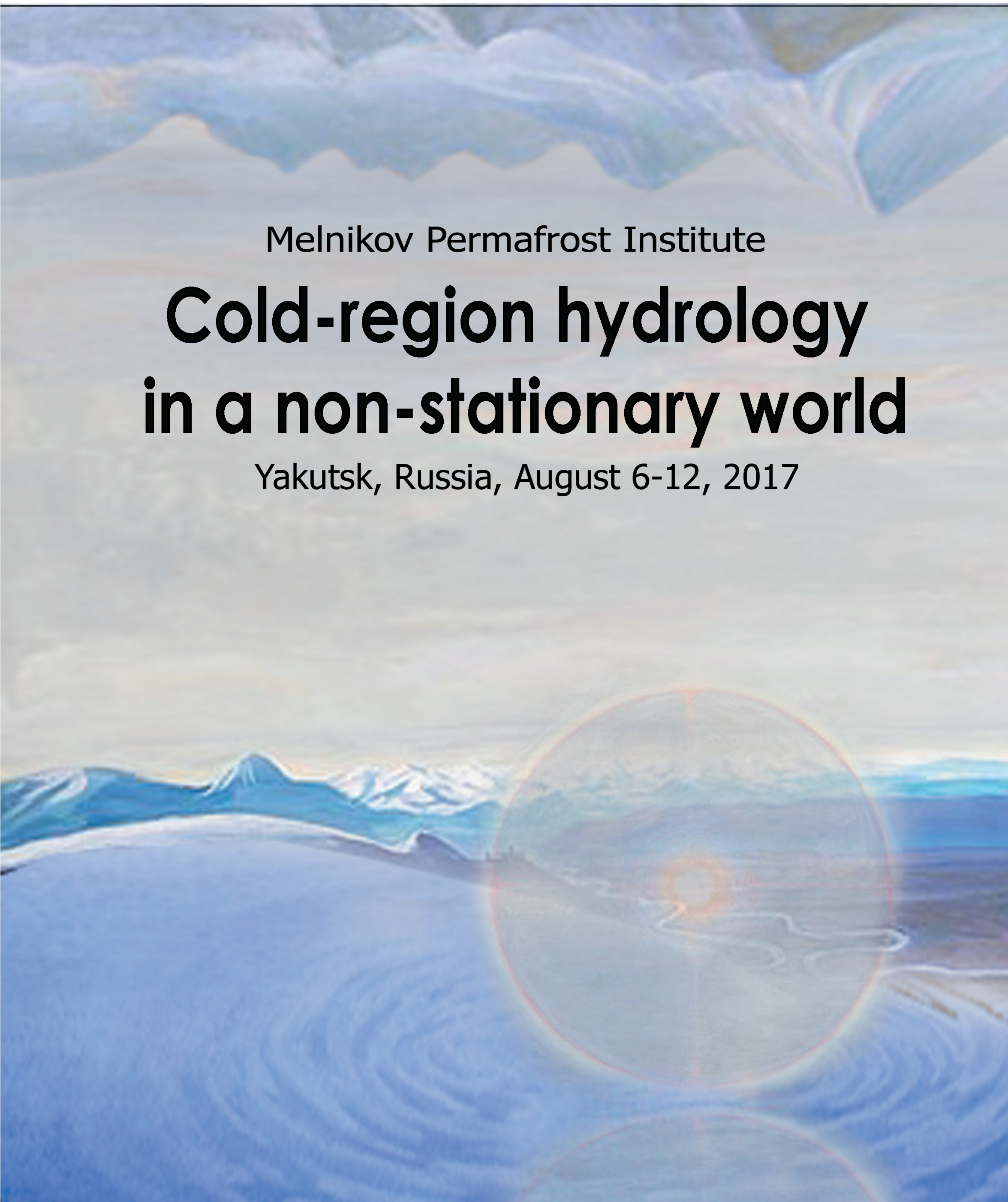
21st Northern Research Basins Symposium and Workshop

Melnikov Permafrost Institute

Cold-region hydrology in a non-stationary world

Yakutsk, Russia, August 6-12, 2017



Melnikov Permafrost Institute
Siberian Branch, Russian Academy of Sciences

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Труды XXI Международного симпозиума по исследованиям северных речных бассейнов «Гидрология холодных регионов в нестационарных условиях»

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Preface

From the old times, cold regions have always been the attraction for strong, determined and free people. Their will for exploration and adventure has led to many great scientific discoveries which would be impossible without self-sacrifice, friendship and mutual help, the admiration of natural beauty, dreams and hopes.

We live in a non-stationary world, not only in the terms of climate and natural environment changes, but also in the sense of the global political situation and general uncertainty of the future. Though the Arctic is still in many aspects a blank spot on the map of our scientific understanding, it has become the field of battle for mineral resources and global dominance.

At the same time, to assess the situation in hydrology, in the last 30 years we have lost about half of our research stations in the cold regions, meaning that even footworn paths are being recently grassed, and the important decisions are made mainly based on momentary profit, rather than on sound science.

In the light of this reality, we host the 21st International Northern Research Basins (NRB) Symposium and Workshop. This year it is organized in Yakutsk, Russia practically without any state support reflecting the world-wide lack of understanding of the importance of northern research basins. However, the fact that the NRB Symposium and Workshop has been running for more than 40 years and that hydrologists continue to be interested in keeping this initiative alive give hope.

Our main hope is that “the fire of exploration” embracing experience, knowledge, inspiration, broad outlook, and love of nature of NRB participants is strong enough to be passed further and enkindle young generations of hydrologists.

In 1975, the International Hydrological Program (IHP) National Committees of Canada, Denmark/Greenland, Finland, Norway, Sweden, the USA and the USSR established the IHP working group on Northern Research Basins. The overall objective of the NRB working group is to encourage research in hydrological basins at northern latitudes where snow, ice and frozen ground play a dominant role in the hydrological cycle. In 1992, Iceland joined the group and Russia took over the responsibilities of the former USSR. In addition, countries with polar research programs are eligible for associate membership. Twenty productive symposia/workshops have been held to date: Edefors, Sweden (1975); Fairbanks, USA (1977); Québec City, Canada (1979); Ullensvang, Norway (1982); Vierumäki, Finland (1984); Houghton, USA (1986); Ilulissat, Greenland (1988); Abisco, Sweden (1990); Whitehorse-Dawson-Inuvik, Canada (1992); Spitsbergen, Norway (1994); Prudhoe Bay-Fairbanks, USA (1997); Reykjavik-Kirkjubærjarklaustur, Iceland (1999); Saariselkä-Murmansk, Finland/Russia (2001); Kangerlussuaq, Greenland/Denmark (2003); Luleå-Kvikkjokk, Sweden (2005); Karelia, Russia (2007); Eastern Arctic, Canada (2009); Bergen-Geiranger-Loen-Fjærland-Voss, Norway (2011); South-central Alaska, USA (2013), and Kuusamo, Finland (2015).

The location of the 2017 NRB Workshop and Symposium is, as always, unique. The Republic of Sakha (Yakutia) is the largest region of Russia (total area 3.1 million km²). More than 40% of its territory is situated within the Arctic Circle. Yakutsk is the oldest (almost 400 years), largest (300 000 inhabitants) and coldest (average January temperature -39°C) city built in the zone of continuous permafrost. Yakutsk is built as a port at the Lena River, the tenth largest river in the world. This land is unique also by people. They need to have hot hearts to live in such a cold place. They keep traditions and are strongly tied to their motherland.

The main topic of the 21st NRB Workshop is “Cold-region hydrology in a non-stationary world” which addresses the issues of climate and environment change impacts on the hydrological cycle in Arctic regions, with implications for society. The Symposium handles the problems of hydrological research, both in fundamental scientific and applied aspects, including the studies of snow, glaciers, permafrost, frozen ground, groundwater, and seasonally frozen rivers and lakes.

Observational techniques have evolved dramatically in recent years, including remote sensing technologies, the expansion of isotope tracers methods, etc. Despite this, research watersheds are still the main scientific laboratories for gaining understanding and insights into the amazing journey of water on our planet.

Acknowledgments

First of all the organizers would like to thank Melnikov Permafrost Institute of the Siberian Branch of Russian Academy of Science (MPI) in the person of Dr. Mikhail Zheleznyak for hosting the 21st International NRB Symposium and Workshop. We must also recognise the general support of MPI for field and modelling hydrological research in permafrost environment.

Our special recognition is given to the Working Group of MPI on organization of the Workshop. Anna Fedorova did proof and layout of the Proceedings. Transport Department of MPI was responsible for transport organization. Printing Office of MPI helped with publishing of the Workshop Proceedings.

Graphic designer Elizaveta Ivanova-Efimova created the logo and all symposium visual materials.

Gidrotehproekt Ltd (St. Petersburg) helped with sorting out finances and all the fiscal matters for the event.

Yakutniproalmaz Research and Design Institute (Yakutsk) has funded simultaneous translation organized by the Department of International Relations of the Sakha Republic (Yakutia).

The organization of the 21st International NRB Symposium and Workshop would not be possible without the initiative and help of Hydrograph Model Research Group, especially junior researcher of MPI Liudmila Lebedeva.

On behalf of the 21st International NRB organizing committee,
Olga Makarieva

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Symposium Papers

SUPRAPERMAFROST TALIKS IN THE SHESTAKOVKA RIVER CATCHMENT, CONTINUOUS PERMAFROST ZONE, INVESTIGATED BY ERT TECHNIQUE

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ABSTRACT
Suprapermafrost groundwater and taliks are abundant yet poorly studied phenomena in continuous permafrost zone. In this study, we estimate talik depth and geometry in the two key sites in Central Yakutia using electrical resistivity tomography. It was shown that talik thickness reaches 3-7 m at 2-10 m depth in the Left Shestakovka River key site. ERT results suggest that talik in the Malaya Chabyda Lake key site could nowadays be gradually freezing. These results highlight the added value of geophysical techniques for assessing spatial patterns of permafrost, groundwater and talik distribution in continuous permafrost environment.

KEYWORDS
Permafrost; ERT; talik; suprapermafrost groundwater

1. INTRODUCTION
Geophysics techniques could provide spatially and vertically distributed information about permafrost, taliks and groundwater which cannot be obtained from drilling data alone. Non-frozen water-saturated deposits are characterized by substantially lower electrical resistivity than frozen sediments. Electrical survey could be used for estimation of geometry of aquifers and frozen aquitards. Electrical resistivity tomography (ERT) is a recent modification of vertical electrical sounding method. Large amount of electrodes allows prompt estimation of electrical resistivity of the studied section. The method was successfully used in different geological environments characterized by complex structure: permafrost, inclined bedding, karst, etc. (Bobachev, 2006).

The study aims at 1) assessment of ERT applicability for investigation of shallow (less than 10 m) suprapermafrost taliks in conditions of poor ground connection of electrodes; 2) define geometry of water-saturated suprapermafrost taliks in sand deposits in Central Yakutia.

2. FIELD SITE
The study area is located in 20 km to south-west of Yakutsk within the erosion-denudational slope of the ancient accumulative plain with absolute elevation of 190-210 m (Fig. 1). The first key site is situated on the right bank of the Left Shestakovka River. The second key site is located on the eastern shore of the Malaya Chabyda lake. The distance between the sites is 5 km. Both sites are characterized by gentle slope with western and north-western aspect. Suprapermafrost subaerial water-bearing taliks were discovered in the depth interval from 2.5 to 7 m. Both sites have a similar geological structure. The upper 40 m of the section are represented by quartz-feldspar sands, sometimes with inclusions of silty sandy loam and loam. The drilling showed that the water-saturated talik thickness varies from 0.1 m to 4.5 m, and its upper boundary is at 1.5-4.1 m depth. The seasonally thawing layer varies from 0.5 to 3 m. Water mineralization typically varies between 20 and 60 mg/l. Fully saturated deposits could be ferruginous (Anisimova, 1980).

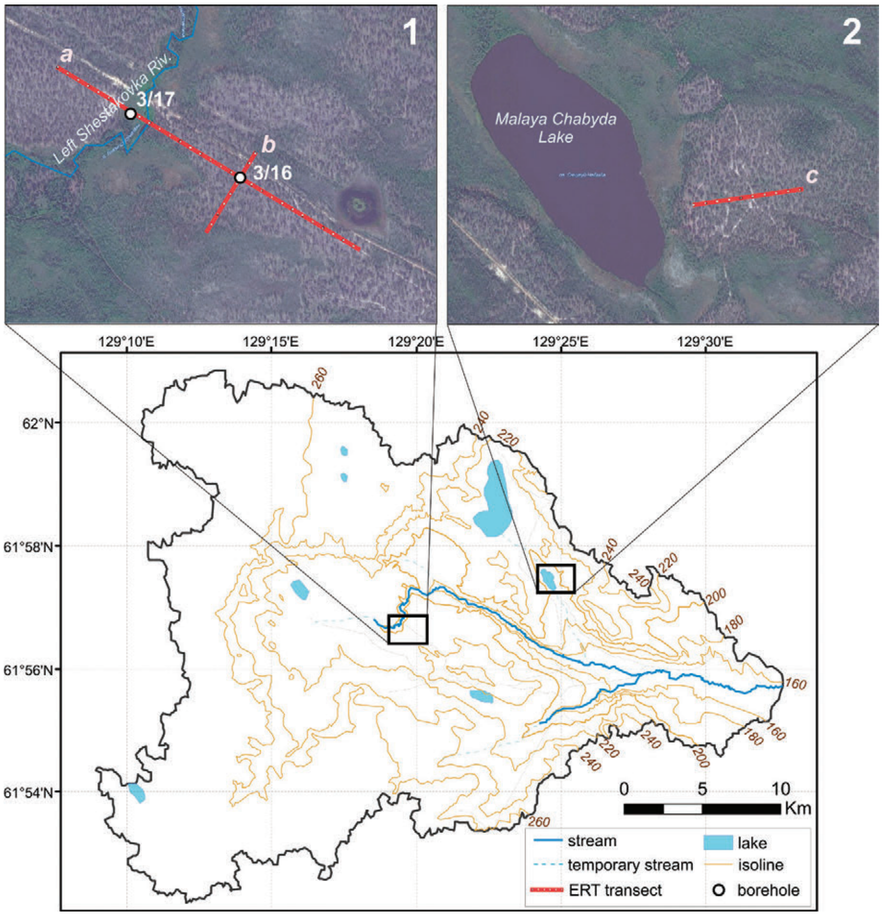

Figure 1. Scheme of the studied area: 1 - Left Shestakovka River key site, 2 - Malaya Chabyda lake key site.

3. RESEARCH METHODS

16-channel ERT and IP instrument SibER-64 produced by "Siber-instruments" (Novosibirsk, Russia) was used for ERT measurements. Two cables consisting of 32 copper wires were employed. Ground connection was performed by means of steel rods 300 mm long. Unit electrode spacing varied from 1 to 5 m depending on required level of detail and studied depth. Dipole-dipole array, gradient array and Schlumberger array with four electrodes were used for measurements. Data processing and interpretation were performed using RiPPP and ZondRes2d software packages.

Two ERT transects in the Left Shestakovka key site were conducted on 8th of June and 18th of August 2016. The first transect was oriented perpendicular to the channel from NW to SE and had a length of 955 m with 5 m unit electrode spacing. Second one had a length of 255 m with 5 m unit electrode spacing and was oriented from NE to SW in parallel with the Left Shestakovka River in 300 m from the channel.

The ERT transect in the Malaya Chabyda key site was done on 23rd of May 2017. It was 300 m long with 2 m unit electrode spacing.

Borehole drilling was performed by drilling device URB-2 installed on the KamAZ truck. The boreholes No 3/16 and No 3/17 were drilled in April 2016 and April 2017 accordingly at the Left Shestakovka study site on the ERT transect. There is no recent drilling at the Malaya Chabyda Lake key site.

Interpretation of the geoelectric section was performed based on the assumption that electrical resistivity of water-saturated sand is less 2000 Ohm*m. Deposits with electrical resistivity

higher than 2000 Ohm*m were treated as frozen or extremely dry. Such approach to the data interpretation was based on the drilling results. The borehole 3/16 intersected a water-saturated talik from 1.7 to 7.2 m depth. Borehole 3/17 showed that sediments in the Left Shestakovka River valley are frozen from the surface to 20 m depth.
Garmin GPS map 62st navigator was used to reference geophysical transects.

4. RESULTS

Extremely dry sand deposits near the land surface were characterized by high earth resistance of electrodes that prevented high-quality measurements. The problem was partially solved by salt water pouring. Another problem was high resistance layer in the frozen sand deposits below the talik. It limited maximum depth of reliable data to the 15 m even using the maximum generator power. Repeated measurements of the same transect using three different arrays showed that gradient array provides the highest data quality. Such conclusion was drawn on the base of analysis of deviations between several measurements.

According to the first ERT transect in the Left Shestakovka key site the upper 2 m layer is composed by dry sand with electrical resistivity from 2 000 to 10 000 Ohm*m. It corresponds to seasonal thawing or freezing layer. Frozen sand deposits lie beneath the seasonal thawing layer at the left bank of the Left Shestakovka River. Talik is found on the right bank of the river from the terrace edge to the end of the transect (appr. 650 m long). Electrical resistivity of the talik deposits is less than 2 000 Ohm*m. Talik thickness is 3-7 m. By the time of ERT measurements seasonally frozen layer above the talik had entirely thawed. In the upper part of the gentle slope in 400-630 m from the terrace edge seasonally frozen layer had low water content. Its electrical resistivity was 2 000 – 10 000 Ohm*m. Talik is underlined by frozen deposits with resistivity of 2 000 – 10 000 Ohm*m. Areas with resistivity below 1 000 Ohm*m are probably related to zones with higher water filtration and ferruginous deposits. There is river talik with thickness of 2.5 m in the valley.
Geoelectric section along the second ERT transect in the Left Shestakovka key site showed talik at 2-10 m depth. Talik is divided into three water-saturated fragments. They are oriented from NW to SE and have maximum thickness of 5.8-6.5 m.
Two ERT transects in the Left Shestakovka key site have a good agreement with each other. Talik depth is verified by drilling of the two boreholes along the ERT transect.
ERT transect in the Malaya Chabyda lake key site is not verified by drilling. Interpretation results are preliminary. There are several separate zones on the geoelectric section with resistivity lower 2000 Ohm*m that could contain water. Only in 260-300 m from the left side of the transect there is zone with resistivity below 1000 Ohm*m. It most probably corresponds to the water-saturated talik. Zone with resistivity between 2000 and 4000 Ohm*m is allocated on the whole transect with thickness from 3-4 m to 10 m. The zone could probably relate to recently frozen talik. By the moment the only drilling along the transect was performed in 1980s. According to historical data (Anisimova, 1980) talik was revealed at 1.5-4 m depth and had thickness 0.2-4 m.

5. CONCLUSIONS

It was shown that water-saturated taliks are abundant phenomena in the pine forests in Central Yakutia, continuous permafrost zone. ERT technique was proved to be as effective method to explore geometry and structure of the taliks. Repeated measurements of the same transect using three different arrays showed that gradient array provides the highest data quality in conditions of poor ground connection of electrodes in extremely dry sand deposits. It was revealed that talik in the Left Shestakovka key site spreads over 600 m along the north-western gentle slope and is divided into several water-saturated fragments. ERT measurements showed that talik in the Malaya Chabyda lake key site most probably is nowadays gradually freezing. These results highlight the added value of geophysical techniques for assessing spatial patterns of permafrost, groundwater and talik distribution in continuous permafrost environment.

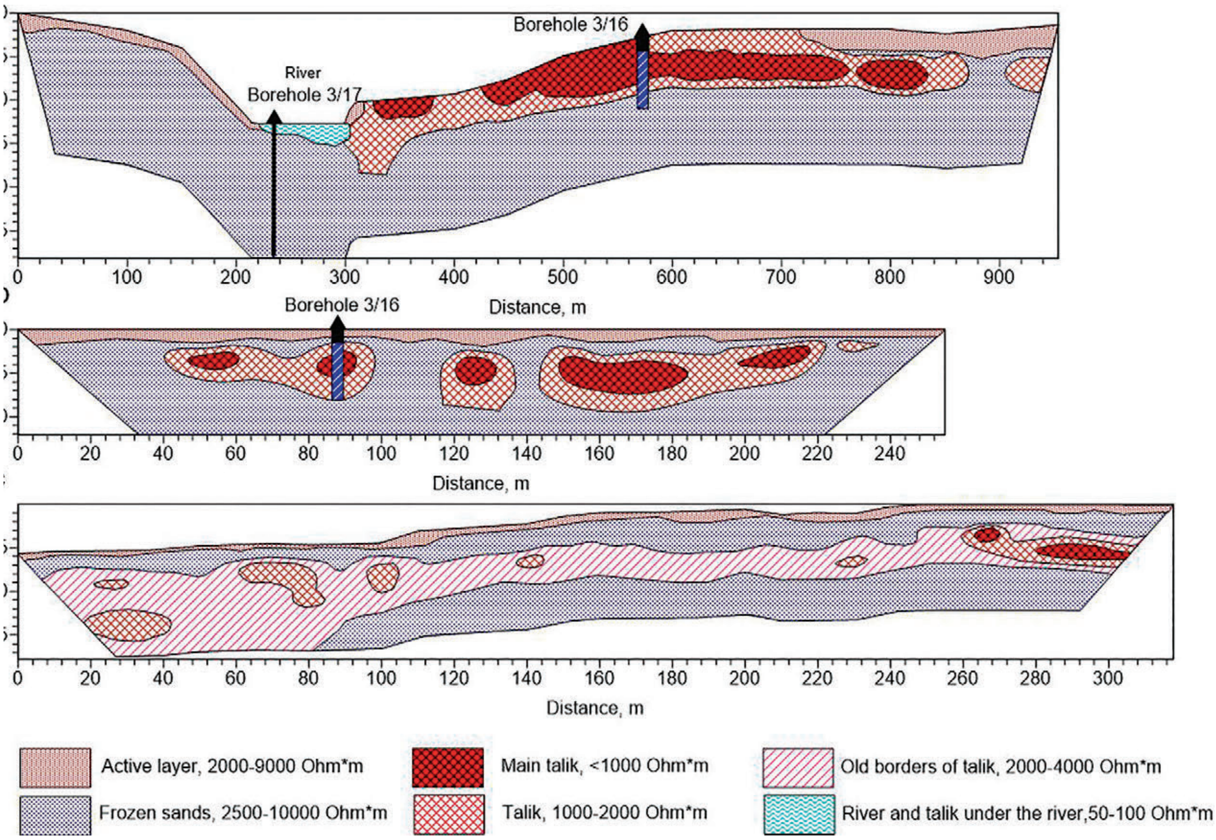

Figure 2. Permafrost, geological geoelectric sections developed on the basis of ERT measurements: a – first ERT transect on the Left Shestakovka key site; b – second ERT transect on the Left Shestakovka key site; c – third ERT transect on the Malaya Chabyda lake key site.

6. AKNOWLEDGMENT

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THE AEOLIAN-CRYOGENIC ORIGIN OF THE INTER-PERMAFROST TALIKS AND UNDERGROUND WATER SOURCES IN CENTRAL YAKUTIA

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ABSTRACT
The alternative mechanism of origin of the inter-permafrost taliks and underground water streams is based on complex study of the structure, absolute age and the spatial distribution of dune formations in Central Yakutia. The main idea is a relatively fast burial of the primary drainage (river) net and sub-aerial taliks of the MIS 3 (thermochron) with active dune massifs during MIS 2 (cryochron). The dune's sands are characterized by extremely low thermal conductivity, the deep foot of AL and very small AHRL depth, as it was pointed above, that is the main reason of inter-permafrost taliks preservation.

KEYWORDS
Interpermafrost taliks, tukulans, dune deposits, Central Yakutia, sources of drinking water, permafrost

1. INTRODUCTION
Completely organized massifs of parabolic sandy dunes (tukulans) occupy large areas in middle basin of Lena river and its trubitaries - Vilue, Linde, Tyung, Blue and others. They are divided on actively moving unvegetated dunes of different hierarchy and more ancient sand massifs, vegetated with pine and birch forests. First one area reaches up to 3200 km^2, the sercond exceeds of 60000 km^2 in Central Yakutia (Galanin et all., 2016).
Central Yakutia is usually characterized by Continuous Permafrost (CP) up to 300-400 m thick with temperatures -3 ... -5 °C at the base of Annual Heat Rotation Layer (AHRL). The thickness of the Active Layer (AL) varies from 0.3 to 1.5 meters (Geocryology, 1989). The groundwater of the Region is represented by above-permafrost, inter-permafrost and sub-permafrost types. The last one lies below the foot of CP and it usually has a high mineralization of 1000-1300 mg/l, reaching in some cases up to 30-40 g/l (Ponomarev, 1960). The above-permafrost (aerial) type is low-mineralized subsurface waters, which are forming at the foot of AL due to atmospheric precipitation.

2. METHODS
To substantiate the hypothesis of the eolian-cryogenic origin of inter-permafros taliks and underground sources of Central Yakutia discussed below, we carried out the integration of collected field and published data, including cryolithological and paleogeographic analysis of the key sections, radiocarbon data (Galanin et all., 2015, 2016), the results of geospatial analysis of the aeolial landforn and river network distribution in the region (Urban & Galanin, 2013).
The phenomenal feature of the permafrost structure of Central Yakutia is specific inter-permafrost (sub-aerial) taliks and associated high-yield groundwater outlets of high-quality low-mineralized drinking water. Most of them confined to local catchments on the surface of high terraces of Lena and Vilyui rivers (Boitsov & Shepelev, 1976). The depth of inter-permafrost taliks reaches up to 70 m and more, the groundwater is usually gydrocarbonic-magnesium-

calcium with mineralization about 20-80 mg/l (Shepelev, 2011). Located in downstream of Viluy river, the Makhatta inter-permafrost outles has discharge 760 l/s, the talik's depth reach about 100 m.

Well watered inter-permafrost taliks and corresponding springs are commonly associated with the Late Pleistocene sand dune massifs that occupy up to 30% of the territory of Central Yakutia. There are well-known ground streams Eryu, Ulakhan-Taryn, Bulus ect. that located at the foot of high Lena terrace near Yakutsk. Significant part of inter-permafrost ground streams discharge directly into river flow, and their outlets outcrop in the summer low water. Sites of year-round underground discharge are accompanied with thermoerosion, suffosion and large frazils (Shepelev, 2011; Gagarin. 2012). The permafrost conditions inside sand dune massifs cardinally differ from adjacent territories.

3. RESULTS AND DISCUSSION

Grain size and mineralogy

Late Quaternary dune deposits of Central Yakutia are composed of fine-grained sands and sandy loams, quartz (75-87%) and minerals of light fraction are dominated. There are some secondary minerals (ferromanganese tubes and quartz fulgurites). The dune sand are characterised by the following statistics of the grain size: average grain size x=301±41 (μm); sorting koeff. σ=1,6±0,1 (φ); asymmetry α=-0,3±0,1 (φ), excess τ=0,94±0,2 (φ). The constitution, mechanical and thermal properties of dune deposits provide good filtration and reservoir properties.

Structure and texture

Dune deposits, except for aquiferous talik zones, are characterized by flat (subhorizontal) and cross-bedded sedimentary textures, very small ice or water content humidity 1-5% in comparison with common frozen fine grained soils (15-25% and more) (Katasonova & Tolstov, 1963). The moisture, contained in the dune sediments, is usually presented in the form of contact or thin-schlieren (very thin membrane-shaped lenses up to 1 mm depth) cryotextures, formed by the condensation of water vapor. In comparison, sandy alluvial deposits are characterized by the more diverse grain size composition, oblique and lenticular stratification, they are intensely saturated with ice-cement and have massive cryotexture (Galanin et all, 2016).

Ground temperatures

The temperature regime of vegetated and unvegetated dune massifs is also specific and radically different from other types of frozen soils of adjacent landscapes (Figure 1). The thickness of AHRL does not usually exceed 4-5 m, but the thickness of AL reach the sane values. A the foot of AHRL temperature ranges from 0 to -1 °C, while the temperature of alluvial deposits with massive cryotexture varies from -2.5 to -3.5 °C (Katasonova & Tolstov, 1963).

Radiocarbon age

Based on radiocarbon data it was concluded that aeolian processes and landforms reached maxima on last cryochron (MIS 2) between 24-12 kyr [Galanin et al., 2016]. In first half of the Holocene (12-6 kyr) significant part of the dune massifs was vegetated, and during the last climate cooling (1-0,5 kyr) aeolian processes reactivated.

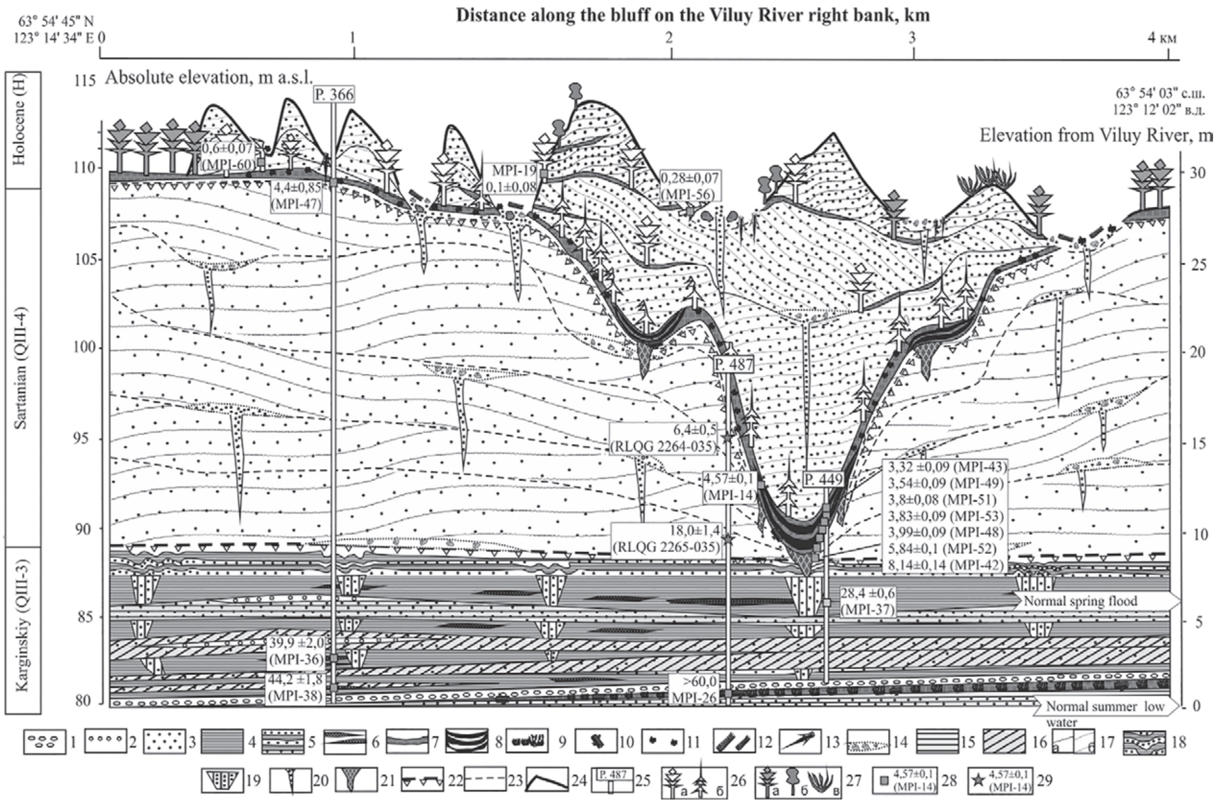

Figure 1. The structure and absolute age of the key-section of Kyzyl-Syr dune massif (tukulan) in the outcrop of the right bank of Vilyuy River bank. Lithology (1-14): 1- pebbles, 2 – fine gravel, 3 – medium sand, 4 – silty sandy loam, 5 – alternation of thin sand and loamy layers, 6 – thin interlayers and lenses of pre-plant detritus, 7 – soil horizons, 8 – lenses of peat bogs, 9 – mature horizons of shrub peat, 10 – fragments of wood, 11 – accumulations of charcoal, 12 – soil Fe-Mn orshtein and pedotubes, 13 – accumulations of fulgurites, 14 – accumulations of small windmills. Syngeneic structures (sedimentary): 15 – horizontal; 16 – cross-bedded; 17 – sloping (a) and steep (b) cross-bedded. Epigenetic structures: 18 – wavy cryoturbated; 19 – wide polygonal sandy veins; 20 – thin sandy veins along cracks of settling; 21 – frozen infiltration-humus veins; 22 – established major deflationary disagreements; 23 – alleged secondary deflationary disagreements; 24 – the surface of the modern mantle; 25 – supporting sections; 26 – dead vertically buried pine (a) and larch (b); 27 – live pines (a), birch (b) and cedar curtains (c); 28 – radiocarbon dating; 29 – OSL-dating

Hypothesis of the inter-permafrost talik's origin

In spite of the good permafrost-hydrogeological study of intermodal taliks and related sources (Shepelev, 2011), their genesis remains controversial. The official theory is formulated in the collective monograph "Geocryology of the USSR" (Geocryology, 1989). It is based on the assumption of the possibility of deep epigenetic thawing of originally frozen sandy sediments in the first half of the Holocene during the warm Boreal Optimum and subsequent partial freezing from above in the second half of the Holocene.

At the same time, the well-known thermophysical features of the sand dune massifs themselves seem doubtful on the possibility of their thawing to a depth of up to 100 m in a relatively short (several thousand years) time interval, taking into account the fact that the entire sand massif was intensively frozen and cooled throughout the Last cryochron (MIS 2). It is also surprising, that in sand massifs the evidence of secondary permafrost deformations (cryoturbation) of the initial stratification were not established, despite the fact that they usually accompany the permafrost thawing. In addition, other types frozen sediments of surrounding permafrost landscapes, including the ice-rich formation ("ice complex") with polygonal ice wedges, they have no sign of significant degradation during the Holocene optimum.

It seems the urgently low thermal conductivity of sandy dune deposits itself is a big obstacle not only for the hypothetical thawing, but also for re-freezing during the Late Holocene. Moreover, the paleoclimatic parameters of the Holocene climatic optimum remain not completely definite and controversial in the Region.

Exothermic processes of moisture condensation from circulating atmospheric air were discussed by some researchers to explain the temperature non-stationarity of dune massifs in Central Yakutia. However, even the presence of condensation processes, that experimentally established in dune formations (Shepelev, 2011), do not at all clarify the initial cause of the formation of such deep inter-permafrost talik net both during the Holocene optimum and after it. Undoubtedly, the condensation processes could play a significant role in the modern thermal and water balance of the dune massifs, but they are more a consequence of the functioning and dynamics of this thermal system than the initial cause of its origin.

Aeolian-cryogenic hypothesis

The following alternative mechanism of origin of the inter-permafrost taliks and underground water streams is argued on the base of complex study of the geomorphology, stratigraphy and the spatial distribution of dune formations in Central Yakutia (Galanin et. all, 2015; 2016) . The main idea is a relatively fast covering of the low-order initial drainage (river) net and its sub-aerial taliks of the MIS 3 (thermochron) with active dune massifs during the climate desertification in MIS 2 (cryochron). Small river valleys and their thalwegs have been systematically overlapping with the moving dunes during dry seasons, thus some watered underflow taliks were covered with sands. As it was pointed above, the dune's sands are characterized by extremely low thermal conductivity, the deep foot of AL and very small depth of AHRL, that was the main reason of the inter-permafrost taliks preservation. This conclusion is not only the result of cryofacial analysis of the key sections of Late Pleistocene dune formations (Galanin & Urban, 2013; Galanin et. all, 2016). It also follows by comparative analysis of the modern dune morphosculpture and buried paleorelief at their bed as well as it is confirmed by the structural-geomorphological and morphometric characteristics. Most bright feature is clearly observed recent eolian destruction of the drainage-erosion network on high Lena's and Viluy's terraces (Bestakh, Tabaginskaya, Tungulinskaya ect.). The fact of intensive disturbance of the ancient drainage (river) network as a result of significant surface modeling by epigenetic eolian processes is proved by methods of modern structural and morphometric hydrology, as well as by results of the geomorphological analysis of satellite images (Figure 2).

4. CONCLUSION

The radiocarbon data (Galanin et al., 2015; 2016) evidence the formation of inter-permafrost talik network in dune massifs of Central Yakutia occured since the end of MIS 3 thermochron (the pluvial stage) during MIS 2 cryochrone (cryoarid stage) and was completed in the Holocene (pluvial stage).The mechanism of formation is advisable to consider in the form of the several successive stages (Figure 3), each of which can be clearly observed on example of modern unvegetated dune massifs of Central Yakutia.

The first stage (Figure 3) characterizes the conditions of the pluvial epoch of the Karginskiy (MIS 3) thermochron (65-30 kyr B.P.), when the depth and thickness of the underflow and underlake taliks reach their maximum characteristics, the moderate activity of the precipitation and the run-off were sufficient to clean the thalwegs from annual sediments. More humid climate led to a decrease in the activity of aeolian processes and landscape stabilization in the region.

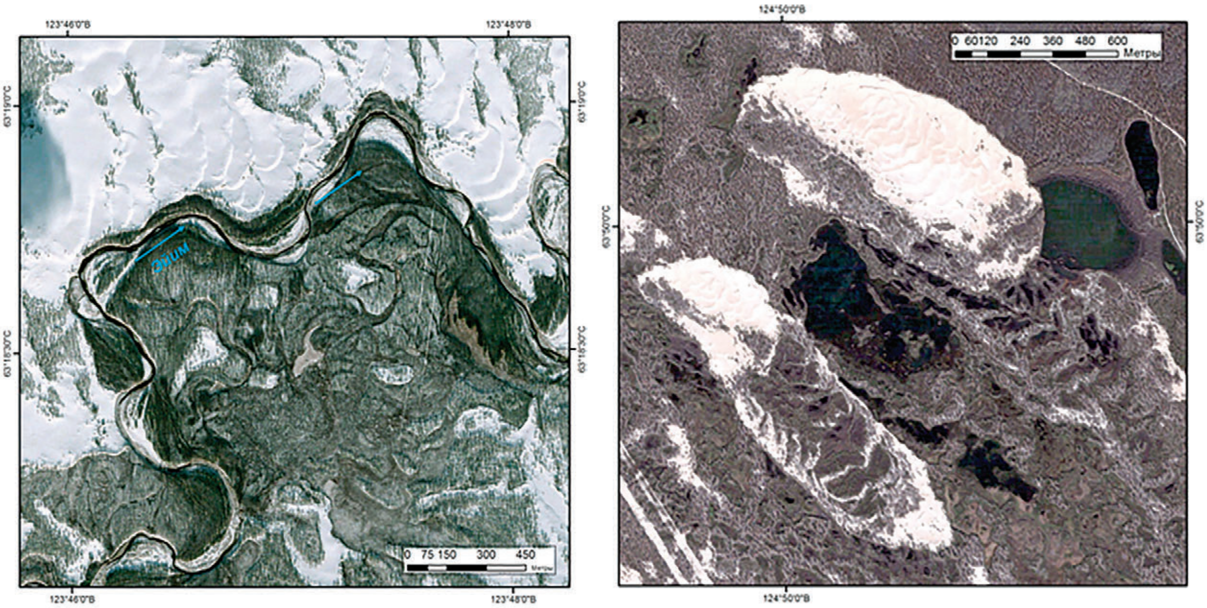

Figure 2. Examples of the modern eolian disturbance of the low-oder drainage network on the terraces of Viluy trubitaries (Cenfral Yakutia). Wight landforms are unvegetated dune massifs moving in South-East direction. the Eyim river valley is blocking with parabolic dunes (right picture); the polychronous dune massifs are overlapping of the lake depressions (left picture)

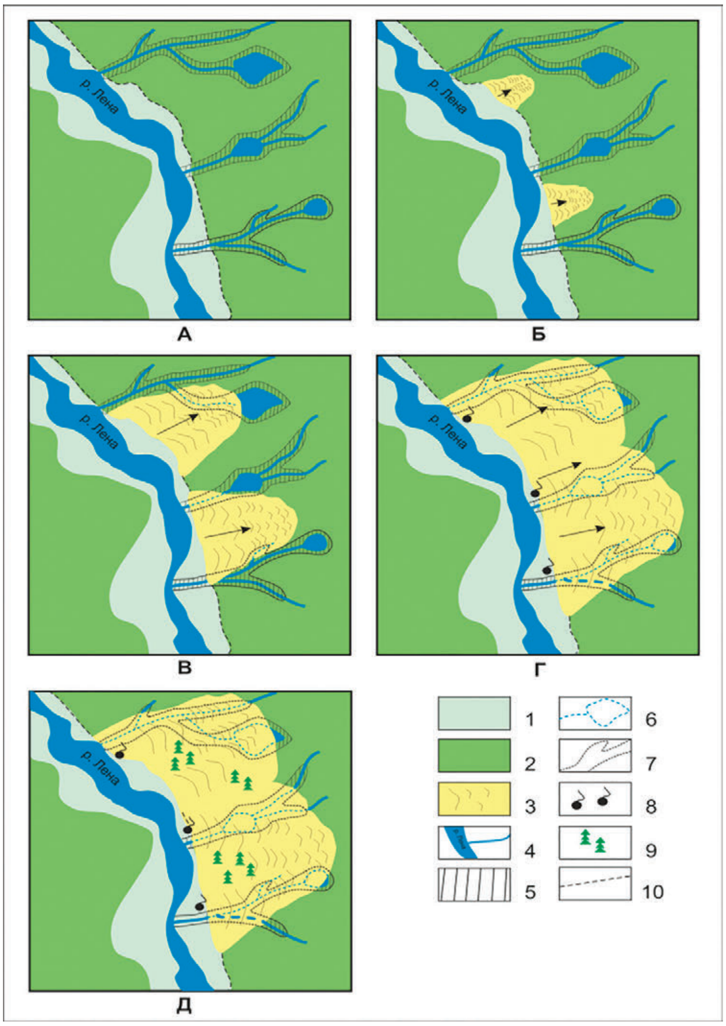

Figure 3. Principal mechanism and main stages of origin of inter-permafrost taliks and ground sources of Central Yakutia (on the example of the Bestyah terrace of the Lena river). Legend: 1 – modern floodplain, 2 – the initial Lena Plain with a drainage-erosion network in the end MIS 3 (Karginskiy thermochron), 3 – sandy dune massifs of MIS 2 (Sartan cryochron), 4 – the modern riverbed of the Lena river and its right tributaries, 5 – zones of the above-permafrost (aerial) taliks of MIS 3 (Karginskiy thermochron), 6 – talik zones buried under the sand dunes, and thawing basins under the lakes, 7 – boundaries of the buried taliks, 8 – sites of the inter-permafrost talik discharge (underground water outlets), 9 – dune massifs vegetated with pine forests, 10 – the edge of the Bestyah terrace of the Lena river

The second stage is associated with the beginning of the global cooling about 28 kyr B.P. (MIS 2, Sartan cryochron) that accompanied by reducing precipitation, and landscapes desertification. The role of forests and shrubs vegetation is reduced, partial degradation of the soil cover occurs. Primary single embryonic dunes began to form on the edges of river terraces due to their special aerodynamic properties. On the different modern examples you can see that the actively moving dune massifs in Central Yakutia are formed as a result of the destruction of the edges of river terraces. An importantissue is the development of embryonic dune massifs in the lower part of the tributaries basins with the subsequent movement and expansion upwards along valleys.

The third stage reflects the significant spread and expansion of dune massifs and their merging into systems of parabolic dunes of the higher orders. During the MIS 2 cooling and lowering of precipitation, the small rivers and streams were dehydrated for most of the year, and their valleys disturbed and blocked with moving dunes that could cause some flooding of the upper parts of the catchments which have been yet preserved. In other words, the equilibrium profiles of the watercourses of the low-level hydrosystem were destroyed, which led to the formation of a large number of closed lake basins (alasy) within the Prilenskaya Plain. Due to nice filtering features of aeolian sands the main part of the runoff began filtered through dune massifs following the original thalweg network.

The fourth stage characterizes the maximum of eolian activity and corresponds to the Global Thermal Minimum of MIS 2 (Sartan cryochron) that occured 20-18 kyr ago. There is a maximum disruption of the initial drainage net of the Kargin thermochron. Only the valleys and channels of large watercourses are preserved, inside Prilenskaya Plain the small-valley network is destroyed almost completely, with the preservation of severely deformed closed and semi-closed fragments of the primary thalweg net of the drainage basin. Simultaneously with the expansion of the dune massifs, their surface was subjected to maximum cooling and freezing, which leads to a partial isolation of the inter-permafrost aquifer zone from above and from the sides. Thus, the newly formed system of subaerial taliks partially inherited the structure of the initial surface drainage network. It is likely that at the peak of a strong cooling the frosts of this system of inter-permafrost runoff could occur and cause sudden flooding of the upper basins. It is likely that at the peak of a strong cooling the freezing of this system of inter-permafrost runoff could occur, and it cause sudden flooding of the upper basins and the appearance of ephemeral shallow lakes.

The fifth final stage reflects the end of the MIS 2 (Sartan cryochrone) about 12 kyr B.P., the warming and stabilization of the dune complex by the vegetation cover during the Boreal optimum of the Holocene (10-5 kyr B.P.). The increase of precipitation and climate humidity contributes to significant increasing of the runoff. In turn, this led to increased water erosion, the release of river channels from eolian deposits and new processing of the equilibrium profiles. This is clearly expressed at the present time in the form of a broad spectrum of thermodenudation, suffosion and erosion processes on the surface of the Prilenskaya Plain above the inter-permafrost taliks and on the sites of groundwater outlets. In general, the final stage is destruction of inter-permafrost taliks and their sources.

5. ACKNOWLEDGMENT

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RECENT CHANGES IN THE HYDROLOGIC CYCLE OF THE LENA RIVER BASIN, EASTERN SIBERIA

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ABSTRACT
The Arctic freshwater budget is a critical component to understand global climate, water cycling, and ocean circulation in the Arctic. The Lena River is one of the major sources of terrestrial freshwater inflow into the Arctic Ocean. In this presentation, several recent researches focusing on the hydrological issues will be revisited including: interannual variability in precipitation and atmospheric circulation patterns over northern Eurasia, trend detection of terrestrial water storage (TWS) changes in the Lena River basin, determination of permafrost thawing trends in several sub-basins of the Lena River, and the estimation of permafrost groundwater age in the middle part of the Lena River basin. Papers shown in the references will be mainly introduced in this presentation.

KEYWORDS
Arctic freshwater; atmospheric circulation pattern; interannual variability; permafrost groundwater; permafrost thawing trends; summer precipitation; terrestrial water storage (TWS)

1. INTRODUCTION
The Arctic freshwater budget is a critical component for the understanding global climate as well as water cycling and ocean circulation in the Arctic. Among all the rivers that flow into the Arctic Ocean, three Siberian rivers (Lena, Yenisei, and Ob) are the largest in terms of fresh water discharge (Oshima et al. 2015). Based on a decomposition analysis of the atmospheric moisture flux, Oshima et al. (2015) revealed that moisture transport associated with cyclone activity dominates the climatological features over the Lena, whereas that associated with seasonal mean winds dominates over the Ob. And both transport processes affect over the Yenisei. Because moisture transport over the three Siberian rivers is the result of atmospheric circulation pattern over northern Eurasia mainly in summer, it is meaningful to investigate the interannual variability in summer precipitation and atmospheric circulation patterns over the regions.

If we focus on the landscapes (surface and subsurface) of the Lena River basin, permafrost has been reported to be degrading at increasing rates over wide areas in the basin (Fedorov et al. 2014). The evidence has come mainly from in situ observations in the soil profile (e.g. Iijima et al. 2010), which have limited spatial and temporal coverages. And detection of changes in the hydrologic cycle of the basin is a critical issue in hydrology. Additionally, better understanding of groundwater dynamics in permafrost regions is needed to assess the vulnerability of the cryolithic water environment to changing climate (Hiyama et al. 2013).

In this study, we revisit recent papers listed in the references focusing on the atmospheric-terrestrial hydrologic cycles including TWS change and permafrost groundwater ages in the basin.

2. INTERANNUAL VARIABILITY OF SUMMER PRECIPITATION

We first investigated the interannual variability in summer precipitation and atmospheric circulation patterns over northern Eurasia using long-term Precipitation REConstruction over Land (PREC/L) and atmospheric Japanese 55-year Reanalysis data (JRA-55) from 1958 to 2012 (Hiyama et al. 2016). A special emphasis was placed on the recent increase in summer (June, July, and August) precipitation around the Lena River basin (Figure 1). We found interdecadal modulation in the relationships between interannual variability in summer precipitation and atmospheric circulation patterns among the three Siberian basins (Lena, Yenisei, and Ob). The interannual variations in summer precipitation over the Ob and Lena were negatively correlated from the mid-1970s to the mid-1990s. However, after the mid-1990s, this negative correlation became insignificant. In contrast, a significant positive correlation was apparent between the Yenisei and Lena. We also found that there has been a significant increasing (positive) trend in geopotential height in the low-level troposphere since the mid-1980s over Mongolia and European Russia, resulting in an increasing trend of westerly moisture flux into the Yenisei and Lena basins. Summer precipitation in both basins was continuously high from 2005 to 2008 under a trough that broadly extended from the Yenisei and Lena River basins, which has been a typical pattern of interannual variation since the mid-1990s. This trough increased the meridional pressure gradient between Mongolia and eastern Siberia in combination with the trend pattern. This further enhanced the eastward moisture flux towards the Lena River basin and its convergence over the basin, resulting in high summer precipitation from 2005 to 2008.

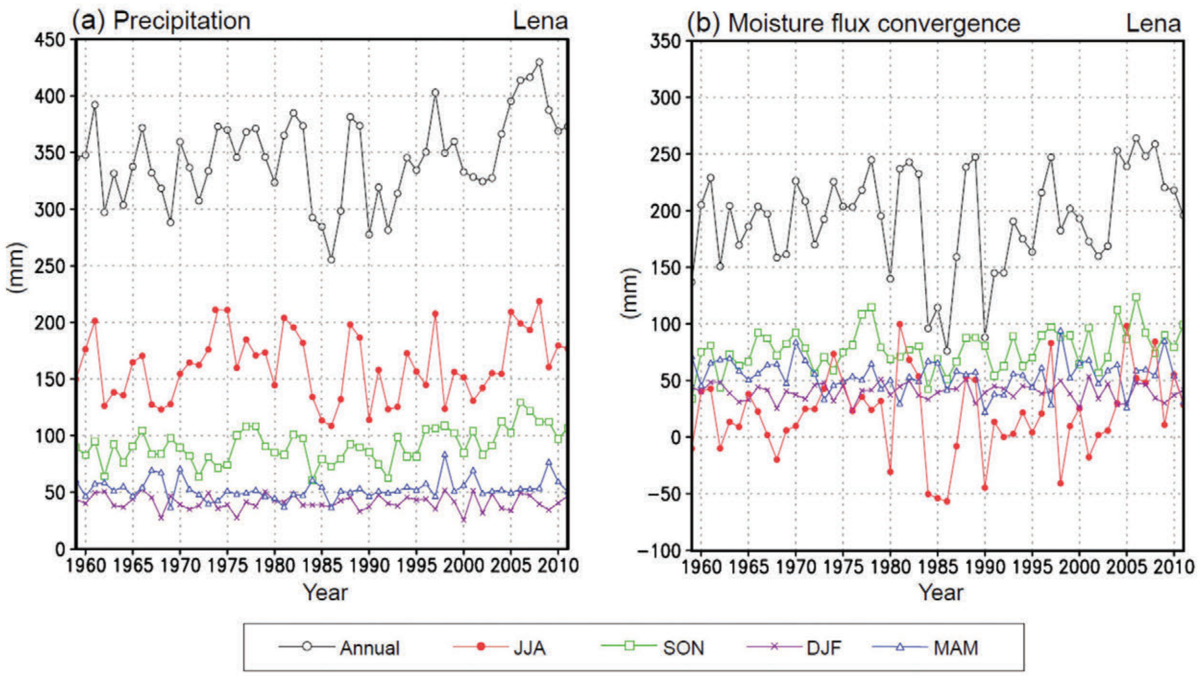

Figure 1: a) Time series of annual total precipitation (open circles; black) from 1959 to 2011 plotted together with summer (JJA) (closed circles; red), autumn (SON) (open squares; green), winter (DJF) (cross marks; violet), and spring (MAM) (open triangles; blue) precipitation in the Lena river basin. Annual values were cumulative from December of the previous year to November of the current year. The unit for annual values is mm year^{-1}, and for seasonal values is mm 3 months^{-1}. The precipitation data are from NOAA's Precipitation REConstruction over Land (PREC/L). b) Same as a) but for moisture flux convergence. The 6-hourly atmospheric reanalysis data from the Japanese 55-year Reanalysis (JRA-55) project was used for this calculation (see Hiyama et al. 2016).

3. TERRESTRIAL WATER STORAGE (TWS)

Then we used 146 months of data from the Gravity Recovery and Climate Experiment (GRACE) spanning the period from April 2002 to August 2015 in order to analyse TWS in the Lena River basin (Suzuki et al. 2016). We examined the lag correlation between TWS values and the annual run-off from the Lena River and found a strong linear relationship between a given year's river run-off and the TWS during November of the previous year. This relationship persisted throughout the winter until the following May. We also found a negative trend in TWS in the downstream portion of the Lena River basin, which might be explained by increasing evapotranspiration associated with warming summer air temperatures.

4. PERMAFROST THAWING TRENDS

In order to detect permafrost thawing trends, three methods were proposed to relate low river flows (or base flows) during the open water season with the rate of change of the active groundwater layer thickness resulting from permafrost thawing at the scale of the upstream river basin (Brutsaert & Hiyama 2012). The methods were tested with data from four gaging stations within the Lena River basin, one in the Upper Lena basin, and three in two of its tributaries, namely the Olyokma and the Aldan basins. The different results were mutually consistent and suggested that over the 1950–2008 period the active layer thickness has been increasing at average rates roughly of the order of 0.3 to 1 cm year^{-1} in the areas with discontinuous permafrost and at average rates about half as large in colder more eastern areas with continuous permafrost. These rates have not been steady but have been increasing. Thus, it appeared that in the earlier years over the period 1950–1970, some large regions have not been undergoing active layer thickness increases and perhaps even decreases, whereas from the 1990s onward vast areas have experienced larger average layer thickness increases, especially those with continuous permafrost. Interestingly, Tananaev et al. (2016) pointed out that no significant trends over either soil temperature at 3.2m or active layer thickness were found in Transbaikalia region (with discontinuous permafrost) in the past decades. These contradicting results between Brutsaert & Hiyama (2012) and Tananaev et al. (2016) should be investigated in future more intensively. Positively we agree to the opinion of Tananaev et al. (2016) who suggested local increase in active layer thickness, leading to talik development in discontinuous permafrost was important hydrologic processes in the region. In the next section, we will focus on the age detection of permafrost groundwater (including talik water) in the middle part of the Lena River basin with continuous permafrost.

5. RESPONSE OF PERMAFROST GROUNDWATER

In order to determine the residence time (age) of permafrost groundwater in the Lena River basin, transient tracers including tritium (^3H), chlorofluorocarbons (CFCs), and sulfur hexafluoride (SF_6) were used to analyse a mixture of supra-permafrost and intra-permafrost groundwater in the middle of the Lena River basin (Hiyama et al. 2013). Tritium analyses showed that the concentration ranges from 1.0 to 16.8 TU, and the apparent age of groundwater ranged from around 1 to 55 years. One of the spring waters appeared to contain more than 90% water recharged by precipitation before the 1960s nuclear testing era, and the water could be partly sourced from thawing permafrost. Comparisons of apparent groundwater ages estimated from different tracers implied that ^3H and CFC-12 were the most suitable for groundwater vulnerability assessments in this region. Because the apparent age was a mixture of those from supra-permafrost and intra-permafrost groundwater, further analysis would be required to assess the contribution ratio of the two types of groundwater.

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CLIMATE CHANGE IMPACTS ON SNOWMELT RUNOFF AND RIVER BREAK-UP ON THE PORCUPINE RIVER IN NORTHERN CANADA

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ABSTRACT
Because of the history of river transportation most Yukon communities are situated on the floodplain and subsequently are prone to flooding. Potential flooding mechanisms include spring ice jams, snowmelt, rain on snow, intense summer rain, glacier melt and surges, and freeze-up ice jams. The earliest Yukon flooding events in the annual cycle are triggered by spring ice jams which typically occur in late April or May. Ice jam flooding is also the most severe mechanism in terms of impact. With respect to timing spring break-up and associated flooding is followed by the snowmelt freshet several weeks later, typically at the end of May or early June. The village of Old Crow in Yukon's Arctic region has had several significant ice jam flooding events where substantial portions of the community have been inundated by flood waters. A change in spring snowmelt dynamics has been observed throughout Yukon in recent decades. Snowmelt and associated runoff is occurring earlier and has a shorter duration resulting in a compressed runoff period. Spring river ice break-up and the snowmelt freshet on the Porcupine River have been normally separated by up to several weeks. With changing snowmelt dynamics overlapping river ice break-up and snowmelt freshet events are becoming more common. The combined processes are generally producing more severe events which has significant implications respecting public safety and impacts to infrastructure.

KEYWORDS
River ice break-up; Snowmelt freshet; Climate warming; Compressed runoff

1. INTRODUCTION
River ice is an important component of both socio-economic and environmental features of cold regions. Of significant importance in remote, sparsely populated areas, frozen rivers are frequently used for transportation purposes for the construction of ice bridges and as transportation networks. In northern North America river ice is frequently relied upon to act as a platform for fishing and trapping purposes. Freeze-up and break-up processes often produce ice jams which may result in flooding with significant implications for public safety, and economic impacts associated with damage to property and infrastructure, road and rail networks and hydroelectric operations. Ice jams and subsequent backwater and ice jam release waves (javes) also affect aquatic ecosystems through impacts on biological and chemical processes. This paper attempts to provide a summary of climate warming impacts on the river ice regimes of the Yukon Territory with specific reference to the Porcupine River at Old Crow.

2. SETTING
Yukon Territory is situated in northwestern Canada, bounded by Alaska and the Northwest Territories to the west and east, respectively, and the 60th parallel of latitude and the Arctic Ocean to the south and north, respectively (Figure 1). The climate is characterized as subarctic in the south and arctic in the north, with some maritime influence from the Gulf of Alaska in the

Figure 1. Location Plan

southwest regions (Wahl *et al*., 1987). Annual mean daily air temperatures range from -1ºC in the south to -10ºC in the north. While, annual precipitation amounts are significant in the Coast and Saint Elias Mountains with amounts up to 2000 mm, precipitation throughout much of the Territory ranges from 300 to 600 mm, with annual amounts declining to approximately 150 mm on the Arctic coast. Much of the Yukon is underlain by permafrost subdivided into continuous, discontinuous and sporadic zones, representing approximately 30, 45 and 25 percent of Yukon, respectively (NRC, 1995). Streamflow response is controlled by the underlying permafrost or lack of it (Janowicz, 2004). The response is characterized by a rapid increase in discharge in the late spring in response to snowmelt at lower elevations followed by runoff from higher elevations with peak freshet flows generally occurring in early June. Hydrologic response is closely tied to the relative extent and location of the three permafrost zones. Peak flow volumes are directly proportional to the amount of underlying permafrost. Greater amounts of permafrost shorten the pathways to the stream channel as a result of limited infiltration rates (Janowicz, 2008). The controlling influence of the underlying permafrost on hydrologic response is extreme in Arctic regions. Peak flows exhibit very quick response times because of the shallow active layer. Secondary peak flows throughout Yukon occur during the summer months as a result of rainfall. Occasionally smaller systems will have the dominant peak resulting from rainfall. Annual minimum discharge occurs in March or April, coinciding in timing with minimum annual groundwater inputs. Annual minimum flows decrease moving northward due to lesser groundwater contributions to winter streamflow. Many smaller streams within the continuous permafrost zone are completely dominated by underlying permafrost and have no observed flow during the latter part of the winter.

3. CLIMATE AND RIVER ICE TRENDS

Annual, winter and summer temperatures have generally increased in all regions, with greater increases observed in central and northern regions. Annual precipitation trends are not consistent. Winter precipitation has generally increased in northern regions and decreased in southern regions. Summer precipitation has generally increased slightly throughout, with greater increases in southeast and central areas (Janowicz, 2010).

Climate warming is affecting the ice regimes of cold regions. In subarctic regions break-up is typically a spring event, with the timing generally a function of latitude. The length of the ice cover season has shortened, with later occurrence of freeze-up and earlier break-up events. Freeze-up observations were sporadically made in Yukon Territory since the 1890s primarily for river transportation reasons (Fountain & Vaughn, 1984). The data was initially collected by the transportation shipping companies, with the Atmospheric Environment Service taking over this role in later years. This practice was discontinued in the mid-1990s. Freeze-up of the Yukon River at Whitehorse has been delayed by approximately 30 days since 1902 (Janowicz, 2010).

Because of the river transportation history there is an excellent record of break-up dates for the Yukon River at Dawson. Break-up signaled the end of winter isolation for Dawson, which would be followed by the arrival of the first steamship from the south in about two weeks. A lottery to predict the exact minute of break-up has been held in Dawson since 1896 and continues today. Over the period of record, break-up at Dawson has ranged from 23 April to 29 May, with a mean date of 8 May (Janowicz, 2010). Jasek (1999) carried out an assessment of the data to 1998 and observed that the break-up date advanced 5 days per century. The last two decades have seen an unprecedented advancement of the break-up date. Prior to 1989, only two April break-ups have been observed, while after 1989, eight April break-ups have been observed, including the 2016 April 23 event which shattered the previous record by five days . A similar trend is noted for the Porcupine River at Old Crow, though the record begins in 1961. The break-up date ranges from 2 May to 30 May, with an overall mean of 16 May. The mean break-up date has advanced from 18 May during the first 20 years of record to 14 May in the last 20 years. (Janowicz, 2010)

Mid-winter break-ups have also occurred in recent years. Dawson experienced the warmest winters in the 115 year record during 2002–3 and 2016-17. Unusual periods of warm weather and rain during December resulting in an early winter break-up event and subsequent formation of an ice jam on the Klondike River. There was minor winter flooding in both years, but after the earlier event, the 3 km long jam which had formed subsequently refroze, creating “jumble” ice with thicknesses up to 3 m. During spring break-up at the end of April 2003, the lower Klondike valley experienced one of the most severe break-up floods on record, with a number of residences, businesses and the Klondike Highway affected.

Winter ice cover is becoming thinner as a result of increased winter temperatures, reducing downward freezing of the water column and less frazil ice generation. Increased winter discharge may also contribute to a thinner ice cover through the transport of frazil ice from the bottom of the ice cover. In some cases greater snow cover is contributing to the development of thinner ice covers through its insulating effect (Michel, 1971). There is some evidence to suggest that duration and severity of break-up events are also being affected. The ice cover period is becoming shorter; with the break-up period likewise, shorter (Janowicz and Hinzman, 2017). Break-up severity is a function of ice cover strength and integrity balanced by hydrometeorological conditions which control streamflow discharge and water level. At the lower range of severity, a thermal break-up occurs when the ice cover has deteriorated sufficiently by radiation from above and erosion from below to allow it to break up with little increase in discharge (Andrishak & Hicks, 2005). In contrast, a mechanical break-up occurs when streamflow discharge and water level increase rapidly, causing the ice cover to break its bond with channel banks while it is still strong and competent. Such events occur when a colder-than-normal period is followed by the rapid melt of a greater-than-normal snowpack and subsequent runoff. The most common break-up events occur are somewhere between the two extremes; however, these smaller events appear to be increasing in severity in some regions of Yukon Territory (Janowicz, 2010). Numerous Yukon communities have historically experienced ice jam flooding, with the most severe floods having occurred at Dawson City on the Yukon River and Old Crow on the Porcupine River. Dawson City and Old Crow have experienced six and four major ice jam floods in the last century, respectively. There appears to be a trend of increasing elevations from the early 1970s to the present, possibly exhibiting the influence of climate warming. Though there is both considerable range and scatter, the trend of increasing elevations is evident at both low and high elevations. A similar trend is not evident for the Porcupine River at Old Crow, possibly due to the paucity of data; however, the characteristics of the break-up event appear to be changing. Greater energy inputs, as represented by higher winter and spring air temperatures are producing an earlier onset of, and more rapid snowmelt events, resulting in a compressed runoff with higher peak flows in some regions (Janowicz and Hinzman, 2017). Figure 2 is a graphical illustration of a pair of streamflow hydrographs. The flatter hydrograph is typical of a nival streamflow regime prior to climate warming. The post climate warming hydrograph shows the peak to be greater and to occur sooner. With this scenario it is possible to have greater peak events even when the streamflow volume is less, due to the more rapid (compressed) runoff period.

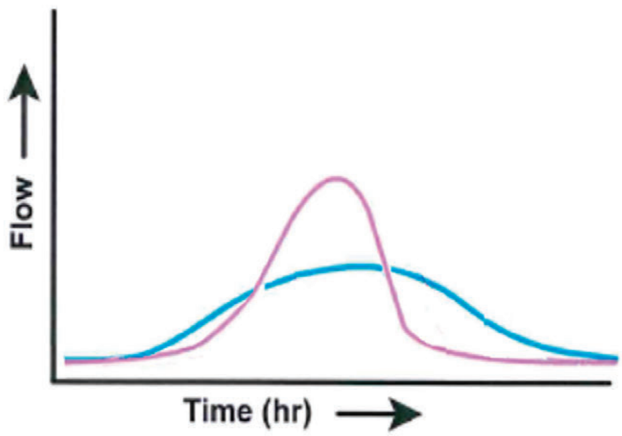

Figure 2. Typified hydrographs – pre-climate warming (blue); post- climate warming (pink).

There is some indication that these observed changes in runoff and streamflow dynamics are affecting the relationship between river ice break-up and the snowmelt freshet in northern Yukon. Typically these events are quite distinct. The timing of river ice break-up in Yukon occurs in early to mid-May, which is then followed by the snowmelt freshet peak two to four weeks later. Figures 3 and 4 illustrate typical spring and summer hydrographs for the Porcupine River at Old Crow. In 2009 the break-up peak occurred on May 7 followed by the larger freshet peak on May

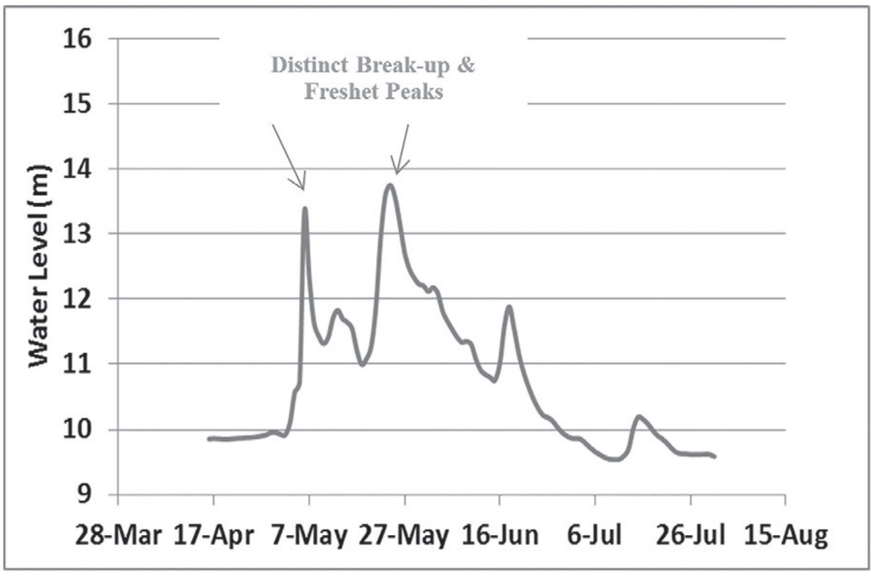

Figure 3. Porcupine River at Old Crow hydrograph – 2009.

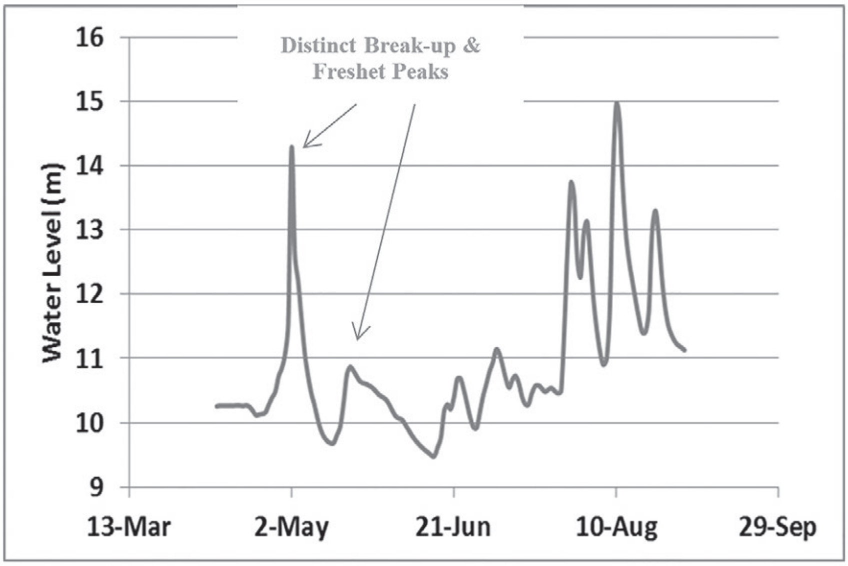

Figure 4. Porcupine River at Old Crow hydrograph – 2010.

24. In 2010 the break-up peak occurred on May 2 followed by the much smaller freshet peak on May 20. It is noted that the 2010 annual peak water level was generated by a rainfall event in August. Recent observations indicate that there is an increasing frequency of overlapping river ice break-up events and freshet peaks.

Figures 5 and 6 illustrate the summer hydrographs for 2011 and 2015. In 2011 and 2015 the combined break-up and freshet peaks occurred on May 23 and 15 respectively. A detailed illustration of the 2011 and 2015 events are presented in Figure 7 and 8 respectively. The 2011 local break-up at Old Crow occurred at 0915 on May 20, with the shifting of the ice in front of the community. Previous to this event, the Porcupine River water level had risen 1.8 m since the minimum winter level on April 13. Break-up was initiated by rapid snowmelt produced by significantly above normal temperatures which resulted in a rapid water level rise. After the onset of break-up the water level rose 3.8 m as a result of a bank to bank ice run over the next 5 hours which was backing up behind a major ice jam at the Bluefish River approximately

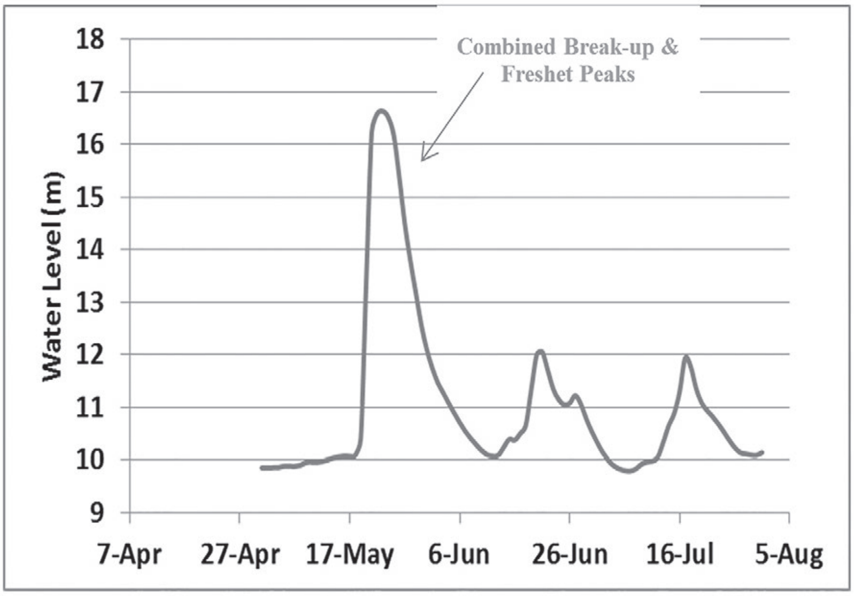

Figure 5. Porcupine River at Old Crow hydrograph – 2011.

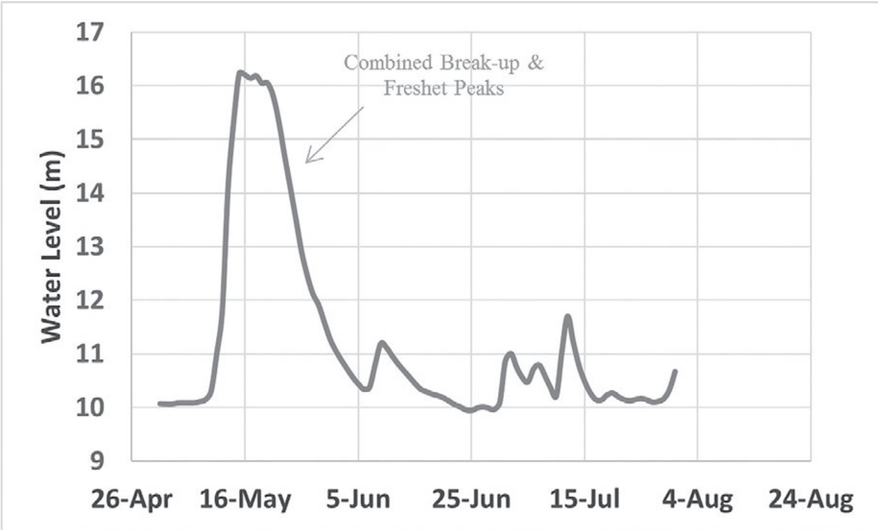

Figure 6. Porcupine River at Old Crow hydrograph – 2011.

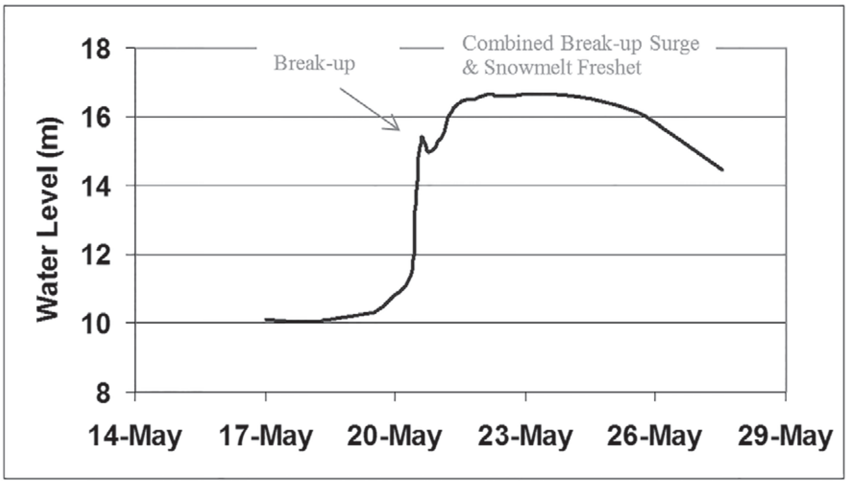

Figure 7. Porcupine River at Old Crow break-up hydrograph – 2011.

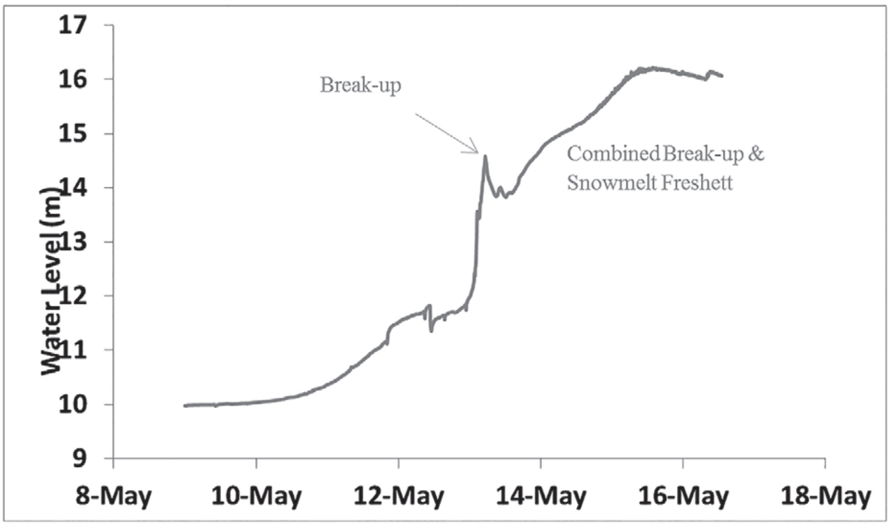

Figure 8. Porcupine River at Old Crow break-up hydrograph – 2015.

40 km downstream. A second major ice run beginning 0400 May 21 brought the water level up slightly. The water level dropped slightly with release of the Bluefish ice jam and the passage of the ice run. Continuous running ice with snowmelt contributions over the next 24 hours maintained the water level at a high level. A maximum ice related water level of 16.65 m occurred at 0500 on May 22 with snowmelt runoff contributions to this peak. Snowmelt runoff produced a slightly higher peak of 16.69 m at 1900 on May 23.

The 2015 local break-up at Old Crow occurred at 1215 on May 12. Prior to this event, the Porcupine River water level had risen 1.95 m since the minimum winter level on April 18. The local break-up was initiated by rapid snowmelt produced by above normal temperatures for the five days preceding break-up, resulting in a rapid water level drop. The 2015 break-up event was very rapid in progression. With the onset of break-up on May 12 the water level dropped 0.33 m as a result of the release of a downstream ice jam. This was followed by a 3.1 m increase to the peak break-up water level of 14.58 m at 0615 on May 13. This event was initiated by the simultaneous release and movement of the ice cover from the 147 km reach of river downstream of the Bell River. The passage of the ice run past Old Crow was followed by a 0.7 m drop in water level which was initiated by the release and movement of ice cover as far downstream as the Bluefish River, which is the traditional anchor point for severe ice jams. A heavy bank to bank ice run moved past the community for approximately 14 hours, slowly bringing the water level up to a new peak ice related level of approximately 14.7 m at 0045 on May 14. By May 15 ice from above the Bell River, as well as the Bell and Old Crow Rivers, moved past the community maintaining high water levels. Subsequent to break-up the weather became significantly warmer than normal, with air temperatures 5 to 15 degrees above the seasonal mean. These temperatures produced a very rapid melt of record high snowpack; which, on top of the break-up surge, resulted in a gradual and sustained water level increase to a combined break-up surge and snowmelt peak of 16.22 m at 1450 May 15. This event resulted in minor flooding of parts of Old Crow.

4. INTERRELATIONSHIP BETWEEN RIVER ICE BREAK-UP AND THE SNOWMELT FRESHET

Recent observations indicate that there is an increasing frequency of overlapping river ice break-up events and freshet peaks. Snowmelt dynamics can be described in various forms. The rate and quantity of snowmelt is determined by the amount of energy provided to the snowpack. The heat content of the snowpack over time is dependent on the summation of incoming solar radiation, sensible and latent heat transfer, the heat transfer through advection, and ground heat transfer (Pomeroy et. Al., 2003). Solar radiation accounts for the greatest portion of heat flux, followed by combined sensible and latent heat fluxes (NEH, 2004). These heat fluxes are often characterized by air temperature (Gray and Prowse, 1992). Energy inputs, as represented by air temperature, are increasing significantly in northern regions. While summer and winter temperatures are increasing in most Yukon regions, summer (April to October) temperatures at Old Crow are increasing at a greater rate than most Yukon communities (Janowicz, 2010). having increased 2.5 degrees in the last sixty years, while April, May and June temperatures have increased 4.5, 5 and 3 degrees respectively. Both river ice break-up and the snowmelt freshet are occurring earlier, often both in mid to late May with overlapping events.

Figures 9 to 14 present air temperature trends for consecutive five day periods from May 1 to May 30. May 1 to 5 and May 6 to10 indicate there is no visible change in air temperate during these periods. The five day periods between May 11 and 30 all exhibit progressively increasing. The average date of break-up has advanced from May 19 to May 13, while the average date of the freshet peak has advanced from May 28 to May 22.

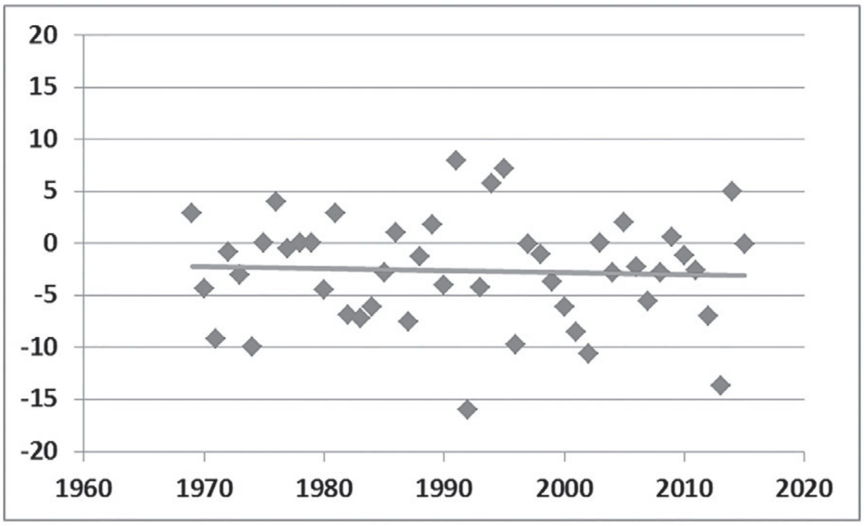

Figure 9. Old Crow Daily Air Temperature (May 1 – 5).

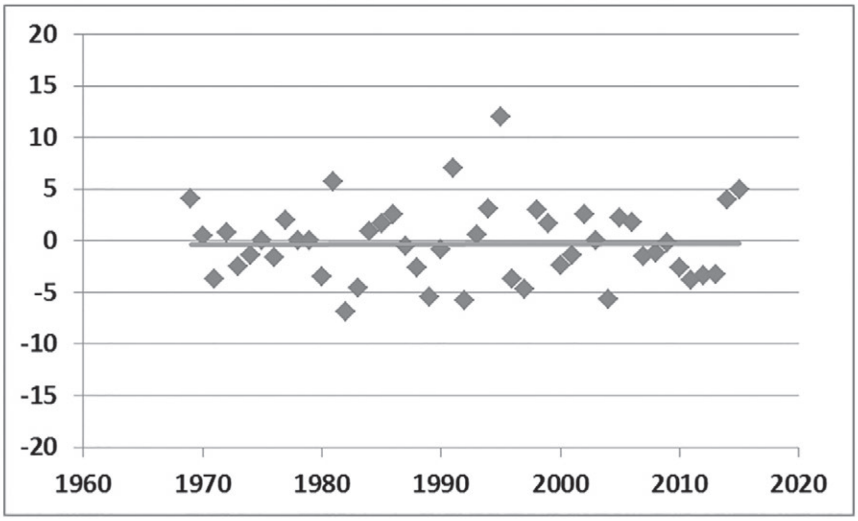

Figure 10. Old Crow Daily Air Temperature (May 6 – 10)

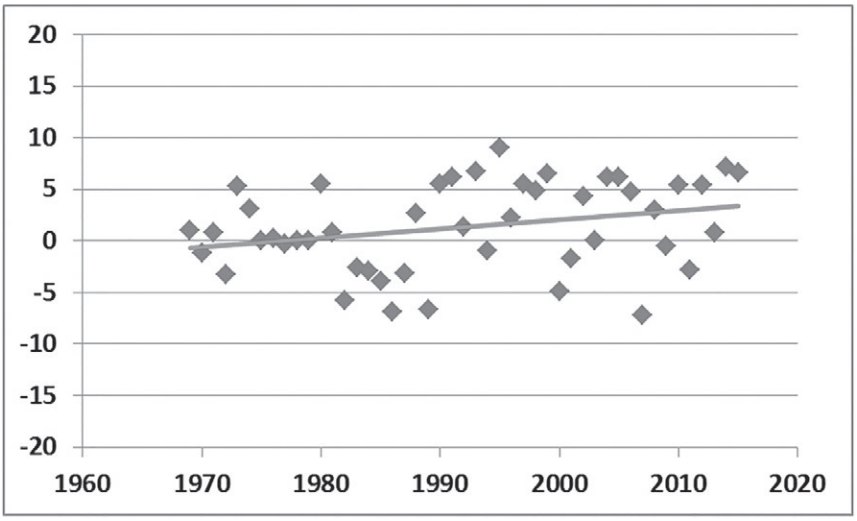

Figure 11. Old Crow Air Temperature (May 11 – 15)

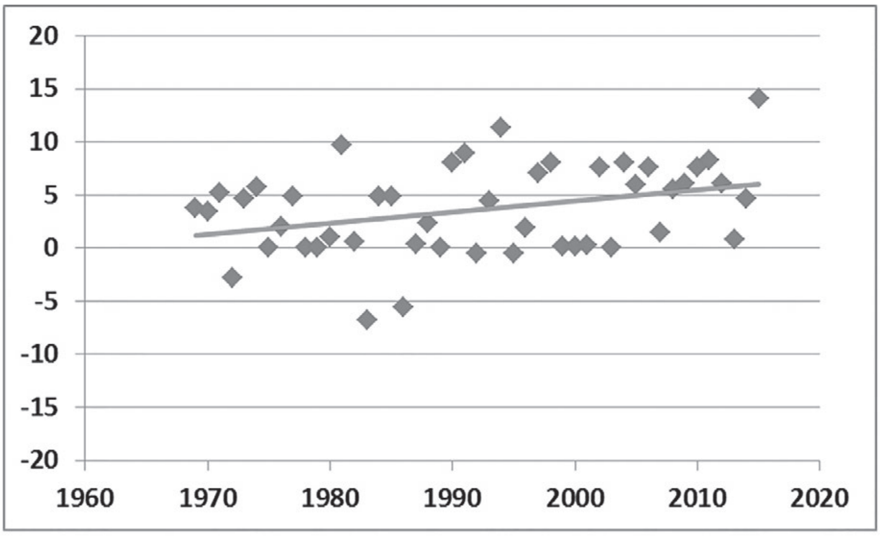

Figure 12. Old Crow Air Temperature (May 16 – 20)

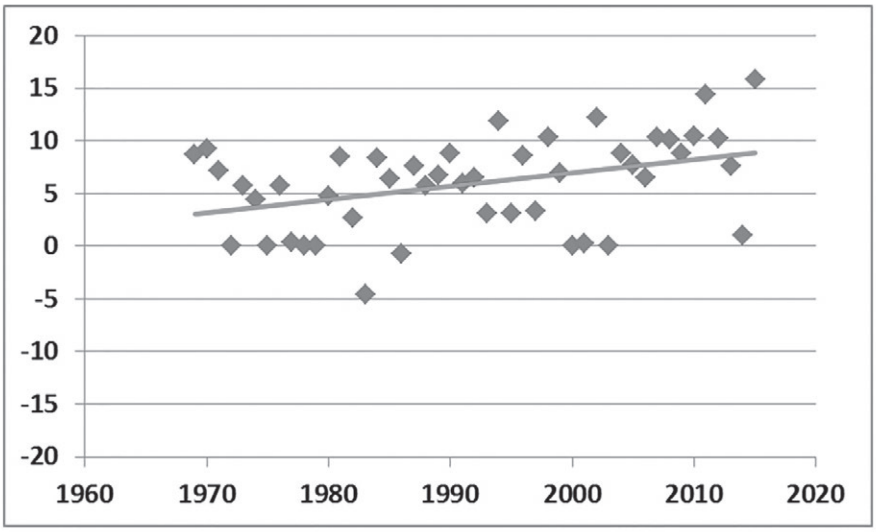

Figure 13. Old Crow Air Temperature (May 21 – 25)

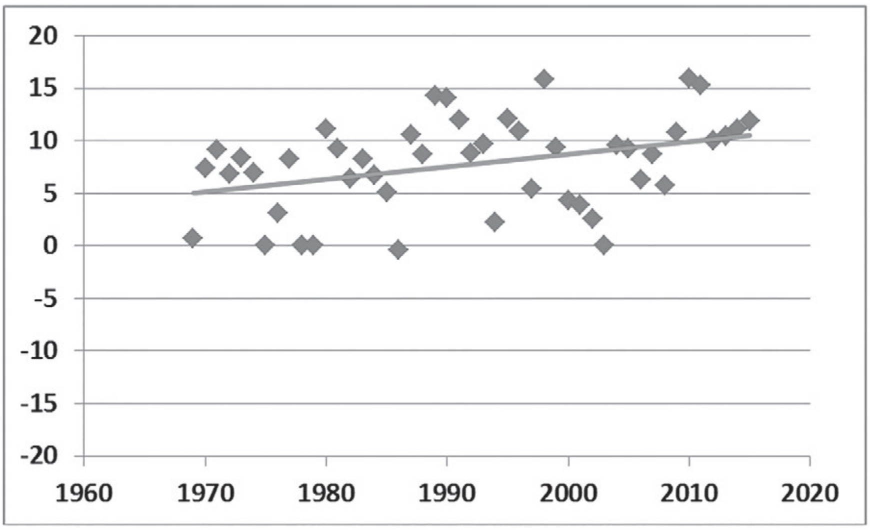

Figure 14. Old Crow Air Temperature (May 26 – 30)

5. CONCLUSIONS

Yukon air temperatures have increased significantly in the last century. Precipitation is not as consistent with greater precipitation in some regions and less in others. The length of the ice cover period is becoming shorter with later freeze-up and earlier break-up dates. Mid-winter break-up events and associated flooding have been observed on two occasions. Break-up water level trends on the Yukon River suggest that break-up severity is increasing. There is also some indication that changes in runoff and streamflow dynamics are affecting the relationship between river ice break-up and the snowmelt freshet. Where these events have been quite distinct, recent observations indicate that there is an increasing frequency of overlapping river ice break-up events and freshet peaks. The observed changes have significant implications associated with public safety and economic impacts to property and infrastructure, transportation networks and hydroelectric operation. In addition there are potentially serious environmental impacts on fish and wildlife habitats.

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A REMOTE ASSESSMENT OF ICE-RICH PERMAFROST LOSS ACROSS A ~500M ELEVATION GRADIENT IN THE MACKENZIE AND SELWYN MOUNTAINS, CANADA

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ABSTRACT

In the eastern Mackenzie and Selwyn Mountains, NT, Canada, five ice-rich permafrost ice-rich permafrost features have decreased in areal extent as permafrost temperatures and supra-permafrost thickness have increased since the mid-1990s. Among these features, thaw rates have been more rapid at higher elevations from the 1940s-1980s, while lower elevation features have more recently been decreasing in extent as peat block calving and subsidence accelerate. A comprehensive regional survey of ice-rich permafrost is necessary to assess whether these five fine-scale observations are representative of broader spatial and temporal changes in the area.
This paper explores the potential of remote sensing data to discern trends in permafrost extent across a ~500m elevation gradient within a 1757 km^2 study area. Our sample of 350 ice-rich permafrost features identified on 2010/2013 satellite images exhibited a decline in log area (log_A) with increasing elevation, but no discernable association with shape index (SI) values. Aerial photographs from 1944 and 1974 were also used to assess temporal trends in permafrost extent for a subset of 33 features, including three of those previously reported on. There was a significant decline in log_A with time, and a complementing pattern in SI was observable among features >1400 m.a.s.l.. In the fall of 2017, field surveys will be conducted to confirm the elevation, dimensions, and site characteristics of each sampled feature. Further research concerning the driving variables of permafrost distribution and thaw rate variance should focus on parameters reflecting the local hydrology, vegetation, snow pack, and soil characteristics of each site.

KEYWORDS

Canada, alpine, permafrost, remote sensing, climate change

1. INTRODUCTION

Throughout the circumpolar, permafrost is thawing and receding poleward in response to a warming climate ((Osterkamp and Romanovsky, 1999; Beilman and Robinson, 2003; Saito *et al.*, 2007). This change is expected to accelerate in future (Saito *et al.*, 2007). Isolated permafrost (<10% of exposed ground surface) (Van Everdingen, 2005)held in Trondheim (Norway is the most vulnerable to this climate induced thaw as its temperature rarely deviates from just below the freezing point depression (Sannel and Kuhry, 2011) and a minimal shift in the annual energy balance of the soil can thaw what small volume of permafrost remains.
Montane systems, with their steep elevation gradients, have been experiencing more extreme changes in climate than lowland areas over the past 150 years (Mountain Research Initiative EDW Working Group, 2015). Assuming climate is the primary factor driving permafrost

distribution, a change in elevation is expected to result in a more pronounced change in permafrost's distribution than a similar magnitude shift in latitude. For example, changes in ground temperature across 1 km elevation have been reported as high as 15°C, roughly comparable to a 1000 km latitudinal shift (Riseborough *et al.*, 2008). This creates a vital research opportunity, as the distribution of climate sensitive landforms and ecology will be more pronounced in montane settings, yet these areas remain relatively data poor (Hilbich *et al.*, 2008; Smith *et al.*, 2010). But montane permafrost models often rely on empirical relationships to air temperature (Riseborough *et al.*, 2008; Bonnaventure and Lewkowicz, 2013), as they are difficult to parameterize and execute due to landscape heterogeneity in soil characteristics, climatic conditions, plant community, slope and exposure. As a result, these models often have increased error and decreased transferability for their results (Riseborough *et al.*, 2008).
Previous studies in the Mackenzie and Selwyn Mountains have described the conditions of ice-rich permafrost features between ~1200-1600 m.a.s.l., reporting general trends of linearly increasing permafrost temperature parallel to non-linearly decreasing feature extent. Detailed accounts of changing supra-permafrost thickness, aerial extent, and shape complexity continue to be made annually, while local air, ground, and soil temperature records are continuous from 1990s onward (Kershaw, 2003; Mamet *et al.*, 2017).
In areas where the peat layer is thick, substrate fine textured, saturated, and seasonal descent of the freezing front slow, permafrost often forms thick ice lenses, heaving the ground upward as the volume of water increases with phase change (Seppälä, 2011). This permafrost is more resistant to a warming climate as a large amount of latent heat is required to thaw the excess ice. When such ice-rich permafrost thaws, it often results in thermokarst pond depressions, which in turn accelerate thaw of the surrounding area via increased lateral heat conduction from the newly formed water bodies (Kurylyk *et al.*, 2016). As thermokarst areas expand, they eventually merge to create hydrological connections via fen networks where before elevated plateau's and isolated bogs prevailed (Quinton and Baltzer, 2013). In turn, the plant community previously perched above the water table must adapt, as many lichen, moss, shrub and tree species become replaced by aquatics in the process of paludification (Jorgenson *et al.*, 2001).
In this study, we use high resolution panchromatic satellite imagery from the Mackenzie Mountains to discern spatial trends in the variability of ice-rich permafrost feature extent and shape complexity across a ~500m range of elevation. We then apply a series of historical aerial photographs from the area to compare temporal changes in permafrost extent. Temporally, ice-rich permafrost features were expected to decline in size across the study area, while spatially, the size and complexity of features are expected to vary such that small, compact features are found at higher elevation while low elevation features are relatively large with complex perimeters. This study is exploratory and it's statistical techniques descriptive in nature. Future research is discussed, specifically what field assessments are necessary to parameterize the local ecological, hydrological, and soil characteristics moderating permafrost freeze/thaw processes.

2. METHODS

1.1. Study Area

The area of the Mackenzie Mountains considered in this study was most recently glaciated during the Gayna River Glaciation 22000 years B.P. (Kershaw and Kershaw, 2016). The ice-rich permafrost features we focus on established post glaciation, less than ~1182-1165 years ago, as evidenced by White River Tephra below aquatic and emergent plant peat layers deposited before ice lensing heaved the ground surface above the water table (Kershaw and Gill, 1979). These features are found in the valley bottoms and across broad perched plateau wetlands bog-fen-permafrost complexes (Kershaw and Gill, 1979) (Figure 1). Today these features have ice-rich soils (typically >20% ice content) with peat overburden (typically <5m) (Skaret, 1995), anomalous for alpine permafrost which is typically found in well drained and organic poor soils or exposed bedrocks (Zhang *et al.*, 1999)together with ancillary data sets of the global land cover characteristics data base and the Global Land One\u2010kilometer Base Elevation data base, are used to investigate the distribution of permafrost and ground ice in the Northern Hemisphere. Our study indicates that permafrost underlies approximately 22.79 \u00d7 106 km2 or 23.9% of the exposed land area of the Northern Hemisphere. Permafrost extends from 26\u00b0N in the Himalayas to 84\u00b0N in northern Greenland. Approximately 70% of the permafrost is distributed between 45\u00b0N and 67\u00b0N. Generally, permafrost with high ice content (>20% by volume.

While some assert such features follow cycles of formation and decay (Seppälä, 2011), this is not obvious in our study area where long-term monitoring efforts have observed permafrost thaw of ~1% per year over the past half century, with some features disappearing completely and no new features forming (Kershaw, 2003). This is consistent with observed permafrost loss throughout the circumpolar and is most likely the result of a larger climatic shift towards warmer conditions. Tree-ring records as proxy for temperature trends suggest a warmer climate occurred in the region during the late 18th century, and that it is again warming since the mid-20th century (Mamet and Kershaw, 2012).

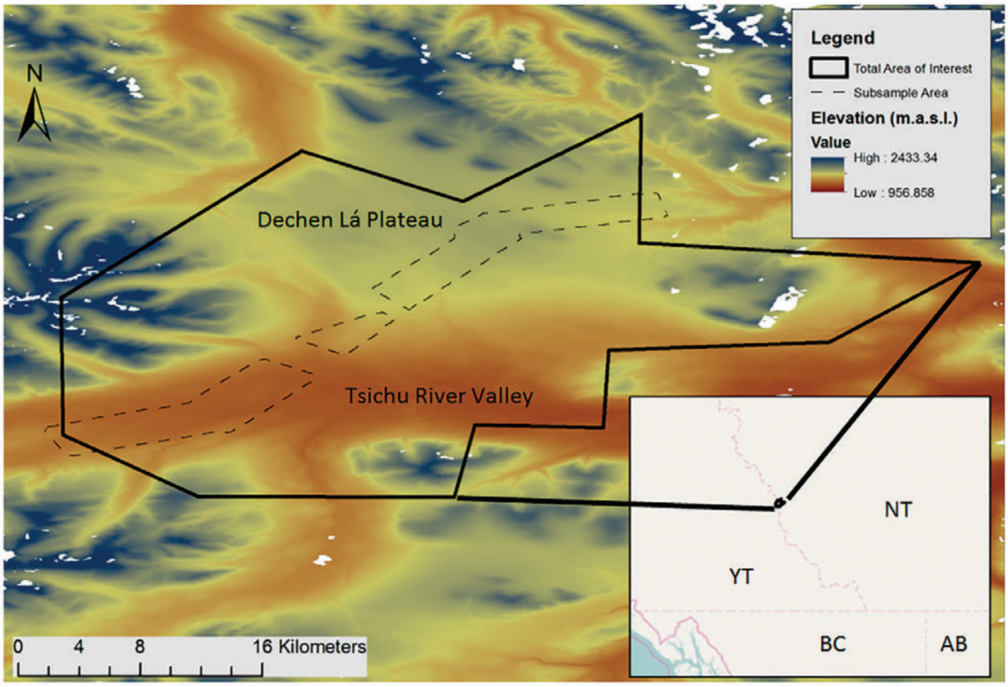

Figure 1. Map of study area. Total area of interest was limited to the maximum extent of overlapping air photos and satellite imagery, while the subsample area was limited to air photos of sufficient quality for feature delineation. The Digital Elevation Model (DEM was created by the Polar Geospatial Center from DigitalGlobe, Inc. imagery.

1.2. Data Processing

1.2.1. Image Acquisition

1944, 1948, 1949, and 1974 air photos of the study area were acquired from the National Air Photo Library (NAPL). 295 Air photos were digitized with an Epson 12000 XL scanner set at 1200 dpi. 56 air photos were georeferenced to Worldview-2 panchromatic satellite imagery from 2010/2013 (DigitalGlobe, Inc., Westminster, CO, USA). The overlapping extent of images from 2000s, 1970s, and 1940s was assigned as our total study area to survey, covering 1757 km^2. A series of 1944 and 1974 images taken at lower elevation (6000-2750 m.a.s.l.) were also selected for a 202 km^2 subsample survey as they had image quality consistent with Worldview's spatial resolution of 0.5m.

All spatial inputs were processed with ArcMap 10.3.1 GIS (ESRI, Inc., Redlands, CA, USA) projected in the datum WGS_1984. Resampling with the nearest neighbor method was done to generate digital pixel values from the scanned air photos, allowing us to georeference them to the 2010/2013 Worldview base map. Spatial distortion and scaling of each photo was corrected with 5 ground control points and an affine 1st order polynomial transformation. The Root Mean Square Errors (RMSE) resulting from this rectification were calculated to assess the accuracy of georeferencing.

1.2.2. Feature Identification

Ice-rich permafrost features were identified by their morphological traits of increased relief and distinct reflectance due to differing surface vegetation. While categorical distinctions have been made among the ice-rich permafrost features we are considering, namely palsas, peat plateaus, and pingos, this distinction cannot be made reliably with remote sensing means.

Our sample included 350 Polygons defining the perimeter of ice-rich features visible on 2010/2013 satellite images. Each feature was assigned a mean elevation as calculated with the Zonal Statistic tool in ArcGIS, linked to a Digital Elevation Model (DEM) generated by the Polar Geospatial Center from DigitalGlobe, Inc. imagery. A subsample of 33 features from across the elevation gradient had additional polygons delineating feature extent in 1944 and 1974 air photos. This subsample included three features with ongoing microclimate and active layer monitoring across the elevation gradient (Mamet *et al.*, 2017).

1.3. Data Analysis

The feature attributes considered in our analysis were log aerial extent (log_A) and shape index (SI). The skew and kurtosis for area lead us to apply a log transformation to better establish normality. A Shapiro-Wilk test was run to confirm the effectiveness of the transformation. Spatial autocorrelation of feature location and areal extent values were assessed with Moran's I coefficients using the global Moran's I tool in ArcMap. SI relates the complexity of the perimeter relative to the area of the feature with the following calculation:

$$SI = 0.25 \times \frac{P}{\sqrt{A}},$$

where P is perimeter (m), and A is area (m^2). A SI value of 1 relates a maximally compact square and values >1 reflect increasing complexity (McGarigal *et al.*, 2012). Spatial and temporal trends in log_A, and SI were assessed across the elevation gradient and from 1944-2013 using simple linear regression. Temporal trends were limited to the subsample of 33 features while spatial trends considered all 350 features identified in 2010/2013 satellite imagery.

2. RESULTS AND DISCUSSION

2.1. Spatial Trends

A histogram of area values exhibited negative exponential form (y = -6.196ln(x) + 26.071, $R^2 = 0.7211$). This was corrected with log transformation to assure normality and homogeneity of variance, as confirmed with a Shapiro-Wilk test (α 0.05) (data not shown). The elevation values were skewed negatively (-0.277), relating a greater concentration of features on the higher Dechen Lá plateau than the lower Tsichu valley or intervening slope (Figure 1). For $\log_A$ and elevation values, specific features were identified as potential outliers (>3x Interquartile range). After reassessing the location and extent of these features, each was found to be accurate and remained included in the sample data.

The Moran's index reported for the sample of 350 features was 0.090 with a z-score of 3.7 (p-value <0.01). Given this value, the null hypothesis was rejected and features considered spatially autocorrelated. This could result from internal or external mechanisms acting on the population. Internally, as larger coalesced fields thaw, they form groups of smaller detached features which are highly spatially autocorrelated. Externally, the conditions for formation and persistence of ice-rich permafrost could be limited to specific areas on the landscape where the slope, snow pack thickness, and substrate consisting of thick organic overburden with highly saturated fine grained clay below are conducive to ice lens formation (Seppälä, 2011).

The combination of Worldview images (0.5m resolution) and Polar Geospatial Centre DEM (5.0 m resolution) allowed for an assessment of spatial trends in $\log_A$ and SI across a ~500m elevation gradient. Log_A reported a significant negative association with elevation ($F(1,350)=11.695$, $p<0.01$). This is consistent with the initial observations of Mamet et al. (2017), though SI did not report a significant association with elevation, as expected ($F(1,350)=0.109$, $p=0.745$) (Figure 2,3).

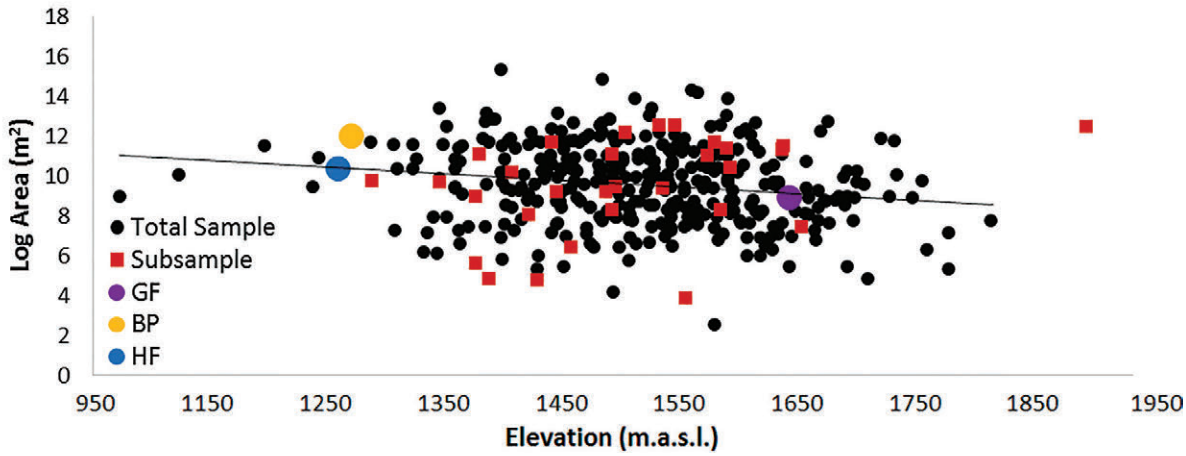

Figure 2. Elevation vs $\log_A$ in 2010/2013 for the total sample population (n=350) with a linear regression trend line of $y = -0.0839x + 24.492$ ($R^2 = 0.0324$). Subsample features and features previously assessed identified.

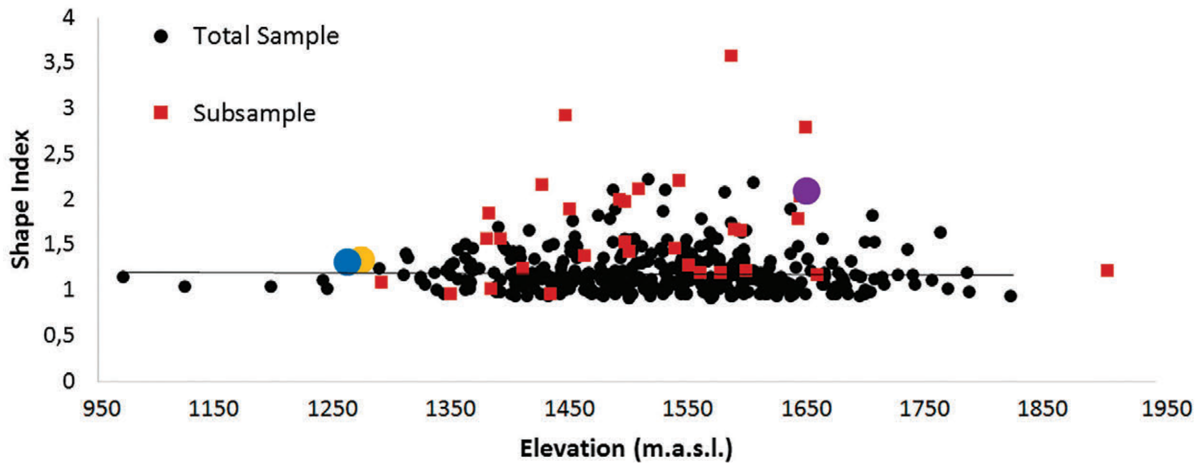

Figure 3. Elevation Vs Shape Index (SI) in 2010/2013 for the total sample population (n=350) with a linear regression trend line of $y = -0.0012x + 1.3478$ ($R^2 = 0.0005$). Subsample features and features previously assessed identified.

2.2. Temporal trends

Given the limited image quality of many of the historical air photos, we were unable to directly assess temporal trends in ice-rich feature geometry across the entire study area. Instead, we chose a subsample of features from within a reduced survey area where air photos were taken at lower elevation, providing better image quality sufficient for feature delineation (Figure 1). The features selected within this area were spaced widely in order to correct for the spatial autocorrelation reported by the larger sample. The subsample also deliberately included three features with previous monitoring of microclimate and active layer thickness for comparison with this larger sample. This technique is problematic due to the non-random nature of the subsample, but we considered it reasonable for the purposes of data exploration in this paper. The resulting Moran's index for the subsample was -0.045 with a z-score of -0.17 (p-value 0.8609) allowing us to assume spatial independence.

Previous research suggested a temporal shift towards less permafrost in our study area, with thaw more rapid at higher elevations in the 1940s-1980s while lower elevation features more rapidly decreased in extent from the 1980s onwards (Mamet et al., 2017). This is consistent with our subsample data as there was a significant decline in $\log_A$ through time ($F(1,108)=22.0433$, $p<0.01$), following the regression $y = -0.0161x + 43.136$ ($R^2=0.0916$). When separating the subsample into high and low elevation features, regressions remained significant ($F(1,108)=16.7175$, $p<0.01$ and $F(1,108)=2.8225$, $p=0.03$ respectively) with a greater rate of aerial loss for high elevation features (Figure 4). We were unable to discern different rates of thaw between the two time periods of 1944-74 and 1974-2010 though, likely due to the limited sample size of the data.

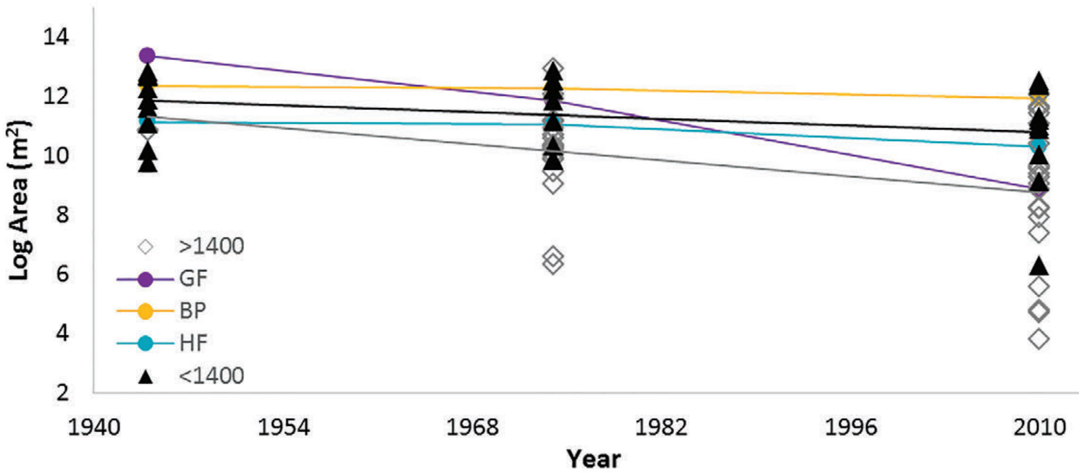

Figure 4: Log area Vs time for the subsample of ice-rich permafrost features (n=33). Goose Flats (GF) at 1625 m.a.s.l., Beaver Pond (BP) at 1275 m.a.s.l., Hare Foot (HF) at 1260 m.a.s.l. identified, as well as features >1400 m.a.s.l. ($y = -0.0357x – 80.581$ ($R^2 = 0.2237$)) and features <1400 m.a.s.l. ($y = -0.0161x + 43.136$ ($R^2 = 0.0916$))

Previous research also identified a temporal shift in shape complexity of ice-rich permafrost features, such that low elevation features were increasing in complexity from the 1980s onwards while high elevation features have transitioned through this period before the 1980s and are more recently reducing in complexity (Mamet et al., 2017). In this study, there was no significant linear trend to the subsample's change in SI ($F(1,108)=3.3308$, $p=0.07$), but when low and high elevation features were considered separately, features >1400 m.a.s.l. reported a significant increase in SI through time ($F(1,59)=16.7175$, $p=0.01$) (Figure 5). This discrepancy is interesting, but may in part be due to limited sample size as only 10 features were identified

<1400 m.a.s.l. compared to the 20 at >1400 m.a.s.l.. It is also important to consider that the observed changes may only be a small portion of the trajectory of permafrost thaw for these features. The maximum extent of these ice-rich permafrost features most likely was much reduced before our period of assessment. In Alaska for example, tree ring and carbon isotope analysis has been generated an estimate that 83% of permafrost's maximum extent had been lost before 1949 (Jorgenson et al., 2001). This would affect the SI values we could expect to observe as the features are likely small and simple relative to their previous maximum extent.

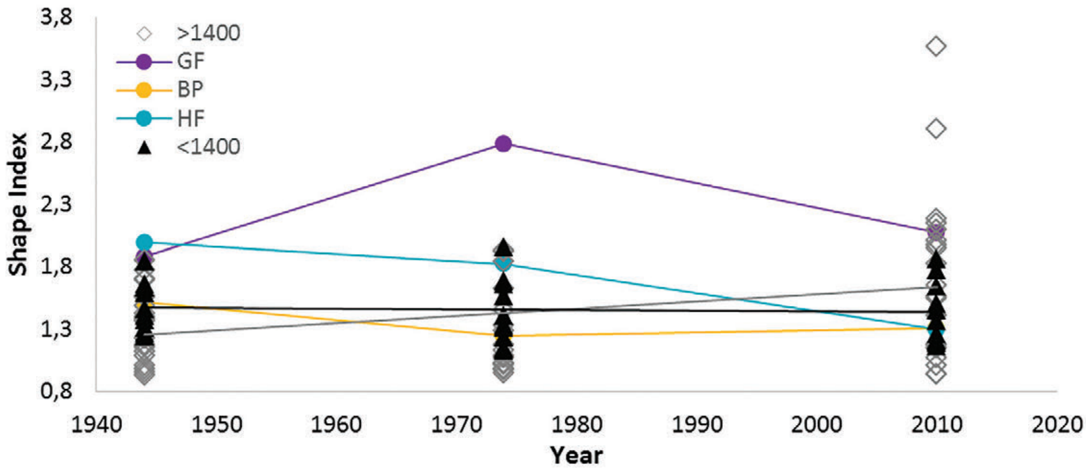

Figure 5. Shape index Vs time for the subsample of ice-rich permafrost features (n=33). Goose Flats (GF) at 1625 m.a.s.l., Beaver Pond (BP) at 1275 m.a.s.l., Hare Foot (HF) at 1260 m.a.s.l. identified, as well as features >1400 m.a.s.l. ($y = 0.0058x - 10.19$ ($R^2 = 0.1013$)) and features <1400 m.a.s.l. ($y = -0.0005x + 2.5171$ ($R^2 = 0.004$))

2.3. Future Research

As this paper is exploratory and descriptive in nature, there is a series of analytical options for expanding the study design. Statistically, spatial autocorrelation was an issue with our total sample. While this was corrected with a selective subsample maximizing coverage and minimizing autocorrelation, data transformations like the generalized least squares and conditional autoregressive models may be more effective at this task as they do not compromise randomness of sample selection (F. Dormann *et al.*, 2007)i.e. locations close to each other exhibit more similar values than those further apart. If this pattern remains present in the residuals of a statistical model based on such data, one of the key assumptions of standard statistical analyses, that residuals are independent and identically distributed (i.i.d. Also, the process of feature identification could be expanded to include proximity to open water, slope, estimated elevation temperature lapse rates, and other attributes that can be defined by remote means.

The use of trained categorization techniques based on pixel brightness values could also help reduce error in feature delineation, depending on how effectively the program could be trained and automated. Field verification could then follow to confirm the presence/absence of permafrost and open water where shadowing and aquatic vegetation obscure feature edges. Such a ground-truthing campaign would require a more recent set of satellite images as the features in question may have changed in the four years since worldview last produced images of the area.

Field verification would also be an opportunity to distinguish palsa, peat plateau, and pingo feature types, as well as parameterize components of mass-energy coupled equations applied in models of permafrost thaw. This parameterization would include soil characteristics such as thermal conductivity, porosity, ice and moisture content, as well as vegetation characteristics such as surface roughness and albedo.

3. CONCLUSIONS

In the valley bottoms and shallow sloped plateaus of the Mackenzie Mountains, ice-rich permafrost features form in wetlands under thick insulating layers of peat, thin snow pack, and where subsurface water is readily available for ice lens formation. This project provides an exploratory assessment of the potential remote sensing data has for quantifying changes in the distribution of ice-rich permafrost features in montane environments. The acquisition, processing, and analysis of historical air photo and satellite images from 1944-2013 is described and the limits of such an approach discussed - namely issues of historical image quality, spatial autocorrelation, and sample size. We describe spatial and temporal patterns in the distribution of ice-rich permafrost across a ~500m elevation gradient. Spatially, Log_A decreases with elevation. Temporally, Log_A has been in decline since 1944. The general trend of declining extent is consistent with previous observations across the elevation gradient. Temporally, SI was unresponsive to changes in elevation and time, unless features >1400 m.a.s.l. were considered alone, whereby an increase in SI through time was observed. A larger sample of features across the elevation gradient, particularly at lower elevations, would likely increase confidence in the observed temporal trends. The discrepancy between changes in low and high elevation Log_A and SI raises the question of what factor(s) are responsible for the distinct trends in thaw rate and shape complexity. Future research should focus on field verification and parameterization of the factors considered in soil freeze/thaw models, providing the data necessary for inferential statistical tests based on explanatory variables of microclimate, soil, hydrology, and vegetation.

4. ACKNOWLEDGEMENTS

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CLIMATIC CHANGES IN THE TERRITORY OF YAKUTIA AND THEIR INFLUENCE ON FOREST FIRES AND MAXIMUM DISCHARGES AND WATER LEVELS OF SNOWMELT FLOODS

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ABSTRACT
The current climate warming is shown in the dynamics of such basic climatic indicators as the number and area of fires, and the maximum discharges and water levels of snowmelt floods in the Republic of Sakha (Yakutia). It has been established that the variances of fires in the center and in the southeast have sharply increased. At the same time, the maximum discharges and water levels in rivers are still stationary. The obtained empirical relationships between climatic indicators and meteorological factors make it possible to obtain scenario estimates of future changes in these indicators.

KEYWORDS
Climate change; Climate change indicators; climate models; spring flooding; forest fires

1. INTRODUCTION
Climate change is already an almost indisputable fact, which manifests itself in the changes of the air temperatures at the different levels: global, annual, monthly and their extremes (extreme values) (Roshydromet 2014, IPCC 2013a, Kalinin 2015, IPCC 2014b). Disputes can be only about the causes of warming (is it only anthropogenic or combined with natural); and the dynamics of temperature growth rate (is it monotonous growth, a step change in the transition from one stationary conditions to another, or we are already close some temperature growth limits after which can happened a temperature drop (Abdysamatov 2013, Dymnikov et al. 2006, Klimenko 2009, Lobanov et al. 2010)). Precipitation change patterns are more ambiguous. They can be determined by characteristics of moisture transfer, precipitation patterns and local geographical conditions. Another level of factors is the climate change indicators, i.e. those natural parameters which depend on the climate and whose variation has a big importance for economy and human well-being. Exactly these climate change indicators were considered in the IPCC reports which investigating climate change adaptation, vulnerability assessment, and climate change impacts mitigation (IPCC 2014c, IPCC 2014d, Kattsov et al. 2011, IPCC 1994).
In this article, will be investigated the dynamics of the three main climate change indicators for the Republic of Sakha (Yakutia). It is:
- forest fires, this indicator tracks the frequency, extent, and severity of forest fires (was applied to all Yakutian forestries);
- soil temperature at the different depths, which determines permafrost properties;
- maximum discharges and water levels of snowmelt flooding, which characterize the intensity of spring floods.

Each from the considered climate change indicators is an important factor for the economy of the republic. For example, forest fires are not only a dangerous natural phenomenon, but also a factor which affects forestry and agriculture sectors, republic's population health and well-being. Soil temperature and its dynamics is an important factor for building and construction, and transport sectors. The maximum discharges and water levels of snowmelt flooding are defined as hazardous flood conditions, as well as indicators of hydropower energy production, agriculture and communal service.

2. METHODS

We made the evaluation of manifestations of climate warming on the example of the set of climate change indicators including forest fires and maximum discharges and water levels of snowmelt flooding. This assessment was made on the basis of statistical modeling and definition of the type of time series model.

3. RESULTS AND DISCUSSIONS

3.1. Forest Fires

The features of the present and future changes of meteorological variables as air temperature and atmospheric precipitation on the territory of Yakutia have been investigated and presented in (Kirillina and Lobanov 2015, Kirillina and Lobanov 2015, Kirillina et al.2015), from which follows that the air temperature has increased and became more non-stationary in recent years. Regarding the atmospheric precipitation some stable patterns were not found, although there is also existing a tendency for their growth. At the same time, the response of other natural indicators to climate changes is not less interesting.

Primarily, were analysed the forest fires with natural origin (their number and burnt area) occurred on the territory of the republic for the historical period from 1955 to 2014 based on the official Government Statistics from the Department of Forestry of the Sakha Republic. Due to the fact that forestries have different area, the number of forest fires and burnt area parameters have been converted per unit area - 1 km^2, with the aim to compare data across the whole Yakutia and identify areas with the highest intensity of forest fires. Then were calculated the correlation coefficients between the number of forest fires and the burnt area which indicate that these two indices are not related to each other, the highest correlation coefficients were around 0.48-0.55 and they can be applied to only six from nineteen forestries. The highest relative intensity of forest fires was found in the Yakutsk, Vilyuysky and Amginsky forestries and it was 0.61%, 0.42% and 0.36%. In the northern forestries (Zhigansky, Indigirsky, Tomponsky), burnt area calculated in relative values was the smallest with values 0.02%, 0.02%, 0.06%, which can be related both to their big area and relatively lower temperature during summer.

The analysis of the long-term dynamics of the number of forest fires allows to report about the absence of trends in the growth of number of forest fires on the territory of Yakutia, and it is possible to conclude only about their steady-state oscillation mode, which was typical for following list of forestries: Lensky, Verhnevilyuisky, Vilyuisky, Myrninsky, Nyurbinsky, Gorniy, Khangalassky, or growth of the dispersion of the number of forest fires for Yakutsk and Amga Forestries. For other forestries, was found even a decrease in the number of forest fires, which was occured in Aldan (since 1998), Neryungri (since 2001), Ust-Maysky (since 1995), Tomponsky (since 1998), Olekminsky (since 1997), Zhiganskiy (since 1985), and Indigirsky (since 1997) Forestries.

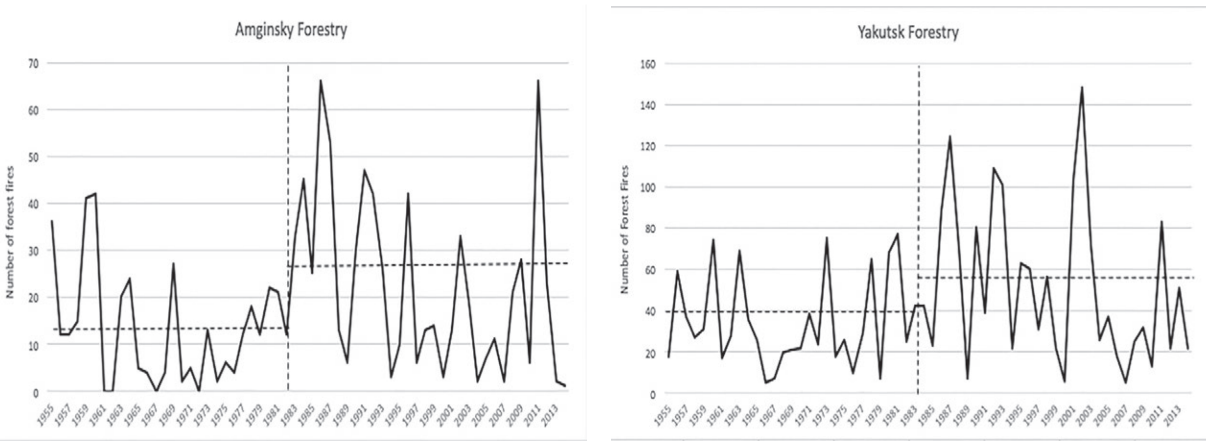

Figure 1. Dynamics of the number of forest fires in the Yakutsk and Amga Forestries.

The analysis of the long-term observational series of the burnt area also showed that their ranks do not contain any directed trends of growth or decrease, but there are periods with different dispersion values, as well as a period with almost a complete lack of burnt area, which differs depending from forestry with the earliest beginning at 1965 and the end in the middle or the late 1980s, and sometimes in the 1990s (Fig.2).

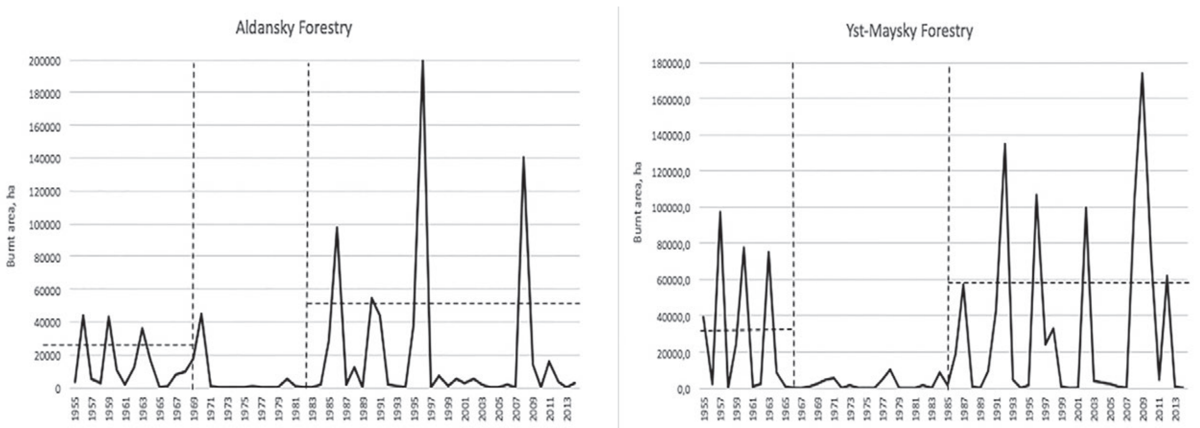

Figure 2. Dynamics of the burnt area in Aldansky and Yst-Maysky Forestries.

As follows from Fig. 2, the dispersion of the burnt area changes significantly, which in turn leads to a change in their mean values. At the same time, was not found any trend in the changes of burnt area in the northern forestries (Indigirskoye, Tomponskoe, Zhiganskoye). The greatest growth of the dispersion of the burnt area was observed in the forestries located in the Central and South-Eastern Yakutia. Tab.1 presents the calculated statistics of Fisher's criterion (F) - the ratio of dispersions, the ratio of the mean values (f_{m1}/ f_{m2}) and the ratio of the largest values (f_{max1}/ f_{max2}) of the burnt area in two homogeneous periods. Moreover, were given the calculated correlation coefficients R between the number of forest fires and burnt area, the low values of which confirm that the number of fires is unrelated to burnt area from forest fires. In the time interval between these two quasi-stationary periods, the burnt area was very small and hence it can be considered as a period with the absence of forest fires.

The results of Tab.1 shows that for eight from ten forestries represented in this Table besides Amginsky and Ust-Maysky Forestries, the calculated values of the Fisher statistics exceed critical even with a significance level of 1%. Although the average values for two homogeneous periods differ significantly (2.5 times on average), but their evaluation by the Student's criterion

Table 1.

Characteristics of burnt area for two quasi-homogeneous periods.

Forestry	*Time period 1*	*Time period 2*	*F*	F_{m1}/f_{m2}	f_{max1}/f_{max2}	*R*
Amginsky	1955-67	1987-2014	1,85	0,89	2,1	0,51
Verkhnevilyuisky	1955-75	1991-2014	52,5	6,84	7,16	0,30
Vilyuisky	1955-74	2001-2014	14,08	4,28	3,63	0,10
Gorny	1955-71	1992-2014	8,62	1,91	3,89	0,35
Neryungrinsky	1975-97	2007-2014	5,55	1,99	1,53	0,22
Yst-Maisky	1955-66	1987-2014	1,84	1,23	1,80	0,20
Yakutsk	1955-67	2000-2014	3,19	1,20	1,88	0,48
Megino-Khangalassky	1955-75	1986-2014	7,05	2,61	3,02	0,38
Aldansky	1955-72	1984-2014	8,07	1,58	-	0,55

was not made, because these changes in the mean values were induced by the heterogeneity of the variances. For eight forestries with statistically significant increase in dispersion of burnt area, the assumed meteorological factors were chosen: air temperature and atmospheric precipitation for each summer month and average for the season, as well as for the whole continuous historical period from 1955 to 2014, and only for two quasi-stationary half-cycles. As a result, the following most effective regression dependencies for homogeneous periods were obtained:

F_{ff}=26,18T_6 – 11,64T_5-0,166P_8-4,438T_7-202,8 (Neryungrinsky Forestry, 1999-2013), R=0,93 (1),

F_{ff}=10,157T_{sum} – 7,394T_5-113,33 (Vilyusky Forestry, 1955-1974), R=0,88 (2),

F_{ff}=-1,322P_9 + 20,60T_8-254,2 (Yakutsk Forestry, 2000-13), R=0,68 (3),

F_{ff}=25,6T_5+8,62T_6+16,13T_8+12,4T_9-80,57T_{seas}-0,792P_7-1,02P_9+471,
(Yst-Maisky Forestry, 1987-2013), R=0,65, (4),

where F_{ff} – burnt area, in th. ha, Tsum, Tseas – mean temperature for summer and whole forest fire season from May to September, Ti – mean air temperature for each i-month, Pi – mean atmospheric precipitation for each i-month, R – multiple correlation coefficient .

From equations (1) - (4) basically follows that the burnt area will be increase with the temperature rise and decrease of atmospheric precipitation. For the remaining forestries and for the whole period of observations, the empirical equations connecting the burnt area and meteorological parameters turned out to be less effective with correlation coefficients R = 0.4-0.6. Such a low efficiency of the equations can be connected with the fact that only one weather station was chosen for the entire area of each forestry, which is definitely not enough to generalize the atmospheric precipitation patterns.

3.2. Maximum Discharges and Water Levels of Snowmelt Floods

The next considered climate change indicator was the maximum discharges and water levels of snowmelt floods in some of the republic's rivers (Fig.3), for which various models and their characteristics were also calculated, which results are given in the Tab.2.

Figure 3. Scheme of the location of the observational stations which monitoring the soil temperature (circles) and maximum discharges and water levels (triangles) in Yakutia.

Table 2.

Characteristics of non-stationary models of the yearly maximum water levels and discharges.

Code		Δ_{tr}	Δ_{step}	F_{tr}	F_{step}	Beginning year	End year	n	R
Yearly maximum water levels									
1018	Kolyma River – п. Zyryanka	0.9	2.7	1.02	1.06	1938	2014	77	-0.13
1367	Berezovka River – Berezovka	0.3	1.3	1.01	1.03	1963	2014	52	0.07
1801	Kolyma River – Srednekolymsk	0.3	1.7	1.01	1.04	1927	2014	86	-0.08
1802	Kolyma River – Kolymskoe	0	0.8	1	1.02	1965	2014	50	0.01
1805	Kolyma – Chersky	0.6	1.7	1.01	1.04	1960	2014	48	-0.11
3029	Lena River – Krestovsky	4	6.7	1.09	1.15	1950	2013	64	0.28
3030	Lena River – Lensk	0.8	7.6	1.02	1.17	1962	2014	53	0.12
3035	Lena River – Olekminsk	1.9	7.1	1.04	1.16	1950	2014	65	0.19
3036	Lena River – Solyanka	2.8	8.6	1.06	1.2	1950	2013	63	0.23
3405	Olenek River – Olenek	0.9	2.8	1.02	1.06	1936	2013	77	0.14
3414	Yana River – Verkhoyansk	0.1	3.6	1	1.08	1926	2013	82	0.04
3416	Yana River – Batagai	0	4.8	1	1.1	1954	2013	60	-0.01
3491	Indigirka River - Yst-Moma	21.7	19.4	1.63	1.54	1965	2014	50	0.62
3801	Anabar River – Saskylakh	0.2	0.9	1	1.02	1936	2013	75	-0.06
3814	Olenek River – Taymylyr	1.4	4.1	1.03	1.09	1950	2013	57	0.17
3816	Olenek River – Yst-Olenek	2.4	4.3	1.05	1.09	1950	2013	62	0.22
3864	Yana River – Nizhneyansk	3.2	9.5	1.07	1.22	1970	2013	41	-0.25
3881	Alazeya River – Argakhtaakh	2.6	7.6	1.05	1.17	1962	2013	50	0.22
3882	Alazeya River – Andryushkino	0.9	3.8	1.02	1.08	1962	2013	38	-0.14
Yearly maximum water discharges									
1367	Berezovka River – Berezovka	1.2	1.8	1.02	1.04	1965	1999	33	0.15
1801	Kolyma River – Srednekolymsk	0.4	2.4	1.01	1.05	1927	2014	75	-0.09
3029	Lena River – Krestovsky	1	2.9	1.02	1.06	1950	2014	65	0.14
3036	Lena River – Solyanka	3.9	6	1.08	1.13	1950	2014	65	0.28
3405	Olenek River – Olenek	0.4	2.5	1.01	1.05	1936	2013	75	0.09
3414	Yana River – Verkhoyansk	0.4	3.5	1.01	1.07	1936	2013	78	0.08
3801	Anabar River – Saskylakh	1.1	2.2	1.02	1.05	1954	2013	60	-0.15
3881	Alazeya River – Argakhtaakh	6.1	8.8	1.13	1.2	1962	2013	50	0.34

The results placed in Tab.2 show that only in one case (Indigirka River – Hydrological Station Ust-Moma) the non-stationary model is effective and statistically significant. Close to nonstationary models with Δ about 10% and statistically significant R can be considered the series of maximum discharges and water levels on the Alazeya River –Arghakhtakh Hydrological Station, maximum levels on the Yana River – Nizhneyansk Hydrological Station (fall), Olenek River - Ust-Olenek Hydrological Station, the maximum discharge on the Lena River – Solyanka Hydrological Station. Their graphical representation is shown in Fig.4.

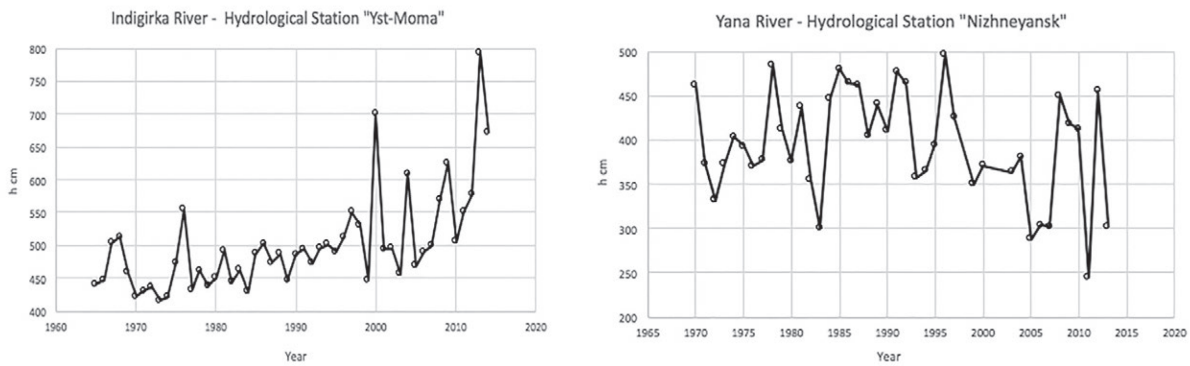

Figure 4. Time-series of non-stationary and close to non-stationary observational ranks of maximum discharges and water levels.

This analysis shows that the growth trend of the maximum water levels is most pronounced on the Indigirka River – Ust-Moma Hydrological Station (Δ_{tr} = 21.7%> Δ_{step} = 19.4%), and also on the Alazeya River – Argachtak Hydrological Station for both water levels and water discharges. At the Lena River – Solyanka Hydrological Station, fluctuations of the maximum water discharges have a stepwise growth in 1990, and for the Olenek River – Ust-Olenek Hydrological Station the upward trend was replaced by a recession in the most recent years, and on the Yana River - Nizhneyansk Hydrological Station was observed the water level decrease in 2000.

Despite the fact that the temperature of May is one of the most non-stationary with average for all meteorological stations of the republic with Δ_{tr} = 9.1%, Δ_{step} = 9.2% and R = 0.41 and the zone of nonstationarity covers almost the half of the territory of Yakutia (Kirillina and Lobanov 2015), manifestations of these temperature changes in the maximum water discharges and water levels are practically absent.

4. CONCLUSIONS

In general, based on the results of the assessment of climate change indicators, the following conclusions can be drawn:

– the dispersion of the burnt area in the forestries located in Central and South-Eastern Yakutia was increased from the end of the 20th and the beginning of the 21st century, which also affected the growth of average burnt area;

– the obtained empirical dependencies, which can be used for scenario assessments of the impact of future climate changes, indicate that the growth of burnt area is directly proportional to the growth of temperatures and inversely proportional to the decrease in precipitation in the months of the warm season;

– the growth in the hydrological response of the climate for the series of maximum water charges and water levels is practically not observed except for some cases, which, however, may be associated with local causes, characteristic for water level changes.

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INCONSISTENCIES IN FINNISH HYDROLOGICAL OBSERVATIONS

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ABSTRACT
Several hydrological analyses have been done based on the observations, showing seasonal changes in the Finnish hydrological regime. Less attention has been paid to the measuring methods and practices or correction methods of observations in time and their possible impacts on hydrological analysis. In this study, known but not yet documented inconsistencies related to Finnish hydrological observations are discussed. Furthermore, the aim is to examine whether these inconsistencies in time series have considerable effects on the outcome of hydrological analysis. The specific focus is on time series of maxima of river flow, ice cover and water temperature. Case studies show that flood frequency analysis based on probability distributions is sensitive on inconsistencies in the time series. Sensitivity analysis did not reveal any changes in the interpretation of the trend analysis with respect of correcting and shifting individual observations of discharge but the cases for water temperature and ice thickness were different. Based on the case studies inconsistencies discussed in this paper can mask the possible trends in the time series or can lead to incorrect rejection of the null hypothesis of "no trend". However, a more thorough analysis is needed.

KEYWORDS
Hydrology, discharge, ice, water temperature, inconsistencies, flood frequency analysis, trends

1. INTRODUCTION
Finnish hydrological observations have been used in several studies related to climate change, modelling and water resources assessment (e.g. Korhonen, 2006; Korhonen & Kuusisto 2010; Veijalainen et al. 2012). Several trend analyses have been done based on the long time series in Finland, showing seasonal changes in hydrological regime. Winter flows have increased, spring peaks became earlier and ice cover shortened. Less attention has been paid to the measuring methods and practices and correction methods of records in time and their possible impacts on outcome of hydrological analysis. In this study, known but not yet documented inconsistencies related to Finnish hydrological observations are discussed. The aim of this research is to examine whether these inconsistencies in the time series have considerable effects on the outcome of hydrological analysis. The specific focus is on time series of maxima of river flow, ice cover and water temperature. A sensitivity analysis is conducted and the effects of inconsistencies on results of hydrological analysis are estimated. The study is based on case studies of certain hydrological stations in northern Finland.

2. FINNISH SURFACE WATER HYDROLOGICAL OBSERVATIONS
2.1. Discharge
Rating curves can be used for estimation of discharges when discharge depends on stage alone. With good site selection same rating can be used for years. However, by analysing the measurement practises and correction procedures in time reveal at least three inconsistency issues in Finnish daily discharges. Firstly, measurement techniques have changed. Before the

year 1986 daily discharge is based on single water level reading at 8 am. Later values in the time series are daily averages from the chart recorder or averaged from momentary values (pressure transducers). In addition to change in the way the daily value is estimated, also the accuracy of the water level reading may have been different during the years.

Secondly, hydraulic properties of the rivers are affected by ice cover and open water rating curve cannot be used during winter. In Finland correction has always been based on the subjective process of the responsible hydrologist and there is no guarantee that another hydrologist would correct exactly the same way. Indeed, it has been shown that they don't (e.g. Hamilton et al. 2000). This issue is extremely likely to cause big uncertainties in the annual maxima in northern Finland as in some years the maximum flood must be ice-corrected. Uncertainties related to ice cover corrections should also always be carefully studied before making any conclusions on long-term changes in winter discharges.

Thirdly, there is a possible inconsistency in the time series related to the different rating curves used. Based on the careful analysis of the actual discharge measurements of some of the Finnish stations, it is possible to argue that instead of many there has actually always been only one rating that should have been used for the whole observation period. Changes in the ratings affect especially the highest flows as rating curves have been extrapolated well beyond the highest ever measured discharge and these parts of the different curves have big differences. Same applies also to very low flows.

2.2. Ice thickness

Observations on lake ice thickness are made usually rather near to a water level gauge. The ice thickness is nowadays measured in the wintertime every 10th, 20th and 30th of the month. This measuring practice has been used since the late 1970s. Earlier measurements were made twice a month every 15th and 30th of the month. Nowadays, manual gauge and a drill are used when measuring ice thickness and the average value from three different holes is defined as the official observation. Before the year 1982 only one ice drilling hole was used.

2.3. Water temperature

The surface water temperature measurements are made at several water level gauges during the open water season at 8 am. These measurements are done in the depth of 20 cm in the proximity of the shoreline. Measurements were made manually by observers in the 20th century. During the last decades, sites have been automated with floating temperature sensors. Even though manual measurements should have been done always in the morning, it is possible that not all observations are exact 8 am observations. Diurnal fluctuation of surface water temperature is large during the timing of warmest days of the season and measuring time has an importance (Korhonen et al. 2015). Automatic sensors measures exactly at 8 am but location of the sensor cannot be that close to the shore due to lowering of water level during the summer. Automatic devices are thus located bit further from the shoreline and water depth below automatic sensor is usually greater than in the manual sites. However, measurement depth is the same.

2.4. Case studies and data used

2.4.1. Discharge

To study the possible effects of inconsistency issues in discharge two northern stations were selected for a sensitivity study. Case studies were delimited to northern stations and certain analysis, although effects may be different for other sites and hydrological analysis. Case studies were constrained to uncertainties related to flood peaks, specific focus on the effect of inconsistencies on flood frequency and trend analysis.

Two northern rating curve stations were analysed, namely 7101100 Repojoki and 7101320 Ivalojoki, Pajakoski. Sites selected are located at the same pristine basin (no regulation etc.) and have rather long records. They are partly located in a national park and thus have relatively small human impact caused by land use changes.

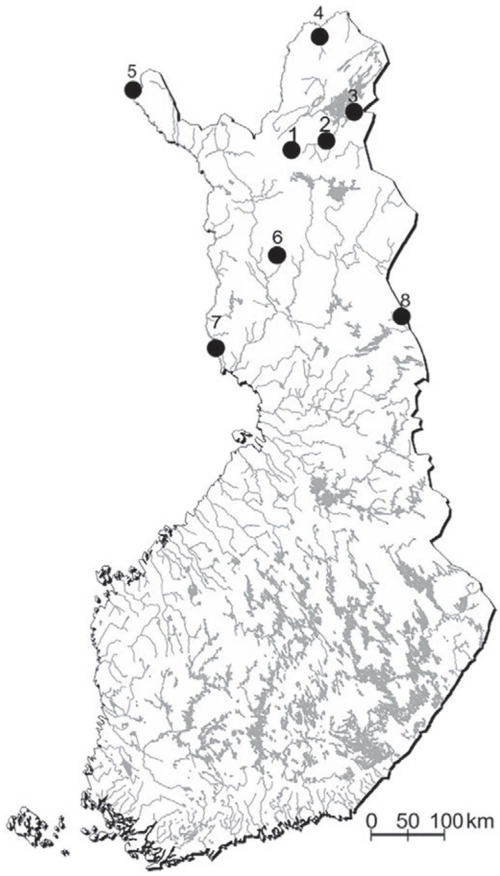

Figure 1. Locations of the stations.

Station 7101110 (Point 1 in Fig. 1) is used to study the effect of rating curve uncertainties and the effect of ice correction of the winter discharges on the hydrological analysis. This station was suitable for such analysis as a total of four different rating curves have been used during the observation period (1974-2016). In addition, 7 out of 43 annual flood peaks are in fact ice corrected and thus have big uncertainties.
Station 7101320 (point 2 in Fig. 1) is used to study the sensitivity of the measurement technique and length of the time series on trend analysis and flood frequency analysis of flood peaks. There are momentary values available for the station since 1989 and thus it is suitable for comparison analysis between 8 am momentary values and daily averages. The time series starts from 1961 but years of 1970-1973 are missing. Furthermore, years 2002 and 2014 (not extreme years) are excluded from the study as their flood peak observations have been corrected because of the problems of automatic measurement device.

2.4.2. Ice thickness and water temperature
The case study of maximum ice thickness is based on four stations in northern Finland with long time series. Stations selected are lakes Inari, Kevojärvi, Kilpisjärvi and Unari (Points 3, 4, 5, 6 in Fig. 1). All sites have continuous records of ice thickness since the 1960s.

The case study of maximum surface water temperature included four stations in northern Finland with long time series. Stations selected are lakes Inari and Kevojärvi and rivers Tornionjoki and Oulankajoki (Points 3, 4, 7, 8 in Fig. 1). These stations have continuous record since the 1960s except Oulankajoki since 1970. These sites were automated in following years; Inari 2011, Kevojärvi 2016, Oulankajoki 2013. Tornionjoki station has not yet been automated. Data until 2016 was used for both variables.

3. METHODS

3.1. Mann-Kendall for trend analysis

Trend analysis was conducted with non-parametric Mann-Kendall (M-K) trend test (Mann 1945, Kendall 1955). The M-K trend test does not require data to be normally distributed. The null hypothesis for the M-K test is that the data are independent and randomly ordered, i.e. there is no trend or serial correlation structure among the observations.

The level of 5% was used for critical significance. M-K trends were calculated with 'MAKESENS', an excel program provided by the Finnish Meteorological Institute (Salmi et al. 2002).

3.2. Flood frequency analysis

Several different extreme value distributions can be used for flood frequency analysis. At least Kuusisto & Leppäjärvi (1979) and Veijalainen (2004) have studied different distributions with Finnish discharge data. Both of these have concluded none of the studied distributions is generally best for the Finnish data. Gumbel distribution is, however, widely used in Finland and it is the default distribution to be used in flood frequency analysis made by the Finnish environment authorities.

In this study two parameter Gumbel distribution is used for flood frequency analysis. Parameters are calibrated with the method of moments and specific focus is on floods occurring on average once every 50 years. This threshold is used by the insurance companies whether to compensate on flood damages or not.

3.3. Rating curves

Within time rating curves have been estimated in Finland by using all the available gauging data, and specific knowledge of the site. Curves have then been estimated by drawing into millimetre papers and then later digitized from these drawings. Theoretical background on estimating the curves have been on power law.

To analyse uncertainties of the rating curves, historical curves are compared with the curve that is estimated based on Bayesian statistics utilising all the available gaugings. Bayesian Rating Tool by Hrafnkelsson et al. (2015) available online is used to estimate this rating curve.

3.4. Ice correction

In Finland estimation of streamflow during the winter season is still based on the subjective process by a hydrologist. Nowadays, a graphical interface is used providing visual tool with all the relevant data. In this process winter discharge during ice cover season is subjectively interpolated by utilizing winter season discharge measurements, simulation results of the conceptual hydrological model, temperature and precipitation observations, and stage observations. Furthermore, streamflow observations of nearby gauging points and stage-discharge curve for winter season are used whenever available. Also freeze-up and break-up observations are available for some sites. Despite of all the data available, the process of estimating winter streamflow is still subjective and time-consuming.

During the years there have been at least three different hydrologists that have been doing ice corrections for the study sites. On the other hand conceptual hydrological model has been available for only for some last 15 years. To analyse the effect of all these changes ice correction of flood peaks were re-estimated for the whole time series (1974-2016) with current tools. This “other truth” is used to compare whether such change would have consequences in analysis of flood peaks.

3.5. Sensitivity analysis

Following data sets are used to analyse the sensitivity of the trend analysis and flood frequency analysis on the data used.

3.5.1. Flood peak data sets for 7101100 Repojoki

Data set 1: Flood peak time series from the data base. Values for years 1974-1985 are maximum momentary values at 8 am each year and 1986-2016 are maximum daily averages each year from chart recorder or from automatic device (pressure transducers.)
Data set 2: original flood peak time series (Data set 1) with re-evaluation of ice-corrections.
Data set 3: flood peak time series with Bayesian rating curve and original ice corrections.
Data set 4: flood peak time series with Bayesian rating curve with re-evaluation of ice-corrections.

3.5.2. Flood peak data sets for 7101320 Ivalojoki, Pajakoski

Data set 5: time series for 1961-2016. Values for years 1961-1984 are maximum momentary discharge at 8 am each year and 1985-2016 are maximum daily average each year from chart recorder or from automatic device (pressure transducer.)
Data set 6: time series 1961-2016. Maximum value of momentary discharge at 8 am each year.
Data set 7: time series 1985-2016. Maximum value of discharge at 8 am each year.
Data set 8: time series 1985-2016. Maximum value of daily average each year.
Data set 9: time series 1985-2016. Maximum momentary peak discharges each year.

3.5.3. Monte Carlo simulations – ice thickness and water temperature data

Inconsistencies in maximum ice thickness and water temperature time series were investigated with Monte Carlo simulations. In addition, the effect of the length of the time series on trends is studied.
The inconsistency between values of one and average of three drilling holes is studied. It is assumed that measurements of one drilling hole result higher maxima than average of three holes. On the other hand, two measurements per month capture actual maxima less likely than three measurements per month. Analysis of few northern stations showed that maximum difference between three holes is approximately 2 cm in the time of maximum ice thickness. Earlier in the winter it is larger. The sensitivity analysis was done by adding a random whole number between 0 and 2 to the maximum values of ice thickness for 1982-2016. This was done for twenty times (MC simulation). For the years 1961-1981 original one drilling hole values were used. After that, trends of these different time series were tested with M-K trend test for different record lengths (1961-2016, 1971-2016 and 1981-2016). Random number generator of MS Excel was used.
It is known that diurnal variation of surface water temperature can be on a sunny day several degrees (Korhonen et al. 2015). The differences are larger in lakes than in rivers where water mixes. Based on the data, diurnal variation is typically max 2 degrees during summer maximum in lakes of northern Finland, in rivers less. Considering the diurnal variation of water temperature

and possibility that some of the manual measurements have been made later than 8 am a sensitivity analysis was done. This was done by creating twenty samples by Monte Carlo simulation. A random whole number was subtracted from the original maximum water temperature time series during manual measurement years (automation year varied between sites, see 2.4.2). Since river Tornionjoki site is not yet automated, only effect of possible delayed observation time is studied . For rivers, numbers between 0 and 1 were used and for lakes 0 and 2. After that, trends of these different time series were tested with M-K trend test for different record lengths (1961-2016, 1971-2016 and 1981-2016). Random number generator of MS Excel was used.

4. RESULTS AND DISCUSSION

4.1. Maximum discharge

Discharge data from station 7101100 were used to examine whether uncertainties related to ice correction of flood peaks have considerable effect on the results of flood frequency analysis and flood peak trend analyses. In addition, data was used to study the sensitivity of the hydrological analyses on issues in validity of different rating curves. Data sets 1 and 4 are presented in Figure 2a, all the rating curves and the years they are valid in Figure 3. Results of the trend analysis and flood frequency analysis based on Data sets 1-4 are presented in Table 1.

Although they were some big changes in the flood peaks especially because of the re-estimation of ice-correction there were no changes in the results of the trend analysis. With 5 % significance level null hypotheses of "no trend" could not be rejected in any of the four cases.

As expected changes are more visible in flood frequency analysis as parameter estimation of the Gumbel distribution with small number of observations is sensitive on the largest floods. Mean estimates of $HQ_{1/50}$ vary between 246 and 314 m^3/s for data sets 1-4 and difference is just on the threshold of two-times the standard error of estimate.

By using a single rating curve (Bayesian statistics) for the whole period the mean estimate of $HQ_{1/50}$ dropped from 314 to 292 m^3/s but the difference is within reasonable uncertainty bounds of the original $HQ_{1/50}$.

Re-estimation of ice-corrected flood peaks has a lot bigger effect on the results of flood frequency analysis. Some of the biggest floods were scaled down and consequently flood estimates dropped respectively. With original rating curves $HQ_{1/50}$ drops from 314 to 246 m^3/s and with a single Bayesian rating curve from 292 to 262 m^3/s.

Generally, changes in mean estimates are so large that this inconsistency issue in the data should be carefully considered before any conclusions are made on flood frequencies at Repojoki station.

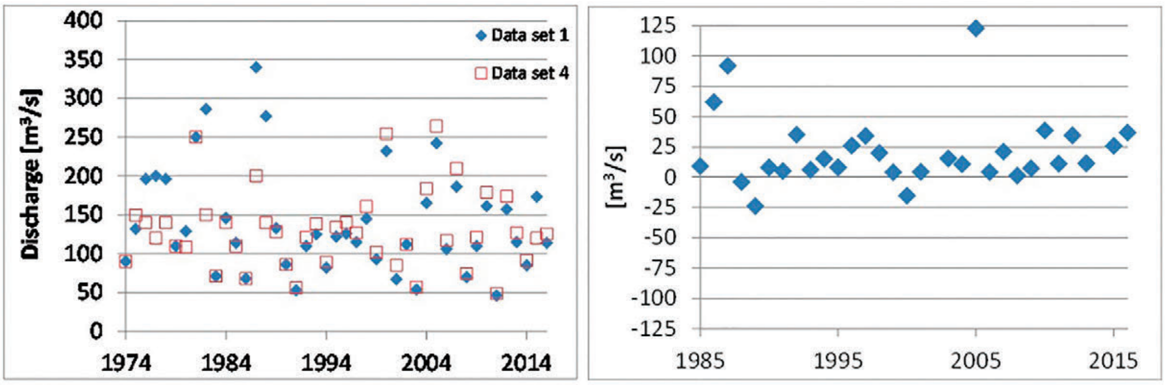

Figure 2. On left (2a) two data sets of 7101110 Repojoki. On right (2b) the difference between maximum momentary discharge at 8 am and maximum daily average each year at 7101320 Ivalojoki, Pajakoski.

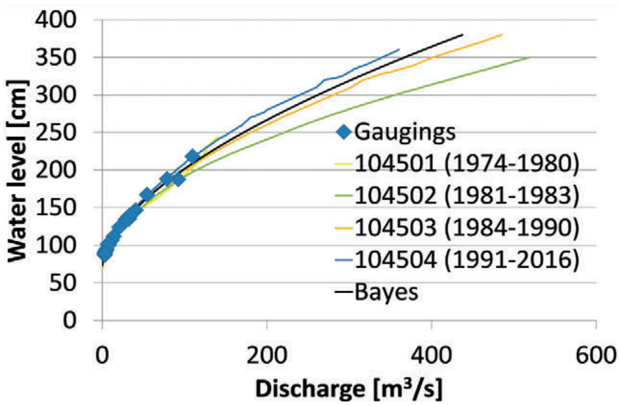

Figure 3. Rating curves for 7101110 Repojoki for different years and a single rating curve for the whole period estimated with Bayesian statistics by using all the gaugings.

Table 1.

Results of hydrological analysis for different data sets. Data set 1-4 are related to 7101100 Repojoki and data sets 5-9 to 7101320 Ivalojoki, Pajakoski.

Data set	M-K test statistic	$HQ_{1/50}$ [m^3/s]	$HQ_{1/50}$ standard error of estimate [m^3/s]
1	-1,21 (No trend)	314	35
2	-0,73 (No trend)	246	24
3	-0,61 (No trend)	292	30
4	-0,14 (No trend)	262	26
5	0,54 (No trend)	990	93
6	0,81 (No trend)	1036	99
7	-0,29 (No trend)	1113	143
8	-0,39 (No trend)	1044	131
9	-0,39 (No trend)	1183	154

Difference between 8 am yearly maximum values and daily average yearly maximum values at station 7101320 can be seen in Figure 2b. It was expected that 8 am values would randomly vary around daily average values. This was not true however. In the data set (1985-2016) 8 am value was smaller than daily average only 3 times out of 30 years. It seems that timing of the highest flood and shape of the hydrograph are reasons for this phenomena but such conclusion would require a more thorough analysis. If daily average values are used for more recent years instead of 8 am at this station, it will lead to slightly smaller value of $HQ_{1/50}$ (990 vs. 1036 m^3/s) but mean estimates are well within uncertainty bounds. For comparison, also momentary flood peak data were studied. Logically, use of momentary flood peak data will increase the $HQ_{1/50}$ greatly compared to use of maximum of daily averages. At this station with data set 1985-2016 from 1044 to 1183 m^3/s. Difference is a bit smaller if compared against momentary flood peaks at 8 am each year (1113 vs. 1183 m^3/s). Differences between the outcome of the flood frequency analysis of daily average and momentary flood peaks are, however so large that it might have consequences in water use and land use planning if not taken into account properly.Discharge data from station 7101320 were also used to examine whether inconsistency in the measurement technique has effect on outcome of trend analyses of flood peaks and flood frequency analysis. Results of the analysis are presented in Table 1. Similarly to case studies related to data from station 7101110, the outcome of the trend analysis did not change between data sets. There was no trend in the flood peak time series.

4.2. Maximum ice thickness

Twenty Monte Carlo simulations for lake ice thickness maximum and the original time series on different time periods (1961-2016, 1971-2016, 1981-2016) were analysed for the study sites. The results show that random correction of max 2 cm to maximum ice thickness to the latter part of the time series does not change the interpretation of trend analysis (5 % significance) from the original one (Table 2) except for lake Kevojärvi (1963-2016). The selection of time period for the trend analysis led to different interpretation of the trend analysis at lakes Inari, Kevojärvi and Unari. Trends were stronger for shorter time periods. All trends were negative resulting decreasing maximum ice cover. Taken into account that random correction was towards positive trends, this declining is strong trend. For the latter period starting 1981 measurement method has been same for the most of the period.

4.3. Maximum surface water temperature

Twenty Monte Carlo simulations of maximum surface water temperature in addition to the original time series for different time periods (1961-2016, 1971-2016, 1981-2016) were analysed. The results show that random shift of 0 to 2 centigrade downwards to the maximum surface temperature of manual measurements changed the interpretation of trend from the original data set (Table 3) except for Tornionjoki (which is not automated). None of original data series did reveal any statistically significant trends but after manipulation of individual observations, some time series showed trends. At lake Kevojärvi and river Oulankajoki, also the length of the time series affected the interpretations of the trends.

Table 2.

Ice trends for different sites and time periods: the original time series (italic) and twenty Monte Carlo simulations varying range (). Trends of * refer for 0.05, ** 0.01 and *** 0.001 significance level. Lake Kilpisjärvi observations started in 1965, Kevojärvi in 1963. M-K is Mann-Kendall test value. No refers no trends.

Site	1961-2016	M-K	1971-2016	M-K	1981-2016	M-K
lake Inari	* (*)	***-2.53*** (-2.47 … -2.11)	*** (**/***)	***-3.39*** (-3.44 … -3.03)	** (*/**)	***-2.84*** (-3.10…. -2.57)
lake Unari	***No*** (No)	***-1.78*** (-1.56 … -1.27)	* * * (**/***)	***-3.58*** (-3.49 … -3.19)	*** (***)	***-3.97*** (-3.97… -3.67)
lake Kevojärvi	* (No/ *)	***-2.29*** (-2.24… -1.85)	* (*/**)	***-2.58*** (-2.58… -2.14)	* (*/**)	***-2.41*** (-2.72… -2.08)
lake Kilpisjärvi	***No*** (No)	***+0.86*** (+1.05… +1.42)	***No*** (No)	***+1.10*** (+1.11… +1.54)	***No*** (No)	***-0.27*** (-0.72… -0.08)

Table 3.

Water temperature trends for different sites and time periods: the original time series (italic) and twenty Monte Carlo simulations varying range (). Trends of * refer for 0.05, ** 0.01 and *** 0.001 significance level. River Oulankajoki observations started in 1970, lake Kevojärvi in 1962. M-K is Mann-Kendall test value. No refers no trends.

Site	1961-2016	M-K	1971-2016	M-K	1981-2016	M-K
lake Inari	***No*** (No/*)	***+1.10*** (+0.13… +2.00)	***No*** (No/*)	***+1.21*** (+0.24… +2.30)	***No*** (No/ **)	***+1.36*** (+0.46… +2.77)
lake Kevojärvi	***No*** (No/*)	***+1.59*** (+0.42… +2.14)	***No*** (No)	***+0.47*** (-0.46… +1.22)	***No*** (No)	***+1.37*** (+0.36… +1.84)
river Oulankajoki	No data	No data	***No*** (No)	***+0.95*** (+0.26… +1.37)	***No*** (No/ *)	***+1.60*** (+0.91 +2.22)
river Tornionjoki	***No*** (No)	***+0.57*** (-0.40… +1.29)	***No*** (No)	***-0.41*** (-1.33… +0.20)	***No*** (No)	***-0.01*** (-0.95… +0.82)

5. CONCLUSIONS

Known but not yet documented inconsistencies related to Finnish hydrological observations were discussed. Inconsistencies in the time series are mainly caused by changes in measurement technique but also on processes on correcting values and estimating rating curves (discharge). The specific focus in this study was on time series of maxima of river flow, ice cover and water temperature. Case studies were used to examine whether inconsistencies are so large that they might affect interpretation of hydrological analysis. Results showed that flood frequency analysis is sensitive on the highest floods in the record. Possible inconsistencies in the time series may change the results of the analysis so much that it might have consequences e.g. in water use, land use planning and e.g. compensations of flood damages. Sensitivity analysis did not reveal any changes in the interpretation of the trend analysis with respect of correcting and shifting individual observations of discharge. For ice thickness and surface water temperature the interpretation of the trend analysis changed between the original time series and different Monte Carlo simulations at some sites. It was shown that inconsistencies discussed in this paper can mask the possible trends in the time series or can lead to incorrect rejection of the null hypothesis of "no trend". However, a more thorough analysis is needed.

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RUNOFF CHANGES AT THE POLE OF COLD OF NORTHERN HEMISPHERE

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ABSTRACT

Analysis of daily runoff data for 19 hydrological gauges in the basins of the Yana and Indigirka Rivers has shown the presence of statistically significant ($p < 0.05$) positive trends in monthly runoff in the autumn-winter period. In October, runoff increases at 11 of 19 rivers, in November and December – at 7 and 5 of 11 non-freezing rivers. The trend values in these months are 62% /1.9 mm, 64% /0.4 mm and 84%/0.1 mm, respectively. At 10 watersheds with an area of more than 7,600 km^2, spring flood increases in May with a median trend value of 93% or 7.8 mm. The annual air temperature has risen by an average of 2.0 °C in 1966-2012. The analysis of monthly precipitation revealed the absence of systematic statistically significant trends, however, at some stations in the basin of the river Indigirka a decrease in precipitation in January and February is observed. 5-7 days earlier shift of spring flood dates has been revealed.

KEYWORDS

Yana and Indigirka Rivers, runoff, changes, permafrost, trend, spring flood, warming, winter precipitation

1. INTRODUCTION

Numerous studies have shown that river streamflow in Northern Eurasia and North America is increasing (Holland et al 2006; Shiklomanov & Lammers 2009, 2013; Rawlins et al. 2010). Most of them are focused exceptionally on “Big 6” Arctic rivers – (the Ob’, Yenisey, Lena, Mackenzie, Yukon, and Kolyma; Holmes et al. 2013). Large Arctic river basins are characterized by a great variety of climatic, landscape and permafrost conditions and the mechanisms of the observed changes could hardly be understood on the scale of “Big 6”. Although climate model simulations show increased net precipitation over the pan-Arctic watershed it is not always supported by ground meteorological data analysis (Rawlins et al. 2010). Runoff change does not necessarily coincide with precipitation and evaporation potential changes (Gao et al. 2016) but usually agree with increase of air and soil temperature. In some cases there is an opposite change direction in runoff from that in precipitation (Karlsson et al. 2015).

River runoff changes estimates in Eastern and North-Eastern Siberia are limited and contradictory. Magritsky et al. (2013) reported an increase in the total runoff of Yana and Indigirka during the period 1976-2006 by 1.5-3% compared to the period before 1976 and noted that the runoff of these rivers increased in summer and autumn by 20-25% and did not change in winter. According to Georgievsky (2016), on the contrary, there is an increase of Yana and Indigirka rivers winter runoff to 40% over the period 1978-2012 compared with the period 1946-1977, and an increase in the spring flood. Majhi and Daqing (2011) concluded that the Yana River monthly flow rises at the Jubileynaya gauge (224,000 km^2) for the period of 1972-1999 for June, August, September, October and April, while May, July and March monthly flow have decreased. Bring and Destouni (2014) reported discharge increase in the Yana, Indigirka and

most of other major Arctic rivers. They noted that it has been greater than the increase in precipitation for period 1991–2002 comparative to 1961–1991.
The flow of small and medium rivers in the cold regions has been studied fragmentarily. Tananaev et al. (2016) found for the Lena River basin that thirty small and medium-sized rivers out of 100 showed trends in the mean annual daily flow and 35 out of 55 – in the minimum daily flow. Assessments of streamflow changes of the small and medium-sized rivers in the Yana and Indigirka river basins do not exist.
The objective of the research was quantitative assessment of current changes in the hydrological regime in two large arctic river basins – the Yana and Indigirka, which basins are completely located within the continuous permafrost zone and which have long-term runoff observations along the main rivers and their tributaries of different sizes.

2. STUDY AREA

The research territory is known as the region, where the Northern Hemisphere's Cold pole is located. Absolute minimum can reach record levels: down as far as -71 ˚C in Oymyakon and -68 ˚C in Verkhoyansk (Ivanova 2006). The research region's climate is distinctly continental. Long-time average annual air temperature changes from -16.1 ˚C (Oymyakon, 726 m, 1930-2012) to -13.1 ˚C (Vostochnaya, 1288 m, 1942-2012). Minimum mean temperatures are typically observed in January and can peak at -47.1 ˚C (Oymyakon) and at -33.8˚C (Vostochnaya). Maximum average annual temperatures are observed in July, averagely 15.7 ˚C and 11.8˚C for stations Yurty (590 m, 1957-2012) and Vostochnaya, correspondingly. Steady cold weather begins in the first decade of October, spring starts in the second part of May or at the beginning of June, when seasonal snow cover starts to melt. Snow accumulation is relatively small and accounts for 25-30 cm. Annual average precipitation sum at Verkhoyansk meteorological station (137 m, 1966-2012) is about 180 mm, at Vostochnaya station (1 288 m, 1966-2012) is 280 mm. Most precipitation (over 60%) occurs in summer, in July their amount reaches the peak, up to 70 mm per month (Vostochnaya, 1966-2012).
The distributional pattern of the permafrost defines rivers hydrological regime, stipulating interaction mechanisms between groundwater and surface flow, water-bearing rocks forms and permafrost. In the studied river basins, permafrost depth can reach over 450 m at watershed divides and up to 180 m in river valleys and intermountain depressions. At the water-permeable sites, taliks interrupt permafrost into fractured rocks zones with the rocks depth exceeding the permafrost depth.
High absolute elevation points of the Verkhoyansk (Orulgan, 2 389 m), Cherskiy (Pobeda, 3 003 m) and Suntar-Khayata (Mus-Khaya, 2 959 m) ranges go with their slopes being broken down by river valleys. Landscapes of the studied territory include mountain larch open woodland and boreal larch taiga with continuous moss and lichens cover. In river valleys, grass-moss larch forest on bog soil is widespread. Goltsy and small glaciers are typical for high altitude mountain areas. Average snow line altitude at glaciers accounts for 2 350-2 400 m (Grave et al. 1964). Aufeises (naleds), which form at mountain foothills, as well as in sub-mountain and intermountain depressions, are another distinguishing feature of the region (Sokolov 1975).
The rivers have Eastern Siberian type of the water regime, characterized with typical spring-summer flood, high summer-autumn rainfall floods and low winter flow. In winter, small and medium-sized rivers freeze thoroughly. Spring flood starts in May-June and lasts approximately for a month and a half. In summer, deglaciation waters as well as ones from melting aufeises and snowfields add to rainfalls.
Daily discharges series for 19 hydrological gauge stations of Roshydromet's network in the Yana and Indigirka river basins were analyzed. Most of the stations had been under operation until

2014-2014, two of them were ceased from operation in 2008 (Table 1, Figure 1). Ten out of 19 examined catchments are less than 10 000 km^2, seven of them – less than 100 km^2. Maximum catchment area is 89 600 km^2 (riv. Adicha – sett. Yurdyuk-Kumakh). Average elevation above sea level of the examined catchments varies from 320 m (Khoptolooh stream – Verkhoyansk) to 1 410 m (riv. Suntar – riv. Sakhariniya mouth). Thus, the study covers not only large, but also small and medium-sized rivers over a broad range of elevations typical for the studied mountain region.

Table 1

Analyzed gauging stations

Code	River – gauge*	Start of discharge observations (year)	Basin area (km^2)	Outlet/average catchment elevation (m)	Long-time average annual runoff depth (mm)
3516	Dunai (Ambar-Yuryuete) –Rempunkt	1964	16.6	494/1060	362
3433	Khoptolookh –Verkhoyansk	1968	18.3	133/320	58
3527	Blizhniy – 0.3 km upstream of the mouth	1945	23	578/900	108
3431	Tinege – Ekchyuchyu	1980	59.2	192/520	136
3501	Sakharinya –stream mouth	1957	84.4	833/1120	98
3480	Turagas – 1.2 km upstream of the mouth	1969	98	187/570	81
3510	Artyk-Yuryakh – 3.5 km upstream of the mouth	1946	644	591/900	89
3479	Borulakh – Tomtor	1956	7570	175/540	71
3499	Suntar river – riv. Sakharinya mouth	1956	7680	828/1410	189
3424	Sartang – Bala	1957	16700	136/700	94
3507	Elgi – 5.0 km upstream of the river Artyk-Yuryakh mouth	1946	17600	594/1140	210
3518	Nera – Ala-Chubuk	1945	22300	568/1150	174
3430	Dulgalaakh – Tomtor	1956	23900	145/930	143
3483	Bytantay – Asar	1945	40000	80/750	123
3414	Yana – Verkhoyansk	1936	45300	125/740	112
3488	Indigirka – Yurty	1956	51100	578/1330	155
3443	Adycha – Ust'-Charky	1960	52800	259/960	203
3489	Indigirka – Indigirskiy	1944	83500	482/1250	168
3445	Adycha – Yurdyuk-Kumakh	1937	89600	108/880	192

3. DATA AND METHODOLOGY

The study used daily discharge data for the entire observation period from 1936 up to and including 2014, published in Hydrological Yearbooks (Hydrological Yearbooks 1936-1980; State Water Cadastre, 1981-2007) and available for the period 2008-2015 on the website of the Automated infrmation data system for state monitoring of water bodies (AIS SMWB) (URL: https://gmvo.skniivh.ru). They were used to form series for the analysis: monthly and annual flow depth (mm) and flood starting dates (their counting numbers from the beginning of the year). A day was considered as a flood starting date if its discharge reached or exceeded 20% of the average discharge value in the studied year.

Daily air temperatures and precipitation data series, observed at eight weather stations (the elevation range varies from 137 to 1288 m), located in the studied basins, over different periods from 1930 (but not later than 1966) to 2012 were reduced to average monthly values series.

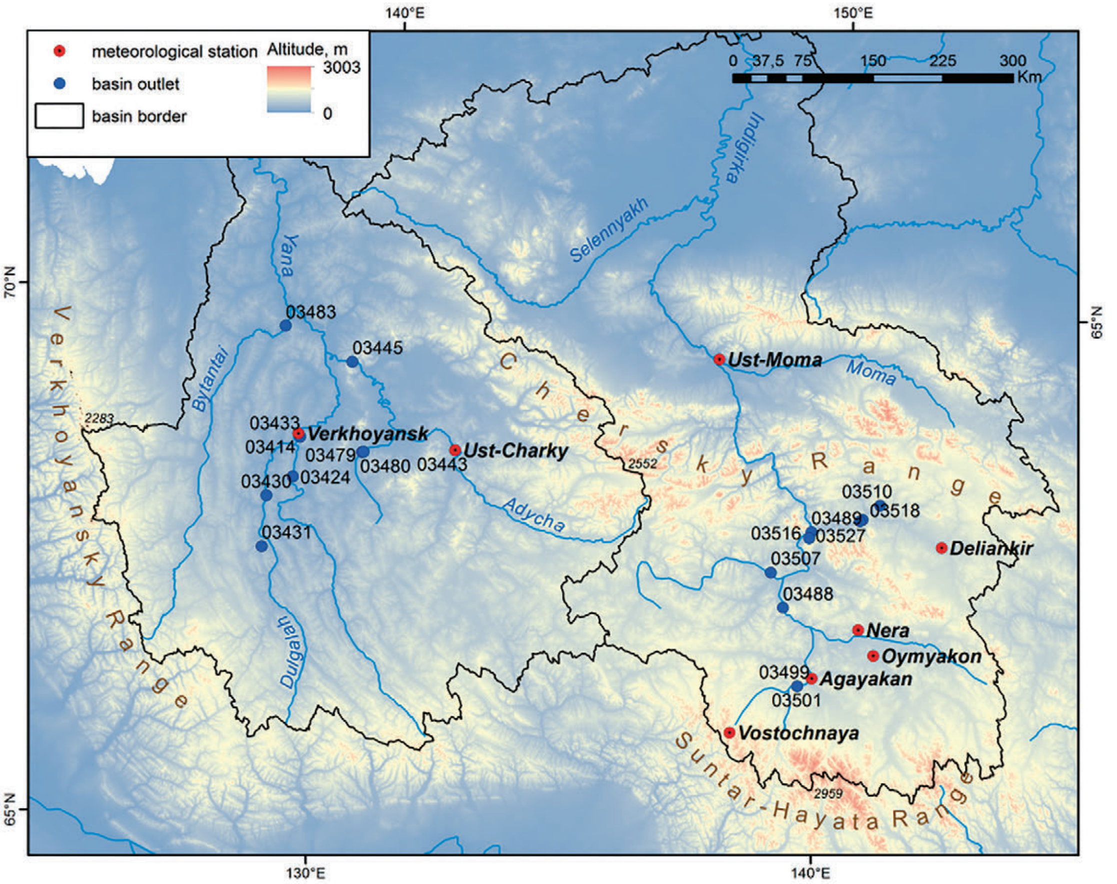

Figure 1. Study area

3.1. Methods

Time series of runoff characteristics (monthly runoff depths and flood starting dates) and meteorological elements (air temperature and precipitation) were evaluated for stationarity, in relation to presence of monotonic trends, with Mann-Kendall and Spearman rank-correlation tests, at the significance level of (Mann 1945; Kendall 1975). If both tests proved a trend at the significance level of , a serial correlation coefficient was tested. With the serial correlation coefficient r < 0.20, the trend was considered reliable. In the case of r≥ 0.20, to eliminate autocorrelation in the input series «trend-free pre-whitening» procedure (TFPW), described by Yue (Yue et al. 2002), was carried out. «Whitened» time-series were repeatedly tested with Mann-Kendall non-parametric test at the significance level of . Trend value was estimated with Theil-Sen estimator (Sen 1968).

4. RESULTS

Positive statistically significant trends ($p< 0.05$) in monthly runoff depth were identified during two periods: in autumn-winter and in the first month of a spring flood – in May for most stations. In May, a reliable increase in streamflow is observed at 10 catchments and in June, at one of the 19 studied ones; median trend rates are 93% or 7.8 mm in May and 64% or 12.8 mm in June. Areas of catchments, where spring flow increases, exceed 7 600 km^2. At small catchments with areas less than 700 km^2 there is no increase in spring flow. There is no up-to-date runoff data for the rivers with catchment areas 644 to 7 680 km^2.

In August, an increase in streamflow was identified at 6 out of 19 rivers (median runoff depths change throughout the entire period of observations by 57% or 21 mm). Areas of the catchments, where discharge increases in August, vary from 644 to 89600 km^2.

In September, a positive trend was identified at 14 out of 19 gauging stations (median is 64% or 9.8 mm); two of these rivers are small ones (catchment areas are less than 30 km^2), the others are medium and large rivers, except for the Bytyntay river, for which the trend is identified only at the level of significance of $p< 0.27$.

In October, streamflow increases at 11 out of 19 rivers (median trend rate is 62% or 1.9 mm), in November – at 7 out of 11 non-frozen streams (median trend rate is 64% or 0.4 mm), in December – at 5 out of 11 non-frozen streams (median trend rate is 84% or 0.1 mm).

For the Indigirka river basin, a positive trend in January-April, before the flooding start, is observed at two gauging stations: the Indigirka river (township Indigirskiy) and its tributary – the Elgi river – 5.0 km above the Artyk-Yuryakh river's mouth, with the median monthly trend rate of 85 % (0.06 mm). Therefore, at these two gauging stations low flow increases throughout the autumn-winter period.

Runoff changes in different months have not led to significant trends in annual runoff for most studied rivers. At 4 gauging stations (the Elgi and Adycha (2 stations) rivers and Dunai stream) out of 19 studied ones, annual runoff has increased significantly (from 32 to 47%). Months of August and September have contributed to the annual runoff rate changes at these gauging stations most dramatically. Median trend rate accounted for 39% or 87.4 mm per year. It can also be noted that these rivers have the highest annual runoff rates among all the studied ones (Table 2).

Table 2

Changes of monthly (May to Dec) and annual runoff (mm) from the beginning of observations*

Gauge	Basin area, km^2	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
3516	16.6	**25.7**	12.4	-9.1	18.3	**36.8**	**1.6**	0.0	0.0	**115**
3433	18.3	4.3	4.8	0.0	1.7	1.5	0.0	NA	NA	25
3527	23	3.9	-11.1	10.9	6.4	**8.8**	0.0	NA	NA	21
3431	59.2	7.0	16.0	-3.7	-1.2	3.4	0.0	NA	NA	45
3501	84.4	0.0	0.0	-2.6	-6.1	0.0	0.0	0.0	0.0	29
3480	98	2.0	11.0	0.9	13.9	4.8	0.0	NA	NA	38
3510	644	5.2	-14.2	5.3	**14.0**	**6.3**	0.1	0.0	NA	21
3479	7570	5.6	**12.8**	4.6	**8.3**	**5.5**	0.1	0.0	NA	38
3499	7680	**6.8**	11.6	-17.6	-2.6	**9.9**	**3.3**	0.4	0.0	26
3424	16700	2.9	7.3	-5.5	0.4	**5.7**	**0.8**	0.1	0.0	9
3507	17600	**15.3**	4.0	2.9	**23.4**	**16.5**	**2.1**	**0.6**	**0.10**	**81**
3518	22300	**11.8**	-12.3	-3.2	3.9	**9.1**	**0.9**	**0.2**	0.0	14
3430	23900	**5.5**	12.6	-2.2	0.4	**7.6**	**2.0**	**0.4**	**0.08**	20
3483	40000	**3.6**	3.4	1.8	9.0	5.7	**1.3**	0.3	0.0	24
3414	45300	**4.3**	7.6	3.2	5.5	**5.8**	**1.0**	**0.3**	**0.14**	24
3488	51100	**7.6**	9.9	-4.4	7.6	**12.5**	**2.3**	**0.4**	0.0	43
3443	52800	**15.5**	18.6	**17.4**	**27.0**	**19.0**	**3.3**	**0.2**	0.0	**104**
3489	83500	**7.4**	-0.6	-6.0	**13.2**	**11.4**	**1.4**	**0.4**	**0.34**	33
3445	89600	**12.4**	**17.9**	7.4	**24.7**	**15.7**	**2.1**	**0.5**	**0.14**	**82**

**the cells filled with grey color and large bold font correspond to statistically significant trends. NA means that river is completely frozen at this month.*

Monthly precipitation sums analysis for 8 weather stations in the region, located at altitudes from 137 to 1 288 m, 1966 – 2012, has shown no evidence of a systematic positive trend. On the contrary, at 4 weather stations in the Indigirka river basin (out of total 8 studied stations) a statistically significant precipitation sums decrease has been observed during winter months, especially in January and February – by up to 69-121%, what accounts for 3.3-6.7 mm a month or up to 11 mm in total for first two winter months. Precipitation totals negative trends at the rest of the stations are unreliable. In March and April precipitation totals trends are also negative, but unreliable. In May, a slight positive (up to 4 mm), but statistically unreliable, precipitation totals trend was identified.

Annual air temperature for the last 49 years (1966-2015) has increased on average by 2.1 °C. Air temperature rise in the Indigirka river basin was more significant than for the Yana river basin. During the period from April to July, there is a statistically valid air temperature trend at most of the weather stations of the region; in May and June average trends values account for 2.9 and 1.8 °C correspondingly (Table 3). Positive winter air temperature trends (up to 6.3 °C) are significant at 3 weather stations in October, December and January, at 4 stations – in November, at one station – in February, and at 5 stations – in March.

Air temperature increase in the last decades has led to a significant flood starting dates shift towards earlier dates. Flood starts 4-8 days earlier (the trends are statistically significant) in 8 rivers with the identified runoff changes in May, as well as in 3 rivers, for which statistically insignificant monthly runoff increase in May is identified; two of these rivers have catchment areas smaller than 1 000 km^2 (Table 2). In small rivers, flood starts in the second decade of May (May 11-18), for the large rivers this date shifts to the middle of the third decade (May 25, on average).

5. CONCLUSION

The performed analysis of the runoff data for the Yana and Indigirka river basins has shown statistically significant positive discharge trends in May and autumn-winter period for the last decades (accompanied with significant warming). Such trends exist mainly for medium and large rivers with the catchment areas over 7 000 km^2. At the same time, negative winter precipitation trends at some weather stations over the studied region have been identified, as well as no statistically significant precipitation trends in other seasons. Average annual air temperature increases at all the meteorological stations of the region by 1.1-3.1 °C.

An annual average discharge increase was identified only for three rivers, what, as we assume, can be mostly caused by typical geotectonic conditions of the runoff formation processes (riv. Adycha and Elgi) or by local features (the Dunai stream). Arjakova (2001) emphasizes that riv. Elgi (the Indigirka river basin) and riv. Adycha (the Yana river basin) cross tectonic dislocations zones. Taliks, which form within faults and excessive jointing zones, supply rivers even in extreme winter conditions (Romanovsky 1983; Piguzova 1989). Correlation between winter runoff and sub-permafrost waters is also confirmed by Glotov et al. (2011), who demonstrated that in particular years, the Kolyma river winter runoff losses at cross sections downstream in comparison with cross sections upstream depend on the current extensive extensions and compressions of underflow through talik space during sublittoral seismic activities periods.

Since there is no research data on dynamics of the water-bearing horizons capacity in the Yana and Indigirka river basins, it does not seem possible to clearly identify the reasons for runoff changes during autumn-winter period. Gurevich (2010) has substantiated the hypothesis on a regulatory impact of ice cover in the regions with long-lasting winter on ground water feeding into rivers. Subsurface ice melting along with air temperature increase, widespread over the studied basins (Brown et al. 1998), can contribute to the streamflow increase (Resources… 1972; Frey & Mclelland 2009).

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IMPACT OF CLIMATE WARMING AND GREENHOUSE GAS EMISSIONS ON H^+ MIGRATION IN THE BIOSPHERE OF YAKUTIA

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ABSTRACT
This paper discusses the results of long-term monitoring of precipitation acidity in central and eastern Yakutia. The increases in atmospheric carbon dioxide concentrations and in mean annual air temperature are accompanied by increasing acidity of precipitation. The increased atmospheric deposition of H+ affects the acidity of soils and lake water. Snowcover acidity has been steadily rising in the taiga and mountain landscapes of eastern Yakutia since the late 20th century. If the current trend of precipitation acidity continues, by the year 2020 the snow pH is projected to lower to the levels that can adversely affect the biosphere.

KEYWORDS
Climate; greenhouse gases; precipitation; acidity; Yakutia

1. INTRODUCTION
With global climate change and increasing atmospheric concentration of carbon dioxide, it becomes increasingly important to understand the migration of H^+ (pH) in various environmental compartments.
Human activities, primarily burning of fossil fuels, emit greenhouse gases. They affect the energy balance of the Earth, which in turn can affect temperature. The observed warming is thought by most scientists to have been caused by the increasing concentrations of carbon dioxide in the atmosphere which has led to a 0.6° C rise of global-mean air temperature.
Regional changes in climate (air temperature) and atmospheric greenhouse gas concentrations are also evident.

2. RESULTS AND DISCUSSION
Monitoring of atmospheric greenhouse gases conducted by N.F. Fedoseyev, Melnikov Permafrost Institute, near Yakutsk since 1997 [Kim et al. 2016] shows a steady increase in carbon dioxide concentration in the near-surface atmosphere from 363 ppm in 1997 to 398 ppp in 2011, with a stable positive trend of about 2 ppm/year (Figure 1).
Over the last three decades, there has been a marked increase in air temperature in Yakutsk. According to Skachkov (2017), the mean annual air temperature has increased by 0.7°C during this period (Figure 2).
The migration of H^+ (pH) in precipitation, lake water and soil was studied near Yakutsk and in eastern Yakutia as part of the long-term eco-geochemical monitoring program (Makarov 2006, 2007). The objective of the program was to obtain a comprehensive geochemical assessment of the pathway and recipient environmental compartments, including the atmosphere, snow, soil, surface water and groundwater.
Of natural agents, CO_2 has the strongest effect on the pH value, because the concentration of hydrogen ions in water depends on the amount of carbon dioxide. All other things being equal, the higher the CO_2 concentration, the higher the hydrogen ion concentration and the lower the pH.

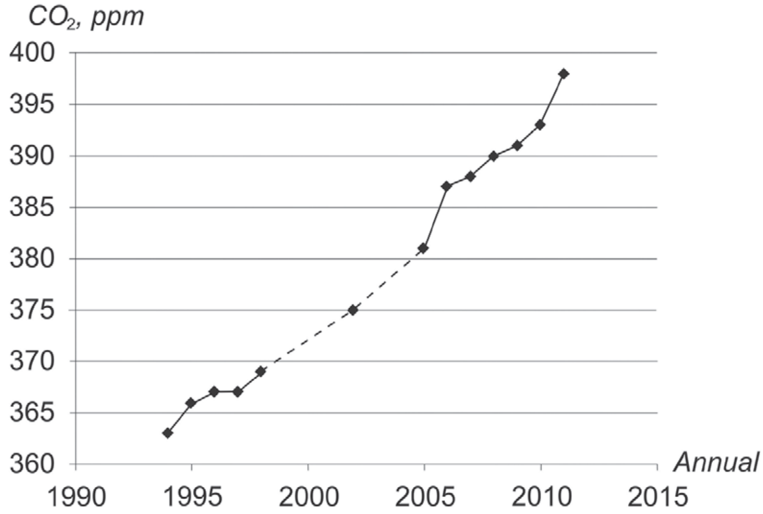

Figure 1. Average annual concentration of CO2 in the atmosphere in the vicinity of Yakutsk [Kim et al. 2016].

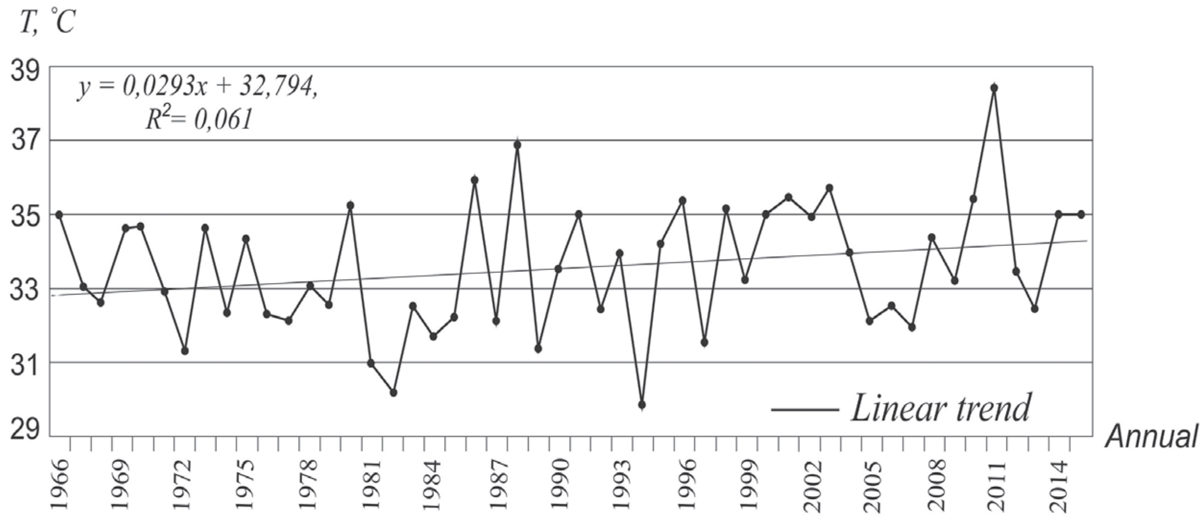

Figure 2. Long-term variability of mean annual air temperature [Skachkov 2017].

The relationship between the concentrations of hydrogen ion and other ions in precipitation (neglecting the contribution of weak organic acids) is expressed as:

$$(H^+) = 2(SO_4^{2-})+(NO_3^-)+(Cl^-)-2(Ca^{2+})-(K^+)-2(Mg^{2+})-(Na^+)-(NH_4^+)+2.5\cdot10^{-6}.$$

The residual term in this expression, $2.5\cdot10^{-6}$ mol/l, is the H+ concentration in equilibrium aqueous solution (pH = 5.6) at average atmospheric carbon dioxide concentration of 330 ppm at temperature of 20°C. This value is normally characteristic of unpolluted precipitation.
The long-term monitoring of precipitation in Yakutsk allowed us to capture temporal variations in H+ concentration (pH) in precipitation. The atmospheric concentration of carbon dioxide was found to steadily increase over 20-25 years since the end of the 20th century (Makarov 2007), which was accompanied by a reduction of precipitation pH and a change from slightly alkaline to slightly acidic (pH_{1989} = 8.0; pH_{2012} = 6.66). The deposition of H+ has increased almost seven-fold between 2000 and 2012: H^+_{2000} = 0.14; H^+_{2012} = 0.97 g/m^2 /yr (Figure 3).

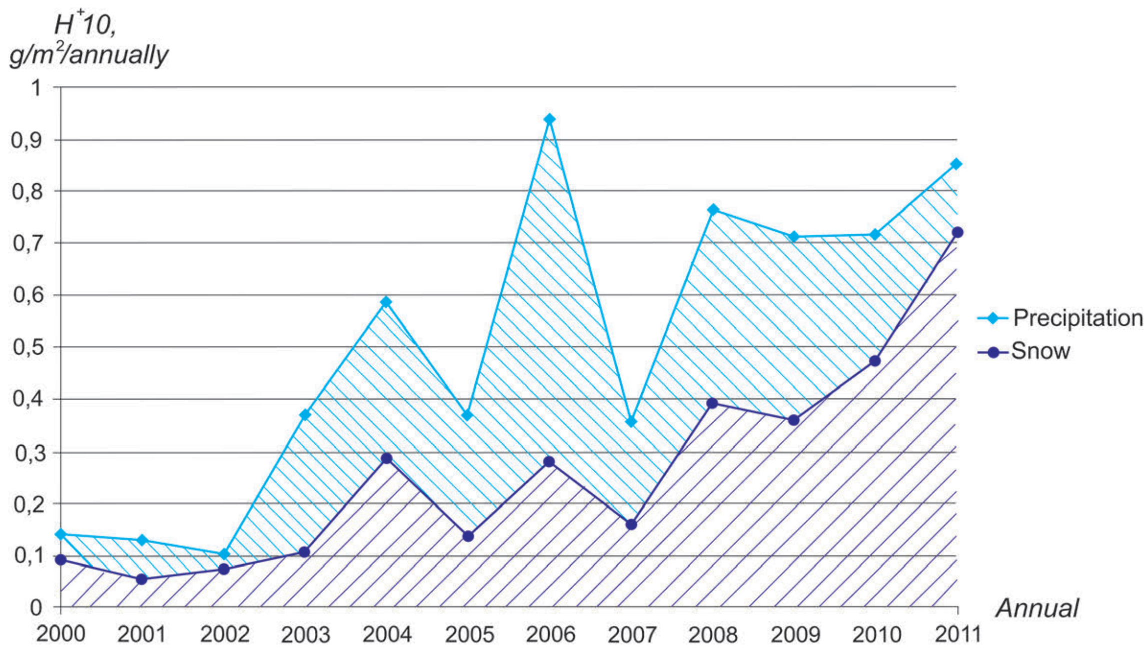

Figure 3. Variation in wet deposition of H^+.

The increased deposition of H^+ cation causes an increase in acidity of the urban soils.
Our observations indicate that the average pH of soils in Yakutsk changed from alkaline values – 8.10 in 1982-1984 to 7.51-7.56 in 2008-2011. The area of acidic soils with pH<7 increased from 0-2% in 1982-1984 to 8% in 2008 and 10% in 2011. Strongly alkaline soils with pH levels above 8.5 have almost disappeared in Yakutsk, and in some areas of the city the soils have pH values 6.8 to 5.6 (Figure 4).

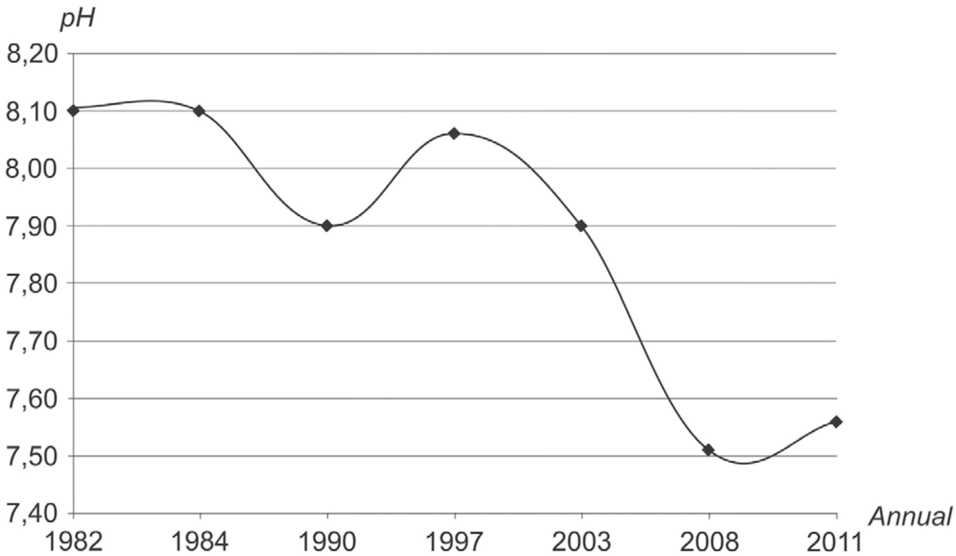

Figure 4. Variation of soil pH in Yakutsk.

Under the influence of acidifying agents, the soils become depleted of exchangeable bases and the weak natural acids (bicarbonates) are displaced with anthropogenic sulfates and nitrogen compounds reducing the pH of water. Acidification is greatest near the soil surface soils, with no change in acidity at the depths of 1-2 m in the active layer of permafrost.

Along with wet deposition and dry absorption of anthropogenic acids, catchment geology and its buffering capacity are known to influence the process of water acidification. Moiseenko (2001) believes that the geochemical structure of a catchment plays a primary role in water acidification. Under the influence of the acid is depleted the soil active layer of the exchangeable bases, a decrease in their income from the catchment area, the displacement of weak natural acids (bicarbonates) anthropogenic sulfates and nitrogen compounds and, consequently, reducing the pH of the water.

The increased atmospheric deposition of the H^+ cation and the increased soil acidity have caused changes to the acidity of lake water in the city of Yakutsk. The average pH of the lakes lowered from 8.1-8.8 in 1986-2003 to 7.6-7.8 in 2009-2012 (Figure 5).

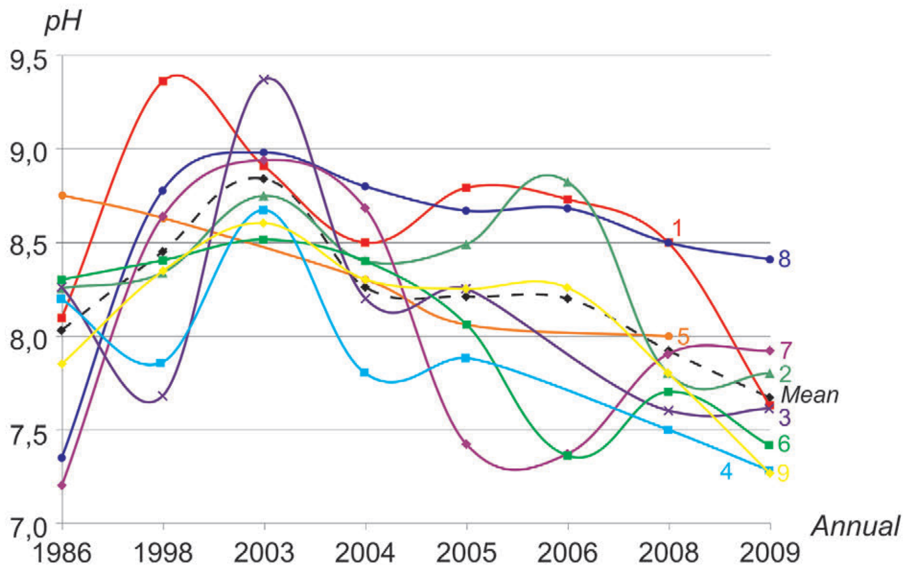

Figure 5. Variations of lake-water pH in Yakutsk.
Lake: 1 – Saysary; 2 - Teploe; 3 - Taloe; 4 – Khomustakh; 5 – Krugloe, 6 – Sergelyakh; 7 - Ytyk-Kyuyel; 8 –Beloe; 9 – Khatyn-Yurekh.

In Yakutsk, the trend of increasing acidity in the environmental compartments over the 30-year period is 0.3 pH/yr for precipitation, 0.2 pH/yr for soils, and 0.1 pH/yr for lake water.

Monitoring observations of snow chemistry in the areas of latitudinal (middle taiga) and altitudinal (mountain-desert, mountain-tundra and mountain woodland) permafrost landscapes in eastern Yakutia show a steady decrease in pH (Figure 6).

In mountain rivers of eastern Yakutia, spring floods caused by rainfall and snowmelt comprise 33-50% of the annual discharge. Most of the rivers in the region have a mixed rainfall and snowmelt hydrological regime, with rainfall being a dominant runoff source (Table 1).

In eastern and southern Yakutia, water in most rivers that drain magmatic or terrigenous rocks is acidic (pH =6.0-6.5). Where sulfidized deposits are widespread in the basin, e.g., in the Upper Burgali area, water is strongly acidic (Table 2).

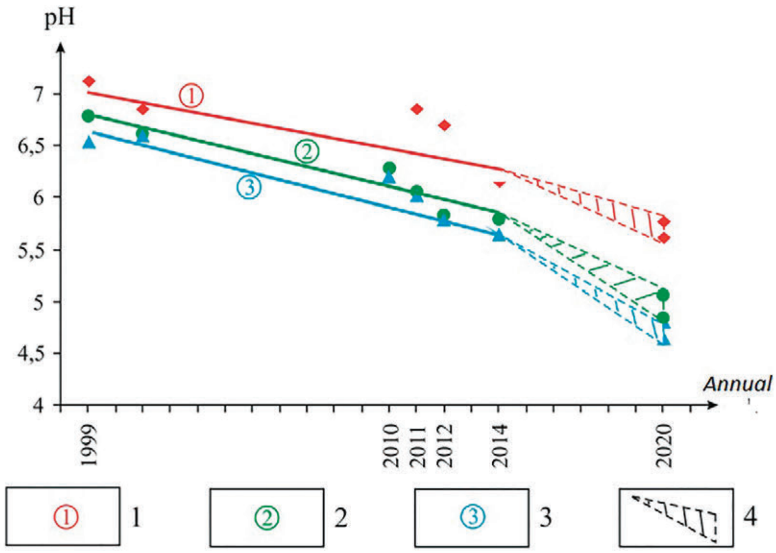

Figure 6. Observed variations of snow pH in permafrost landscapes, eastern Yakutia and predictions up to year 2020.
1 –Yakutsk; 2 – middle taiga, 3 – mountain landscapes; 4 – prediction to 2020.

Table 1.
Seasonal flow and snowmelt contribution to rivers in eastern Yakutia (Protasyeva 1972; Korzhuev 1965).

River	Gage	Seasonal flow,% of annual total				Snowmelt contribution,%
		Spring V-VI	Summer VII-VIII	Autumn IX	Winter X-IV	
Aldan	Tommot	52.9	21.4	10.9	14.8	35.0
Yana	Dzangky	40.1	47.4	10.8	1.7	14.3
Maya	Chabda	49.9	27.3	11.6	11.2	-
Adycha	Kurdyuk-Kumakh	40.7	46.6	10.5	2.2	-
Indigirka	Vorontsovo	31.0	52.2	11.6	4.2	16.9
Nera	Andagachan	46.8	43.9	7.9	1.4	-
Kolyma	Ust-Srednekan	48.2	35.9	11.0	4.9	30.4

Table 2.
Acidity of water in the mountain rivers of eastern and southern Yakutia.

River	Date	pH
Indigirka (Indigirsky)	04.06.1987	6.5
Timpton (Ust-Baralas)	11.06.1967	6.0
Ebithia (Ebetem)	26.06.1967	6.4
Upper Burgali	19.08.2012	4.2
Upper Burgali	12.08.2008	6.2
Kondekan	01.08.1975	6.0
Dhzarbang	17.08.1970	6.6
Middle Adycha	21.06.1973	6.0
Elga	20.06.1970	6.0
Inyali	12.06.1974	6.5
Lower Maya	08.07.1959	6.3
Lower Yudoma	07.08.1959	6.2

Spring snowmelt and input of acidic snowmelt water can significantly reduce the acidity of river water during this period.

Possible adverse effects may be associated with the occurrence of acidic geochemical barriers in areas of higher groundwater acidity. Moreover, the acidic barrier may be present under alkaline conditions on transition from strongly alkaline to slightly alkaline environment. The acidic barrier concentrates the anionic elements, such as silica, molybdenum, and selenium, the mobility of which is reduced in an acidic environment. Elements which readily form soluble anionic and carbonate complexes in an alkaline environment (B, V, Mo, As, S, and Cr) can significantly extend the range of migration.

Most plants prefer pH near neutral and further increase in acidity may lead to degradation of vegetation.

A study by Baker & Harvey (1985) on the relationship between lake water pH and fish population status showed that for lakes with pH of 5.6 or higher the probability of a viable fish population is high, while for lakes with pH <5.6 the probability is lower.

3. CONCLUSIONS

Long-term observations of snow pH in permafrost landscapes of eastern Yakutia indicate increasing acidity of precipitation.

Several authors (e.g., Balobaev 2009; Neradovsky & Skachkov 2011) predict that the period of a warmer climate will probably end by about 2015-2020. If we accept that the climate warming and anomalously high CO_2 emission will continue for another 10 years, and the 1999-2015 trend of increasing acidity of winter precipitation persists, snowcover pH in permafrost landscapes of eastern Yakutia will be reduced to 5.25-5.26 by 2020-2025 or to 4.70-4.92 based on the 2010-2012 data (Table 3).

Table 3.

Variation of snowcover pH in permafrost landscapes in eastern Yakutia and forecast to 2020-2025.

Year	Permafrost landscapes, elevation, m asl	
	Middle Taiga, *100-400*	Mountain, *400-1200*
1999	6.78	6.58
2001	6.62	6.61
2010	6.28	6.17
2011	6.06	5.85
2012	5.83	5.76
2014	5.45	5,69
2015	5,20	5,32
Trend pH/year	-0.099	-0.079
Prediction for 2020-2025		
1999-2015	4.70	4.92

This will lead to a significant increase in river water acidity during floods when the river chemistry is dominated by snowmelt. This may have adverse effects on fish communities.

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SCALING ISSUES RELATED TO SNOW STORAGE AND MEASUREMENTS

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ABSTRACT

A method for the design of snow measurement systems has been applied to several catchments in Norway over the last 14 years, with the objective of establishing an effective snow measurement system. The method studied here is based on GIS analysis of grid maps (raster) of the terrain as well as other characteristics in each catchment. The method is used to establish representative snow measurement courses which have characteristics that are comparable to the characteristics of the catchment. This will lead to determining representative snow data for the whole catchment. The following seven terrain parameters were defined as relevant for the comparison of catchment and snow courses: elevation, slope, aspect, curvature, location (x- and y coordinates), and vegetation.

Resolution, as a measure of scale, has a potentially large influence on the measurement of snow storage, especially for the terrain parameters derived from the digital elevation model (aspect, slope and curvature). Different resolutions will reflect different snow storage and distribution mechanisms, taking place at micro-, meso- and macroscales with different precision. Previously the available geographical information was rather coarse, but in later years digital elevation models with finer resolution have become available.

In the presented work, the relationship between terrain characteristics and snow depth was investigated at grid cell resolutions of 2, 10, 50 and 100 meters. The snow data were collected in the Swedish basin Överuman, during a field campaign in March 2017. The snow measuring system was designed with the method described above, and measurements were taken with snowradar (georadar). The total length of the measured snow courses was 74.2 kilometres.

Most of the seven terrain parameters showed a significant, but weak to moderate relationship to the snow depth. An exception is seen at high elevation in forest regions, which demonstrated a strong relationship.

KEYWORDS

Snow measurements, snow measurement system, snow distribution, snow depth, terrain characteristics, terrain parameters, DTM, snow radar, Överuman

1. INTRODUCTION

Snow storage is a very important issue for hydropower production and flood forecasting in many countries. In the described work the relationship of snow depth data to terrain characteristics was analyzed at different grid scales. It would be valuable if a relationship could be established to predict the snow depth based on terrain characteristics and a few calibration measurements.

Marchand and Killingtveit, 2005, investigated the topic at grid scales from 25 to 1000 meters. In that study only a weak relationship could be established.

Due to advances in LIDAR technology as well as more extensive use, better terrain models have become available. In the area of the investigated catchment, Överuman, a digital elevation model (DEM) is available at a 2 meter grid cell size. The map in Figure 1 shows the location of the Överuman catchment within Scandinavia.

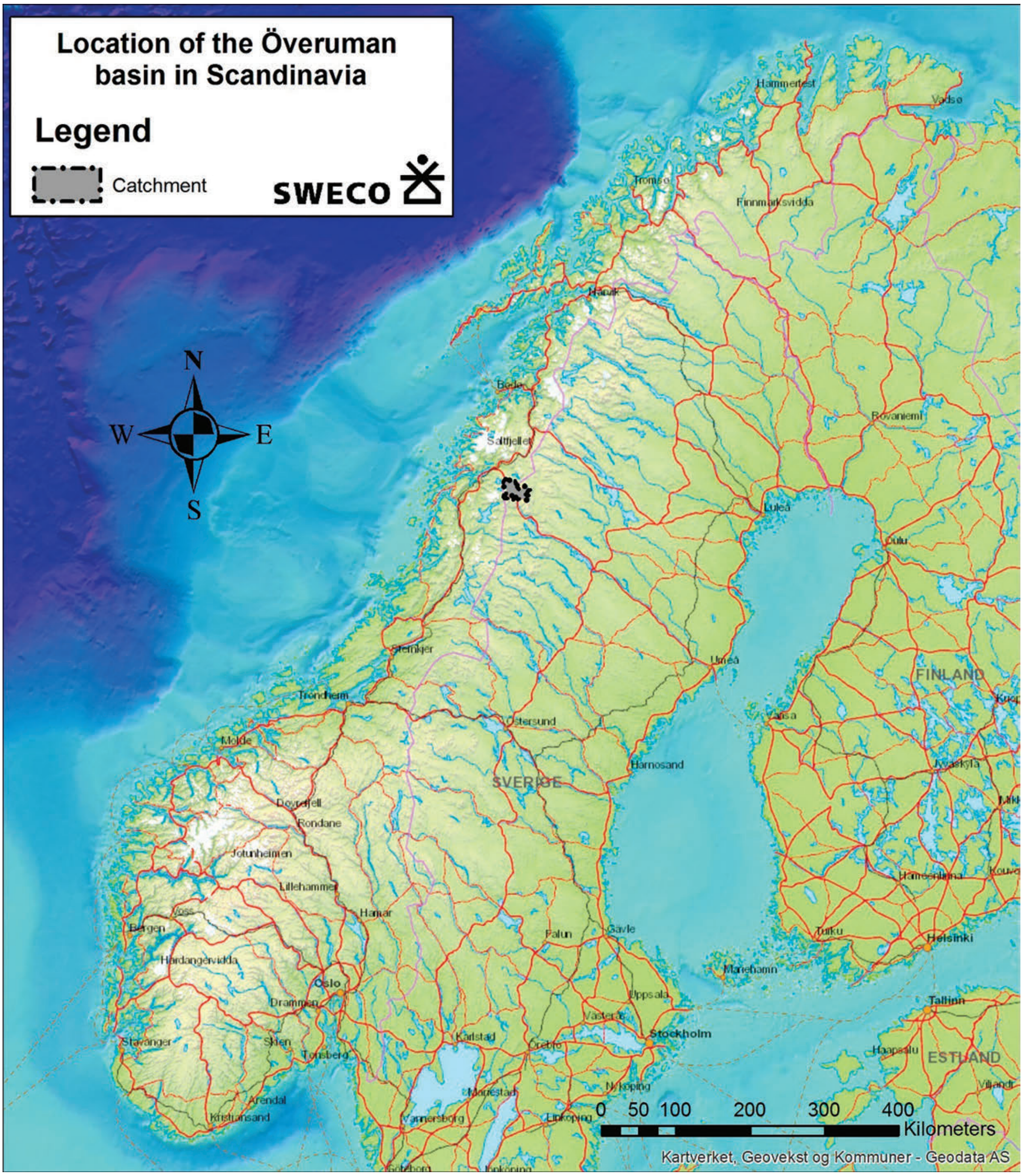

Figure 1. Overview of the location of the Överuman basin in Scandinavia

The majority of the catchment area is located in Sweden, with the remaining portion in Norway. The outflow from the catchment drains into the river Umeälv, which runs through Sweden to the Baltic Sea. The catchment has a size of 652 km^2. The elevation ranges from 524 to 1575 masl, and large parts of the area have alpine character with little vegetation.

1.1. Snow measurement system

To improve hydropower production planning, a snow measurement system was developed for the Överuman catchment during the winter of 2016-2017. Data collection using the new measurement system was performed for the first time in March 2017.

The theory behind the development of the measuring system is described in detail in Marchand, 2003. In short, snow measurement courses are distributed over the whole catchment in a way that snow distribution variations are accounted for at different scales: micro-, meso- and macroscale. This is obtained through comparing the following seven terrain characteristic parameters, elevation, X and Y location, aspect, curvature, slope, and vegetation (forest or no forest) for the whole catchment, with the same parameters at the snow courses. Based on an iterative process and statistical analyses the locations of the snow courses are adjusted until a satisfactory fit between catchment and snow courses is obtained. For Överuman, the resulting 8 snow courses have a total length of 74.2 km and are located as shown in the map in Figure 2.

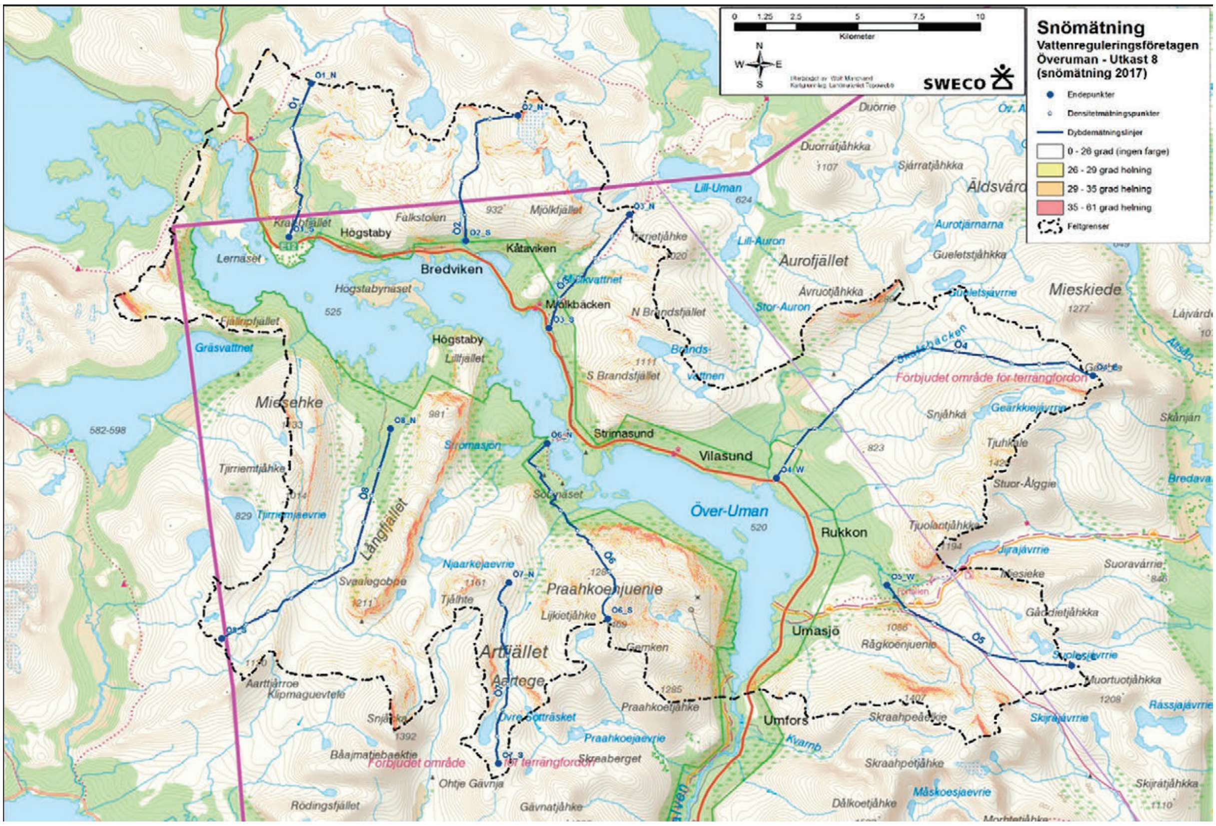

Figure 2. Map of the Överuman catchment and the resulting snow courses (blue lines). Steep slopes are indicated in the map with yellow, orange and red color, for snow avalanche awareness.

The obtained fit between catchment and snow courses is illustrated in Figure 3.

1.2 Data collection

Data collection was performed by pulling a snowradar (georadar) behind a snowmobile. The radar device measures the two-way travel time (TWT) of the radar signal. This is the time for the signal to travel from the emitting antenna to the ground and back to the receiving antenna. Radar measurements are taken with a fine sampling rate along the driving track. Calibration measurements for snow depth were taken with traditional rods. Additionally, many bulk snow density samples were collected at given locations along the snow courses. Density measurements are required for conversion of snow depth to snow water equivalent (SWE), but in this study only snow depth is considered. The resulting radar data were post-processed using custom-made software, Sirdas.net (Albrektsen Consulting AS). With minimal input from the user Sirdas can determine the snow ground interface. Resultant data files of location and snow depth for each

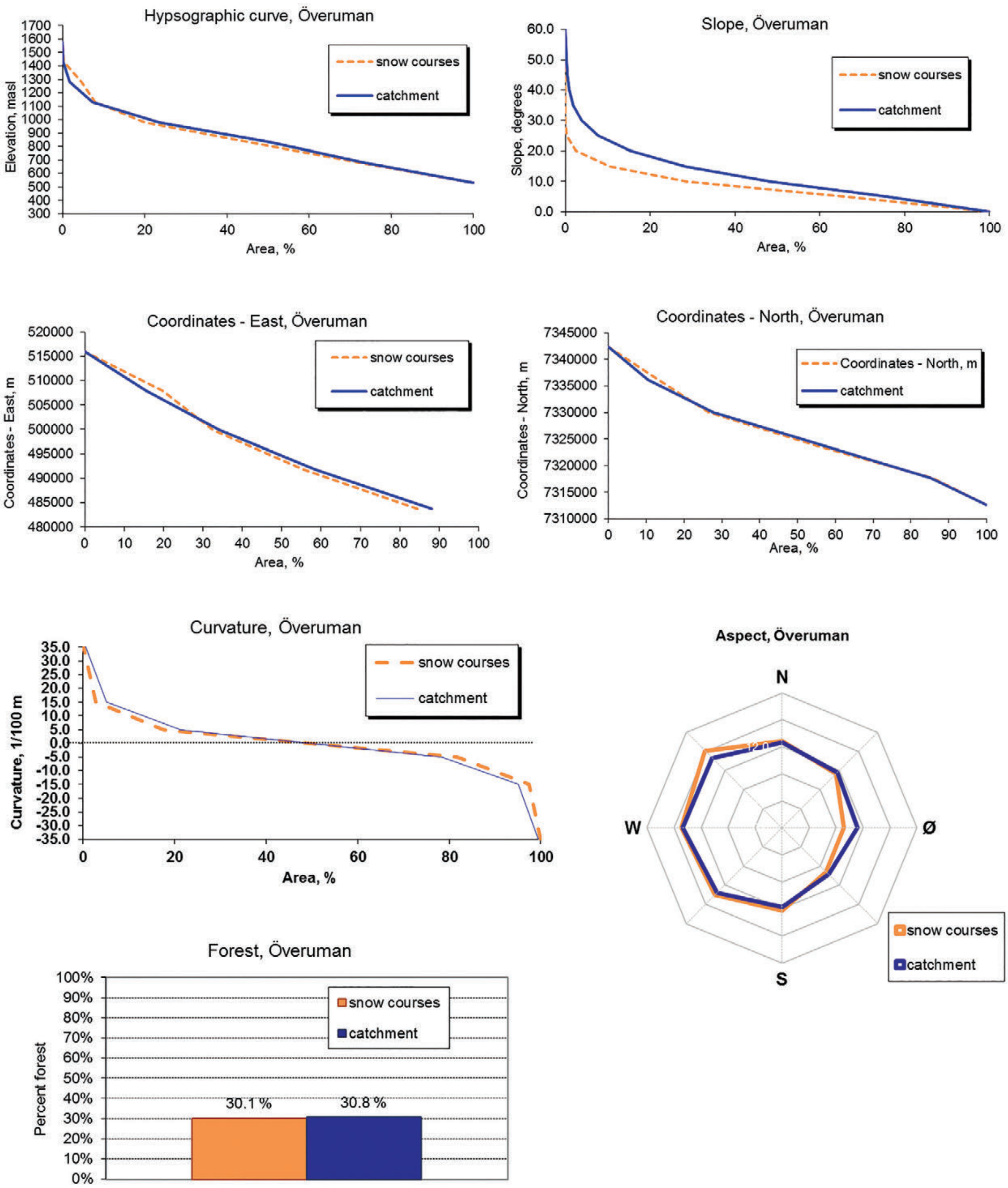

Figure 3. Obtained fit between terrain characteristics of the snow courses with the ones from the entire catchment

radar reading are then produced. The mean snow depth in the catchment was approximately 179 cm, with values ranging from 0 to 923 cm.

2. METHODS

In approaching the goal of the study, the investigation of the relationship of snow depth to terrain parameters at different scales, several analyses were performed:

First, Arc GIS (ESRI) analysis was performed. The 2 m grid DEM and the accompanying six terrain characteristics from the establishment of the snow measurement system were used as a starting point. Snow depth data with coordinates for every measurement were imported into the GIS. Next, the snow depth values were converted to a 2 m grid, wherein every grid cell that contained snow samples was assigned the mean snow depth of the values in the cell. Then, each snow depth cell was converted to a point, representing the geographic mid-point of the cell, with the snow depth value. This point dataset was then matched with the seven grid cell layers. Based on that the attribute table of the point dataset with snow depth received seven more columns, one for each of the associated terrain parameters at the specific location of the snow depth sample.

The generated attribute table was then imported into Excel for further processing and statistical analyses. The first step was to transform the aspect values, since the circular character of this parameter is difficult to handle otherwise. Aspect values were divided into North-South and East-West directions by assigning values from 0 to 1 depending on how much a grid cell is facing in either of the two directions. This process increased the number of parameters from 7 to 8.

Secondly, snow samples in forest were separated from those in open field since snow distribution in these two terrain types typically is very different, see Marchand and Killingtveit, 2004. In creating two data sets the number of parameters was reduced for each dataset. The two obtained datasets were then used in correlation and multiple regression analyses.

Furthermore, using the 2 m DEM, grid layers for all terrain parameters were generated at 10, 50 and 100 m grid cell size. The procedure described above was then repeated for each of these additional three scales.

3. RESULTS AND DISCUSSION

All the major results from the described analyses are presented in Table 1 and Table 2. The correlation analyses show that most of the terrain parameters have a significant correlation to snow depth. An exception is the parameter aspect N-S, which is either not significant or just above the threshold at all grid scales for both forest and open field. This might indicate that solar radiation induced snowmelt is not relevant at this time of the year for high latitudes. Conversely, a much stronger correlation seen between aspect E-W and snow depth in open field could be caused by the predominant westerly wind direction. Curvature has a generally weak correlation to snow depth. An exception is in the 10 m scale, where the correlation is much stronger for both forest and open terrain. Considering the location of the snow samples, expressed by the x and y coordinates, it is obvious from the results that the east-west direction is much more correlated to snow depth than the north-south direction. This corresponds well with the precipitation gradient which varies a lot from west to east, both in Norway and Sweden. This phenomenon is related to westerly winds which bring precipitation to the coast of Norway. Precipitation typically decreases towards inland regions, such as those in Sweden. It is noteworthy that slope is correlated to snow depth, both for forest and open field, but the correlation strength is weaker at larger grid scales. By far the strongest correlation values can be found for the elevation parameter in forested areas, whereas this correlation is rather weak in open field. However, the latter correlation increases at larger grid scales. Increased precipitation with increasing elevation is a well-known relationship. This is confirmed by the results seen in forest regions. However, in the open areas a weak correlation is most likely caused by redistribution due to wind.

Table 1

Correlation factors between snow depth (sd) and terrain parameters at different grid scales

	2 m grid		10 m grid		50 m grid		100 m grid	
	sd. open	sd. forest	sd. open	sd. forest	sd. open	sd. forest	sd. open	sd. forest
aspect N-S	-0.01	-0.06	0.03	-0.02	0.03	0.02	0.02	0.05
aspect E-W	-0.21	-0.06	-0.29	-0.12	-0.25	-0.07	-0.27	-0.01
coord-x	-0.18	-0.18	-0.18	-0.20	-0.20	-0.22	-0.20	-0.29
coord-y	0.07	0.11	0.05	0.12	0.06	0.15	0.04	0.30
curvature	-0.07	-0.04	-0.21	-0.17	-0.03	0.00	-0.06	0.12
elevation	0.06	0.58	0.10	0.52	0.12	0.54	0.17	0.48
slope	0.17	0.16	0.17	0.15	0.13	0.06	0.13	0.04

green color = significant at the 5% level

The results from the multiple regression analyses are presented in Table 2. At the 2 m grid scale all terrain parameters are significant at the 5% level. However, the adjusted R-square value shows that the degree of explanation from the model is much higher in the forest, compared to that in open field. This is valid for all grid scales. In the forest, the regression model can explain from 47 to 42% of the variation in snow depth, with the highest adjusted R-square values for the lowest resolutions. This reverse proportional relationship between adjusted R-square and grid scale is not present in the results for open field. The parameters aspect N-S, curvature and slope are not significant predictors at the larger scales.

Table 2

Results from the multiple regression analyses of snow depth (sd.) and terrain parameters: P-values and regression statistics

	2 m grid		10 m grid		50 m grid		100 m grid	
	sd. open	sd. forest	sd. open	sd. forest	sd. open	sd. forest	sd. open	sd. forest
aspect N-S	0.000	0.000	0.260	0.000	0.445	0.030	0.274	0.188
aspect E-W	0.000	0.000	0.000	0.000	0.000	0.047	0.000	0.942
coord-x	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
coord-y	0.000	0.000	0.000	0.000	0.000	0.000	0.036	0.000
curvature	0.000	0.000	0.000	0.000	0.313	0.628	0.226	0.444
elevation	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
slope	0.000	0.000	0.000	0.000	0.004	0.994	0.387	0.719
green color = significant at the 5% level								
Multiple R	0.112	0.686	0.449	0.659	0.394	0.658	0.418	0.661
R^2	0.112	0.471	0.202	0.435	0.155	0.433	0.175	0.437
Adjusted R^2	**0.111**	**0.470**	**0.201**	**0.433**	**0.150**	**0.426**	**0.166**	**0.422**
Standard Error	136	72	123	70	118	61	111	47
Observations count	12183	6088	5851	2553	1287	560	650	277

The standard error of predicted snow depth values is largest for small scales. This is most likely caused by the large variations of snow depth at small scales, whereas the averaging of snow depth for the large grid cells smoothed the extreme values, resulting in a smaller standard error. The important role of the parameter elevation, especially in forest, is illustrated by the line fit plots in Figure 4. In open terrain, the result is quite different, as shown in Figure 5. Both figures indicate a large degree of linearity. However, for the highest elevation the pattern is different. It can be assumed that this is reflective of the fact that at the highest mountain tops much of the snow is removed by the wind. This means that the relationship is not linear and therefore difficult to explain with a linear model. One possible approach to overcome this could be to split the terrain into elevation zones and to apply different models. To do this the elevation at which

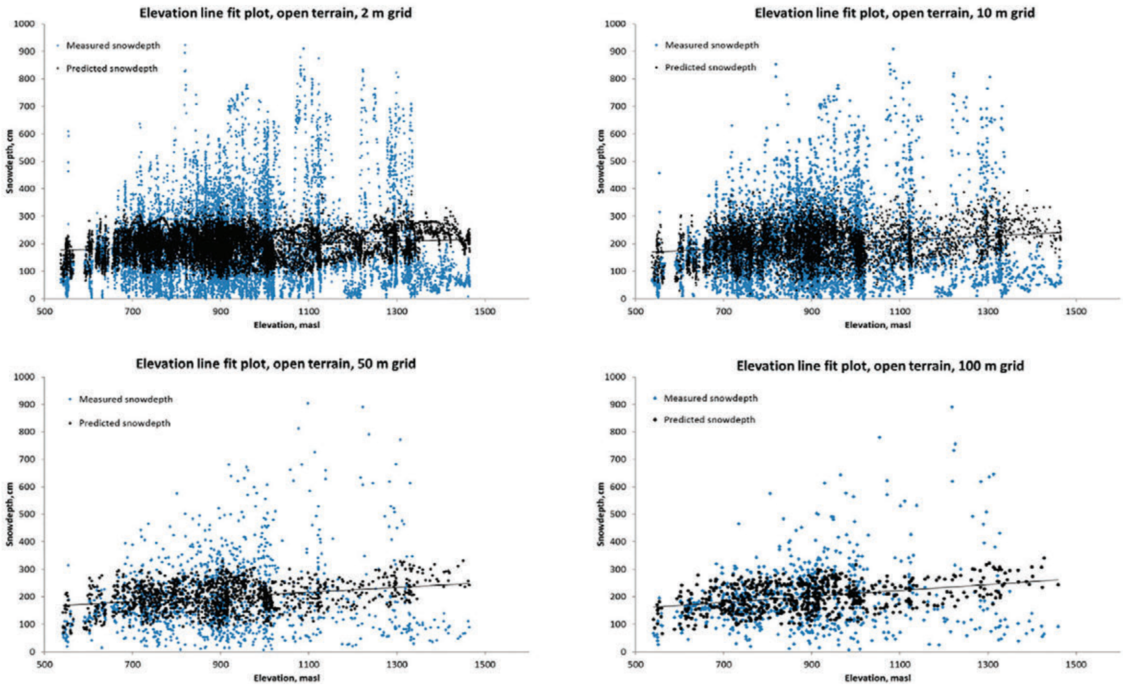

Figure 4. Resulting elevation line fit plot from the multiple regression in forest

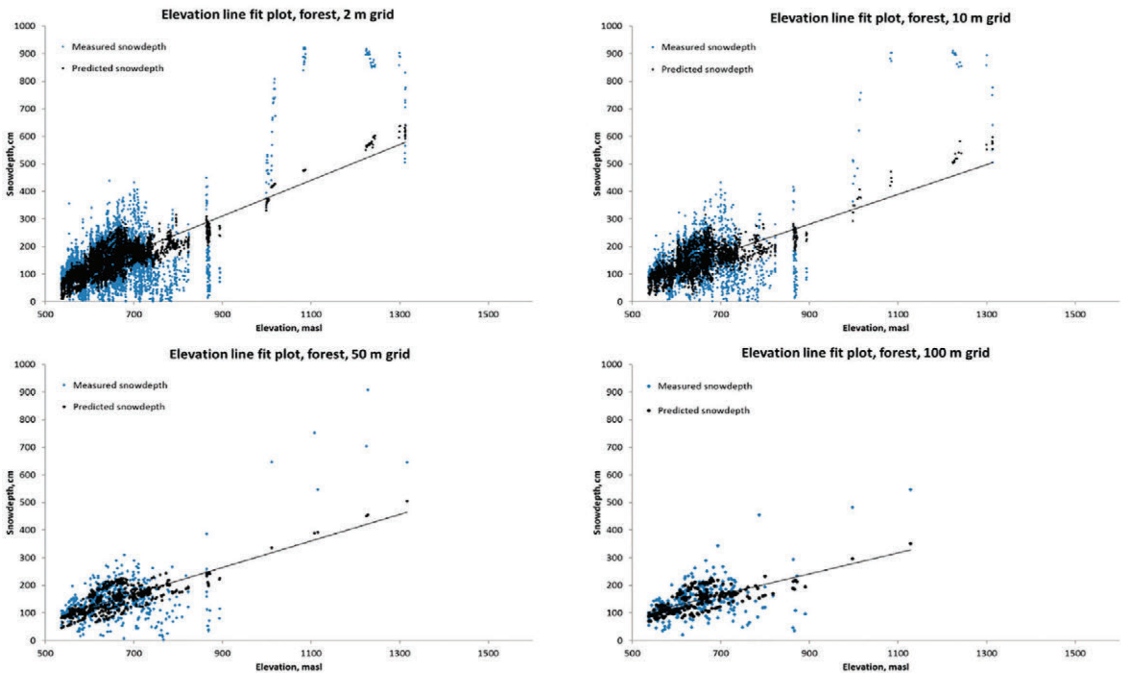

Figure 5. Resulting elevation line fit plots from the multiple regression in open terrain

its relationship with snow depth is not linear anymore must be determined. From the plots in Figure 4 and Figure 5 it seems that snow depth increases with elevation up to 1100 masl. At higher elevations the snow depth decreases. This feature is most visible at smaller grid scales and in open terrain. At larger grid cell sizes the slope of the line in the plots decreases, presumably as a result of the averaging effect when using mean values for larger grid cells. This removes the extreme snow depth values. This effect is clearest in the forested terrain where the snow distribution is more normal, versus those in the open terrain which are more skewed (Marchand and Killingtveit, 2004).

The gradient of snow depth in the x-coordinate, with decreasing depth from west to east, is visible in the line fit plots in Figure 6 and Figure 7. There is little variation in the slope of the line at the different grid scales.

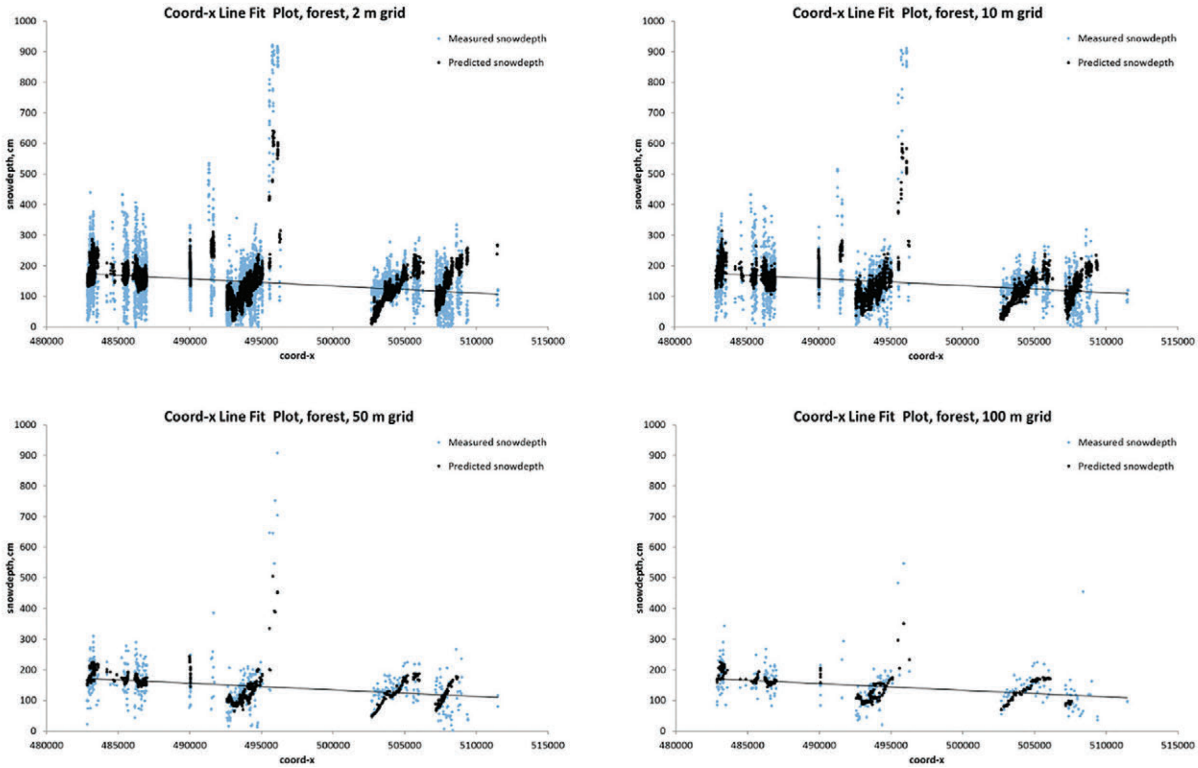

Figure 6. Resulting coordinate-x line fit plot from the multiple regression in forest

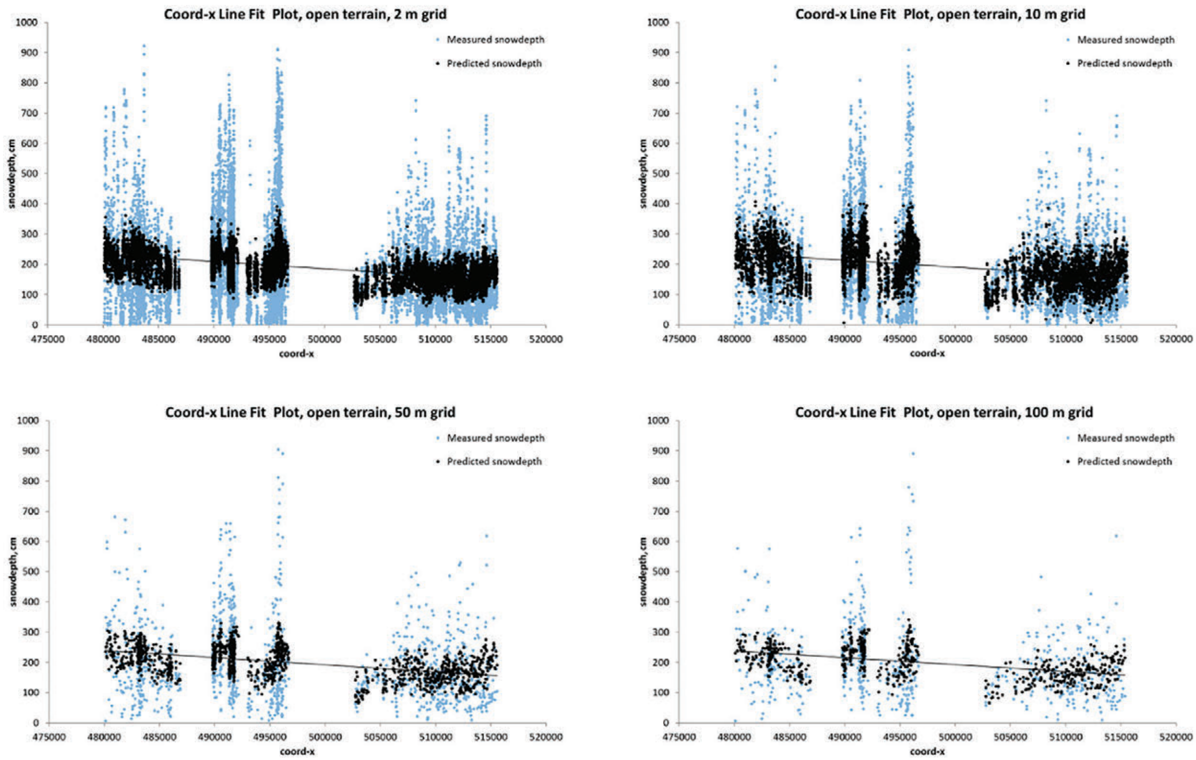

Figure 7. Resulting coordinate-x line fit plots from the multiple regression in open terrain

4. CONCLUSIONS

The relationship between terrain characteristics and snow depth was investigated at grid cell resolution of 2, 10, 50 and 100 m. The analysis was based on snow data, collected with snowradar (georadar) at snow courses, with a total length of 74.2 kilometres,

Most of the seven terrain parameters investigated showed a significant, but weak to moderate relationship to the snow depth. An exception is elevation in forest, which had a strong relationship. However, it seems difficult to fit a linear model.

The error when predicting snow depth from terrain characteristics can be very large. Standard deviation is largest at small grid scales and in open terrain, whereas forested areas and larger grid scales produced a lower standard deviation.

5. ACKNOWLEDGEMENT

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HYDROLOGICAL PROCESSES AT THE SUNTAR-HAYATA RIDGE (EASTERN SIBERIA)

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ABSTRACT
The detailed studies of permafrost, hydrological and glaciological processes at the Suntar-Hayata Ridge started in 1957 year with in the Program of International Geophysical Year (1957-1958). We used the archives of Melnikov Permafrost Institute in Yakutsk for assessment of runoff formation processes and the factors affecting them in remote high-elevation permafrost area, and estimated the parameters of hydrological model based on those findings. At the first stage we conducted the simulation of individual processes and compared the results with the observations. Such, the comparison of simulated and observed values of height and snow cover indicates that the model correctly describes snowmelt processes. Also the results of soil thaw/freeze and evaporation for the Suntar-Hayata site were considered satisfactory. At the second stage we conducted runoff modelling with initially assessed parameters. The results of modeling are satisfactory. For vast regions of the world with sparse monitoring networks like much of Russia, detailed catchment data only exists in small-scale research basins. There are very few such catchments, especially in the high-mountain areas of Siberia. Process based model parameterizations that are confirmed and refined in this basins can be applied in hydrologically similar regions, for example, the regions with high rocks and glaciers.

KEYWORDS
Hydrological modeling, Suntar-Hayata ridge, parameterization, Hydrograph model, permafrost

1. INTRODUCTION
East Siberian mountain regions are specifically characterized by their unique diversity of landscapes, terrain, climatic conditions and subsequently by predominant hydrological processes.
The main objective of the research is to develop and test methods for modeling of hydrological processes in different mountain landscapes in permafrost region, taking into account explicit influence of variable states and water-and-thermal regime of active layer on the runoff formation processes, which can be used in the face of extreme lack of observation data.
While modeling characteristics of subsurface permafrost and runoff formation processes, the role of landscapes is often being ignored, though various studies have proven it as determining (Antipov et al. 2007).
Novelty of the research, based on using the *Hydrograph* hydrological model (Vinogradov & Vinogradova 2010), is presented by the fact that analysis and quantitative evaluation of the frozen ground water-and-thermal regime and runoff formation processes are suggested to be performed at scale, comparable to the scale of a process. First, a single soil column is examined, then the result is spread over an elementary slope or a typical landscape, after that – modeling for a small catchment and transferring parameters onto medium and large basins. At each of the stages modeling results are verified against corresponding observation data, received, for

example, at observation network and experimental sites. It allows to track adequacy of the estimated values in comparison with natural environment, as well as to transfer parameters estimations from the studied catchments to unstudied ones, which have similar characteristics. In the study were used the reports on the comprehensive observations at the high-altitude Suntar-Khayata station under the program of the International Geophysical Year from the archive of the SB RAS Melnikov Permafrost Institute in Yakutsk (Grave 1959; Grave & Koreisha 1957; Grave & Koreisha 1960; Koreisha 1963; Koreisha 1957).

2. SUBJECT OF THE RESEARCH

The catchment of a tributary of the river Indigirka – the Suntar river – is taken as the research subject, at the channel cross section nearby the Sakhariniya river outlet, which heads from the Suntar-Khayata Range, the catchment area is 7680 km2. The presence of special observation data conditions this choice: in the Suntar river basin during 1957-1959 period, under the program of the International Geophysical Year, the high-altitude Suntar-Khayata station was operating, where glaciological, geomorphological, geocryological and hydrological observations were carried out (Grave et al. 1964).

The territory of the study is situated in the region of continuous permafrost, its thickness under watersheds is 400-600 m, and under stream valleys – 200-300 m (Geocryology 1989).

High peaks of the Suntar-Khayata Range (Mus-Khaya town, 2959 m above sea level (m a.s.l.)) go together with their slopes being dissectioned with stream valleys. The average altitude of the Suntar river catchment is 1410 m a.s.l., peaking at 2794 m a.s.l.

The climate of the research region is extremely continental with altitudinal zonation and temperature inversions. Average annual temperature is -13,8 and -14,1 ºC ºC (in July +6,4 and +17,5 ºC, in January -28,0 and -39,6 ºC) at the stations of Suntar-Khayata (2068 m a.s.l.) for the period of 1957-1964 and Agayakan (776 m a.s.l.) for the period of 1957-2012 correspondingly. Snow cover at the range starts in September. In winter, there is no runoff. Maximum runoff occurs in summer months. Usually a spring flood begins in the third decade of May. In the Suntar river basin, which is made up of soil and subsoil with low infiltration capacity, the role of the ground runoff is insufficient (Grave et al. 1964).

Near the Suntar river head, the glacier №31 is located, its area is 8,48 km2, what accounts for 0,11% of the river's catchment area. The glacier's runoff into the study region can exceed 3,8% of the overall annual runoff and reach 6,1% of the runoff in July and August, as, for example, in the neighboring basin of the Agayakan river, where glaciers cover over 1,35% of the catchment (USSR 1966). In the last decades, a steady decreasing trend of the Suntar-Khayata Range glacierization is noticed (Lytkin 2016).

The region's distinguishing characteristic are aufeises, formed at mountain ranges, in submountain and intermountain troughs. In the Suntar river basin aufeises can cover up to 0,5%, and the runoff from them is around 7% of the annual runoff. The feeding from melting aufeises is the most significant in May-June (Sokolov 1975).

On the slopes of the Suntar-Khayata Range, perennial snow fields (USSR 1966) and rock glaciers are widespread. They, along with the ice of the active layer and summer atmosphere precipitation, may represent a significant source of the local rivers feeding, however in this respect they are studied poorly (Zhizhin et al. 2012; Lytkin 2016).

Water balance of the research territory is understudied. Estimating the water balance elements – precipitation and evaporation for high-mountain regions, the Yana and Indigirka rivers headstreams, published in references on hydrometeorology (Reference Book 1968; USSR 1966), – was performed based, among others, on observation data from the high altitude Suntar-

Khayata station, 1957-1964, and the rivers runoff, but they are highly controversial.

Based on the data (Reference Book 1968), dependencies between the precipitation increase and elevation for warm (May – August) and cold (September – April) periods of the year were developed. Annual precipitation at the Suntar-Khayata station exceeds the precipitation amount observed at foothills more than twofold. Precipitation gradient at the altitude range 777 to 1350 m a.s.l. is 7 mm (5-7 %) per 100 m, and at the altitude range 1350 to 2068 m a.s.l. it exceeds 35 mm (15-16%). Snow surveys data received by Grave and Koreisha (Koreisha 1963) had demonstrated that, on average, during 1957-1959 period between altitudes of 2068-2257 and 2257-2477 m a.s.l. altitudinal gradients of precipitation increase are steady and equal to 35 (5-8%) and 30 (4-5%) mm per 100 m correspondingly. Solid precipitation depth at 777 m a.s.l. is approximately 25% of the annual total, and at 2068 m a.s.l. – approximately 60%. With linear extrapolation based on these two points, the percentage of the solid precipitation at 2900 m a.s.l. reaches up to 90% of the annual total (Grave 1960).

Mean annual precipitation sum from 1957 to 1964 at the Suntar-Khayata station according to the rain gauge data is 555 mm. However, in (Reference Book 1968) some adjustments are recommended for wind underestimation and wetting loss, which can reach up to 1,7 times (1,6 on average) for solid precipitation, and 1,3 times(1,16 times on average) for liquid precipitation, what leads to the annual precipitation amount of 688 mm at 2068 m a.s.l. (Reference Book 1968), and 800 mm at the ranges' peaks (Vasiliev & Torgovkin 2002; Hydrological Yearbook 1983).

Average annual runoff depth for the Suntar river at the channel cross-section of the Sakharinya river's mouth (catchment average altitude of 1410 m a.s.l.), 1957-1964, was accounted for 180 mm, and for the entire observation period until 2014 – 186 mm, maximum recorded discharge – 1910 m^3/s. During the same time periods, runoff of the Sakharinya river – a small tributary of the Suntar river (catchment average altitude of 1110 m a.s.l.), reached 71 and 98 mm correspondingly. While evaluating the actual runoff depth value, it is necessary to take into consideration such phenomenon as water drainage in mountain rivers' beds peculiar to this region and noted, for example, in the description of the Sakharinya river channel cross section (USSR 1966). Glotova and Glotov (Glotova & Glotov 2015) also point out that one of the important factors of the flood runoff formation in Northeast Russia is seepage losses into alluvial deposits dried out in winter, capturing part of the snowmelt runoff.

Annual evaporation value of 200-250 mm for the studied region, published in (USSR 1966), apparently, is calculated as a difference between precipitation (with adjustments described above) and runoff values. The authors of the study consider such evaporation estimates, as well as adjustments for winter precipitation, as unreasonably high and unrealistic. For example, according to the observation data and modelling results at the streams, carried out at the Kolyma water-balance station, annual evaporation under similar, but less severe conditions, is less than 150 mm (Lebedeva et al. 2015).

3. PARAMETRIZATION OF THE HYDROLOGICAL MODEL FOR THE SUNTAR RIVER BASIN

Choice of the model of runoff formation processes *Hydrograph* (Vinogradov & Vinogradova 2010) as a research tool is conditioned by its several advantages. The model contains algorithms that describe dynamics of heat and moisture in soil profile, which includes the active layer, depending on their physical characteristics (Vinogradov et al. 2015). The model's robustness towards input meteorological information (air temperature and humidity, precipitation) enables to estimate runoff from poorly gauged basins.

Parametrization of the *Hydrograph* model was performed based on joint analysis of soil hydrothermal regime and consistent patterns of runoff formation in typical landscapes (Lebedeva et al. 2015). In order to perform modeling, the catchment was divided into runoff forming complexes (RFC), or primary types of landscapes. Properties of the model characterize RFC in general, are fixed within its bounds and change steeply at its borders (Vinogradov & Vinogradova 2010).

According to altitudinal zonation the catchment of the Suntar river is divided into 4 RFCs (Figure 1). For each RFC, a schematization of the elevation profile is developed that considers soil composition, vegetation, snow accumulation and runoff formation processes.

Goltsy complex is located at the altitude range 1900 to 2700 m a.s.l. (average height is 2040 m), its share accounts for 7% of the Suntar river catchment.

Soil profile of this complex consists of macro fragmental argillite broken stone (specific density 2920 kg/m^3) with admixed loam materials, cemented together with ice and with layers of clean ice up to 2 m depth. Vegetation is absent. Broken stone is presented in the form of glacial frost-split boulders as well as diluvia soil of the valley slopes with admixed loam material. Seasonal thaw depth of soils within the goltsy complex at elevation over 1700 m a.s.l. varies from 55 to 75 cm (Grave et al. 1964). For the upper layer of diluvia, insignificant humidity and its barely visible variations during the warm season is typical, despite significant amount of precipitation and its irregular distribution. It is explained with high permeability of broken rocks. Unevaporated moister easily infiltrates deep down and flows along the frozen bedrock. The unsatisfying experience of runoff site construction had shown that the bedrock is a complex surface for it has deep splits and hollows, and even though their temperature is below zero, they are not filled with ice (Grave 1959).

Alpine tundra belt is located within the altitudes of 1450-1900 m a.s.l. (1630 m on average) and takes 37 % of the basin. It is characterized by distribution of a tight and depressed layer of grass and moss with bushes. Upper 10 cm layer of the alpine tundra profile consists of moss cover, under which there is rock formation with some ice with admixtures. The seasonal thaw depth in alpine tundra is 80-90 cm.

Taiga (1100-1450 m a.s.l., 1310 m on average) consists of thin larch wood on the north slopes and dense larch forest on the south slopes, and covers 42% of the catchment.

Swamped sparse forest and meadow moors are typically situated within rivers valleys and floodplains (828-1100 m a.s.l., 1060 m on average) and covers 14%.

A distinguishing feature of larch taiga and floodplain on boggy soils is the presence of peat layer, which is located under the moss cover and is spread as deep as 20 and 40 cm deep correspondingly. This layer is characterized by low heat-conducting, increased porosity and water retaining capacity. Seasonal thaw depth changes within 85-150 cm in taiga and 30-115 cm in rivers valleys and floodplains.

For modeling the whole catchment is presented as a hexagonal grid, with 32 representational points (RP) as grid points. Three of these points match goltsy mountains, 11 – tundra, 13 and 5 – taiga and swamp sparse forest/meadow moors correspondingly.

Assuming that runoff formation processes in mountain regions of the Kolyma river headstreams and in the Suntar river basin are similar, parameters for RFC 2-4 were derived from the Lebedeva's et al. publication (Lebedeva et al. 2015), who determined them with use of the data from the Kolyma water-balance station. Parametrization of the goltsy landscape is examined more detailed in the following section.

3.1 Preparation of input meteorological data

Such data as daily air temperature and humidity, precipitation sums for each station within or nearby the catchment are set as input meteorological information for modeling. Data from 4 weather stations (Suntar-Khayata, Nizhnyaya Baza, Agayakan and Vostochnaya) were interpolated for each RP using triangulation method. While interpolating precipitation daily sums were normalized in accordance with the earlier created dependencies between both liquid and frozen precipitation amount and terrain elevation at meteorological stations. While interpolating temperature and saturation deficit, annual change in both values with the height was considered. Annual average monthly temperature and saturation deficit gradients, calculated with the data from the Suntar-Khayata (2068 m a.s.l.) and Agayakan (776 m a.s.l.) meteorological stations, change from +1,1 ºC and +0,01 mbar at 100 m in January to -1,3 ºC and -0,35 mbar at 100 m in June.

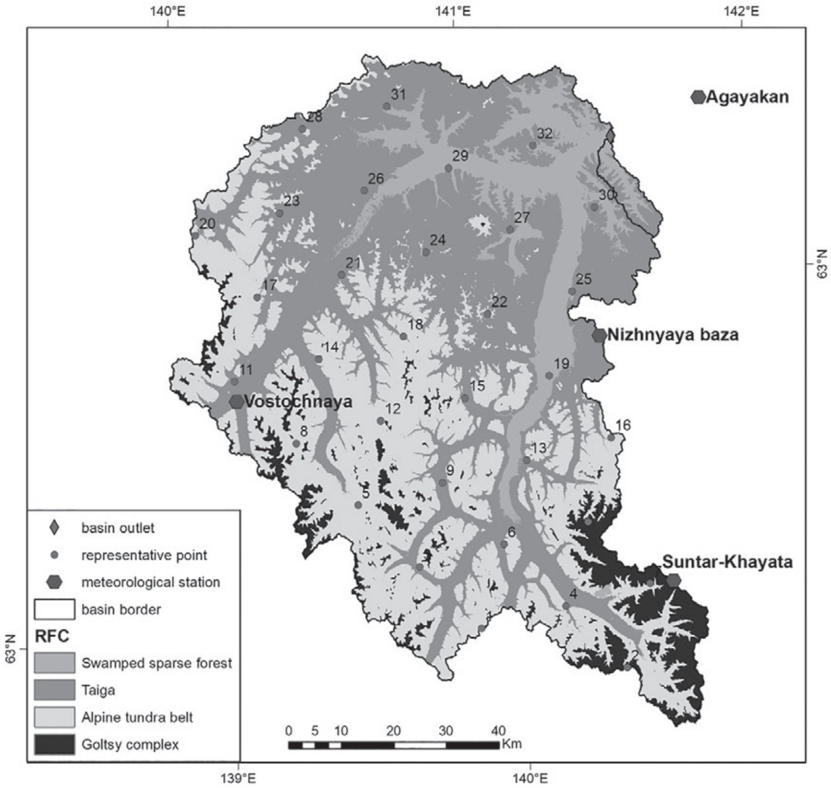

Figure 1. Study basin

4. MODELING STATE VARIABLES OF GOLTSY LANDSCAPE

The Suntar-Khayata station site was located in goltsy landscape in the north mountain group of the range at 2040 m a.s.l. in a through glacial valley. Here various observations were carried out, including soil temperature measurements at different depths, snow cover and evaporation observations. These data were used to evaluate and correct the model parameters that describe the goltsy complex.

4.1 Soil temperature

Geothermal measurements were carried out at the Suntar-Khayata station site in three bore holes at depths from 10 to 20 m. Soil temperature modeling with the daily calculated interval at different depths was done using thermal-physical soil properties, generalized according to the materials from Grave and Koreisha reports (Grave 1959; Grave& Koreisha 1957; Grave& Koreisha 1960; Koreisha 1963). Assessed values of average volume density with 42% porosity and natural moisture content, with pores fully saturated with ice and in dry condition are equal to 1700, 1930, 1580 kg/m^3 correspondingly. Specific heat capacity of soil particles in dry condition accounts for 840 J/kg °C, and specific heat conductivity – 1,5 w/m °C.

The *Hydrograph* model calculates heat balance for different layers of soil for each estimation period – that is daily. Verification of the results of the soil heat balance calculations was performed based on observed average monthly soil temperatures at numerous horizons (down to 2 m deep) at the Suntar-Khayata station site during 1958. Mean absolute deviations of the calculated monthly temperatures accounted for 1,4 ºC, 1,5 ºC, 1,1 ºC and 0,6 ºC, and their maximum values – +3,8 ºC (June), +4,0 ºC (November), +3,2 ºC (June) and -1,6 ºC (January) at 5, 50, 100 and 200 cm depths correspondingly. Overall, calculated and observed soil temperature values at different depths fit together (Figure 2A).

4.2 Snow cover

Data on snow depth and storage at the Suntar-Khayata weather station were used to verify parameters of the *Hydrograph* model. Snow depth was measured by means of three rods, placed in the corners of a triangle 12 m on a side. Water equivalent was estimated based on mean snow depth and bulk density (Koreisha 1963). Comparison of the calculated and observed values of snow depth and storage during winter seasons 1958-1959 is presented in Figure 2B and proves the model adequacy.

Base on the snow surveys data which were carried out nearby the station along the 3-km long route with elevation range over 400 m, a coefficient of snow depth variation was calculated equal to 0,57. It was used for snow distribution modeling for the goltsy landscape.

4.3 Evaporation

Evaporation observations were carried out by the means of two GGI-500-50 land evaporimeters that were installed at the Suntar-Khayata station site in the early June 1958, while the snowpack was continuous, didn't start melting yet, and soil temperature was below zero. Evaporation tanks were filled with rank soil the site consisted of, set full height into dig out slots and left under snow until it completely melted at the site, what happened on June 20-27. Evaporation observations continued throughout August 1958. Evaporation tanks were weighted every 5 days, precipitation was registered daily in direct proximity to them (Grave 1959).

The precipitation, observed in August 1958, accounted for 77 mm, infiltration rate – 36 mm, average evapotranspiration – 44 mm. Calculated evapotranspiration in August of that year accounted for 37 mm, and during warm seasons 1957-1964 – 50 mm on average. Calculated data do not contradict the observations from loosen rocks of the goltsy complex at the Kolyma water balance station, where annual average evaporation at the altitude range 1200-1700 m a.s.l. (1979, 1983-1984) was equal to 72 mm according to Lebedeva's et al. data (2015).

5. RESULTS OF MODELING RUNOFF FORMATION PROCESSES

Continuous runoff modeling with daily temporal resolution was carried for the Suntar river basin in the Sakharinya river's mouth cross-section, for 1957-1964, using data from four weather stations (Figure 1). Key water balance components are presented in Table 1, and comparison of the observed and calculated runoff hydrographs – in Figure 2C.

For modeling, there were used adjusting coefficients 1,1 and 1,15 to solid and liquid precipitation correspondingly. Analysis of the reports (Grave 1959; Grave & Koreisha 1957; Grave& Koreisha 1960; Koreisha 1963) and water balance modeling do not confirm the adjusting coefficient of 1.6 for solid precipitation recommended in (Reference Book 1968), and the authors of the study presented consider it unreasonably overstated.

For the 1957-1964 period, the calculated annual precipitation depth for the Suntar river basin is 344 mm on average, the calculated runoff depth – 199 mm, which is on average 10% higher than observed value of the runoff depth (180 mm). Evaporation from the catchment equaled 143 mm, which is 50% lower than the value stated for this region in the study (USSR 1966). The average Nash-Sutcliff coefficient of efficiency of calculating hydrographs for the main-stream outlet was 0,75. Overall, despite minor overvaluation of the runoff during flood period, the calculated runoff hydrographs match the observed ones quite well, both in phases and absolute discharge values.

Based on the results of the modeling, elements of the water balance for each runoff formation complex (RFC) were evaluated, and was also estimated each RFC contribution to runoff formation at the main-steam station (Table 1). Goltsy complex that covers 7% of the catchment provides 20% of the total runoff at the main-steam outlet of the Suntar river, and a runoff coefficient reaches 0,92. Tundra is the largest contributor to the runoff formation at the Suntar river catchment – 49 % of the total runoff, with a runoff coefficient – 0,74. The total runoff from the taiga and swamped sparse forest landscapes, which take 56% of the territory, is around 31%. Contribution of the goltsy landscapes increases in low water years and in the Suntar river basin it can be 28% (for example, in 1963 the total annual runoff depth was just 130 mm, while the calculated one for the goltsy complex – 513 mm).

Table 1

The water balance of RFC at the Suntar river, 1957-1964

	Total	Goltsy complex	Alpine tundra belt	Taiga and swamped sparse forest
Simulated flow, mm	199	567	263	105
Precipitation, mm	344	618	356	292
Evaporation, mm	143	50	86	186
Area percentage, %	100	7	37	56
Flow percentage, %	100	20	49	31
Coefficient of flow, m^3m^{-3}	0,59	0,92	0,75	0,36

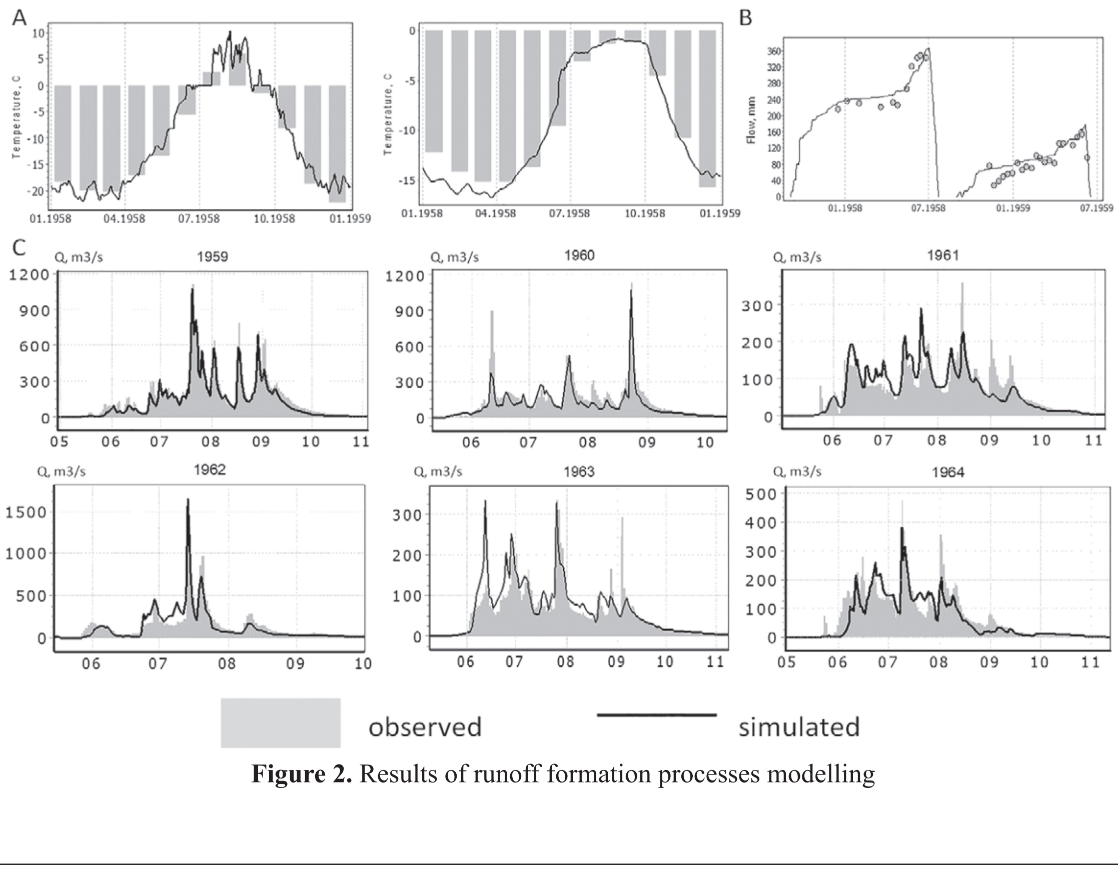

Figure 2. Results of runoff formation processes modelling

6. CONCLUSION

Based on the observation data at the high altitude Suntar-Khayata station (the Indigirka river upper reaches) under the program of the International Geophysical Year in 1957-1959 parameters for the *Hydrograph* hydrological model were developed, which describe the runoff formation processes in the high mountain goltsy zone of the Suntar river basin. Various states of snow cover and heath dynamics in soil profile in the goltsy zone were modeled, as well as runoff formation process throughout the whole catchment of the Suntar river. Modeling results are considered acceptable.

Model calculations have allowed to evaluate long-term average annual values of water balance for different landscapes, their contribution to runoff formation in the mountain river outlet, and offer updates for reference values of long-term average annual precipitation and evaporation for the research region.

Total runoff from the goltsy and alpine tundra that take 44% of the catchment area, on average accounts for 69% of the runoff in the Suntar river outlet, reaching 75% in low water years.

Nowadays, in the mountain regions of the Yana, Indigirka and Kalyma rivers basins there isn't any research station left to perform a comprehensive study of runoff formation processes in different permafrost landscapes. Therefore, development and verification of methods for hydrological processes modeling that are used while data is extremely scarce, become more of great current interest. The study presented has demonstrated that the *Hydrograph* hydrological model can become a foundation for solving scientific and practical issues in the research region.

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MONITORING OF THE UNUGESTYAKH LAKE WITH OUTLET IN CENTRAL YAKUTIA

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ABSTRACT
Features of the Unugestyakh lake with outlet and its talik are studied. The lake talik connects surface flow with suprapermafrost and intrapermafrost groundwater. Seasonal and interannual variability of water chemical compositions of lake and outflowing creek are characterized. Lake water TDS has decreased for last 6 years. It suggests increase of the suprapermafrost groundwater contribution to the lake.

KEYWORDS
lake, spring, lake talik, groundwater discharge, chemical composition

1. INTRODUCTION
Water-saturated suprapermafrost and intrapermafrost taliks are widely developed in the terraces of the Lena River in Central Yakutia. The region is characterized by continuous permafrost with a thickness of up to 250-400 m. The talik aquifers contain high quality drinking water, which is bacteriologically sterile and is protected from surface contamination by permafrost with a thickness of 15-50 m. The recharge of intrapermafrost aquifers is conducted mainly by the infiltration of the suprapermafrost waters through subaquatic and subaerial taliks (Efimov, 1952; Anisimova, 1981; Boytcov, 2002; Pavlova et al., 2016). Suprapermafrost waters turn into the category of intrapermafrost waters on sections of transit and discharge. The natural groundwater outlets are usually formed as concentrated groups of springs at the foot of the fourth (bestyah) Lena River terrace. Some suprapermafrost and intrapermafrost groundwater discharge into the lakes in the form of underwater sources. Underwater sources are not well studied since they could not be directly observed. It is not understood how suprapermafrost and intrapermafrost waters impact on the heat and hydrochemical regime of lakes and streams. The spring water contribution to the water bodies has not been estimated. To fill this gap hydrological and hydrochemical studies have been conducted at the Unugestyakh source lake since 2010.

2. METHODS
The work included the discharge measurement of a creek flowing out of the Unugestyakh lake, the sampling of lake, creek, suprapermafrost and intrapermafrost water. Suprapermafrost water was sampled from pits in the seasonal thaw layer. Samples of the intrapermafrost waters were taken from a well located in the lake watershed (Fig. 1). The studies were carried out during the winter low-flow period (March-April) and autumn (September-October) before freeze-up. In some years, discharge measurements and water sampling were carried out in the summer. Two boreholes with a depth of 15 m were drilled on the lake watershed. Samples for the ground water/ice content were taken from the cores. Boreholes have been used for ground temperature observations. The geological structure of the territory is characterized by the drilling data of the Yakutsk exploration and survey expedition. Meteorological data from the Pokrovsk weather station was used in the study. Meteorological data source is the website “Reliable Prognosis” Ltd. (https://rp5.ru).

3. RESULTS

The Unugestyakh lake is located on the gentle sandy slope of the bestyah Lena River terrace, facing south-west towards the valley of the Menda river, the right tributary of the Lena river. The shape of the lake is almost round. The dimensions are 0.7 x 0.9 km (Fig. 1). Lake bottom is funnel-shaped. The maximum lake depth is 4 m. The lake depth in 50 m from the coast is 1.3-1.5 m. The lake does not have any traces of desiccation. The north and north-eastern banks are overgrown. The larch-birch forest surrounds the lake with a 30-150 meter ring. Sparse pine forest with traces of fires is spread behind the larch-birch forest. The perennial Unugestyakh Creek flows out of the Unugestyakh Lake in its southern part. Despite the rather low water (0-0.5 ° C) and air temperature (-30 – -50°C) in winter, a polynya is preserved in the lake near the creek source. The length of the polynya reaches 50 m in some years. The aufeis is formed downstream. In the summer, the creek joins the Malyi Unugestyakh Lake, and then flow to the Menda River.

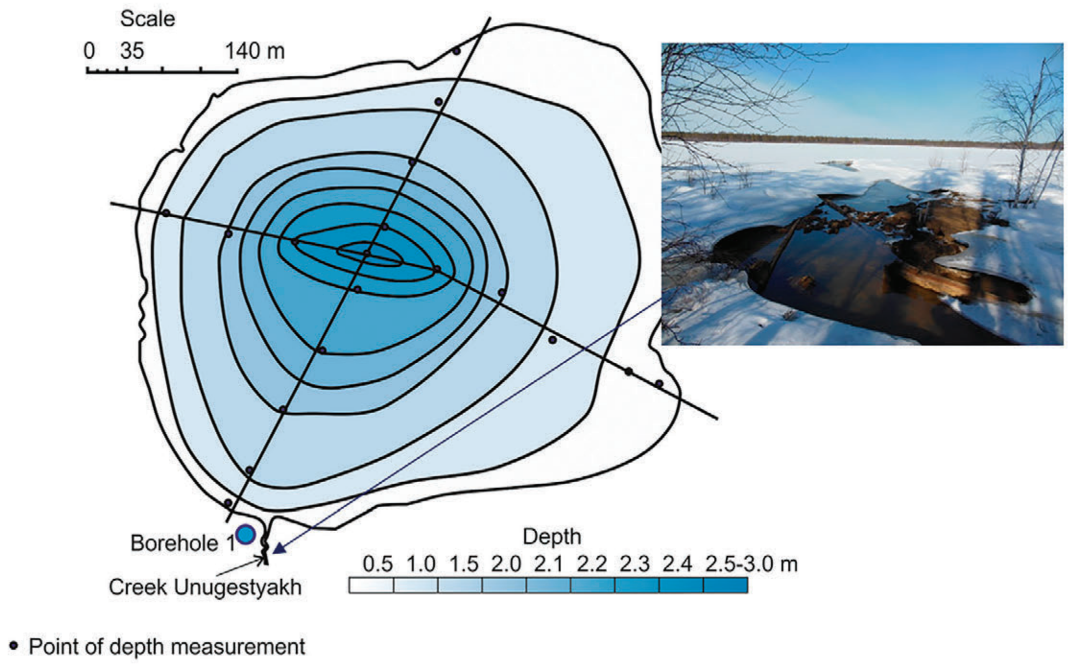

Figure 1. Bathyorographical scheme of the Unugestyakh lake and lake polynya near the creek source.

The base of the geological section is formed by middle-upper Pleistocene fine and medium-grained sands with gravel. Quaternary deposits are underlain by Middle Cambrian limestones at a depth of 37 m. The seasonal thawing depth varies from 3 m on sandy slopes up to 1.5 m near the Unugestyakh Lake. There are suprapermafrost taliks with thickness of 6-10 m in some parts of the lake watershed. In the upper part of the taliks, the sands have extremely low moisture content (2-6%). Near the talik base, it approaches the maximum water capacity (15-20%).

Lake talik is developed beneath the Unugestyakh lake. The talik thickness exceeds 100 m. In the coastal parts the upper 24 m are frozen. Drilling in the lake bed showed that the level of groundwater in quaternary sediments was established at a depth of 1.5 m, and groundwater from the Cambrian deposits - 0.8 m above the land surface. It indicates that the groundwater from the Cambrian deposits contributes to the Unugestyakh Lake through the quaternary aquifer and sand deposits. A preliminary analysis of the 7-year observations of streamflow from the lake in different seasons showed that groundwater contribution to the lake, estimated on the basis of the winter streamflow measurements of the Unugestyakh creek, is at least 4000 m^3/day.

The results of hydrochemical studies suggest mixing of waters of different genesis in the talik and lake. All sampled water relates to hydrocarbonate type. The groundwater from the Cambrian sediments has the highest TDS (Table). They belong to the sodium (47% of the main cations sum) and magnesium (36%) group. Underground waters from the Quaternary aquifer have a TDS of 0.2 g/l. Magnesium (36%) and calcium (56%) ions prevail. Suprapermafrost waters of the seasonally thaw layer are ultra-fresh and mixed in cations. Their chemical composition is stable during the summer.

Table

Water chemical composition of the Unugestyakh study site, mg/l

Sampling date	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^{-}	CO_3^{2-}	SO_4^{2-}	Cl^{-}	TDS	pH	Sr	Eh, mv
Borehole 1, Quaternary aquifer, depth interval 24-37 m												
02.07.2004	29.5	11.5	11.6	1.2	189.1	0.0	2.7	3.4	207.1	7.6	0.22	-
Borehole 1, Cambrian aquifer, depth interval 37-100 m												
21.08.2004	22.5	26.3	60.0	3.5	378.2	0.0	1.7	4.1	501.4	7.9	5.51	183
Suprapermafrost waters of the seasonally thaw layer												
22.06.2012	2.9	1.6	2.4	0.7	27.4	0.0	1.6	0.5	37.5	5.6	0.06	570
30.09.2012	5.2	1.4	2.0	1.5	21.2	0.0	1.6	1.0	34.7	6.0	н/о	433
Unugestyakh creek												
28.03.2012	38.0	16.0	9.0	1.0	222.2	0.0	4.9	1.4	298.8	7.0	0.31	-
21.06.2012	19.6	7.8	4.5	0.2	93.5	6.6	1.6	1.2	135.3	8.8	0.18	-
30.09.2012	31.2	11.3	5.0	0.4	115.6	14.4	0.7	1.0	179.7	8.7	н/о	383
19.03.2016	32.7	17.0	6.0	1.20	212.9	0.0	0.2	0.8	272.2	7.0	0.32	230
22.06.2016	20.9	7.5	3.4	0.40	54.1	28.6	0.3	0.1	115.8	9.1	0.08	412
06.09.2016	17.6	6.11	3.0	0.30	93.7	2.85	0.1	0.2	124.0	9.3	0.20	409
24.03.2017	39.4	10.0	6.5	1.20	194.7	0.0	0.2	0.4	254.2	7.0	0.32	220

Note: - – chemical parameter was not analyzed

TDS of the lake and the creek varies seasonally, but the ratio of the main cations throughout the year is stable: calcium (51-52%) and magnesium (30-35%) ions prevail. Minimal TDS (0.1 g/l) of the lake and creek water occurs during the snow melt and early summer. TDS slightly increases during the late summer and autumn. At the end of winter the lake and creek TDS reaches 0.3 g/l. In March-April, the increase of strontium concentration is observed in the creek. Higher concentrations of strontium are characteristic of Cambrian aquifer. This confirms the contribution of Cambrian aquifer groundwater to the Unugestyakh lake.

TDS and water discharge of the Unugestyakh Creek directly correlates with each other in autumn (Fig. 2). TDS of the Unugestyakh creek in October corresponds to the precipitation amount for August and September. Autumn streamflow agrees with sum of summer air temperatures: an increase of air temperature leads to an increase of creek discharge.

There is an inverse relationship between TDS and streamflow in winter: increase of streamflow is accompanied by TDS reduction. Desalination of surface waters has been observed for the last 7 years. The possible reason could be a change of the seasonal thawing depth, and, accordingly, of the suprapermafrost groundwater flow into the lake. Additional research is required to reveal the mechanism of change.

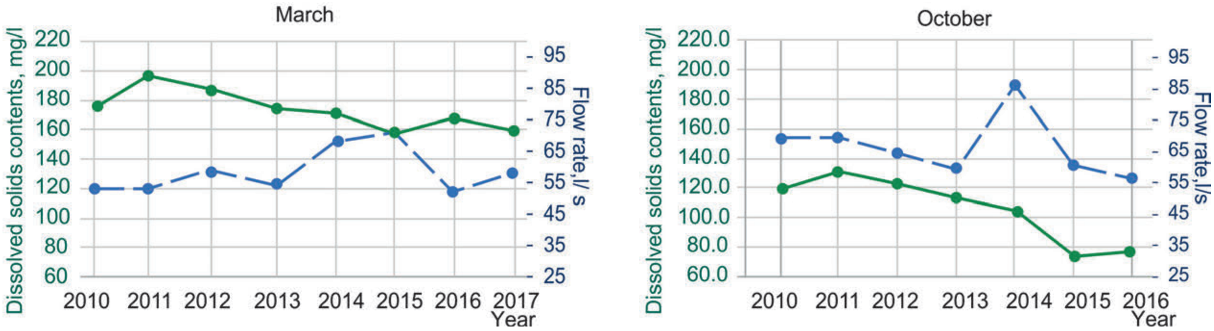

Figure 2. TDS and water discharge of the Unugestyakh creek for March and October, 2010-2017 (2016). Solid line corresponds to TDS, dotted line – to water discharge.

4. CONCLUSION

The study results are preliminary. Comprehensive research is required to establish the basic regularities of the geothermal, hydrochemical and level regime of the lake, creek and their interactions with suprapermafrost and intrapermafrost groundwater. Chemical composition of the lake and lake talik is formed by mixing of groundwater of different genesis. The Quaternary aquifer connects the underground waters of the Cambrian aquifer with the suprapermafrost water of the seasonal thaw layer and the lake. The creek water desalination observed in recent years indicates an increase of suprapermafrost water contribution to the lake.

Permanent pressure regime in the talik suggests that part of the intrapermafrost groundwater flow is transit and discharges in the valleys of nearby rivers.

The Unugestyakh lake is not unique for the studied region. There are also a number of water bodies with year-round outflow. Regular observations will allow assessing the influence of climate change on the variability of water resources in the region.

This territory is intensively developed and there is increased demand for potable water. Lakes with underwater sources could be considered as an alternative source of water supply.

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TOWARD UNDERSTANDING THE TRAJECTORY OF HYDROLOGICAL CHANGE IN THE SOUTHERN TAIGA PLAINS, NWT, CANADA

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ABSTRACT

Climate warming in the southern Taiga Plains ecoregion of northwestern Canada has led to unprecedented rates of permafrost thaw and a myriad of land-cover changes with uncertain impacts on hydrology. As a result there is growing uncertainty in regards to the future availability of water resources in this region. This paper synthesises key findings of recent hydrology field studies and remote sensing analyses of land-cover change in the southern Taiga Plains to improve the understanding of the trajectory of land-cover change in this region and how such change can be expected to influence water flow and storage processes.

KEYWORDS

Permafrost thaw; landcover change; peatlands; Boreal; hydrology.

1. INTRODUCTION

Northwestern Canada is one of the most rapidly warming regions on Earth, and permafrost thaw is one of the most important and dramatic manifestations of warming in this region. Permafrost thaw has wide ranging environmental impacts that include rapid changes in geomorphology (*e.g.* Jorgenson *et al*., 2013); browning of the forest (Michaelian *et al*., 2011); frequency and severity of wild fires (Flannigan *et al*., 2009); soil drainage (Jorgenson *et al*., 2013) flow path lengths and transit times (Jones & Rinehart, 2010); biogeochemical fluxes (Gordon *et al*., 2016), thermakarst lake storage (Korosi *et al*., 2017), groundwater fluxes (Bense *et al*., 2009), and other impacts. Thaw is particularly pronounced in the southern margin of discontinuous permafrost (Kwong & Gan, 1994), where permafrost is thermally insulated by an organic cover of dry peat, allowing it to persist even where mean annual air temperatures are positive (Smith & Riseboroug, 2002). Such "ecosystem-protected permafrost" (Shur & Jorgenson, 2007) is particularly susceptible to thaw since it is already at the melting point temperature, and its discontinuous nature enables energy to enter individual permafrost bodies not only vertically from the ground surface (as for continuous permafrost), but also laterally from adjacent permafrost-free terrain. Furthermore, because such bodies are relatively thin (<10 m), permafrost thaw in this southern margin often quickly leads to local disappearance of permafrost (Beilman & Robinson, 2003). Permafrost thaw leads to ground surface subsidence which transforms landscapes and ecosystems and ultimately affects the distribution and routing of water. As such, permafrost thaw is confounding the prediction of hydrological responses, a situation made worse by uncertain environmental feedbacks on the rates and patterns of such thaw.

Despite the large number of studies documenting the environmental impacts of permafrost thaw, there remains little consensus on the trajectory of the thaw-induced land-cover change in the southern Taiga Plains. Since hydrological functions vary among land-covers, a change in their relative proportion influences the water balance of drainage basins. To properly manage the water resources of the southern Taiga Plains, decision makers in both government and industry and in local communities require an understanding of 1) the hydrological differences among the major land-cover types, and 2) how permafrost thaw is changing the relative proportions of these land-covers. An understanding of these two factors will allow new insights into the trajectory of land-cover change and its implications on regional water resources. By synthesising recent hydrological field and remote sensing studies, this paper provides key insights into the trajectory of land-cover change in the southern Taiga Plains, and how this may affect water resources.

2. METHODS AND SITE DESCRIPTION

This study is focussed on the southern Taiga Plains ecoregion in northwestern Canada (Figure 1a), and draws mainly from studies conducted at Scotty Creek (61°18' N, 121°18' W), a 152 km^2 drainage basin 50 km south of Fort Simpson, Northwest Territories (NWT) (Figure 1b). Scotty Creek basin is underlain by discontinuous permafrost (Hegginbottom & Radburn, 1992) and is covered by peatland complexes typical of the 'continental high boreal' wetland region (NWWG, 1988). The peat thickness at Scotty Creek ranges between 2 and 8 m (McClymont *et al.*, 2013) below which lies a thick clay/silt-clay glacial till deposit of low permeability (Aylesworth & Kettles, 2000). Most of the Scotty Creek basin is a heterogeneous mosaic of forested peat plateaus underlain by permafrost, and treeless, permafrost-free wetlands (Figure 1c), typical of the southern fringe of discontinuous permafrost (Helbig *et al.*, 2016). The 1981-2010 climate normals indicate that Fort Simpson and has a dry continental climate with short, dry summers and long, cold winters. Fort Simpson has an average annual air temperature of -2.8° C, and receives 388 mm of precipitation annually, of which 38% is snow (MSC, 2013). Snowmelt usually commences in early to mid-April and continues throughout most of the month, so that by May, only small amounts of snow remain (Hamlin *et al.*, 1998). This paper draws on numerous published and unpublished studies involving both hydrometric field observations and aerial/satellite image analysis. Collectively these studies were used to inform conceptualisations of coupled land-cover and hydrological change presented herein.

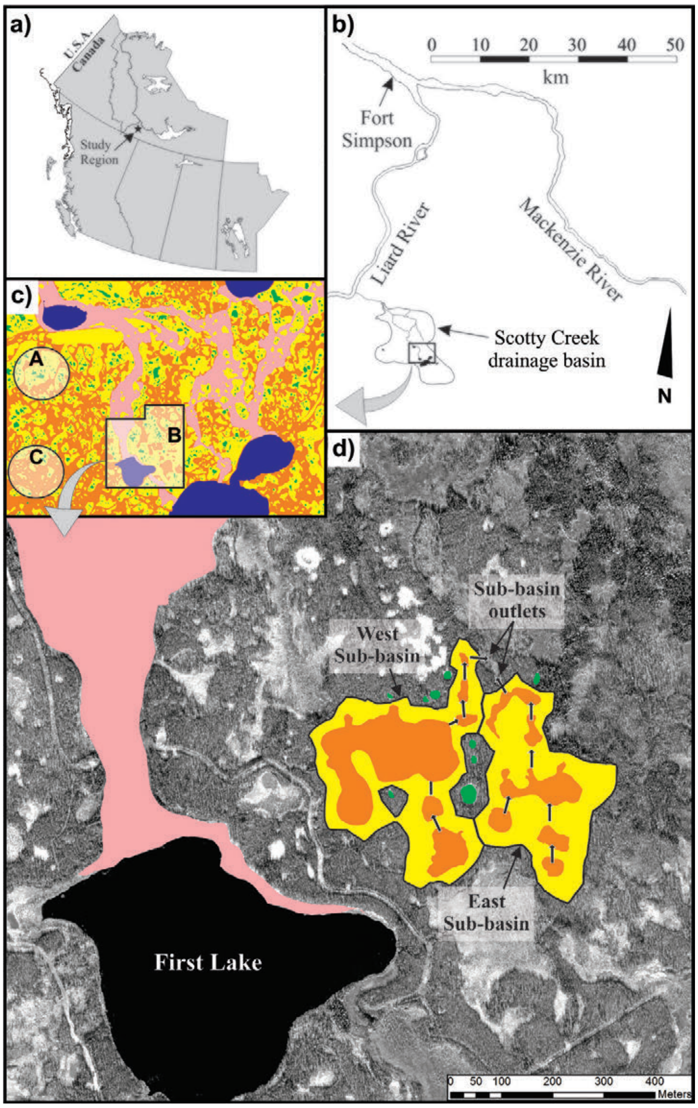

Figure 1. a) study region and b) Scotty Creek; c) peat plateaus (yellow), isolated (green) and connected (orange) bogs, channel fens (pink); d) enlargement of "B" showing cascade bogs.

3. RESULTS AND DISCUSSION:

Ice-rich permafrost in the form of tree-covered peat plateaus dominates much of the fringe zone, where plateaus rise 1 to 2 m above the surrounding wetland terrain of flat bogs and channel fens. Figure 1c shows a classified Ikonos image acquired in 2000, of a 22 km^2 subarea within the Scotty Creek basin where permafrost plateaus occupy the largest (43%) portion of the subarea, followed by expansive bogs that are hydrologically connected to channel fens (23%), channel fens (21%), lakes (9%) and isolated flat bogs, also known as collapse scar bogs (4%). Unlike the expansive bogs, the isolated bogs are surrounded by raised permafrost and as such are unable to exchange surface or near-surface flows with the fens. The contrasting biophysical properties of these peatland types, gives each a specific role in the water cycle (Quinton *et al*., 2003). The plateaus function primarily as runoff generators, given their relatively high topographic position and limited capacity to store water. The isolated bogs, being internally drained are predominantly areas of water storage. The expansive bogs exchange surface and near surface flows with channel fens during periods of high moisture supply, but otherwise predominantly store the water they receive. Water draining into channel fens from the surrounding plateaus and (during periods of hydrological connection) expansive bogs, is conveyed laterally along their broad (~50-100 m), hydraulically rough channels to streams and rivers (Quinton *et al*., 2003).

Permafrost thaw increases the cover of the bogs and fens at the expense of the forested peat plateaus. Using tree-cover as a proxy for the presence of permafrost, the area underlain by permafrost at Scotty Creek decreased from 70% in 1947 to 43% in 2008 (Quinton *et al*., 2011), with degradation rates increasing in recent decades (Baltzer *et al*., 2014). This rate is consistent with that estimated for the larger southern Taiga Plains region, where 30%-65% of the permafrost has degraded over the last 100-150 years (Beilman & Robinson, 2003). The above mentioned percentages of Figure 1c occupied by each cover type should be assumed to be in transition.

Recent hydrological field studies at Scotty Creek (Connon *et al*., 2014) provides valuable insights into the nature of this transition and how it affects water flow and storage processes. They described ephemeral flow from bogs that were previously assumed to be hydrologically isolated. Specifically, during periods of high moisture supply, water was found to cascade bog-to-bog and then into channel fens. It was also found that the ephemeral channels connecting the bogs were areas of preferential permafrost thaw. Two bog cascades, one draining the West Sub-basin and the other draining the East Sub-basin, are identified in Figure 1d. The hydrographs of the two sub-basins (Figure 2) show the amount of water that would otherwise have remained on the plateau in the absence of the bog-to-bog drainage process. The annual drainage from the slightly smaller East Sub-basin is substantially larger since its bogs are smaller and therefore more readily filled (Connon *et al*., 2015), a condition that must be reached before bog-to-bog flow can commence.

Connon *et al*. (2015) compared historical images of Scotty Creek, and showed that numerous bogs that were isolated from the basin drainage network in 1970 (Figure 3a) had become connected to it by 2010 (Figure 3b). This 'bog capture' process increases basin runoff by increasing the basin's runoff contributing area (Connon *et al*., 2014). Therefore, in addition to initiating or at least enhancing bog-to-bog drainage cascades, permafrost thaw also transforms hydrologically-isolated bogs into 'open' bogs by removing the permafrost that once separated such a bog from the a near-by channel fen. This land-cover transformation is important hydrologically because it adds to the basin drainage network 1) runoff arising from direct precipitation falling onto the captured bog (*i.e.* bog drainage), and 2) runoff from the 'captured' bog's watershed (*i.e.* slope drainage). As captured bogs expand due to permafrost thaw at their margins, they merge into other bogs, a process that increases both the bog and slope drainage contributions to fens.

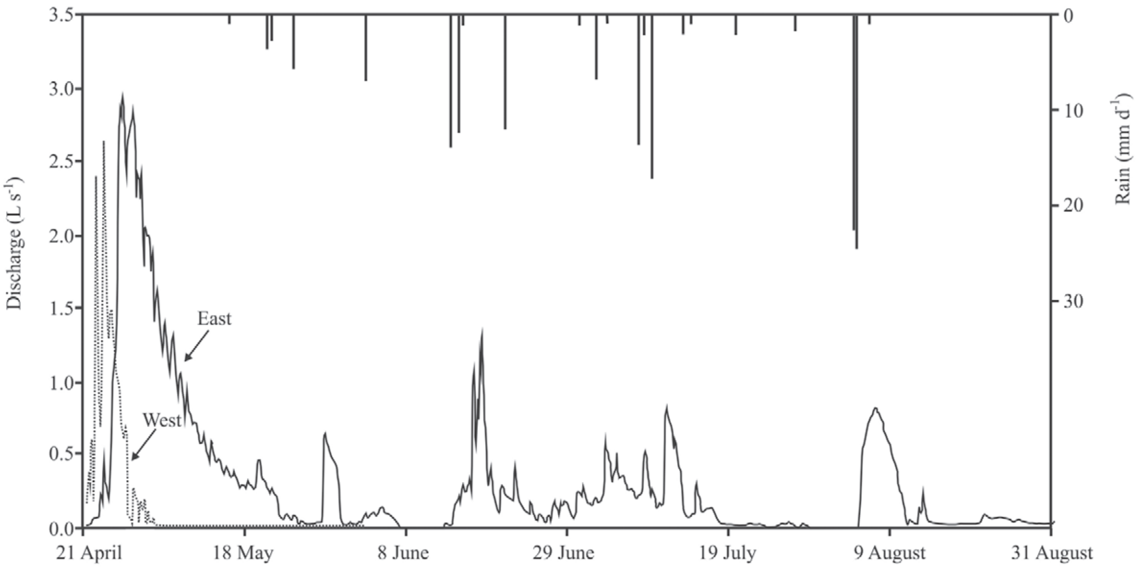

Figure 2. Runoff hydrographs measured at outlets of East and West bog cascades for 2014.

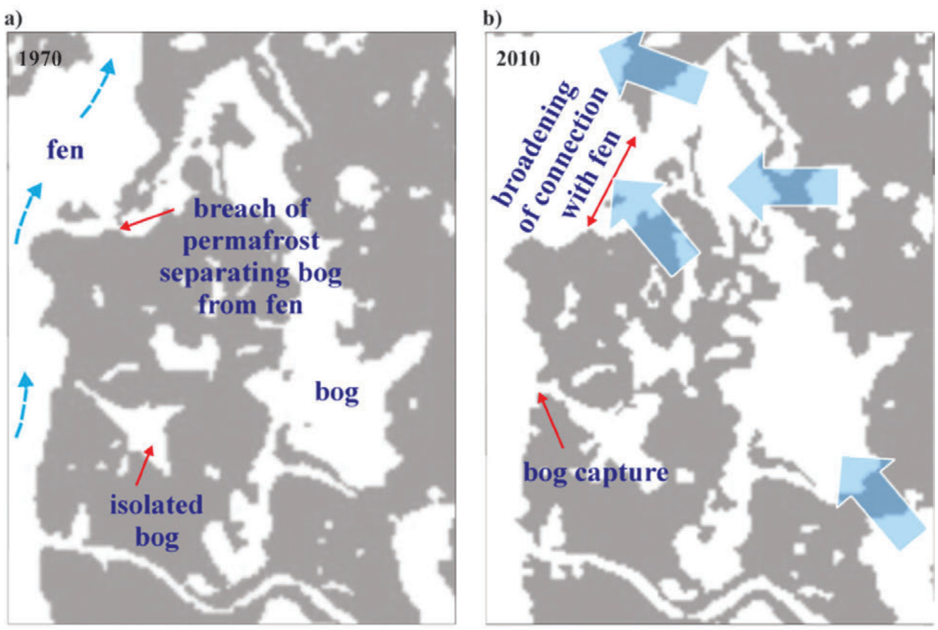

Figure 3. Classified images for a 1 km^2 area in the Scotty Creek catchment showing the change in permafrost coverage (grey) between 1970 (a) and 2010 (b). The large arrows in (b) signify subsurface flow through taliks to the basin drainage network of channel fens.

Figure 4a depicts an example of the bog capture process as it evolved between 2006 and 2015, based on detailed ground surface elevation and permafrost table depth surveys. As the permafrost table lowered between these two years, the plateau surface subsided and was flooded by the adjacent bog or fen, a process leading to loss of forest and expansion of the wetlands. By 2015, the permafrost table was below the elevation of the water tables of the bog and fen, and as a result, the permafrost body no longer obstructed subsurface flow from the bog, through the plateau, to the fen (Figure 4b). Furthermore, by 2015, a talik (*i.e.* perennially unfrozen layer) had formed which enabled the plateau to conduct subsurface flow throughout the year from the bog, down-gradient to the fen. Subsurface flow through the talik (Figure 3b) augments the surface and near surface flows into channel fens.

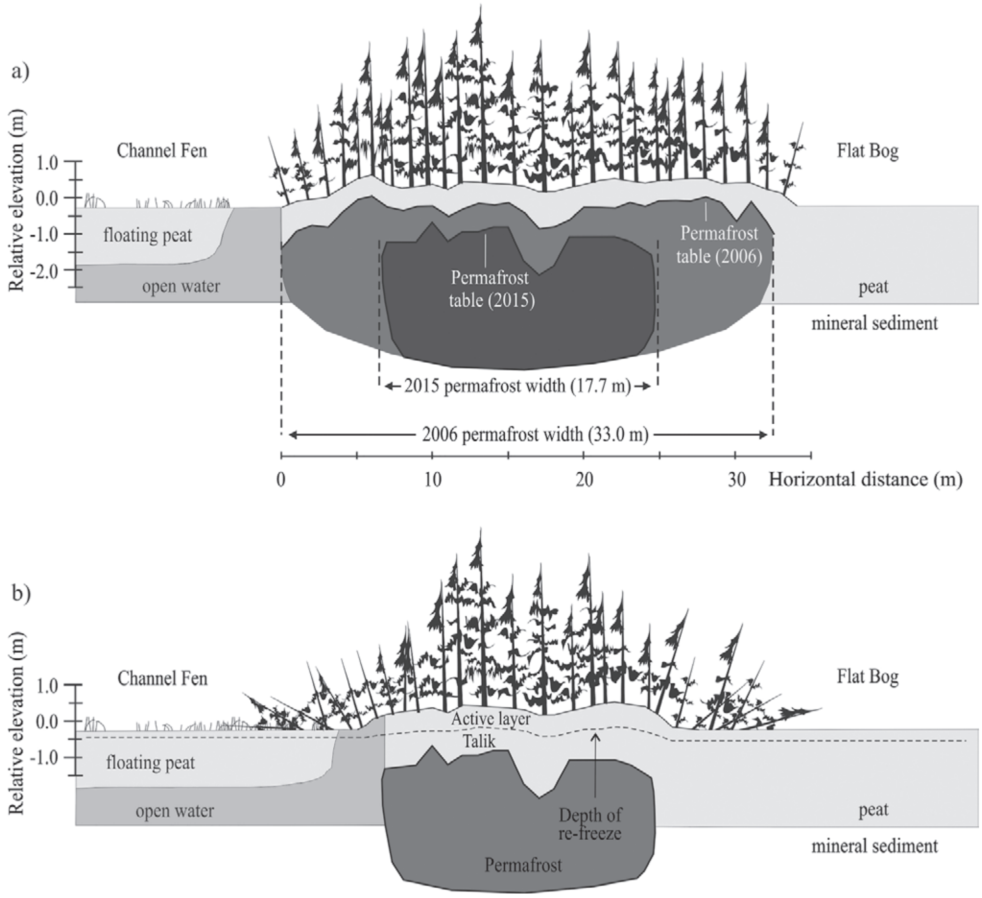

Figure 4. Cross section of a peat plateau at Scotty Creek based on measurements of supra-permafrost thickness and ground surface elevation at 1 m intervals showing the difference in depth and lateral extent of the permafrost table between 2006 and 2015 (a) and the development of a talik in 2015 (b).

There are strong indications that permafrost thaw and the resulting land-cover changes have affected basin water balances, as suggested by rising river flows throughout the border region (St. Louis & Sauchyn, 2009). Most notably, the total annual runoff from all gauged rivers in the lower Liard River valley of the NWT has steadily risen since the mid-1990s (Connon *et al*., 2014). The current understanding of water flow and storage processes in wetland dominated, discontinuous permafrost terrains, and how climate warming and the resulting ecological changes affects these processes, cannot explain this rise in flows, nor is it sufficient to predict future flows. Rising flows from subarctic rivers are often attributed to 'reactivation' of groundwater systems (*e.g.* St. Louis & Sauchyn, 2009), but the very low hydraulic conductivity of the glacial sediments below the peat, precludes appreciable groundwater input. Permafrost thaw-induced changes to basin flow and storage processes offers a more plausible explanation for rising river flows in this region (Connon *et al*., 2015).

Field observations and image analyses (Baltzer *et al*., 2014) suggest that plateaus contain two distinct runoff source areas separated by a break in slope approximately 10 m inland from the fen-plateau edge (Figure 5). Primary runoff drains from the sloped edges of plateaus directly into the basin drainage network (*i.e.* a channel fen). Field measurements suggest that the entire primary area supplies runoff to the fen throughout the thaw season. Secondary runoff drains into the interior of the plateau toward the topographic low often occupied by a bog. If the receiving bog is hydrologically isolated, the runoff it received will remain in storage, evaporate or recharge the underlying aquifer. If the receiving bog is part of a bog cascade, and if its storage capacity is exceeded, then the secondary runoff it receives will be routed toward the channel fen via the down slope bog or bogs. Secondary runoff is therefore, neither direct nor continuous. The rate of secondary runoff is greatest during periods of high moisture supply and minimal ground thaw when the hydrological connection among the bogs of a cascade, and between individual bogs and their contributing "bog-sheds" is maximised. As the active layer thaws and drains, the contributing area shrinks and secondary runoff decreases. Large rain events can temporarily reverse this decrease.

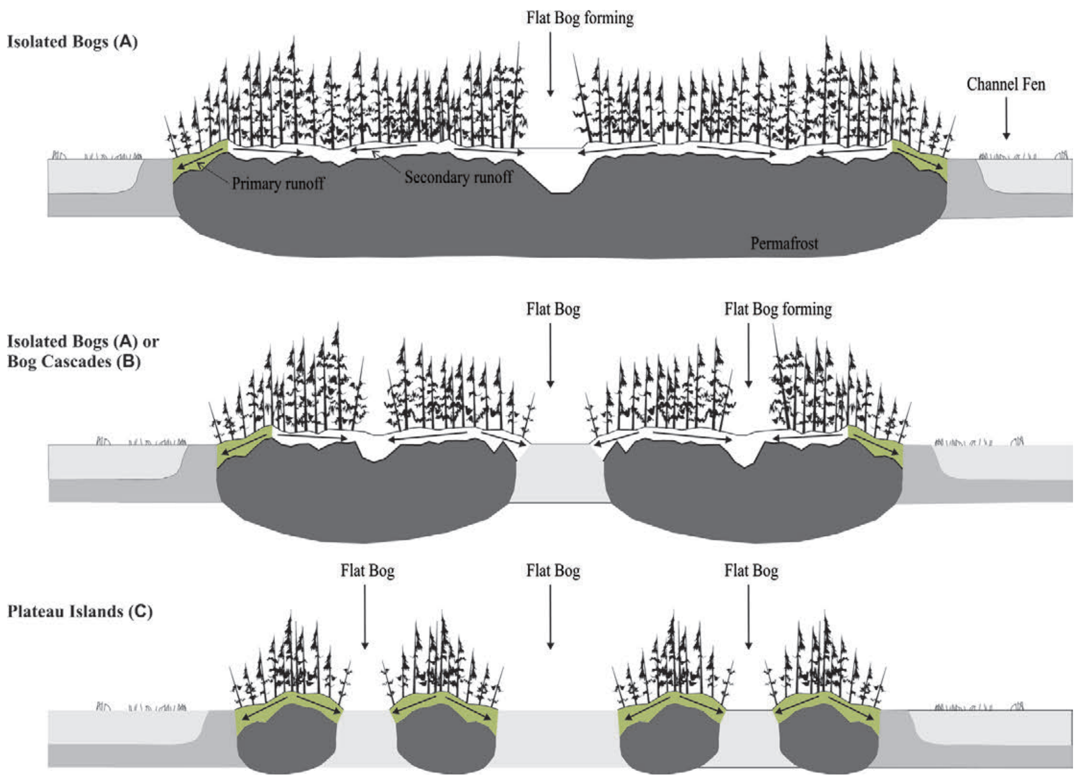

Figure 5. Conceptualisation of permafrost thaw induced land-cover transformation in the wetland-dominated zone of discontinuous permafrost typical of the southern Taiga Plains. Green areas represent the areas producing primary runoff.

Over a period of decades, the plateaus conducting primary and secondary runoff transform as a result of permafrost thaw (Figure 5). Three general stages can be seen from Figure 1c. In chronological sequence, the area indicated by "A" represents an early stage of permafrost thaw where bogs are mostly hydrologically-isolated, and as such, drainage into the fen is supplied only by primary runoff from the margins of the plateaus. "B" represents an intermediate stage of permafrost thaw where primary runoff is augmented by secondary runoff from the ephemeral bog cascades. The activation of secondary runoff arises from the greater hydrological connectivity of land-cover "B" than "A". As a result, a greater proportion of the snowmelt and rainfall arriving on land-cover "B" is converted to runoff than in the previous stage (Figure 6). Because B is transitional between A and C, some bogs are hydrologically connected (via surface or near surface flow and/or talik flow), while other bogs remain hydrologically isolated. "C" represents an advanced stage, where the shrinking peat plateaus occur as islands within an expansive bog complex. Interestingly, this stage is a near mirror image of "A" where it is the bogs that occur within an extensive plateau complex. By stage "C", plateau diameters are on the order of a few tens of metres and as such contain no secondary runoff and no interior bogs.

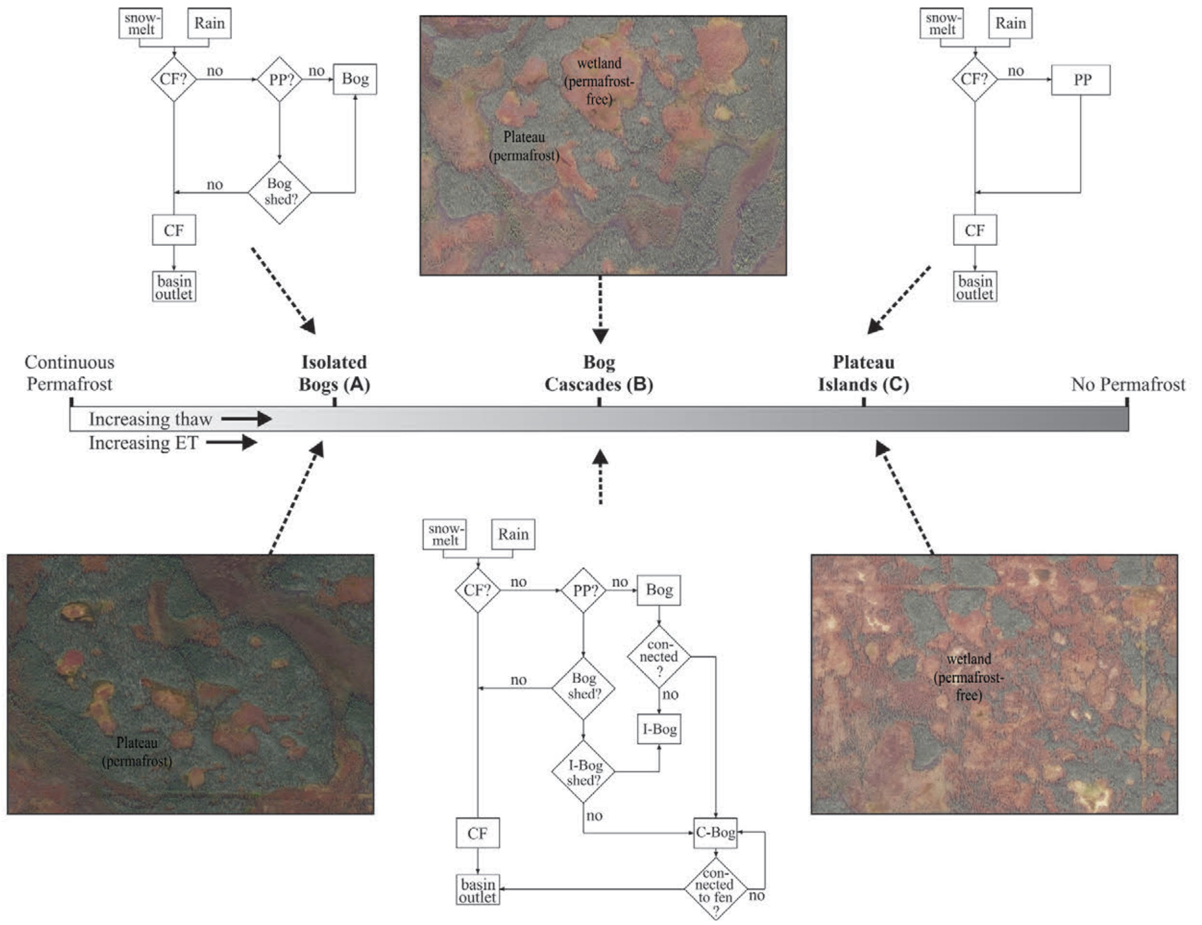

Figure 6. The transformation of peat plateau runoff generation processes with increasing thaw from left to right. PP = peat plateau, CF = channel fen, I = isolated, C = connected.

As the land-cover transitions from the isolated bog (A) to the plateau islands (C) stages, the way in which peat plateaus generate runoff changes dramatically, with direct consequences on their runoff pattern and rate (Figure 6). For each stage, water (snowmelt or rainfall) arriving in the channel fen is conducted directly to the basin outlet. Likewise, water arriving on a plateau but not within a bog catchment (*i.e.* bogshed) is routed directly to the adjacent fen. Water arriving directly into bogs or their bogsheds is prevented from reaching channel fens in stage A, but can reach the latter in stage B depending upon the degree of hydrological connectivity in the bog cascades. Activation of secondary runoff therefore increases the amount of runoff between stages A to B. Primary runoff may also increase between these two stages since the fragmentation of plateaus increase can increase the length of the overall plateau-fen edge. Water arriving onto plateaus in stage C is neither stored nor routed as secondary runoff through bog cascades and as such this stage provides the most direct runoff response. However, stage C has the lowest plateau runoff since it is capable only of generating primary runoff, and this runoff is generated from the relatively small total surface area of the remaining plateaus.

4. SUMMARY & FUTURE DIRECTIONS

Studies at Scotty Creek have recently expanded to include remote sensing and ground-based observations along a ~200 km transect that extends from Scotty Creek southward to the NWT-British Columbia border. By substituting space for time, the land-cover characteristics near the southern end of this transect suggests that the trajectory of Scotty Creek is towards increasing fragmentation and eventual disappearance of peat plateaus. Less clear is the trajectory of the intervening wetland (*i.e.* bog and fen) terrains. The transect studies suggests that a concomitant expansion of the wetland area with the shrinkage and loss of peat plateaus would initially produce a wetter land-cover characterised by expansive wetland with little forest cover. Although the hydrological connectivity of this stage would be high, the reduction of the plateau area reduces the impact of their relatively rapid flowpaths, and as a result, such a land-cover may produce less runoff than presently observed at Scotty Creek. However, recent studies in the Scotty Creek region (*e.g.* Helbig *et al*., 2016) indicate increased basin average evapotranspiration (ET) as the relative coverage of wetland terrain increases (Figure 6). The transect studies also suggest that this initial wet stage is superseded by a drier land-cover of the type presently observed near the NWT-British Columbia border. In that region, the permafrost-free terrain is sufficiently dry to enable the regrowth of forest covers that include black spruce (without permafrost) and a greater proportion of deciduous species. Although this synthesis provides some insights into the trajectory of land-cover and hydrological change in the southern Taiga Plains, there remain several significant unknowns. For example, the time scale over which the land-cover transitions will occur is not well understood. There is also a dearth of knowledge on how possible ecological and/or hydrological feedback mechanisms may affect trajectories of land-cover change. There is also little understanding of how such trajectories may change in response to changes in precipitation regimes, such as total annual precipitation, the proportion of the latter occurring in the form of snow, the number of multi-day events and other precipitation distribution characteristics, and the timing of snowmelt.

5. ACKNOWLEDGEMENT

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TRACER STUDIES OF PREFERENTIAL WATER FLOW PATHS IN MOUNTAIN SLOPES, PACIFIC RUSSIA

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ABSTRACT

The investigation of valley slopes in Sikhote-Alin' Mountains and in Upper Kolyma Highlands (Pacific Russia, Northeast Asia) in 2015 and 2016 has revealed comparatively big preferential water flow paths (PFP) – seasonal subsurface watercourses where slope flow mostly occurs. These watercourses were found within bodies of active and relic block streams – usually narrow stone stripes of deposits on the slopes. The flow rates in some of those watercourses were estimated as 0.23–6 l/s, and flow velocities – as 0.4–7.2 cm/s. Those values approach to the low-water flow rates in headwaters streams. Evidently, the relic block streams and, consequently, the biggest slope subsurface watercourses are to be controlled by comparatively small faults. Accounting the generalized data on the spring catchments dynamics and patterns in the karst areas of Piedmont, Italy (by Vigna et al. 2010, 2015) and mean transit time estimates in talus water flow in Rocky Mountains, Oregon, USA (by McGuire & McDonnel 2010), we offer to divide the subsurface PFP network onto three main groups: 1) main conductors – conduits; 2) filtration flow domains; and 3) matrix flow paths. Supposedly, the biggest slope subsurface watercourses are the main form of slope flow and river flow interaction within a given catchment.

KEYWORDS

Pacific Russia; slope flow; preferential flow paths; tracers

1. INTRODUCTION

The water flow generation processes and mechanisms of interaction between subsurface and surface flow remain a key problem in hydrology and hydrogeology, despite the abundance of datasets from standard observations, the results of special studies in experimental and representative catchments, and scientific publications in this field as well.

In mountain landscapes where excessive events regularly occur, surface temporary streams on slopes are extremely rare due to turfiness and high permeability of soil (Gartsman et al 1971; Zhil'tsov 2008; Glotova & Glotov 2012; Shepelev 2011). Atmospheric water, as a

rule, quickly penetrates into the slope deposits and then flows as more or less concentrated subsurface streams. These streams are denoted in plenty publications in various terms as "fast groundwater flow" (Vinigradov 1967), "contact flow" (Befani et al. 1966), "macropore flow" (Pierce et al. 1986; Weiler 2017), pipeflow (Uchida et al. 2005), "flow in subsurface temporary channels" (Vasilenko 2013), "rapid subsurface storm flow" (Vigna & Banzato 2015) et al. The slope flow are currently poorly studied (Angermann et al. 2016), and, obviously, in connection with this, are usually not considered as order-forming tributaries of rivers. There is also no clarity in understanding the conditions and causes of flow concentration in slopes. At the same time, the study of processes of this concentration is consistent with the current concept of preferential flow paths (PVPs) in catchments. Moreover, their undercounting can lead to errors in calculations and forecasts of extreme flood characteristics based on popular runoff generation models (Brutsaert 2005; Uhlenbrook 2006; Weiler 2017; Gerke et al. 2010; Vasilenko 2013; et al.). Subsurface slope flow often rapidly enters the mountain streams, thereby contributes to extreme floods, especially in the regions impacted by the powerful extratropical cyclones. The established increase in the proportion of subsurface slope water in low-order streams during rain floods (Gubareva et al. 2015, 2016) as well as two peaks in flood hydrographs caused by a single rain (Gartsman & Shamov 2015), confirm this fact.
Dynamic characteristics of the slope subsurface streams observed in small mountain river valleys as well as geological and geomorphological conditions of these streams are discussed below.

2. OBJECTS, METHODOLOGY AND STUDY RESULTS

Three areas of study are located in Pacific Russia (Figure 1): 1) the upper reaches of the Kolyma R., (Upper Kolyma Highlands), 2) headwaters of the Ussuri/Wusuli R. basin (central part of Southern Sikhote-Alin Mountains), and 3) the Rudnaya R. basin (eastern edge of Southern Sikhote-Alin Mountains).

Figure 1. The areas of study. The symbols are given in the text body above.

In 2015 and 2016, the authors investigated the slopes of small river valleys in those areas. Rather narrow – with width of some meters to the first dozens of meters – coarse clastic stripes descending down the slopes investigated were found there. In the South of Pacific Russia, these stripes often, especially in foothill, are turfy and covered with abundant vegetation. That is why most of them are hard-identified in relief by common observation tools. In the north, the blocks forming such stripes, due to their low slipping down the slope, are found to be covered mostly with lichens and, consequently, more apparent and identifiable. In some of these formations, via making a number of pits of 70–100 cm depths, a free or practically free water flux was observed. Those pits also served as the observation sites to make a series of the slope flow measurements.

Permafrost landscapes in the Upper Kolyma Highland (site 1, Figure 1) are mainly underlain by Upper Permian sedimentary rocks (clayey shales). The stone debris and placers of intrusive granodiorite rocks are widespread at the altitude higher 1000 m above sea level. Fine-grained matter is practically absent in such deposits, the coarse-clastic layer is well aerated, and in some places the depth of thawing reaches 2.5 m or deeper, while at the foothills, ice is found under the rocks near the earth surface, even at the end of summer (Alekseev et al. 2011; Mikhailov 2013).

The site 2 (Figure 2) includes several small typical catchments in the Pravaya Sokolovka R. basin. The youngest deposits are represented there by acid late Cretaceous effusive rocks (tuffs, tuff sands, ignimbrites and rhyolites). Their thickness reaches 600–650 m. The upper fractured zone, due to weathering processes, can be traced to a depth of 30 m in valleys and deeper – in watersheds. Deeper fracturing is associated with faults. The right side of the Pravaya Sokolovka R. valley is mainly formed by the Jurassic and Triassic sedimentary rocks (sandstones, siltstones with fragments and blocks of limestones and flint) (Triassic… 2004). Effusive rocks of basic composition have a local distribution. Generally, the Sikhote-Alin hydrogeological massif is characterized by a wide spread of fissured and fissured-veined water. A more detailed landscape characteristic of this territory is given in one of our works (Boldeskul et al. 2016).

The site 3 (Figure 1) is located in the Pad' Vas'kova R. basin that joins the Rudnaya R. mouth reach and falls into the large structural-facies zone – the Coastal anticlinorium (Khanchuk et al. 1995). The rocks in that area are sedimentary siliceous terrigenous ones of the Carboniferous, Permian and Mesozoic ages. The main part of this territory is underlain by effusive rocks, intrusions and volcanic-sedimentary formations of the Upper Cretaceous and Paleogene age. The effusive rocks (quartz porphyrites and rhyolites, their lavas and tuffs) have medium and acidic composition, and intrusive rocks are Paleogene granitoids that somewhere create large tracts through the effusive rocks [Arzhanova & Elpatyevskii 1990].

The authors performed more than 70 determinations of flow rates in some of identified slope subsurface streams. The measurements were carried out using the method of indicator mixing, or the ion flood method, which is used in the cases of flow rate measurements in the boulder-stony mountain river channels (Karasev & Shumkov 1985). A widely common artificial tracer – NaCl solution with a given concentration, usually equal to 100 g/l, was used. Durable continuous (6 hours to 1 day) registration of water mineralization in the flux in the observation pit was accomplished with the YSI Professional Plus water quality control equipment. The time step of recording was established as 5, 10 or 20 seconds as depends on the flow rate and the distance of observation pit from the place where the tracer was filled, whereas the solution volume ranged from 0.5 to 2 liters. In the Kontaktovyi Creek basin only speed of subsurface flux was measured by the float method.

To avoid errors in the calculations, those experimental results when the hydraulic connection of the stream at the place where solution was filled with the stream in the control site was reliably established were used. The measured discharges and velocities are given in the Table below.

Table

Dynamic characteristics of slope subsurface streams from the field measurements. Numerator is the arithmetic mean; denominator is the limits of variation.

Name of slope stream	Location	Number of measurements	Discharge, l/s	Flow velocity, m/s
South Sikhote-Alin Mountains, headwaters of the Ussuri/Wusuli R., Primorskii Krai, Russia				
Botanical drain	Pravaya Sokolovka R. basin	7	$\frac{0,48}{0.148–1.19}$	$\frac{0.004}{0.002–0.008}$
Antropogenic drain	Pravaya Sokolovka R. basin	16	$\frac{1.19}{0.056–6.95}$	$\frac{0.010}{0.001–0.042}$
Zhil'tsov drain	Pravaya Sokolovka R. basin	10	$\frac{1.24}{0.15–4.81}$	$\frac{0.018}{0.003–0.039}$
Deep-forest drain	Pravaya Sokolovka R. basin	12	$\frac{0.23}{0.041–1.55}$	$\frac{0.007}{0.002–0.030}$
South Sikhote-Alin Mountains, lower reach of the Rudnaya R., Primorskii Krai, Russia				
Marmalade drain	Pad' Vas'kova R. basin	3	$\frac{0.82}{0.504–1.27}$	$\frac{0.022}{0.012–0.034}$
Wafer drain	Pad' Vas'kova R. basin	2	4.12–6.62	0.020–0.030
Kissel drain	Pad' Vas'kova R. basin	2	5.85–7.36	0.031–0.050
Marchpane drain	Pad' Vas'kova R. basin	3	$\frac{1.94}{1.51–2.44}$	$\frac{0.020}{0.008–0.029}$
Dry Creek	Pad' Vas'kova R. basin	1	3.91	0.072
Headwaters of the Kolyma R., Upper Kolyma highlands, Magadan Oblast, Russia				
Squirrel drain	Kontaktovyi Creek basin	4	Not measured	$\frac{0.006}{0.005–0.008}$

So, the mean discharges measured vary in the range 0.25–6.61 l/s at average flow velocities of 0.4–7.2 cm/s. These values are comparable to the flow rates in the adjacent 1-order open streams in low water periods. It should be noted the Dry Creek is considered a transitional form from a slope subsurface stream to an open watercourse, and therefore its dynamic characteristics in the series of data obtained are expected to be maximum among all streams observed.

3. THE DISCUSSION OF THE RESULTS

The most significant and apparently regular in the frost-free period subsurface streams in slopes have a constant position and are confined to linear zones of slope flow concentration (preference) – natural underground channels, or *drains*. Even a little information obtained when examining a dozen subsurface drains reveals their association with well-washed coarse clastic deposits of effusive rocks elongated downward the slopes – block streams (kurums) described in the literature (Khudyakov et al. 1972; Tyurin et al. 1982; Korotkii 1984; Govorushko 1986; etc.).

The block stream is a linearly stretched downward the slopes a cluster of stone blocks and rubble (colluvium), usually slowly moving down under the influence of cryogenic deserption, solifluction, crushing and other processes (Geological… 1978; Mudrov 2007). The watercourses is often found under the lumps. Block streams are widespread in goltsy altitudinal belt and often descend rather far down into the forest belt in the mountainous countries of Pacific Russia. They are far from always distinctly definable in relief that complicates seriously their detection in space images. In particular, we have found weakly cut divergent double thalwegs that determine the W-shaped cross-sections of the block stream "hollows" in the slopes, while the block streams themselves perform convex cones framed by thalwegs and reach dozens of meters in width at their foothills.

It is necessary to distinguish block streams from a typical decolmatized washout fractolite eluvial horizon where the amount of rock fragments larger than 10 cm is 38–52%. The latter is discovered in the undisturbed soil profile of mountain landscapes (Arzhanova & Elpatyevskii 1990). Block streams have much higher thickness and predominantly boulder composition (block size more than 30 cm is predominant), which indicates a much higher permeability. Outcropped by road notches in the upper reaches of the Ussuri/Wusuli R., such block streams are U- or V-shaped, practically devoid of sandy-loamy aggregates, blocky lenses in the detrital deposits (Figure 2 left). The width of the lenses varies from 1–3 m to the first dozens of meters, thickness – within 1–2 m. In the upper parts of some slopes, there are slightly faded, coarse-grained deposits washed from fine particles.

Figure. 2. Left: a tilled stone stream in the headwaters of the Ussuri/Wusuli R., Southern Sikhote-Alin (outcrop in the roadway). Photo by V. Shamov. Right: the block stream in the Kontaktovyi Creek valley (the Squirrel drain), Upper Kolyma Highlands. Photo by A. Tarbeeva.

In tundra and woodland landscapes in the Upper Kolyma Upland, block streams are usually open (Figure 2 right) and are obviously confined mainly to dike and stock-like intrusive Early Cretaceous age bodies, which according to (Govorushko 1986) are often found in this region.
Our observations reveal fine soil matter accumulates in the lower part of the block stream body profile to provide a weakly permeable surface, on which percolating water quickly concentrates and flows down the slope. Due to this, subsurface streams, or drains, flow freely in block stream bodies, through the PFP network, or the network of drains.
Any block stream has a recharge zone, a transportation zone and a substance accumulation zone (Polunin 1989), and these zones can differ in specific facies and sub-facies structure (Tyurin et al. 1982), which determines the slope relief. The genesis of block streams, their evolution and their connection with geotectonic and geomorphological processes on the slopes is considered an independent task (Korotkii 1984; Tyurin et al. 1982; Suzuki et al. 2013; Seto et al. 2015; Tarbeeva et al. 2015). According to (Polunin 1989; Gavrilov 2006) gravitational geomorphological processes (including block streams) are closely related to the crushing, cataclasis, mylonization of rocks in fault zones, on the boundary of differentially moving blocks. Outside the river valleys and hollows, these zones also determine the places of migration and discharge of fissured underground water and form the cones of removal in foothills and the

linear weathering crusts. The latter according to (Geological… 1978) appear along tectonic fractures or on contact of rocks that are different in composition. The linear weathering crusts have the shape of elongated veined bodies. Such bodies extend in plan in hundreds of meters, and to a depth – usually in several dozens of meters or more. Tyurin et al. (1982) and Korotkii et al. (1976) indicated the assignment of block streams (kurums) to the tectonic fracture zones and frost cracks stretched down the slope. Notably, in the conditions of permafrost or long-term seasonal frost in frost cracks that penetrate to considerable depth into rocks inheriting their primary tectonic fissuring, the slope water definitely concentrates and moves down (Glotova & Glotov 2012; Tyurin et al. 1982).

It is known that the regional structure of hydrogeological reservoirs is genetically and spatially related to long-lived tectonic dislocations and their nodes, which determine zones of connection of hydrogeological massifs and artesian basins (Sorokina 1992, 2006). It should be assumed that small slope-scale faults which frame local faults can, in turn, govern stable subsurface water streams on slopes. Block streams on the valley slopes, in our opinion, can indicate such "sub-local" faults and, consequently, a most significant subsurface drainage network, in which rapid concentration and discharge of slope water into rivers occurs. Evidently, it occurs owing to good permeability of well-washed lumpy deposits and the appearance of shallow impermeable surfaces folded by finely dispersed matter accumulating in the bottom of the block stream bodies' profile. As our experiments showed the flow velocities and rates are quite comparable with those in small open watercourses.

Based on the results of our observations and accounting the generalized data on the regime and structure of recharge plots at the karst areas in Piedmont, Italy (Galeani et al. 2011; Vigna & Banzato 2015) and also estimates of mean transit time of slope flow into watercourses in Rocky Mountains, Oregon, USA (McGuire & McDonnel 2010) we propose to divide the sub-surficial PVP network into three main structural groups: 1) a system of drains, or main conductors, or conduits; 2) a system of filtration flow domains; and 3) a system of matrix flow paths.

The drainage network of the first type provides the most rapid response (in a few hours to the first dozens of hours) of a watercourse to a storm, and, as a rule, their warming and dilution with rainwater entered. In this network the slope water freely flows, and according to our data, after a series of heavy storms (with a total of 120–150 mm), the water volume in such conduits can increase by 2 orders or more.

The system of filtration domains collects and pours out water more slowly (along several to many days), displacing the previously accumulated water reserve according to the piston effect principle (Uchida et al. 2005; Galeani et al. 2011). The total dissolved substances in river increases at the same time as more cold "old water" is displaced by leaking "new water" due to the sufficient time for leaching of these substances from the rocks drained and for lowering the temperature (Vigna & Banzato 2015).

A system of matrix (disperse) drainage paths, obviously, corresponds to the PFP in the soil profile scale (Umarova 2011; Angermann et al. 2016). It determines the least mobile part of soil water and, generally speaking, is practically insensible in the response of rivers to inter-annual and multiyear precipitation regime. The velocity of water movement in the system of dispersed paths is about 0.01 mm/s (Angermann et al., 2016).

Taking into account the classifications of rock fractures (Stepanov 1989), there is reason to believe that the block streams, and, consequently, the associated systems of subsurface drains, are probably confined to the faults of the slope scale. According to Stepanov (1989) length size of such fractures in this case are estimated in hundreds and first thousands of meters, and width – the first meters. Domain PFP systems are evidently concentrated in the fractures (macropore flow (Weiler 2017)) and, probably, soil pipes (pipeflow (Uchida et al. 2005)) of various origins within the eluvial soil horizons. The absolute length of such cracks ranges from 10 cm to the first dozens of meters (in basins of fracture waters), and the width is 0.1 to dozens of centimeters (Stepanov 1989; Gavrilov 2006). The systems of dispersed flow paths are likely controlled by a network of micro-cracks and capillary pores that are much less than 10 cm in length and less than 0.1 cm in width (Stepanov 1989), providing the form of “festoons” (or separate “tongues”) in the vertical percolation front within a soil profile (Umarova 2011).

4. CONCLUSION

The data obtained and analyzed above show that the slope flow occurs in the form of a practically constant network of subsurface streams – drains – controlled by a system of preferential flow paths within weathering crust. The largest of these paths, in terms of their dynamic characteristics, are comparable to 1-order surface watercourses in mountain regions during low water periods. They obviously represent a main form of interaction between slope flow and riverbed flow.

The largest slope preferential flow paths are confined to block streams – narrow stripes of coarse clastic deposits, often associated with faults and fractures mostly of tectonic and/or frost origin. Those block streams are not always visible in relief, and it is difficult to identify them by means of remote sensing under dense vegetation canopy.

To develop research in this direction, adaptation and development of technical and model aspects of the tracer methodology (Gartsman & Shamov 2015; Gubareva et al, 2015) as well as geographical extension of studies of slope preferential flow paths are expected. The further detailed investigation of texture of slope deposits and also hydrological and hydrochemical conditions in mountain slope streams would make it possible to formulate a hypothesis on the significance of active and relict block streams in the concentration of slope flow and, with that, rain floods generation processes in mountain catchments, especially in areas with severe seasonal moisture fluctuations as, i.e. in Pacific Asia.

5. AKNOWLEDGMENT

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CHARACTERISTIC FEATURES OF THE WATER-EXCHANGE FUNCTION OF PERMAFROST

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ABSTRACT
This paper highlights the importance of understanding the water exchange function of permafrost, as well as its role in the formation of surface and ground water resources and regime. The role of the active layer and its long-term dynamics in the water cycle and its direction and magnitude is discussed. Geocryological processes which increase the water-exchange function of permafrost are briefly described.

KEYWORDS
Permafrost; water exchange; water phase changes; climate

1. INTRODUCTION
Permafrost, which underlies about a quarter of the Earth's land surface and extends to depths of 1.5 km or more in some areas, is estimated to contain about 400,000 km^3 of ground ice. However, it is not only a great storehouse of ice, but also a highly dynamic, open dissipative system. The main characteristic feature of permafrost as a cryogenic geosystem is the phase changes of water which dictate to a large extent its complex structure, as well as the high dynamics and specificity of processes occurring in it. In this system, water changes its phase from liquid to solid (ice formation or freezing), solid to liquid (melting), gas to solid (desublimation or deposition), solid to gas (sublimation), liquid to gas (evaporation), and gas to liquid (condensation).

2. WATER-EXCHANGE FUNCTION OF THE ACTIVE LAYER
The water-exchange function of permafrost is related primarily to the phase changes from water to ice and back to water, i.e. with freezing of water-saturated sediments and thawing of ice-saturated sediments. In the upper part of permafrost, seasonal transitions of ground ice to liquid phase and back occur annually in huge quantities (Vtyurin 1975; Shumilov 1986; Kotlyakov 2002; Alekseev 2005). The total volume of water from melting of the ground ice accumulated in winter in the active layer of permafrost is estimated to be about $4 \bullet 10^{12}$ m^3, or 3.3 times the water contained in all rivers of our planet (Shepelev 2011). A significant portion of this water forms various types of suprapermafrost water which is lost by transpiration and discharged to rivers and lakes. Our estimates indicate that the suprapermafrost water flow rate is $19 \bullet 10^9$ m^3/day, or 220,000 m^3/sec. For comparison, this value is over three times the total annual mean discharge of the Lena, Yenisei, Ob, Amur and Volga, the largest rivers in Russia and in the world.

Thus, the water-exchange function of the uppermost layer of permafrost governed by the seasonal phase changes of water from liquid to solid and back is very significant. For this very reason, it seems appropriate to distinguish a cryolithogenic component in the hydrologic (climatic) circulation of natural waters, which involves seasonal changes of groundwater from liquid phase to solid and back in the active layer of permafrost.

3. INFLUENCE OF LONG-TERM PERMAFROST DYNAMICS ON WATER-EXCHANGE PROCESSES

Far more significant and interesting is the water-exchange function of permafrost related to its long-term dynamics under the effect of large-scale climate fluctuations. The thickness and extent of permafrost increase significantly during colder climatic periods (cryochrons) and decrease during warmer periods (thermochrons). In West Siberia and East Siberia, for example, about 20 fluctuations in permafrost distribution (cryocycles), each lasting 2,000 to 40,000 years or more, are thought to have occurred during the last 800,000 years (Fotiev 2005).

The last strong cooling occurred in the Sartan cryochron (37-11 kyr ago) referred to as the main climatic minimum in the Pleistocene (Zubakov 1986). This cooling peaked at about 18 kyr ago when permafrost occupied approximately $110 \bullet 10^6$ km^2 of the Earth's land surface, covering nearly all of Europe, most of the Asian continent and North America. This period was followed by a general warming trend with its maximum at about 7-6 kyr ago (Holocene Climatic Optimum). During this thermochron, the southern limit of permafrost shifted to the north by 1,300 to 1,500 m, permafrost completely degraded, and extensive water-bearing talik zones formed up to 150-300 m in thickness.

It is roughly estimated that about $4.5 \bullet 10^{15}$ m^3 of ground ice was transferred to liquid water during the Holocene thermochron. The input of water from melting ground ice to surface and subsurface runoff and storage was on the order of 820 km^3/yr. This is comparable to water-exchange intensity in the lithogenic, metamorphogenic and magmatogenic components of the geological water cycle. Considering this fact, the author proposed earlier that a cryolithogenic component be added to the geological water cycle (Shepelev 2000, 2008).

The interaction of groundwater with surface water under large-scale climatic changes occurs in a very specific way. During colder periods, freezing of aquifers causes part of the groundwater to be expelled away from the freezing front, since the volume of ice is about 9% greater than that of liquid water. This crystallization-and-compression effect leads to a considerable increase in hydrostatic pressure in freezing aquifers and rising of the piezometric surface of subpermafrost water. The increase of hydrostatic pressure in the subpermafrost zone during colder periods, as shown by a modelling study, can be up to 32-34 MPa (Balobaev 2003). This causes a considerable increase in subpermafrost water discharge through open taliks existing beneath major rivers and lakes (Figure 1).

During long-term warming periods, the phase boundary (frozen ground – groundwater) moves toward the surface accompanied by lowering of hydrostatic pressure in the subpermafrost zone. As a result of this crystallization-and-vacuum effect, large depressions of the piezometric surface are formed in subpermafrost aquifers, intensifying both the horizontal groundwater flow and infiltration of surface water through open talik zones. The recharge of subpermafrost water from infiltration through open taliks may reach 40-60 m^3 per 1 km^2 of talik area during warming periods (Balobaev 2003).

Thus, long-term climate cooling results not only in the transition of tremendous amounts of groundwater into a solid phase, but also in the depletion of subpermafrost aquifers due to the increased discharge through open talik zones by the crystallization-compression effect. During periods of long-term warming, the subpermafrost water storage is recovered both by melting ground ice and increased recharge by infiltration through open taliks. This is the active water-exchange role of the latter that prevents their freezing even during periods of significant and lasting climate cooling.

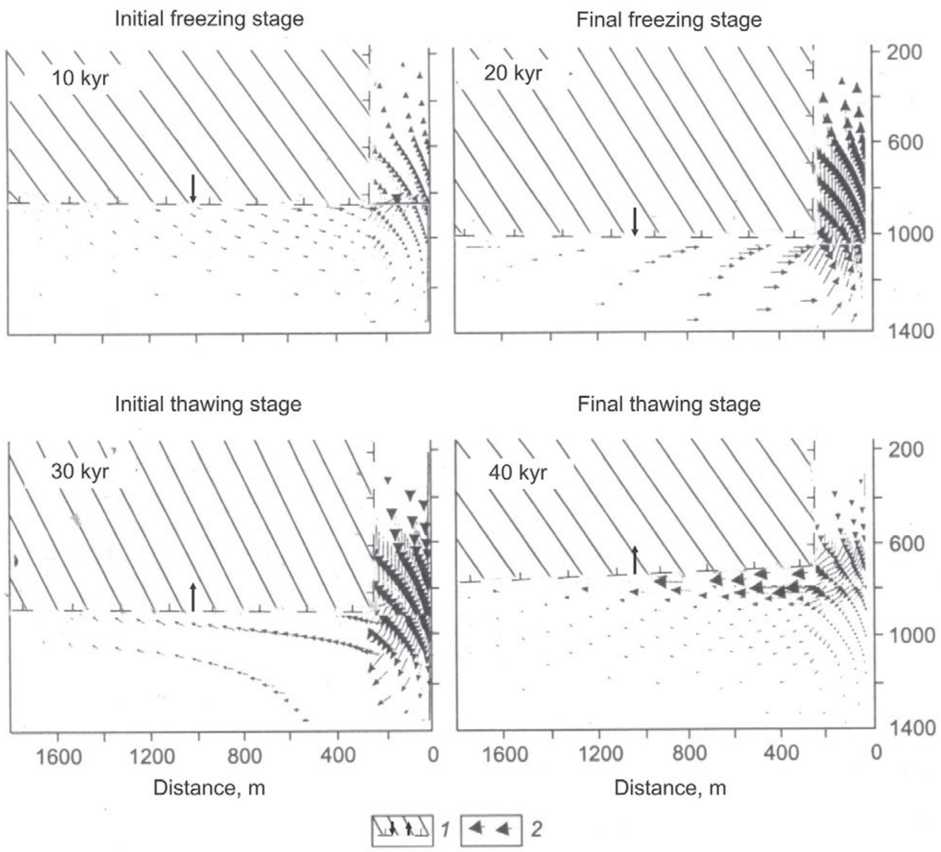

Figure 1. Groundwater flows through an open talik zone at different climatic periods with cyclic changes in permafrost thickness. (Schematized after Balobaev (2003).
Minimum thickness – at 0 and 40 kyr, maximum thickness at 20 kyr.
1 – permafrost, its boundary and movement direction; 2 – direction of water movement; density and size of arrows are proportional to groundwater flow rate.

4. GEOCRYOLOGICAL PROCESSES INCREASING THE WATER-EXCHANGE FUNCTION OF PERMAFROST

The high water-exchange activity of permafrost is obviously related to the fact that the periodic changes of groundwater to a solid phase and back increase fissuring and effective porosity of soils and rocks, increasing their hydraulic conductivity. Frost destruction of soils and rocks is most intensive in the upper layers of the ground where seasonal transitions between liquid and solid water phases occur. Widespread frost fissuring promotes the development of a peculiar polygonal-localized type of recharge and discharge of the active-layer waters, intensifying their interaction with surface waters.

Frost destruction occurs also in the middle and lower parts of the permafrost section, resulting in the zones of secondary fissuring (frost disintegration). Periodic fluctuations in the lower boundary of permafrost and in the plan dimensions of open taliks lead to the formation of highly watered zones at the contact between thawed and frozen ground, which promotes greater water exchange in the hydrogeological structures and better interaction between subpermafrost water and surface water (Alekseev 2009; Fotiev 2009).

Other phase changes of water, along with freezing and thawing processes, play some role in the water-exchange function of permafrost. However, their role remains poorly understood. Limited available studies suggest that such processes as desublimation (ablimation), evaporation and condensation are quite intensive in permafrost. Water-balance observations in Central Yakutia, for example, have shown that about 40 to 60 mm of water is accumulated as desublimation ice in winter in the active layer consisting of fine to coarse sands (Shepelev 2011). In the subpermafrost zone, ice may be formed by desublimation process where the unconfined groundwater table lies below the permafrost base, as, for example, in some high-mountain hydrogeologic massifs and sub-massifs.

5. CONCLUSIONS

It is worth noting in conclusion that the active water-exchange role of permafrost associated with the phase changes of water has a tremendous effect on the qualitative composition of groundwater and surface water. Crystallization of liquid water and subsequent melting of ice, for example, lead to significant changes in the chemical, gas and isotope compositions of water, improving, in particular, its drinking quality and biological activity. The cryolithogenic components in the climatic and geological water cycles present unique, very effective and highly productive natural mechanisms for continuous water purification and regeneration. All this highlights the need for further comprehensive research on the water-exchange function of permafrost.

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LONG-TERM FLUCTUATIONS OF ARCTIC RIVER RUNOFF IN CENTRAL SIBERIA UNDER A CHANGING CLIMATE

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ABSTRACT
This article presents the results of analysis of the long-term runoff fluctuations of the medium and small rivers, which belong to the lower Yenisei and Pyasina watersheds in the Arctic Zone of the Central Siberia. Dataset of annual and seasonal river runoff was formed from 11 hydrological stations, located within forest-tundra and tundra zones. Using graphical and statistical methods it was found that fluctuations in the river runoff occur asynchronously between the forest-tundra and forest zones.
There was a transition from a period with the annual runoff below normal to the period with the runoff higher than normal in the 1987/1988 hydrologic year, this regularity is common for most of the rivers within the forest-tundra zone. High-water period was observed until the mid-1980s, and then it was changed to a period of low-water period until 1995 for the rivers Dubches and Yeloguy within the forest zone.
Trends over the entire observation period and trends for the period since 1981 have mostly positive sign for the forest-tundra rivers, while trends have a negative sign for the forest rivers. Asynchronous nature of runoff fluctuations proves the usefulness of water resource border of the Arctic zone for future integrated assessments of water resources under the influence of climate change. Small and medium-sized rivers of the North Central Siberia are characterized by different trends in the dynamics of annual runoff, but there is widespread increase in winter runoff. This result confirms earlier studies on the variability of the Siberian rivers winter runoff.

KEYWORDS
Central Siberia, water resources, river runoff, rivers, Arctic Zone, long-term variability, climate change

1. INTRODUCTION
Surface water bodies contain water resources that are used in a various economic branches. Quantitative and qualitative characteristics of water resources are the result of natural transformations and anthropogenic impact.
The fact of the global climate change is defined in modern studies. The Arctic region is most vulnerable to climate change. There are estimates of a very significant warming in the Arctic with the mid-20th century and a higher rate of warming compared to other regions of the planet. Natural transformations are reflected by changings in sea level, sea ice area, snow cover height, river runoff, permafrost conditions, frequency of dangerous natural phenomena (Kattsov, Porfiryev 2012; IPCC 2013; Roshydromet 2014; Alekseev 2014). The river runoff is the basis for assessing the quantitative changes in resources. Early studies provided by Ivanov et al. (2004) characterize the variability of river runoff in the Arctic Ocean by different trends. Other researchers (Peterson et al. 2002; Shiklomanov I.A., Shiklomanov A.I. 2003; Berezovskaya et al. 2004) formulated a conclusion on the increase in runoff of large Arctic rivers from the 70s of the XX century, especially in winter runoff.

As a result of the state monitoring of the Arctic since the beginning of the 20th century long-term information has been accumulated. In the context of climate change, the study of the dynamics of water resources is important for making managerial decisions.
The purpose of the article is to identify the spatio-temporal changes in river runoff of medium and small rivers in the north of Central Siberia.
Research on medium rivers of the north of Central Siberia are represented only in some modern works on the runoff fluctuations (Bolgov & Korobkina 2012; Rumiantseva 2012; Burenina et al. 2015).

2. DATA AND METHODS

The observation data of the hydrological posts of the Roshydromet were used in this study. The data were collected from the State Water Cadaster editions.
The choice of the hydrological posts is defined by the presence of data, which includes the current climatic period from 1981. For the analysis, information was used for medium and small rivers. The variability of medium rivers runoff is defined by zonal factors, including climate, while rivers with a small drainage area are more susceptible to the influence of local factors. Due to comparative analysis of long-term runoff variability of the medium and small rivers it is possible to assess the scale of the impact of climate change on rivers with different water resources.
The dataset of seasonal and annual runoff was formed for 11 hydrological posts located in the state border of Arctic zone of Russian Federation according to the Presidential Decree (2014). Table 1 lists the hydrological posts on which data were used. It should be noted that the Dubches and Yeloguy Rivers do not belong to the Arctic zone according to water resource border (Ivanov & Tretiakov 2015).

Table 1

Hydrological posts on the study arctic rivers in the Central Siberia

№	Code	River – hydrological post names	Drainage area, km^2	Period of observation, time series length (year)
1	09455	Norilka – Valek	19800	1938 –2013 (76)
2	09432	Khantayka – source	12300	1970 –2001 (32)
3	09550	Tukalanda –Tukalanda	860	1984 –1996 (13)
4	09445	Kulyumbe –Kulyumbe	3160	1982 –1995 (14)
5	09183	Kureyka –«Ozernaya»	28100	1976 – 1990 (15)
6	09498	Gorbiachin–Gorbiachin	3670	1981 –1997 (17)
7	09431	Graviyka – Igarka	323	1938 –1992 (55)
8	09427	Sovetskaya Rechka – Sovetskaya Rechka	1430	1973 –2012 (40)
9	09425	Turukhan – Yanov Stan	10100	1941 – 2013 (73)
10	09401	Yeloguy – Kellog	16300	1960 –2012 (53)
11	09385	Dubches – Sandakches	8360	1964 – 2013 (50)

The figure 1 shows the location of the hydrological posts and boundaries of the Arctic zone on the territory of the Central Siberia.
The investigated rivers belong to the basin of the Kara Sea. Some rivers flow down from the western slope of the Putorana Plateau. Basically, rivers are located in the forest-tundra zone. The rivers Dubches and Yeloguy belong to the forest zone. The basin of the Turukhan River is located on the border of these zones.

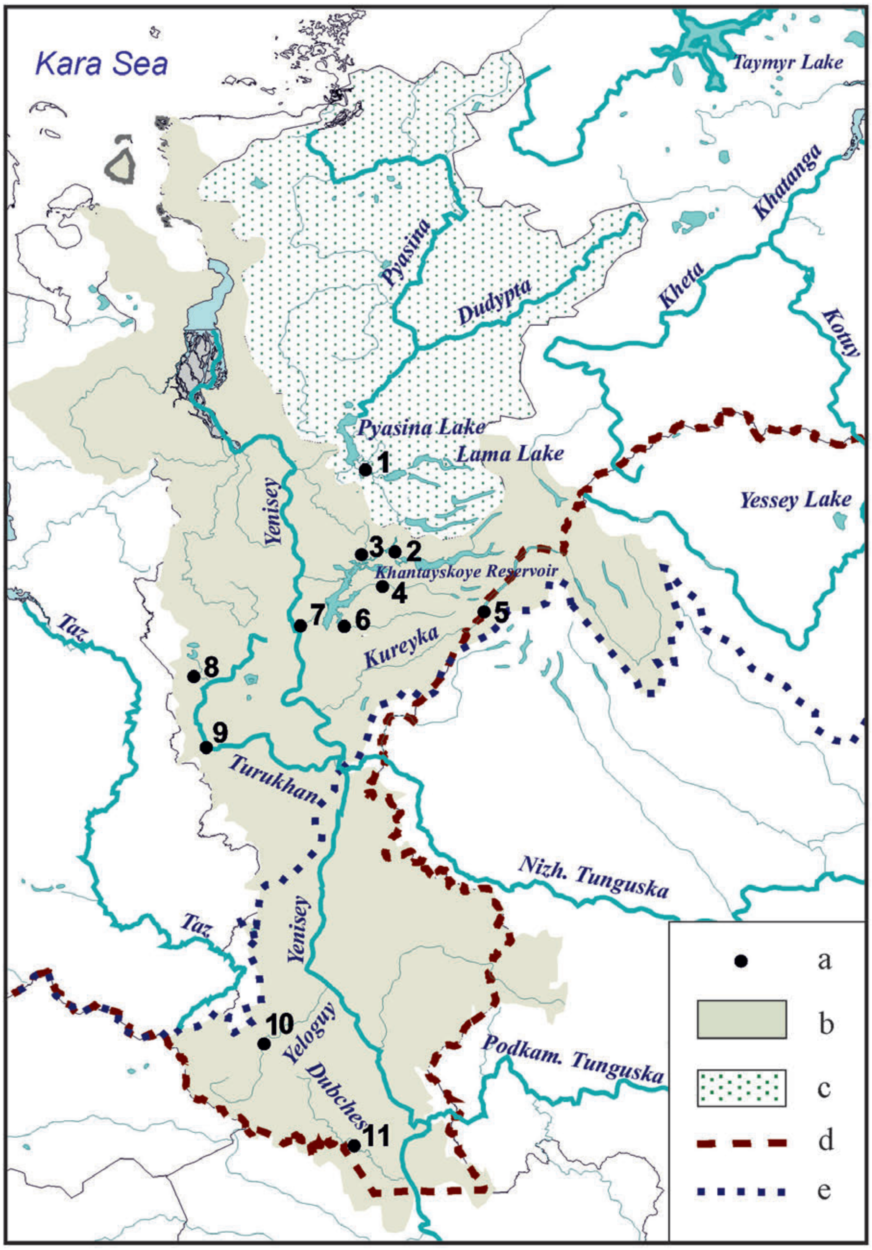

Figure 1. Map of the Arctic zone in the Central Siberia
a – hydrological posts (according to the table 1); b – drainage area of the lower part of the Yenisei, excluding the basins of the Nizhnyaya Tunguska and Podkamennaya Tunguska; c – drainage area of the Pyasina; d – the state border of the Artic zone of Russian Federation; e – water resource border of the Artic zone of Russian Federation

The hydrological regime of the rivers in the north of Central Siberia is characterized by spring flood, long period of freeze-up and prevalence of snow feed. Permafrost affects the amount of underground runoff resulting from seasonal thawing. There are many lakes in the drainage areas; some of them are regulated by rivers.

Data gaps were recovered using the method hydrological analogy. Graphical methods of difference integral curves and smooth curves as well as statistical methods of linear trends evaluation and their significance were applied for data analysis and identification of spatio-temporal changes. Mathematical processing of observational data was carried out using specialized software for hydrological calculations

3. RESULTS AND DISCUSSION

As a result of processing data of river runoff in the north of Central Siberia, dynamics over a long period of time has been discovered. For the analysis, the volume of river runoff was calculated for the hydrological year, the beginning of which was determined from October 1st. The analysis of the runoff for the cold (winter) period was carried out according to the water volume for October-April.

The main characteristics are calculated for all analyzed series of annual runoff: mean annual, minimum and maximum, coefficients of variability (C_v) and skewness (C_s). The results of calculations are presented in Table 2.

Table 2

The basic statistical characteristics of the long-term series of annual water runoff

River	Runoff volume, km^3	Minimum, km^3	Maximum, km^3	C_v	C_s
Norilka	14,2	6,96	21,8	0,20	0,09
Khantayka	8,52	5,79	13,1	0,19	1,23
Tukalanda	0,64	0,39	0,98	0,26	0,68
Kulyumbe	2,13	1,44	2,89	0,20	0,22
Kureyka	11,4	7,93	16,6	0,19	1,12
Gorbiachin	2,28	1,70	3,01	0,17	0,45
Graviyka	0,16	0,06	0,23	0,23	-0,16
Sovetskaya Rechka	0,49	0,25	0,65	0,17	-0,33
Turukhan	3,51	2,04	4,90	0,16	0,09
Yeloguy	5,07	3,93	6,75	0,14	0,52
Dubches	2,78	1,64	3,48	0,13	-0,19

The most full-flowing rivers are the Norilka, Kureyka, and Khantayka Rivers. The rivers of low water content include the Graviyka, Sovetskaya Rechka and Tukalanda Rivers. In general, the variability of annual runoff is not high. The coefficient of variability does not exceed 0.3. The greatest variability of the runoff is typical for the small rivers Tukalanda and Graviyka. Their coefficients of variability are 0.26 and 0.23 respectively. The coefficient of skewness of the annual runoff varies from -0.33 to 1.23. A significant coefficient of asymmetry (± 0.5) is determined for the rivers Khantayka, Tukalanda, Kureyka and Yeloguy.

The homogeneity of the annual runoff series is confirmed by the criteria of Student and Fisher. An exception is the heterogeneity of the Norilka annual runoff in terms of variance. It should be noted that there was an increase in the frequency of positive flow deviations from normal for the Norilka in the early 1990s. There was a decline in water flow in 2013, which decreased by half, so there is a change in the variance of the time series. The inhomogeneity is also defined for the Turukhan River on the average value, which is due to a gradual increase in the volume of annual runoff in the late 1980s.

Analysis of long-term variability of annual runoff by difference integral curves shows the general trend of changing low-water and high-water periods for the rivers of the forest-tundra zone. Figure 2 shows the difference integral curves of the annual river runoff in the north of Central Siberia.

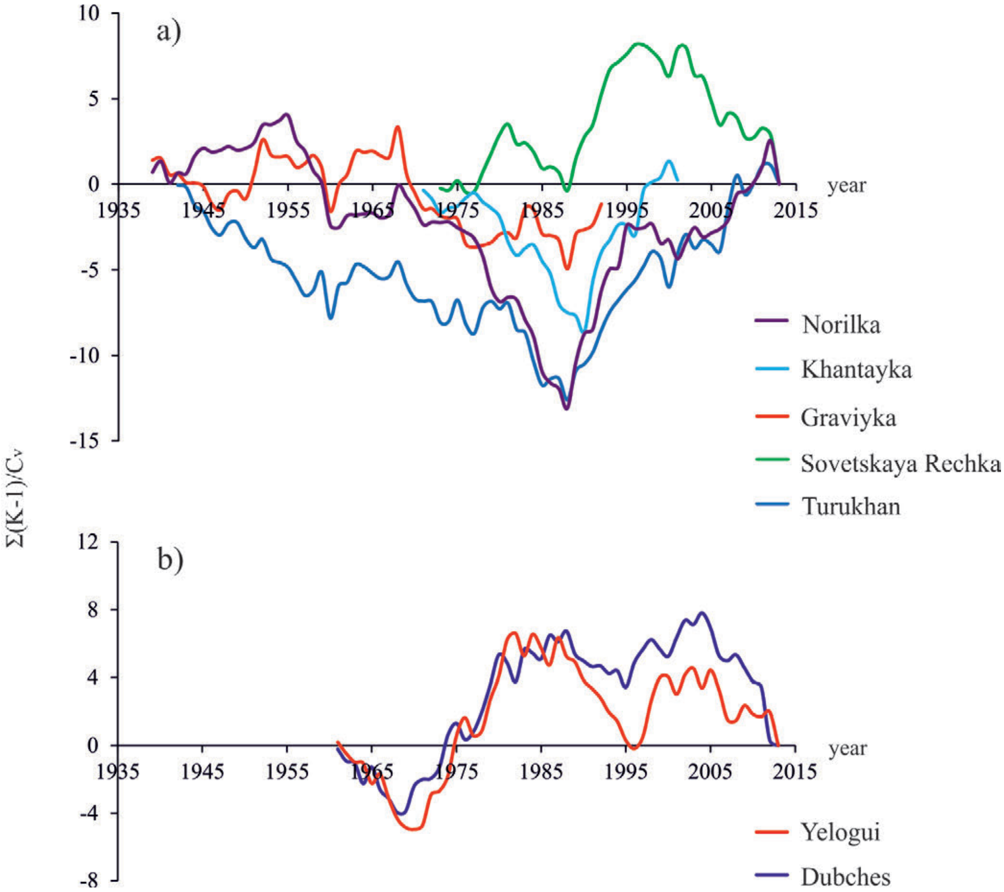

Figure 2. Difference integral curves of annual runoff volume of:
a) rivers of forest-tundra zone, b) rivers of forest zone

Three periods of water availability have been identified for the Norilka during the whole period of instrumental observations: the high water periods of 1938-1955 and 1988-2013, the low-water period of 1956-1987. The low water period is characterized by an abrupt decrease in the river runoff. Low water period of 1940-1987 and modern high-water period are defined for the Turukhan River. The general trend of the change of low-water and high-water periods in the 1987-1988 hydrological year can be identified for the small rivers Graviyka and Sovetskaya Rechka, but it is difficult to reveal the clear boundaries between the water phases before this transitional year. In general, 1987-1988 hydrological year is a transition from low-water to a high-water period for the rivers of the forest-tundra zone.

There is a reverse trend for the rivers Yeloguy and Dubches of the forest zone. The high water period was observed until the mid-1980s and it was replaced by a low water period until 1995. The current high water period was replaced by a low-water period in 2004.

The applied method of difference integral curves has the main disadvantage, which consists in the dependence of calculations on the selected period. Nevertheless, the method allows to reveal the cycles of river runoff.

An analysis of long-term variability of annual runoff by the method of smoothing time series was performed. Figure 3 shows the graphs of moving average curves for 5 years for the rivers Norilka, Turukhan, Yeloguy.

There was the prevalence of positive anomalies in annual runoff in the 1990s for the Norilka and Turukhan Rivers. The Yeloguy is characterized by a shift in this period towards the end of the 1990s - the middle of the 2000s.

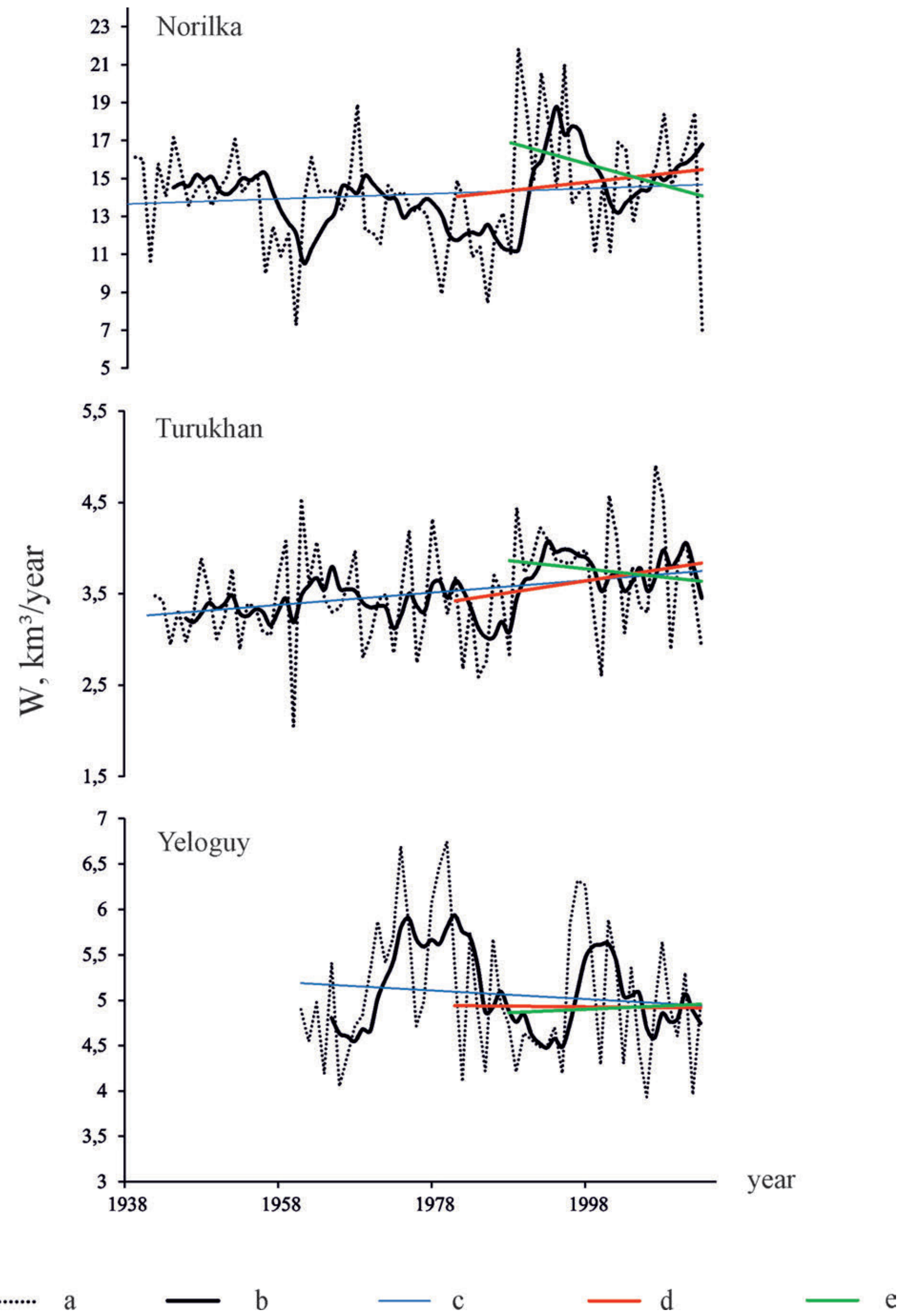

Figure 3. Long-term fluctuations in the annual runoff of the Norilka, Turukhan, Yeloguy Rivers, a - the chronological curve, b - the moving average curve for 5 years, c - the trend for the entire observation period, d - the trend for the period from 1981, e - the trend for the period from 1988.

Analysis of trends allowed us to identify general trends in fluctuations in annual runoff over the entire period of instrumental observations. Figure 3 also shows the linear trends of the annual runoff for the entire observation period and for individual periods. Trends were analyzed for the period from 1981 in accordance with the generally accepted time frame of the current climatic period (Roshydromet, 2014). The analysis was also carried out for the period since 1988 hydrological year.

The significance of river runoff linear trends was estimated at the 5% significance levels. Table 3 presents the results of these estimates for rivers with the longest observation periods.

Table 3

Assessment of linear trends in river runoff volume

River	Characteristic	Period	a,km^3/(year·10^3)	R^2·100	Significance
Norilka	Hydrological year	1939-2013	13,67	1,00	-
		1981-2013	54,79	2,00	-
	Cold period	1939-2013	-1,01	0,00	-
		1981-2013	20,77	5,00	-
Sovetskaya Rechka	Hydrological year	1972-2013	-2,14	9,00	+
		1981-2013	-2,96	8,00	-
	Cold period	1972-2013	0,70	10,00	+
		1981-2013	0,59	5,00	-
Turukhan	Hydrological year	1942-2013	6,77	7,00	+
		1981-2013	13,62	4,00	-
	Cold period	1942-2013	2,27	7,00	+
		1981-2013	9,66	21,00	-
Yeloguy	Hydrological year	1961-2013	-4,74	1,00	-
		1981-2013	-9,81	0,00	-
	Cold period	1961-2013	4,16	4,00	-
		1981-2013	5,31	3,00	-
Dubches	Hydrological year	1961-2013	-4,51	4,00	-
		1981-2013	-12,65	7,00	-
	Cold period	1961-2013	2,99	10,00	+
		1981-2013	2,85	3,00	-

Over a long period of more than 70 years along the Norilka and Turukhan rivers, a positive trend has been identified, which is also observed for the current climatic period. There is a negative trend in annual runoff during the high water period of 1988-2013. The negative trend of annual runoff is recognized both for the whole period of observations and for the current climatic period for the Yeloguy River.

Analysis of the significance of linear trends in the annual runoff of water shows that trends are not significant both for the entire observation period and for the current climatic period. The statistical significance of the trend was confirmed only for the Sovetskaya Rechka and Turukhan Rivers at a level of 5% for the entire observation period. There is a negative trend for the small river Sovetskaya Rechka, and there is a positive trend for the medium river Turukhan. Summer-autumn tendencies have the same trends as river runoff volume of the hydrological year.

Variability of the winter runoff is characterized by positive trends for all of rivers, including significant trends for the Dubches, Turukhan and Sovetskaya Rechka. Figure 4 shows the graphs of the relative change of annual and winter runoff for the current climatic period to the previous period.

4. CONCLUSION

The overall results show different trends in the dynamics of the annual runoff for small and medium rivers, but there is a general increase in winter runoff. The conclusion on the river runoff increase over the cold period is confirmed by early studies on the variability of the winter runoff of large Siberian rivers.

The statistical significance of the linear trends in the annual runoff has not been confirmed. The exception is the Turukhan and Sovetskaya Rechka. There is a significant tendency to reduce the annual runoff for the small river Sovetskaya Rechka.

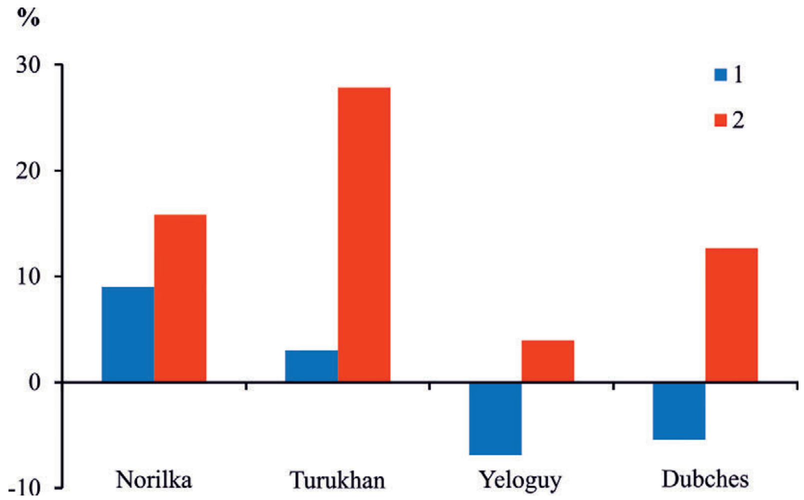

Figure 4. The relative change of annual (1) and winter runoff (2) for the period 1981-2013 to the period 1961-1980.

Asynchronous fluctuations of the annual river runoff in the forest-tundra and forest zones of Central Siberia were found. A transitional year was determined from the low-water phase to the high-water phase in 1987/1988 for the rivers of the forest-tundra zone. A similar increase in the river runoff in the late 1980s is consistent with the analysis of the current trends in the river runoff in Russia as a whole. There is an opposite tendency to a decrease in river runoff for the rivers of the forest zone in Central Siberia. This general conclusion can be considered as a quantitative confirmation of the usefulness of the water resource border of the Arctic zone to further estimates of climate change influence and water management planning.

5. ANKNOWLEDGMENT

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70 YEARS ANNIVERSARY OF THE KOLYMA WATER-BALANCE STATION, THE PIONEER OF HYDROLOGICAL RESEARCH IN PERMAFROST

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ABSTRACT
In 2017, 70 years have passed since the beginning of work at the Kolyma water-balance station (KWBS), a unique scientific research hydrological, hydrogeological and permafrost station. The volume and duration (more than 40 years) of the gained hydrometeorological standard and experimental data, characterizing the natural conditions and processes occurring in the permafrost zone, significantly exceeds foreign counterparts. The data are representative of the vast territory of the North-East of Russia. Since 1997 the station has been terminated, thereby leaving Russia without operating research watersheds in the permafrost zone. This article highlights the main historical stages of the station's existence, its work and scientific significance, and outlines the prospects for the future of the station, where the Kolyma water-balance station could restore the status of a scientific research watershed and become the international center for hydrological research in permafrost.

KEYWORDS
Permafrost hydrology, experimental research, North-East of Russia, Kolyma water-balance Station, KWBS, runoff, permafrost, active layer

1. INTRODUCTION
In 2017 we celebrate 70 years since works on organizing the Kolyma Water-Balance station (KWBS) began. This hydrological and permafrost research station has accumulated standard and experimental data unique both in terms of their amount and duration.

In the paper «Save northern high-latitude catchments» Laudon et al. (2017) recognize the KWBS as a currently functioning scientific station, even though scientific research was suspended here 20 years ago, and nowadays only standard observations at the meteorological site and the only water level gauge are carried out. As it is also mentioned by Laudon et al. (2017), number of scientific and hydrological research stations in the Northern regions of the world has decreased by 40%, and it happened alongside with the most significant climate change in the Arctic.

The KWBS is located in the headwaters of the Kolyma River, in a mountain region of continuous permafrost (Fig. 1). Runoff formation conditions at the station are representative for an immense territory of the upper Kolyma River basin and mountainous regions of the North-East Russia.

From 1948 to 1997 at the KWBS there were 10 hydrological gauges at catchments ranging between 0,27 and 21,2 km^2, two meteorological plots, 55 (in total) precipitation gauges, over 30 frost tubes (cryopedometers), several groundwater wells, evaporation, water-balance and runoff plots, also regular snow surveys were conducted, as well as experimental investigation of specific hydrological and permafrost processes.

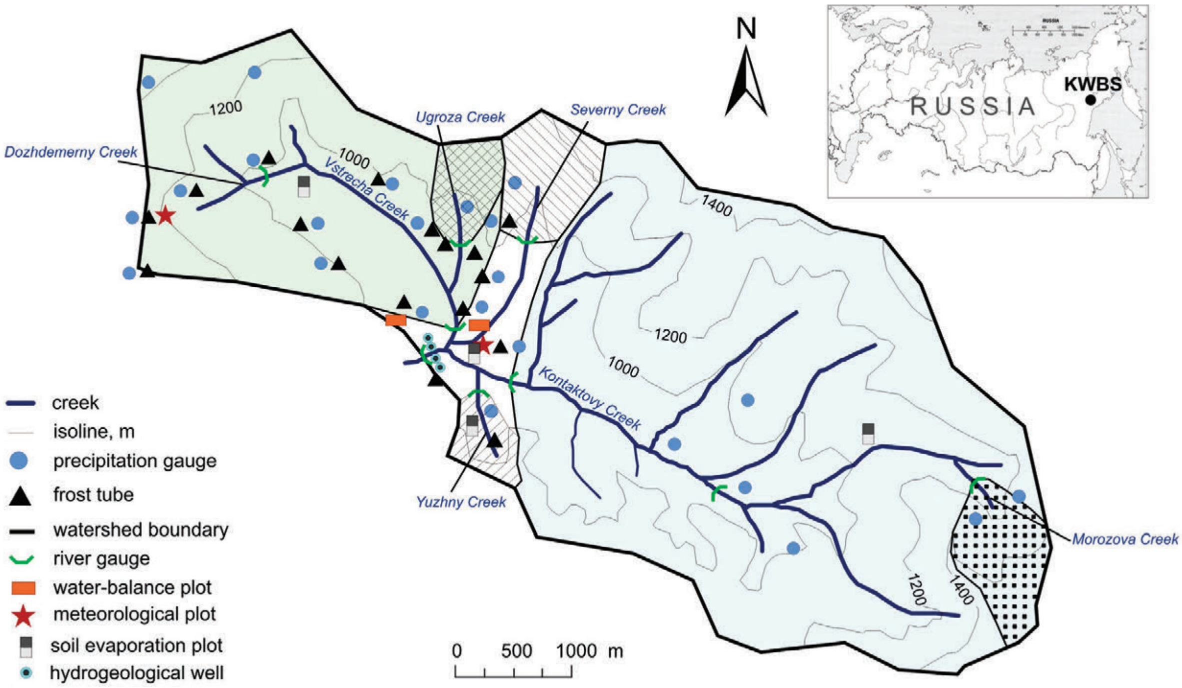

Figure 1. Scheme of the Kolyma Water Balance Station (KWBS) indicating the location of observation sites in 1988. Inset shows the location within Russia.

2. NATURAL CONDITIONS

The Kolyma water-balance station is located in Tenkinsky district of the Magadan region of Russia within Upper-Kolyma highland. The station's territory – the Kontaktovy Creek catchment – is a part of the Kulu River basin, which is the right tributary of the Kolyma River, and is located 16 km away from Kulu village settlement. It is characterized with a mountain landscape, typical for the upper reaches of the Kolyma River. The territory of the basin is severely cut up with creek valleys. These valleys are narrow, with steep slopes, watershed lines are mostly well marked. Absolute elevations of the basin range between 800 m asl near the Kontaktovy Creek outlet and 1700 m asl at watershed divides.

The climate is severely continental – with harsh long winters and short but warm summers. Average annual temperature at the Nizhnaya meteorological plot during 1948-1995 is -11.4 °C (Sushansky, 2002). Average monthly temperature in January is -33.7 °C, in July +13.2 °C. The absolute minimum temperature of -55.5 °C was registered in 1982 and the absolute maximum temperature is +32.0 °C (1951).

The period of negative air temperatures lasts from October to April, freeze-free period is, on average, 130 days long. Annual amount of precipitation at the catchments increases with the elevation of the basin, and averagely accounts for 314 mm (according to the data from Nizhnaya meteorological plot). Precipitation distribution is highly uneven through the year. About 70% of precipitation falls in warm season.

Snow cover is formed in the first decade of October, and melts in the third decade of May. Maximum snow water equivalent increases with the elevation and reaches 110-150 mm.

The station is located in the continuous permafrost zone. Permafrost thickness varies from 120 to 210 m in valleys and can reach 300-400 m in highlands, following the relief. Seasonal soil thaw depth depends on slope exposition, altitude and landscape and changes from 0.2-0.8 m on north-facing slopes to 1.5-3.0 m on south-facing ones.

KWBS is situated in the transitional zone between forest-tundra and coniferous taiga. Soil types vary from rock debris to clayey podzol with partially decayed organic material underlain by frozen soil and bedrock. Most of the KWBS area is covered by rocky talus, practically without vegetation (31 %). Dwarf cedar and alder shrubs are spread at south-facing slopes and cover about 27% of the territory. Larch sparse woodland with moss-lichen cover is typical for steep north-facing slopes (14%). Open terrain larch wood (11%) and swampy sparse growth forest with minimal permafrost thaw depth, constant excessive stagnant moisturizing, tussock or knobby microrelief (9%) are characterizing creek valleys.

3. HISTORY OF KWBS

The Kolyma water balance station (KWBS), previously known as the Itrikanskaya runoff station of the Dalstroy Hydrometeorological Service (HMS), and later as the Kulinskaya runoff station of the Upper-Kolyma regional geological-survey department (1948-1956) and the Kolyma runoff station (1957-1969), was established by the Dalstroy HMS.

The station was founded on October 15, 1947. The primary goal of this station was studying runoff formation processes in small rivers catchments in the context of mountain landscapes and permafrost distribution, typical for Northeast USSR.

As soon as in May, 1948, the first runoff observations at the Kontaktovy and Vstrecha Creeks were launched, as well as regular observations at the «Nizhnaya» weather station (850 m asl). A bit later, on September 1, 1948, observations at the «Verkhnaya» weather station (1220 m asl) were started. In 1948, stage gauges Sredny, Nizhny and Vstrecha were equipped with automatic water level recorders, gauging footbridges and flumes.

During the period of 1949-1957, at the Vstrecha Creek catchment a rain-gauge network was organized. Channel cross-sections at the Severny, Dozhdemerny, Vstrecha creeks were equipped with various hydrometric facilities. Observations on soil, water and snow evaporation, soil freezing and thawing were commenced, as well as experimental observations at a runoff plot.

In 1957 the station was handed over to the jurisdiction of the Kolyma Hydro-meteorological service administration, and in 1958 it was partially electrified. At that time there were taken active steps toward fitting up the station with new types of devices and equipment, engaging new specialists in hydro-meteorology, building accommodation facilities.

In 1960 runoff observations at the Yuzhny Creek were begun, rationalization of the precipitation network was continued, radio rain gauges were installed. In 1963 two new water-balance sites were organized. In 1968 runoff measurements were started at the unique research object, at the Morozova Creek catchment, which has no vegetation cover and is composed of rocky talus.

In 1969 the Kolyma runoff station was renamed into the Kolyma Water Balance station (KWBS). In these years there was transition to broad experimental water balance observations of all of its elements and to enhanced technical level of research.

Since 1970 the KWBS carried out snowpack observations at avalanche catchments of the Tenkinskaya road, as well as stratigraphy, temperature and physical and mechanical properties of snow at four sites. Since 1980 there were introduced additional observations on dynamics of icing formation at the Kontaktovy creek. In 1982 observations on soil moisture were started at 3 agro-hydrological sites at the fields of the «Kulu» state farm.

In 1976 the station hosted a delegation of the USA scientists. They highly praised professional and personal qualities of the station's staff members, their commitment, on which extensive field studies and theoretical works were based, despite the equipment being rather simple and extreme living conditions. According to Slaughter and Bilello (1977), the data, received at the KWBS, were unique and unprecedented for the world practice.

Since the beginning of the 1990s, research program at the station has been gradually cutting back, and after 1997 water balance observations at the KWBS were ceased. One weather station and five level gauges functioned at the KWBS up to mid of June, 2013, when an extreme flashflood destroyed four level gauges. Nowadays only standard observations are conducted at the meteorological site and at the Kontaktovy (Nizhny) level gauge.

4. PROCESS STUDIES AT KWBS

4.1. Runoff measurements

Most of runoff in the catchments of this territory occurs in summer. In winter the creek freezes all the way through to the bottom. Spring freshet starts in May. For the summer period frequent rainfall floods are typical. Runoff observations were carried out at 11 catchments: creeks Kontaktovy (cross-sections Verkhny, Sredny, Nizhny), Morozova, Yuzhny, Vstrecha, Vstrecha (the mouth), Dozhdemerny, Severny, Ugroza (Fig. 2). Key characteristics of the catchments are listed in Table 1. All the water level gauges were equipped with «Valdai» water level recorders, as well as needle and hook water level gauges. In spring and autumn, when recorders did not work due to ice phenomena on the creeks, discharge was measured more frequently, every 4 hours. To prevent the recorders floats from freezing, the wells were heated with electric bulbs. At the small Morozova and Yuzhny creeks runoff was measured by means of a V-notch weir, at the Severny creek – with a flow measuring flume.

Table 1

Characteristics of KWBS runoff gauges

Catchment (creek – outlet)	Period	Area, km^2	Average slope, ‰	Catchment altitude (max-min, average), m	Average annual flow, mm	Maximum observed daily discharge, m^{3}s^{-1}
Yuzhny	1960-2012	0.27	303	1110-917, 985	202	0.14
Severny	1958-2012	0.38	388	1300-880, 1020	248	0.18
Morozova – Vodopadny	1968-1996	0.63	649	1700-1100, 1370	453	0.44
Ugroza	1983-1991	0.67	461	1270-914, 1100	354	0.27
Dozhdemerny	1952-1971	1.43	432	1450-950, 1180	208	0.31
Vstrecha	1949-2013	5.35	401	1450-833, 1070	266	3.15
Kontaktovy – Verkhny	1973-1980	5.53	537	1700-909, 1170	317	2.52
Vstrecha – the mouth of Ugroza Creek	1984-2012	6.57	406	1450-831, 1060	320	2.6
Kontaktovy – Sredny	1948-2012	14.2	413	1700-842, 1120	323	7.02
Kontaktovy – Nizhny	**1948-present**	21.3	407	1700-823, 1070	322	8.15

4.2. Meteorological observations

In 1948 in the Vstrecha Creek catchment measurements were carried out at two weather stations: Nizhnaya, that was opened for studying regional meteorological characteristics, and Verkhnaya (till 1972), organized to study regularities of elevation wise distribution of meteorological elements. At the sites, the observations on the main meteorological elements were carried out according to the weather station program (Fig. 2). In 1960 actinometrical and gradient observations were organized at the stations. In 1980, as a consequence of the KWBS and Kulu hydro-meteorological services merging, actinometrical, gradient and snow evaporation observations were transferred to the «Kulu» meteorological site, which became the main station of the KWBS, as a result. Rain-gauge stations for measuring daily precipitation

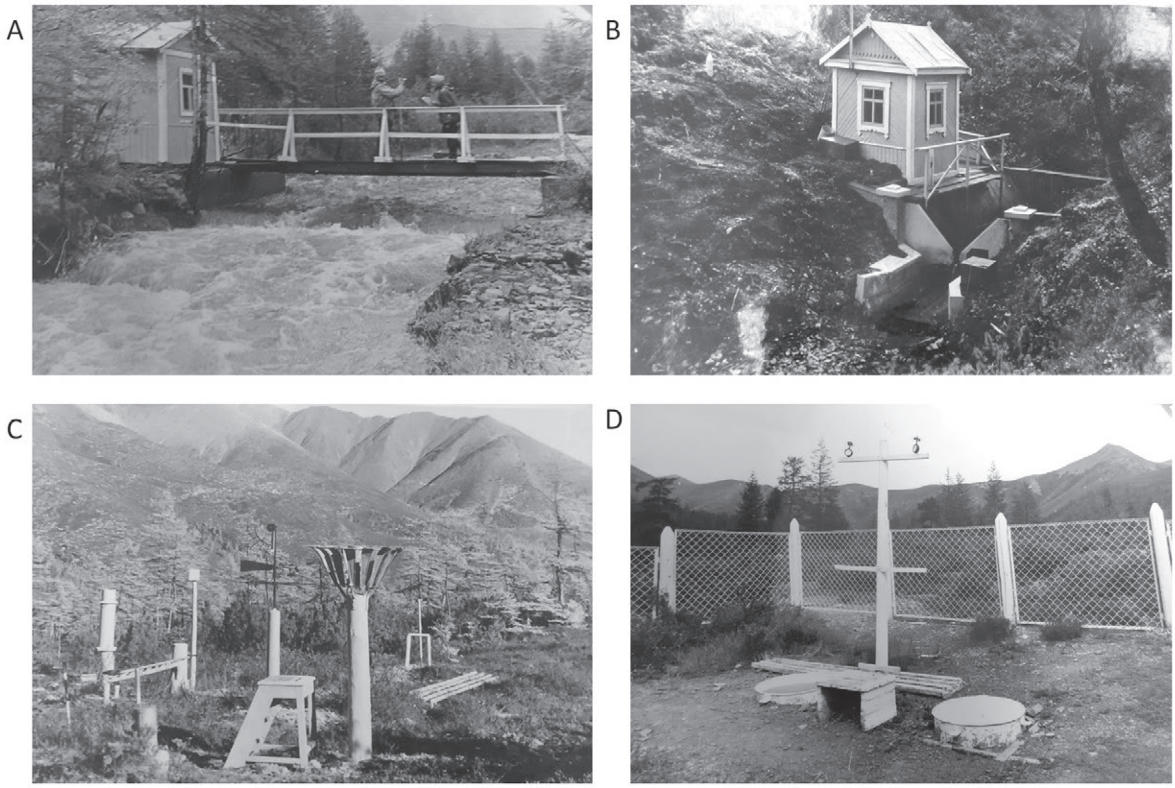

Figure 2. A –The runoff gauge at the Kontaktovy creek, 1979; B – The runoff gauge at the Yuzhny creek, 1960; C – The precipitation site (the bridge for Danilin frost tube can be seen at the right), 1959; D –The evaporation site, 1988

sums at the opening in 1948 were equipped with the Nipher shielded rain gauges and later – with the Tretyakov precipitation gauges (Fig.2). For a long time observations were carried out with both devices parallel. Continuous precipitation recording was performed with automatic rain-gauges, installed at 12 sites. In the 1960-s in the basin of the Yuzhny Creek four Tretyakov precipitation gauges were installed, and in the Kontaktovy Creek basin – six Kosarev total precipitation gauge and three Tretyakov precipitation gauges. At the same time there were attempts to register precipitation with automatic radio-precipitation gauges, but due to improper performance of the devices those observations were stopped. In 1988 precipitation observations at the KWBS were carried out with 36 precipitation and rain gauges, distributed relatively even throughout the area and altitudinal zones. Average density of the precipitation network accounted for 1.6 units per 1 km^2.

4.3. Evaporation

From 1950 to 1953, five soil evaporation sites were opened at plots with diverse underlying surface, on the slopes with different expositions and absolute altitudes. The sites were equipped with the Rykachev and Gorshenin evaporimeters with evaporation area 1000 cm2. Observation results till 1958 are considered to be approximate due to the absence of rain gauges and scales of required accuracy. From 1958 to 1966, the measurements were conducted at the soil evaporation site, located near the Nizhnaya weather station (Fig. 2). Observations on evaporation were carried out with two evaporimeters GGI-500 and the Rykachev evaporimeter, on precipitation – with a land rain gauge. To cover with soil evaporation observations diverse underlying surfaces, in different time at the KWBS territory three more soil evaporation sites were established: in the Vstrecha (1967), Morozova (1971 and Yuzhny

(1977) creeks basins. Measurements were carried out with standard weighting evaporimeters GGI-500-50, which, due to the physical proximity of permafrost, were changed to GGI-500-30, meaning that evaporimeters height has decreased to 30 cm. Also, observations on water surface evaporation (started in 1951 and made with the use of irregular devices) and on snow evaporation were carried out at the territory of the station. In 1959 the evaporation site was equipped with an evaporimeter GGI-3000 and a soil rain gauge with collecting area 3000 cm^2. Snow evaporation observations during the period of 1951-1955 were carried at five sites, equipped with Gorshenin evaporimeters (15 cm height with evaporation area 1000 cm2). From 1958 to 1966 snow evaporation observations were only conducted at the Nizhnaya meteorological site and exclusively during the thaw period. From 1968 to 1981 observations were made with standard evaporimeters GGI-500-6 routinely. In 1981 observations were transferred to the Kulu weather site.

4.4. Active layer depth

Since 1952 the observations on permafrost seasonal thaw dynamics were conducted at KWBS. Most permafrost observation sites equipped with Danilin cryopedemeters were located in approximate vicinity of Nizhnaya and Verkhnaya weather stations on the slopes with different expositions and various types of underlying surface. Despite the fact that permafrost observation sites were equipped with special bridges for observers to come close, eventually surface damage in the area where the device was installed began to influence thaw depth (Fig. 2). To understand spatial variability of thawing, special route surveys were carried out, for example, on September 1-3, 1983, seven route surveys on soil thawing from the watershed to the thalweg, considering types of underlying surface and slope exposition were carried out.

4.5. Snow cover

Snow cover observations were started in 1950 and initially were conducted at two catchments, at two meteorological sites and four typical squares. The observations at the Nizhnaya and Verkhnaya meteorological sites were carried out every 10 days along the routes, which total length at every site accounted for 1000 m (Fig. 2).

Snow surveys at the Vstrecha Creek catchment from 1950 to 1957 were conducted along 16 parallel routes, and since 1958 –along six routes. The studies at the Severny creek catchment were conducted along ten routes till 1958, since 1958 –along five ones. In 1968, snow cover observations at the Morozova creek basin were begun. During the period of 1968-1974, the number of snow-measuring lines in the Morozova creek basin increased from two to five, the number of snow depth measuring points – from 960 to 2640. Snow depth was measured every 5 m, snow density – every 50 m.

5. SCIENTIFIC VALUE OF KWBS

For the period of 1948 to 1997 the KWBS accumulated a huge amount of data on hydro-meteorological and special observations of a unique duration (40-50 years) that characterize the natural settings, which, on the one hand, are nearly ungauged, and on the other hand, appear to be representative for vast territory of North-East Russia. They were published in 40 issues, the first one covers the period of 1948-1957. Following issues were published annually (KWBS Observation's materials, 1959-1997). Precipitation and discharge observations, combined with rather rarely observations on evaporation, water yield from snow, surface runoff, soil thawing, etc. provide a chance for a detailed study how individual processes of the hydrological cycle interact with each other and with landscape components. Based on the analysis of the KWBS results the regularities of runoff formation processes in ungauged basins of the permafrost region of Russia have been revealed.

Observation results were reflected in more than 250 publications, dedicated to different aspects of runoff formation in the continuous permafrost region, active layer dynamics, underlying surface structure and its influence on hydrological processes.
Based on the KWBS materials, particular aspects of water balance formation were studied (Boyarintsev, Gopchenko, 1992; Suchanskiy, 2002; Zhuravin, 2004; Lebedeva et al., 2017), peak and spring flood runoff in small rivers in the permafrost zone (Boyarintsev, 1988), their base flow (Glotov, 2002), principles of runoff cryo-regulation (Alekseev et al., 2011), ice content dynamics of rocky talus deposits (Bantsekina, 2003), floodplain taliks in continuous permafrost (Mikhaylov, 2013) and many others.
Collected data are used for development and testing different geoscience models: runoff formation (Gusev et al., 2006; Kuchment et al., 2000; Vinogradov et al., 2015; Lebedeva et al., 2015; Semenova et al., 2013), climatic (Shmakin, 1998), land surface, vegetation dynamics (Tikhmenev, 2008).

6. PERSPECTIVES OF KWBS

In summer 2016, with the assistance of Melnikov Permafrost Institute SB RAS, a group of specialists, consisting of different scientific Institutions representatives, conducted a reconnaissance survey of the KWBS in order to find out if it was possible to carry out scientific research and stationary monitoring of permafrost and hydrological processes at the station. Despite difficult logistic access to the KWBS it was considered possible to organize accommodation and provision of the station for the period of summer expeditions (Fig. 3).
At first, the main goal of research resumption at the station can be renewal of regular observations of runoff, meteorological elements and active layer dynamics at three small catchments (Morozova, Yuzhny, Severny) and the KWBS main-stream outlet (Nizhny cross-section) using advanced equipment with automatic data recording. As a result, some unique runoff observations series – over 60 years long – will be continued, what will allow for evaluation of climatic impact on permafrost and provide a scientifically based forecast on current and future climate change impact on hydrological regime. Runoff measurements would be accompanied by collection of water samples from creeks, precipitation, water, snow and ice of seasonal thaw depth in different landscapes to identify chemical composition and stable isotopes content for identification feeding sources of the KWBS creeks (Tetzlaff et al., 2015).
During short 3-4 week field trips at the beginning and at the end of the warm season it would be also possible to study specific processes of runoff formation under permafrost conditions. Slope runoff occurs unevenly, and is concentrated in particular areas (drainage zones or preferential pathflows), where subsurface water flow speeds up enough to transfer matter, both as dissolved and as soil particles. Reconnaissance surveys of the Kontaktovy creek catchment at the KWBS territory, 2016, revealed there several types of such zones of slope runoff concentration.
Another possible scientific goal is to evaluate the role of cryogenic redistribution of runoff in rivers feeding, which occurs due to ice freezing-melting in coarse-grained slope deposits. Similar studies have already been carried out in mountain regions of permafrost, including the KWBS (Sushansky, 1999; Bantsekina, 2003). Other research issue is the study of floodplain taliks (Mikhaylov, 2013) and aufeises (Alexeev, 2016) and their impact on hydrological processes in continuous permafrost zone.
Field trips for a limited group of scientists could be covered with relatively modest financial support through research grants. In the future, the KWBS could get back its status of a research station, to receive state funding, sponsor support from gold mining companies of the Magadan region and become an international center for complex studies in the field of permafrost hydrology.

The KWBS is situated in the region where natural processes monitoring network is extremely sparse. From 1986 to 1999, number of hydrological gauge stations in Far-East parts of Siberia decreased by 73% (Shiklomanov et al., 2002).

Resumption of water balance observations and organization of complex research of permafrost, climate, and landscape, hydrological and hydrogeological processes on the basis of the KWBS would make it possible to get new data, representative for understudied territory of the Arctic in the context of environmental changes. Considering insufficient knowledge about this territory, the KWBS has prospects to become a highly demanded complex international stationary center for testing natural processes models at different scales – from point to regional, –validation of remote sensing products and a place for multidisciplinary field research.

Though the largest rivers of the Arctic Ocean basin flow through Russia, it has no stationary centers that could conduct focused studies of hydrological processes at catchments in the permafrost region. The KWBS incorporation into the international network for monitoring natural processes in cold regions (Interact, SAON, CALM, GTN-P, etc.) could enhance international cooperation.

Nowadays, resumption of continuous observations and research at the Kolyma station appears to be a critical task due to increased interest in natural processes of the Arctic region. Present-day data, following the KWBS long-term observations series, could become a valuable indicator of climate change and a basis for studying its impact on the state of permafrost and hydrological regime.

Right now, until the station infrastructure is not completely destroyed, some of the specialists who worked at the KWBS are still active, it is necessary to gain attention and support from Russian and international scientific community regarding the renewal of the KWBS.

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COMPARISON OF RUNOFF AND RIVER FLOW IN TWO LARGE NORTHERN BASINS

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ABSTRACT

The magnitude and timing of water delivery in two large northern basins are analysed to clarify where runoff is generated and how their rivers acquire comparable regimes (or seasonal rhythms) of flow. These two rivers, the Mackenzie in Canada and the Yenisei in Russia, traverse equivalent latitudes, physiographic provinces, vegetation zones and climatic regions. Within the basins, mountainous terrains and high precipitation sections usually yield large runoff, but low runoff comes from the plains, low plateaus and areas of aridity. Winter runoff is commonly low and snowmelt is responsible for annual peak runoff in most parts of these basins, though rainfall is a prominent runoff source in southern Yenisei. Many rivers in the drainage networks display a seasonal pattern that suggests the dominance of snowmelt to produce a spring freshet followed by a general decline in summer that diminishes to winter low flows. Artificial regulation of reservoir outflow greatly distorts the natural flow regime. Yet, along the main river downstream of the reservoirs, the influx of tributary discharge can dilute such human influence. Like these two rivers, most large northern rivers exhibit a snowmelt-dominated (or nival) flow regime at their outlets to the polar seas, but this is only an apparent nival regime through an integration of sub-basin discharge. To truly understand how water is produced and transferred, the spatial and temporal complexity of flow-generation mechanisms and storage effects need to be unravelled.

KEYWORDS

Large basins; Mackenzie River; Yenisei River; runoff; river flow; regime

1. INTRODUCTION

Several mega-basins in North America and Eurasia discharge directly or indirectly northward, bringing substantial amounts of continental water to freshen the sea-water of the Arctic Ocean. The importance of this river water in affecting the temperature and salinity of the surface ocean-water layer, sea ice growth and decay, and thermohaline circulation of the ocean have been discussed in the literature (e.g. Aagaard and Carmack 1989; Peterson et al. 2002). Since these northern rivers of sub-continental scale traverse multiple latitudinal and altitudinal zones, the magnitude and timing of water delivery vary greatly within the river system. To know how large rivers attain their discharge magnitude and how they acquire their regime of flow, it is necessary to analyse the contribution of runoff from various parts of the basin and the modification of streamflow regime along the drainage network.

Two mega-basins that are continents apart are chosen for this comparative study of their hydrological behaviour. These are the Mackenzie River Basin in Canada and the Yenisei River Basin in Russia (abbreviated as MRB and YRB in the rest of this paper). Together, their rivers provide about one-third of the flow that directly enters the Arctic Ocean. Using these basins as examples, this study considers where and when runoff is produced and relates it to the physical setting of various hydro-physiographical regions. We examine the amount and seasonality of water delivery from the drainage networks and analyse how the rhythm of river flow is modified as water moves downstream. Since the two selected basins have physical settings shared by other major northern basins, information obtained from this investigation is pertinent to others that discharge to the polar sea.

2. DATA SOURCE

This paper utilizes discharge data available from public domain websites. Discharge data were obtained from HYDAT, the National Water Data Archive compiled by Water Survey of Canada for the MRB and R-ArcticNet (version 4.0) for YRB. Climate data were obtained from the Meteorological Service of Canada and from the Daily Temperature and Precipitation Data for 518 Russian Meteorological Stations available from the Carbon Dioxide Information Analysis Centre. The contributing drainage areas above each station were delineated manually using sub-basins obtained from the HydroBASINS and HydroSHEDS projects. The spatial precipitation plots were generated using the Global Precipitation Climatology Project (GPCP) Version 2.3 Combined Precipitation Dataset Data, which combines observations and satellite precipitation data.

3. COMPARISON OF BASINS

3.1. Physical Setting

The two basins cover large areas: 1.8 million km^2 for MRB and 2.55 million km^2 for YRB. They straddle between latitudes 45ºN and 70ºN. Both basins have large altitudinal ranges that give rise to contrasts of uplands and lowlands (Fig. 1), but the topographical configuration differs between these two basins and that has a bearing on their hydrology.

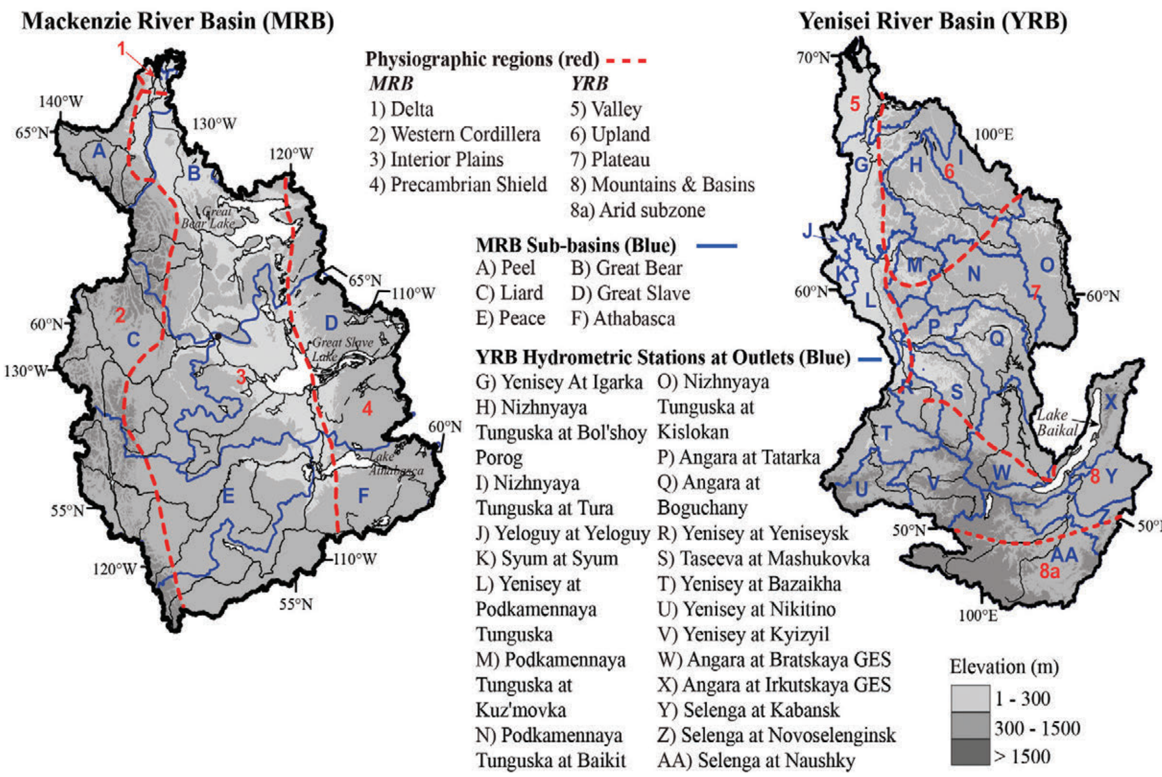

Figure 1. Mackenzie and Yenisei River Basins showing their hydro-physiographical regions, rivers and lakes. Also shown are the hydrometric stations mentioned in this study.

Around the mouths of the Mackenzie and the Yenisei rivers, networks of distributaries drain the deltas and wetlands. For the MRB inland of the coastal belt, the topographical alignment is predominantly north-south, partitioned into the rugged mountain chains of the Western Cordilleras, the subdued relief of the Interior Plains and the rolling topography of the Canadian Shields in the east. In contrast, the topography of YRB has a preferred east-west orientation. Inland from the coastal zone is the Central Siberian Plateau which can be distinguished into an upland and a low plateau section. Putorana Mountains are part of the upland that rises sharply from the Yenisei River, with its valley extending to the plains in the Ob River Basin to the west. South of the Putorana Moutains, the Yenisei valley narrows as it intersects the Yenisei Ridge, and the Ridge is treated as a part of the upland. The region in southern YRB comprises of mountain chains and basins. The depressions and many mountain slopes in southern YRB (in Mongolia and Russia) are predominantly arid areas that affect their hydrological behaviour. In this study, we deal with areas upstream of the lowest hydrometric stations that collect most of the flows in MRB (at the village of Arctic Red River) and in YRB (at Igarka). Figure1 shows the major physiographical regions.

As both basins extend from cold temperate to the Arctic domain, they have comparable latitudinal ranges in vegetation. The southern tip is mainly grassland, with agricultural practices at limited locations. North of it lies the boreal and subarctic forests. Beyond the forests lies the tundra north of the treeline. In addition to latitudinal change in vegetation, elevation gives rise to vertical zonation in the mountains, with forests at the lower elevations and alpine tundra and rock outcrops exposed above the alpine treeline.

3.2. Precipitation

Comparing the two basins, one common feature is the pronounced interaction of airflows with topography. Both MRB and YRB lie in the latitudinal belt of the westerlies, but airflows from other directions also reach them during different times of the year. These atmospheric dynamics give rise to seasonal and spatial variations in moisture influxes and effluxes, which strongly influence precipitation. Annual precipitation patterns of MRB and YRB are shown in Figure 2. In general, higher amounts fall on high altitudes and large differences exist between slopes. The Sayan Mountains and Tannu-Ola ridge of YRB exhibit pronounced contrasts in precipitation between windward and leeward slopes. Topographic depressions and deep valleys are precipitation shadow areas, many of which are semi-arid. Cyclones from the Atlantic bring high precipitation to the Yenisei Mountains and Putorana Plateau in northern YRB, yielding precipitation to the western parts of the plateaus but decreasing eastward in the adjacent Lena River basin. In a similar fashion, considerable quantities of snow and rain are frequently deposited on high altitudes of the Cordilleras of MRB, but precipitation diminishes greatly at the eastern foothills downwind of the mountain chains (Hydrological Alas of Canada 1978). The least amount of precipitation reaches the lowlands and low plateaus.

Winter (November-March) is the season of low precipitation for both basins. The Cordilleran region of MRB has >100 mm, depending on elevation. Elsewhere, winter precipitation decreases to about 80 mm in the north. For YRB, high areas receive over 150-200 mm of winter precipitation, and it is 100 mm or less for other areas. For both basins, most precipitation falls between April and October, dominating the annual total. Rainfall constitutes a large portion of annual precipitation for both MRB and YRB, with snow representing about 25% of total precipitation at most places. The snow cover duration increases northward and with elevations. MRB has snow on the ground for 150 days in the southern plains and 270 days at its Arctic sea coast. The northern half of YRB has >200 snow covered days and the duration reaches nine months in the Arctic coast.

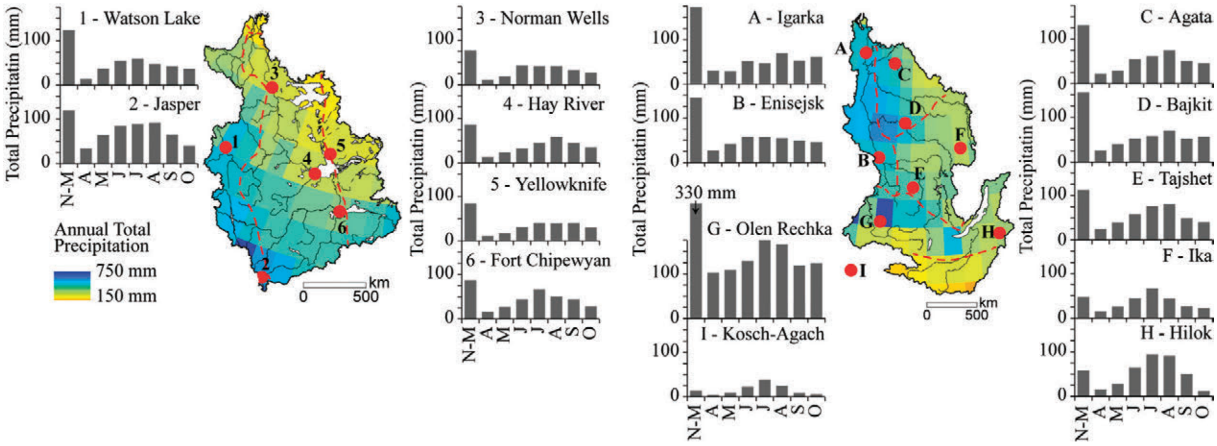

Figure 2. Annual precipitation of MRB and YRB, with hyetographs for selected stations.

3.3. Storage and Release Functions

Situated at continental high latitudes, the storage of snow and ice is an important consideration in basin hydrology. Snowfall accumulates in winter to build up a seasonal snow cover that undergoes re-distribution, mainly through drifting by wind and interception by vegetation. For both basins, snowmelt and associated runoff is a major event in the spring, and streamflow is often accompanied by the breakup of river ice to generate annual peak flows.

Rivers and lakes usually acquire an ice cover that lasts for several months. This feature is insignificant in magnitude compared with other forms of storage. Ground ice in permafrost is a long-term storage but a warming climate and disturbance resulting from geomorphic processes or human activities lead to permafrost thaw and the release of water from melted ground ice. Groundwater storage and release is less well measured than surface water. Karst terrains are especially favourable for groundwater discharge, sometimes maintaining perennial springs to feed the river system. Winter base flow of rivers is sustained by groundwater, sometimes building up icing (or naled) in river channels.

Lakes are particularly effective for surface water storage. In addition to numerous small lakes, several of the world's largest lakes are found in YRB (Lake Baikal) and in MRB (Lakes Athabasca, Great Slave and Great Bear). Sizeable human-made lakes (reservoirs) have also been created in both basins and they are a significant component of the hydrological landscape (Jaguś et al. 2015; Peters & Prowse 2001; Yang et al. 2004). YRB has a longer history and larger number of reservoirs than MRB. All these reservoirs are operated for the generation of hydro-electric power.

4. RUNOFF GENERATION

Figure 3 presents the pattern of runoff in MRB for the period 1973-2015, mapped using data from sub-basins with areas of <100 000 km^2.The flow of Mackenzie River along its main trunk will be treated in the next section, which incorporates the effects of large lakes on river flow. Annually, the western mountains with high precipitation consistently produce the highest runoff among all regions, generally >400 mm. Within this region, the southern zone usually has the largest runoff that exceeds 500 mm, supported by rainfall, snowmelt and glacier melt at high altitudes. Runoff diminishes in the foothill areas, dropping to the lowest amount of <100 mm in the southern Interior Plains where summer evaporation is particularly intense. The Shield region has 100-200 mm of annual runoff, values that are intermediate between the mountains and the plains.

Owing to the many missing winter discharge data, values from November to March are lumped to give runoff for the winter season. Winters in MRB are intensely cold. Winter runoff is provided mainly by groundwater and in some cases, by discharge from lakes and reservoirs (e.g. Great Bear Lake, Williston Lake that is a reservoir on the Peace River) (Woo and Thorne 2014). The mountain region with steep terrain yields high runoff of >50 mm. Similarly, the Shield produces about 50 mm because the storage effect of its many lakes sustains moderate outflows during the winter. In contrast, the plains have very low runoff of 10 mm.

Snowmelt comes early in southern MRB, producing >10 mm of runoff in April. Snowmelt intensifies in May and the southern mountains can yield 100 mm while much of the plains produce 10-30 mm. On the other hand, the Shield has low runoff, likely because the lakes can withhold much of the snow meltwater in storage. Snowmelt is delayed in the northern mountains until June or even July, when most the snow has already disappeared at all low elevations. After the snowmelt period, runoff diminishes in July, particularly as evaporation increases (Hydrological Alas of Canada 1978). The plains have the lowest runoff of <10 mm/month. This pattern persists into August and September. Freeze-up occurs in October at most parts of MRB.

Runoff of YRB is calculated using the discharge of headwater catchments or by considering the difference in discharge for the inter-station area between two adjacent stations along a river (Fig. 4). Stations along the Angara and Yenisei rivers that lie immediately below reservoirs are excluded to reduce the signal imparted by flow regulation on runoff calculation, though we used the post-dam periods (1971-2006 for Angara and 1981-2005 for upper Yenisei) that have more data available to us.

At low temperatures, winter precipitation is stored and does not melt until spring. Disregarding segments of the Angara and upper Yenisei, where reservoir operation masks the rhythm of natural runoff, the unregulated sub-basins of the Angara system (e.g. Taseeva River) and of upper Yenisei (e.g. Kyizyil River) produce low runoff throughout October to April. Annual peak runoff in these catchments is generated by snowmelt, first in southern YRB and on south slopes of the high plateau in May, then proceeds to other plateau areas in June. During these months, the Taseeva yields about 30 to >45 mm/month and the Yenisei between Kyizyil and Bazaikha around 25-30 mm/month. Snowmelt in the Tannu-Ola and Sayan Mountains produces 40-50 mm/month in May and June, and >50 mm/month from the northern upland, the latter benefits from the accumulated snow deposited by winter storms. Summer is the season with the highest precipitation of the year. Kyizyil basin yields 40-45 mm/month in July and August and Taseeva gives about 25 mm/month, while rainfall also augments runoff from the low plateau. The Selenga basin that drains into Lake Baikal, however, is semi-arid with low precipitation and high evaporation, giving the lowest summer runoff (5-9 mm/month) in YRB.

The annual pattern shows high runoff from the western upland areas in northern YRB and from the mountainous areas in southern YRB, but the latter area has a complex relationship with topography. Large contrasts exist between the rivers fed by much snowmelt and rainfall and those in the rain shadow areas. Elsewhere, the high plateau yields high runoff of >200 mm while less comes from the lower plateau areas (125-200 mm), and low precipitation and high evaporation in the southern end of YRB result in <100 mm/year of runoff.

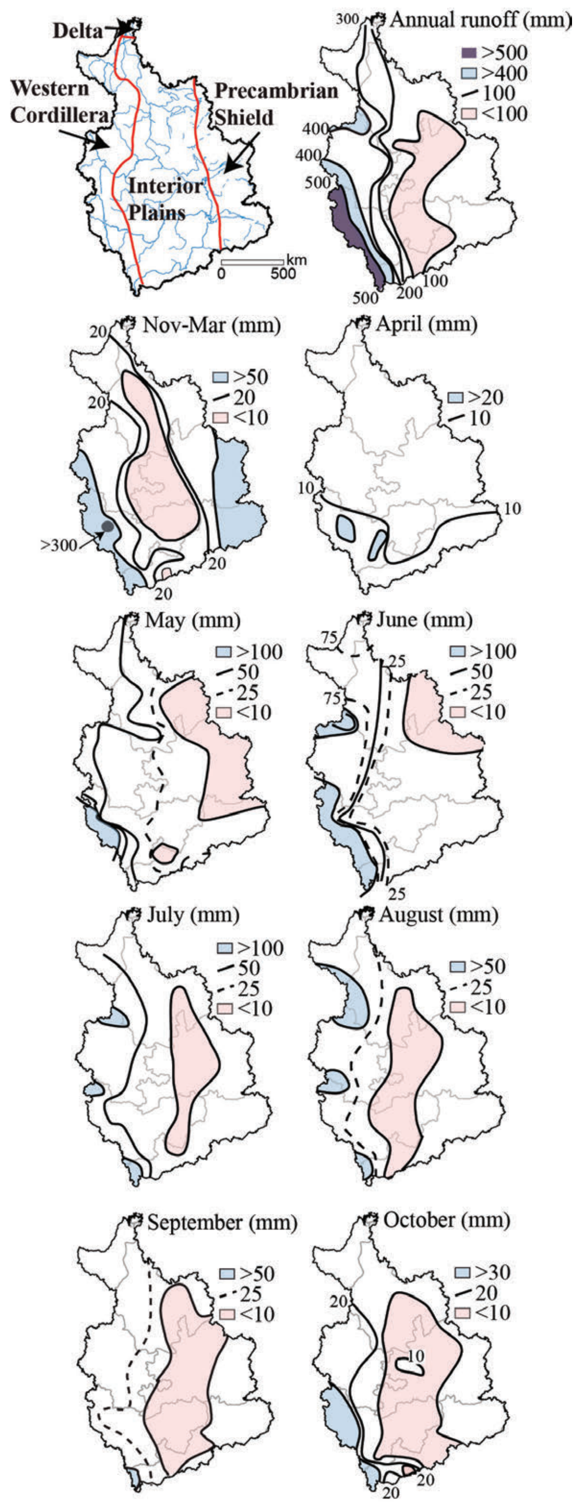

Figure 3. Runoff of Mackenzie River Basin.

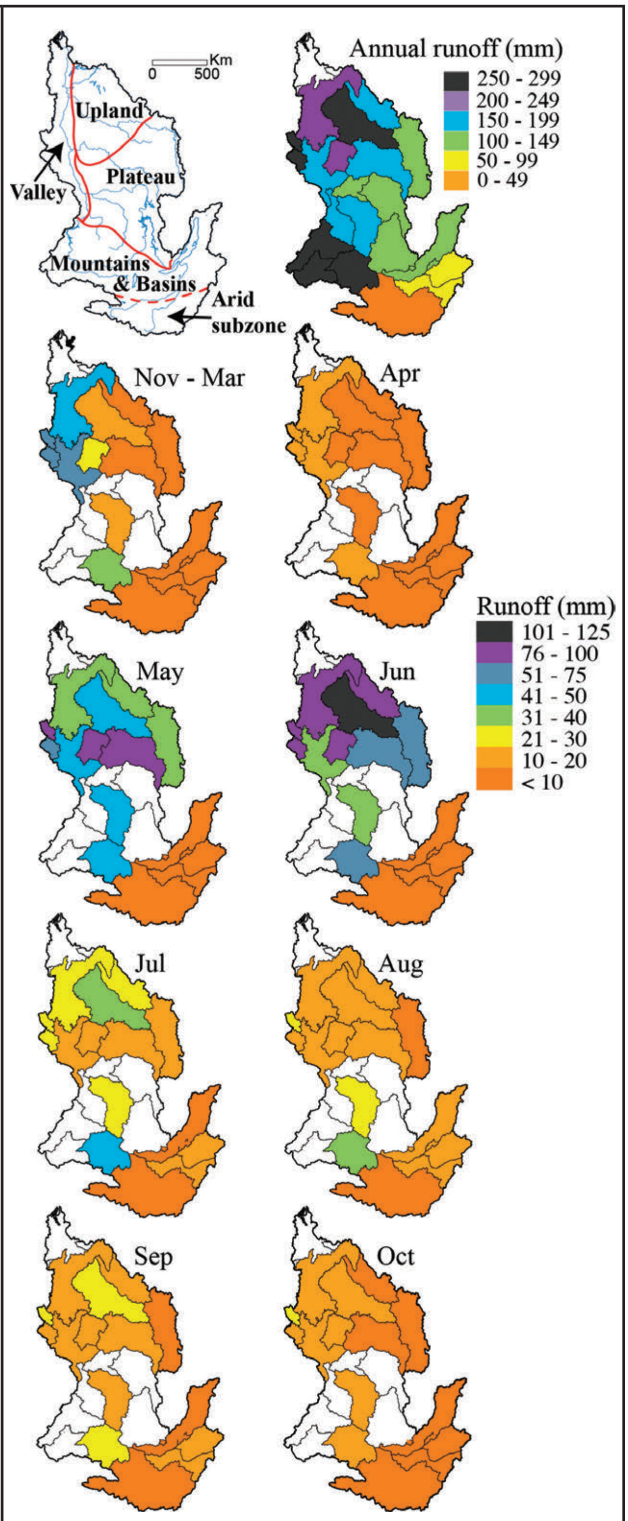

Figure 4. Runoff of Yenisei River Basin. Monthly runoff from inter-station areas strongly affected by reservoirs is not shown.

5. RIVER FLOW

5.1. Comparison of Storage and Release Functions

Through runoff production, water balance governs the net amount of water available. Storage, both in the land phase and along the channels, dictates the timing of river flow.

Snow storage delays runoff generation until melt events, as noted previously. For both MRB and YRB, the most common natural flow rhythm is the nival regime in which the melting of winter snow in conjunction with river ice breakup gives rise to annual high flow, followed by a general flow recession in the summer, ending with low flow in the winter (Brown et al. 2008). In large basins like the MRB and YRB, the timing of snowmelt differs, arriving first in the south. Thus, high flow occurs in Taseeva and Podkamennaya Tunguska rivers in May but for Nizhnyaya Tunguska River further north, it occurs in June. Where snowfall is less important than rainfall, snow storage plays a lesser role. Then, both runoff and river flow respond readily to the timing of precipitation. In southern YRB that receives most of its precipitation in summer, the pluvial regime prevails. On a local scale, other storage considerations including the presence of glaciers, extensive wetlands and the abundance of small lakes lead to adjustments to the nival and pluvial regimes (Woo 2012).

The storage function of large lakes significantly impacts the timing and magnitude of their outflow, producing a prolacustrine flow regime. Lake storage generally absorbs water influx from the basin and gradually yields relatively more uniform and delayed outflow, dampening or removing short-term fluctuations in the inflow hydrograph and extending the period of water release from the lake. Large natural lakes, including the Athabasca, Great Slave and Great Bear in MRB and Lake Baikal in YRB perform this function of flow retention and release.

As artificial lakes, reservoirs seriously modify the natural flow regime of a river (Vyruchalkina 2004). Annual discharge can be reduced through increased evaporation from the reservoir water surface, and the construction and initial operation phases of the upper Yenisei entailed a drop in annual discharge as far downstream as Igarka, as indicated by Steufer et al. (2011). In the operational phase, the amount of flow release is dictated by human demand (e.g. outflow from Williston Reservoir on Peace River at Hudson Hope in Fig. 5a). Discharge fluctuates from day to day to suit the need for power production, though on a seasonal basis the pattern of regulated flow indicates that less water is released in the summer than in the winter, except for special occasions. On the whole, because the water is not for consumptive use, flow regulation alters the seasonal rhythm of flow and would less seriously affect the flow amount totaled over an extended time period (such as a year).

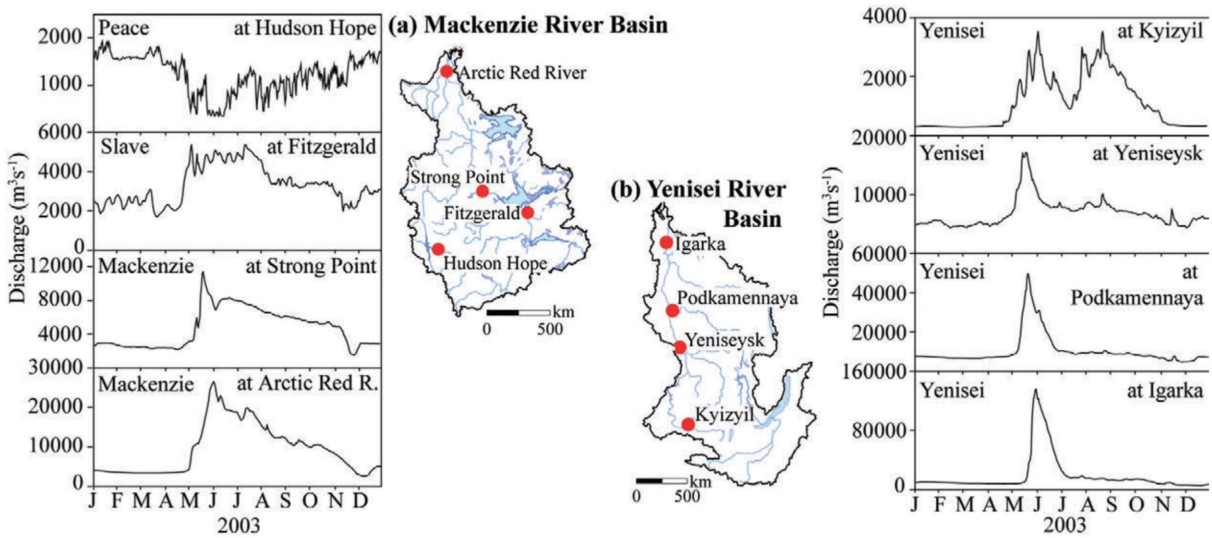

Figure 5. Integration of flow in large river systems as shown by the 2003 daily discharge of rivers in (a) the Mackenzie Basin from Peace River at Hudson Hope to Artic Red River, (b) the Yenisei Basin from Kyizyil to Igarka.

The regime exhibited at the main trunks of Mackenzie and Yenisei rivers is continuously modified by runoff inputs and tributary influxes when these large rivers flow towards their mouths. The effect of natural lake storage or human manipulation is diluted when downstream tributaries join the main stem. Figure 5a traces the downstream integration of flow along these mega-rivers. Williston Reservoir on Peace River in MRB distorts the original nival regime, which is restored by the natural inflow downstream. After joining the Athabasca River at Lake Athabasca, it becomes the Slave River (at Fitzgerald in Fig. 5a) before entering Great Slave Lake. The river that flows out from this lake is the Mackenzie River, which displays a prolacustrine regime as shown at Strong Point. However, in meeting the Liard River, the pronounced nival flow regiment of this tributary overwhelms the lake effect on Mackenzie flow. This flow pattern is conveyed all the way downstream to Arctic Red River station before the Mackenzie branches into its delta. In a similar fashion (Fig. 5b), the flow regime found at Kyizyil is altered by reservoir operation downstream and is changed again at Yeniseysk where it is joined by the Angara River, itself having a combination of flows from such unregulated river as the Taseeva and the highly regulated Angara above Boguchany. Further downstream, the Yenisei receives natural inflows from the Podkamennaya Tunguska and Nizhnyaya Tunguska rivers, which further adjust the regime of the Yenisei such that it resembles that of a nival regime at Igarka. Thus, the regime of Mackenzie and Yenisei rivers at their mouths is an amalgamation of different flow patterns.

5.2. Comparison of Flow Magnitudes

Table 1 groups the flows of the major sub-basins and inter-station sections along Mackenzie and Yenisei rivers into three seasons: winter (November to March), snowmelt (April to July) and summer (August to October). In terms of contribution to total flow of MRB and YRB, the relative importance of each subsection changes seasonally. On an annual basis, the mean flow of MRB at Arctic Red River is 290 km^3 and for YRB at Igarka, it is 610 km^3. Total flow is larger for YRB because it has higher mean runoff and substantially bigger basin area (173 mm runoff from 1.68 million km^2 for MRB and 249 mm runoff from 2.44 million km^2 for YRB). Within each basin, the amount of annual flow varies (Table 1). Rivers in mountains and high plateaus generally have the largest flow. They include part of the Athabasca, the Liard and lower Mackenzie basin, with most of their drainage catchment located in the Western Cordilleras. Together, they provide almost half of the total Mackenzie River flow. In YRB, the lower Yenisei (from the confluence with Podkamennaya Tunguska River to Igarka) and Nizhnyaya Tunguska River yield 24% and 19% respectively of the Yenisei total. A notable departure of mountainous terrain as a high flow zone is the arid range and basin topography in southern YRB, where the Yenisei above Kyizyil, the Taseeva River and the Selenga River each produce only 4-5% of the Yenisei flow. The next group with high annual flow consists of the regulated rivers: Peace River gives 23% of total Mackenzie flow; the Angara produces 16% though the reservoir section along the Yenisei River provides only 9% of total Yenisei flow. The remaining parts of the two basins, consisting of large natural lakes, plains, low plateau of Central YRB, and rolling areas such as the Canadian Shield, are responsible for 20-25% of the flows in MRB and YRB.

Table 1

Flow contribution from sub-basins and inter-station areas to seasonal flow of Mackenzie and Yenisei rivers. Numbers in brackets are areas in thousands of km²

Mackenzie River Basin	Flow in km³				Flow as % of seasonal total			
	Nov-Mar	Apr-Jul	Aug-Oct	Annual	Nov-Mar	Apr-Jul	Aug-Oct	Annual
Athabasca (133)	2.58	10.79	5.69	19.06	4.8	7.4	6.2	6.6
Peace (293)	20.37	31.07	14.59	66.03	37.6	21.2	16.0	22.7
Liard (275)	9.34	48.72	23.05	79.45	17.2	33.3	25.2	27.4
Great Bear (146)	6.71	5.59	4.54	16.83	12.4	3.8	5.0	5.8
Slave (180)	8.49	2.56	9.10	20.15	15.7	1.7	10.0	6.9
Mid-Mackenzie (389)	4.49	15.90	14.37	34.76	8.3	10.9	15.7	12.0
Lower Mackenzie (264)	2.27	31.80	20.05	54.12	4.2	21.7	21.9	18.6
Seasonal Total	54.23	146.43	91.39	290.40				
Yenisei River Basin	Flow in km³				Flow as % of seasonal total			
	Nov-Mar	Apr-Jul	Aug-Oct	Annual	Nov-Mar	Apr-Jul	Aug-Oct	Annual
Selenga (360)	1.87	11.04	11.18	24.08	1.6	3.0	9.1	4.0
Taseeva (127)	2.37	14.17	7.03	23.58	2.0	3.8	5.7	3.9
Angara main trunk (553)	42.02	38.21	18.65	98.88	35.9	10.4	15.2	16.2
Podkamennaya Tunguska (232)	5.27	42.41	8.91	56.60	4.5	11.5	7.3	9.3
Nizhnyaya Tunguska (447)	4.88	87.59	21.23	113.69	4.2	23.7	17.3	18.7
Yenisei at Kyizyil (115)	3.73	18.53	9.97	32.23	3.2	5.0	8.1	5.3
Yenisei reservoir section (245)	29.46	12.42	13.20	55.08	25.2	3.4	10.8	9.0
Mid-Yenisei (128)	4.92	46.30	8.86	60.08	4.2	12.5	7.2	9.9
Lower Yenisei (233)	22.41	98.53	23.43	144.37	19.2	26.7	19.1	23.7
Seasonal Total	116.92	369.21	122.45	608.59				

6. DISCUSSION AND CONCLUSION

The Mackenzie River basin (MRB) in Canada and the Yenisei River basin (YRB) in Russia are significant suppliers of freshwater to the Arctic Ocean. These basins have equivalent hydrological setting. Both basins receive their atmospheric moisture from two sources: advected from outside and recycled from within. Topography interacts with large-scale atmospheric flows, either blocking them or facilitating their movement, to influence the regional climates; and enhances moisture recycling through mountain-plains circulation.

High runoff comes from their mountainous zones and low runoff comes from areas with low precipitation and high summer evaporation. Large lakes in these basins provide notable storage that detains high inflow and sustains moderate outflow in the winter months. Large reservoirs have been built in both basins for hydro-power production, which significantly alters the natural flow regimes.

The seasonal rhythm of discharge of the two large rivers, the Mackenzie and the Yenisei, are similar, peaking in late spring to early summer and diminishing to their annual minima in winter. This pattern appears superficially to resemble a nival regime with the high flow suggestive of meltwater release from the snow accumulated over their long winter. Such an apparent nival regime is not entirely related to spring snowmelt, but is the product of integration of flow contribution from their sub-basins and of flow modification along their drainage networks. Knowledge of spatial variations in runoff generation and flow contribution from different parts of a mega-basin can be helpful to future studies for isolating human and natural influences and for identifying areas in the basin that are sensitive to expectant changes.

MRB and YRB have hydro-physiographical attributes typical of river basins in the circumpolar region. The general conclusions drawn from this study regarding runoff generation, flow contribution and river discharge characteristics are of relevance to other northern mega-basins. Thus, our approach and findings have circumpolar applications.

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SWINGS IN RUNOFF AT POLAR BEAR PASS: AN EXTENSIVE LOW-GRADIENT WETLAND, BATHURST ISLAND, CANADA

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ABSTRACT

Polar Bear Pass is a large low-gradient wetland situated in the middle of Bathurst Island, Canada (75°40'N, 98°30'W). The watershed (422.1 km^2–77.7% uplands, 22.3% lowlands) is a National Wildlife Area providing food and shelter for migratory birds and larger fauna such as muskox, caribou and polar bears. Since 2006, our studies here have focused on snow and pond hydrology, biogeochemistry and runoff from small hillslope catchments. As yet, we have not quantified runoff from the wetland into nearby coastal arctic waters. Recent research suggests that there is a critical need to explore how arctic stream discharge patterns and water budgets may be shifting in response to climate warming. Hence, the focus of this study was to (1) assess the pattern of runoff out of PBP watershed (eastern sector-102.6 km^2) during two contrasting spring/summer seasons: 2012 – warm, early melt versus 2013 – cold, late melt; (2) quantify the seasonal water budgets for these two contrasting seasons; and (3) place these results in the context of other arctic basin studies. The end-of-winter snowpack was quantified using a terrain-based approach. A physically-based snowmelt model using weather station data provided daily melt estimates. Direct measurements of snow ablation were also made. Daily rainfall was measured with recording rain-gauges, and streamflow was estimated at the outlet using the mid-section velocity approach.

The end-of-winter snowpack was slightly deeper in 2013 than in 2012. Due to warmer and drier conditions in 2012, snowmelt began and ended earlier than 2013. Spatially, the melt pattern was similar in both years; here the north part of the Pass melted out earlier than the southern part. In 2012, in response to warm/dry conditions the stream hydrograph showed a rapid rise in flow driven by meltwater from the north part of the Pass. This was followed by a series of secondary peaks driven by snowmelt from the southern part of the Pass. Due to a late and slower melt in 2013, the largest peaks in the stream hydrograph came instead from the southern sector (mid-July). Overall, the runoff ratios varied between the two years as did the stream water budgets. The 2013 streamflow pattern was more typical of high arctic watersheds in the early 70's.

KEYWORDS

Arctic Snowcover; Arctic Wetland; Snowmelt; Streamflow

1. INTRODUCTION

Wetlands are critical landscapes in High Arctic landscapes, providing food for migratory birds and larger fauna such as caribou and muskox. They also serve to store and replenish freshwater supplies and recently they have been the focus of interest in terms of their role in up-taking and releasing greenhouse gases (CO_2, CH_4 and H_20). Snow remains an important source of water into these ecosystems, often replenishing ponds and lakes and re-saturating wet meadow areas at the end of a cold winter season. Evaporation remains a key loss of water from these wetlands during the spring and summer seasons (*e.g.* Miller & Young 2016; Young *et al*. 2016). Runoff out of these wetlands in coastal arctic waters is less well known; hence our ability to evaluate the flow of carbon, sediments and nutrients out of these wetlands is limited. Streamflow studies in Northern Canada (Déry & Woo, 2005; Déry *et al.* 2009) have shown an intensification of streamflow patterns such as earlier freshets, flashier peak flows, and increases in baseflow in response to climate warming. Karlsson *et al.* (2015) remarks that that interest in Arctic river discharge changes is growing, as these changes can signal alterations in the terrestrial hydrological cycle. Whether recent climate warming in the High Arctic (Woo & Young 2014) is translating into flow regime shifts for extensive wetland systems is not yet apparent. The objectives of this study are to (1) assess the pattern of runoff out of the eastern sector of Polar Bear Pass, a wetland watershed during two contrasting spring/summer seasons: 2012 – warm, early melt versus 2013 – cool, late melt; (2) quantify the seasonal water budgets for these two contrasting seasons; and (3) place these results in the context of other arctic basin studies.

2. STUDY AREA

The study took place at Polar Bear Pass (PBP) which is located in the middle of Bathurst Island (75° 40'N, 98° 30'W) during two spring/summer seasons, 2012 and 2013 (Figure 1). In previous years a range of hydrology, biogeochemistry, climatological and remote sensing studies have been conducted here (*e.g.* Young *et al.* 2010; Young & Labine, 2010; Abnizova *et al.* 2014; Muster *et al.* 2015; Young *et al.* 2016). Polar Bear Pass is a designated National Wildlife Sanctuary and Ramsar site with a watershed area of 422.1 km^2. The low-lying wetland (22.3 % of the total basin area) runs east-west and, as is typical of wetland complexes, comprises two large lakes, a mosaic of ponds and areas of wet and dry ground. Wet meadows here are

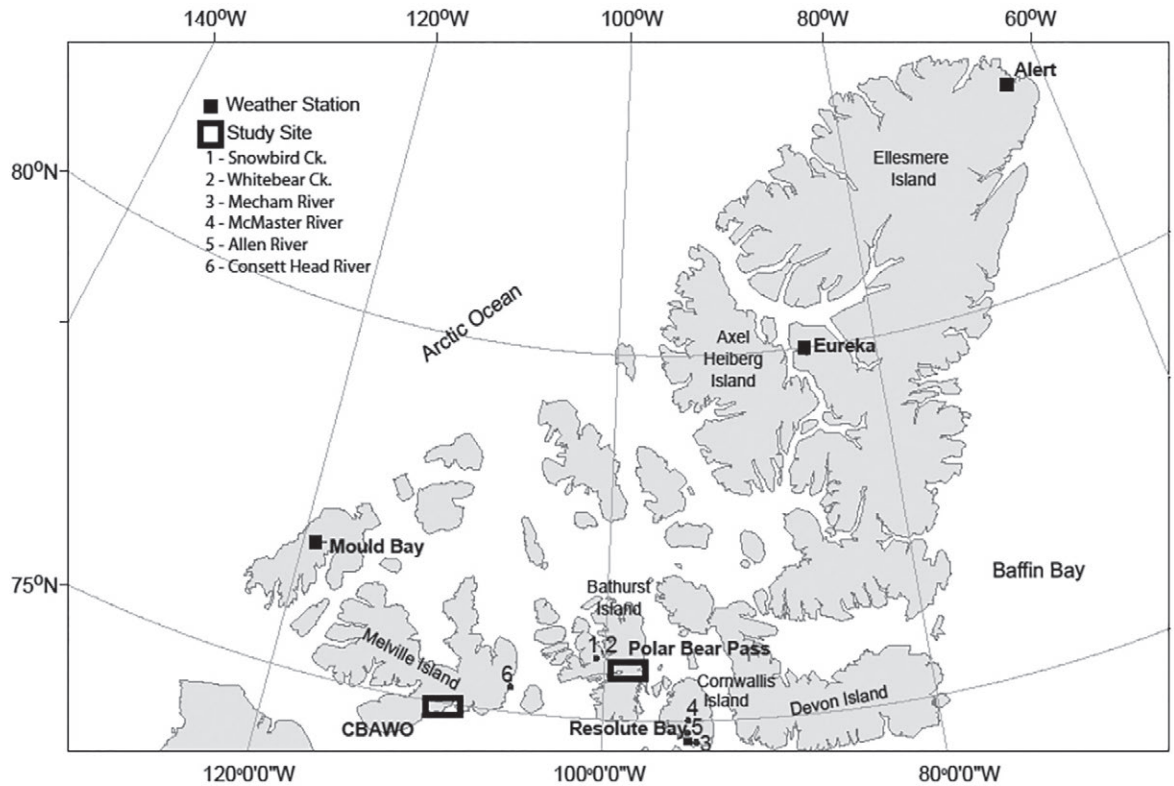

Figure 1. Location of Polar Bear Pass, Bathurst Island, Nunavut, Canada (75° 40'N, 98° 30'W). Small watersheds mentioned in the paper are also indicated and more information about these basins can be found in Young *et al.* 2015. Map modified after Young *et al.* 2015).

lush having a rich cover of grasses, sedges often with *Salix arctica* hummocks. Hillslopes and barren uplands (77.7% of the area) boarder the wetland rising from about 23 m up to about 150 m a.s.l. V-shaped stream valleys (about 50 of varying order) notch the bordering hillslopes and are effective in transferring water and nutrients to the low-lying wetland (Young *et al.* 2010; Abnizova *et al.* 2014). Late-lying snowbeds typically occur in the lee of slopes and in the incised valleys after the main snowmelt period has finished. Two large rivers cut through the wetland at its periphery. The Goodsir River, a braided gravel stream, flows from the north, transitioning into a well-defined channel as it moves east into Goodsir Inlet. It does not flood the wetland in the spring but runoff from the eastern sector of wetland drains into the Goodsir River via rivulets. Caledonian River (not named-Figure 2) is a well incised stream which flows north and then heads west where it joins the western outlet of the wetland (as marked) to empty into Bracebridge Inlet. It also does not flood the wetland during spring melt but does drain the western sector of the wetland (Figure 2). Figure 2b provides an aerial view of the wetland while Figure 2c is a photograph of the eastern sector outlet, the focus of this particular study.

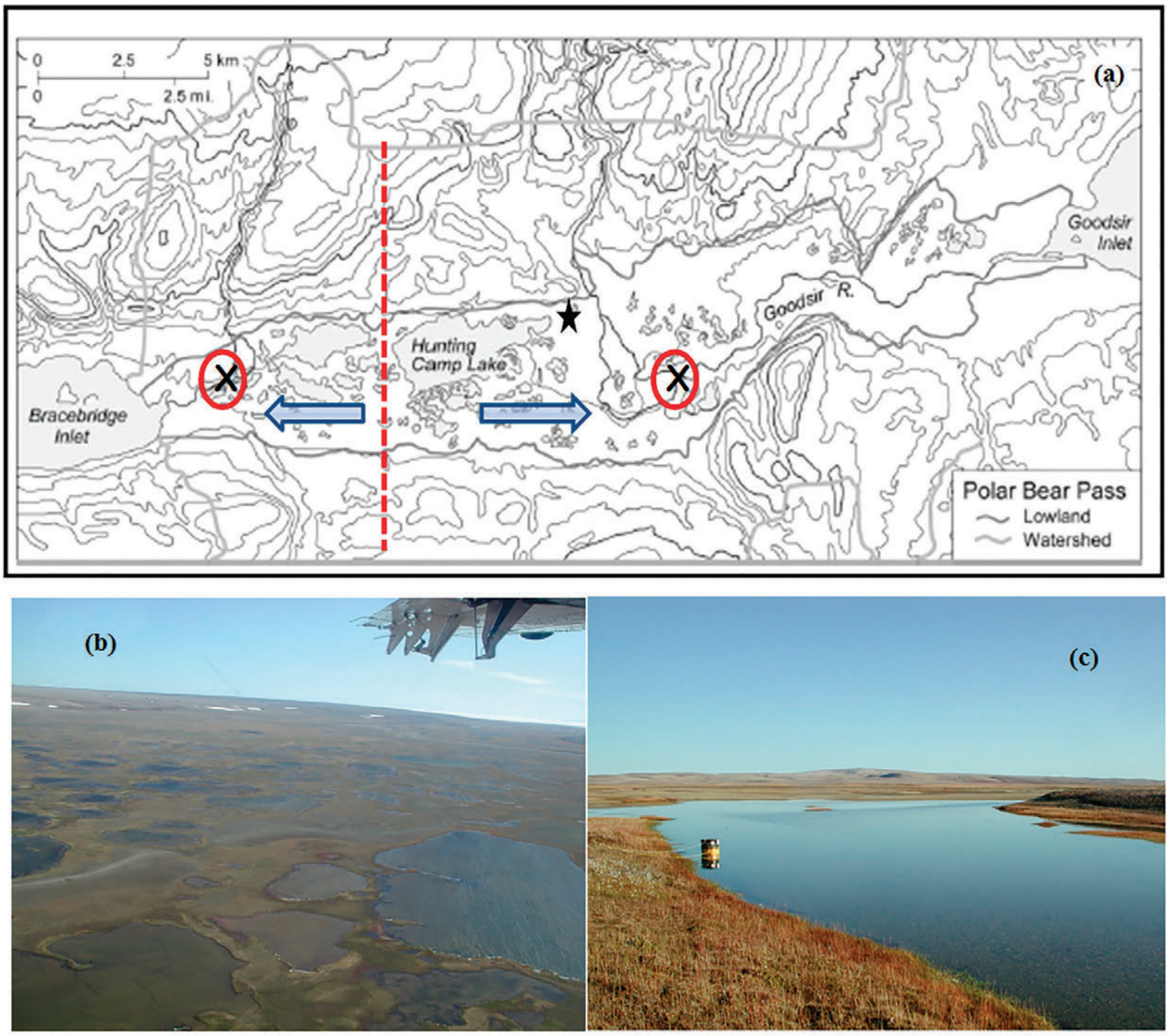

Figure 2. Detailed map of PBP study watershed (a) including wetland and upland region (source: Young *et al.* 2014). The vertical dashed line signals the 'likely' drainage divide in the watershed and the star, the wetland Automatic Weather Station (AWS). Flow direction and wetland outlets gauged are also indicated. Aerial view of the wetland (b) and photograph of the eastern outlet taken on July 4th, 2012 (c) are also provided.

3. METHODOLOGY

3.1. Snowcover

Detailed end-of winter snowcover measurements (depth, density) were carried out from mid- to late May in each year. A terrain-based snow survey of representative sites (plateau, wet meadow, pond, lake, late-lying snowbed, stream valley, *etc*) was followed (see Woo 1997). The results of the local snow survey were then up-scaled to the regional scale utilizing two topographical maps covering the area (Caledonian River, 1985, 68H/11, 1:50 000 and McDougall Sound, 1994, 68H, 1:250 000). A 100 × 100 m grid was applied over the digitized maps, and each 1000 m^2 grid cell (small to medium-sized pond) was classified as one of the main terrain types. There were 25 possible values, based on terrain type and geographical location in the eastern sector of PBP. This allowed snow survey estimates to be interpolated across this part of the watershed.

3.2. Snowmelt

Snowmelt measurements at PBP consisted of both direct measurements of surface ablation of key terrain types (*e.g.* pond, wet meadow, late-lying snowbed, plateau, *etc.*) using the approach outlined by Heron & Woo (1978) and employment of a physically-based snowmelt model (see Woo & Young 2004). Specifically, direct measurements of ablation were made daily at four sites: pond, wet meadow, late-lying snowbed and plateau. This involved measuring the distance (±5 mm) from the top of the snowpack to a stable reference point (*i.e.* a string held taut between two dowels). An average of 10 height measurements at each site were recorded daily and then averaged. Daily surface density was also recorded in order to determine surface melt in snow water equivalent units (mm/day) (see Young *et al.* 2010).

The physically-based snowcover model (Woo & Young 2004) was useful in distributing the snowcover and melt across the eastern sector of the watershed for two contrasting seasons: 2012 and 2013. Initially, this model builds up the snowcover at varying terrain units with respect to a base climate station; here, a centrally - located wet meadow site. It considers slope, aspect and lapse rates when determining initial snowcover depth and cold content. Meteorological inputs into the snow model include hourly incoming solar radiation ($K\downarrow$, W/m^2), air temperature (Tair, ^oC), relative humidity (%), precipitation (mm) and atmospheric pressure (Pa). Radiation is adjusted for aspect, temperatures and relative humidity are adjusted for elevation, and melt will only occur once the daily cold content of the snowpack is fulfilled. Adjustments can be made to the albedo algorithm for different snowpacks (seasonal vs. semi-permanent) as can the wind function (sheltered vs. exposed landscapes). A Hobo pressure transducer provided hourly air pressure data (Pa). Meteorological information for the model came from the centrally located wetland AWS and the main AWS situated on the Plateau < 1 km away (see Young & Labine 2010 for details on the instrumentation used). The model outputs daily melt (mm) for various terrain units. The snow model has shown its reliability in simulating daily snowmelt at PBP and other diverse terrain across the Canadian High Arctic, though it can over/underestimate melt for different terrain by a few days (*e.g.* Woo & Young, 2004; Young *et al*. 2013). This is a limitation shared by other snowmelt models in arctic landscapes (*e.g.* Pedersen *et al*. 2016) and can be related to errors and assumptions inherent in the snow model and fieldwork results. Young *et al.* 2013 provide a thorough assessment of the limitations of the snowmelt model used in this study.

3.3. Streamflow

Seasonal runoff from the outlet of the eastern sector watershed (see Figure 2) was measured using the mid-width velocity approach. Stage (H) was monitored by recording water level sensors with Hobo pressure transducers (± 3 mm) (Young *et al.* 2015). Frequent direct measurements of stilling well stages were made with a metric ruler (± 5 mm). They provided an additional check on the reliability of continuous stage measurements at different time intervals, and/or were used to correct stage when values drifted. Direct discharge measurements at both low and high flows were made to develop reliable stage-discharge curves for each year. Generally, due to shifting conditions in the channel two equations were developed, one for an ice-filled channel and another for an ice-free channel. These equations with an R^2 typically > 0.80 were then used to adjust the water stage into a continuous discharge record. In 2012, from Jday 168 to 170, $Q=24.60H^{3.2231}$, $R^2=0.99$, n=3; from Jday 171 to 176, $Q=10.55H^{2.6461}$, $R^2=0.99$, n=6. In 2013, from Jday177 to 183, $Q=17.096H^{3.3614}$, $R^2=0.99$, n=4; and from JD183 and onwards, $Q=12.537H^{3.0878}$, $R^2=0.85$, n=10. Errors in streamflow estimates can amount to 10 to 14% on average (Young *et al.* 2010).

4. RESULTS & DISCUSSION

General weather conditions at Polar Bear Pass (PBP) during the period from 2012 and 2013 are comparable to those of Resolute Bay confirming a polar desert climate designation (Table 1), though rainfall was much greater in 2013 at Resolute Bay than PBP. Air temperatures and wind speeds at PBP were similar to Resolute Bay in both years. Air temperatures fell above the long-term mean at Resolute Bay in 2012 and below the long-term mean in 2013. Rainfall totals at PBP were comparable between the two years: 2012 – 42 mm; 2013 – 39 mm, though the frequency distribution of the daily precipitation totals differed (data not shown). Resolute Bay recorded more rainfall than PBP in both years (2012 – 58 mm; 2013 – 72 mm) though the higher receipt in 2013 was below the long-term mean at Resolute Bay (1982-2010). Positive degree days (June, July and August-*JJA*) for PBP were 490 in 2012 and 176 in 2013. For Resolute Bay, estimates of *JJA* were 430 in 2012 and 169 in 2013. Determination of positive-degree days followed after Woo & Young (2014).

Table 1

General Climatic data at PBP Plateau versus Resolute Bay, June-August 2012, 2013. Wind speed adjusted to 1 m height, assuming a surface roughness (z_o) of 0.001 m.

		June			July			August		
	Year	Avg. Tair	PPT	U	Avg. Tair	PPT	U	Avg. Tair	PPT	U
		°C	mm	m/s	°C	mm	m/s	°C	mm	m/s
PBP	2012	3.7	0.2	4.2	8.7	27.0	4.1	3.4	14.8	3.5
	2013	-1.8	11.8	4.7	3.1	5.4	4.8	0.7	21.8	5.2
Resolute Bay	2012	3.3	12.6	4.2	7.3	28.7	3.2	3.2	17.0	4.1
	2013	-1.2	34.2	4.5	3.15	21.4	4.0	0.9	16.4	4.2
	1981-2010	0.4	14.6	4.1	4.5	28.1	4.1	2.0	33.8	4.3

4.1. End-of Winter Snowcover

Table 2 illustrates the end-of winter snowcover results for 2012 and 2013 and includes snow depth (cm), snow density (kg/m^3) and snow water equivalent (mm) for typical terrain units comprising the PBP watershed. As expected, wind-blown terrain (plateau, ponds, lakes) tend to accumulate less snow than sheltered valleys and the lee of slopes where winds are dampened. Like other environments including temperate ones, snowcover variability in terms of water equivalent units largely depends on the variation in snow depth which ranges from 63 to >800 mm in 2012 and 140 to > 600 mm in 2013. Snow density tends to be more uniform, ranging from 200 to 300 kg/m^3 for the terrain units in both years (Table 2). Higher densities occur in deep snow (valleys, lee of slopes) or shallow snowpacks (plateau) where strong winds enhance wind slab. Areally weighted snow (SWE, mm) across the eastern watershed sector was slightly higher in 2013 (81 mm) than in 2012 (72 mm) with the southern end generally capturing more snow (see Table 2). Surprisingly, the estimates for the wet meadow, ponds, lake and plateau areas (depths, density and SWE, mm) are akin to wetlands having a polar oasis-type climatic regime (Woo & Guan 2006) rather than ones influenced by a polar desert climate (Abnizova & Young 2010). Polar oasis-type regions (*e.g.* Truelove Lowland, Devon Island; Fosheim Peninsula, Ellesmere Island) are typically sheltered from poor weather and often receive lower amounts of snowfall and rainfall than polar desert landscapes. They often experience warmer springs and summers due to elevated solar radiation levels. Arctic plants still thrive in this area tapping into ground ice melt supplies during drought conditions (Edlund & Alt 1989; Edlund *et al.* 1990; Woo & Young 1997).

Table 2

Average snow survey and SWE index results for main terrain types at Polar Bear Pass.

Terrain	2012				2013			
	Snow Depth (mm)	**Snow Density (kg/m^3)**	**SWE (mm)**	**SWE index** (*relative to wet meadow*)	**Snow Depth (mm)**	**Snow Density (kg/m^3)**	**SWE (mm)**	**SWE index** (*relative to wet meadow*)
Plateau (N)	63(62)	252	16	0.50	140(99)	314	44	0.76
Plateau (S)	331(292)	313	99	3.09	286(315)	315	90	1.55
Stream Valley (N)	440(534)	282	123	3.84	559(496)	277	146	2.52
Stream Valley (S)	815(598)	252	219	6.84	610(379)	316	194	3.34
Late-lying Snowbed (N)	383(245)	228	89	2.78	454(419)	269	126	2.17
Late-lying Snowbed (S)	399(272)	242	101	3.16	312(151)	280	86	1.48
Ponds (N)	136(46)	218	31	0.97	217(59)	246	55	0.95
Ponds (S)	229(52)	273	62	1.94	304(101)	252	77	1.33
Wet Meadow	138(34)	230	32	1.00	225(50)	257	58	1.00
Polygon Area	279(50)	277	77	2.41	354(228)	262	93	1.60
Hunting Camp Lake (HCL)	189(97)	312	59	1.84	219(74)	336	73	1.26

(N/S) denotes North/South part of the Eastern Sector. Values in brackets ( ) are standard deviations of snow depth.

4.2. Snowmelt

4.2.1. Eastern watershed

Due to a deeper snowpack and a cooler spring in 2013, active snowmelt was strongly delayed in comparison to 2012 (Figure 3). In fact active melt in 2013 did not begin until well after most of the snowcover had disappeared in 2012. As observed elsewhere in the Arctic, deeper snow lingers in the lee of slopes and in the stream valleys as the shallow snowpack across uplands, wet meadows, ponds and lake disappear first. Aspect plays a defining role in Pass, as the snowpack is slower to leave in the southern part of the Pass (north-facing). This snowmelt pattern is a regular occurrence each spring (Assini & Young, 2012; Young *et al.* 2013) and in some years, it is accelerated by aeolian processes. In 2009 strong, north winds eroded the North plateau blowing sediment onto the northern half of the Pass. This dirt accelerated melt here while the southern part of the Pass remained relatively clean. This erosion/deposition pattern failed to emerge in 2010 arising from a deeper and more extensive snowpack on the upland; implying that this phenomenon only occurs in low snow years when the North plateau is relatively barren or else blown free of snow by strong winter winds. The acceleration of melt due to aeolian erosion and deposition has been found elsewhere in the Canadian High Arctic (Lewkowicz & Young 1991; Woo *et al.* 1991).

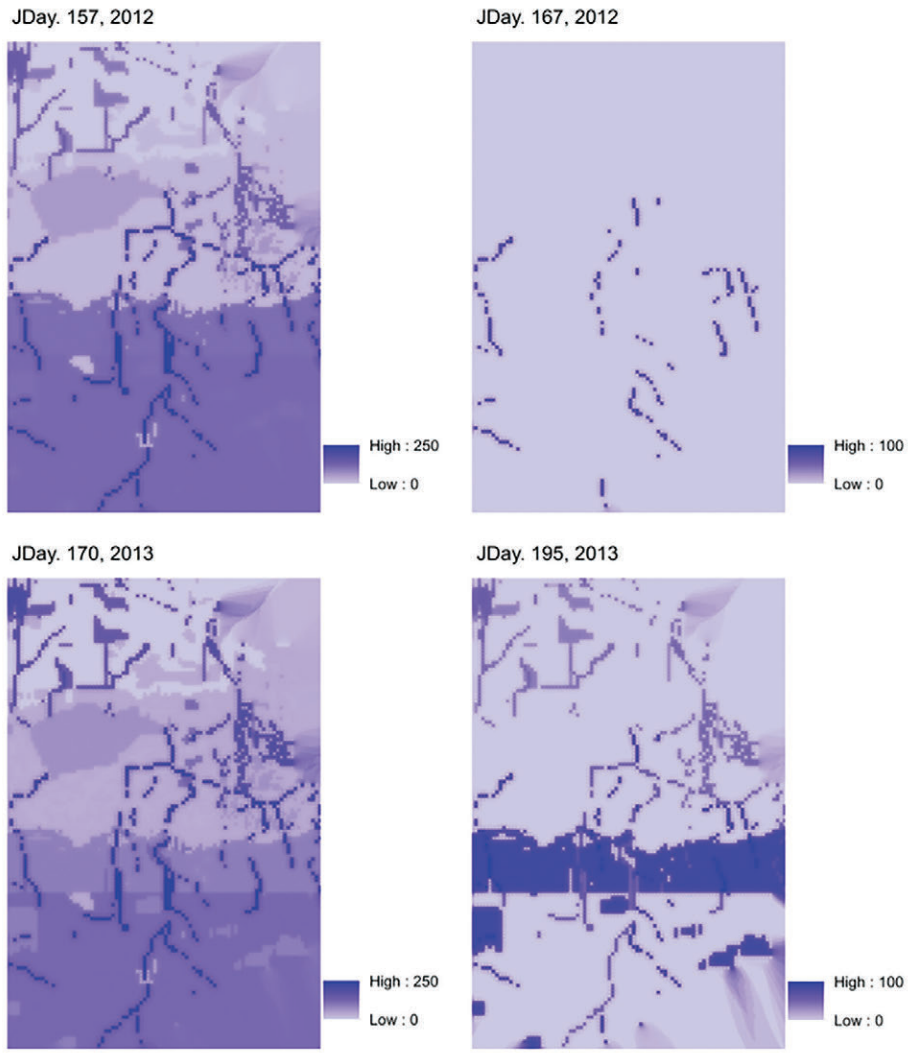

Figure 3. Modelled snowcover patterns (SWE, mm) for PBP (eastern sector) during the main snowmelt periods in 2012 and 2013. North is to the top of each image and the outline of Hunting Camp Lake (HCL) is visible in the left-hand plots.

4.3. Streamflow at PBP-2012, 2013

Streamflow out of the eastern sector of PBP was quite different in 2012 and 2013 owing to variations in snow and climatic conditions (Figure 4). Initiation of streamflow was earlier and more dramatic in 2012, with the early peaks in runoff being driven by the rapid melt-out of the northern part of the Pass (Figure 3). A series of secondary peaks occurs when the southern part of the Pass (northern aspect) started to melt out (Figure 3). By the last week in June, the outlet was already approaching baseflow conditions. In 2013 streamflow did not begin until late June; a delayed start considering that streamflow in 2012 was already at baseflow conditions by then.

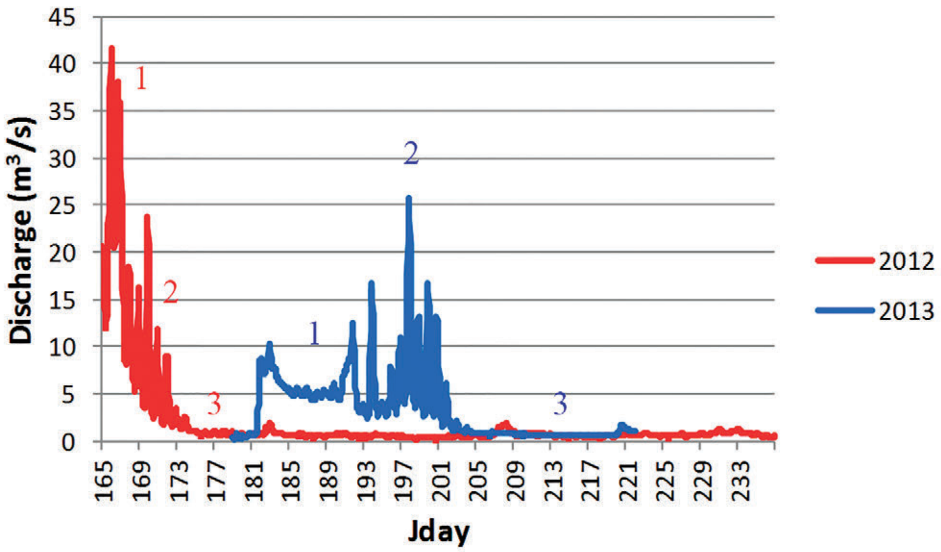

Figure 4. Streamflow hydrograph for the eastern sector outlet, PBP in 2012, 2013. The influence of snowmelt from the northern half of the Pass is designated 1, the southern half snowmelt 2, and baseflow is 3.

The 2013 streamflow hydrograph is different from 2012. The early start to the hydrograph is still being controlled by snowmelt coming from the north but the biggest peaks in mid-July and strong diurnal rhythm are indicative of active snowmelt (see Figure 3) from the southern part of the Pass. Despite the variations in timing and patterns of the hydrographs, an exceedance probability analysis of the runoff data (graphs not shown) revealed that streamflow out of the PBP eastern sector watershed is characterized by a nival-type regime. Runoff ratios differed between the two years. In 2012, the freshet Q/P=79% and the seasonal Q/P=89%, while in 2013 the ratios fell above 100% with the freshet Q/P=133%, and the seasonal Q/P lower – 112%. The 2012 ratios are typical for other high arctic catchments, and while runoff ratios can be found greater than 100% owing to deep snowbeds in stream channels or on slopes lingering from the previous year (Woo, 2012), an estimate of 133% for the freshet runoff suggests problems in our methodology. There could be an error in areal estimate of the eastern watershed or terrain units, and/or an underestimate in the snowpack. Deep snowpacks in incised valleys or hillslopes are frequently, inadequately measured (Assini & Young, 2012; Young *et al.* 2015). Assumptions concerning the distribution of the winter snowpack (SWE, mm) across terrain units may have also played a factor in the elevated runoff ratio in 2013.

4.4. Comparison of streamflow to other high arctic catchments

The early timing of streamflow and peak streamflow out of the PBP wetland (eastern sector) in 2012 was similar to nearby small hillslope catchments at PBP and to recent studies at Melville Island (East, West Rivers). Since 2007 these studies suggest an earlier start to runoff and peak response across High Arctic islands. However this pattern varied in 2013, where initiation of runoff was quite late and more akin to high arctic streamflow studies in the 70's and early 80's – a much colder period than today (see Woo & Young, 2014). These strong swings in streamflow response from an exceptionally warm spring/summer to below average cold seasons is a reflection of the streamflow intensification process, where northern Canadian basins are responding to extreme shifts in climate (temperature) and precipitation (snow or rainfall receipt), a characteristic of the climate warming signal (Déry & Woo, 2005; Déry *et* *al.* 2009; Karlson *et al.* 2015). Recent studies (*e.g.* Liljedahl *et al.* 2012; Karlson *et al.* 2015) have also raised the importance of understanding the present status and shifting conditions of landscape micro-topography (% of high or low-centered polygons) and geomorphology (*e.g.* depth of active layer thaw, thermo-erosion processes), as these factors can help modify rates of evaporation, water storage and runoff from arctic catchments.

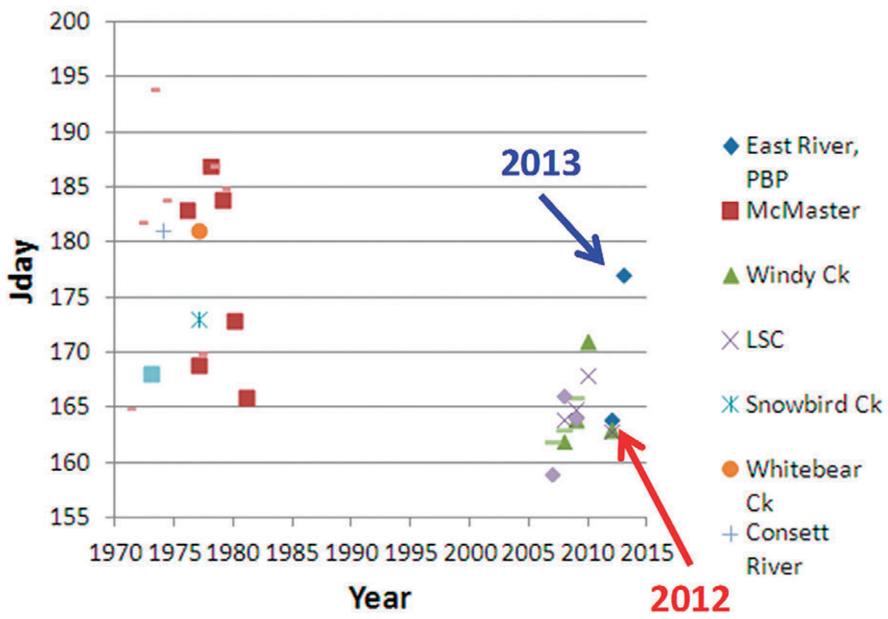

Figure 5. First day of flow for PBP from the eastern sector watershed (East River, PBP) in comparison to other catchment and watershed studies. Diagram adapted after Young *et al.* (2015). Details about watersheds other than East River, PBP can be found in Young *et al.* (2015).

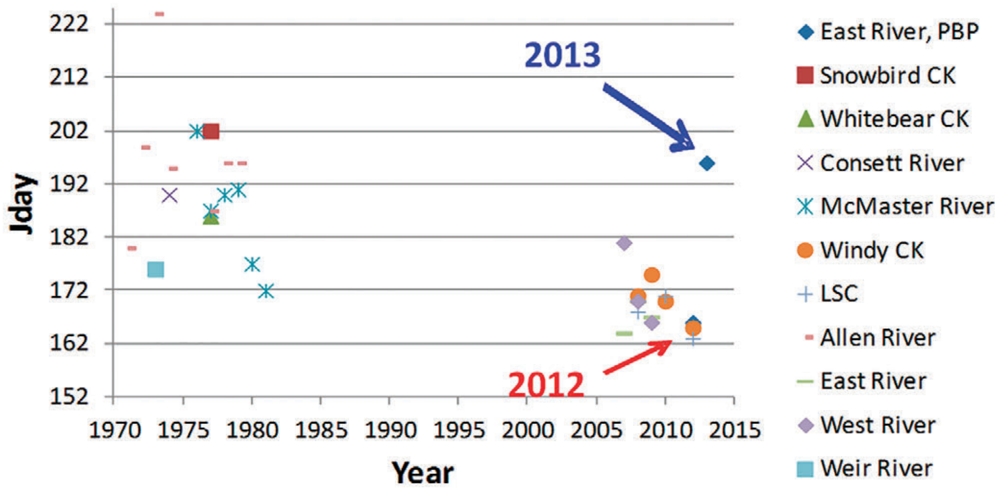

Figure 6. Day of peak discharge from the eastern PBP outlet designated here as East River versus other arctic drainage basins. Details about watersheds other than East River, PBP can be found in Young *et al.* (2015). Diagram adapted after Young *et al.* (2015).

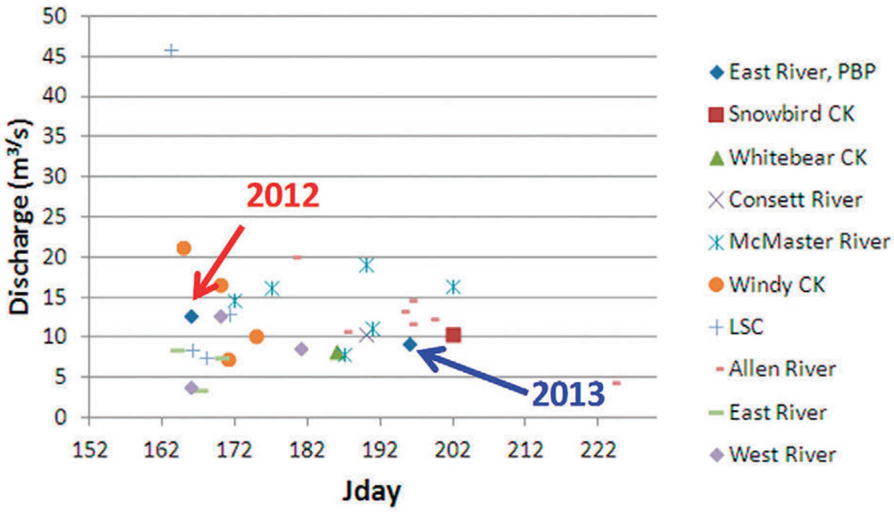

Figure 7. Peak discharge (mm/d) at the eastern outlet (East River, PBP) versus other Arctic drainage basins. Diagram adapted after Young et al. (2015). Details about watersheds other than East River, PBP can be found in Young *et al.* (2015).

4.5. Seasonal water budgets (2012, 2013)

Given the differences in climate and snowcover in 2012 and 2013, it is reasonable to expect that the seasonal water budgets would vary (see Figure 8). In response to an early and warm spring, snowmelt in 2012 began earlier than in 2013. This was followed by a longer duration for runoff, evaporation and changes in wetland storage which aided in higher evaporation losses and greater wetland storage changes in 2012 than in 2013.

As is typical of other arctic water budgets studies (see Kane and Daqing, 2004), evaporation losses can sometimes exceed precipitation inputs during warm seasons and this occurred in 2012 as well (E/P=1.12). In 2013 evaporation losses were instead met by precipitation inputs as energy receipt was diminished (E/P=0.37). The storage term was large in 2012 reaching 125 mm, almost 2.5 x greater than in 2013. Evaluation of the storage term, which is the residual in this study but also includes the error term, can be upwards of 20 to 25% for arctic basins (Young and Woo, 2004a, b).

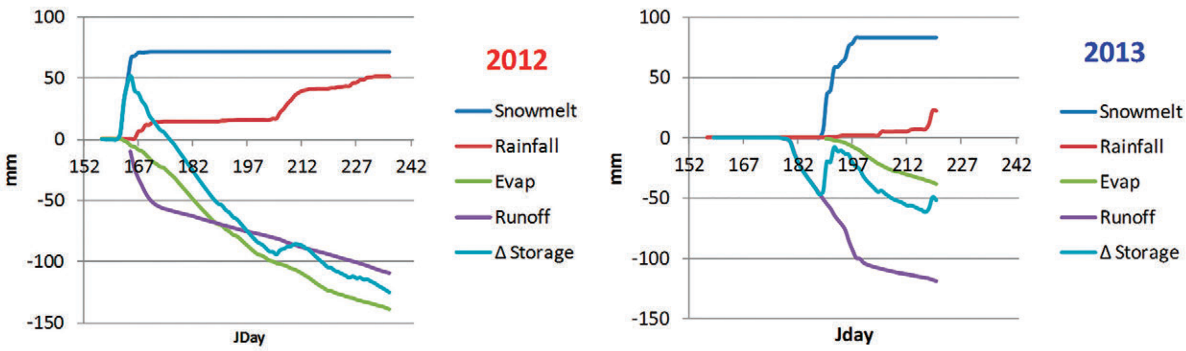

Figure 8. Seasonal cumulative water budget of PBP eastern watershed sector: 2012-Jday 157 to 236 (June 5-Aug. 23) top diagram and 2013-Jday 171-220 (June 20-Aug. 8) bottom diagram. In 2012: ∑Snowmelt=+72 mm; ∑Rainfall=+52 mm; ∑Evap=-139 mm; ∑Runoff=-109 mm and ∑ΔStorage=-125 mm. In 2013: ∑Snowmelt=+83 mm; ∑Rainfall=+23 mm; ∑Evap=-39 mm; ∑Runoff=-118 mm and ∑ΔStorage=-52 mm. Note that in 2013, runoff (measured) begins prior to basin snowmelt. This indicates an error in the snowmelt model (see text). The shorter field season in 2013 is due to arctic logistics (a lack of helicopter support at the end of August).

5. CONCLUSIONS

Like other High Arctic basins, streamflow out of the eastern sector of PBP is controlled by the seasonal snowmelt, thereby experiencing a nival regime. In 2012, due to an early and quick snowmelt season, the first and highest runoff peaks responded to snowmelt from the northern half of the Pass. In contrast, during a cold and prolonged snowmelt season, initial runoff was fed by snowmelt from the northern half of the Pass but it was not until snowmelt from the southern part reached the outlet, that the highest peaks were reached in mid-July. In fact, streamflow response in 2013 was more akin to arctic catchments in the 70's and 80's, a period which was characterized by higher precipitation and colder conditions. Subsequently, the runoff ratios and water budgets differed between the years based on the differences in precipitation and variations in climate. Given that future climate warming is expected to trigger extreme conditions (Stiegler *et al.* 2016) and the two years shown here (2012, 2013) provide a reasonable picture of this scenario, strong swings in streamflow out of large, low-gradient wetland watersheds in the High Arctic can be expected.

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Symposium Abstracts

ABOUT HYDROLOGICAL PROBLEMS OF THE LAKE BAIKAL

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In recent years in the basin of Lake Baikal and on the lake the difficult water management situation develops as a result of long lack of water and not enough reasonable restrictions for the range of fluctuations of water level. Water resources of the Lake Baikal are used for power generation and water supply therefore often emergence of the negative ecological phenomena coordinates with possible violations of the natural hydrological regime. The existing approaches to water resources management of Baikal, and the recommendation about their "greening" are considered in the article. Dates of approach of the maximum and minimum levels, amplitude of fluctuations of level, speed of filling of the lake and other parameters which essential change can influence a condition of an ecosystem of the lake are used as a set of ecological indicators (restrictions). The choice of management strategy of the water level of the Lake Baikal is executed on the basis of the analysis of results of water management calculations for ecological indicators. Results of calculation of security with water resources for water users and ecosystem requirements are presented as distributions of probabilities of the corresponding indicators of the water level fluctuations of the Lake Baikal. Conclusions are drawn on preference of one of the considered schemes of a runoff regulation.

PAN-ARCTIC MODELING OF PERMAFROST DOC AND ITS LATERAL TRANSPORT AND EVASION IN A GLOBAL LAND SURFACE MODEL

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The modelled production, transport and atmospheric release of dissolved organic carbon (DOC) from high-latitude permafrost soils into inland waters and the ocean is explicitly represented for the first time in the land surface component (ORCHIDEE) of a CMIP6 global climate model (IPSL). This work merges two models that are able to mechanistically simulate complex processes for 1) snow, ice and soil phenomena in high latitude environments, and 2) DOC production and lateral transport through soils and the river network, respectively, at 0.5° to 2° resolution. We present results for the Pan-Arctic and Eurasia, with a focus on the Lena River basin, and show that soil DOC concentrations, their riverine transport and atmospheric evasion are reasonably well represented as compared to observed stocks, fluxes and their respective seasonality. Surface warming caused by anthropogenic climate change can be reasonably expected to destabilize permfrost stores via microbial or hydrological mobilization following thaw, and as the permafrost line migrates pole-ward over time (Schuur et al., 2015). Most global climate models do not represent the unique permafrost soil environment and its respective processes. This significantly contributes to uncertainty in estimating their responses, and that of the planet at large, to warming. The riverine component of this 'boundless carbon cycle' is likewise little-recognised in global climate modelling (Regnier et al., 2013). Hydrological mobilisation to the river network results either in sedimentary settling or atmospheric 'evasion', the latter amounting to 480-850TgC yr^{-1} globally (Lauerwald et al., 2015). Potential feedbacks owing to such a response are of particular relevance, given the magnitude of the permafrost carbon pool. Our work aims at filling in these knowledge gaps, and the response of these DOC-related processes to thermal forcing.

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SEDIMENT FLUXES FROM THE OUTER AND INTERIOR SHORES OF THE LENA DELTA

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Coastal and riverbank erosion studies in the Lena Delta were conducted at more than 60 key sites. The results obtained indicate that average erosion rates are 0.7 to 1.5 m/yr for the coastline and 1 to 3 m/yr for the actively eroding riverbanks. The magnitudes of coastal erosion and riverbank erosion are comparable, with riverbanks eroding slightly faster. Extremely high erosional activity along the distributaries is observed during the flood period. The highest erosion rates of up to 15 m/yr are observed where the major distributary branches off near Gogolevsky Island. Coastal change in the Lena Delta region has been studied better than riverbank erosion.

The study has also found that erosion of the coastline, 1900 km in length within the Lena Delta region, supplies 6.5×10^6 t/yr of sediments, including 67×10^3 t/yr of organic carbon into the Laptev Sea shelf. Sediment inputs derived from riverbank erosion were studied only at 42 sites located in the central and eastern sectors of the delta, totaling 72.7 km in length. The amount of sediment released to the distributary channels from the studied bank stretches was estimated to be about 2.1×10^6 t/yr, including almost 60×10^3 t/yr of organic carbon.

The study has shown that riverbank erosion discharges large amounts of sediment, part of which remains in the delta and with the remaining part deposited in the underwater delta and further delivered to the Laptev Sea. Unfortunately, the lack of reliable field data prevents quantification of the contribution of riverbank erosion to the sediment input to the sea. It is evident, however, that riverbank erosion along the distributary channels should be taken into account when estimating the total annual terrigenous flux to the Laptev Sea basin. This flux is composed of sediment inputs of the large rivers (28.6×10^6 t/yr), seacoasts (62.2×10^6 t/yr) and some minor sources, such as eolian material, bed load and ice-transported load. The study of riverbank erosion in the Lena Delta region will help generate more precise estimates of the inputs of terrigenous sediment, including organic carbon, to the Laptev Sea basin and the Arctic basin on the whole.

HYDROLOGICAL FORECASTING IN PERMAFROST DOMINATED CATCHMENTS – IMPACTS OF PERMAFROST PROCESS REPRESENTATION, FORECAST DATA UNCERTAINTY AND SPATIAL SCALE

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Climate change impact on hydrological regime in Arctic rivers and fresh-water inflow to the Arctic Ocean is evident from observations showing an increase in river discharges over the latest decades. Permafrost is existing in large part of the Arctic ocean drainage basin, and it is of highest interest to assess the impact of thawing permafrost on the observed changes. Processes of soil freezing and thawing, seasonal development of active layer and talik zones are controlling surface and subsurface water flowpaths, horizontal and vertical connectivity, and fluxes, and climate driven thawing of permafrost is expected to change the hydrological response in many regions. Similar to the long-term changes, it is interesting to assess to impact of permafrost process representation on the skill of hydrological forecast models used for instance for flood warning and hydropower management. The current trend in hydrological forecast research is on assessing the relative importance of uncertainties in initial states and meteorological forecasts used to force the hydrological model. The objective of this study is to include also the representation of permafrost processes, spatial scale and landscape properties in the assessment of forecast skill. We will use data from five river basins in Central Yakutia, Eastern Siberia of varying size (170 to 65400 km^2) and landscape characteristics. In particular, the study areas have differences in permafrost features such as thermokarst lakes and ground ice, and large differences in observed runoff behaviour. We will compare short-term and seasonal hydrological forecasts using two hydrological models of different spatial and process representation: The Hydrological Predictions for the Environment (HYPE) model and the Hydrograph model. Both models have been setup for the Lena River basin in previous project, but with very different approaches. The HYPE model is a semi-distributed model without permafrost processes (reduced infiltration capacity in frozen soil is the exception). The pan-arctic application Arctic-HYPE will be used, which is based on globally available data and calibrated with river discharge observations from the entire Arctic region without special attention to performance in the Lena River basin. The Hydrograph model is a distributed process-based model, representing the permafrost processes in much more detail. Model parameters are adjusted on the slope scale and are not calibrated on runoff data. In addition to assess role of initial hydrological state and meteorological conditions, we will test the hypotheses that the permafrost process representation is more important for the river discharge forecasts on longer lead times and smaller spatial scales, whereas initial conditions and overall water balance would be dominating the forecast skill on shorter times and larger scales, respectively.

SEEMINGLY SIMILAR PERMAFROST RIVER BASINS: DATA ANALYSIS AND EVALUATION BY DIFFERENT MODELLING APPROACHES

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The basis of the hydrological modelling in ungauged basin is transfer of algorithms and parameters from similar studied basin. It is assumed that catchments characterized by similar climate, geology, surface conditions and proximity generate similar streamflow. Modelling performance for such basins is expected to be of the same level of agreement. The goal of the study was to analyse long-term runoff series for five river basins in Eastern Siberia and compare two different modelling approaches for runoff simulation. Studied rivers are tributaries of the Lena and Aldan Rivers in Central Yakutia. They are characterized by cold and dry climate with mean annual air temperature and precipitation varying from -7.7 to -11.6°C and from 240 to 450 mm accordingly. The potential evaporation measured at the Yakutsk meteorological station averaged 360 mm over the period 1993–2008. Predominant landscapes are larch forests, pine forests and tussock bogs. There is continuous permafrost zone with active layer depth from 0.5 to 3 m depending on surface conditions. Areas of studied basins vary between 3 380 and 65 400 km^2. Additionally we used data from small research watershed (area 170 km^2) near Yakutsk. Preliminary data analysis showed runoff increase in last decades in response to air temperature growth. Precipitation doesn't show any significant trend. Although precipitation within the studied basins varies only in two times the runoff depth differs in two orders from 1 to 100 mm/year. Our hypothesis is that such permafrost features as thermokarst lakes and ground ice could significantly influence runoff generation. Such features are not accounted for in traditional hydrological modelling approaches. The Taatta River basin with the lowest runoff depth (1 mm/year) has more than 4 000 small thermokarst lakes that are associated with ground ice. Lakes could contribute much water to evaporation and not feed river. To evaluate runoff generation processes in studied basin and speculate about possible reasons of large runoff difference two modelling approaches were applied. Hydrograph model is a distributed process-based model, explicitly representing the soil thawing and freezing. Model parameters are adjusted on the slope scale and are not calibrated on runoff data. Opposite to the bottom-up approach of the Hydrograph model the Hydrological Predictions for the Environement (HYPE) model was used. The HYPE model is a semi-distributed model without permafrost processes (reduced infiltration capacity in frozen soil is the exception). The pan-arctic application Arctic-HYPE will be used, which is based on globally available data and calibrated with river discharge observations from the entire Arctic region without special attention to performance in the Lena River basin.

MELTING OF ICE IN LAKES: MEASUREMENTS AND MODELLING

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Melting of lake ice is a complex process driven primarily by solar radiation. Ice melts at the boundaries and in the interior, with fractions depending on the solar radiation balance and surface heat balance. Complications arise from large variability of optical properties of ice in the melting season. To gain more understanding of the melting process, field experiments have been performed in Finnish lakes from the boreal zone to Arctic tundra. The heat budget in the melting season is dominated by the radiation balance with positive albedo feedback, while turbulent heat fluxes are normally small but occasionally can be large. In dry climate, sublimation can be significant in the ice mass budget. Strong solar radiation leads to internal melting, and under the ice water warms up resulting in convective mixing. Internal melting leads to weakening and breakage of the ice cover that further speeds up the decay of ice. Based on the collected field data, development work of ice cover decay model is ongoing.

MOUNTAIN GLACIER CONTRIBUTION TO ARCTIC AND SUBARCTIC RIVER DISCHARGE

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Arctic and subarctic mountain glacier coverage has decreased up to 70% in recent decades, while discharge has increased in most rivers. Still, less is known about the contribution of glacier change on high-latitude river flows and long-term trends in river discharge. Here we combine long-term analyses in river discharge and changes in glacier coverage within Russian, Canadian and Alaska Arctic and subarctic watersheds. We hypothesize that in the semi-arid subarctic and Arctic climate, the increasing air temperatures and limited commensurate increase in precipitation, has allowed glaciers to serve as contributors to the observed increases in river discharge. We calculated excess discharge, which is the glacier-derived runoff that is due to net glacier mass loss (negative volume change), and normalized excess discharge to watershed area. Preliminary results suggest that, of the large Russian watersheds (Ob, Yenisey, Lena, Indigirka, and Kolyma), which glacier coverage range from 0.0001 (Kolyma) to 0.042% (Indigirka), the excess runoff ranged from 0.01 (Yenisey) to 5.1 mm/yr (Indigirka). The limited glacier area in Kolyma River watershed (<1 km^2) limited the amount of excess discharge despite the largest relative glacier coverage reduction (<70%). If assuming a 12% glacier area decrease, which was observed in the Tanana River basin (7.3% glacier coverage), the excess runoff ranged from 1.7 (Mackenzie River, 0.103%) to 36.4 mm/yr (Yukon River, 1.278%) in the large North American rivers. The estimates of excess runoff can be further refined by more effective quantification of the watershed boundaries, which is where the mountain glaciers are located. The initial results suggest that the excess runoff exceed the observed increase in annual flow in some rivers, which may indicate that glacier melt is masking a decrease in another source. In other rivers, the change in annual runoff is a magnitude larger than the excess runoff suggesting that another source primarily explains the increased river discharge. Further, the relative change in winter river discharge is positively correlated to recent watershed glacier coverage if less than 0.1%. Changes in discharge in watersheds with glacier cover above ~0.1% all show positive changes in winter discharge. Our analysis emphasizes the importance to consider high-latitude mountain glaciers, albeit their small coverage, as potential contributors or modifiers to the documented changes in Arctic and subarctic river discharge.

THE INFLUENCE OF CHANGES OF ACTIVE LAYER THICKNESS ON THE RUNOFF GENERATION IN A PERMAFROST CATCHMENT

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Changes in active layer thickness over Arctic and permafrost regions have important impact on rainfall-runoff transformation. General warming is observed across Svalbard Archipelago and corresponds to the increases of ground temperatures. Permafrost thaw and changes in active layer thickness due to climate warming alter how water is routed and stored in catchments, and thus impact both surface and subsurface processes. The overall aim of the present study is to examine relationships between temporal changes of active layer depth and hydrological model parameters together with variation in the catchment response (observed and simulated runoff). The analysis is carried out for small unglaciated catchment Fuglebekken, located in the vicinity of the Polish Polar Station Hornsund on Spitsbergen. For hydrological modelling the conceptual rainfall-runoff HBV model is used. The model is calibrated and validated on flows within subperiods. A moving window approach (two weeks long) is applied to derive temporal variation of parameters. Model calibration together with an estimation of parametric uncertainty was carried out using the SCEM-UA algorithm (Shuffled Complex Evolution Metropolis). This allowed the dependence of HBV model parameters on active layer depth to be analysed. In addition we tested an influence of parametric uncertainty of the estimated correlations by a comparison of the results obtained for the Pearson correlation coefficient and weighted Pearson correlation coefficient.

ICE REGIME OF ESTONIAN RIVERS AS AN INDICATOR OF CLIMATE CHANGE

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The climate of Estonia have significantly changed over the last century. Year mean air temperature has increased by 1.0-1.5 ^{0}C (J. Jaagus 2016). Estonian rivers are small and therefore very sensitive to any impacts, especially changes of ice regime. Characteristics of ice phenomena as dates of ice-on and ice-off, durations of ice events, maximum ice cover thickness are very important for many sectors of water industry.

Studies have shown that climate change has an impact on ice events formation in rivers.

Observation data series of fifteen hydrological stations the periods 1920-2014 have been used for the analysis of ice regime characteristics that include the dates of ice freeze-up, ice-breakup and ice cover duration. Analysis of maximum ice cover thickness covers the period 1946-2013. The analysis of changes in ice cover shows systematic values variations depending on the regions of Estonia. Average dates of ice-on of the Northern part of Estonian rivers moved to 3-4 days later and dates of ice-off appeared to be earlier in 1-2 days. Dates of ice-on of the West-Southern parts of Estonian rivers were 4-5 days later and dates of ice-off appeared to be earlier in 2-3 days. Thus the duration of ice cover was decreased by 6-8 days. The trend of maximum annual ice thickness was decreasing, by 2-5 cm of average for whole rivers, expect at some hydrological stations in Northern karstic area where trends were increased or not changed. The study results show that ice cover duration and maximum ice thickness of Estonian rivers is decreased that is in agreement with analogical results obtained within the Nordic countries studies (BACC II).

SUB-SEA PERMAFROST MODELLING: 1D VERTICAL EXTENT, SEDIMENT REACTION- TRANSPORT SCHEME AND *IN SITU* GHG FORMATION

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The role that submarine permafrost may play within global carbon cycle and its possible feedback effects on climate system in perspective of current global warming change represent a challenge for the research of the field. Subsea permafrost is in fact thought to store a large amount of carbon (~ 1400 PgC) but it is also believed to determine the physical conditions for the formation and the preservation of gas hydrates as well as to act as a cap preventing upward seepage of gas coming from beneath. Due to the paucity of measurements very little is known about subsea permafrost but it is expected to be particularly susceptible to changes and a comprehensive carbon budget is lacking.

As a first step to fill this gap we have modified the terrestrial permafrost module already present in the land-surface model JSBACH to account for a submarine environment: we developed a 1D vertical scheme, ran an offline spin-up phase using MPI-ESM climatology for pre-industrial state and then we simulated the submergence imposing the uppermost boundary conditions (temperature and salinity) taken from output of MPI-OM and ran it for a centennial time scale to check for permafrost thermal degradation. To account for carbon soil accumulation in subaerial condition an already existing Yedoma model will be employed jointly with bioenergetic reaction- transport model taylored for high-latitudes shelf environment to account for the period after sea transgression. The latter can possibly be refined to consider carbon (gas, DIC and DOC) exchange with overlaying water column.

The outcomes of the simulations highlight the prominent role of the first phases after sea transgression (characterized by a higher thawing rate) in determining subsea permafrost thermal profile. This stresses the importance of focusing on submergence details and time sequence in order to get plausible model outcomes. Freshwater influxes (riverine and groundwater – measured or modelled within MPI-ESM or other ESMs) and their influence on subsea permafrost condition and carbon release may also be considered in future developments The flooding phase at early Holocene was associated with drastic permafrost changes (thermokarst formation, change of hydrological connectivity, Yedoma degradation and coastal collapse) which should also be taken into account. The importance of instantaneous soil response after sea level rise is also remarked in a series of lab experiments we performed, where ice or a frozen sand sample was overlayered by warmer salty water, in order to i) measure its degrading rate, ii) test a conceptual model (based on Stephan's problem) and iii) simulate a frozen soil degradation.

TERRESTRIAL ENVIRONMENTAL OBSERVING NETWORK (TEON) – THE KUPARUK RIVER, ALASKA'S ARCTIC, USA

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There are two major difficulties that are widely acknowledged when the Arctic observing systems are being discussed: 1) sparse observational network and 2) changes in the observational network over time. This presentation provides overview of the long-term hydrologic monitoring network in the Kuparuk River and adjacent watersheds. This network has been maintained by the Water and Environmental Research Center (WERC) at University of Alaska Fairbanks led by Dr. Douglas Kane and funded by numerous research projects. The data collection was initiated in 1985 in the small Imnavait Creek watershed just north of the Brooks Range. Over the years, data collection extended to include the Upper Kuparuk River watershed in the early 1990s, the entire Kuparuk River watershed in the late 1990s and then the adjacent watersheds in 2000s. As of today, observational network includes the Kuparuk River, Putuligayuk River and Roche Mountannee Creek watersheds to obtain continuous hydro-climatological data streams for a new program funded by the Arctic Landscape Conservation Cooperative (U.S. Fish and Wildlife Service). This newly directed observation effort, the Terrestrial Environmental Observing Network (TEON), joins the legacy of hydro-climatic monitoring and research in the Kuparuk with new permafrost and vegetation observation programs to start building integrated datasets to benefit a variety of Arctic Alaska stakeholders.

IMPACTS OF SOIL ORGANIC MATTER ON PERMAFROST SOIL THERMODYNAMICS BY THE ORCHIDEE LAND SURFACE MODEL

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Soils in the northern permafrost regions contain large quantities of organic carbon, formed under cold climates with limited decomposition; meanwhile, soil organic matter (SOM) significantly modifies soil thermal and hydraulic properties, and in turn affects soil thermo- and hydro-dynamics, vegetation growth, and soil carbon accumulation. Typically, the presence of SOM, with its lower soil thermal conductivity and higher heat capacity, modulates the heat transfer from the soil surface along the depth and leads to a cooler soil temperature during summer than without organic matter (e.g. Lawrence and Slater, 2008; Decharme et al., 2016). SOM also alters hydraulic properties towards higher soil porosity, thus higher saturated hydraulic conductivity and higher plant-available water capacity (e.g. Huntington, 1994; Morris et al., 2015). Both processes may enhance soil organic carbon accumulation and therefore form a positive feedback loop.

In this study, we incorporate the effects of SOM on soil thermal and hydraulic properties in the ORCHIDEE land surface model. ORCHIDEE computes the principal soil–atmosphere–vegetation energy and water exchange processes in 30 min time steps. The ORCHIDEE high-latitude version used here includes vertically resolved soil carbon and cryoturbative mixing (Koven et al., 2009), a scheme describing soil freezing-thawing and its effect on soil thermal and hydrological dynamics (Gouttevin et al., 2012), and a multi-layer snow scheme (Wang et al., 2013). We show that the inclusion of SOM effects significantly decreases modeled summer soil temperature and active layer thickness in permafrost regions, which is in better agreement with site observations compared to without the SOM effects. Consequently, modeled total soil organic carbon in northern permafrost region increase by 27%, from 800 PgC to 1016 PgC. We further test the hypothesis that the potential loss of permafrost organic carbon due to future climate change could be accelerated through the SOM-thermodynamics coupling, a pathway generally missing in current model projections.

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