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BY

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AN INVESTIGATION INTO PROCESSES OCCURRING IN SOLIFLUCTION

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ABSTRACT. The mechanics of solifluction downslope of a late snow patch has been investigated using instrumental techniques new in this field, for the determination of soil pore-water pressures, density and movement. Substantial changes in water content and considerable sub-surface water flow occur together with changes in state of stress. Low and sub-atmospheric pore-water pressures were observed. To explain solifluction as being analogous to soil movement induced by impeded drainage over an impermeable layer is an over-simplification, which ignores fundamental characteristics of newly-thawed and frozen soils. It is doubtful to what extent concepts of soil strength applicable under normal conditions apply during freeze-thaw processes.

INTRODUCTION AND ACKNOWLEDGEMENTS

Preliminary investigations of solifluction features were made in 1953, 1954 and 1955 in central Norway. These involved several new techniques and repeated observations (Williams 1957a, 1957b). The importance of detailed instrumental studies of processes in solifluction was obvious. Accordingly, during five weeks in June, July and late August, 1956 a team with an average strength of five maintained an instrument layout and recorded readings and observations on solifluction below a late snow patch. A cyclostyled report (Williams, et al., 1957) detailed these methods, and stimulated discussion. Subsequently, further field observations were made in March, June and August, 1957.

The field work was part of a program supported by the Royal Geographical Society, Norwegian Nansen Fund and others. Willing Cambridge and Oslo students, of whom Mrs. Kari Williams was the most persistent, were field assistants during the work. Mr. J. A. Pihlainen, National Research Council, Canada, carried out the statistical analyses. To these, thanks are extended.

SITUATION AND GENERAL DESCRIPTION

The occurrence of solifluction investigated (pl. 1), on the bank of the Einøvling stream near Svaanalegret, about 8 kilometers southeast of Snøhetta, is similar to many others in Dovrefjell, Trollheimen, Rondane, and elsewhere above the tree line. Permafrost is absent and the mean annual temperature is –0.9°C (Det Norske Meteorologiske Institutt 1949-1955). In this example, the snow patch persists only on the upper part of the 30-meter high valley side. Solifluction of the same type occurs for some 400 meters along the valley side and there is a conspicuous break of slope about two-thirds of the way up (see pl. 1 and fig. 1). This break of slope might be interpreted as remains of a glacio-fluvial terrace of the type commonly found in this area (Strøm, 1952). However, the rate of soil movement is great enough at present that it alone could be sufficient to produce the break of slope. The soil is till, with many stones and boulders, clay and sand lenses, in a silty-clay matrix. Depth to bedrock is not known but is greater than 2 meters.

Six resistance thermometers consisting of glass-encased thermistors built into simple probes of 1 to 2 meters length were placed at various depths in

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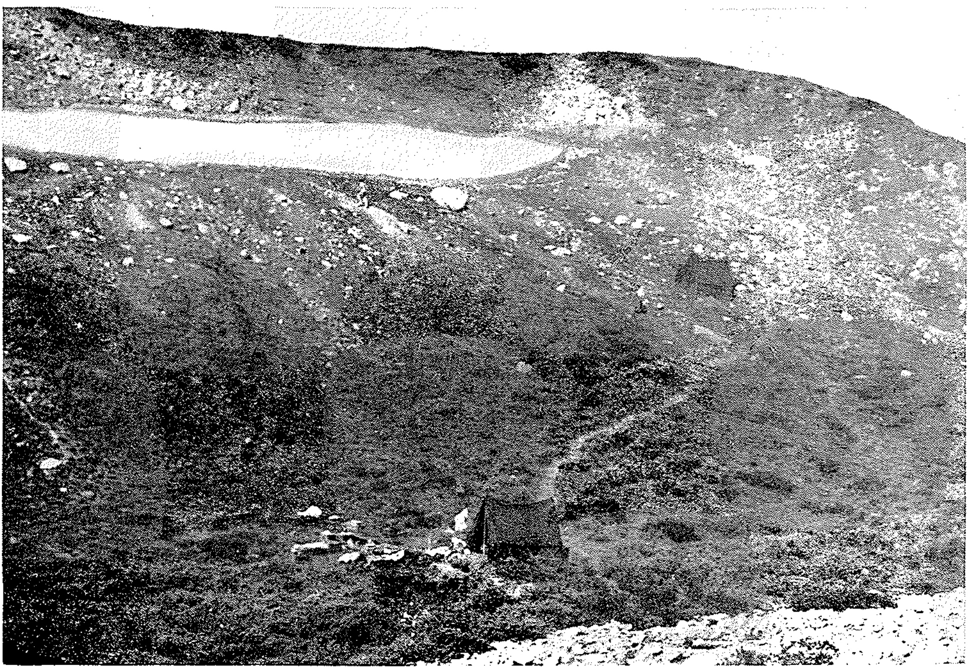
PLATE 1

A. Site of the solifluction investigations described. (photo: D.R.M. Lillistone)



B. The leads from the thermistors and soil movement probes were run to the tent seen at right center left. (photo: Kari Williams)

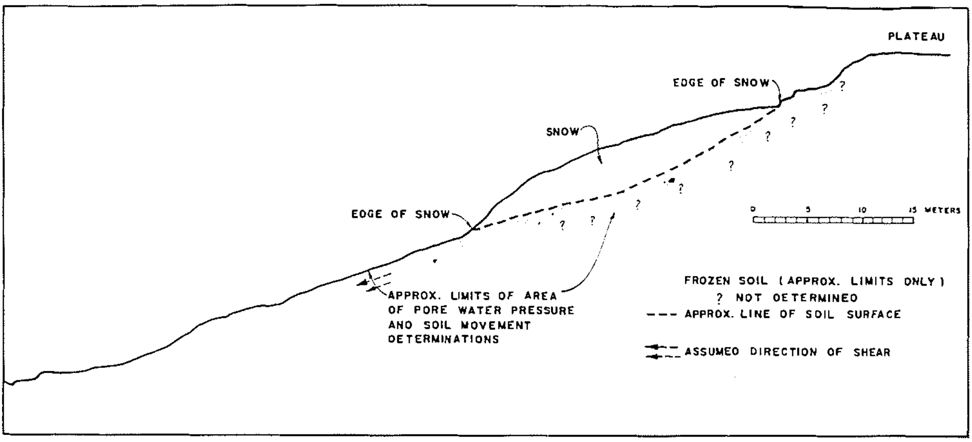


Fig. 1. Section through slope showing site investigated, late June condition. The external soil and snow surface is a true profile surveyed by Abney level, about June 20, 1957. The other data are based on various observations in 1956-1957.

holes drilled in the still-frozen slope (fig. 2). Eighteen mercury soil temperature thermometers were placed at various positions up to 40 centimeters deep in the soil. Results of the temperature recordings were limited but together with some probing, the general shape of the still-frozen soil mass was established (see fig. 1). The temperature of the still-frozen layer was virtually constant throughout (the thermistor probe sensitivity being about $\pm$ 0.05°C).

The snow accumulates in the winter because of wind blowing it from the plateau at the top of the slope. This plateau was observed to be free of snow in March, and the presence of stony earth circles—a form of patterned ground developing only where there is no snow cover—(Williams, in press) verify its general absence there.

INVESTIGATIONS INTO FACTORS INFLUENCING SOIL STABILITY

Theoretical background.—General explanations of solifluction involving concepts such as "lubrication of soil particles by water" are incomplete and often misleading. Terzaghi (1950) citing Hardy (1919) points out that water acting on quartz is not a lubricant. The study of processes in solifluction phenomena is a study of soil mechanics, and such special conditions are met with under freezing and thawing as to demand investigation of many basic soil properties.

According to soil mechanics theory, movement in soil occurs when the forces tending to produce shear (shear stress) exceed the strength of the soil (its resistance to shear). The latter is assumed to consist of two components, that due to internal friction and that due to cohesion. Cohesion in soils is ascribed severally to the properties of the adsorbed water in close proximity with the soil grains, the mutual attraction of particles, a grain structure inherent in clay soils, apparent cohesion which is due to the capillary forces in the pore-water in unsaturated soil, etc. According to the work of Bjerrum (1954), based on that of Hvorslev (1937) and others, in saturated soils the

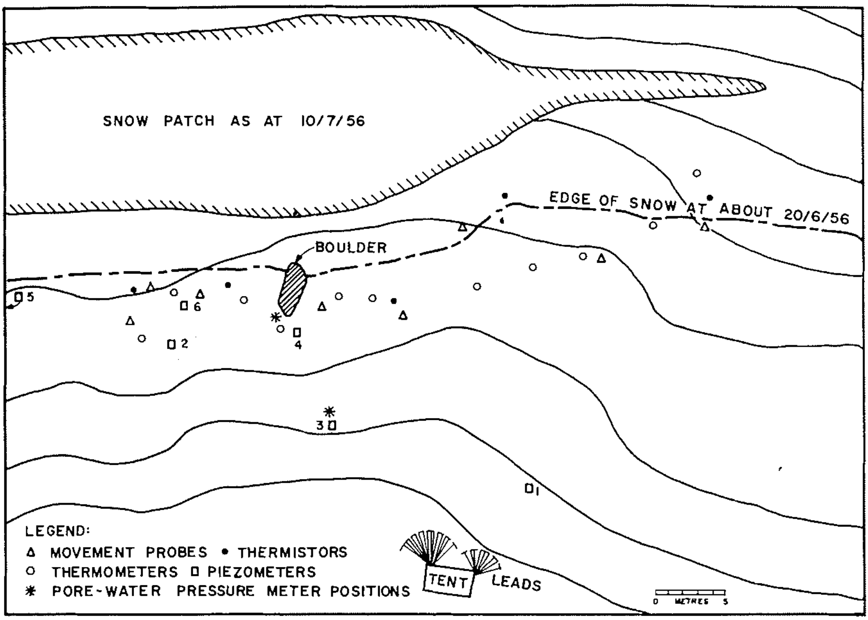


Fig. 2. Map of instrument layout on solifluction slope below late-lying snow (drawn from compass survey and field sketches). Contour interval—2 m.

magnitude of cohesion is dependent on water content and therefore on void ratio $= \left(\frac{\text{vol. voids}}{\text{vol. solid}}\right)$ The internal friction developed depends on the pressure at the intergranular contacts, $\bar{\sigma}$, which varies with U_w, the pore-water pressure, according to the relations $\bar{\sigma} = \sigma - U_w$ where σ is the total weight of wet soil above. Pore-water pressure, therefore, also directly affects soil strength.

Soil movement.—Soil movement was recorded directly by the use of special electrical probes described fully elsewhere (Williams, 1957b). Movement was also apparent through visible distortion of the probes and thermistor probes. In any one year, soil movement on the slope is largely restricted to several patches of a few square meters in area, although a large part of the slope is in an unstable condition. Maximum surface movement during the spring of 1956 was about 25 centimeters downslope. It is not possible to distinguish those parts actively moving, except by repeated observations. The movements occur at progressively greater depth as the season advances, and to a depth of at least 75 centimeters. At any given position, movements occur gradually, taking place over several weeks during the spring. Other than upslope from the snow patch, no distinct shear planes were visible. The movements recorded showed no regular or periodic variations of a diurnal or similar nature.

Effect of snow patch.—The late-lying snow patch in summer to some extent loads the slope, and the possibility of this increasing the shear stress must be considered. The initial solifluction movements will tend to remove this increment of the shear stress from the soil downslope of the snow (see fig. 1). It will only persist or be reapplied if there is also downslope movement of the still-frozen soil, which is unlikely. In other words, the weight of the snow patch is probably supported almost entirely by the still-frozen soil at least after the initial soil movements have occurred and has, therefore, little significance in causing movement. The movement is also substantially independent of the size of the snow patch which is steadily decreasing during thaw. The effects of the snow in causing movement are probably indirect.

Pore-water pressure determinations.—Pore-water pressure was determined at locations in the slope (fig. 2) with piezometers using simple mercury manometers attached to lengths of brass tubing with an internal diameter of about 7 mm, whose lower ends were at a depth of 50 to 115 cm. The brass tube was pushed into the thawed soil (in two cases, partially into holes drilled into still-frozen soil) while a rod was held in position inside the tube to prevent the entry of mud. Before attaching the filled manometer, the tube was filled with boiled water. Difficulty was experienced because of the rapid drainage from the tube into the soil. An electrical apparatus was also used, and confirmed the general nature of the pore-water pressures, although being far less sensitive.

Since water could be seen on the soil surface around most of the tubes and the soil was apparently saturated, it would be expected that the pressure of water (in centimeters of water) in the soil at any depth would equal approximately the depth to which the tubes were inserted, at least at those points. If complete drainage of excess water (above saturation point) did not occur, it might be greater than this. This latter condition (hydrostatic excess pressure in soil mechanics terminology) follows the transfer of some part of the weight of mineral soil to the entrapped pore-water. As explained above the greater the pore-water pressure, the smaller the component of soil strength due to internal friction. The pressures found, however, were considerably lower than expected, varying between –103 cm of water at 90 cm depth (sub-atmospheric pressures being expressed as negative) to +41 cm of water at 44 cm depth. About 250 readings, well distributed between these extremes, were taken at seven points, at intervals of half to several hours for periods up to three weeks. The lowest pressures (i.e. with the numerically highest, negative value) were recorded from depths in the proximity of the boundary between the thawed and frozen soil; surface water was also visible at these locations. Rhythmic diurnal changes of about 10 cm of water pressure were commonly observed. A gradual trend towards a minimum pressure and a subsequent rise was observed at most points during the three weeks, the total range at a single point being about 30 to 60 cm (see fig. 3; air temperatures for the period are shown in Williams, 1957b, fig. 3). Several more weeks elapsed before the soil had reached its driest summer condition. There was no clear relationship of pressure variations to rainfall.

These unexpected observations are of considerable significance and their causes are discussed below. The resulting effects are not easily interpreted. In general, lower pressures give greater strength by increasing internal friction, but even larger pore-water tensions are produced by drying in the summer

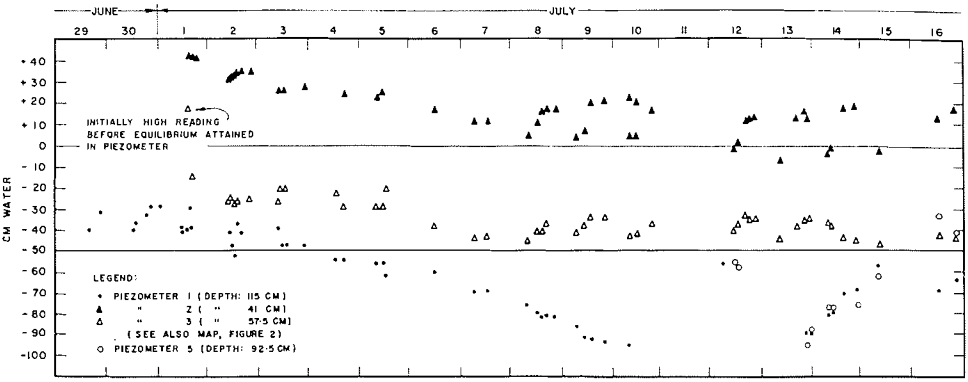


Fig. 3. Pore-water pressures at four locations in the slope.

(see Aitchison, 1957), so that the spring condition is to be interpreted as one of *relatively high* pore-water pressures.

It is often tacitly assumed that, in thawing of soil where considerable water accumulation has occurred at freezing, conditions of hydrostatic excess pressure, or at least a significant positive pressure, will arise generally and contribute in considerable measure to the instability. This is shown not to be the case. It is particularly tempting to believe that spontaneous liquefaction might occur (Terzaghi, 1950) where there is a sudden collapse or settlement of the soil grains with a transfer of a considerable part of their weight to the pore-water. However, failure to detect such high pore-water pressures in this slope indicates this to be of little significance in causing movement.

Frost heave.—Soils within the fine-sand-clay range, provided that sufficient water is available, show expansion at freezing several times greater than could be explained on the basis of the conversion to ice of the initial water content. The subject of much recent research, this expansion is due to accumulation of water during freezing, producing discrete ice layers, or individual crystals and masses which are often, but not always, visible to the naked eye (see for example, Beskow, 1935; Johnson, 1952). Such accumulation occurs as long as there is a sufficient supply of water available in the soil. The exact relation of frost heave to water availability is not fully understood but substantial heave occurs in the zone of capillary saturation. Recent work shows that heave may be caused by migration of water in soils that are only partly saturated (Penner, 1956). Most theories of frost heave attach great significance to the observed depression of the freezing point in soils which is generally supposed to be related to a considerable extent to effects of the surface forces of the particles, and hence to pore size.

Expansion per unit volume of the soil on this slope was calculated from density (gms. wet soil per cc) determined on soil samples from a depth of 5 to 20 cm. Using a modified form of the sand replacement method (West, 1953) determinations were made in the field in the spring (still-frozen and newly-thawed condition) and late summer, spread over two week periods. Subsequently, water contents were determined in the laboratory, permitting calculation of *in situ* dry densities (grams of mineral and organic soil per cc).

The ratio of average dry density for the spring series to that for the late summer series gives the contraction occurring during the summer, and is hence a measure of expansion due to frost heave.

Since the soil is heterogeneous, analysis of a number of samples from both periods is necessary. A maximum error of 6 percent is assumed possible in a single determination. The observations are summarized in the following table:

	No. of Determinations	Average Wet Density (grams wet soil per cc)	Average Dry Density (grams dry soil per cc)	Average Water Content (grams water per cc wet soil)
Year: 1956 Spring	14	1.74	1.37	0.37*
Late Summer	13	1.74	1.46	0.28
Year: 1957 Spring	10	1.75	1.30	0.45*
Late Summer	7	1.77	1.52	0.25

* The low water contents result from selecting some samples from the *thawed* soil, which had to be sufficiently drained to permit use of the sand replacement method.

The following conclusions can be drawn:

(a) Frost heave occurs in this slope. This is shown by the significantly different average values for spring and late summer dry densities and by the occasional observation of ice layers in the frozen soil. The slope is rather dry in late summer but the incomplete drainage due to a frozen layer at some centimeters depth during autumn freeze-thaw cycles, of rain or melt water liberated from early snow on the slope, may provide a source of water for the ice formation and frost heave.

(b) It is also established statistically that the heaved condition persists into the newly thawed state. A significance test on the difference between dry density in the spring (thawed soils only) and late summer of 1957, was based on the following data:

	Spring	Late Summer
M = samples =	7	7 (total samples taken)
$\bar{x}$ (average)	1.32	1.52
S (sample variance)	0.0451	0.0188

The number of samples, however, is insufficient to derive a value for the percentage expansion with any certainty.

(c) The *in situ* wet densities (unit volume weight) for spring and late summer periods are very similar. It is often stated that increased water content increases the likelihood of movement, because its weight adds to the shearing stress. In this case of heaved soil, there is little general increase in soil weight during the thaw period.

(d) In spite of the high water content, air filled some of the voids.[1] This was shown by specific weight determinations made on solid material from many of the frozen samples, enabling calculation of volume of voids per unit volume of soil. The latter is compared with water (ice) content.

An increase in the void ratio has several results in the thawed soil. Permeability is increased substantially; for example, Taylor (1948, p. 116) cites a 15 percent increase in void ratio of a cohesive soil, which increased permeability by ten times. Where ice-layer formation has occurred, the increase in permeability is likely to be primarily in a direction parallel to the ice layers, and hence parallel to the surface. It is not possible to get a quantitative value of permeability changes from calculations of void ratio in this case.

EFFECTS OF FROST HEAVE ON PORE-WATER PRESSURE AT THAW

That this increased permeability persists after thawing (although to an unknown degree) and is effective in a direction likely to be followed by melt water flow, partly explains the unexpectedly low pore-water pressures observed in this slope. Consolidation of the soil (the packing closer of grains) and liberation of water by thawing takes place only slowly and drainage of the water occurs sufficiently rapidly for pore-water pressures to remain low. Entry of air or surface water which would dissipate negative pressures is thought to be prevented because of the low permeability assumed to exist normal to the surface and the partial closure of capillaries by the snow mass or frozen soil higher up the slope.

A further explanation is supported by the work of Nersessova, as described by Tsytovich (1957). In her laboratory experiments, she found thawing of soil takes place progressively as the temperature approaches 0°C. Water is liberated according to temperature and the manner in which it is held in the soil, the most firmly held (bound) water thawing at the lowest temperature. The percentage of unfrozen water at a given negative temperature also varies with grain size composition.

Obviously the most firmly held water is that least likely to drain, but according to Tsytovich (1947) migrations of unfrozen water in frozen soil occur. The writer suggests that during the several weeks when the frozen layer in this slope has a temperature approaching 0°C it has a definite permeability, so that some drainage occurs over a long period prior to complete thaw. This will tend to lower the pressure in the water. It is well known that subatmospheric pore-water pressures are developed in the soil below the frozen layer, during freezing (see Penner, 1956). Water may then be drawn down from the frozen layer during partial thawing. The pressures may be further lowered by the volume decrease of the water (ice) taking place on the melting of some of the ice, while the expanded (heaved condition) of the soil is maintained by the longest remaining ice structures. These effects are likely to be accentuated by the heterogeneous nature of the soil and the range of pore sizes present. The existence of sub-atmospheric pressures in the unfrozen water of frozen soils was pointed out by Lovell (1957), and it is probable that these persist from the time when freezing occurred. To what extent the low pore-

[1] The term "void" includes all pores filled either with water or air.

water pressures observed should be related to each of these possible causes is not clear.

During the most active thaw period, therefore, the soil is in a unique condition. The water content is extremely high, and far above the normal saturation level, but because of phenomena associated with the thawing and frost-heaved soil, the pore-water is under low or sub-atmospheric pressures in much of the soil. Relatively high pore-water pressures must occur near the surface where water flows over it, and necessarily where upward flow is evidenced by soil particles washed out from the soil (see below).

EFFECT OF FROST HEAVE ON COHESIVE STRENGTH

The distribution of void space is unusual in thawing soils. It is clear that the numerous planes of discontinuity left by ice layers and the disruption caused by small ice masses and crystals represent a very severe disturbance and an increase in void ratio, that would greatly reduce cohesion phenomena. The complex phenomenon of strength loss following frost heave is as yet little studied. It is also possible, for example, that the desiccation of clay layers between discrete ice layers caused by the suction of the developing layer produces high shearing strengths within these layers (because of the extremely low pore-water pressures), but in most cases, the effect of such strength increases must be far outweighed by the loss of strength due to discontinuities in the soil mass.

Melt-water flow.—The evidence for high permeability introduces the possibility of melt-water flow within the slope occurring with enough speed as to introduce considerable movement of soil grains. Fine sand and silt particles were also observed being carried downslope in surface water.

GENERAL CONCLUSIONS

Solifluction of this type cannot be considered analogous to soil failures due solely to increase in water content over an impermeable layer.

1. The pore-water pressures in the thawed layer are generally far lower than would occur, could the soil be saturated to the same degree in an unheaved condition.

This results partly from the drainage and void ratio characteristics of the soil in the newly thawed condition; tensions are also characteristic of unfrozen water existing in the frozen layer.

2. A considerable part of the loss of strength at thaw in frost-heaved soil is due to the reduction in the cohesion component following the separation of particles by frost heave.

3. The shear stresses are not increased proportionately to the weight of excess water present under thawing conditions, since this is counteracted by the decreased dry density of the soil.

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