

Pleistocene Permafrost of West Siberia as a Deformable Glacier Bed¹

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ABSTRACT

Subglacial permafrost is usually referred to as a factor impeding basal glacial processes, and the deformation of soft substrata in particular. In West Siberian, widespread glacial disturbances present in permafrost of Pleistocene age suggest that frozen sediments, if clayey and/or icy, can readily deform, thus translating basal glacial stress into sliding of the entire glacier/sediment complex along subglacial shear zones. Ductile deformations such as folds and diapirs are also widespread. The mode of deformation of frozen subglacial sediments is dependent on their lithology, ice content and temperature conditions. Signatures of former subglacial permafrost in currently thawed sediments are deduced from contrasting deformation behaviours of lithologically different sedimentary formations.

RÉSUMÉ

Le pergélisol sous-glaciaire est d'habitude considéré comme un élément entravant les processus à la base du glacier et en particulier la déformation d'un substrat meuble. En Sibérie occidentale, de nombreuses perturbations glaciaires affectant le pergélisol pléistocène suggèrent que les sédiments gelés, s'ils sont argileux et/ou contiennent de la glace, peuvent être déformés facilement, par l'effort affectant la base du glacier en des glissements du complexe glacier/sédiment le long de zones de rupture sous-glaciaires. Des déformations ductiles telles que des plis et des diapirs sont également nombreuses. Le mode de déformation des sédiments sous-glaciaires gelés dépend de leur lithologie, de leur contenu en glace et des conditions de température.

Des indices de l'existence d'un pergélisol sous-glaciaire peuvent être trouvés dans les variations des déformations que les différentes formations ont subies selon leur lithologie.

KEYWORDS: subglacial permafrost; glaciotectonism; buried glacier ice; till genesis; Pleistocene

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INTRODUCTION

Subglacial permafrost is usually controlled by the thermal zonality of ice sheets and greatly affects glacial erosion and deposition (Sugden, 1977, 1978; Moran *et al.*, 1980; Hughes, 1981). It is often thought that glacial deformation of frozen rocks is hindered by their higher shear strength (e.g. Wateren, 1985). If glacial stress is high

enough to overcome the frozen rock resistance, the frictional heat released precludes the existence of subglacial permafrost (Hart *et al.*, 1990). As a result, glaciotectonism in frozen rocks is considered possible, but unlikely (Hart and Boulton, 1991). Formation of thick till sheets is also thought not characteristic for frozen glacial beds. However, many years of field experience in northern Russia, where glacial drift is up to

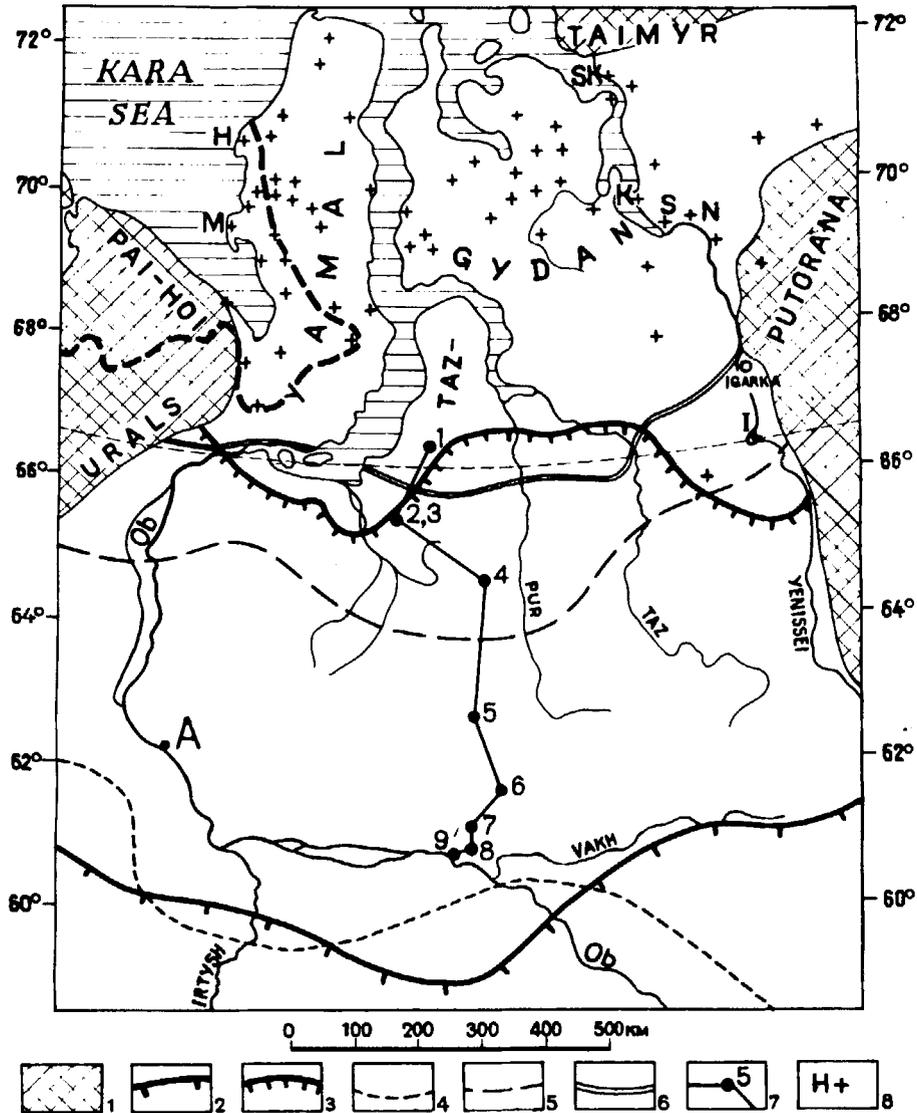


Figure 1 Location map with principal geocryological phenomena. (1) Palaeozoic uplands. Limits of former glaciations (Arkhipov *et al.*, 1986; Astakhov, 1992); (2) Middle Pleistocene. (3) Weichselian. Thick broken line outlines the Baydarata Ice Stream. Limits of present permafrost zones (Yershov, 1989); (4) Relict permafrost at depth. (5) Discontinuous permafrost in mineral near-surface substrate. (6) Continuous permafrost. (7) Borehole profile in Figure 2. (8) Stratiform bodies of massive ground ice (according to Badu *et al.*, 1982; Grosswald *et al.*, 1985; and observations by the authors). Locations mentioned in the text: A, Atlym; H, Harasavey; I, Ice Hill; K, Karaul; M, Marresale; N, Nikitinsky Yar; S, Selyakin Mys; SK, Sopkarga.

300 m thick and there are widespread deformations of soft bedrock, suggest a different conclusion.

Most of the current ideas in the international literature have been derived from geological data on former ice sheets developing either on hard bedrock or in sedimentary basins of temperate climate regions. A different case is the world's largest sedimentary basin of West Siberia, half of which is still perennially frozen (Figures 1,2). It provides an opportunity to better understand the geological consequences of the glacial ice/permafrost interaction during the Ice Age.

This opportunity stems from the well established fact that the present permafrost of West Siberia is principally a Pleistocene phenomenon at various stages of its adjustment to the Holocene climate. South of 59°N it has completely thawed, but in the northern boreal forests it exists beneath a 100–200 m thick layer of thawed sediments as an isothermal zone of sub-zero temperatures up to 150–250 m thick. Close to the Arctic Circle the top of the relict layer rises to coalesce with the modern permafrost, c. 50 m thick, formed in the Late Holocene. Beyond the Arctic Circle, in the tundra regions with mean annual air temperatures below -8°C , small ($1\text{--}2^{\circ}\text{C}$ per 100 m) temperature gradients appear in continuous permafrost, 300–550 m thick (Yershov, 1989).

The Pleistocene age of much West Siberian

permafrost is evident from the existence of old ice wedges, buried or truncated by Holocene deposits (Yershov, 1989), and from relict glacial ice abundant in the surficial till (Kaplyanskaya and Tarnogradsky, 1976, 1986b). Whereas the upper member of the two-layered permafrost of the sub-Arctic (Figure 2) may be ascribed to the cooling of the last 3000 years, the lower layer must be older. It is a Pleistocene relict, modified by Holocene climate, even if it occurs on the surface directly beneath the present active layer, as at the 'Ice Hill' on the Yenissei (I in Figure 1). At the latter, non-finite radiocarbon dates from alluvial and ablation sediments, overlying the surficial till containing large stratiform bodies of glacial ice, indicate that the permafrost has survived at least since the Early Weichselian (Astakhov and Isayeva, 1988). The relict layer, which occurs in the area of the maximum glaciation (Figure 1), may even have formed in the Middle Pleistocene (Zemtsov, 1976).

These data are better understood if the high thermal inertia of the Meso-Cenozoic sedimentary rocks, 3–4 km thick, is taken into account. These are 35–40% marine clay. Geophysical evidence is provided by deep boreholes in which observed geothermal gradients are distorted by 40–60% compared with normal ones (Figure 3). The latter occur only deeper than 3–4 km. Upwards in the succession the gradients increase in the northern glaciated part of the basin and

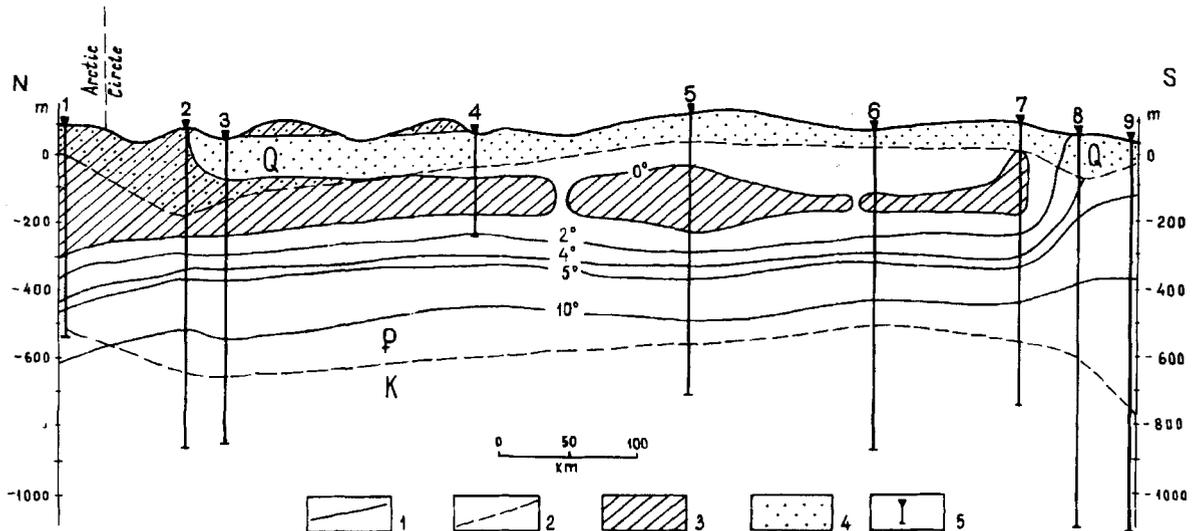


Figure 2 Temperature profile along the axis of the West Siberian basin (after Balobayev *et al.*, 1983 with near-surface permafrost added). For location see Figure 1. (1) Isotherms, (2) stratigraphic boundaries, (3) perennially frozen sediments, (4) Quaternary deposits, (5) boreholes.

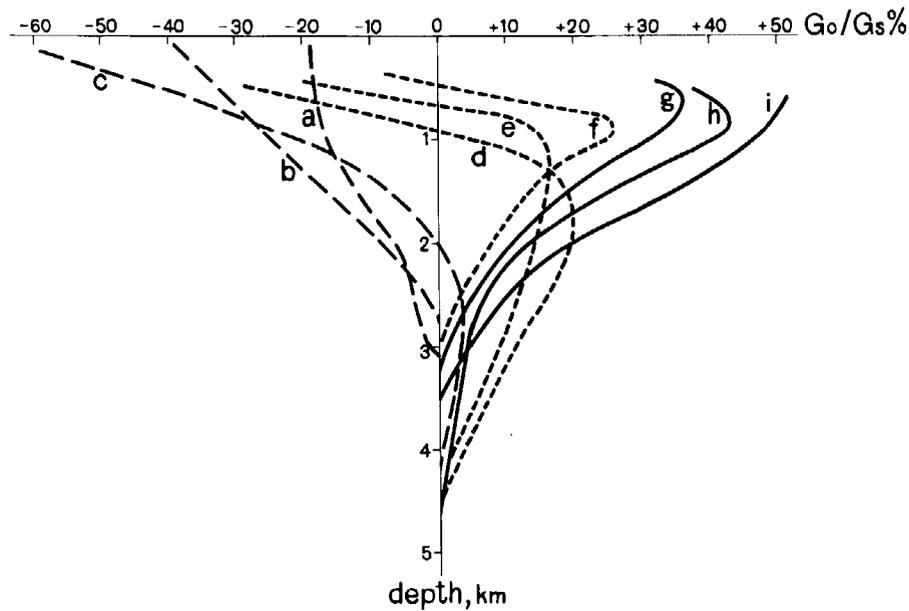


Figure 3 Deviations of observed geothermal gradients (G_o) from stationary gradients (G_s) in boreholes of West Siberia at latitudes: (a) 56° , (b) 58° , (c) 60° , (d) 62° , (e) 64° , (f) 66° , (g) 68° , (h) 70° , (i) 72° (after Kurchikov and Stavitsky, 1987)

decrease in the south. All soft Meso-Cenozoic rocks are cooler by $25\text{--}30^\circ\text{C}$ than expected for stationary conditions (Kurchikov and Stavitsky, 1987).

The distorted gradients, reflecting the climate history along the north–south profile (Figure 3), can be clearly subdivided into three sets according to palaeogeographic zonality. The reduced gradients in the southern periglacial zone (curves a, b and c) are explained by interglacial and Holocene heating from above which led to thaw of the Pleistocene permafrost. Curves d, e and f, belonging to the zone of Middle Pleistocene glaciation, show reduced gradients only in the uppermost part of the sedimentary column, where the relict Pleistocene permafrost still persists (cf. Figure 2). The increased gradients at depths greater than $0.4\text{--}0.9$ km are traces of the Pleistocene cold wave. The gradients in the Arctic regions with permafrost increase northwards (curves g, h and i) in conjunction with the progressively colder environment. It is noteworthy that curves g and h show a slight decrease in the uppermost layer, c. 0.5 km thick. This agrees with the geological evidence on the present permafrost in the Arctic, which, being essentially of Pleistocene age, was much colder prior to the Holocene. Curve i relates to the high Arctic, where the Holocene heating was minimal.

Judging by the depth of distortion of the temperature field of West Siberia (Figure 3), the

accumulation of the cold started long before the maximum glaciation; interglacial warmings were too short for restoration of the normal temperature gradient. Another implication is that the Pleistocene ice sheets, operating in this sedimentary basin, were predominantly cold-based, as Hughes (1985) has suggested.

The northern part of the West Siberian Basin is characterized by numerous displacements of the subsurface Meso-Cenozoic strata. They have been described from everywhere within the drift limit (Rostovtsev, 1982). These innumerable folds, imbricated fold structures, thrust sheets, and diapiric protrusions form arcuate tracts up to $10\text{--}25$ km wide and up to 200 km long. They are accompanied by erratic blocks of soft sediments of various sizes. Soft surficial rocks are almost entirely reworked by epidermal tectonism over large areas of northern West Siberia.

Thus, in West Siberia large-scale glaciotectionic disturbances coexist with old permafrost. One may argue that the existence of Pleistocene subglacial permafrost, deduced from the present temperature gradients, and the ubiquitous deformations of the glacier bed, are incompatible. However, geological observations on the structures formed beneath the former ice sheets definitely indicate that glaciotectionic disturbances in West Siberia developed mainly in perennially frozen sediments. Indirect evidence comes from

the area south of the Arctic Circle, where previous subglacial permafrost can be inferred from sedimentary and tectonic structures analogous to those in the Arctic. These observations show that in West Siberia the ice sheets eroded mostly by glaciotectonism. Also, they deposited mostly by accretion of stagnant debris-laden sheets of basal ice, building up the subglacial permafrost.

GLACIAL DISTURBANCES WITHIN PRESENT PERMAFROST

The upper part of the present permafrost in Arctic West Siberia contains a variety of structures, originating from the base of the Late Pleistocene ice sheet, and still containing remnant glacial ice. The basal ice/sediment complex, studied since the 1970s (Kaplyanskaya and Tarnogradsky, 1976), includes ice-flow features as glaciotectonites with streamlined casts of frozen sediments, changing laterally into normal matrix-supported diamictons of basal till.

The Western Yamal Peninsula

An example of complicated subglacial deforma-

tions has been studied on the western Yamal Peninsula coast near Marresale Polar Station (M in Figure 1). Permafrost is 120–300 m thick here (Yershov, 1989), versus 300–500 m in northern Yamal and further east. The surface temperature is -5 to -8 °C. The permafrost is underlain by a layer, 80–150 m thick, with negative temperatures but unfrozen owing to the high content of soluble salts.

The oldest exposed sediments are laminated clayey silts and silty clays with seams of fine sand and plant detritus, labelled the Marresale Formation (Figures 4,5). Their salinity is 0.4–0.6 wt% (Gataullin, 1991). They do not contain visible ice inclusions, though the moisture content may attain 33 wt%. Gataullin (1991) maintains a deltaic origin and Eemian age for this formation. The Labsuyaha sand, up to 50 m thick, infills wide channels incised into the Marresale silt. The sand contains about 38% of interstitial ice (Gataullin, 1988). There are two radiocarbon dates on driftwood from the sand: $31,000 \pm 400$ and $42,000 \pm 1000$ (Bolikhovsky *et al.*, 1989).

These interstadial deposits are overlain by the Kara till up to 20 m thick (Kaplyanskaya and Tarnogradsky, 1982; Gataullin, 1988). The till occurs in two facies. The first facies consists only of apparently local material: clasts of the

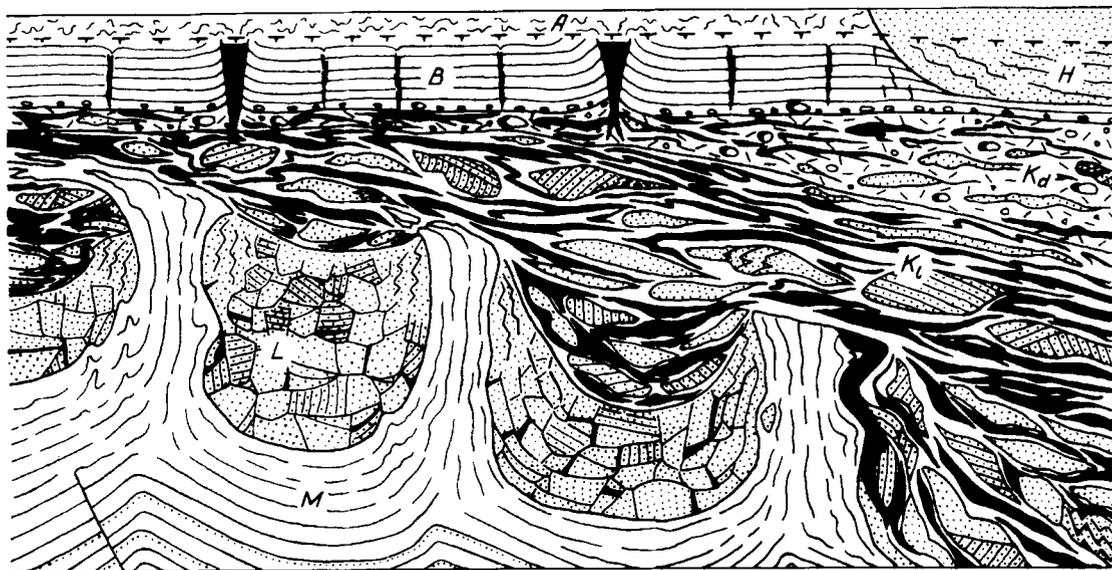


Figure 4 Sketch of sedimentary structures exposed in coastal bluff 30 m high at Marresale, Yamal west coast (M in Figure 1). M, clayey silt of the Marresale Formation, folded and occasionally faulted, in upper part subject to diapirism; L, brecciated Labsuyaha sand; K₁, Kara till, local facies; K₂, diamict facies with foreign material; B, Baydarata sandy silt with large ice wedges and minor syngenetic ice veins; H, Holocene limnic sediments with collapse structures; A, active layer. Thin lines are sedimentary and deformation structures; thick lines and blackened areas are ice; sands are shown by dots.



Figure 5 Folded clayey silts in coastal cliff 17 m high, north of Marresale, western Yamal (M in Figure 1). The hinge parts are pulled out by diapirism to make small tight folds with axes oblique to axial planes of the major folds. Sand seams look white in photo.

Labsuyaha sand suspended in the matrix derived from the Marresale silt, or vice versa. The second facies is a diamicton with clasts of local fine-grained material and an admixture of erratic pebbles. When thawed, both facies, as observed in the seasonally thawed layer at the surface of the coastal cliff, are not appreciably different from analogous sediments occurring in the non-frozen zone of West Siberia. In deeply incised thermocirques, or in special clearings made to remove the present day active layer, the till facies appear primordially frozen (Kaplyanskaya and Tarnogradsky, 1976, 1986a) without any trace of subsequent thawing.

The ice content varies widely and may reach 80%, with layers of almost clear ice up to 1 m thick being visible. The icy till has a striped appearance owing to alternating bands of clear and dirty ice. The bands form deformation structures typical of debris-rich ice of modern glaciers (e.g. Lavrushin, 1976; Lawson, 1979). These same deformed banded structures are clearly seen in the thawed till on cliff headlands, although such steep slopes are unfavourable for preservation of original structures in the sediments.

Formerly thawed but subsequently refrozen till, 5–6 m thick, also displays well preserved

structures of glacier flow when found beneath former thermokarst ponds. This refrozen till now contains much less ice as compared to the primordially frozen till. The two varieties of the Kara till are separated by sharp thaw boundaries. On the contact of the primordially frozen till with overlying subaerial sandy silt there is another, thinner (30–50 cm) horizon of low ice content, which must be a former active layer.

Thin (in one case up to 4 m thick) lenses of ice-poor diamicton with traces of gradational bedding sometimes overlie the basal till and may be interpreted as flowtills. At the northern end of the exposed cliff small depressions on top of the basal till are filled with varved clay.

The described sediments are mantled by the Baydarata sandy silt, 1.5 to 6 m thick. Owing to its banded appearance, this is often interpreted as a fluvial-limnic deposit. Syngenetic ice wedges, only 0.5 to 1 cm wide, and large epigenetic ice wedges, up to 5–6 m high, are numerous in this sediment. The wedges, partly penetrating into the underlying till, differ from the till ice in their vertical banding and yellowish colour. Upright weeds and roots are radiocarbon dated to 13–14 ka (Gataullin, 1988; Bolikhovsky *et al.*, 1989). The weeds *in situ*, the lack of clay seams and the

mantling occurrence of the Baydarata silt on diverse topographic elements point to it being aeolian loess, analogous to the Yedoma Formation of East Siberia and Alaska. The Marresale succession is topped with various Holocene formations, limnic fills of thermokarst sink-holes and fluvial sediments.

Many investigators, including two of the present writers (Kaplyanskaya and Tarnogradsky, 1982), have studied glaciotectonic disturbances in the coastal cliff for 30 km north and 3 km south of Marresale. The cliff cuts into an elongated, slightly sinuous bulge 50–60 m high, striking NW–SE. The most complete geological profile along the entire cliff was obtained by Gataullin (1988, 1991). North of Marresale it reveals multiple open and tight folds accompanied by faults, embracing only the Marresale silt and Labsuyaha sand. The strike of the fold axes and faults is 100–120° SE. The younger formations, truncating the folds at a sharp unconformity, are very thin, or absent at the crest of the bulge and increase in thickness towards adjacent low terrain. The wavelength of the folding is 100–200 to 500 m, the apparent amplitude being 10–20 to 50–80 m. The limbs of the major folds are complicated by disharmonic subsidiary folding and buckling of sand seams. The faults, mostly thrusts, and axial planes of the folds, all dip south by south-west and indicate that a horizontal stress was applied from an ice lobe of the Baydarata Ice Stream (Astakhov, 1979, 1992). The entire folded structure may be perceived as an anticlinorium, most of which has been removed by erosion (Gataullin, 1988).

The Kara till attains 12 m in thickness, and even more if folded, on the southern flank of the anticlinorium, where it is exposed in a cliff 3 km long at Marresale itself (Kaplyanskaya and Tarnogradsky, 1982, figure on p. 78). The description below relates to this section which displays structures of different morphological types and generations, as well as various ground ice structures (Figure 4).

First, large open folds with minor thrusts occur in the clayey sediments of the Marresale Formation. They are preserved in the lower part of the cliff.

A second type of deformation results from the squeezing up of vertically protruding clayey sediments of the Marresale Formation together with subsidence of the overlying Labsuyaha sand and lower layers of the Kara till. Some of the protrusions are mushroom-shaped diapirs, evident

from their overhanging lateral contacts being closer to the top. The apparent maximum height of the diapirs is approximately 20 m. In plan they are elongated with the axes striking 165° SE. Their inferred width is 10–20 m, possibly up to 50 m. The planar bedding of the clays involved becomes progressively more contorted upwards, merging into tectonites with subvertical flow structures, in which the original bedding is almost indiscernible. The boundary between the gently folded clay at the cliff base and the diapirs is not a sharp unconformity, as was erroneously shown in the figure by Kaplyanskaya and Tarnogradsky (1982), but is a transitional zone in which strata from the hinge parts of the major anticlines are gradually extended (Figure 5).

The subsidence basins between the diapirs are filled with the Labsuyaha sand, the upper part of which is often transformed into a sandy till. The sands are brecciated and consist of randomly oriented angular blocks up to several metres across. At the margins of the basins the sands are sometimes broken by normal faults into subparallel slabs tens of metres long, inclined towards the basins lows. The blocks mostly retain the original bedding of the sand, although subvertically oriented in places. On the whole, the observable deformations of the sand are almost exclusively brittle. Only in one case was the sand conformable to an overhanging diapiric contact, making a fold-like structure.

Removal of surficial thawed material in one intradiapir basin has revealed thin, 0.5–4 cm thick, ice veins coating the edges of the sand blocks. Within the layer of seasonal thawing these veins are replaced by fine-grained mineral material. The veins are probably a result of partial friction melting of the interstitial ice during the deformation with subsequent regelation.

In a gully, 0.3 km south of Marresale, a diapiric overhang has been observed in both a thawed and a perennially frozen state (Figure 6a, b). Both cases demonstrate the contrasting deformational behaviour of clay versus sand. Veins of clear transparent ice, probably again of friction-regelation nature, are contained in extension fractures of the sand close to the diapir wall. Apart from this, ice seams, up to 40 cm thick and subparallel to the sand bedding, have been found in some sand blocks. These may be segregation ice veins developed in the Labsuyaha sand prior to its deformation.

The local sandy till contains more ice, sometimes in the form of thick layers and lenses, than

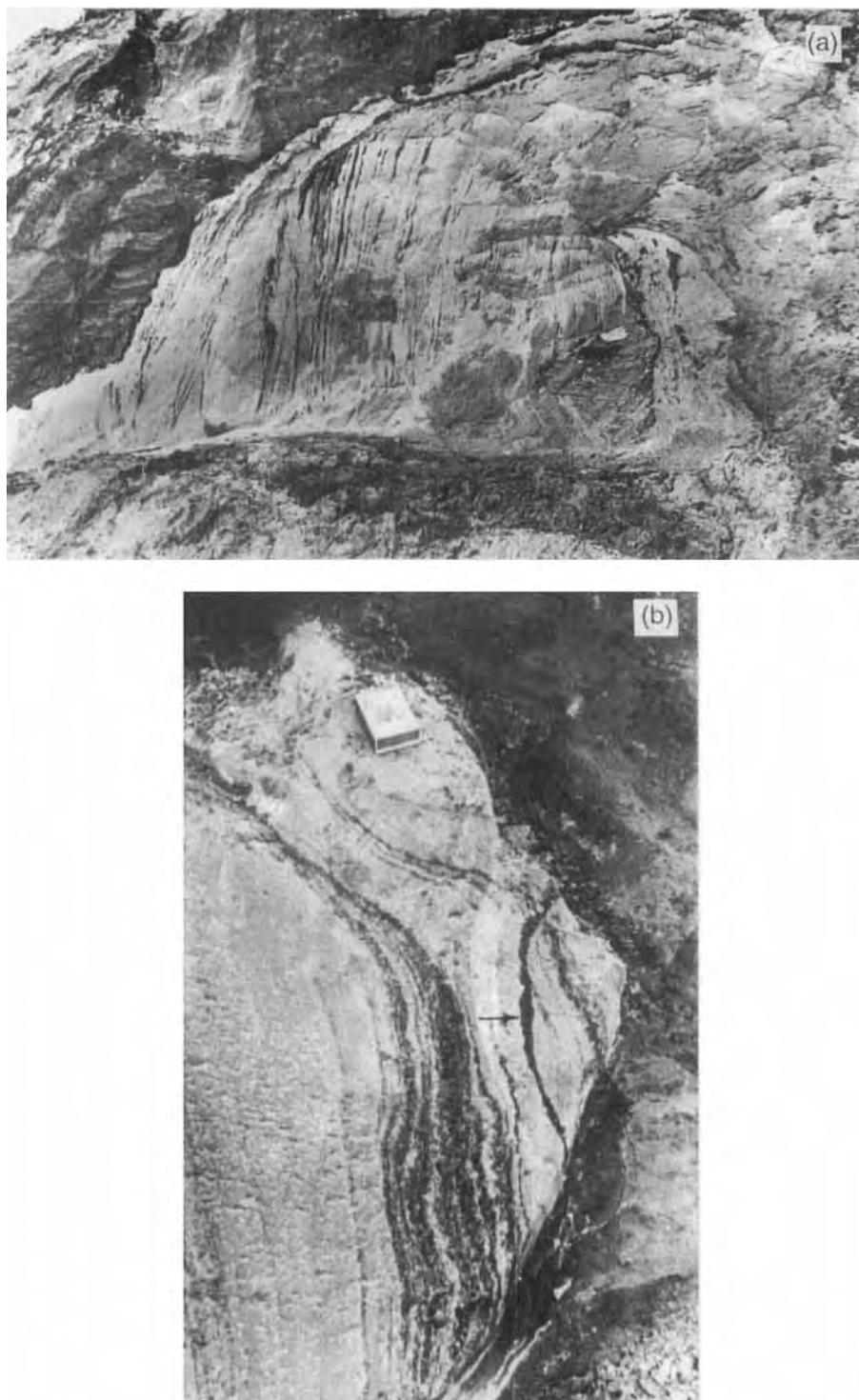


Figure 6 Diapir contact, 0.3 km south of Marresale, 15 m ASL, the western Yamal. (a) Diapir mushroom (dark clay) hanging over brecciated sand with variously dipping remnants of sedimentary bedding; clearing 15 m high viewed from south. (b) Detail of the same viewed from north (matchbox for scale); subvertical stripes are sedimentary bedding in a sand block; arrow is an extension crack filled with transparent ice which separates a slightly displaced fragment of frozen sand.

the parent fluvial sand. Because of this, it shows only ductile (fold) deformations resulting from subsidence. One such ice layer, more than 8 m long and 1 m thick, is folded and dips steeply and conformably to a diapir wall (Figure 7). This was observed 1.5 km south of Marresale (Tarnogradsky, 1982). The ice has been identified as buried glacial ice on account of its glaciodynamic foliation and angular foreign clay debris. The difference between the local till and its parent Labsuyaha sand is easily seen in the perennially frozen state, but when thawed the boundary between the formations is indistinct.

Glaciodynamic structures, typical for the basal layers of glaciers (Lavrushin, 1976, 1980), constitute the third type of deformation in this section. They are best seen in the upper part of the sequence, where both local and diamict facies of the Kara till are horizontal, truncating the tops of diapirs. These small-scale structures of permanently frozen icy till consist of alternating bands of debris and ice (Figure 7b), gently dipping shear planes, rounded sandy clasts, recumbent flow folds, crenulations, and so forth. The crenulated, dirty ice looks the same as that observed by French and Harry (1990, Figure 3A) in the buried glacial ice of Banks Island, NWT, Canada. The form of the recumbent folds and their noses indicate an ice flow to the NE (Gataullin, 1990; see also Figure 1).

Distortions of the original glaciodynamic structures by ice wedges penetrating the till may be considered a fourth type of disturbance. Also, collapse structures can be observed in the Holocene limnic sequences of lakes formed on top of icy sediments.

The stages and depths of deformation distinguished in this section reflect the varying response of frozen sediments to the stress and temperature fields caused by changing thicknesses and flow patterns of the Baydarata Ice Stream. The lead factor was probably advection of warm basal ice flowing from the deepest part of the western Kara shelf; this produced reverse temperature gradients in the subglacial permafrost of the Yamal Peninsula.

Thus, the recent glacial history of the site must have commenced with low temperature permafrost formed immediately after deposition of the Labsuyaha sand. The clayey sediments of the Marresale formation, heavily loaded by a subsequent ice advance, flowed laterally to form the folded compression bulge along the ice margin. Its proglacial position makes a frozen state of the folded sediments very likely. This is also indi-

cated by the generally brittle deformations of the Marresale sand. Occasional ductile deformations in the sand are probably due to a low strain rate and/or locally high ice content. The incompetent behaviour of the clay is explained by its persistent plasticity at a wide range of negative temperatures (Tsytovich, 1973), which in this case is enhanced by the high salinity of the clay.

Later, an encroaching ice stream overrode the proglacial 'forebulge'. The warmer basal ice deformed more readily than the frozen bedrock. The debris-laden basal ice is partly preserved in the intradiapir basins as frozen local till with large clasts of Labsuyaha sand. This must be a product of plucking which never occurs under sliding, wet-based glaciers. The frozen Marresale clay might have participated in basal motion, but only involved a thin subglacier layer where plasticity was greatest.

At the end of the first ice advance the insulation by glacier ice and the release of deformation heat would have increased the temperatures of the frozen bed, at least in the upper 20 m. The diapiric structures are the result of this warming and associated reduction of shear strength of the clay, whereas the sand stayed solidly frozen. Their deformation in a frozen state is evident from the ice layers, seams and veins in the surrounding sand, which sank into the intradiapiric depressions along with dirty basal ice. The straightforward diapirism without any drag features suggests a temporary stagnation of the ice lobe. The diapirs may also have originated at the intersections of large cracks in the stagnant ice and on the crests of anticlines. Crevasses are the normal cause of subglacial diapirism which is usually associated with water-soaked sediments (e.g. Sharp, 1985) but we have no indications of wet-based ice conditions at Marresale.

A second influx of basal ice at near-zero temperature restored the erosive activity of the ice stream with a resultant sharp unconformity between the diapirs and overlying frozen till with glaciotectionic banding. The diapirs survived mostly south of Marresale, but in other places along the cliff only the roots of the diapirs are visible. The extension strain of moving ice was concentrated in its basal layer, which left the previous structures below the tops of the diapirs almost undeformed.

The final stagnation of the ice stream was followed by melt of the majority of the clean ice. The debris-rich basal ice survived, protected by the cold permafrost beneath and by a thin layer



Figure 7 Bed of glacier ice (pitted surface) plunging along with local sandy till into an intradiapir basin, 1.5 km south of Mar-resale, the western Yamal. (a) General view; shovel 1 m long is at a vertically oriented block of stratified sand. (b) Large scale view of the downward continuation of the ice bed; ice bands with different debris content and melt-out fragments of clay are visible; the section illustrated is 1.5 m high.

of melt-out debris above. The subsequent accumulation of loess with ice wedges has enhanced the protective quality of the ablation mantle. Holocene thermokarst produced only localized melt-out of the ice/debris mixture and many melt-out patches of the originally icy till refroze again at the end of the Holocene.

The above reconstruction of deformation events is not necessarily accurate. However, two inferences are certain. First, the large disturbances imposed by glaciers or perennially frozen sediments result in all major types of glacial deformations – proglacial, subglacial and dead-ice related (Hart and Boulton, 1991). Second, the preservation of ice-flow structures in melt-out but originally frozen tills was inferred long ago by Lavrushin (1969, 1976) from a comparison of ancient tills with active basal ice. In the western Yamal Peninsula the same phenomenon may be observed *in statu nascendi*. It is clear that the basal till there could not be deposited by particle-by-particle lodging but only via stagnation of entire layers of dirty basal ice with their incorporation into pre-existing permafrost.

Late Pleistocene till is also exposed farther north in coastal cliffs mapped in detail by Gataullin (1984). At Harasavey settlement (H in Figure 1), and at a locality about 20 km north, it forms a discontinuous layer 1–2 to 15–20 m thick (Kaplyanskaya, 1982; Tarnogradsky, 1982). In its lower part the till is primordially frozen with varying ice content. The ice occurs in bands and laminae participating in small and generally planar glaciodynamic structures. The upper part of the till has reticulate and massive cryogenic structures and is obviously refrozen after melt-out beneath small water bodies. The till is again underlain by glacially eroded Middle Weichselian Lapsuyaha sand, which lies mostly below sea level. The visible, upper part of the sand is very ice-rich. In places, ice prevails with the sand particles suspended within it.

The icy sand and sandy ice form multiple diapiric protrusions, only the tops of which are observable in the cliffs. The rest of the piercing ice/sand mass has been studied by boreholes (Krapivner, 1986). The sand-ice strata dipping divergently from the diapir axes are contorted into small isoclinal folds showing upward flow direction. In places, the cores of the diapirs are built almost entirely of pure ice (Figure 8). The tops of the diapirs are glacially reoriented in a subhorizontal plane with many complicated re-fold and drag structures (Figure 9).

The ice of the Harasavey diapirs is poor in water soluble salts (0.1 to 0.19 g/l) and is thought to be segregation ice formed from glacial melt-water (Tarnogradsky, 1982). Probably here (and at Marresale) the lower boundary of the permafrost due to insulation by glacial ice before its stagnation extended into the Lapsuyaha sand. Below the permafrost boundary the sands were water-soaked and constituted an aqueduct for pressurized subglacial water discharging from beneath the central wet-based part of the Baydarata Ice Stream. With thinning of the stagnant glacier the frost front moved downwards. This, coupled with the influx of pressurized water, facilitated the formation of abnormally large volumes of segregated ice. The resultant ice/sand sandwiches, loaded by fractured stagnant ice, produced the present icy diapirs. Then the re-advancing glacier reoriented the diapir caps and deposited the overlying till.

At Harasavey, as well as at Marresale, exposures show that the melt-out sand, even if formerly very icy, retained its original intricate structure (Figure 10).

The Lower Yenisei Region

Traces of subglacial permafrost also occur in the Lower Yenisei region, where permafrost is up to 560 m thick and has surface temperatures -7 to -9 °C (Yershov, 1989). The type locality is Selyakin Mys (S in Figure 1), where a network of large, partly syngenetic ice wedges (Figures 11, 12) and small tabular pond ice bodies (Solomatin, 1986) have survived under epigenetically frozen Zyryanka (Weichselian) till, up to 10 m thick. In most cases the tills are primordially frozen and contain well preserved glacial ice (Figure 13) up to 60 m thick (Kaplyanskaya and Tarnogradsky, 1976, 1993; Solomatin, 1986; Karpov, 1986; Astakhov and Isayeva, 1988). The tills are Late Pleistocene in age, although preservation of older primordially frozen tills within deep (up to 300 m) buried troughs is not excluded.

The till sheets with occasional inclusions of buried glacial ice are normally represented by icy diamict beds intercalating with chaotic assemblages of balls, angular blocks and giant rafts of stratified sand and clay containing Quaternary and Cretaceous marine fossils. The matrix of these 'glacio-mixtures' (cf. glaciomélangé, Komarov, 1987) is partly diamict with ice and erratic stones, partly tectonically deformed clayey



Figure 8 Ice/sand complex steeply dipping in the core of a diapir, at Harasavey, western coast of the Yamal (H in Figure 1). Formerly ice-cemented contorted sand is seen in the seasonally thawed layer in left upper corner. The stick is 1.2 m long.

sediments. The sand clasts are mostly of marine sediments with interglacial (boreal) fauna or of stratified drift. The clayey clasts, commonly consisting of marine sediments, often contain Arctic fauna. The fauna are the reason why the entire succession is perceived by some geologists (e.g. Bryzgalova and Bidzhiyev, 1986; Zarkhidze *et al.*, 1991) to be a marine formation *in situ*.

However, most of these sedimentary inclusions are foreign to the area and must have been delivered by glaciers from the Kara Sea shelf and adjacent lowlands (Kaplyanskaya and Tarnogradsky, 1975). The predominant north-south direction of Middle and Late Pleistocene ice streams from the Kara Sea ice dispersal centres is amply indicated by the pattern of ice-pushed

ridges, till fabrics, striations, mineralogy of tills, etc. (Astakhov, 1976, 1979). Deeply penetrating glaciotectionic disturbances of both Quaternary and pre-Quaternary sediments have long been known in the region (Troitsky, 1975).

The structure of drift terrain built of tills and glacio-mixtures is not completely understood yet. However, most known glacio-mixture assemblages constitute folded and imbricated structures. Such formations imply at least two stages of subglacial deformation. The first is a longitudinal extension characteristic of non-marginal glacial transport. The second stage is longitudinal compression typical of ice margins. Both types of deformation must have occurred in a frozen state because: (i) sand blocks could have been



Figure 9 Upper part of the diapir in Figure 8 which has been subhorizontally reoriented by glacial stress. The protruding ledges are almost pure ice contained in icy sand; the section is about 3 m high.

detached only when the parent sand was frozen solid; and (ii) icy beds participate in the compression structures.

Folded glacial ice in a marginal position has been described at the 'Ice Hill' (Astakhov and Isayeva, 1988, location I in Figure 1). Upthrust and contorted debris-rich ice (Figure 14) is also part of a glacio-mixture occurring within an imbricated stacking of various Quaternary sediments about 50 m thick at Karaul (K in Figure 1). In addition, 2–5 m thick slices of diamictons and Kazansevo (Eemian) sands with shells of boreal molluscs, dipping west at angles 25–35°, participate in the imbricate structure. The topmost diamict sheet contains numerous large blocks of the Eemian sand and bodies of glacial ice. The upthrust beds are expressed on the adjacent plateau as an east-facing bow of small arcuate ridges indicating glacial movement from the west. Such movement could result from the eastern flank of an ice lobe which occupied the Yenissei valley.

The glacio-mixtures on the Lower Yenissei, as in other regions of West Siberia, reveal the rheological difference between frozen sand and clay. Frozen sand and even sandy silt (if they are poor in clay particles) show almost exclusively brittle deformation, provided the ice content is low.

Ductile deformations predominate in clay and ice-rich sediments.

For example, sand transformed into large-block breccia (Figure 15a) forms a layer between two diamict beds at 20–31 m from the top of a 54 m high bluff at Nikitinsky Yar (Figure 1). A detailed description of this key section is available in Kaplyanskaya and Tarnogradsky (1975, Figure 3). Downwards in the succession (34–41 m from the top) a layer of allocthonous mylonised silt and clay with marine fauna (Figure 15b) has been encountered. Judging by the meltwater seeping through the talus, the sequence also contains buried ice. The topmost diamict bed exposed in the thermocirques has been observed in a primordially frozen state with spectacular recumbent folds in relict glacial ice (Kaplyanskaya and Tarnogradsky, 1978, Figures 1, 2; Solomatin, 1986, Figure 34, 36). When exposed in the layer of seasonal thawing the till reveals numerous clasts of loose sand.

GLACIAL DISTURBANCES IN THAWED ROCKS

South of 64–65°N, where permafrost is sporadic or non-existent, structures very similar to those

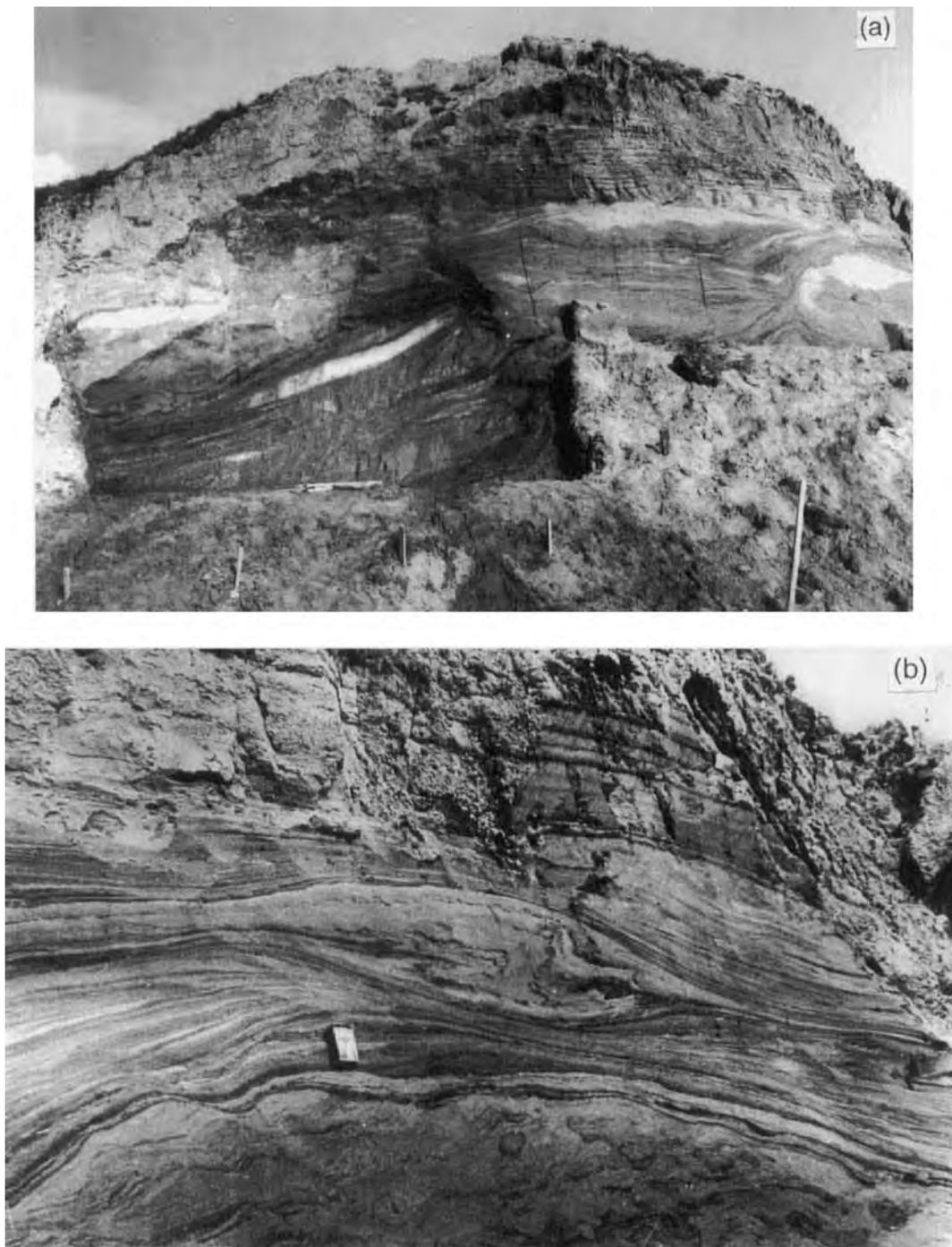


Figure 10 Glaciotectionally displaced and reoriented material of ice/sand diapiir at Harasavey, the western Yamal. (a) general view of exposure, vertical lines are 1 m apart; left – frozen sediments: lower part – primordially frozen glaciotectionite with ice bands; milk-white ice band terminates at the base (partly dotted) of a refrozen talik beneath a former lake, in which sand/silt rhythmite seen at the top was deposited; right – same on the seasonally thawed surface of the cliff; (b) close-up of right (thawed) part of the exposure featuring fluidal structure of the glaciotectionite preserved after two thaw events induced by: i) the former lake (above dotted line in Fig. 10(a)) and ii) seasonal thaw on the surface of exposure; bottom of 10(b) shows the upper part of a large sand ball with palimpsest bedding.

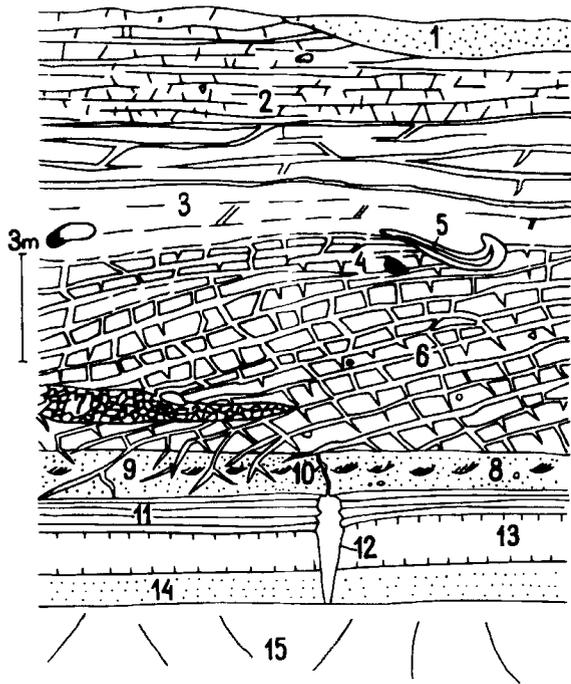


Figure 11 Sketch of cryogenic structures exposed in a thermocirque in the upper part of a Yenissei bluff at Selyakin Mys (S in Figure 1). (1) Iceless sandy drift. (2) Diamicton with vague, tightly spaced reticular structure. (3) Diamicton with bedded structure. (4) Inclusions of dark iceless clay. (5) Inclusions of sand with bedded structure. (6) Diamicton with widely spaced reticular structure. (7) Inclusions of sandy silt with tightly spaced reticular structures. (8) Sand with plant detritus, massive structure. (9) Same with widely spaced reticular structure. (10) Crack above ice wedge. (11) Sandy silt with bedded structure. (12) Ice wedge, syngenetic in the upper part. (13) Massive transparent ice. (14) Iceless sand. (15) Talus. (2)–(7) Epigenetically frozen till.

described above are widely observed in thawed sediments, but it is not always easy to connect them with former subglacial permafrost. The most spectacular evidence of former subglacial permafrost is in the form of large inclusions of non-glacial stratified sand within Pleistocene glacial drift. They range from sand balls several centimetres across up to large stratiform rafts hundreds of metres long. They have been described throughout glaciated West Siberia from within the drift limit (Kaplyanskaya and Tarnogradsky, 1974) up to the northern tip of the Yamal Peninsula (Astakhov, 1981). The sands, if thawed, are so loose that they obviously must have been detached, entrained and deposited in a frozen state. This is especially evident in many cases of long-travelled rafts composed of pre-Quaternary sand with their primary sedimentary

structures preserved. One of the largest is the sand slab from 80 to 120 m long resting upon basal till in the right bank of Irtysh River close to the village of Semeyka near the limit of the maximum glaciation (Kaplyanskaya and Tarnogradsky, 1974). The quartz sand with lignite seams originated from a Palaeocene formation (Nikitin, 1988), which occurs *in situ* at least 650 km NE.

Although ubiquitous, such erratics are evidence only of a frozen sand substrate which does not occur everywhere within the basin; in many places the substrate is composed of various clays and silts, or sand underlain by clayey rocks. It is in those areas of lithologically diverse substrate with underlying clay that the most conspicuous and large-scale glaciotectonic disturbances can be observed.

An instructive case is presented by the well studied Atlym disturbances on the right bank of the Ob river (A in Figure 1). At this site glaciotectonic structures have been directly measured in river bluffs up to 90 m high, providing a detail profile 15 km long. Most of the observed lithological boundaries dip westwards at various angles. The same directions are recorded for numerous overturned folds, thrust planes, imbricated slices and clay injections. The contacts of the local sedimentary formations have been extrapolated below water level using borehole data (Figure 16). The resultant generalized profile shows a very complicated zone of deep crumpling of the Palaeogene strata up to 25 km across (Astakhov, 1990).

This zone is of interest not only because of its size, but also because it occurs at the base of the drift and, being truncated by a basal till with a pronounced unconformity, is not expressed in topography. The basal till, containing slabs of the loose Oligocene sand, extends over undisturbed Palaeogene formations southwards to the limit of Middle Pleistocene glaciation. Predecessors perceived the structure as basically folded, either ascribing it to unspecified glaciotectonism (Li and Kravchenko, 1959), or remaining undecided about its genesis (Nalivkin, 1960).

Our detailed survey has revealed that the most characteristic features are listric faults growing steeper and more tightly spaced upstream, i.e. downglacier. In this direction the simple folds change into tight recumbent ones accompanied by numerous zones of mylonisation and thin injections of the underlying Eocene clay. At the distal end subvertically dipping and heavily foliated Palaeogene rocks are abruptly replaced by



Figure 12 Ice wedge (centre) penetrating into massive ice (bottom) in sediments underlying Late Pleistocene till at Selyakin Mys (see Figure 11). Wall is about 3 m high.

subhorizontal strata lying in the normal stratigraphic order with the Eocene clay positioned 250 m below the surface. The general impression is that the increasing downglacier shear strain was suddenly terminated by an insurmountable obstacle. Conversely, in the upglacier direction the structure gradually becomes simpler, with gentle slightly asymmetric folds prevailing and only occasional clayey injections breaking through the overlying Oligocene sand. It is noteworthy that while upglacier the Eocene clay is abnormally thin (about 100 m), at the distal end its thickness may reach 200–250 m (boreholes 1 to 6). This is an indication of extension upglacier and compression downglacier. In borehole 3 steep angles of dip and slickensides have been observed to depths of 310 m below river level (Li and Kravchenko, 1959). Thus, the entire zone of disturbed sediments is *c.* 400 m thick. The underlying Mesozoic-Cenozoic beds down to the Palaeozoic basement, which is 2.8 km deep, are only slightly undulated.

For our discussion the most important feature of the Atlym disturbances is the difference in the

deformation style between the various Palaeogene formations. This difference, dependent on the mechanical properties of rocks, is amply evident in minor structural complications of the extremely deformed distal part of the zone. The most competent behaviour is observed in the Oligocene sand which shows no crenulation but only numerous sharp-edged blocks, often displaced and stacked into a pile of 'chips' (Figure 17, top). Sometimes blocks are thinly sliced along minor upthrusts (Figure 18). In extreme cases the sand is fracture-cleaved or turned into a typical friction breccia consisting of small angular splinters. Such deformations normally develop only in cemented sandstones, which is not the case here. Therefore, the sand must have been cemented by ice during the deformation. This is confirmed by a 120 m long raft of the same sand within the basal till truncating the distorted Palaeogene sediments. The sand raft shows no strong internal deformation, having probably been incorporated into glacial ice before maximum compression was achieved.

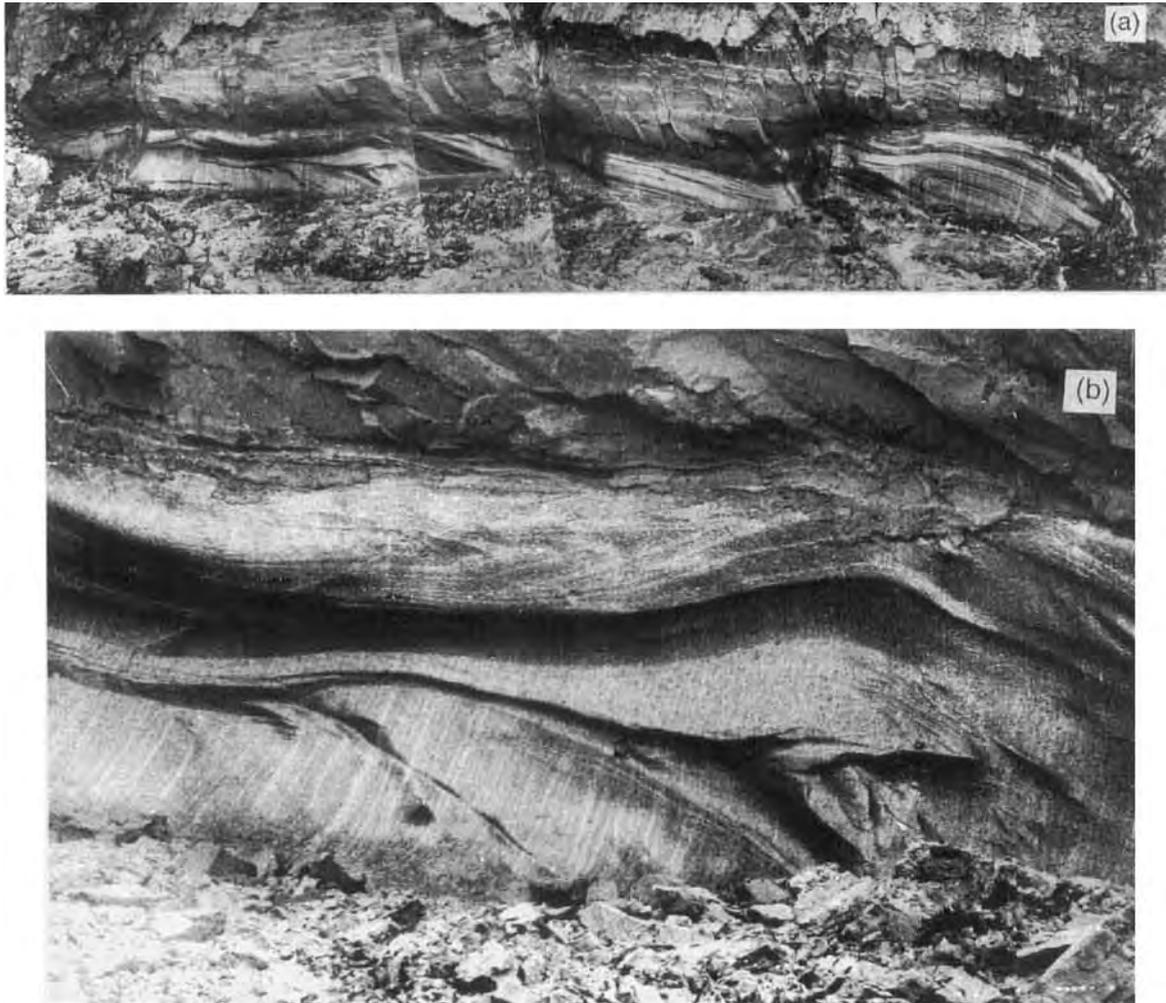


Figure 13 Relict dirty glacier ice with flow structures buried under epigenetically frozen flowtill at Sopkarga, right bank of the Yenisei Estuary (SK in Figure 1). (a) General view of exposure about 20 m long. (b) Close-up of the left end; rare small erratics are visible.

The Upper-Oligocene/Lower-Miocene silty rhythmities feature typical ductile deformations, namely small harmonic and recumbent folds with crenulated bedding (Figure 17, bottom). They also must have been frozen as follows from their normal position on top of the Middle Oligocene sand.

Extreme ductile deformations have been observed in the Eocene montmorillonite marine clay which participates in the above structures only as tabular injections up to 200 m across, or minor clastic dikes 0.3–1 m thick (Figure 17). Similar features could theoretically develop in thawed clays, but in this case it is very unlikely. First, the internal structure of the injected clay is

not just flow foliation but crenulation cleavage. Second, the contacts with the surrounding sand are mostly flat slickensides with practically no sand xenoliths incorporated into the protruding clay. Third, in several places pulled-apart siderite seams within the clay are bent into small tight folds. This demands a relatively competent medium to translate the huge stress. These and similar features, revealing ductile non-fluid rheology, tell us that the clay was frozen when it pierced the overlying sediments.

The ductile deformations of the frozen clay of the Atylm disturbances are normal because clay always retains unfrozen water. According to numerous experiments, this may occur at tem-



Figure 14 Contorted and cleaved debris-rich glacier ice with small erratics within glacio-mixture at Karaul, right bank of the Yenissei (K in Figure 1).

peratures as low as -7°C . At these temperatures, clay shows dynamic viscosity an order lower than pure ice. The rheological behaviour of frozen clay is also more time-dependent (elasto-viscous) than crystalline ice, the shear strength of the former being ten times reduced if a prolonged stress is applied. After the initial resistance is overcome, stressed frozen clay would creep faster than pure ice (Tsyтович, 1973). An important consequence is that maximum deformation with the glacier/frozen-clay complex would occur beneath the ice/rock interface with resultant 'floating' of the glacier upon a clay pillow.

Thus, the mode of deformation of Palaeogene sediments, quite different from their behaviour in their present thawed state, is evidence of former permafrost at least 300 m thick under a Middle Pleistocene glacier at about 300 km from the drift limit. An interpretation in terms of glacial geology is presented in Figure 19. Stage 1 shows a glacier advancing over perennially frozen Palaeogene sediments with shear strength sufficient to support not very thick ice. As the ice

thickness increases (stage 2) the shear resistance limit is first overcome in clay formations closest to the sole of the glacier. This means that the dynamic sole of the glacier (a plane along which the maximum deformation occurs) starts to split with a part of the sliding component being transferred into the frozen clay, which is capable of faster deformation than ice itself. The frozen overlying sand provides a competent dam downglacier, where it is thicker, and gets crumpled and broken into slabs upglacier, where it is thinner.

Stage 3 illustrates the further growth of ice thickness with the dynamic sole being fully established within the frozen clay upglacier. The result is downglacier flow of sheared frozen clay supporting the overburden of the glacier coupled with the frozen sand. Breaking of the competent sand roof reaches a limit downglacier where the sand is thick enough to resist any glacial stress. This means a drastic reduction of the effective velocity of the clay flow and its thickening against the sand obstacle with steep thrusts developing in the ice/sand. This leads to a shift of

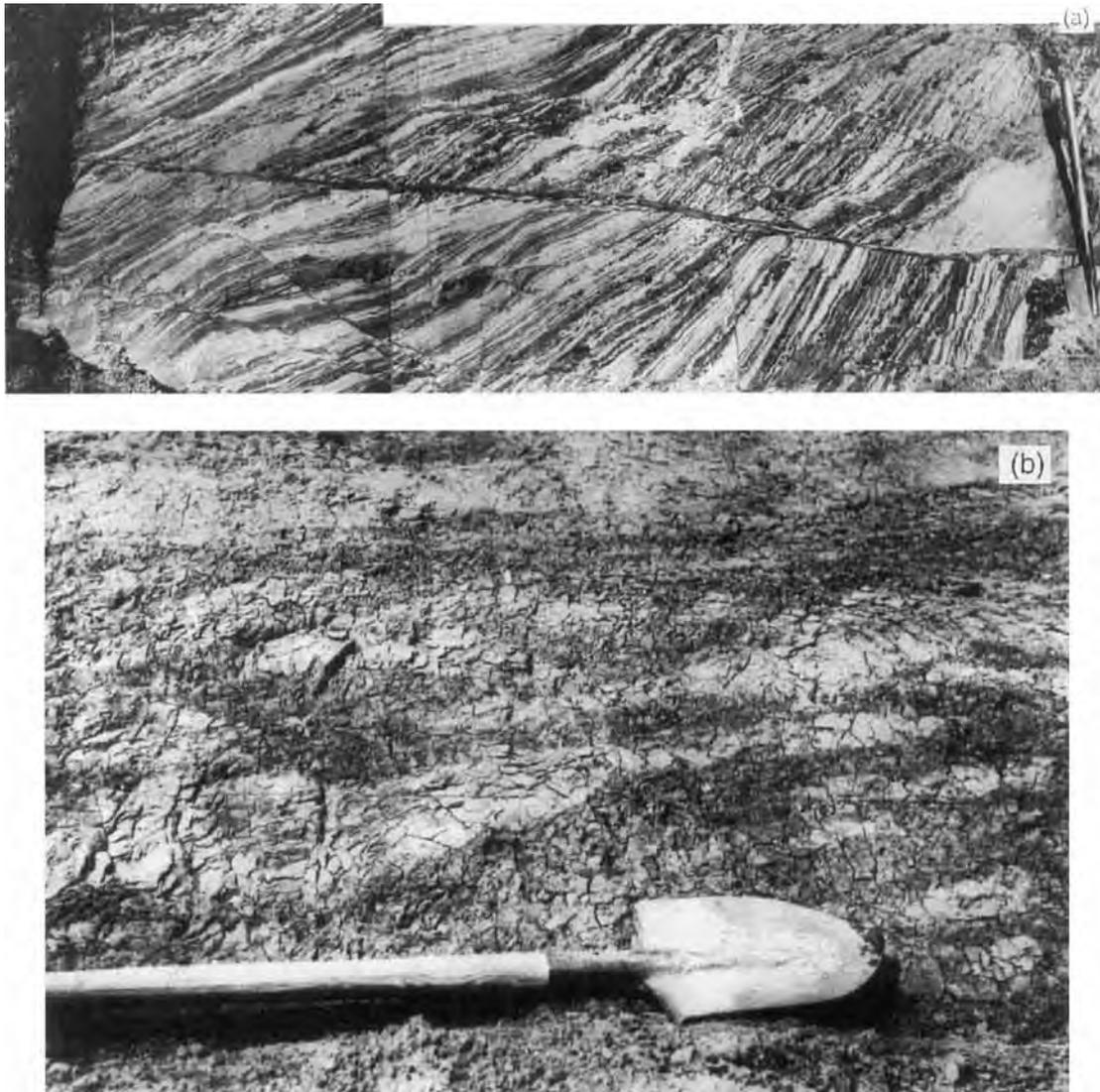


Figure 15 Contrasting deformation behaviour of different rocks participating in glacio-mixture at Nikitinsky Yar, right bank of the Yenissei (N in Figure 1). (a) Large-block breccia consisting of loose sand (light) with silt seams (dark); shovel 1 m long for scale. (b) Mylonised marine sediments: dark is more plastic silty clay; light streaks are clayey silt.

the dynamic sole upwards into downglacier pure ice. Then the glacier advances as a whole block, with two dynamic soles: one within the frozen clay upglacier and one within the ice downglacier. A likely result of the latter is stagnation of the basal or dirty ice downglacier, leading to deposition of frozen tills, as suggested by Lavrushin (1976, 1980).

The model in Figure 19 accounts also for major glaciotectionic features, which in a sedimentary basin do not need a thawed bedrock or an ice margin to develop. In the case of the

Atlym disturbances a marginal environment is unlikely because of the great width and depth of the distorted zone, which would demand large ice thicknesses. The main prerequisite is rheologically contrasting sedimentary formations. Moreover, an originally diverse subglacial topography is not necessary for large-scale deformations of the substrate and entrainment of large erratic blocks. An upward shift of the dynamic sole of the glacier, reacting to the different rheologies of the substrate, is ample reason for detachment and distant transportation of large

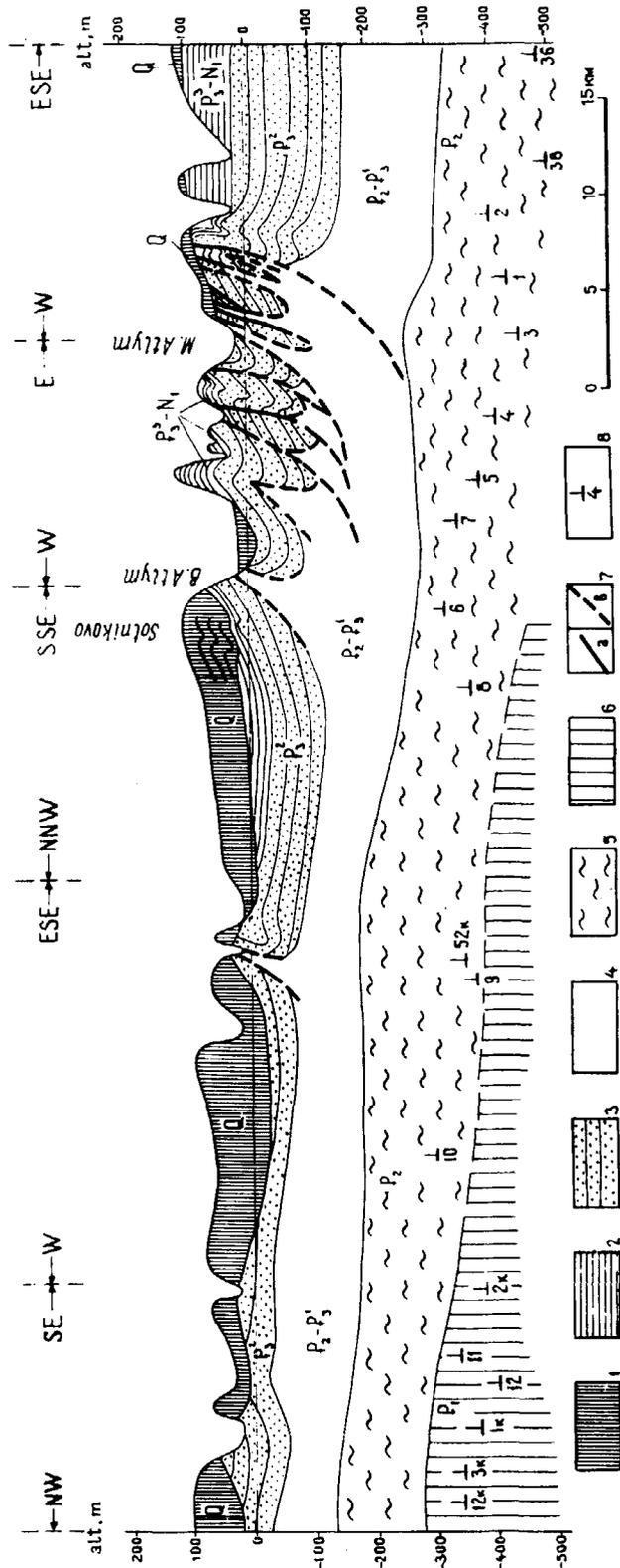


Figure 16 Geological profile along the right bank of the river Ob near Maly Ailym settlement (from Astakhov, 1990). Location is A in Figure 1. (1) Glacial tills and outwash. (2) Terrestrial silty rhythmites. Upper Oligocene to Lower Miocene. (3) Terrestrial sand. Middle Oligocene. (4) Marine clay with siderite seams. Eocene to Lower Oligocene. (5) Eocene diatomites. (6) Palaeocene marine clay. (7) Faults: (a) observed, (b) inferred. (8) Bottom of borehole with its number.

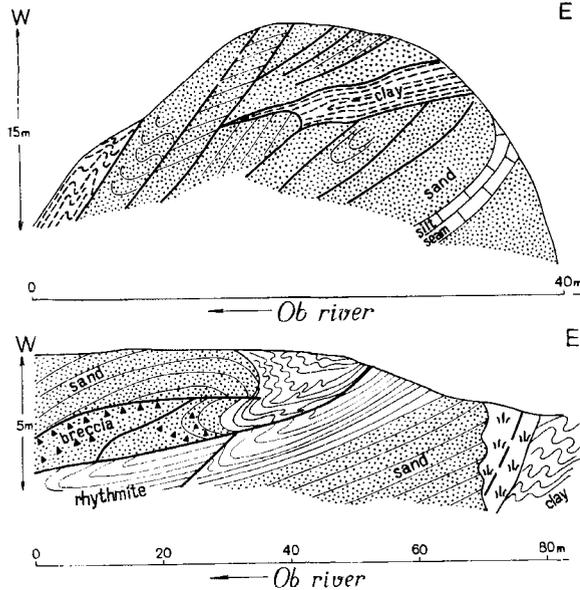


Figure 17 Tectonically deformed Palaeogene sediments observed in river bluffs, 3.5 km (top) and 13 km (bottom) from the eastern end of the Atlym disturbances (see Figure 16). Top section exposes blocks of loose Middle Oligocene sand stacked in a frozen state with small injections of Eocene marine clay. Bottom section displays a ductile thrust in folded and crenulated Upper-Oligocene/Lower-Miocene silty rhythmites; Middle Oligocene sand is not crenulated but brecciated along the thrust plane. The disturbed sediments are positioned 40–50 m below the base of the glacial drift.

E slabs of sediments. For a better understanding of the erosional and tectonic activities of a cold glacier it seems useful to consider a possible split of its dynamic sole which may (sequentially or simultaneously) occur in (i) englacial, (ii) basal, and (iii) subglacial positions. The different positions of the zone of maximum deformation within the ice/permafrost complex accounts for the large-scale shaping of landscape by glaciers, i.e. deepening of upglacier depressions by excavation and building up of downglacier elevations by accretion of till sheets.

DISCUSSION

Palaeoglaciological reconstructions of former ice sheets usually take into account three main forms of ice motion: (i) basal sliding of warm-based glaciers over a rigid bed; (ii) internal deformation of basal ice of cold-based glaciers upon a rigid bed (Drewry, 1986); and, especially lately, (iii) deformation of thawed glacial drift with high pore-water pressure beneath glacier ice (e.g. Alley, 1991). In addition, as the West Siberian data suggest, deformation of frozen sediments under cold-based glaciers was widespread in the Pleistocene. This process has already been described by Echelmeyer and Wang (1987) on a



Figure 18 Middle Oligocene sand with seams of clayey silt exposed, 8 km upglacier from the eastern end of the Atlym disturbances (see Figure 16). The photo shows thin slices of medium-grained quartz sand upthrust along silt seams truncating laminated sand.

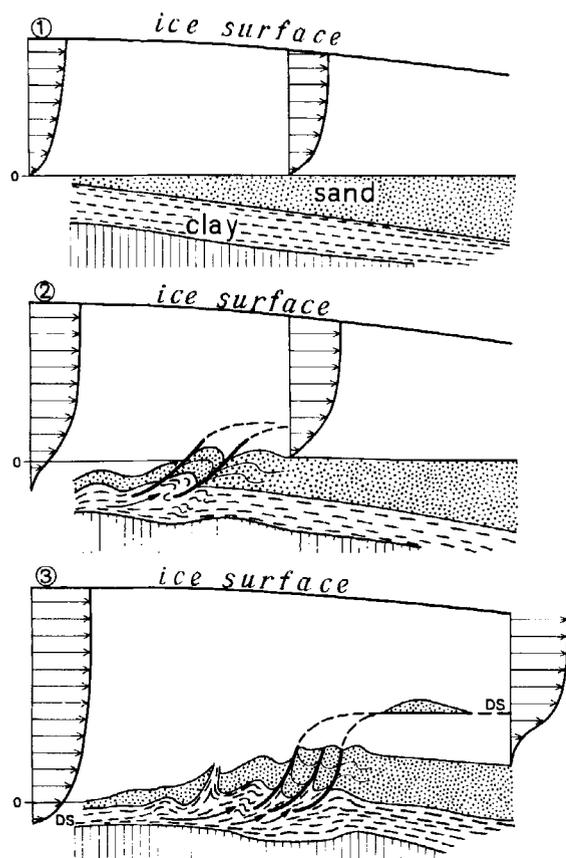


Figure 19 Origin of deformations in frozen glacier bed consisting of clay and sand as inferred from the Atlym disturbances (see Figure 16). Stages 1 to 3 refer to thickening of advancing glacier. Parallel arrows show vertical distribution of summary velocities within ice/permafrost. DS is the dynamic sole of the glacier. See text for explanation.

much smaller scale in a thin layer beneath a cold-based alpine glacier. Our inspection of frozen relicts of the Pleistocene glaciation in Siberia shows ubiquitous traces of this process over vast expanses, sedimentary sheets tens to hundreds of metres thick being involved in subglacial deformation. Such a thick deformable bed means a generally lower shear strength as compared to glacier ice. This could have happened if thick clayey strata and/or other sediments with high ice content were available. Water-saturated clayey sediments can retain their plasticity under a wide range of negative temperatures owing to the preservation of water in a liquid state (Tsytoich, 1973; Williams and Smith, 1989). The high salinity of pore water in West Siberia would also facilitate the plastic-frozen state of clays and retard their freezing at subzero temperatures.

Various types of ground ice, if voluminous enough, would also produce subglacial zones of low shear strength. In such cases even coarse-grained sediments could react to prolonged stress as incompetent rocks. The zones of weakness may consist of stratiform ice as well as ice-supported sands. The latter case (represented by the Harasavey diapirs) resembles dilation decreasing the shear strength of a sand massif. A weak element, namely dead glacial ice, is also present in tills deposited in a frozen state.

Favourable conditions for the downward shift of the zone of maximum deformation are provided by any increase in the temperature gradient in the frozen substrate and overlying ice. This is because it implies colder and therefore more rigid glacial ice. Conversely, if warmer ice flows over colder permafrost, glacial deformation of the sedimentary substrate is insignificant (e.g. the Marresale and Selyakin Mys cases).

Lastly, deformations concentrate at the base of permafrost if it is not too deep for shear strains. Such shallow permafrost might develop as a result of the refreezing of taliks just before an ice advance. Especially favourable conditions for the zone of weakness appear when the permafrost base moves upward under a thickening ice sheet (Romanovsky, 1993), particularly if meltwater discharge is impeded.

Thus, the dynamic sole of a cold-based glacier shifts down into incompetent strata of the substrate to depths allowed by increasing normal stress. Simple shear deformations develop in a weak layer until a plane of decollement emerges in the form of shear zone or drag (ductile) fault. Owing to friction heat and pulsating ice movement a multigelation regime of recurrent freezing and thawing must exist along the fault or shear zone. This regime, even with positive feedback, would hardly produce enough heat for all-round thawing of subglacial permafrost.

More competent, usually sandy, members of the frozen sediment package moving above the subglacial dynamic sole in the substrate were partly transformed into cataclastic breccia. Subsequently, they could be glacially entrained and turned into long-travelling sediment rafts and clasts. Unimpeded subglacial displacement in this basically extension zone supported by glacial shear stress could provide forward motion of the entire glacier/permafrost complex with large glaciotectonic nappes developing. In other cases subglacial extension of incompetent frozen sedi-

ments ceased, changing downglacier into a zone of compression of frozen rocks.

The transition to a compression zone usually happens at the ice margin. This zone may develop either upglacier from the margin or in the proglacial position. In the former case the normal reason for the change of deformation style is the cooling and hardening of subglacial sediments, thus impeding their horizontal flow. The result would be tight folding and imbricate stacking of slabs of frozen sediments (and often ice) separated by thrusts. In the proglacial position, compressional deformation may be topographically expressed as a squeezed-up end moraine with a similar imbricate structure.

The change of subglacial extension to compression deformations often occurred in non-marginal environments, induced by lithological, orographic or thermal barriers. The purely lithological barrier formed by the downglacier thickening of hard frozen Palaeogene sands is exemplified by the Atlym disturbances. A similar effect could be achieved by advection of colder ice producing a rigid dam downglacier.

Large compression disturbances, marginal at Marresale, and non-marginal at Atlym, are ubiquitous in glaciated West Siberia. When studied in boreholes and natural exposures (e.g. Zakharov, 1968; Astakhov, 1979; Sergiyenko and Bidzhiyev, 1983; Kaplyanskaya and Tarnogradsky, 1986; Generalov, 1987), they show deeply crumpled sedimentary successions hundreds of metres thick. In many places they change to horizontally lying strata only below 300–400 m depth. The size and amplitude of these imbricate structures tempt some geologists to connect them with deep-seated tectonism (e.g. Sergiyenko and Bidzhiyev, 1983; Generalov, 1987). Such attempts are refuted not only by the epidermal character of the disturbances, but also by the bow-shaped form of their topographic expression and the fundamental fact that they occur exclusively within the drift limit (Astakhov, 1986).

The argument is usually focused on 'para-autochronous' imbrications of the Atlym type, where the parent undisturbed strata are close by, and where Pleistocene tills are not directly involved in the compression structures. However, the same tectonic style is also observed in Quaternary successions, and in those cases where compressed and folded slices are definitely foreign, i.e. transported from afar along a glacio-thrust. One of the most spectacular examples is known in Arctic European Russia, where an

imbricate assemblage of Mesozoic and glacial sediments has been found stacked against a Palaeozoic ridge, 40–50 km from their *in situ* occurrence (Gornostay, 1990). A bow-shaped zone of erratic Cretaceous sandstones, 0.5–4 km wide and 60 km long, located near the Lyamin River in West Siberia (Generalov, 1987), is probably another.

In West Siberia large imbricate structures are often accompanied by numerous erratic rafts of sedimentary rocks, although many far-travelled rafts are known to lie on top of undisturbed younger sediments. Such are the largest studied sedimentary erratics, the stratiform slabs of Palaeogene (Samarovo) and Jurassic (Yugan) rocks near the limit of the maximum glaciation. The Samarovo erratic is at least 400 m long and 30 m thick, and the Yugan erratic is more than 160 m long and up to 14 m thick (Shatsky, 1965). Transportation of these, and the above-mentioned Semeyka erratic, is conceivable only if a good part of the basal ice was, at least temporarily, relatively rigid and immobile. This allowed the upper ice layers with entrained rafts of frozen sediments to move distances of several hundreds of kilometres.

The subglacial compression in the non-marginal position, expressed as assemblages of tightly spaced, folded slices with protrusions of substrate sediments mixed with glacial ice, is associated with accelerated ice flow and entrained material in the higher layers of the glacier (Weertman, 1976). This may take the form of a large thrust plane inside the glacier, along which blocks of frozen sediments are transported far down-glacier (Lavrushin, 1976) and are transported far downglacier. Such a change of subglacial deformation to englacial sliding over obstacles is described here as the shifting or 'springing up' of the dynamic sole of the glacier (Figure 19). In this context, glaciers on wet beds represent only a special case of glacier sliding, geologically recorded in lodgement tills. Sliding may also occur well below the glacier sole within a frozen sedimentary substrate with resultant ductile thrusts, or englacially, which can be recorded in long-travelled rafts of soft sediments.

Both processes — extensional motion of subglacial rocks over a weak zone and their compression and deformation at an obstacle — lead to unstable, shifting behaviour of the dynamic sole of the glacier. The latter probably provides the most effective mechanism of erosion and distant transportation by cold-based glaciers rather than

the abrasion and regelation adfreezing of debris characteristic of warm-based glaciers.

Subglacial diapirism is widespread in glaciated West Siberia. It includes not only the described mesoscale protrusions but also very large topographically expressed diapiric massifs built of clays and diatomites (Generalov, 1987). The diapirism is generally thought to be induced by differential loading by stagnant glacial ice (Astakhov, 1986). The meso-scale diapirism described in the Yamal Peninsula obviously occurred in perennially frozen rocks owing to the low shear strength of clays (e.g. Marresale) and/or low density of buried ice (e.g. Harasavey).

Vaccillations of the base of thin subglacial permafrost in an area adjacent to a thawed bed could (at a stage of aggradation) intensify the segregation of massive ground ice, which only indirectly, via parental meltwater, is connected with continental glaciation. A similar process has been described for subglacial bodies of intrasedimental ice in the Canadian Arctic by Rampton (1991).

The thermal zonality of Pleistocene ice sheets (Sugden, 1977; Hughes, 1981) is still an open question for West Siberia. A thawed substrate may be suggested for the deeper parts of the Kara Sea shelf, where ice divides were situated. Farther south it must have changed into the freezing zone with intense erosion and glacio-tectonism (Hughes, 1981). Most of the glaciated area was covered by cold-based ice with the exception of large ice streams along the major river valleys. Especially numerous traces of wet-based conditions are observed along the Yenisei. For example, boulder pavements, typical for lodgement by warm-based glaciers (Eyles, 1983), have been described in Middle Pleistocene tills at several localities between 64 and 62°N (e.g. Troitsky, 1975). Drumlinized surfaces are noted in airphotos of the right-bank Yenisei area, where relatively rigid Permian sandstones are close to the surface.

The predominantly frozen glacier bed is closely connected with the manner of deposition of thick successions of Middle Pleistocene tills in the central parts of glaciated West Siberia, which almost invariably contain clasts and rafts of very loose sands. Such tills cannot be deposited by classical lodgement processes in which there is particle-by-particle release of debris from the melting sole of the glacier (Drewry, 1986). The model of frozen basal exfoliation suggested by Shatsky (1966) and elaborated by Lavrushin

(1976, 1980) is more applicable. This model envisages accretion of basal tills by progressive stagnation of thick layers of debris-laden basal ice. A possible cause of this process is the relatively fast influx of cold ice from the upper layers of the ice sheets towards their base (Kaplyanskaya and Tarnogradsky, 1993). Another reason is the stagnation of basal ice over lithological barriers, as suggested in Figure 19.

Exfoliated icy tills incorporated into subglacial permafrost are the last elements of disintegrated Pleistocene ice sheets to thaw because of (1) their basal position and (2) protection by the cold stored in the underlying permafrost. Thus, they lose their ice much later than when the main glacier body disappears. Over great expanses of West Siberia (between 66 and 60°N) these melt-out tills are still underlain by deep-seated relict permafrost blocking terrestrial heat flow. In the area of present permafrost analogous tills have not melted out and are described as 'primordially frozen tills' with relict glacial ice. Therefore, we think that many basal tills are deposited not at the moment of aggregation of particles by melt-out, but when the debris-laden ice becomes stagnant, i.e. at the end of glacial transportation. Melt-out may be considered as a kind of diagenetic process (Astakhov and Isayeva, 1988) dependent on the post-depositional climate history. The slow melt-out of thick sequences of stagnant ice protected from beneath by permafrost has evidently provided the rare hydraulic and gravitational conditions (Paul and Eyles, 1990) necessary for the preservation of glaciodynamic structures.

CONCLUSIONS

The geologic and geothermal data suggest that the Pleistocene ice sheets of West Siberia were mostly cold-based. The subglacial frozen sediments were not a passive, rigid bed but were involved in a glacier strain system. In many cases this resulted in the all-round mobility of large sedimentary masses, which were capable of lateral movement over considerable distances. The dynamic sole of the glacier could dive deeply into a weaker zone of the sedimentary substrate to produce a plane of detachment, thus providing a mighty mechanism of subglacial erosion. The mobilized masses of frozen sediments were stopped by lithological, orographic or thermal barriers and then stacked to form imbricate

structures. Glacier ice participating in such imbricate structures is preserved within the present permafrost zone. These subglacial barriers could make the dynamic sole of the glacier rise up into an englacial position, thereby providing entrainment and distant transportation of giant sedimentary rafts.

In West Siberia glacially entrained debris was mostly deposited by basal exfoliation of stagnating sheets of dirty ice, thus building up the subglacial permafrost. Such icy tills melted out much later, under interglacial conditions, and not everywhere. In the zone of present permafrost the basal tills often retain their primordially frozen state. They are the most conservative element of inland glaciation during interglacial shrinking of the cryosphere. In sedimentary basins frozen basal tills survived longer, being protected from beneath by the cold stored in the sedimentary substrate.

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REFERENCES

- Alley, R. B.** (1991). Deforming-bed origin for southern Laurentide till sheets? *Journal of Glaciology*, **37**(125), 67–76.
- Arkipov, S. A., Isayeva, L. L., Bepaly, V. G. and Glushkova, O.** (1986). Glaciation of Siberia and North-East USSR. *Quaternary Science Reviews*, **5**(1–4), 463–474.
- Astakhov, V. I.** (1976). Geological evidence for Pleistocene ice dispersal centres on the Kara shelf (in Russian). *Doklady, Acad. Sci. USSR*, **231**, 1178–1181.
- Astakhov, V. I.** (1979). New data on the latest activity of Kara shelf glaciers in West Siberia. In *IGSP Project 73/1/24, Quaternary Glaciations in the Northern Hemisphere*, Prague, vol. 5, pp. 22–31.
- Astakhov, V. I.** (1981). Tills in the Northern Yamal (in Russian). *Doklady, Acad. Sci. USSR*, **260**(1), 149–152.
- Astakhov, V. I.** (1986). Glaciotectonics of West Siberia as related to petroleum prospecting problems (in Russian). *Geomorfologia*, **3**, 56–64.
- Astakhov, V. I.** (1990). The tectonic value of aerial and satellite images of petroleum fields in the transverse Ob region (in Russian). In *Aerokosmicheskiye Syomki pri Izuchenii Glubinnogo Stroyeniya Regionov SSR*. VSEGEI, Leningrad, pp. 28–41.
- Astakhov, V. I.** (1992). The last glaciation in West Siberia. In *Sveriges Geologiska Undersokning*, Ser. Ca 81, pp. 21–30.
- Astakhov, V. I. and Isayeva, L. L.** (1988). The 'Ice Hill': an example of 'retarded deglaciation' in Siberia. *Quaternary Science Reviews*, **7**, 29–40.
- Badu, Yu. B., Trofimov, V. T. and Vasilchuk, Yu. K.** (1982). Main regularities of distribution and types of massive ground ice in the north of the West Siberian Plate (in Russian). In *Plastovye Ldy Kriolitozony*. Yakutsk, pp. 13–24.
- Balobayev, V. G., Pavlov, A. V. and Perlstein G. Z.** (1983). *Teplofizicheskiye Issledovaniya Kriolitozony Sibiri (Thermophysical Studies in the Cryolithozone of Siberia)*. Nauka, Novosibirsk (214 pp.).
- Bolikhovskiy, V. F., Velichkevich, F. Yu. and Sulerzhitskiy, L. D.** (1989). Dating Pleistocene and Holocene events in Marresale area of the Yamal Peninsula (in Russian). In *Geokhronologia Chetvertichnogo Perioda*. Abstracts of All-Union Congress. Tallinn, 11.
- Bryzgalova, M. M. and Bidzhiyev, R. A.** (1986). The history of marine basins of the West Siberian North (in Russian). *Izvestia, Acad. Sci. USSR, Geography*, **1**, 81–93.
- Drewry, D.** (1986). *Glacial Geologic Processes*. Arnold, London (276 pp.).
- Echelmeyer, K. and Wang, Z.** (1987). Direct observation of basal sliding and deformation of basal drift at sub-freezing temperatures. *Journal of Glaciology*, **33**(113), 83–98.
- Eyles, N.** (ed.) (1983). *Glacial Geology*. Pergamon (409 pp.).
- French, H. M. and Harry, D. G.** (1990). Observations on buried glacier ice and massive segregated ice, Western Arctic Coast, Canada. *Permafrost and Periglacial Processes*, **1**(1), 31–41.
- Gataullin, V. N.** (1984). The structure of soft sediments in an area on the western coast of the Yamal Peninsula (in Russian). In *Stroyeniye Shelfa Morei SSR kak Osnova Otsenki Inzhenerno-Geologicheskikh Usloviy*. Riga, pp. 26–29.
- Gataullin, V. N.** (1988). *Verkhnechetvertichnyye Otlozheniya Zapadnogo Poberezhya Poluostrova Yamal (Upper Quaternary Deposits of the Western Coast of the Yamal Peninsula)*. PhD thesis, VSEGEI, Leningrad.
- Gataullin, V. N.** (1990). Massive ice on the western coast of the Yamal Peninsula: structure, composition, origin (in Russian). In *Geokriologicheskie*

- Issledovaniya v Arkticheskikh Rayonakh*. International Symposium at Yamburg, Vol. 1, Tyumen, pp. 3–11.
- Gataullin, V. N.** (1991). The Marresale Formation of the western Yamal: deltaic sediments of the Pre-Ob (in Russian). *Bulleten Komissii po Izucheniyu Chetvertichnogo Perioda, Acad. Sci. USSR*, **60**, 53–61.
- Generalov, P. P.** (1987). The secondary tectonism in the surficial sedimentary cover of West Siberia (in Russian). In *Opal'ity Zapadnoi Sibiri*, Tyumen, pp. 48–74.
- Grosswald, M. G., Vtyurin, B. I., Sukhodrovsky, V. L. and Shishorina, Zh. G.**, (1985). Ground ice of West Siberia: the origin and geoecological significance (in Russian). *Materialy Glatsiologicheskikh Issledovaniy*, **54**, 145–152.
- Hart, J. K., Hindmarsh, R. C. A. and Boulton, G. S.** (1990). Different styles of subglacial glaciotectionic deformation in the context of the Anglian ice sheet. *Earth Surface Processes and Landforms*, **15**, 227–241.
- Hart, J. K. and Boulton, G. S.** (1991). The interrelation of glaciotectionic and glaciodepositional processes within the glacial environment. *Quaternary Science Reviews*, **10**, 335–350.
- Hughes, T. J.** (1981). Numerical reconstructions of paleo ice sheets. In *The Last Great Ice Sheets*. Wiley, pp. 221–261.
- Hughes, T. J.** (1985). The Great Cenozoic Ice Sheet. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **50**, 9–43.
- Kaplyanskaya, F. A.** (1982). Stratiform ground ice in the glacial drift of the western coast of the Yamal Peninsula at Harasavey settlement (in Russian). In *Plastovye Ldy Kriolitozony, Yakutsk*, pp. 71–80.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1974). *The Middle and Lower Pleistocene of the Lower Irtysh Area* (in Russian). Trudy VSEGEI, 214, Leningrad (160 pp.).
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1975). The origin of the Sanchugovka strata and problem of interrelation between glaciations and marine transgressions in the north of West Siberia (in Russian). In *Kolebaniya Urovnya Mirovogo Okeana v Pleistotsene*. Geographical Society of the USSR, Leningrad, pp. 53–95.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1976). Relict glacial ice in the north of West Siberia and its role in glaciated regions of the cryolithozone (in Russian). *Doklady, Acad. Sci. USSR*, **231**(5), 1185–1187.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1978). Relict glacier ice and its role in the Quaternary mantle and relief of the permafrost area (in Russian). In *Trudy VSEGEI*, 297, Leningrad, pp. 65–76.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1982). Glacial formations in the vicinity of the Marresale Polar Station, the Yamal Peninsula (in Russian). In *Trudy VSEGEI*, 319, pp. 77–84.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1986a). Preglacial marine sediments in the north of West Siberia: the Leskino formation (in Russian). In *Kainozoi Shelfa i Ostrovov Sovetskoi Arktiki*. Sev-morgeologia, Leningrad, pp. 100–109.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1986b). Remnants of the Pleistocene ice sheets in the permafrost zone as an object for palaeoglaciological research. *Polar Geography and Geology*, **10**(4), 65–72.
- Kaplyanskaya, F. A. and Tarnogradsky, V. D.** (1993). *Glatsialnaya Geologia (Glacial Geology)*. Nedra, St Petersburg (328 pp.).
- Karpov, Ye. G.** (1986). *Podzemnye Ldy Yeniseiskogo Severa (Ground Ice of the Yenisei North)*. Nauka, Novosibirsk (132 pp.).
- Komarov, V. V.** (1987). Glaciomelange on the Lower Yenisei (in Russian). In *Paleogeografia Srednei Sibiri*, Krasnoyarsk, pp. 136–143.
- Krapivner, R. B.** (1986). *Beskornevye Neotectonicheskie Struktury (Rootless Neotectonic Structures)*. Nedra, Moscow (208 pp.).
- Kurchikov, A. R. and Stavitsky, B. P.** (1987). *Geotermiya Neftegazonosnykh Oblastey Zapadnoi Sibiri (The Geothermometry of Petroleum Regions of West Siberia)*. Nedra, Moscow (134 pp.).
- Lavrushin, Yu. A.** (1969). *Chetvertichnye Otlozheniya Spitzbergena (The Quaternary of Spitsbergen)*. Nauka, Moscow (181 pp.).
- Lavrushin, Yu. A.** (1976). *Structure and Development of Ground Moraines of Continental Glaciations* (in Russian). Trudy, Geological Institute of Acad. Sci. USSR, 288 (237 pp.).
- Lavrushin, Yu. A.** (1980). Vital problems of till sedimentogenesis. In *Till and Glacigenic Deposits*, Poznan, pp. 19–40.
- Lawson, D. E.** (1979). *Sedimentological Analysis of the Western Terminus Region of the Matanuska Glacier, Alaska*. CRREL Report, 9 (122 pp.).
- Li, P. F. and Kravchenko, L. M.** (1959). To the question of genesis of the Maly Atlym disturbances (in Russian). In *Trudy SNIIGiMS*, Novosibirsk, pp. 36–38.
- Moran, S. R., Clayton, L., Hooke, R. L. B., Fenton, M. M. and Andriashek, L. D.** (1980). Glacier-bed landforms of the Prairie region of North America. *Journal of Glaciology*, **25**(93), 457–476.
- Nalivkin, V. D.** (1960). The Maly Atlym disturbances on the river Ob (in Russian). In *Trudy VNIGRI*, Leningrad, 158, pp. 256–266.
- Nikitin, V. P.** (1988). On the age of the glacial raft near Semeyka settlement on the Irtysh (in Russian). In *Geologia i Poleznye Iskopayemye Yuga Zapadnoi Sibiri*, Novosibirsk, pp. 166–168.
- Paul, M. A. and Eyles, N.** (1990). Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers. *Quaternary Science Reviews*, **9**, 51–69.
- Rampton, V. N.** (1991). Observations on buried glacier ice and massive segregated ice, western Arc-

- tic Coast, Canada: discussion. *Permafrost and Periglacial Processes*, **2**, 163–165.
- Romanovsky, N. N.** (1993). *Osnovy Kriogeneza Litosfery (Fundamentals of Lithosphere Cryogenesis)*. Moscow University (335 pp).
- Rostovtsev, N. N.** (ed.) (1982). *Geological Map of the West Siberian Plain and Adjacent Regions (Without the Quaternary), Scale 1:1 500 000* (in Russian). VNIGRI, Leningrad.
- Sergiyenkov, V. M. and Bidzhiyev, R. A.** (1983). The Late Quaternary tectonism in the north of the West Siberian Lowland (in Russian). *Bulletin Moskovskogo Obshchestva Ispytateley Prirody, Geology*, **58**(6), 73–82.
- Sharp, M.** (1985). Crevasse-fill ridges – a landform type characteristic of surging glaciers. *Geografiska Annaler*, **67A** (3–4), 213–220.
- Shatsky, S. B.** (1966). The glacially detached blocks in the Quaternary at Yurty Yeutskie on river B.Yugan and close to the city of Khanty-Mansiysk (in Russian). In *Osnovnye Problemy Izucheniya Chetvertichnogo Perioda*, Nauka, Moscow, pp. 206–217.
- Solomatin, V. I.** (1986). *Petrogenез Podzemnykh Ldov (Petrogenesis of Ground Ice)*. Nauka, Novosibirsk (216 pp.).
- Sugden, D. E.** (1977). Reconstruction of morphology, dynamics and thermal characteristics of Laurentide ice sheet at its maximum. *Arctic and Alpine Research*, **87**(A), 89–100.
- Sugden, D. E.** (1978). Glacial erosion by the Laurentide ice sheet. *Journal of Glaciology*, **20** (83), 367–392.
- Tarnogradsky, V. D.** (1982). On the origin of the tabular massive ground ice on the Kara coast of the Yamal Peninsula (in Russian). In *Plastovye Ldy Kriolitozony*, Yakutsk, pp. 80–89.
- Troitsky, S. L.** (1975). *Sovremennyyi Antiglatsializm (Modern Antiglacialism, a critical essay)*. Nauka, Moscow (163 pp.).
- Tsytovich, N. A.** (1973). *Mekhanika Myorzlykh Gruntov (Mechanics of Frozen Soils)*. Vysshaya shkola, Moscow (446 pp.).
- Wateren, D. F. M. van der** (1985). A model of glacial tectonics applied to the ice-pushed ridges in the Central Netherlands. *Bulletin of Geologic Society of Denmark*, **34**, 55–74.
- Weertman, J.** (1976). Age of ice-sheet drill cores. *Quaternary Research* **6**(2), 203–207.
- Williams, P. J. and Smith, M. W.** (1989). *The Frozen Earth*. Cambridge University Press (305 pp.).
- Yershov, E. D.** (ed.) (1989). *Geokriologia SSSR. Zapadnaya Sibir (The Geocryology of the USSR. West Siberia)*. Nedra, Moscow (454 pp.).
- Zakharov, Yu.F.** (1968). Exotectonic disturbances in the sedimentary cover of West Siberia (in Russian). *Geologia i Geofizika*, **6**, 148–155.
- Zemtsov, A. A.** (1976). *Geomorfologia Zapadno-Sibirskoi Ravniny (Geomorphology of the West Siberian Plain)*. Tomsk (344 pp.).