dense contour lines are therefore seen in the thaw-depth map. By contrast, no non-sorted circles were observed in S10G. Numerous water tracks developed into gullies in the hillslope. In some places, the active layer in these gullies was deeper than that on the tundra flat.

The spatial correlation in thaw depth variation was calculated by semivariogram, $\gamma' = 0.5 \text{ var}\{z(s + h) - z(s)\}$, where $z$ is the thaw depth at point $s$, and $h$ is the spatial lag. The spatial variation increases with $\gamma'$. If anisotropy were present, $\gamma'$ has a different form (Mizoguchi et al. 1998, Gomersall & Hinkel 2001). Figures 4b and 4d show $\gamma'$ for F10G and S10G, respectively, associated with each lag distance. In F10G, $\gamma'$ flattens at a lag of about 10 m (range $\approx 10$ m), indicating that sampling of thaw depth at intervals over 10 m are independent observations. No difference in $\gamma'$ was observed in any direction. There appears to be no anisotropy in the thaw depth in the tundra flat. Likewise, no periodicity of $\gamma'$ was detected for the slope direction in S10G. By contrast, $\gamma'$ for the contour direction in S10G had local minima with a periodicity of 7 m (arrows in Fig. 4d). Water tracks meandered through the surface of ice wedges that formed at an angle around the polygons, and some of the water tracks developed into gullies. Although the ground thaw is mainly dependent on the vegetation and the thermal properties of soil (Nelson et al. 1997, 1999, Hinkel et al. 2000, Watanabe et al. 2001), $\gamma'$ observed in the contour direction may be attributed to the water tracks. The drop in periodicity in the direction of the slope can be explained in terms of movement of the subsurface soil by solifluxion.

4.2.2 Scale

The values of thaw depth, such as average and maximum thaw, differed when a different grid interval was
selected (Table 1). The difference between the maximum and minimum thaw depth decreased, and detailed variation in thaw depth was obscured by increasing the grid interval. Here, we compare the spatial variability for each grid size. Figure 5 shows the relationship between variogram, \( \gamma \), and the lag, \( h \), for each scale (F1G, F5G, F10G, & F100G) in the maximum thaw season of 1997, and for F100G during the 1997–2000 thaw seasons. There are no fundamental differences in the variability of thaw depth at each grid scale, which have similar fractal dimensions, \( D_i \) (4–2.2) (Burrough 1981, Zhang et al. 2001). Figure 6 shows the seasonal change in the fractal dimension observed in Figure 5. The fractal dimension decreased with ground thaw, reached a minimum value during the maximum ground thaw period, and then increased with ground freezing, presumably because ground thaw progressed irregularly during the first stage, becoming smooth as it progressed. A similar trend was observed in all observation years. Although the \( D_i \) value is high, it would seem worthwhile to use the \( D_i \) value as a guide to further mapping and observation.

These results suggest the optimal sampling interval for field observations in this watershed. Figures 5 and 6 show that an interval of 100 m or greater is adequate for capturing the primary scale of the variability in ground thaw for this watershed. This interval fails however to resolve some of the minor variability related to localised phenomena. At the scale of hydrological processes associated with cryogenic topography such as gully erosion and solifluxion, sampling at less than 10 m intervals is useful, as shown in Figure 4. Alternatively, to consider the specific mechanism of ground thaw at one point, we may need to determine the soils’ thermal properties, surface conditions, and vegetation cover at a spatial interval that is at least the same as that used for the maximum thaw depth (i.e. 1 m or less for this watershed) (Watanabe et al. 2001).

5 CONCLUSION

Variability in thaw depth occurred at different temporal and spatial scales in a tundra watershed near Tiksi, Siberia. The ground began to thaw in early June and refroze in mid September. The progression of ground thaw was proportional to the square root of the thawing index. The maximum thaw, which was observed at the end of August, ranged from 1.2–0.2 m and averaged 0.4 m. While there was no anisotropy in the variance of thaw depth in the tundra flat, the thaw depth on the hillslope periodically fluctuated perpendicular to the gradient. The ground thaw had a similar fractal dimension (=1.8) that was independent of the sampling interval. Although temporal and spatial variations in thaw depth are important for understanding biological and hydrological interactions, it is difficult to identify the variation in a large watershed over a short time. Our results have significant implications for designing effective strategies for the sampling and construction of regional estimates of thaw depth and closely related phenomena.

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