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Reprint from FOREST SOILS AND LAND USE

Proceedings of the
Fifth North American Forest Soils Conference
Colorado State University, Ft. Collins, Colorado

August 6-9, 1978

Edited by
Chester T. Youngberg

Published by
Colorado State University

SENSITIVE SOILS OF PERMAFROST TERRAIN

S.C. Zoltai and V. Woo[1]

Permafrost in northern soils creates conditions not encountered elsewhere. Great quantities of water may accumulate in the ground in the form of ice. As the brief summer progresses, the thawing of seasonal frost continually releases moisture in the thickening active layer. Permafrost remains within 0.5-2 m of the surface, acting as a virtually impermeable layer to percolating water. It also keeps the soil cool, retarding the evaporation of moisture from the soil.

In the natural state, soil, vegetation and permafrost are thermally balanced. Permafrost occurs where the climatic thermal regime, as modified by vegetation and soil conditions, is favorable for its development and maintenance. In many areas, especially near the southern limits of permafrost, the thermal balance is precarious. Should the vegetation or soil conditions be changed by artificial or natural means, the thermal balance will be altered and the permafrost may thaw completely. Certain soils are sensitive to shock: vibrations can cause them to liquefy and flow as a viscous mass. Permafrost soils are therefore sensitive to disturbance because the consequences of the disturbance far exceed the magnitude of the initial disruptive action.

In permafrost areas there are soils that are prone to gullying, riverbank erosion, wind erosion or landslides familiar from other parts of the world. However, conditions in permafrost terrain create additional, unique hazards which are examined in this paper.

THERMALLY SENSITIVE SOILS

PROPERTIES OF PERMAFROST SOILS

Permafrost, or more precisely perennially frozen ground, is a temperature condition defined as ground that remains below 0°C during two winters and the intervening summer (Brown 1970). It is formed when the ground has a negative heat balance; the annual heat loss being greater than the heat input. When the average annual soil temperature falls below 0°C, permafrost is formed.

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Permafrost can be found in about two-thirds of Canada (Fig. 1). At its southern limits, permafrost occurs in peatlands only, often as thin (1-2 m) patches of frozen peat. This is the Localized Permafrost Zone (Zoltai 1971). Farther north in the Discontinuous Permafrost Zone (Brown 1967), permafrost becomes more common in the peatlands and in moss-covered imperfectly-drained woodlands. Still farther north in the Discontinuous Permafrost Zone, permafrost becomes widespread, occurring in all types of terrain except some very wet peatlands and dry granular deposits. In the far north is the Continuous Permafrost Zone, where all land surfaces are underlain by permafrost.

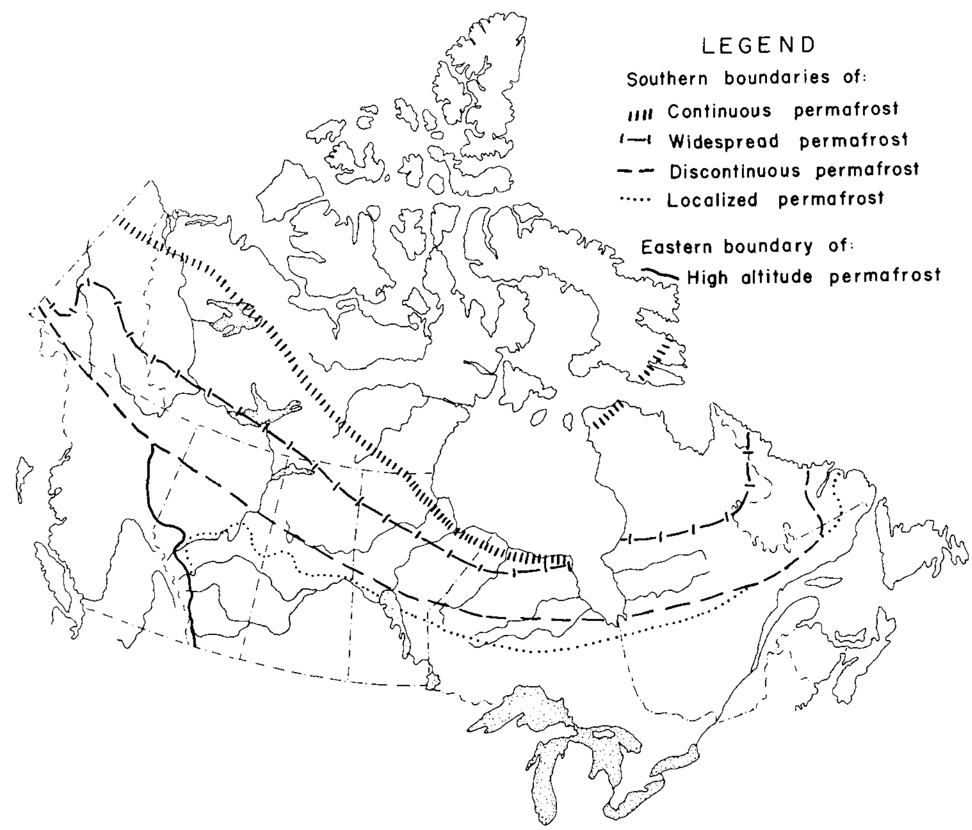

Figure 1. Map of permafrost zones in Canada, modified after Brown (1967).

The temperature of permafrost generally decreases northward and its thickness increases. It is not unusual to find permafrost layers only a few centimetres thick in the Localized Zone (Zoltai 1972a), with a mean annual soil temperature barely below 0°C. In the Discontinuous Zone the thickness of permafrost may reach 10 m, but the temperature is still in the -1°C range throughout the frozen lens (Brown 1970). In the Continuous Zone the thickness of permafrost may reach 450 m at mean annual soil temperatures of -10°C. The temperature of the upper 6-16 m is influenced by seasonal air temperatures (Brown 1970), but below this depth the temperature remains steady, influenced by long-term climatic conditions and geothermal heat.

The surface layer that freezes and thaws every year over the permafrost is the active layer. Its thickness varies from about 2 m on mineral soils in the south to as little as 25 cm in the Arctic islands, and is generally influenced by the heat-conducting qualities of the soil and vegetation layer. Both living and dead organic material form an effective insulation over the soil. Removal of the organic mat by fire resulted in a twofold increase of the active layer thickness within three years in the Discontinuous Zone (Zoltai and Pettapiece 1973). At Inuvik, in the Continuous Zone, active layer thickness increased by 83% one year after stripping of the soil surface (Heginbottom 1971).

The thermal regime of the soil depends on its moisture content. Although the thermal conductivity of water is greater than that of air (Baver et al. 1972), the thermal capacity of water is also substantially higher. The rate of heat flow, or thermal diffusivity (a ratio of thermal conductivity to thermal capacity), in saturated soils is considerably slower than in unsaturated soils. The slower thermal diffusivity results in more net energy accumulation in the wet soils. Field observations in the Continuous Zone confirmed that maximum summer thaw depth is always deeper in saturated peat than in adjacent better-drained peat. For the same reason, saturated peat in the Discontinuous Zone usually contains no permafrost, even though adjacent better-drained peat may be frozen.

The moisture content of the active layer and the ice content of the near-surface permafrost can vary greatly according to internal and external drainage conditions. In sands and gravels the amount of moisture or ice is negligible (Mackay 1971a), although ice may occur in wedges (Fig. 2). However, in imperfectly- to poorly-drained soils, the top 50-100 cm of permafrost often contains over 85% ice by volume. The occurrence of this icy layer in the near-surface permafrost is especially predictable under earth hummocks (Zoltai and Tarnocai 1974) (Fig. 3).

Massive ice bodies can be found in fine-textured soils that either overlie coarse-grained soils or are interstratified with them (Mackay 1971a). Under such conditions, massive ice bodies tens of metres thick can accumulate. The ice may be pure, as in the case of injected ice under pingos, or it may consist of impure ice with a gravimetric moisture content of over 200% (Mackay 1971b).

TERRAIN DAMAGE

In northern Canada there are extremely sensitive areas where even a slight disturbance results in severe terrain degradation, but other areas can withstand considerable abuse. The severity of damage and the ease of rehabilitation depend on terrain (ground ice conditions, drainage, texture, slope, etc.), climate (ground temperature, air temperature, snow cover, etc.) and vegetation (insulating qualities, composition of flora, choice of species, revegetation strategy options, etc.).

Damage to thermally sensitive soils can be grouped into three categories: subsidence, mass wastage and erosion. Subsidence occurs when the insulating quality of the active layer is reduced and summer thawing reaches greater depth. If the excess water produced by the melting near-surface permafrost

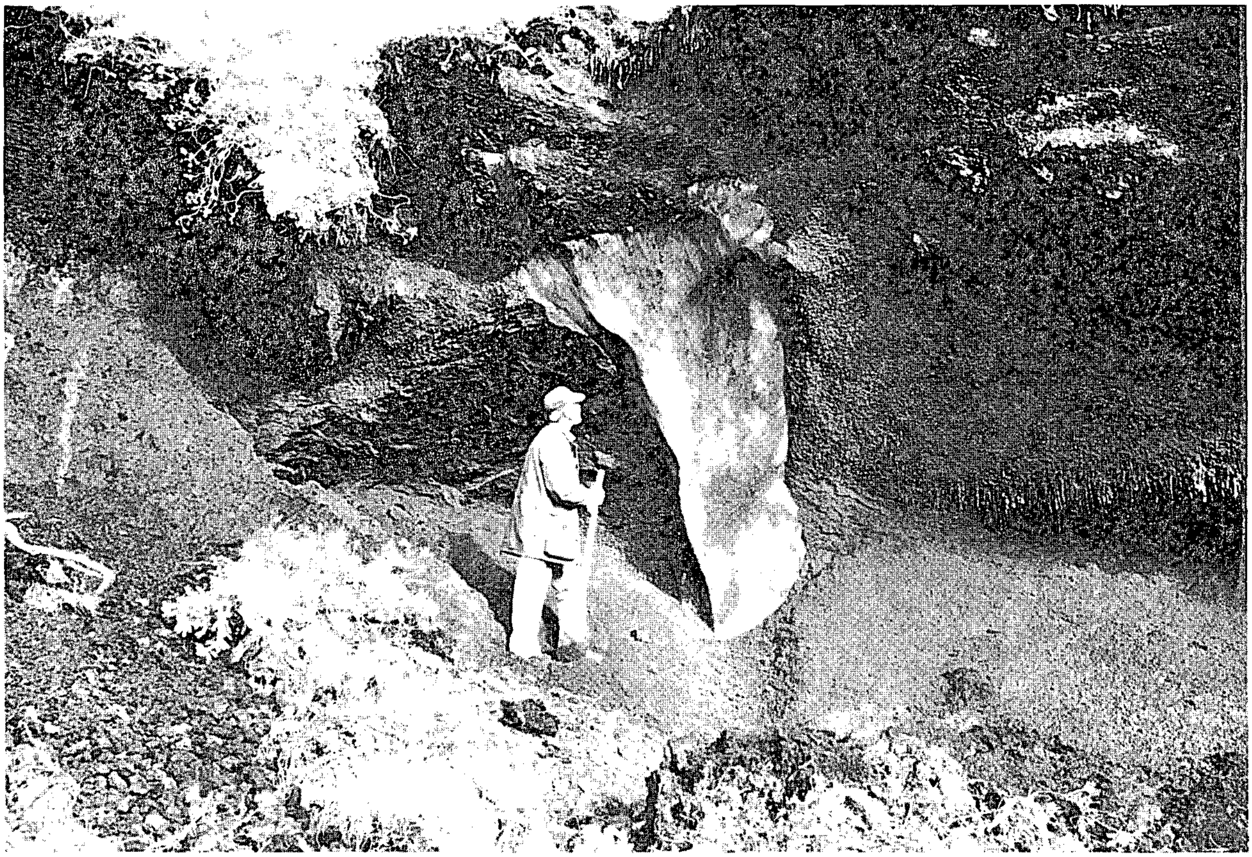

Figure 2. Ice wedge in fine-grained alluvial soil.

drains away, a new active layer will develop at a greater depth relatively rapidly; perhaps within three years. However, the ground surface will be lower than before due to the loss in volume by the melted ice. On the other hand, if the ice content is high, and the surface drainage is poor, water will accumulate on the subsided surface, causing a deeper thawing of the soil through increased heat absorbtion (Fig. 4). Ponds may be formed by this thermokarst process and may continually be enlarged by the retrogressive thawing of the banks.

The severity of environmental damage by subsidence is variable. It may be slight in some areas of the Arctic and Subarctic where only a general lowering of the surface may take place. In the fringe permafrost areas, disturbances may cause complete thawing of permafrost bodies. Structures and buildings are particularly vulnerable to the effects of subsidence through differential settling of foundations.

Mass wasting causes perhaps the most spectacular thermal degradation of the terrain. It occurs when a disturbance exposes an underground icy body. The meltwater, mixed with soil, flows downslope in a slurry, preventing the accumulation of slumped material and thus keeping the icy headwall exposed to

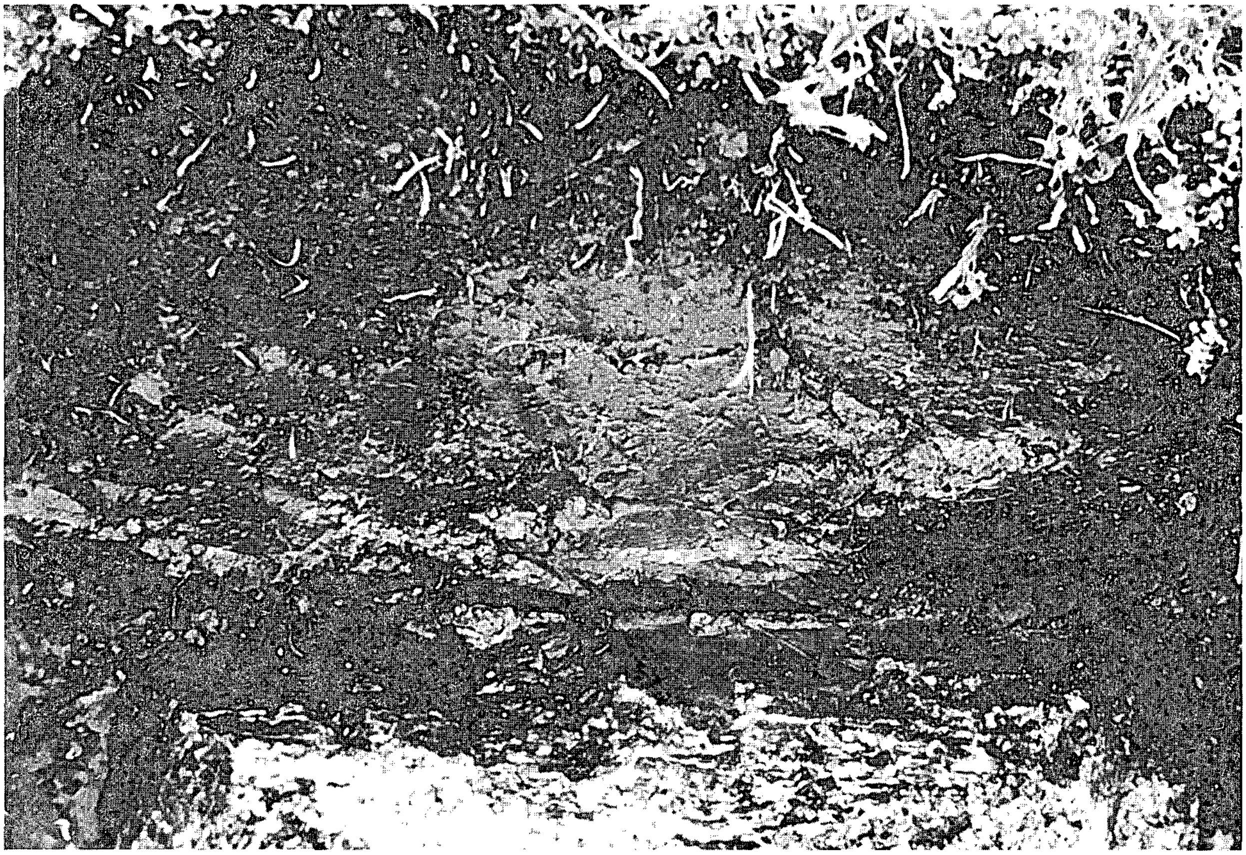

Figure 3. Icy layers under an earth hummock. The light-colored material near the base is 94% ice.

thawing temperatures (Fig. 5). Examination of 35-year-old aerial photos shows that some of these retrogressive thaw flow-slides are long-lived and may expand upslope as much as 10 m a year (Mackay 1971b). Stabilization occurs only if the ice mass is all thawed, or if the surface organic mat, undercut by thawing, slumps and lodges against the headwall, thereby insulating it from further thawing.

Hydraulic erosion may be induced by thermal subsidence, especially if the shape of the disturbed area is linear, such as roads or tracks. The thaw-settled surface forms a channel that can intercept surface drainage and divert small creeks. If the soil is not ice-rich and thaw settlement is minor, the thickened active layer under the track can still form a groove in the permafrost table, which may in turn channel sub-surface drainage and initiate gully erosion under certain conditions.

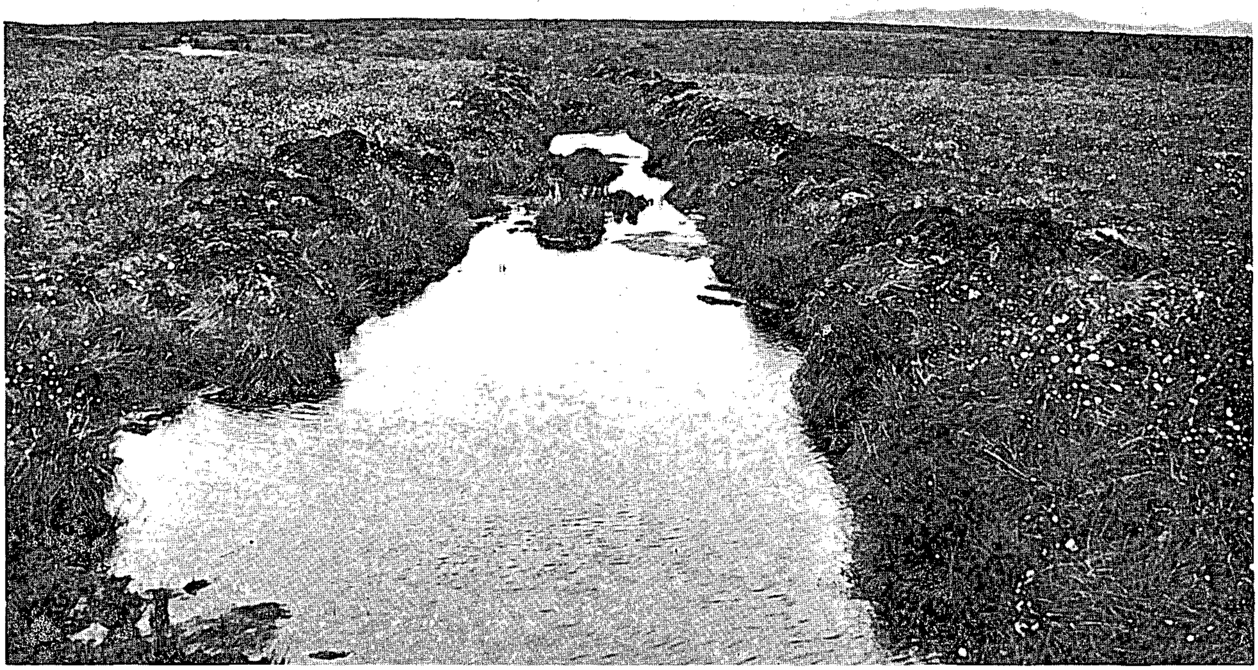

Figure 4. A subsidence of 1 m on a seismic line after the removal of the insulating organic mat.

SHOCK-SENSITIVE SOILS

The active layer of many permafrost soils becomes liquefied when subjected to mechanical vibrations such as those caused by machines or by repeated shifting of a person's weight (Fig. 6). After a period of rest, these soils regain their original solid state, indicating that the liquefaction process is reversible. Under certain conditions, the resolidified soil mass may become denser than its original undisturbed state. For instance, prolonged vibration during transport of samples causes them to become hard masses from which the water separates and collects on the surface (Fig. 7).

Shock sensitivity in the active layer of permafrost soils is wide-spread in northern Canada. In the Continuous Permafrost Zone, the till derived from crystalline rocks of the Canadian Shield is shock-sensitive, covering about half the land surface (Zoltai and Johnson 1979). The soils of some Arctic islands derived from carbonate rocks are also shock-sensitive (Woo and Zoltai 1977). Shock sensitivity has been reported from the Yukon and Northwest Territories (Shilts 1974, Tarnocai 1977), Labrador (Eden 1977) and northern

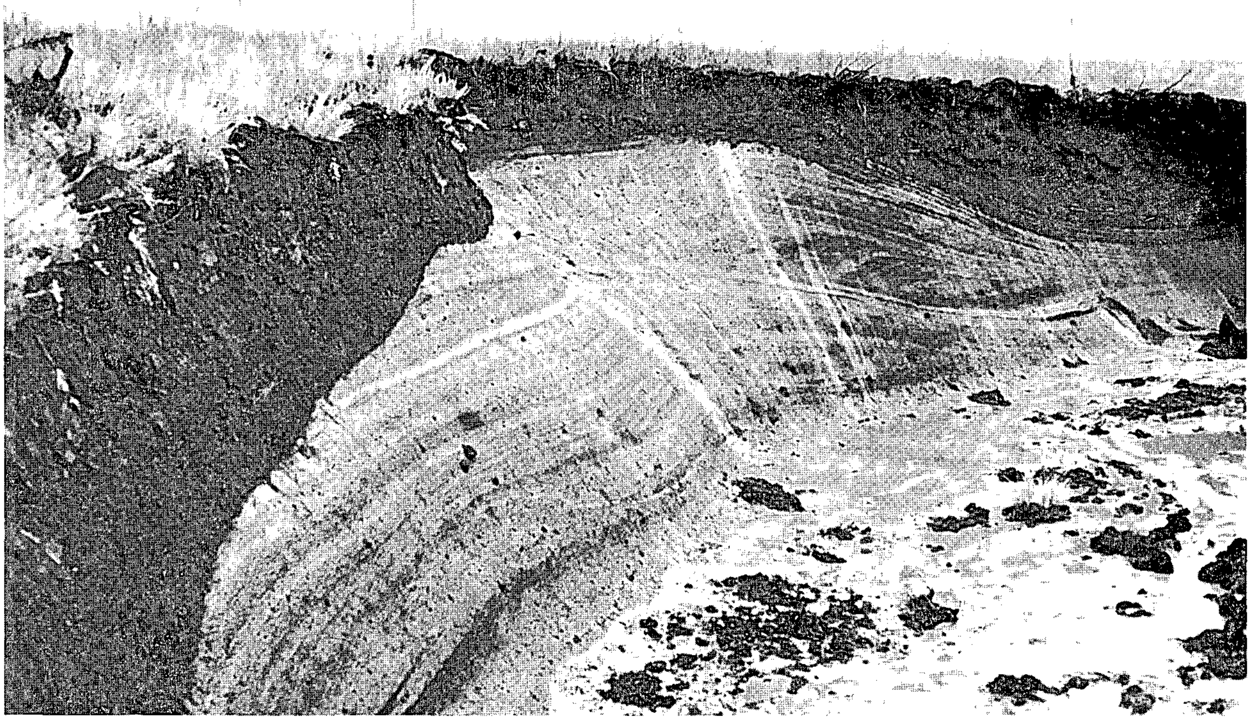

Figure 5. Icy headwall of a thawing massive ice bed.

Manitoba (G. Mills, Manitoba Soil Survey, personal communication), and from tundra (Liverovskaya-Kosheleva 1965) and mountain regions (Mudrov and Tumel 1967) of the Soviet Union.

Textural data on shock-sensitive soils that we collected from widely separated areas in northern Canada show that about two-thirds of the grains fall into the medium size range (fine sand, very fine sand and silt; 0.25-0.002 mm diameter). These grain-size analyses of twenty shock-sensitive soils derived from crystalline bedrock indicated that the medium-sized particles averaged 68% (± 10%), and the medium-sized particles in twenty-five shock-sensitive soils derived from carbonate bedrock averaged 69% (± 14%). The soil from the crystalline rock areas contained 5-10% clay, whereas the calcareous soils contained 12-25% clay; but this difference in clay content did not affect the shock sensitivity of these soils.

The plastic limit and liquid limit of shock sensitive till derived from crystalline rocks is very low (Zoltai and Johnson 1979), with a plasticity index generally below 1%. Since the material readily changes from solid to liquid state, the plastic phase is brief even at low moisture levels (6-12% by weight, Shilts 1974).

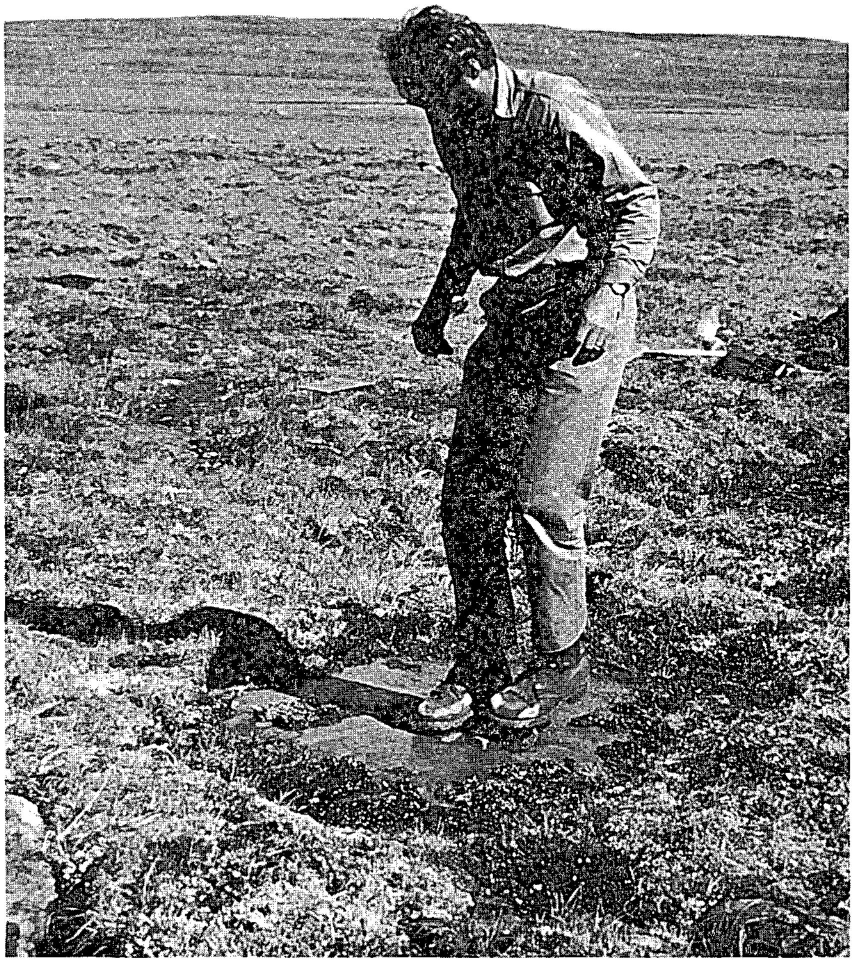

Figure 6. Soil becoming liquefied under the shifting weight of man.

ORIGIN

The origin of shock sensitivity can be explained by conventional soil mechanics concepts. Figure 8 illustrates the behaviour of shock-sensitive soils under different stress conditions. Consider a system of granular soil particles loosely arranged in a "quasi-crystalline" structure; the pores may or may not be saturated with water (Tsuchiya 1977). If a series of shock stresses is delivered by vibrating the sample (Fig. 8, Step 1), the additional energy is initially absorbed by the pore water. When the pore pressure build-up exceeds intergranular friction, the structure collapses, and the soil mass liquefies and flows under self-weight. The occurrence of liquefaction can be predicted by a mathematical equation; it depends on the rapidity of pore pressure buildup and stress history of the soil (Tsuchiya 1977).

In contrast, if stress is applied with a steadily increasing shear rate, such as pressure exerted by a thumb on the surface (Fig. 8, Step 2), water will simply be squeezed out of the pores. The soil mass becomes compacted and a

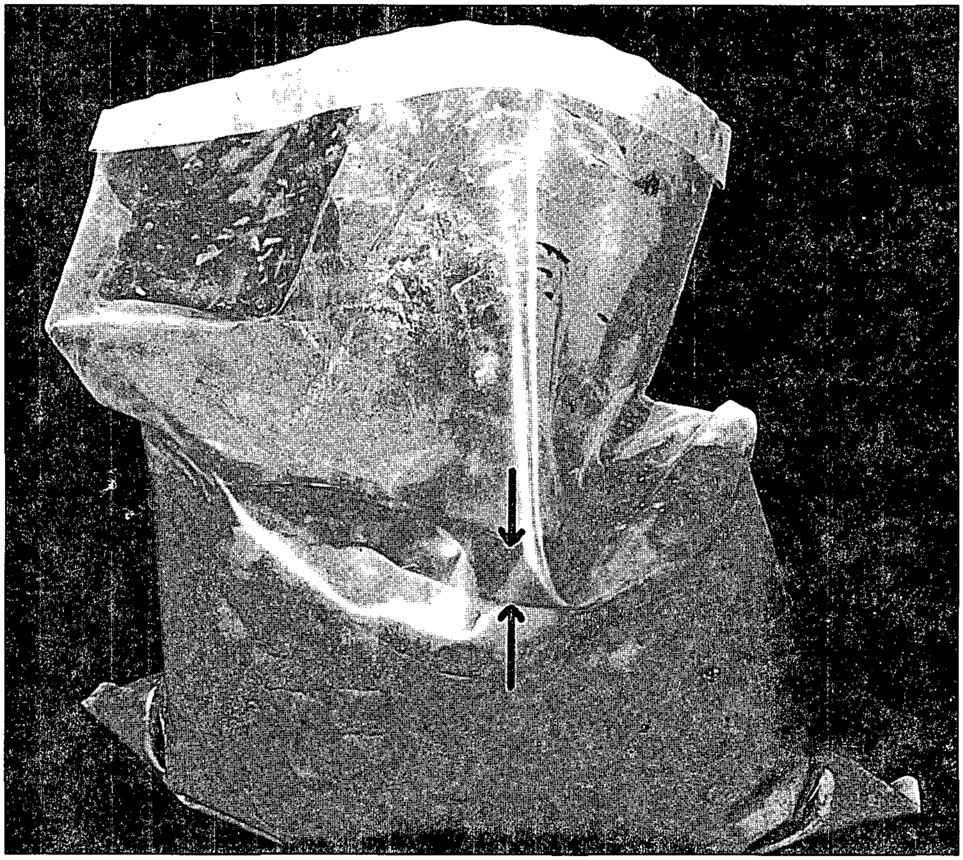

Figure 7. Sample of shock-sensitive soil after transport. Note the 1.6 cm layer of water (between arrows) separated from the densely compacted soil.

resistance builds up against shearing because soil particles are forced to roll over each other. This resistance, called dilatancy, is defined in rheology as an isothermal reversible increase in viscosity when shear rate is increased, with or without volume change and with no measurable time dependence (Bauer and Collins 1967). If shear stress increases indefinitely, the soil will eventually fail. Bauer and Collins (1967) note that dilatant systems often liquefy instantaneously on failure.

Liquefied soil will revert to solid state if allowed to rest. However, we have observed that when shock-sensitive soils are transported from the field, prolonged vehicle vibration will cause overconsolidation of previously liquefied soils (Fig. 8, Step 4). The samples turn into extremely hard masses with a layer of water separated from the soil on top of the sample (Fig. 7). This behaviour can be duplicated in the laboratory by using an automatic shaker set at fast and then slow speed. In rheology, the accelerated recovery of liquefied material by gentle vibrations is called rheopexy (Reiner and Scott Blair 1967). The overall effect is similar to shaking a bag of grain in order to pack it.

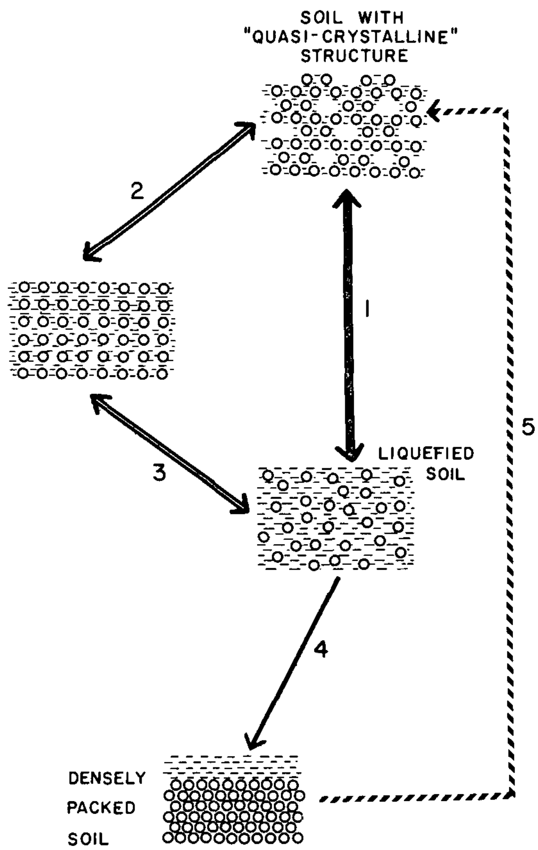

Figure 8. Schematic diagram of the mechanism of shock sensitivity. Single arrows indicate shock stress; the heavier line indicates the most severe shocks. Double line represents a steadily increasing shear stress. Striped line indicates a freeze-thaw process.

Step 1. Liquefaction: time-dependent reversible isothermal decrease in strength in response to shock stress.
Step 2. Dilatancy: reversible increase in strength due to shearing with no measurable time-dependency.
Step 3. Instantaneous liquefaction after failure.
Step 4. Prolonged vibrations cause consolidation of particles and separation of water layer.
Step 5. Repetitive freezing-thawing lead to loose soil aggregation.

If the liquefied soil is allowed to rest, most of its original hardness will be regained after an extended period. The recovery speed is related to the rate of pore pressure dissipation as well as stress history, and it can be calculated by Tsuchiya's equations (1977). In permafrost regions, pore pressure in the active layer is likely to remain high for much of the summer due to the gradual release of water from the thawing seasonally-frozen soil. A resolidified soil mass can be loosened again by the repetitive freezing and thawing cycles in the spring and fall (Fig. 8, Step 5). This action is due to the formation of ice crystals and lenses between soil particles during freezing (Scott 1969).

NOMENCLATURE

In the past, shock sensitivity of permafrost soils has been described as thixotropy (Liverovskaya-Kosheleva 1965, Mudrov and Tumel 1967), thixotropic-quick (Tsytovich and Kronik 1978), liquefaction (Eden 1977, Shilts 1974) and rheotropy (Tarnocai 1977, Woo and Zoltai 1977). Since most of the confusion arises from different usage of the same terms in different disciplines, a few words of clarification are offered here.

In soil mechanics, thixotropy is defined as an isothermal, reversible soil-gel transformation due to stress (Yong and Warkentin 1975, Terzaghi and Peck 1967). Thixotropy is believed to originate from electrical charges of clay particles in the soil. However, as pointed out earlier, most shock-sensitive soils in permafrost regions are granular soils with relatively low clay contents. Therefore, shock sensitivity is not thixotropy as understood in soil mechanics. On the other hand, in rheological nomenclature, thixotropy is defined much more broadly, and electrical forces are not considered to be a necessary condition of the phenomenon (Bauer and Collins 1967). Thus shock sensitivity is in keeping with the rheological concept of thixotropic dilatancy (Weltman 1960).

Furthermore, there are differences in the interpretation of the concept of thixotropy. Descriptions by various investigators (Liverovskaya-Kosheleva 1965, U.S. Soil Survey Staff 1975, Yong and Warkentin 1975) show discrepancies in the method of testing, the liquefaction process and the duration of the recovery period. Such lack of standardization may result in different investigators describing somewhat different phenomena by the same name.

Processes outlined in Figure 8 are equivalent to "dilatancy" and "spontaneous liquefaction" in soil mechanics (Terzaghi and Peck 1967). It is worth noting that "metastructure of true quick sands" (Terzaghi and Peck 1967) is very similar in concept to "quasi-crystalline structure" of thixotropic soils (Yong and Warkentin 1975). Because of these differences, we chose the term "shock-sensitive" to describe the observed behaviour of permafrost soils; the mechanism of shock sensitivity is identical to the one postulated by Tsuchiya (1977).

OCCURRENCE

The mechanism of shock sensitivity described above is not exclusive to permafrost areas. In fact, shock-sensitive soils are widespread in earthquake zones around the world (Tsuchiya 1977). Recently, Taskey et al. (this volume) described soil properties in the mountainous areas of Oregon that are very similar to shock sensitivity. However, the mineral halloysite is thought to be responsible for this soil instability. In view of our findings that shock-sensitive permafrost soils occur in areas of widely-different lithologies, the mineralogical explanation probably does not apply to northern regions.

Shock-sensitive medium-grained soils are particularly common in permafrost areas. Soils having textures identical to the shock-sensitive tills of northern Canada are widespread across the Canadian Shield (Zoltai 1965, 1972b; Eden 1977). Yet none of these soils has been reported to be shock-sensitive in nonpermafrost environments. It appears, therefore, that texture alone does not predispose a soil toward shock sensitivity.

The association of shock sensitivity and permafrost may stem from a delicate balance in the soil moisture regime. The gradual thawing of the active layer provides moisture to the soil throughout the summer. Furthermore, the permafrost table acts as an impermeable layer to water seepage, promoting high moisture levels in the soil; high pore water pressure is conducive to liquefaction under shock. In addition, the loosening of soil aggregation during freezing and thawing contributes to unstable soil structure.

Permafrost and the slow thawing of the active layer maintain the soil at a critically high moisture level for a relatively long period of time during the thawed season. The rebulking by frost action conditions the soil for a repetition of the cycle of liquefaction-solidification-loosening in response to shock. In nonpermafrost regions, although the soil texture may be similar, the critical moisture level may last for only a short period, and some of the other requisites may be absent or be too ephemeral for the initiation of shock-sensitive behaviour. This may explain the prevalence of shock-sensitive soils in permafrost areas.

TERRAIN DAMAGE

Sustained vibrations, such as those caused by moving or stationary machinery, digging, trampling, etc., can cause the shock-sensitive soils to liquefy. Soil liquefaction will take place in the active layer only. Structures impeding the internal drainage of shock-sensitive soils would tend to make them more prone to liquefaction by maintaining the soil at high moisture levels. In the Arctic islands a number of active layer detachment slides were noted near seismic operations, where the vibrations of detonations may have liquefied the active layer, causing it to flow down-slope. The removal of the active layer by slumping will cause further thawing of the permafrost, aggravating the mass wasting process.

CONCLUSIONS

The soils of northern Canada possess properties that make them especially vulnerable to damage and degradation. Such features as ice-rich soils and the consequences of exposing them to thawing are well known to resource managers and engineers. Our northlands are nevertheless replete with scarred terrain where lack of planning or poor operational control caused permanent terrain alteration. Other soil properties, such as shock sensitivity, are often overlooked, despite their importance in geotechnical considerations.

Many deleterious effects of man on the sensitive northern landscape can be avoided if the vulnerable properties of the soils are respected. The maintenance of insulation on ice-rich soils is an elementary precaution. Shock-sensitive soils may be more difficult to handle, but research and field trials will find practical solutions. Once the problem is identified and understood, it can usually be avoided or overcome.

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