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Massive Ground Ice Body of Glacial Origin at Yugorski Peninsula, Arctic Russia

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ABSTRACT

A massive ground ice body at Cape Shpindler on Yugorski Peninsula, southern Kara Sea coast, Russia, was studied with regard to large-scale internal structures, its stratigraphic context and contacts to surrounding sediments, in order to highlight its origin. The massive ground ice contains deformation structures and deformed sediment rafts that show a consistent direction of deforming force. It is bounded upwards with a sharp and unconformable thaw contact to overlying till. The stratigraphical and structural evidence suggests that the massive ground ice body is relict glacier ice. Examination of data from a separate study on ice crystallography and isotopic composition of the massive ice body does not contradict this conclusion. The isotope composition and profiles conform with what can be expected for deformed basal ice. The chronology for the Shpindler Cape sequence implies that the glacier ice might be older than 250 ka years. Consequently, permafrost has preserved the relict glacier ice for the duration of at least two interglacials (Eemian and Holocene), as well as several Saalian and Weichselian interstadials, illustrating the preservation potential of the permafrost. Copyright © 2003 John Wiley & Sons, Ltd.

KEY WORDS: massive ground ice; permafrost; glacier ice; arctic Russia

INTRODUCTION

Thick bodies of massive ground ice have been described from northern Alaska (e.g. Lawson, 1983), the western Canadian Arctic (e.g. Mackay, 1971; French and Harry, 1988, 1990; Mackay and Dallimore, 1992), China (e.g. Wang, 1990) and western Siberia (e.g. Vtyurin, 1975; Kaplyanskaya and Tarnogradsky, 1986; Astakhov and Isayeva, 1988). A favoured genesis is that most bodies of massive ground ice formed through ice segregation and injection processes, where excess pore water froze within the sediments (Mackay, 1971; Mackay and Dallimore, 1992). Russia has had a long discussion on the origin of massive ground ice occurring in western Siberia. The prevailing opinion of the permafrost community in Russia has been that ground ice is primarily the results of ice segregation and ice injection processes (e.g. Lazukov, 1972; Vtyurin, 1975; Melnikov *et al.*, 1990). Many Russian Quaternary geologists have favoured the alternative explanation, originally suggested by Toll (1897), that the massive ground ice in northwestern Siberia is buried glacier ice and remnants of Pleistocene ice sheets (e.g. Kaplyanskaya and Tarnogradsky, 1986; Astakhov and Isayeva, 1988). French and Harry (1990) suggest that both massive segregation-injection ice and buried glacier ice exist in the western Canadian Arctic, but that field differentiation between massive ground ice of the two origins is complicated. Other theories explaining the origin of massive ground ice include buried lake, river or sea ice, and buried snow bank ice (see French, 1996, 98–101). It has been

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pointed out that massive ground ice in the Arctic primarily occurs within the limits of formerly glaciated areas (Harry *et al.*, 1988; French and Harry, 1990; Astakhov and Isayeva, 1988). One explanation for this relationship is that most of these ice bodies are relict glacier ice. However, the most commonly accepted theory is that these ice bodies, more than 10 m thick, are segregation ice, and that glacial meltwater supplied the huge amounts of water required (Rampton, 1988; French and Harry, 1990).

The different ice types relevant to this study are glacier ice (both firnified ice and basal ice (Mackay, 1989; Knight, 1997) and intrasedimental segregation ice (Mackay and Dallimore, 1992). Although both segregation and glacier ice in the modern environment can be distinguished on the basis of diagnostic criteria such as crystallography and geochemistry, and nature of the contacts between the ice body and surrounding sediments (Mackay, 1989; French, 1996), bodies of buried massive ground ice often have undergone post-burial chemical and structural alterations (French and Harry, 1990). This makes any field determination between massive segregated ice and glacier ice very complicated. It is therefore important to consider as many criteria as possible before a genetic interpretation of a massive ice body can be made (Mackay, 1989; French and Harry, 1990).

During Quaternary geological fieldwork on the coastal cliffs of Cape Shpindler (Figure 1), Yugorski Peninsula, on the southern coast of the Kara Sea in arctic Russia, a conspicuous body of massive ground ice was studied. Yugorski Peninsula is situated well within the limits of Late-Quaternary continental glaciations (Svendsen *et al.*, 1999). According to recent reconstructions it was last covered by an ice sheet during an Early-Middle Weichselian glaciation, 90–60 ka years BP, but remained ice-free during the Late-Weichselian glacial maximum about 20 ka years BP (Lokrantz *et al.*, 2003). Goldfarb *et al.* (1985) and Goldfarb and Ezhova (1990) were first to describe the Cape Shpindler massive ground ice. They recognized deformation structures in the ice, and concluded that the massive ground ice could be interpreted as being either of glacial origin or buried firn accumulations. Leibman *et al.* (2000, 2001) conclude that the massive ice has characteristics of both intrasedimental and surface (glacier) ice. They interpret crystallographic and geochemical data in favour of an intrasedimental origin.

The Cape Shpindler ground-ice body is directly overlain by a till deposited by ice advancing on land from an ice dispersion centre in the Kara Sea. Manley *et al.* (2001) and Lokrantz *et al.* (2003) interpret the massive ground ice as relict glacier ice, based on its structural characteristics and stratigraphic position. The discrepancy between these studies and those of Leibman *et al.* (2000, 2001) led us to study the massive ground ice in order to determine if it was relict glacier ice, subsequently buried by sediment, or if it was formed *in situ* by ice-segregation processes. Our approach is to study the stratigraphical context of the ice body, its contacts to surrounding sediments, and internal structures in the ice, to infer depositional history. Also, we briefly evaluate descriptions of the crystallography, chemistry and isotopic composition of the ice by Leibman *et al.* (2000, 2001) and discuss their preferred interpretations in light of our stratigraphical data.

THE STUDY AREA—LOCATION AND STRATIGRAPHY

Cape Shpindler borders the Kara Sea in Russia, and has a flat to gently undulating topography, incised

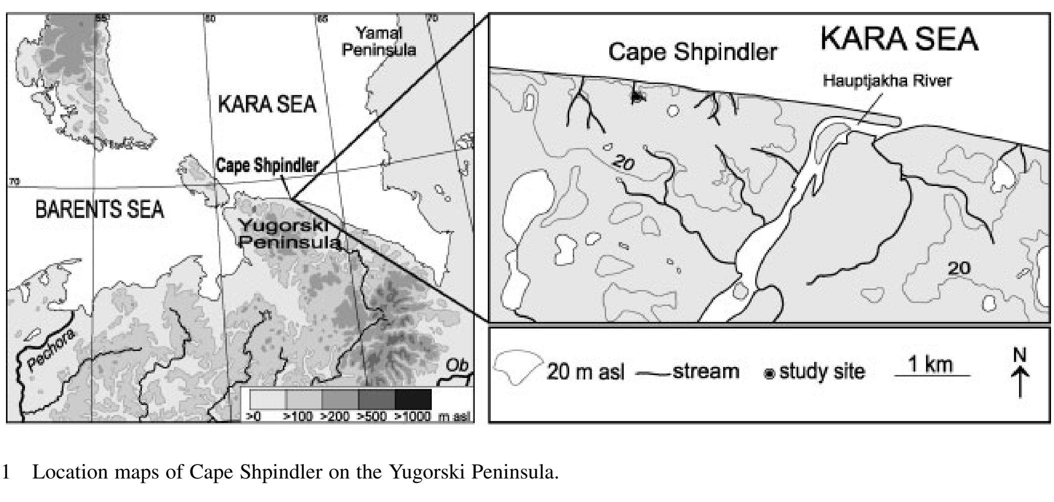

Figure 1 Location maps of Cape Shpindler on the Yugorski Peninsula.

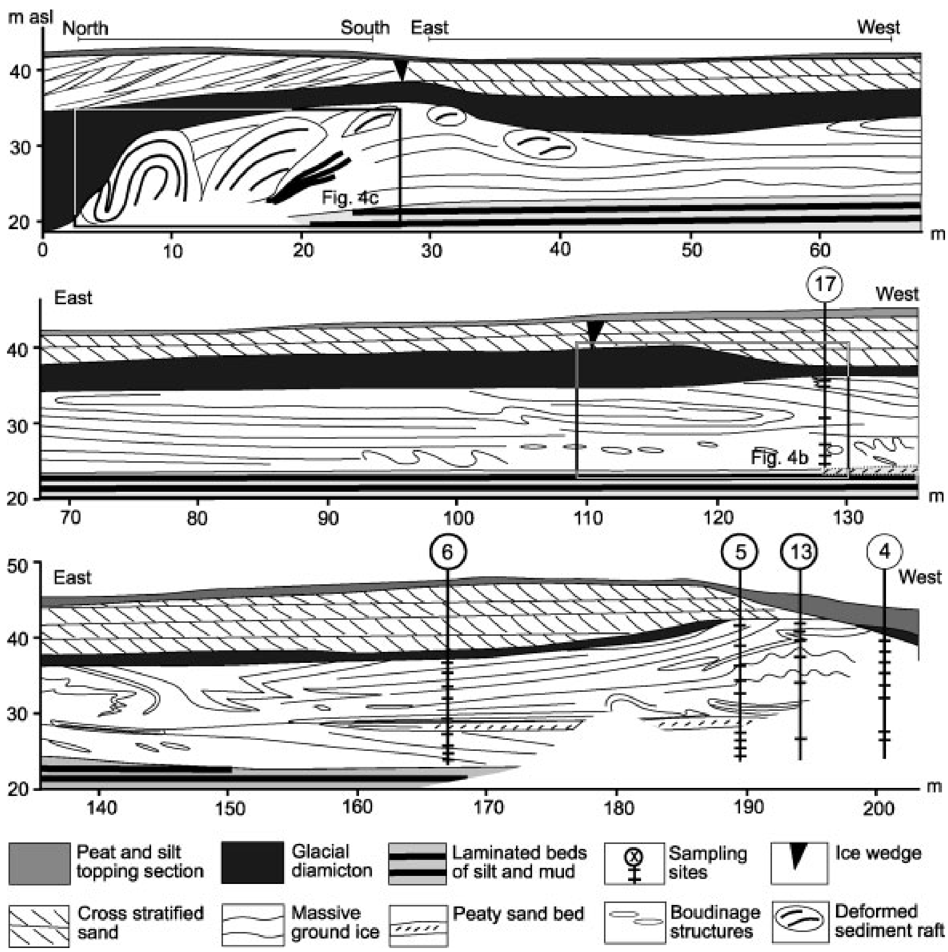

Figure 2 The studied section with massive ground ice and its stratigraphic context. Approximate sites for ice sampling are indicated with numbers and horizontal markings.

with meandering rivers (Figure 1). Coastal erosion and retrogressive thaw slumps have resulted in numerous sediment exposures along the cliff profile. The body of massive ground ice studied is located in the face of a thaw slump about 2 km west of Huptjakha river mouth (Figure 1). The exposure created by the slump is approximately 200 m wide (Figure 2) and more than 40 m high.

Two major post-Eemian glaciations have been inferred for the southern Barents Sea-Kara Sea region (Svendsen *et al.*, 1999; Houmark-Nielsen *et al.*, 2001; Mangerud *et al.*, 2001), with ice moving from a divide in the Kara Sea and flowing landwards from the shallow shelf areas. The local composite event stratigraphy recorded in the Cape Shpindler bluffs (Lokrantz *et al.*, 2003) recognizes six major stratigraphical units, termed A-G, which reflect the Late-Quaternary environmental development (Figure 3). At the base, marine-to-fluvial regressive sediments (units A, B and C) were deposited under interglacial conditions during a period of marine regression. Unit A is silty-clayey marine sediment, containing numerous marine fossils. The rich mollusc fauna lacks warm boreal indicators and together with foraminiferal taxa suggest that unit A is a shallow marine deposit, accumulated in a sub-arctic environment with open marine conditions, potentially similar to present conditions. Unit B is a shallow-marine pro-delta sediment of cross-stratified gravelly sand and beds of laminated silt and sand. Fossil shells testify to its marine origin. Unit C overlies it; a facies association of silts and sands interpreted to represent deposition on a tidal flat on a coastal delta plain (C1) and shallow fluvial channel and overbank environment on a delta

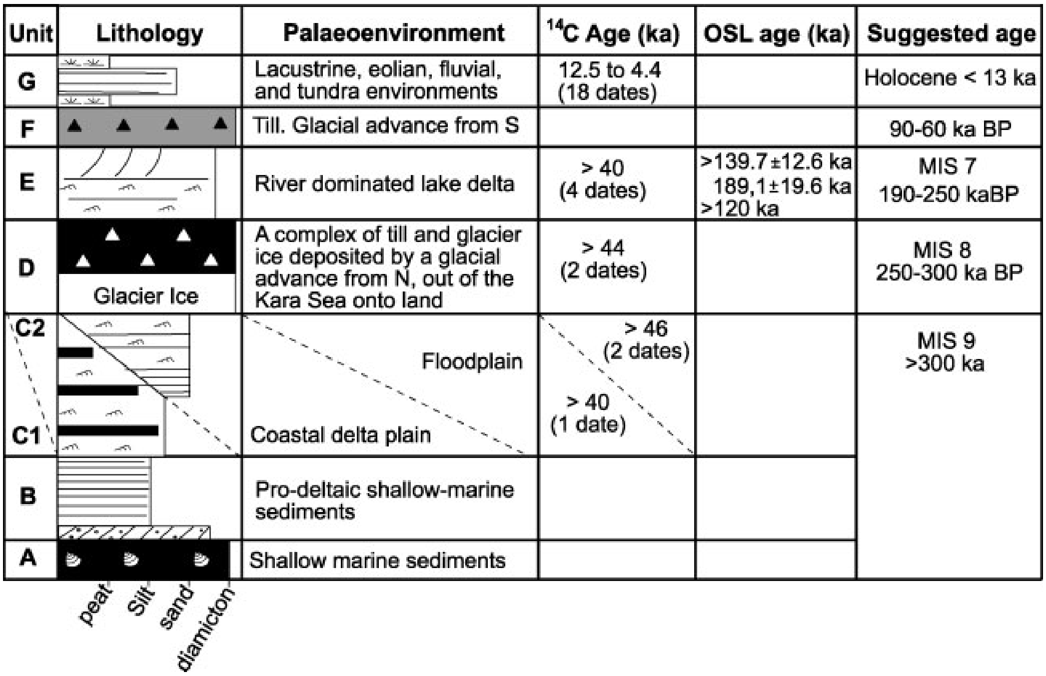

Figure 3 Composite stratigraphy of the Cape Shpindler coastal cliffs. Modified after Lokrantz *et al.* (2003).

floodplain (C2). Moss and vascular detritus from both sub-units yielded AMS radiocarbon ages of >40 ka years. BP. The succession from marine to fluvio-deltaic environments in units A through C reflects a base level fall either reflecting increasing sediment supply and/or a fall in relative sea level, potentially resulting from regional glacio-isostatic unloading. Environmental conditions for this sedimentologic succession with its documented marine fauna probably occurred repeatedly during interstadials or early within interglacials during the past 700 ka years.

On top of the interglacial sediments, a till (unit D) was deposited by a glacier advancing from the north, out of the Kara Sea towards the Pai-Hoi uplands of Yugorski Peninsula. Apart from till, this glacial event is recorded in the cliffs by large-scale glaciotectonic disturbances. The glaciotectonic deformations are described in detail in Lokrantz *et al.* (2003). The massive ground ice constitutes, stratigraphically, the lowermost part of unit D. After deposition of the unit D till there followed a period of fluvial deposition in a river-dominated lake delta (unit E).

The age control for units A-E, based on AMS ^{14}C analyses on samples from units C and E, give infinite ^{14}C ages. Moss and vascular detritus from unit C and three pieces of wood and a mammoth bone from unit E yielded ages of >40 ka years (Manley *et al.*, 2001; Lokrantz *et al.*, 2003). Optically-stimulated luminescence dating was attempted on five sediment samples from the fluvial sands and silts of unit E, in order to constrain the age of the unit D till and the massive ground ice. Two quartz separates showed insufficient growth in luminescence (saturation) with additive beta dose and thus did not yield ages. Two other samples yielded ages of ca. >120 ka years and >140 ka years, respectively and these estimates are considered to be minimum estimates. One sample gave the finite age estimate of 189.1 ± 19.6 ka years. It is prudent to consider luminescence ages >100 ka years to be possible minimum estimates (e.g. Forman *et al.*, 2000) and thus we conclude that unit E probably predates the Eemian and may be associated with non-glacial conditions during MIS (Marine Oxygen Isotope Stage) 7 or possibly earlier. MIS 7 corresponds to Saale/Drenthe interstadial conditions in the Quaternary climato-stratigraphy of Northern Europe, 250–190 ka years BP (Lowe and Walker, 1997, 11). As a consequence, unit D might have been deposited by a regional glaciation during MIS 8, corresponding to the Drenthe glaciation in Northern Europe, 300–250 ka years BP (Lowe and Walker, 1997). The massive ground ice, if it is relict glacier ice, may have survived several interstadial events, as well as at two interglacial periods (the Eemian (ca. 130–115 ka years BP) and Holocene (10–0 ka years BP)).

Some time before 12.5 ka years BP, possibly during the Early-Middle Weichselian (90–60 ka years BP), the area was once again subject to glacial overriding (unit F). During this younger glacial event, ice moved from an ice divide inland, tentatively from the Pai-Hoi upland. It caused deposition of till, as well as conspicuous glaciotectonic disturbances. This ice advance may have overridden western Yamal Peninsula, where Forman *et al.* (1999, 2002) described a complex of till and relict glacier ice, the Kara diamicton, of Middle or Early Weichselian age (60–90 ka years BP). The uppermost stratigraphic unit in the Cape Shpindler cliffs constitutes Holocene peat, aeolian, fluvial and lacustrine sediments (unit G), dated by AMS to have been deposited between 12.5 and 4.4 ka years BP (Manley *et al.*, 2001; Lokrantz *et al.*, 2003).

METHODS

The fieldwork focused on outlining the geometry of the massive ice body, by tracing unit boundaries along the section, as well as documenting the structural characteristics of the ice by describing and measuring tectonic structures (folded, sheared and thrusted structures) displaced in the ice and surrounding sediments. Complementary descriptions and interpretations of strata not accessible for field mapping due to steepness of the exposure or risk for slumps were made from a video and photographic documentation. Leibman *et al.* (2000, 2001) collected samples of ice from different profiles along the outcrop (Figure 2) so that vertical variations in chemistry, isotopic composition and crystallography could be evaluated. At each sample site, the thawing outer layer of ice was removed before sampling to avoid contamination by percolating water. Vertical thin slices were photographed through a polarized filter directly in the field (Leibman *et al.*, 2000). In the laboratory, images of the thin ice slices were digitized and characteristic crystallographic parameters (average maximal diagonal axis and crystal elongation) were calculated (Leibman *et al.*, 2000). Furthermore, the photo images were used to analyse bubble content and distribution and the concentration and mode of dispersion of debris within the ice. Ice samples for chemical and isotopic composition analyses were collected in 0.5 l plastic bottles (Leibman *et al.*, 2001). Ion concentration of different elements were tested at the Shirshov Institute of Oceanology at the Russian Academy of Sciences in Moscow while isotopic analyses were conducted at the Alfred-Wegner-Institute for Polar and Marine Research at Potsdam (Meyer *et al.*, 2000).

RESULTS

Stratigraphy

Description

The stratigraphy of the exposure is shown in Figure 2. The massive ground ice overlies alternating planar parallel beds of silt to fine sand and mud belonging to the interglacial sediments of unit C (Figure 3). The lower contact of the ice body is sharp and unconformable, but ice veins emanating from the lower part of the ice were observed to penetrate down into unit C. Due to slumping and poor exposure the ice veins could not be studied in detail. The ice body attains a thickness of >10 m and consists of alternating layers of bubbly, transparent and laminated muddy (silty-clayey) ice. The massive ice body can be divided into two parts, separated by a deformed bed of sand with peat laminae (Figure 2). Although somewhat folded, this sand layer is continuous in the western part of the exposure (Figure 2) while it gets increasingly broken up into boudins eastwards (Figures 4a, 4b). Boudinage structures develop when competent sediment layers are stretched under conditions of high shear strain and horizontal extension (Price and Cosgrove, 1991; van der Wateren, 1995), and they are considered to be kinematic indicator structures that record the sense of flow in deforming rocks. Above the sand layer, deformation structures are conspicuous in the ice body and appear as one large isoclinal, recumbent fold (Figures 2, 4b), with small-scale folding and various sigma and boudinage structures (Figures 4a, 4b). Both sigma- and boudinage structures are kinematic indicators that develop in ductile shear zones (Twiss and Moores, 1992). Furthermore, the ice contains large (3–5 m diameter) intraclasts of folded sand strata (Figures 4c, 5a, 5b). They have primary sedimentary structures preserved, and have sharp contacts to the surrounding ice. All deformation structures measured in the sand and massive ice, including folded and sheared structures, show convergence of folds towards south and a deforming palaeo-movement from NNW. By contrast, the lower part of the ice below the sheared and deformed sandy bed constitutes a relatively undeformed parallel laminated ice facies. The laminae dip gently towards the west (Figure 2). This facies is thicker in the west and wedges out towards the east. At 130 m along the section, the ground ice unconformably overlies a peaty sand layer, similar to that occurring within the ground ice (Figure 2).

Striated stones, which can suggest a glacial origin of certain massive ground ice bodies (e.g. French and Harry, 1988), were not observed in the exposure examined, but were present in an outcrop of massive

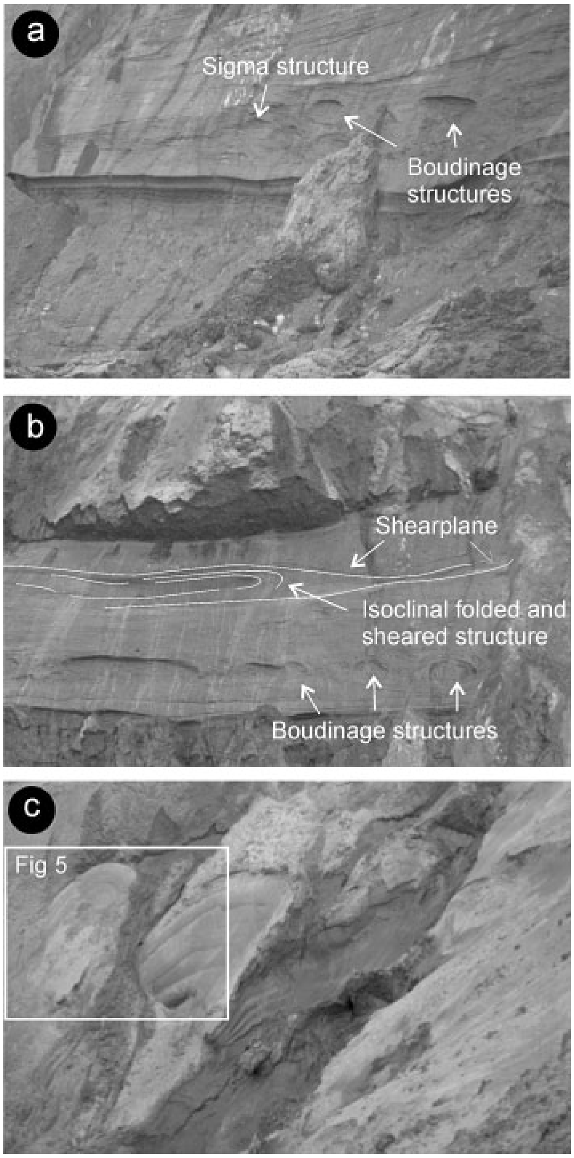

Figure 4 Structures observed in and around the massive ground ice. a) Laminated and bubbly ice facies, with sigma and boudinage structures; b) an isoclinal recumbent fold developed between shear planes in the ice. Notice the sharp thaw-contact between the ice and overlying diamicton; c) imbricated rafts of folded sand within the ice that are overlain by till. For a closer picture of the contact between the massive ice and the till see Figure 5.

ground ice about 8 km further to the west along the coastal cliffs. Although ground ice in that outcrop resembles the deformed ice facies described above, the stratigraphical control does not allow for correlating the two ice bodies.

The massive ground ice is overlain by unit D; a dark, compact, massive, matrix-supported and silty-clayey diamicton. This unit attains a thickness of up to 8 m in the thermocirque. It contains whole and fragmented shells. Striated and bullet-shaped clasts were also observed in the diamicton in sections close to the thermocirque. In the eastern wall of the thermocirque section, three imbricated, folded intraclasts of sand occur within the massive ice. One sand-intraclast extends into the diamicton (Figures 4c, 5a, 5b; Lokrantz *et al.*, 2003). The deformed intraclasts define imbricated folds on the scale of 3–5-m, verging southwards, with sharp contacts to the surrounding ice and sediments (Figures 5a 5b). Here, the contact between the ice and the diamicton is sharp and conformable (Figure 5b), while elsewhere it is an unconformable thaw contact (Figures 2, 4b). The diamicton is overlain by a 10 m thick cross-stratified medium sand and gravel, with secondary occurrence of ripple cross-laminated and planar parallel-laminated sand (unit E), with an erosive lower contact. Orientations of cross bedding indicate palaeocurrent directions from the southeast. Ice wedges were observed within unit E, extending from the top of the unit to a couple of metres depth (Figure 2).

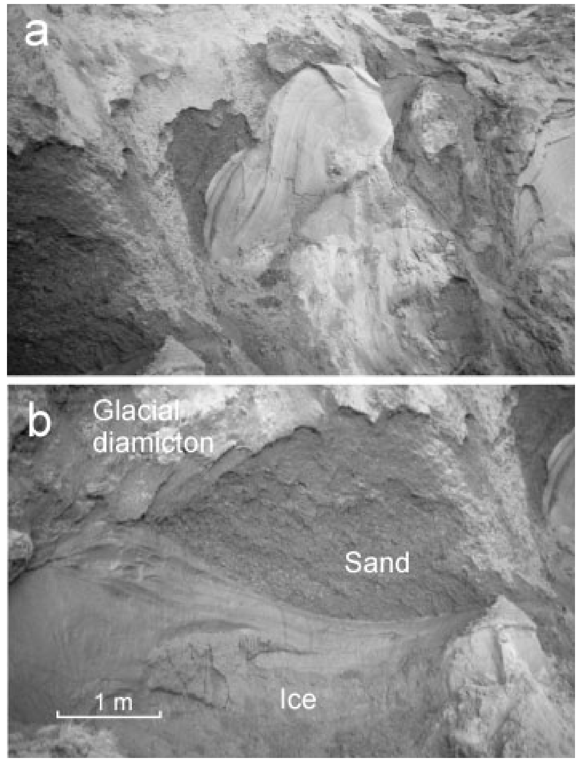

Figure 5 a) Imbricated and dislocated rafts of folded sand at the contact between the ice and overlying diamicton; b) the contact between the massive ground ice, containing imbricated sand rafts, and the overlying till.

Interpretation

Bullet-shaped bedrock clasts and deformed intraclasts of sand in the unit D diamicton indicate that it is a glacial diamicton. Kinematic indicators suggest deposition by a glacier moving out of the Kara Sea onto land, which explains the shells and shell fragments occurring in the diamicton. The intraclasts are interpreted to be subglacially-derived rafts (e.g. Astakhov *et al.*, 1996; Hambrey *et al.*, 1999) from underlying unit C. The preservation of primary structures suggests that the sand was frozen at the time of dislocation.

It has been suggested that an unconformable and sharp contact between massive ground ice and its overlying/underlying sediments argues in favour of a glacial origin for a ground ice body (French and Harry, 1990; Mackay and Dallimore, 1992). This is because massive segregated ice bodies usually have gradational and conformable contacts to surrounding sediments (e.g. see Mackay, 1989; French and Harry, 1990; Mackay and Dallimore, 1992). The ice-diamicton contact at Cape Shpindler is predominantly a sharp, unconformable thaw-contact. Deformation structures, similar to those recognized in the massive ground ice at Cape Shpindler, frequently occur in glacier ice, particularly close to the base of a glacier (Sharp *et al.*, 1994; Knight, 1997) and have been described from relict glacier ice in the western Canadian Arctic (French and Harry, 1990). The orientation of deformation structures in both ice and diamicton consistently suggest a deforming push from the NW, and kinematic indicator structures suggest a strong zonal shear. The observation that folded intraclasts and structures in the ice continue into the overlying diamicton (Figures 4c, 5), i.e. cross a lithostratigraphic boundary, suggests a glacial origin for the ice. An alternative explanation for the deformation structures is that they occurred in pre-existing segregation ice as a result of later glacial thrusting and overriding (e.g. Mackay, 1971; Mackay and Dallimore, 1992; French, 1998). We find this interpretation less plausible for the Shpindler Cape massive ground ice, because it cannot explain the nature of contacts between the ice and overlying diamicton, stratigraphical relationships between the ice and the raft inclusions, or the rafts crossing a stratigraphical boundary between the ice and the overlying diamicton. Hence, post-depositional deformation in the form of either, or both, glaciotectonism and postburial loading and unloading is rejected. A glacial origin of the ice is favoured.

Segregation or intrusion ice is, by definition, younger than the surrounding sediments, whereas the deformation structures that bridge the ice and diamicton in the Cape Shpindler sequence suggest contemporaneous deposition and deformation of the ground ice and diamicton. Mackay and Dallimore (1992) described massive ground ice from the Tuktoyaktuk area, western Canadian Arctic, as an example of massive ground ice forming as a result of freezing of water that flowed under pressure through permeable and unfrozen sand. The stratigraphical setting for that massive ground ice body differs in critical aspects from the Cape Shpindler setting: there, the contact between ice and overlying diamicton is conformable, whereas it is an unconformable thaw contact at Cape Shpindler. The model of Mackay and Dallimore (1992) requires unfrozen, coarse-grained material below the aggrading permafrost, whereas unit C at Cape Shpindler is a stratified fine-grained facies of silts and sands, not very conductive to lateral and upward flowing of groundwater. Also, the inclusions of rafts of unit C in the massive ground ice, with primary structures preserved, suggest the sand was frozen when it was incorporated in the massive ice.

It has long been recognized that ice-sediment coupling occurs below glaciers that advance over soft sediments, and that the styles of glaciotectonic deformations can give information on ice dynamics and basal conditions (Hart and Boulton, 1991). We suggest that the glacier advancing from the Kara Sea across the Cape Shpindler strata entrained frozen subglacial material, sand, marine clay and diamicton, that was transported along thrust planes to a higher position in the ice (e.g. Moran *et al.*, 1980; Hambrey *et al.*, 1999). This diamicton is associated with melt-out processes associated with partly-inherited properties of a primarily basally-transported diamicton (Dreimanis, 1989). Although shearing and entrainment of basal debris can take place under cold-based glaciers (Echelmeyer and Zhongxiang, 1987; Tison *et al.*, 1993; Cuffey *et al.*, 2000), the combination of striated, bullet-shaped stones in the diamicton and subglacial rafts in the ice can indicate a partly warm-based ice sheet (e.g. Weertman, 1961; Hooke, 1970; Hart and Boulton, 1991; Hambrey *et al.*, 1999).

Our reconstruction of the glacial overriding suggests that the ice sheet was probably thin near its terminus and periodically frozen to its bed. When the ice front advancing from the Kara Sea basin onto land reached higher ground, the obstacle formed by both the ice frozen to its bed and the uphill movement likely resulted in compression and high angle thrusting. Overturned folds testify to ductile deformation before shear resistance was overcome and the subglacial sediment was entrained into the ice. The thrust planes probably developed along weak planes in the frozen bed. Boudinage and sigma structures testify to

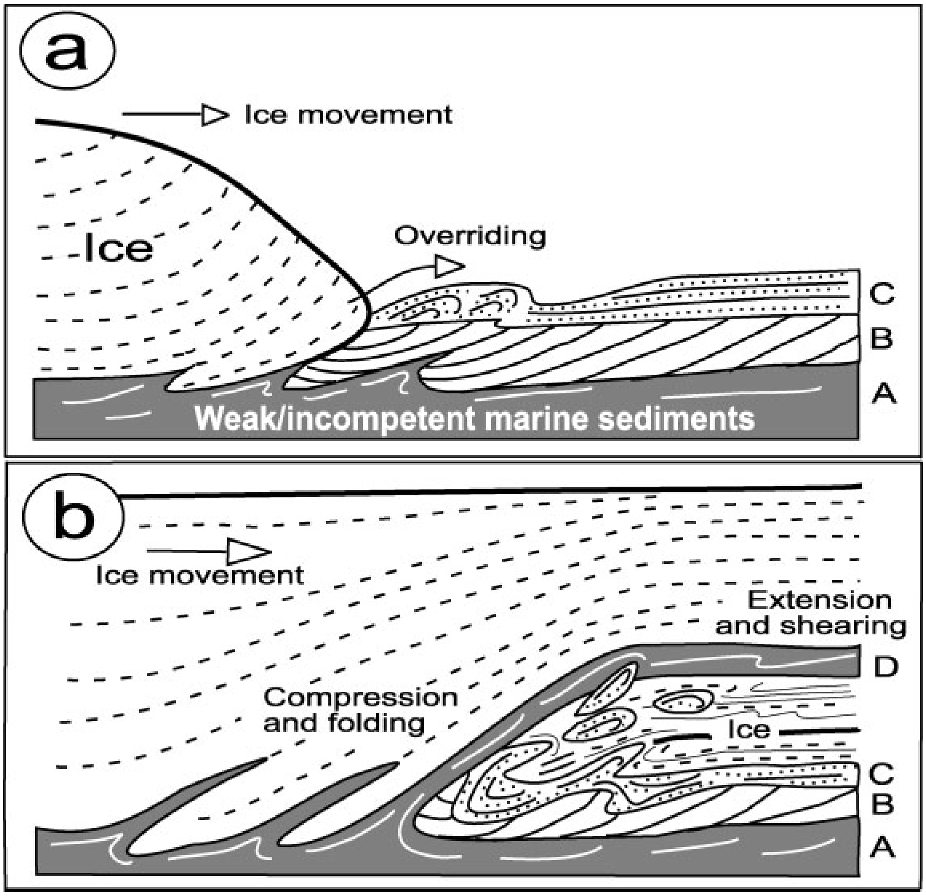

Figure 6 A conceptual model for ice overriding and glaciotectonic deformation of the Cape Shpindler strata: a) ice advances over the sediments at Cape Shpindler, causing both erosion and frontal deformation; sediments transported upwards in the ice via dirt bands; b) a large raft of basal ice, contorted together with surrounding sediments, is detached from the overriding ice.

later subglacial extensional strains. This pattern of compression and then extension of glaciotectonic deformation as an ice sheet advances across sediments has been described by Hart and Boulton (1991). The complex of glacier ice and the unit D diamicton can be compared to the thrusted and contorted glacio-mixtures described from outcrops along the Yenisei River by Astakhov and Isayeva (1988) and Astakhov *et al.* (1996). There, subglacial processes reflect a dynamic sole where sliding of a cold-based glacier may have provoked subglacial extensional shearing in frozen fine-grained sediments and the development of englacial thrust planes within the ice during compression and deformation at an obstacle. According to our reconstruction, the Cape Shpindler massive ground ice is a strained mixture of deformed glacier ice with entrained ice-marginal and subglacial debris. In this environment, elements of entrained snow or proglacial icings and refrozen meltwater might be included in the ice. Figure 6 outlines a simple conceptual model for the stratigraphical associations and glaciotectonic deformations observed at Cape Shpindler.

Ice Petrography and Stable Isotopes

Description of Ice Petrography

The petrographic descriptions are based on polarized photographs of thin slices from the various ice types and data published by Leibman *et al.* (2000). Three different types of ice were recognized: (1) transparent ice, (2) bubbly ice, and (3) laminated ice.

The transparent, vitreous ice crystals average between 3–4 cm in diameter (Figure 7a). They are close to isometric, but also show jagged and interfingering grain boundaries. They have no preferred orientation.

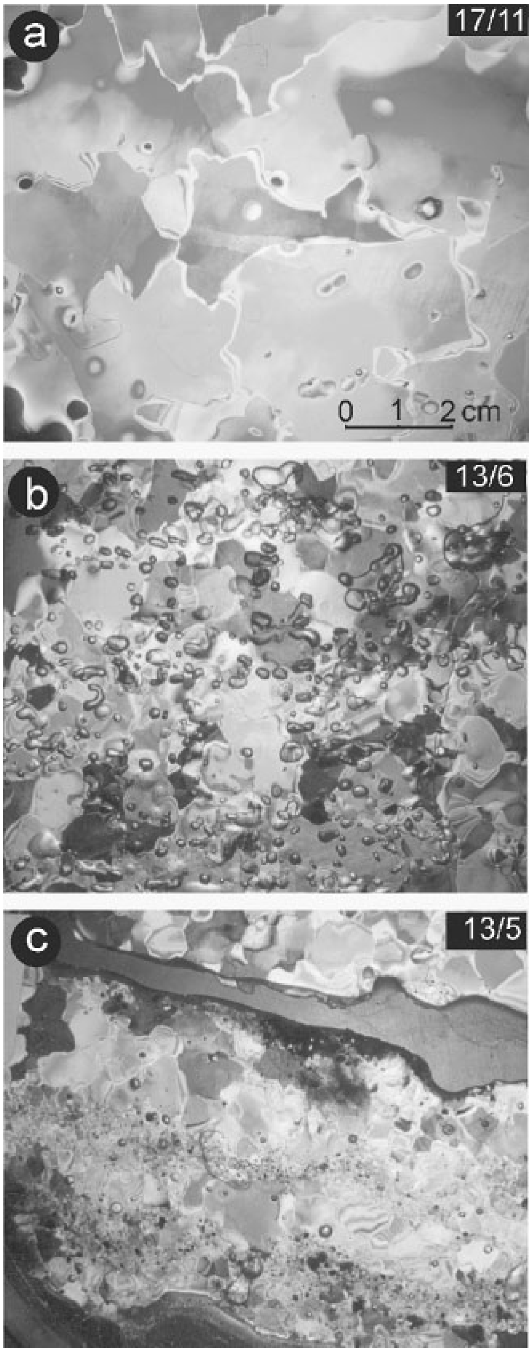

Figure 7 Examples of ice textures of different facies, photographed by Alexander Vasiliev through a polarized filter in the field. For the locations of the sampling sites see Figure 2. a) vitreous ice (17/11), with jagged and interlocked crystal boundaries; b) bubbly ice (13/6), where the zonation of bubble-rich and bubble-poor ice is oriented horizontally; c) muddy laminated ice (13/5), where the mud aggregates are concentrated between crystals. Note that the crystal size is smaller in the muddy-laminae matrix compared to the debris-free ice.

In the bubbly ice, diffuse zones with higher and lower bubble concentrations give the ice an internal banding (Figure 7b). The thickness of bubble-rich and bubble-poor layers are approximately 2–4 cm and 1–3 cm, respectively, and the banding is usually parallel to the bubbly layer. Bubbly ice has an average crystal diameter of 1.5—2.0 cm. The crystals are slightly elongated and often with uneven and jagged grain boundaries. Their long axes are oriented approximately 35° to the horizontal. Gas inclusions are mainly intra-crystalline and confined in the peripheral zones of ice crystals, and more rarely inter-crystalline. They are round to oval and range between 0.2–5 mm in diameter. Oval bubbles are often aligned with their long axes parallel to the horizontal.

In laminated muddy ice, the laminae are expressed as zones with suspended mud aggregates, 0.5–5 mm in diameter, and rounded or amorphous mud fragments, 0.5–3 cm in diameter (Figure 7c). Although occasional grains were observed within crystals, most sediment is inter-crystalline. Within the muddy zones, particle aggregates are sometimes aligned at crystal boundaries as thin strings (Figures 7c, 8), and in one sample these strings were organized into a conjugated set, resembling listric shear planes (Figure 8). The crystal diameters vary with laminae. In muddy laminae they are 0.1–0.5 cm, while they are almost an order of magnitude larger in-between. The crystals are slightly elongated and inclined 30–35° to the horizontal.

Interpretation of the Ice Petrography

The crystal sizes of the massive ground ice at Cape Shpindler are not diagnostic for the genesis of the ice, as they all fall within ranges described both for glacier ice and segregation ice (e.g. Eckerbom and Palosuo,

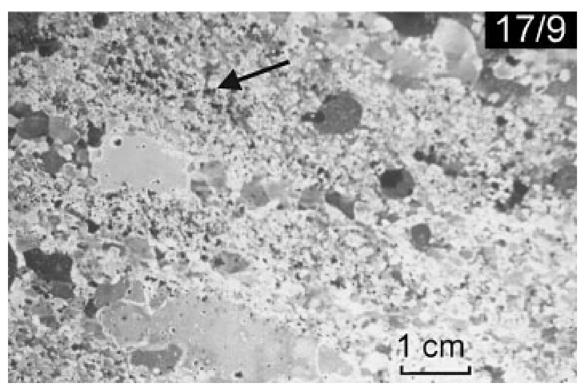

Figure 8 Ice texture of muddy-laminated ice (17/9), sampled close to the lower contact of the massive ground ice body. The arrow points at listric shear planes emanating from what appears to be the same horizon. The shear planes are lined with mud aggregates.

1963; Yoshikawa, 1993; Tison *et al.*, 1993; Tison and Hubbard, 2000). The jagged and interlocked contacts between crystals in vitreous ice and bubbly ice are taken to reflect a degree of recrystallization. In glacier ice, vitreous ice is generally considered as formerly bubbly ice that has lost its bubbles due to gas expulsion during ice deformation and recrystallization or due to meltwater (Souchez and Lorrain, 1991; Hooker *et al.*, 1999). Vitreous ice can also form through the freezing of lake or river water (e.g. Fitzsimons, 1996) or refreezing of meltwater (Souchez and Lorrain, 1991).

Bubbles frequently aligned into foliations of bubble-poor and bubble-rich ice, giving a striped ice appearance, usually characterize meteoric glacier ice (Knight, 1997, 1999). Mackay (1990) described alternating bubble-poor and bubble-rich banding, interpreted as seasonal growth bands, from pingo ice in the western Canadian Arctic. There, the bubble trains tapered upward toward the ground surface. In a ground ice body formed through segregation, i.e. with unidirectional freezing, a subvertical elongation and alignment of bubble trains is to be expected (Mackey, 1989; Hubbard, 1991). This alignment and orientation can be lost over long periods due to recrystallization (Williams and Smith, 1989). Since bubble banding occurs both in glacier and intrasedimental ice, it cannot be used for determining the genesis of the Cape Shpindler massive ice.

The disposition of sediment predominantly between the crystals in the laminated muddy ice facies suggests that the particles were incorporated from below. The sediment could have been incorporated through either segregational processes (including regelation) or shearing at the base of a glacier. The observation of small listric shear planes at the bottom part of the massive ice suggest that some amount of shearing has taken place. Whether this reflects either shearing and debris entrainment at the base of a glacier (Tison *et al.*, 1993), a shearing of material that was incorporated to the glacier by freeze-on mechanisms, or shearing of segregation ice by an overriding glacier cannot be decided from our observations. Petrographic characteristics, such as 3–5 cm-sized pieces of mud within the muddy-laminated ice, could be attributed to segregation ice. However, the mud pieces were not observed to be matched, and they could as well have been entrained through regelation at the base of a glacier.

In summary, the petrographic characteristics of the Cape Shpindler massive ground ice cannot be used diagnostically for deciding if the ice is of either glacial or intrasedimental origin, although both bubble trains and signals of shearing in connection with incorporation of mud in the ice could be attributed to glacial processes.

Description of Stable Isotopes

The chemistry and isotopic analyses of water samples from the Cape Shpindler massive ground ice were described by Leibman *et al.* (2001). Results of isotopic analyses are presented in Figure 9. The data of Leibman *et al.* (2001) show no progressive change in water quality with depth in the massive ice. Instead isotope ratios and ionic composition (Na, Mg, Ca, Cl and SO_4) seem to vary with the different facies of vitreous, bubbly and muddy-laminated ice. Bubbly ice has comparatively lighter δ^{18}O values than vitreous and laminated-muddy ice (Figure 9; sampling profiles 6, 13 and 4). Figure 10 shows the results of stable isotope measurements in a δD-δ^{18}O diagram. All the samples align along the linear regression $\delta D = 7.54\delta^{18}O - 1.46$ with the coefficient of determination $r^2 = 0.97$.

Interpretation of the Stable Isotopes

In a body of segregation ice, vertical profiles of water chemistry parameters through the ice body into the sediment beneath tend to show a continuous trend and a common origin of water (Ad Hoc Study Group, 1984, in Fujino *et al.*, 1988; French and Harry, 1990; Mackay and Dallimore, 1992). Furthermore, no major deviations ought to be expected in the isotope profiles (Vaikmae *et al.*, 1988). Both laboratory and field experiments show that segregational ice formation is accompanied by isotopic fractionation, resulting in about 3‰ heavier ice than the original water (Jouzel and Souchez, 1982; Lorrain and Demeur, 1985). Also, isotopic profiles are expected to show gradual changes with small (<3‰) amplitude range (Vaikmae *et al.*, 1988; Mackay and Dallimore, 1992). The stable isotope properties of the Cape Shpindler ground ice body do not confirm with these signatures of segregation ice. Instead, the ion concentration and isotope values vary with different ice types. Bubbly ice has δ^{18}O values between −18 to −25‰ (average of nine samples is −21.6‰), compared to vitreous and laminated ice, which are isotopically heavier with values generally between −17 to −19‰ (average of 28 samples is −18.4‰).

The relatively large (up to 7–8‰) fluctuations in δ^{18}O values between samples from different ice types could be the combined result of a number of variables. First, there is a significant seasonal variation in oxygen isotope ratios over an ice sheet, so that the range of δ^{18}O between summer and winter precipitation can be in the order of 5–15‰ (Epstein, 1956; Daansgaard *et al.*, 1973). Precipitation on present-day polar ice caps in the Barents Sea area is associated with the passage of cyclones, and rapid changes in δ^{18}O have been observed in winter when the cold arctic air is

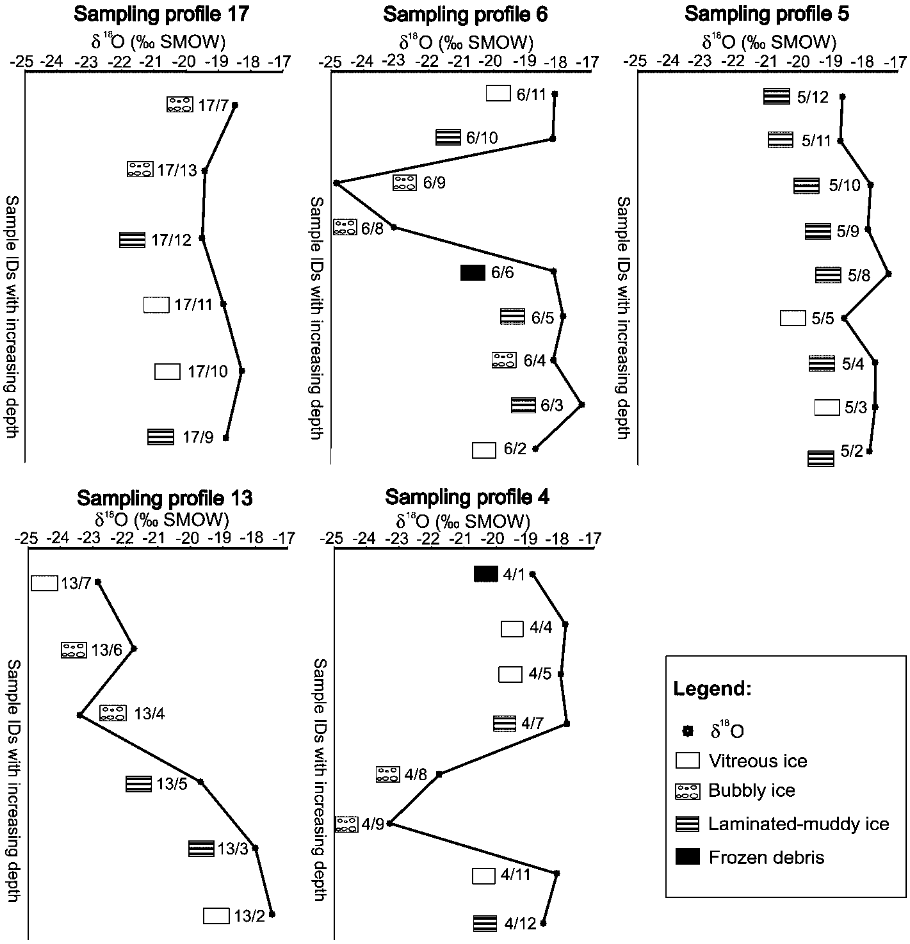

Figure 9 Vertical profiles of the ^{18}O isotope values in the massive ground ice body. The samples are arranged relative to each other, with increasing depth. For sampling sites see Figure 2.

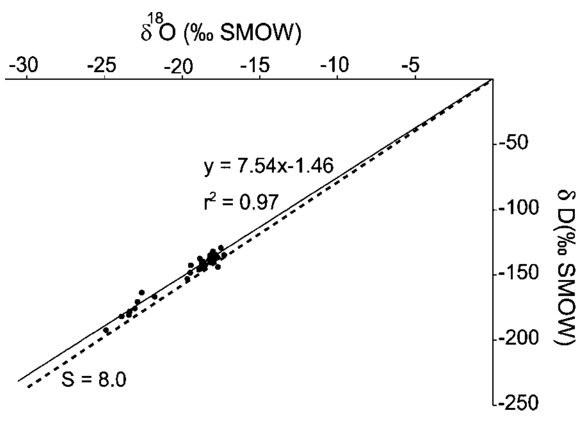

Figure 10 The isotopic composition in $\delta^{18}O$ and δD of the samples from the Cape Shpindler massive ground ice. The solid line defines the slope (7.5) of the linear regression of the ice samples. The broken line shows for comparison the meteoric water-line defined by Craig (1961) where the slope S = 8 is considered as a worldwide mean value.

broken by a cyclone bringing moisture with a high $\delta^{18}O$ from mid-latitudes (Simões, 1990). Although the amplitude of the $\delta^{18}O$ signal decreases with depth in ice due to diffusion effects (Johnsen *et al.*, 1999), some of the $\delta^{18}O$ difference observed in the Cape Shpindler massive ice might be an inherited seasonal signal. Second, silty basal ice from the GRIP ice core on Greenland is significantly isotopically heavier (7–12‰) than the above-lying glacier ice. This has been interpreted to signify that the silty basal ice could be a remnant of a growing stage of the ice sheet (Souchez *et al.*, 1994), reflecting isotopically heavier precipitation during the build-up phase of the ice sheet. Third, fractionation due to freezing leads to an enrichment in heavy oxygen of up to 3‰ in $\delta^{18}O$ (Herron and Langway, 1979). Fourth, folding and faulting in the basal parts of a glacier strongly affect isotopic profiles and can produce complex isotopic patterns (Boulton and Spring, 1986). Fifth, Knight

(1989) found that debris-rich basal ice from west Greenland was isotopically heavier and with greater range of $\delta^{18}O$ than clean ice above the basal zone. He interpreted this to signify isotope fractionation during entrainment of debris through freeze-on.

The significant deviations in $\delta^{18}O$ values in the $\delta^{18}O$ profiles from the Cape Shpindler massive ground ice confirm with what could be expected for deformed glacier ice. Hooke and Clausen (1982) found 5‰ range in $\delta^{18}O$ for glacier ice from the Barnes Ice Cap, Arctic Canada, and Lorrain and Demeur (1985) found 6‰ range in relict glacier ice on Victoria Island, Arctic Canada. We interpret the 7–8‰ deviations in $\delta^{18}O$ values from the Cape Shpindler massive ground ice to favour a glacier origin for the ice. The observed heavier isotope signature for the vitreous and laminated ice compared to the bubbly ice could reflect isotope fractionation due to refreezing and entrainment of debris.

A number of studies have attempted to correlate the co-isotope slope of $\delta^{18}O$ and δD for different kinds of ice in open and closed freezing systems (discussion in: Jouzel and Souchez, 1982; Knight, 1989; Mackay and Dallimore, 1992; Souchez *et al.*, 1990, 1994). Co-isotope analysis has been used to discriminate between different basal ice types in a variety of environments including the European Alps (Jouzel and Souchez, 1982; Hubbard and Sharp, 1995), Arctic Canada (Lorrain *et al.*, 1981), Greenland (Sugden *et al.*, 1987; Knight, 1997), Alaska (Sharp *et al.*, 1994) and Antarctica (Tison *et al.*, 1993). In general, co-isotope slopes greater than about 7.2 have been considered as being associated with atmospheric sources, such as for glacier ice, whereas co-isotope slopes below 6 are to be expected for groundwater-derived intrasedimental ice formed during the latter stages of closed-system freezing (Mackay and Dallimore, 1992). The slope of 7.5 found for the Cape Shpindler massive ground ice is slightly lower than the usual value of 8.0 for the meteoric water line (Craig, 1961), but close to what has been described from other studies of glacier ice. Lorrain and Demeur (1985) found that relict glacier ice on Victoria Island, Arctic Canada, had a co-isotope slope of 7.5. Jouzel and Souchez (1982) found that ice in the Aktineq Glacier, Canadian Arctic, had a co-isotope slope of 7.8, which they considered representative for glacier ice that had not been subject to major isotopic changes during or since its formation. They also found that ice formed by refreezing in the basal layers defined a different slope of 4.9, caused by isotopic fractionation. Similarly, Knight (1989) found that, whereas clean ice in west Greenland had a co-isotope slope of 8.1, samples from debris-rich basal ice lay along a co-isotope slope of 6.5. He suggested that the debris-rich ice had undergone isotopic fractionation consistent with freezing-on at the glacier bed. Mackay (1990) described a co-isotope slope of 7.2 for intrusive ice in a pingo whose water source was groundwater below a drained lake, but usually intrasedimental ice has slopes below that.

We conclude that the co-isotope slope of 7.5 found for the Cape Shpindler massive ground ice suggests primarily glacial origin. Entrainment of debris in the ice is compatible with some re-freezing taking place at the base of the glacier, which could explain the deviation from the ideal co-isotopic slope of 8.0 (Craig, 1961). Our interpretation also allows for a partially more complex origin for the ice. Proglacial segregation ice and sediment may have been accreted to basal ice as the glacier progressively overrode them (e.g. Souchez *et al.*, 1994; Knight, 1999), the advancing ice may have incorporated firn or icings, and ice segregation mechanisms may have operated on the massive ground ice body subsequent to its deposition.

CONCLUSIONS AND DISCUSSION

We favour the interpretation that the massive ground ice body at Cape Shpindler is buried glacier ice, possibly older than 200 ka years. Evidence in support of a glacial origin are: (1) the massive ground ice underlies glacial till deposit; (2) the upper contact to the till is sharp and primarily an unconformable thaw contact; (3) the ice contains deformation structures and deformed sediment rafts that show a consistent direction of deforming push; (4) kinematic indicator structures such as boudins and sigma structures have developed in response to strong horizontal shear; (5) sediment rafts bridge the massive ice and the overlying diamicton, suggesting contemporaneous deformation and deposition of ice and diamicton; (6) changes in $\delta^{18}O$ values in profiles correspond with different ice facies instead of being gradual which is common in segregation ice where freezing continuously progresses downwards and where sediment and water have the same water source; (7) significant deviations in $\delta^{18}O$ profiles can be expected in glacier ice, and the pattern of $\delta^{18}O$ in the Cape Shpindler massive ground ice confirms with what could be expected for deformed glacier ice; and (8) the co-isotope slope of 7.5 is best explained by atmospheric source for the ice, as is typical for glacier ice, whereas lower co-isotope slopes are expected for segregation due to isotopic fractionation. Each of these individual lines of evidence from Cape Shpindler are not conclusive, but, when combined, provide cumulative support for a glacial origin.

Our conclusion is that the massive ground ice of Cape Shpindler is buried glacier ice, and that it possibly is older than 190–200 ka years. Leibman *et al.* (2000, 2001) focused on the petrography, geochemistry and stable isotopes of the massive ground ice body, and concluded that it was intrasedimental in origin. Re-examining their data, we find that neither petrographic characteristics of the ice nor ice chemistry give conclusive results as to the genesis of the ice. The stable isotopic composition and profiles conform with deformed basal glacier ice, and are more difficult to reconcile with intrasedimental ice. Our investigation takes into consideration the geometry of the ice body, as well as the stratigraphical and structural characteristics of the ice. These cannot be overlooked, and studies on geochemistry or crystallography of the ice cannot explain them.

There is ongoing debate as to the genesis of massive ground ice in western Siberia. Many geocryologists, mostly using geochemical methods, deny the possibility of a glacial origin (e.g. Lazukov, 1972; Zubakov, 1972; Ansimova and Kritsuk, 1983; Dubikov, 1983; Dubikov and Ivanova, 1988; Kritsuk and Ansimova, 1985; Kritsuk and Chervova, 1985; Vasilchuk, 1992; Leibman, 1996; Michel, 1998; Leibman *et al.*, 2000, 2001). Many Quaternary geologists, mostly using stratigraphical and sedimentological methods, favour burial of glacier ice based on similarity of sedimentary facies and ice structures to those described from recent glacial environments (e.g. Kaplyanskaya and Tarnogrodsky, 1986; Astakhov and Isayeva, 1988; Gataullin, 1988; Astakhov *et al.*, 1996; Forman *et al.*, 1999, 2002; Alexanderson *et al.*, 2001; Manley *et al.*, 2001). Even though it is acknowledged that the origin of massive ground ice in western Siberia can be attributed to either the burial of glacier ice or *in situ* segregation of intrasedimental ice, few studies combine stratigraphical-sedimentological-structural and geochemical-petrographical studies. Our study adheres to the advice of Mackay (1989) and French and Harry (1990) that it is important to consider as many criteria as possible before a genetic interpretation of a massive ice body is attempted, and shows that a combination of methods can provide cumulative data for interpreting the genesis of massive ground ice bodies.

Our results support the conclusion of Kaplyanskaya and Tarnogrodsky (1986), Astakhov and Isayeva (1988), Astakhov (1992) and Astakhov *et al.* (1996) that massive ground ice of glacial origin probably exists widely inside the limits of former glaciation in western Siberia. Massive ground ice, interpreted as buried glacier ice older than 40 ka years, has been described from a number of sites (e.g. Kaplyanskaya and Tarnogrodsky, 1986; Astakhov and Isayeva, 1988; Gataullin, 1988; Astakhov, 1992; Astakhov *et al.*, 1996; Forman *et al.*, 1999, 2002). Most authors suggest the buried glacier ice to be of Weichselian age (i.e. younger than MIS 5e). French and Harry (1988) suggested that buried glacier ice identified on southern Banks Island, western Canadian Arctic, could have been preserved for tens of thousands of years. Once glacier ice becomes buried beneath sediments and incorporated within permafrost, it might survive for more than one interglacial. Much of the permafrost in Siberia is a relict of Pleistocene climate conditions (Astakhov *et al.*, 1996; French, 1996, 1998). Kudryavtsev *et al.* (1978) mapped the thickness of permafrost in Yugorski Peninsula as being 200–400 m, implying that it is in disequilibrium with present climate and largely relict. In western Siberia, huge bodies of glacier ice may have survived through the Holocene without melting (Astakhov *et al.*, 1996). The present occurrence of sub-sea permafrost in the Kara Sea along western Yamal Peninsula (Kudryavtsev *et al.*, 1978), developed during periods of low global sea levels during the last glacial maximum, also illustrates the preservation potential of relict permafrost.

It has been suggested that relict glacier ice preserved in permafrost in east Antarctica might be of Miocene age (>8 million years (MY) old) and reflects the stability of the Antarctic cold-climate system (Sugden *et al.*, 1995; Schaefer *et al.*, 2000). The history of permafrost in the Russian Arctic has been traced some 2–1.5 MY years back in time and it is presumed that much of the deep permafrost is relict (Arkangelov *et al.*, 1989; Astakhov *et al.*, 1996; Rozenbaum and Shpolyanskaya, 1998). Zemtsov (1976) suggested that relict permafrost in Siberia might have formed in the Middle Pleistocene. Our chronology suggests that the glaciation that deposited the Cape Shpindler massive ground ice might be older than 190–200 ka years, perhaps of Drenthe age (MIS 8; 300–250 ka years, BP). This means that the buried ice body has survived two interglacial periods (Eemian and Holocene), as well as several interstadials. This has consequences for our perception of the potential of relict permafrost to preserve long stratigraphical archives and signs of ancient glaciations.

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